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The Carboniferous-Permian Transition in Central New Mexico

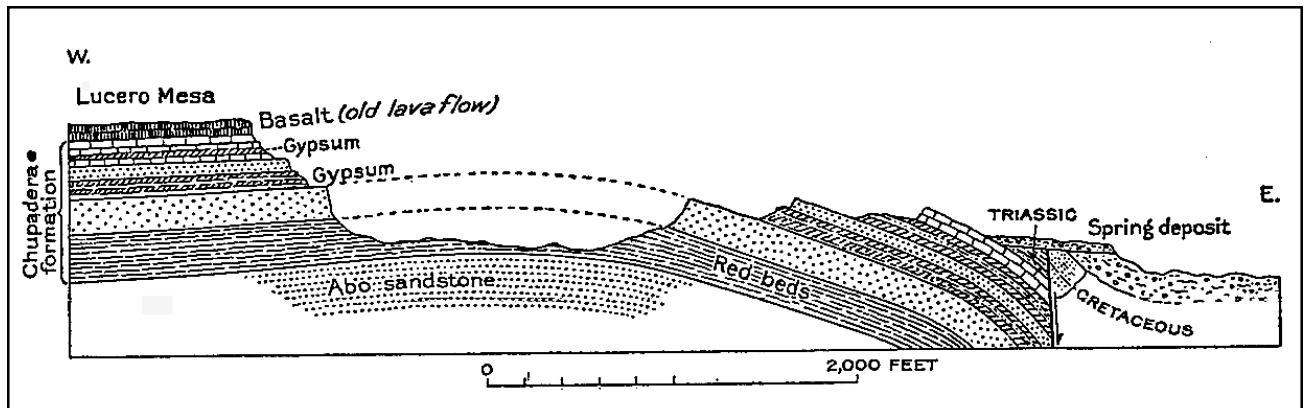
NMMNH&S Bulletin 59

Bulletin 59

New Mexico Museum of Natural History & Science

A Division of the
DEPARTMENT OF CULTURAL AFFAIRS

The Carboniferous-Permian Transition in Central New Mexico



Edited by
Spencer G. Lucas, W. John Nelson, William A. DiMichele,
Justin A. Spielmann, Karl Krainer, James E. Barrick,
Scott Elrick and Sebastian Voigt

Albuquerque, 2013

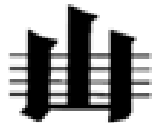
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**IUGS
Subcommission on Carboniferous Stratigraphy
Subcommission on Permian Stratigraphy**

New Mexico Museum of Natural History & Science

Albuquerque, 2013

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Cover illustration: Geological cross section from Mesa Lucero on west through Carrizo Arroyo on east, Valencia County, New Mexico. From Darton (1928, U. S. Geological Survey Bulletin 794). Note that Darton's "Abo sandstone" includes strata of the Red Tanks Member of the Bursum Formation, which crosses the Carboniferous-Permian boundary. Darton's "Chupadera formation" encompasses strata now referred to the Lower Permian Yeso Group, Glorieta Sandstone and San Andres Formation.

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THE CARBONIFEROUS-PERMIAN TRANSITION

MAY 20-22, 2013

An international meeting devoted to all aspects of Carboniferous-Permian geology with special emphasis on the Carboniferous-Permian transition.

Hosted by the New Mexico Museum of Natural History and Science, Albuquerque, New Mexico, USA

Organizing Committee: Spencer G. Lucas (Albuquerque), James E. Barrick (Lubbock), Vladimir Davydov (Boise), William DiMichele (Washington, D. C.), Karl Krainer (Innsbruck), John Nelson (Champaign), Joerg W. Schneider (Freiberg) and Sebastian Voigt (Thallichtenberg)

Schedule:

19 May

Pre-meeting fieldtrip to the Carboniferous-Permian transition section at Carrizo Arroyo, central New Mexico (limited to 25 participants).

20-22 May

Talks and posters.

21 May

Afternoon fieldtrip to Late Pennsylvanian Kinney Brick Quarry.

23-25 May

Post-meeting fieldtrip to Pennsylvanian-Permian rocks exposed in Joyita Hills-Cerros de Amado east of Socorro, New Mexico

Fieldtrips:

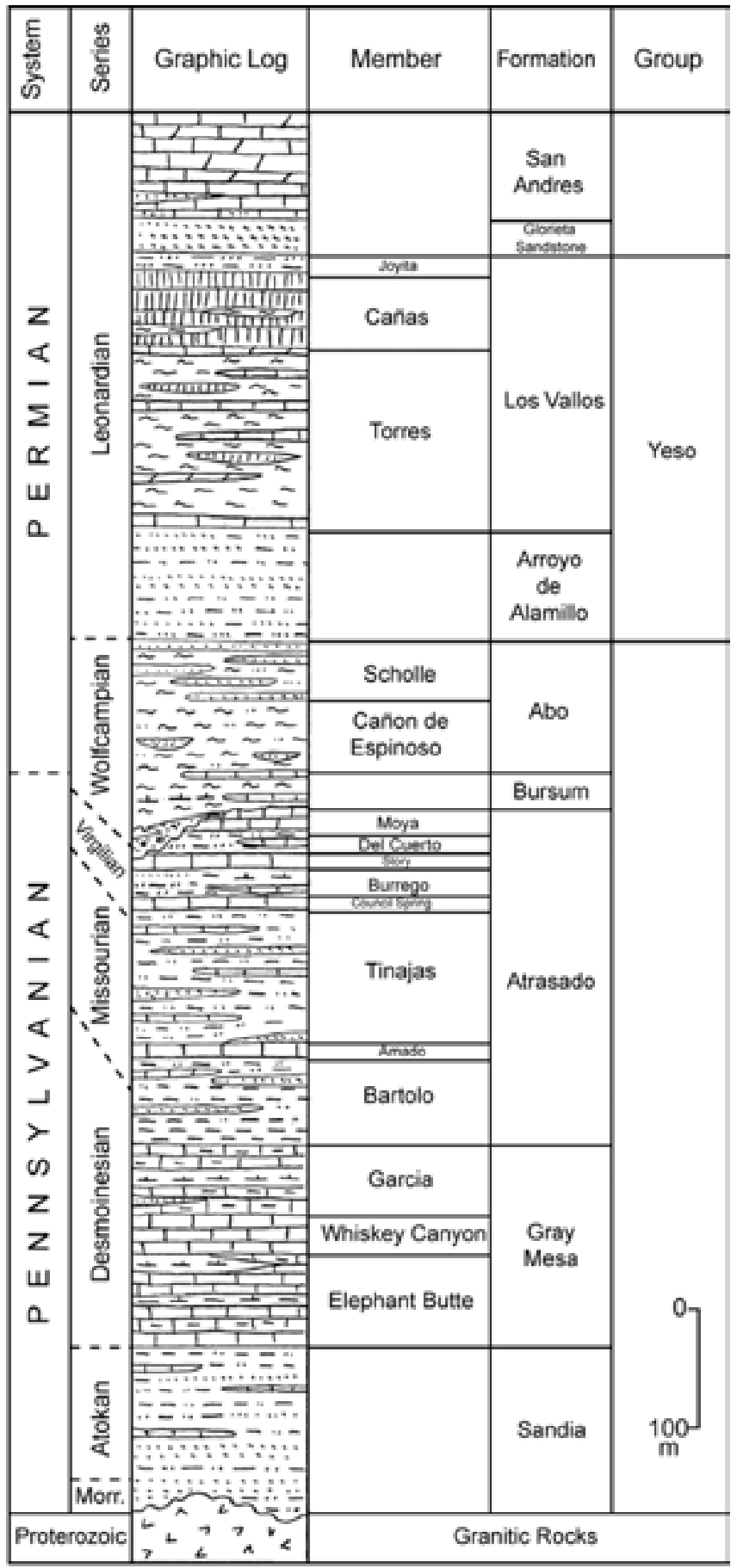
Trip 1: Carrizo Arroyo is one of the most paleontologically diverse localities across the Carboniferous-Permian boundary. It exposes mixed marine and non-marine strata of the Bursum Formation that yield everything from plants and insects to fusulinids and brachiopods. This section plays a key role in global marine/non-marine correlations because of the co-occurrence of conodonts and insect-zone species. Access is difficult, by 4-wheel-drive vehicle over difficult roads, so the number of participants is limited to 25 persons.

Trip 2: The Kinney Brick quarry is a world class Late Pennsylvanian Lagerstätte, located just east of Albuquerque. It is also important for marine/non-marine correlations due to the occurrence of conodonts, fusulinid, branchiosaur and insect zone species. All participants will take an afternoon excursion to the quarry as a break in the meeting technical program

Trip 3: East of Socorro, marine and nonmarine sedimentary rocks of Middle Pennsylvanian-Early Permian age are exposed along the eastern margin of the Rio Grande rift. This is one of the best exposed and most studied Pennsylvanian-Permian sections in New Mexico, and recent work has brought forth diverse paleofloras, detailed conodont biostratigraphy, extensive ichnofossil assemblages, and much more. The three-day trip, headquartered in Socorro, will work through this entire section, focusing on issues of stratigraphy, sedimentation and paleontology.

Symposium proceedings:

Proceedings of the symposium and a field guide will be published by the New Mexico Museum of Natural History and Science. Contributions on all aspects of Carboniferous and Permian geology are appropriate for the proceedings. Contributions to the proceedings can range from abstracts to full length articles.



ROAD LOG FROM ALBUQUERQUE TO UPPER PENNSYLVANIAN STRATA IN THE KINNEY BRICK QUARRY, MANZANITA MOUNTAINS, NEW MEXICO

SPENCER G. LUCAS¹, BRUCE ALLEN², KARL KRAINER³ AND JUSTIN A. SPIELMANN¹

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SUMMARY

This trip is to the Kinney Brick Quarry, a world-famous Late Pennsylvanian Lagerstätte in the Manzanita Mountains, just east of Albuquerque, New Mexico.

Mileage Description

- | | | |
|--|--|--|
| <p>0.0 Parking lot, east exit, New Mexico Museum of Natural History and Science. Turn right to proceed south on 18th Street. 0.1</p> <p>0.1 Traffic light at Mountain Road. Turn right to go west on Mountain Road. 0.3</p> <p>0.4 Traffic light at Rio Grande Boulevard. Turn right to go north on Rio Grande Boulevard. 0.3</p> <p>0.7 I-40 on ramp to right. Enter I-40 and go 2-3 lanes to the left. 2.1</p> <p>2.8 I-25 interchange, go straight. 7.2</p> <p>10.0 Pass under Juan Tabo Boulevard bridge. Continue east on I-40. 1.0</p> <p>11.0 Tramway Boulevard exit 167 to right; continue east on I-40. 1.1</p> <p>12.1 Sandia Granite here and in roadcuts for next few miles. This Proterozoic granite (ca. 1.4 Ga) forms the core of the Sandia Mountains. 0.4</p> <p>12.5 Carnuel Exit, number 170. Continue east on I-40. 0.3</p> <p>12.8 Tall roadcut in Sandia Granite to left. Just ahead, I-40 crosses under old US-66, which ran from Chicago, Illinois, to Santa Monica, California, crossing 8 states and 3 time zones--the country's major east-west highway until the 1970s. See Connolly et al. (1982) and Lucas et al. (1999) for detailed road logs of the route from Albuquerque to Tijeras following US-66. 1.0</p> <p>13.8 Cross Tijeras Arroyo. 1.2</p> <p>15.0 Proterozoic (ca. 1.65 Ga) Cibola Granite/Gneiss in road cuts on right here and for next 0.2 mi. After rounding the big turn in the highway, roadcuts across the highway to the left are also Cibola Gneiss. The large, north-northeast trending ridge to the right is capped by Pennsylvanian-age strata, underlain by the Proterozoic Tijeras Greenstone. The two Proterozoic terranes are separated by the Tijeras fault (which follows the highway in this stretch of I-40), a locus of recurrent movement during much of the geological record of this area (Karlstrom et al., 1999). The Tijeras fault zone trends northeast-southwest, and continues northeastward from I-40 just before the next big turn in the highway ahead. 2.3</p> | <p>17.3 To left (after rounding turn in the highway) are roadcuts in I-40 and a frontage road above where an excellent cyclic sequence of Pennsylvanian strata, and a diverse, mollusc-rich fossil locality, were formerly exposed. About 20 years ago these roadcuts were covered with gunnite (a kind of liquid concrete) by the New Mexico State Highway Department, presumably to protect the highway from falling rocks. For a photo and description of these roadcuts before they were "improved," see Connolly and Kues (1987). 0.4</p> <p>17.7 Exit 175 to Tijeras on right; leave I-40 at this exit and proceed toward southbound NM-337 (bear right after entering off ramp). 0.4</p> <p>18.1 Four-way intersection with traffic light in Tijeras. Continue straight through intersection, heading south on NM-337; cross Tijeras Arroyo. 0.3</p> <p>18.4 On right, entrance to GCC Cement Company plant. Pennsylvanian limestone has been mined here since 1956, when the plant was first built by the Ideal Cement Co. In 1994, Grupo Cemento de Chihuahua became the plant's parent company. About 120 employees work at the plant during full production operations, 8-10 of which drill and shoot limestone in the large open-pit quarry that adjoins the plant. Throughout the 1990s and early 2000s, approximately 500,000 short tons of cement were produced here annually with a gross value of about \$40-50 million. As of 2012, current employees number approximately 70 and current production is about half of capacity due to relatively low market demand. The cement from this plant is mostly sold in New Mexico and southwestern Colorado. The roadlog route now will proceed up the winding course of Cedro Canyon. 0.1</p> <p>18.5 Entering Cibola National Forest; Sandia Ranger Station (with a visitor center and walking tour of pre-historic Tijeras Pueblo) a short distance ahead on the left. We will be driving downsection through the Pennsylvanian stratigraphic sequence for the next few miles. Roadcuts on right are mixed terrestrial-marine deposits of the Pennsylvanian Bursum Formation, which overlie predominantly marine deposits of the Pennsylvanian Atrasado Formation (Fig. 1). 0.2</p> <p>18.7 Roadcut to right exposes limestone beds of the Moya Member of the Atrasado Formation. Pennsylvanian stratigraphic nomenclature employed in the Manzanita Mountains (Fig. 2) is that of Lucas et al. (2009, 2011), which replaces that of Myers (1973). The oldest Pennsylvanian strata here are olive-drab micaceous sandstones, siltstones, and mudstones, with a few marine limestones, assigned to the Sandia Formation, of Atokan age. Thickness of the Sandia Formation ranges from only a few m to over 120 m in the Manzanita-Sandia Mountains (Krainer et al., 2011), but we will not be encountering the Sandia Formation on this field trip.</p> | <p>The overlying remainder (majority) of the Pennsylvanian section consists of the Gray Mesa Formation (mostly</p> |
|--|--|--|

Sandia Ranger Station Roadcut

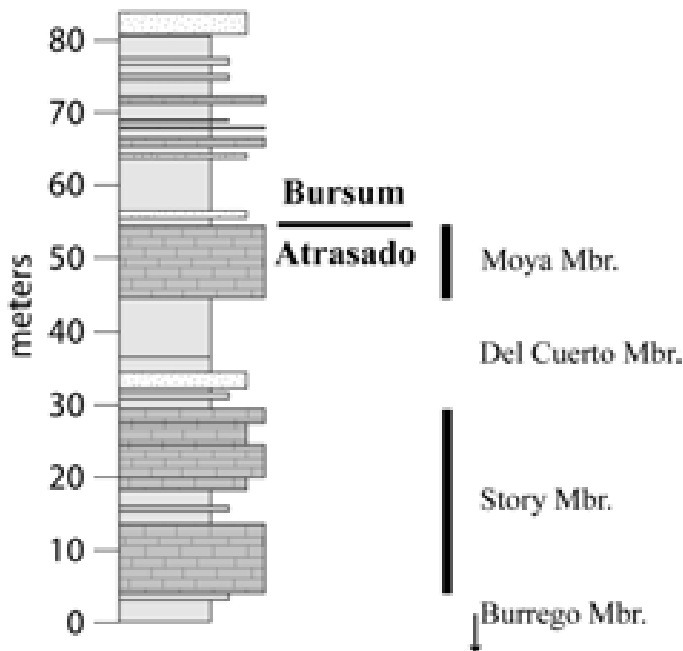


FIGURE 1. Measured stratigraphic section of the upper part of the Atrasado Formation and lower part of the Bursum Formation across the highway from the Sandia Ranger Station.

Desmoinesian in age, based on fusulinids; about 120 m thick on Cedro Peak to the southeast: Fig. 3), overlain by the Atrasado Formation (Missourian-Virgilian age; about 200 m thick on Cedro Peak: Fig. 3), and finally the Bursum Formation (Virgilian to lower Wolfcampian; up to about 90 m thick along I-40 northeast of here). We will be seeing roadcuts and exposures of the Atrasado and Gray Mesa formations as we drive southward from here to the Kinney Brick Quarry. **0.1**

18.8 Roadcut to right exposes limestone beds of the Story Member of the Atrasado Formation.

Our stratigraphic studies in central New Mexico over the past 15 years have rekindled the mid-20th century effort by M.L. Thompson (1942) to provide a detailed stratigraphic framework and nomenclature for Carboniferous rocks in the region (e.g., Lucas et al., 2009, 2011, 2012; Lucas and Krainer 2009, 2010). These early attempts by Thompson were largely ignored by U. S. Geological Survey and other geologists, resulting in a crude lumping of stratigraphic units on the one hand, and a plethora of names for correlative units in neighboring mountain blocks in central New Mexico on the other hand. Ongoing work in the Manzano-Manzanita-Sandia Mountains (e.g., Lucas and Krainer, 2010; Lucas et al., 2011; Vachard et al., 2012, 2013) has demonstrated that a member-level lithostratigraphic subdivision of the Pennsylvanian System is indeed applicable over a broad area of central New Mexico. Thus, the Atrasado Formation in the Manzanita Mountains can be divided into eight members (Fig. 2) that are recognizable in mountain uplifts from here to the vicinity of Truth or Consequences over 200 km to the south. The upper four members of the Atrasado Formation (Moya, Del Cuerto, Story and Burrego members, in descending order: Fig. 2) are present in the roadcuts between here and the (southern) turnoff to Forest Road 462 ahead. A number of small-offset faults and a broad monoclinial flexure are visible in these

roadcuts. Overall, however, the strata are dipping toward the northwest, and the southward winding road-log route is proceeding gradually downward through the stratigraphic section. **0.2**

19.0 Enter private land, with several mountain homes. Limestones exposed in roadcuts are still in the upper part of the Atrasado Formation. **0.4**

19.4 To the left, Forest Road 462 leads to trails popular with off-road bicycling enthusiasts. Cedro Peak (elevation 2373 m, 7785 ft) with radio towers on its summit, is visible at 10:00, about 2.9 km in the distance.

In the Manzano Mountains, because of structural complications and forested cover, only in a few local areas can the entire Pennsylvanian section be examined (Myers, 1973, 1982). One such area is Cedro Peak, a conical mountain near the northern end of the Manzano uplift, in the part of the range usually termed the Manzanita Mountains. Here, a Pennsylvanian section about 340 m thick is preserved with only minor fault interruptions, and this section is relatively well exposed along the steep flanks of Cedro Peak (Fig. 3). We assign this section to the (ascending order) Sandia, Gray Mesa (= Los Moyos), Atrasado (= Wild Cow) and Bursum formations. We divide the Gray Mesa Formation into the (ascending order) Elephant Butte, Whiskey Canyon and Garcia members, and we divide the Atrasado Formation into the (ascending order) Bartolo, Amado, Tinajas, Council Spring, Burrego, Story, Del Cuerto and Moya members. We thus reject the names Sol se Mete, Pine Shadow and La Casa for member-level subdivisions of the Atrasado Formation.

At Cedro Peak, the ~ 14 m thick Sandia Formation is almost entirely of nonmarine origin and is assigned an Atokan age based on regional correlations. The ~ 120 m thick Gray Mesa Formation records normal marine deposition. It contains fusulinids of latest Atokan? to Desmoinesian age. The ~ 200 m thick Atrasado Formation is a complex succession of marine and nonmarine (mostly fluvio-deltaic) strata. It contains fusulinids of Missourian and Virgilian age. Only the lowermost 6 m of the Bursum Formation are exposed at Cedro Peak, but nearby sections indicate a Bursum thickness of ~ 90 m and yield Virgilian-age fusulinids from a limestone bed near its base. **0.2**

19.6 Roadcuts in Council Spring Member of the Atrasado Formation for next ~0.3 km (with minor fault repetition). The lower four members of the Atrasado Formation are, in descending order, the Council Spring, Tinajas, Amado and Bartolo members (Fig. 2), which we will traverse over the next 1.1 km. **0.5**

20.1 Limestone cliff on right is Amado Member of the Atrasado Formation, with covered slope of Bartolo Member below. A down-to-the-northwest fault zone is present ahead; road cut exposures to the south of that are in Gray Mesa Formation. **0.4**

20.5 Roadcuts are now in the upper part of the Gray Mesa Formation. **0.2**

20.7 To right, mouth of Tunnel Canyon. Just beyond, in limestone roadcut, is an outcrop that yielded an upper Desmoinesian marine invertebrate fauna, within the upper part of the Gray Mesa Formation, described by Kues and Koubek (1991). The fauna, from a thin calcareous shale unit, is dominated by brachiopods (especially *Phricodothyris perplexa* and *Hustedia mormoni*) and bryozoans, with fewer fusulinids, solitary rugose corals, gastropods, trilobites, and partial crinoid

Marcou (1858)	Herrick (1900)	Gordon (1907)	Read et al. (1944)		Myers & McKay (1976)		this paper														
New Red Sandstone	Manzano Series	Manzano Group	Abo Formation		Abo Formation	Wolfcampian	Abo Formation		Wolfcampian	Perm											
Mountain Limestone	Coal Measures	Magdalena Group	Madera Formation	Madera Limestone	arkosic limestone member	Wild Cow Formation	La Casa Member	Bursum Formation	Virgilian	Bursum Formation	Permian										
												gray limestone member	Sol Se Mete Member	Atrasado Formation	Moya Member	Virgilian					
															Del Cuerto Member						
															Story Member						
												Sandia Series	Sandia Fm.	Sandia Formation	upper clastic member	Los Moyos Limestone	Desmoinesian	Desmoinesian	Gray Mesa Formation	Burrego Member	Missourian
																				Council Spring Member	
																				Tinajas Member	
												Sandia Formation	Sandia Formation	lower limestone member			Atokan	Atokan	Sandia Formation	Amado Member	Atokan
																				Bartolo Member	
																				Coyote Sa Bed	
								Garcia Member													
								Whiskey Canyon Member													
								Elephant Butte Member													

FIGURE 2. Development of Pennsylvanian lithostratigraphic nomenclature in the Manzano uplift.

calyxes. **0.5**

21.2 Large Gray Mesa Formation limestone cliff on left. To right, note old road bed with low limestone cliffs and recent rock slide debris. **0.1**

21.3 Thick, well-exposed section of Gray Mesa Formation within a large natural amphitheater to the left, east of the valley floor. Old road bed west of the highway leads to trilobite locality, marked by a talus pile just beyond a medium-sized tree on the road bed.

This is the Cedro Canyon trilobite locality/section of Szabo (1953), which was subsequently described by Kues (1982). Although trilobites were common members of early Paleozoic mud-bottom marine communities, by the late Paleozoic (Carboniferous-Permian), trilobites were not common, so this Pennsylvanian Cedro Canyon locality, where trilobite fossils abound, is an unusual occurrence.

The trilobites at this locality, *Ditomopyge scitula* (Meek and Worthen), occur in large numbers in a 0.2-m-thick interval of relatively soft, dark gray shale in the Gray Mesa Formation, together with abundant and diverse brachiopods and bryozoans, and rarer bivalves, and solitary rugose corals. Kues (1982) interpreted the facies as representing deposition in normal marine, moderately deep water, many km offshore. Most of the fossils represent epifaunal filter feeders.

The abundant trilobites are anomalous here; they are typically absent to minor elements of Pennsylvanian marine assemblages elsewhere in New Mexico. The concentration of trilobites here reflects a favorable microhabitat, but the parameters of this microhabitat are difficult to identify.

Although this locality has been collected for decades, fragments to complete specimens of the trilobites can still be found, although the producing horizon is mostly covered by talus now. **0.3**

21.6 Small quarry in Gray Mesa Formation to left. **1.6**

23.2 Pass junction with Forest Road 242 (Juan Tomas Road, leading to Cedro Campground) to left; continue on NM-337. Highway has crossed the down-to-the-east Cedro fault a short distance back; route will now begin to climb upward through the top of the Gray Mesa Formation and back into the overlying Atrasado Formation. **0.4**

23.6 To left, a long roadcut composed mainly of shaley beds, with limestone beds at the top. These exposures are near the contact between the Gray Mesa and Atrasado formations. A number of relatively small-offset faults shift the boundary between the Gray Mesa and Atrasado up and down by several meters in adjacent fault blocks along this stretch of the highway.

At the southern end of this road cut is a unit of fissile, soft, micaceous greenish-gray shale, near the base of the Atrasado, which contains a sparse assemblage of marine fossils dominated by the chonetoid brachiopods *Mesolobus* and *Chonetinella*, and diverse bivalves, but also including sparse bryozoans, gastropods, ammonoids, nautiloids and echinoderms (Kues, 1983). Also present are well-preserved specimens of cephalopod aptychi, which are seldom preserved opercula or mouth parts of cephalopods. **0.4**

24.0 To right is El Cedro Road, leading to small village of Cedro,

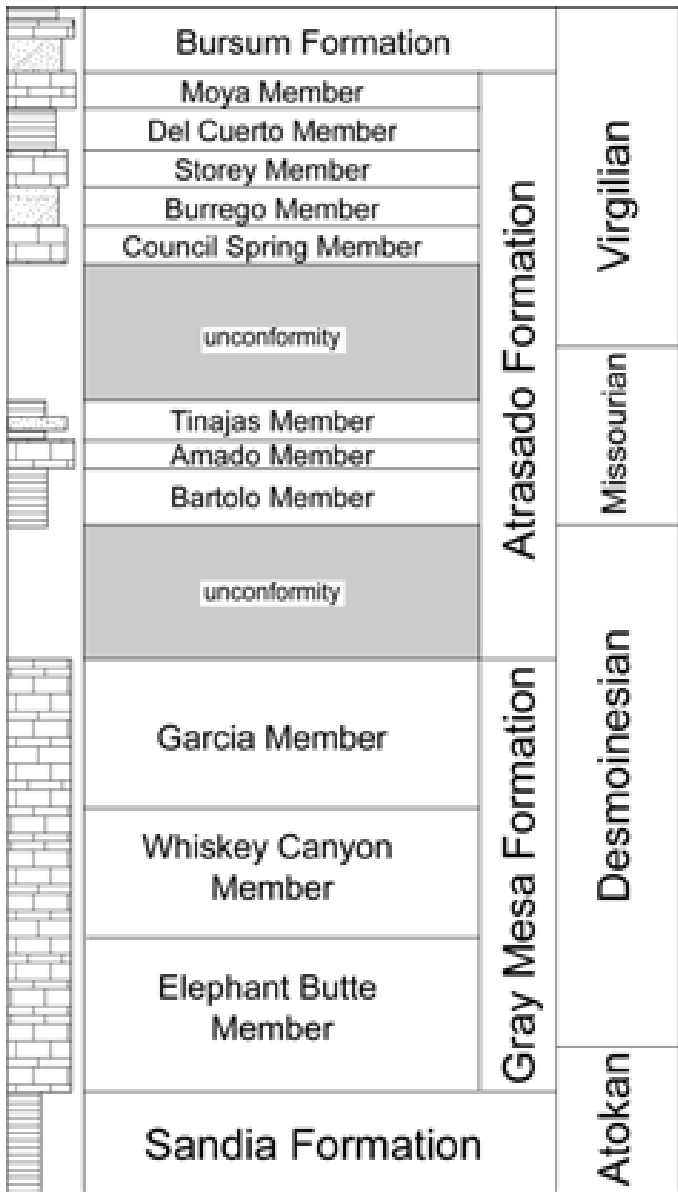


FIGURE 3. Summary of the Pennsylvanian stratigraphy and biostratigraphy at Cedro Peak.

visible in the hills near NM-337. The long roadcuts to the left display small-offset faults and thin, gray, micritic, locally concretionary limestone beds, alternating with dark-gray shale and calcareous shale interbeds near the boundary between the uppermost Gray Mesa Formation and basal Atrasado Formation. **0.1**

24.1 In lower part of roadcut to left note fault and a thin (0-1 m thick) stringer of gray-green shale sharply bounded above and below by thick limestones. The shale thickens gradually to almost 3 m at the south end of the road cut, and represents the deposition of a slug of siliciclastic mud in the prevailing carbonate environment. The shale unit contains a diverse marine assemblage dominated by brachiopods (e.g., *Mesolobus*, *Neochonetes*, *Orbiculoidea*, *Derbyia*), bivalves (*Euchondria* and other pectinids, *Pernophorus*, *Parallelodon*) and fenestrate bryozoans (Lucas et al., 1999). **0.2**

24.3 Ramblewood Blvd. and mailboxes to right. **0.2**

24.5 To left in roadcut are dark green shales and a sandstone bed of the basal Atrasado Formation (Bartolo Member). Highway has now climbed out of the Gray Mesa Formation for the remainder of the roadlog. **0.5**

25.0 On left is entrance to Pine Flat picnic area (entrance by vehicles requires a reservation from the Sandia Ranger Station). No outcrops are exposed along the highway here, but the long, gentle rise to the south is held up by the Amado Member of the Atrasado Formation. Exposures of the Amado Member can be seen along the drainage walls of Cedro Canyon in the northern part of the Pine Flat picnic area. **0.7**

25.7 Entrada de Cibola Road to right. Highway is now in the basal Tinajas Member of the Atrasado Formation, which contains a marker bed of fusulinid-bearing limestone indicating a Missourian age that can be traced throughout the Manzanita Mountains. **0.2**

25.9 **Turn right on unmarked gravel road**, the entrance to the Kinney Brick Quarry. **0.1**

26.0 On left is the old quarry pit where the first paleontological collections were made at the quarry. **0.1**

26.1 **STOP 1 is at the active pit of the Kinney quarry.** The Kinney Brick Quarry in the Manzanita Mountains of central New Mexico is a world-famous locality for Late Pennsylvanian fossil plants, invertebrates and vertebrates. The quarry is actively mined for clay utilized to manufacture bricks by the Kinney Brick Company in Albuquerque, New Mexico.

A classic Konservat Lagerstätte, Kinney preserves soft tissues and other delicate structures of plants and animals not well known from correlative deposits (Lucas and Huber, 1991; Kues and Lucas, 1992). Fossils documented from the Kinney Brick Quarry are palynomorphs (Willard, 1992), a diverse, conifer-rich megaf flora (Mamay, 1981, 1990; Ash and Tidwell, 1982; Mamay and Mapes, 1992), a shelly marine invertebrate assemblage that includes a few ammonoids but is dominated by brachiopods and the pectinacean bivalve *Dunbarella* (Clark, 1978; Archer and Clark, 1992; Kues, 1992a, b; Mapes and Boardman, 1992), syncarid and hoplocarid crustaceans (Schram and Schram, 1979), eurypterids (Kues, 1985), conchostracans (Kozur et al., 1992), ostracods (Kietzke and Kaesler, 1992), terrestrial arthropods, mostly diplopods and insects (Carpenter, 1970; Shear et al., 1992), conodonts (Krukowski, 1992; Lucas et al., 2011), a diverse assemblage of fishes, mostly acanthodians and palaeoniscoids (Zidek, 1975, 1992b; Gottfried, 1987a, b, 1992; Bardack, 1992; Huber, 1992; Schultze, 1992) and amphibians (Berman, 1973; Hunt et al., 1992, 1996; Wernneburg and Lucas, 2007), as well as coprolites (Hunt, 1992; Hunt et al., 2012) and “fish eggs” (Mamay, 1994). The depositional setting of Kinney is interpreted to be that of an estuary fed by a river delta (Feldman et al., 1992; Lorenz et al., 1992).

The Kinney Brick Quarry is a clay pit developed in Missourian strata of the Tinajas Member of the Atrasado Formation (Fig. 4). Strata exposed here provide a unique glimpse of a diverse and well-preserved Late Pennsylvanian lagoonal biota, and the quarry stratigraphy and paleontology was the subject of a symposium volume about 20 years ago (Zidek, 1992). Students at the University of New Mexico discovered the rich fossil biota of the Kinney quarry in 1961. During the 1960s, large collections were made of fishes, plants and other fossils from the quarry, especially by D. Dunkle and S. Mamay of the National Museum of Natural History. Extensive fossil collections from the Kinney Brick Quarry are now housed at the Carnegie Museum of Natural History,

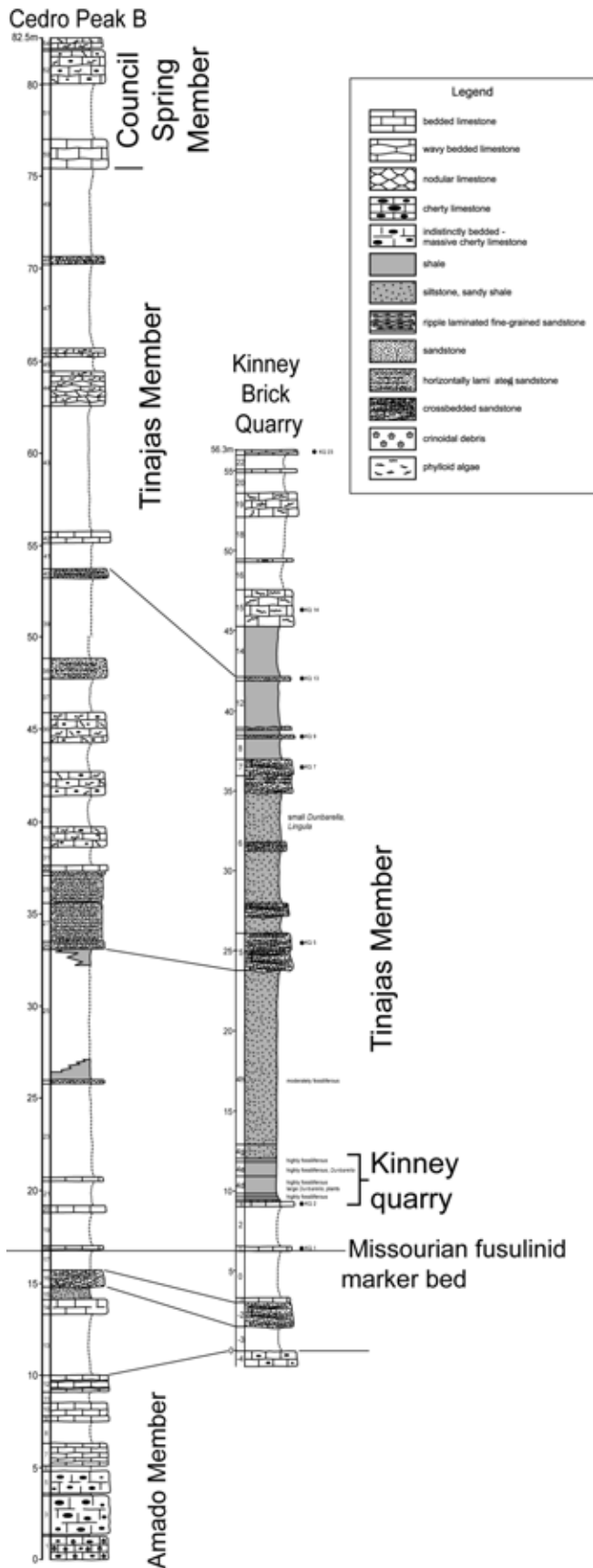


FIGURE 4. Measured stratigraphic section at Kinney Brick Quarry correlated with part of the Cedro Peak section, showing Missourian fusulinid level (after Lucas et al., 2011).

National Museum of Natural History, New Mexico Museum of Natural History (NMMNH), University of Kansas Museum of Natural History, and University of New Mexico.

The quarrying operation has exposed about 30 m of the upper part of the Tinajas Member of the Atrasado Formation (Figs. 4-5). The active quarry floor typically is developed on black, platy and laminar micrite with a high content of black clay containing an invertebrate fauna indicative of abnormal marine conditions (Fig. 5, units 1-2). The upper 10 cm of this limestone are finely laminated and grade into the overlying shale. The basal shale exposed at the quarry is 0.4 m thick, olive gray to olive black, well indurated, calcareous and highly fossiliferous. Part of this interval is highly fissile and not calcareous. Above it is 2 m of olive-gray, calcareous shale that contains thin bands of plastic clay. Individual laminae in these shales are continuous on strike for at least 10 m. Above, and in gradational contact with the underlying shale, are ~ 11 m of olive-gray and greenish-gray, laminar calcareous silty shale that are moderately bioturbated. These strata coarsen upward slightly and are marked by a rapid decrease in floral and faunal diversity near their base. Above them are ~ 13 m of olive-gray, silty shale and claystone interbedded with lenticular, yellowish-orange and grayish-brown, subarkosic and sublitharenitic, laminated and ripple-laminated sandstone ledges. The section is completed by 1.0+ m of limestone-cobble conglomerate with clasts as much as 8 cm in diameter, and grayish-yellow, trough-crossbedded subarkosic-sublithic arenite.

Some early workers (e.g., Stucky, 1967; Berman, 1973) assigned a Permian age to the Kinney fossils, based primarily on an unsubstantiated report of the "Permian index fossil" *Callipteris* (cf. Read and Mamay, 1964) from the quarry (Kelley and Northrop, 1975). In the 1990s (see articles in Zidek, 1992), most workers considered the quarry to be Virgilian in age, based on the incorrect stratigraphic placement of the quarry in the geologic mapping of Myers and McKay (1976), but not based on any direct fossil evidence at the quarry. However, fusulinids of the Tinajas Member from stratigraphically just below the quarry (Fig. 4) include *Triticites* species indicative of the Missourian, and conodonts from the quarry also indicate a Missourian age (Lucas et al., 2011). Therefore, the Missourian age of the Kinney fossils is now well established.

Combining physical stratigraphy with the fossil assemblages, we can interpret local sea level trends and identify several distinct depositional environments at the Kinney Brick Quarry (Fig. 5). These make up a regressive sequence in which limestone grades up through prodelta and deltaic clastics with a capping delta-plain facies.

Thus, the basal micrite (Fig. 5, units 1-2) represents deposition in a nearshore marine environment that received some input of freshwater and clastic sediments. Note its lithology (especially the high black-clay content of this micrite) and unusual fauna (some stenohaline brachiopods and other groups, but dominated by euryhaline taxa, such as the inarticulate brachiopod *Lingula* and bivalves *Myalina* and *Solemya*). The black-clay content, terrestrial plant debris, and euryhaline elements of the fauna (especially abundant *Lingula*) are consistent with deposition near the shoreline with a significant freshwater input.

The overlying highly fossiliferous shales (Fig. 5, units 3-4) were deposited in a calm lagoonal or estuarine environment with a significant fresh-water input. Uniform, fine grain size, fine lamination and lack of bioturbation, dark colors, and preservation of soft-bodied forms suggest deposition in quiet, oxygen-poor waters with restricted circulation. A lowland, pteridosperm-dominated flora and freshwater faunal elements

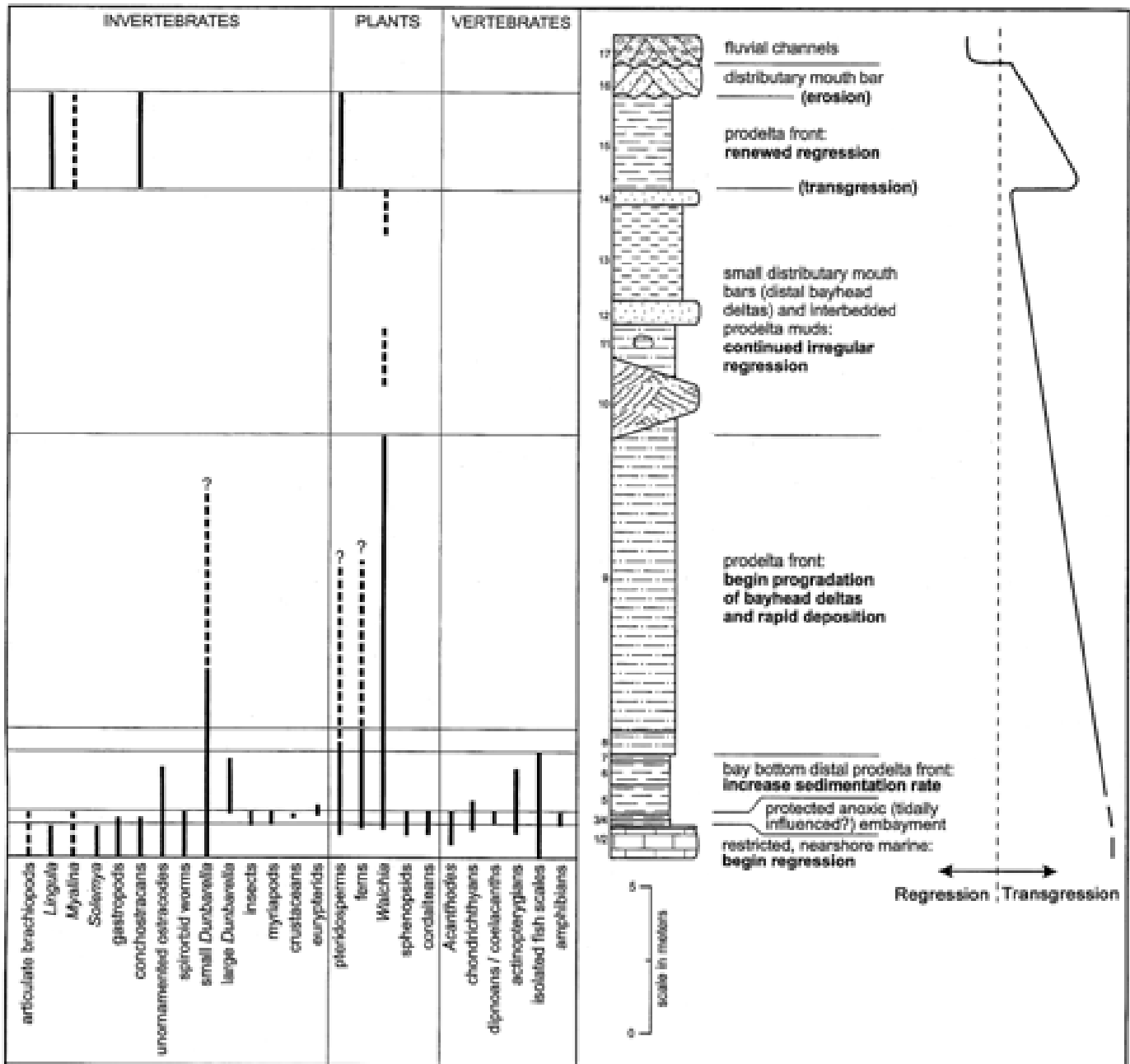


FIGURE 5. Summary diagram of the paleontology, stratigraphy, depositional environments and sea-level changes at the Kinney Brick Quarry (after Lucas et al., 1999).

(especially conchostracans and a few salamander-sized amphibians) suggest low salinity. Overlying shales (Fig. 5, units 5-7) represent a similar facies, but probably with a greater fresh-water influence. Dominant elements are *Dunbarella*, an euryhaline bivalve, and terrestrial plants.

Overlying silty shales (Fig. 5, units 8-9) are interpreted by us to represent increased sedimentation rates in a prodelta estuarine environment, brought about by the onset of a significant fluvial discharge. A conifer-dominated floral assemblage abounds here, with sparse, small *Dunbarella*. Most of the relatively xerophytic floral assemblage may have been floated a short distance into an environment characterized by frequent shallow ponding and deposition on a surface better drained than the underlying shales.

Overlying laminated and ripple-laminated sandstone ledges and intercalated shales and claystones (Fig. 5, units 10-14) are

interpreted by us as delta front, distributary mouth bars, and associated deposits. Unit 15 is a shale that shows marine influence indicated by the presence of *Lingula* and *Myalina*. This unit, and the overlying fluvial sandstone/conglomerate, may be the base of another transgressive sequence.

We thus interpret the stratigraphic sequence at the Kinney Brick Quarry as mostly reflecting a marine regression corresponding to the progradation of a clastic delta (Fig. 5). Shifting sediments from the delta probably isolated an embayment from normal marine conditions as a clastic wedge developed and extended seaward. Lagoonal conditions were established, and clastic input was initially restricted to clay-size particles. Eventually, the embayment was filled by silty shales from an advancing delta plain on which sand was later deposited. The onset of a subsequent transgression is documented by the highest strata in the quarry section. **End of log.**

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Along the Arroyo de la Presilla, a few miles east of Socorro, the base of the Phanerozoic section is the Middle Pennsylvanian Sandia Formation resting with depositional contact on Proterozoic granite. The brown, blocky granite forms the base of the cuesta in the left foreground, overlain by gray quartz sandstones and conglomerates at the base of the Sandia Formation. Overlying Sandia Formation strata are variegated white, red and brown siltstones, sandstones, shales and thin limestones. The mine adit in the upper right of the photograph is where “fire clay” was mined before World War One to make bricks at a plant in Socorro.

CLOSING THE COMMUNICATION GAP BETWEEN THE SCIENTIST/RESEARCHER AND THE SCIENCE TEACHER: HOW TO TEACH THE KINNEY BRICK QUARRY LAGERSTÄTTE

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The Kinney Brick Quarry Lagerstätte, located near Albuquerque, New Mexico, is a world-class Upper Carboniferous (Pennsylvanian) age (~ 320 Ma) Lagerstätte – a sedimentary deposit that exhibits extraordinary fossils with exceptional preservation.

One of the challenges faced by students in science classes is reading professional scientific reports. Most professional scientific reports are written for other professional scientists with advance degrees and post-doctoral work experience, not for mid-school or high school students. Mid- and high school students find things like stoichiometry, crystalline structure, kinematics, and algebra to be challenging subjects. Reading and interpreting technical graphs and charts, often the “meat” of a professional scientific work, is an advanced skill for many high school students. Any “community” of scientists has its own “jargon,” words that have specific meaning to those within the community but are often vague or even meaningless to “outsiders.”

After reading most of the scientific reports on the Kinney Brick Quarry Lagerstätte, an effort was started to “translate” the technical information contained in the reports into an educational document that would close the gap between the highly technical reports and an easy-to-understand package that would be instructional and hopefully inspire students to consider earth sciences as a future career path.

A project was conducted involving interviews with the Kinney Brick Quarry Lagerstätte experts about their personal research and special projects related to the specimens recovered from the quarry and held at the NM Museum of Natural History and the Smithsonian Institution (National Museum of Natural History). The project resulted in a 3-part product designed for educational use (Burton, 2012).

This educational product has not “dumbed-down” the scientific reports. It has, instead, supplied the “gap” information – the information assumed by the scientist but unfamiliar to the student and even unfamiliar to many science teachers. The package includes a glossary of terms and concepts complete with pictures and simple diagrammatic explanations where necessary, for example, transgressing/regressing shorelines and associated deposits.

There are three parts to this educational product:

Part 1: Printed Field Trip Guide. The guide includes suggested format and content for a physical field trip to the quarry. It also has suggestions for the integration of non-science disciplines (language, civics, biology, history).

Part 2: PowerPoint Presentation – Overview of the Kinney Brick Quarry Lagerstätte: Included with the Field Trip Guide is a set of PowerPoint slides to serve as an overview of the Kinney Brick Quarry Lagerstätte. This PowerPoint presentation may be used independently or in conjunction with the Field Trip Guide.

Part 3: PowerPoint Presentation – Depositional History of the Kinney Brick Quarry Lagerstätte: Pangaea (Upper Carboniferous) to Present; The 17 individual layers of strata are addressed in four major groups from bottom (oldest) to top (newest). Each band is examined on the basis of lithology (nature of the rock), flora (plant fossils), fauna (animal fossils), and depositional environment. Each of these stages of

examination is accompanied by color photographs of actual rock, plant fossils, and animal fossils from the quarry. The depositional environment is depicted in photographs of areas in the world today that represent good analogues of the depositional environment during the Upper Carboniferous.

If properly used, this material satisfies multiple NM State Educational Standards and should enhance the teaching of earth science and render a sense of reality to the past.

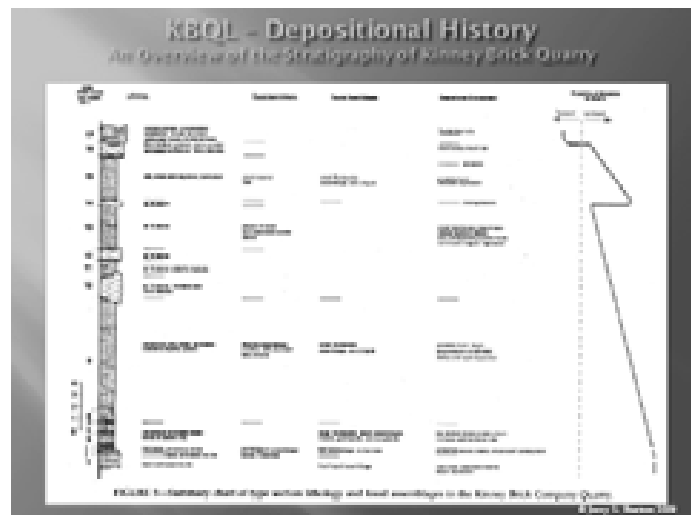
Below is the Table of Contents for the Printed Field Trip Guide: CONTENTS

- SECTION I: BACKGROUND
- SECTION II: PLANNING THE FIELD TRIP
- SECTION III: PREVIOUS FIELD TRIP DETAIL
- SECTION IV: HANDOUT MATERIAL
- SECTION V: WORKS CITED

ADDENDA:

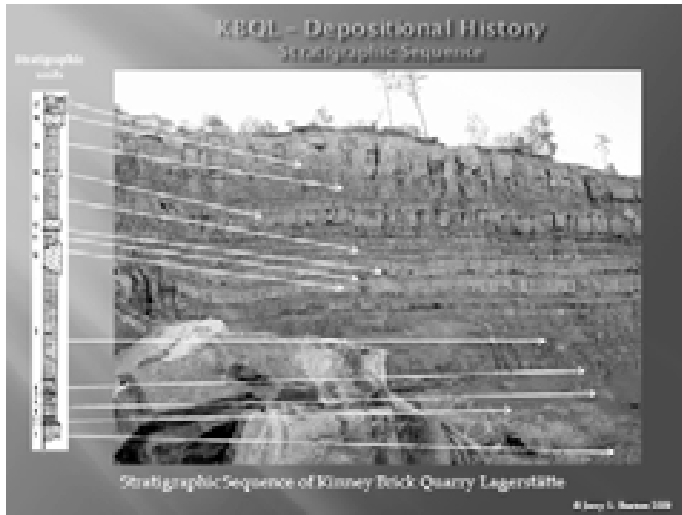
- A: Alignment with New Mexico PED Standards
- B: Glossary of Key Terms and Processes
- C: Depositional History Narrative
- D: Demographics of Albuquerque
- E: Business and Employment - Albuquerque
- F: Common Questions Suitable for Student Assessment
- G: Overview (PowerPoint Presentation)
- H: Depositional History (PowerPoint Presentation)

The depositional history of the quarry is primarily based on Figure 5, Summary chart of type section lithology and fossil assemblages in the Kinney Brick Quarry Lagerstätte (Zidek, 1992). Updates have been made to the stratigraphic naming and dating based on the recent paper, “Precise age and biostratigraphic significance of the Kinney Brick Quarry Lagerstätte, Pennsylvanian of New Mexico, USA” by Lucas et al. (2011).



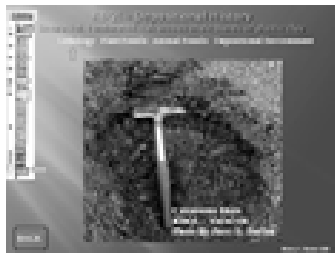
The chart was modified by color coding the units of the stratigraphic column.

The column was then added to a picture of the high wall of the quarry for visual correlation:



The verbal descriptions of the Lithology, Floral Assemblages, Faunal Assemblages, and Depositional Environments (Reference Figure 5) are accompanied by pictures of samples.

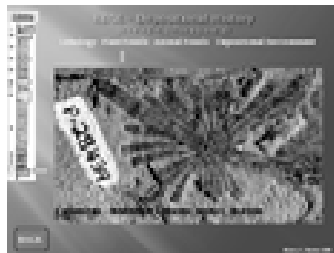
LITHOLOGY



ANIMAL FOSSILS (FAUNAL ASSEMBLAGES)



PLANT FOSSILS (FOSSIL ASSEMBLAGES)



DEPOSITIONAL ENVIRONMENTS



This is a sample page from the Glossary of Key Terms and Concepts:

Lingula, a fossil found at Kinney Brick Quarry Lagerstätte, is an example of an inarticulate brachiopod.



Lithology: (li-thol'-o-gy) The description of rocks, esp. in hand specimen and in outcrop, on the basis of such characteristics as color, mineralogic composition, and grain size.

Lithostratigraphic unit: (lith-o-strat'-i-graph-ic) A body of rock that consists dominantly of a certain lithologic type or combination of types, or has other unifying lithologic features. It may be igneous, sedimentary, or metamorphic, and it may or may not be consolidated. The critical requirement is a substantial degree of overall homogeneity. A lithostratigraphic unit has a binomial designation, preferably consisting of a geographic name from its type area combined with a descriptive term (e.g., Ohio Shale) or with the appropriate rank term (e.g., Rome Formation).

Maceration: (mac-er-a'-tion) The process of disintegrating sedimentary rocks such as coal and shale in order to extract and concentrate acid-insoluble microfossils. It includes mainly chemical treatment by oxidants and alkalies that will remove extraneous mineral and organic constituents.

Marine Regression: A marine regression occurs when sea level is falling. The sea is retreating from the continent, and the energy at any given point on the shelf is increasing as water depth decreases. So a point that is outer continental shelf (limestone facies) is on the mid-continental shelf (shale facies) following sea level fall. With continued regression the depth at this point is more reflective of the inner continental shelf (sandstone facies). So with a fall in sea level we see a seaward shift of facies zones and the stratigraphic section reveals that texture coarsens upward.

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COLORING THE KINNEY BRICK QUARRY LAGERSTÄTTE

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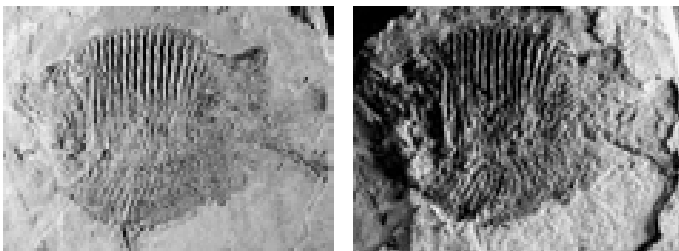
The Kinney Brick Quarry, located near Albuquerque, New Mexico, is a world-class Upper Carboniferous (Pennsylvanian) age (~ 320 Ma) Lagerstätte – a sedimentary deposit that exhibits extraordinary fossils with exceptional preservation.

An effort is currently underway to create exciting new educational material relating to the Kinney Brick Quarry Lagerstätte (KBQL). Part of this effort involves a photo log of all the major fossil specimens taken from KBQL. The photos will present the fossils in “living” color. This project began three years ago at the New Mexico Museum of Natural History. Since that time, over 3100 photos have been taken of approximately 400 specimens.

Photographing the fossils is a complex task. A reference photo is taken of each specimen. The reference is simply a mid-range well lit photograph of the fossil with its museum index card. This allows re-identification of photographs that may become separated from their photo group before the addition of museum identification information and credits.



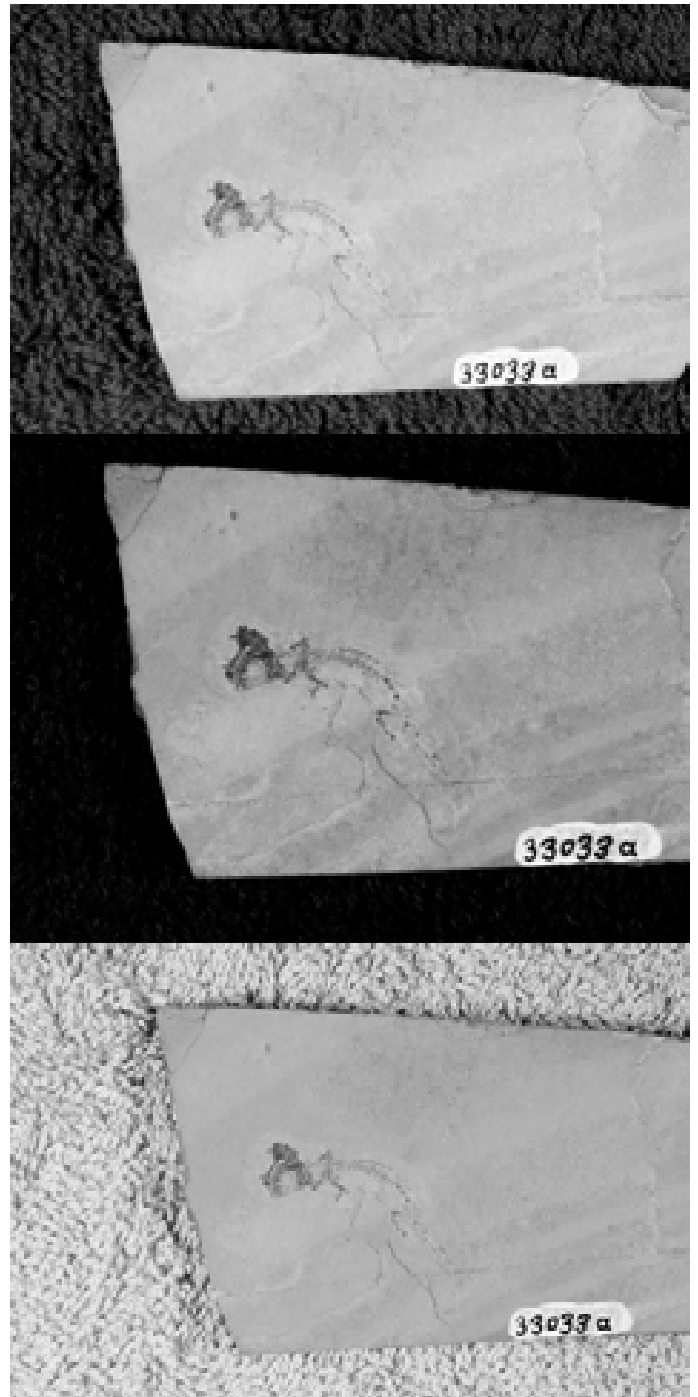
Each specimen is photographed in multiple lighting setups. Various angles and height of the lights are used to maximize the depth of the specimen. If the lights are too close, details of the specimen will be washed out. If the light is too far away, detail will be lost in the shadows. The angle of the side lighting also affects the amount of detail that will be presented in the photograph. Some specimens are photographed with light from only one side; others are photographed with as many as five lights (left, right, top, bottom, and above).



Five-light photo.

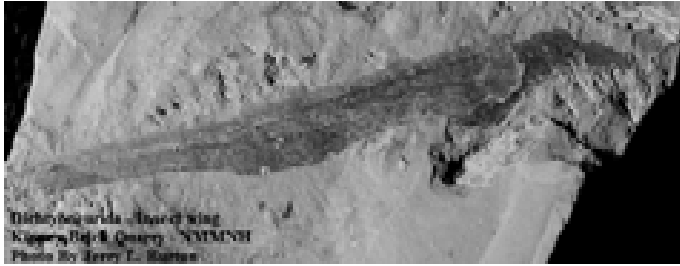
Left side lighting, 30 degree up angle.

Many specimens are photographed on backgrounds of different color. A rough surface, actually colored hand towels, are used as the background. The rough surface diffuses the light, and the reflected light adds a soft hue to the photograph. This hue often accentuates features of the specimen not visible under direct, bright light.



Every photograph of each specimen is then evaluated by several people. Some of the evaluators are scientists. Others, however, are teachers and students. It is important to capture the scientific detail that objectively represents the fossil, but it is also, because of the mixed audience, important that each photograph has esthetic appeal.

More than one photograph may be selected for any fossil if different detail is brought to the foreground. After selection of the representative photograph(s), the photograph is cropped and labeled with the scientific description, the type of collection (museum name or collector's name), and the photographer credit.



This project will be on-going in an effort to fully document for the scientific and academic community the various specimens removed from KBQL. Major contributors, to date, are the New Mexico Museum of Natural History and the National Museum of Natural History of the Smithsonian Institution in Washington, D. C. This summer, the Carnegie Museum of Natural History in Pittsburgh, Pennsylvania will also be participating. We invite all who have specimens from the KBQL to contact us so that we may include those specimens in the photo log. None of the photographs may be used commercially. It is the intention of this project that each contributing museum or similar organization will receive a copy of the photo log and be able to publish it for sale at their facility. The details of this are not yet determined.

The photographic log will be an atlas of Kinney fossils, useful to students and researchers alike, as a way to review the diversity and nature of the incredible fossil record from the KBQL. This is the importance of this project.

It is hoped that the photographic log will soon be available on its own website. Contact information for this project is:

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PENNSYLVANIAN STRATIGRAPHY AND BIOSTRATIGRAPHY AT CEDRO PEAK, MANZANITA MOUNTAINS, NEW MEXICO, USA: A BRIEF SUMMARY

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Cedro Peak is a conical mountain near the northern end of the Manzano uplift, in the part of the range usually termed the Manzanita Mountains, just east of Albuquerque in Bernalillo County, New Mexico, USA. We have undertaken a detailed study of the Pennsylvanian stratigraphy, sedimentology and biostratigraphy at Cedro Peak, some of which has been published (Lucas et al., 2011; Vachard et al., 2012, 2013), and the remainder of which is in a lengthy manuscript in review. Indeed, we regard the Pennsylvanian section at Cedro Peak as a reference section of the Pennsylvanian in central New Mexico, one critical to the interpretation and correlation of the Pennsylvanian stratigraphy in the Manzano and Sandia Mountains. Here, we present a brief summary of the Pennsylvanian lithostratigraphy and fusulinid biostratigraphy at Cedro Peak (Figs. 1-2).

At Cedro Peak, a Pennsylvanian section about 340 m thick is preserved with minor structural complications, and is relatively well exposed along the steep flanks of the mountain. We assign this section to the (ascending order) Sandia, Gray Mesa (= Los Moyos), Atrasado (= Wild Cow) and Bursum formations (Fig. 1). We divide the Gray Mesa Formation into the (ascending order) Elephant Butte, Whiskey Canyon and Garcia members, and we divide the Atrasado Formation into the (ascending order) Bartolo, Amado, Tinajas, Council Spring, Burrego, Story, Del Cuerto and Moya members. At Cedro Peak, the ~ 14 m thick Sandia Formation is almost entirely of nonmarine origin. The ~ 119 m thick Gray Mesa Formation records normal marine deposition. The ~ 200 m thick Atrasado Formation is a complex succession of marine and nonmarine (mostly fluvio-deltaic) strata. Only the lowermost 6 m of the Bursum Formation are exposed at Cedro Peak, but nearby sections indicate a Bursum thickness of ~ 90 m.

We reject the lithostratigraphic names that Myers proposed for this section (see Fig. 1). Indeed, the Pennsylvanian lithostratigraphic nomenclature proposed by Myers (see especially 1973) for the Manzano uplift is fraught with problems, including Myers own inconsistent mapping of the units he named and the fact that more than one of the units he named had been named previously. The lithostratigraphy we use (Fig. 1) is based on an older nomenclature introduced by Thompson (1942) and Kelley and Wood (1946) that well reflects the Pennsylvanian stratigraphic architecture across much of central New Mexico (e. g., Lucas and Krainer, 2009, 2010; Lucas et al., 2009, 2011).

Biostratigraphy of the Pennsylvanian section at Cedro Peak is based on fusulinids. Myers and McKay (1976) showed eight fusulinid levels in the Cedro Peak section (also see Myers, 1988a, b) (Fig. 1). Lucas et al. (2011) documented some Missourian fusulinids from the Tinajas Member at Cedro Peak, and Vachard et al. (2012, 2013) documented extensive microfossil assemblages that include fusulinids, smaller foraminifers, and carbonate algae from throughout the Pennsylvanian section at Cedro Peak.

We recognized no age-diagnostic fusulinids in the Sandia Formation section at Cedro Peak. Fusulinids from the Sandia Formation at the lectostratotype section in the Sandia Mountains, 13 km north of Cedro Peak, are of ?late Morrowan to early Atokan age (Krainer et al., 2011). Myers (1988a) documented Atokan fusulinids (his zone of *Fusulinella*) from the Sandia Formation at Sol se Mete Peak, ~ 9 km southwest of

Cedro Peak. At Cedro Peak, the limestone of the basal Gray Mesa Formation immediately above the Sandia Formation contains fusulinids that indicate a latest Atokan? to early Desmoinesian age (Vachard et al., 2012, 2013). Therefore, at Cedro Peak, we assign an Atokan age to the Sandia Formation (Fig. 1).

In the lower part of the Elephant Butte Member of the Gray Mesa Formation, Vachard et al. (2013) documented a fusulinid assemblage that consists of *Pseudostaffella* sp., *Profusulinella fittsi* (Thompson), *Schubertellina* sp. and *Dagmarella iowensis* (Thompson). They considered this assemblage to be late or latest Atokan?-early Desmoinesian because of the association of *Dagmarella iowensis* and *Profusulinella fittsi* (cf. Wilde, 1990, 2006).

Myers and McKay (1976) listed *Beedeina* aff. *B. arizonensis* (Ross and Sabins) and *Plectofusulina?* sp. from ~ 7 m above the base of the Elephant Butte Member of the Gray Mesa Formation (USGS locality f1087) (Fig. 1). Myers (1988a, pl. 3) later illustrated these fusulinids, identifying them as *Beedeina insolita* (Thompson) and *Plectofusulina* sp. He considered them to be of early Desmoinesian age, assigning them to his subzone of *Beedeina insolita*.

In the middle to upper part of the Elephant Butte Member and overlying Whiskey Canyon Member of the Gray Mesa Formation, Vachard et al. (2013) documented a fusulinid assemblage that consists of *Pseudostaffella* sp.; *Schubertellina bluensis* (Ross and Sabins); *Plectofusulina manzanensis* Vachard, Krainer and Lucas; *Wedekindellina* cf. *W. excentrica* Roth and Skinner; *W. pseudomatura* Ross and Tyrrell; *Beedeina* cf. *B. novamexicana* (Needham); *B.* cf. *B. leei* (Skinner); and *B.* sp. Vachard et al. (2013) correlated this assemblage to the second biozone of the early Desmoinesian, DS2 (e. g., Myers, 1988a; Wilde, 1990, 2006) because of the presence of *Beedeina novamexicana*, a species relatively advanced of the first lineage of *Beedeina* of Groves and Reisdorph (2009).

Myers and McKay (1976) also reported *B. novamexicana* from this stratigraphic interval (USGS locality f10188) (Fig. 1). Myers (1988a, pl. 3) illustrated *Wedekindellina* cf. *W. euthysepta* (Henbest) and *Beedeina* aff. *B. novamexicana* (Needham) from this locality, assigning it to his *Beedeina novamexicana* subzone of early Desmoinesian age.

We did not recognize any fusulinids in our samples from the Garcia Member of the Gray Mesa Formation at Cedro Peak. However, Myers and McKay (1976) reported *Beedeina* aff. *B. sulphurensis* (Ross and Sabins) from a horizon 113 m above the base of the Gray Mesa Formation at Cedro Peak, which is a level in the Garcia Member (USGS locality f10189) (Fig. 1). Myers (1988a, pl. 5) illustrated *Beedeina rockymontana* (Roth and Skinner) from this locality, and assigned it to his *Beedeina rockymontana* subzone of middle Desmoinesian age. The fusulinids thus suggest that the entire Gray Mesa Formation is of Desmoinesian age at Cedro Peak.

However, there is a large stratigraphic gap in fusulinid age control in the Cedro Peak section that spans much of the Garcia Member of the Gray Mesa Formation and the Bartolo and Amado members of the Atrasado Formation (Fig. 1). Based on conodont biostratigraphy in the Cerros de Amado of Socorro County, about 110 km to the south, the Desmoinesian-Missourian boundary is close to the base of the Amado Member (Lucas et al., 2009). However, direct dating of this interval at Cedro Peak will require additional data.

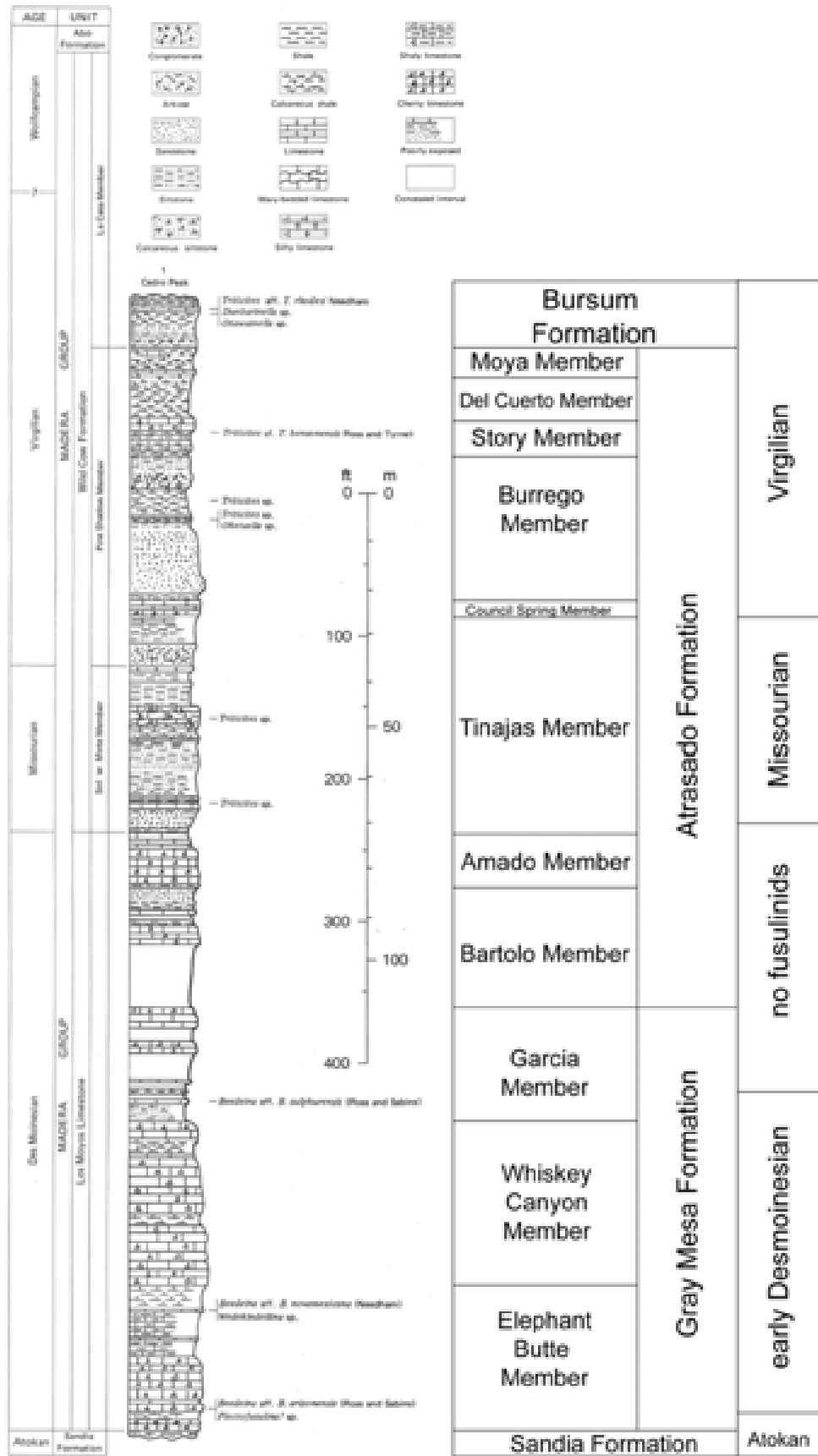


FIGURE 1. Lithostratigraphic section at Cedro Peak (on left) of Myers and McKay (1976) showing their fusulinid sample levels (also see Myers, 1988a, b) and (on right) lithostratigraphic nomenclature of this section and age determinations of Lucas et al. (2011) and Vachard et al. (2012, 2013).

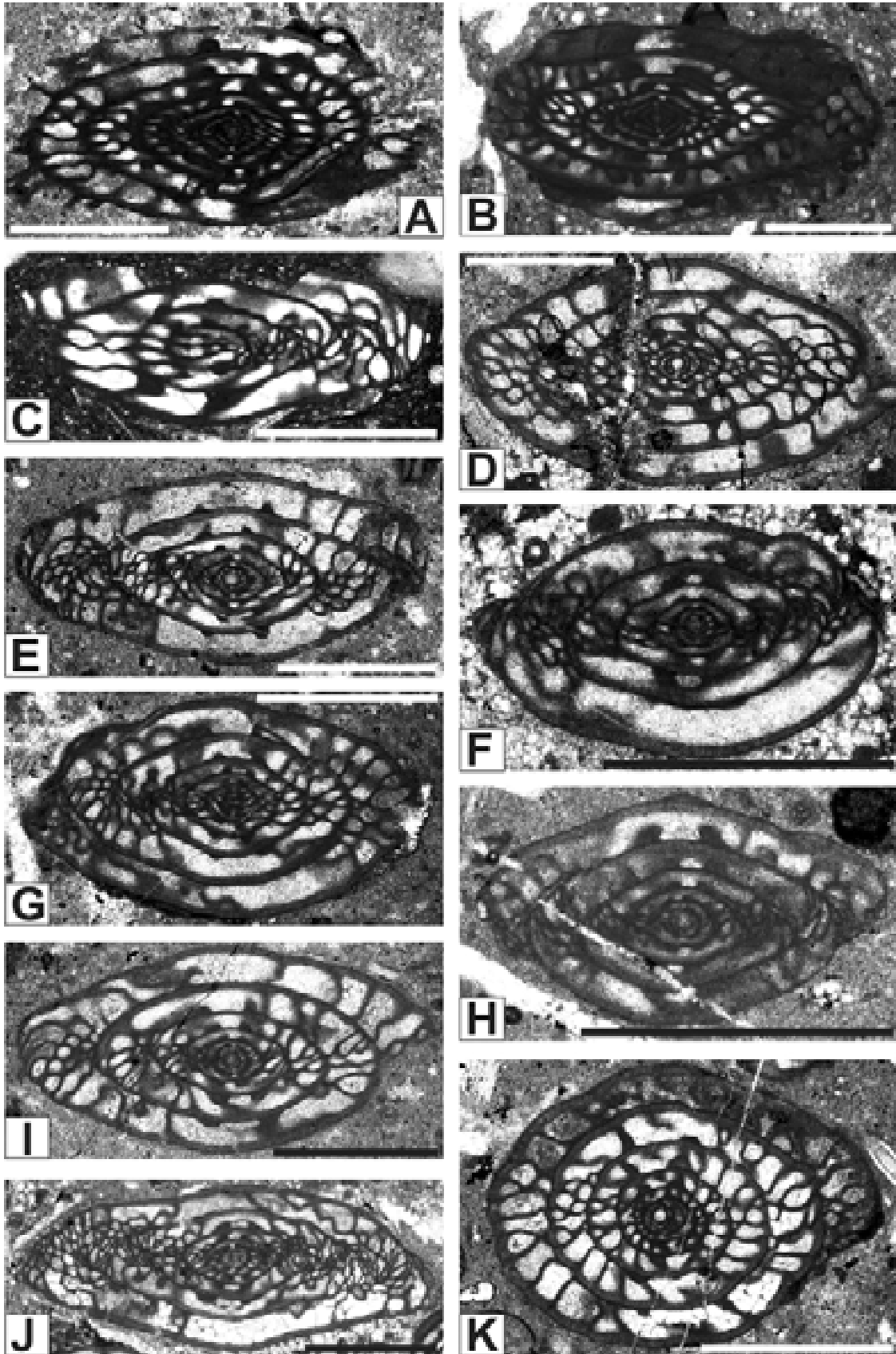


FIGURE 2. Selected Virgilian fusulinids from the Council Spring Member of the Atrasado Formation at Cedro Peak and the nearby section at Tablazon. A-C, Sample CP-B 52 (Cedro Peak, section B). D-K, Sample TAB 14 (Tablazon section). A-B, *Triticites* cf. *T. californicus* Thompson and Hazzard. C, *Triticites* sp. 1. D, *Triticites* cf. *T. legatum* Wilde. E, I, *Triticites* cf. *T. cameratoides* Ross. F-G, *Triticites* aff. *T. bungereensis* Kauffman and Roth. H, *Triticites*? n. sp. J, *Triticites* aff. *T. springvillensis* Thompson, Verville and Bissell. K, *Triticites* sp. 2. Scale bars = 1 mm.

At Cedro Peak, Myers and McKay (1976) reported *Triticites* sp. from strata near the base of the Tinajas Member of our stratigraphy. Myers (1988b, pl. 3) later illustrated *Triticites nebraskensis* Thompson from this locality (USGS locality f10258). This indicates an early Missourian age for a horizon near the base of the Tinajas Member.

Myers and McKay (1976) reported *Triticites* sp. from a horizon 44 m above the base of the Tinajas Member at Cedro Peak. Myers (1988b, pl. 3) illustrated *Triticites ohioensis* Thompson from this locality (USGS locality f10259), which indicates an early-middle Missourian age.

A laterally extensive, fusulinid-bearing limestone of the Tinajas Member at Cedro Peak (which is also a few meters below the level of the Kinney Brick Quarry to the south) yields an early-middle Missourian fusulinid assemblage consisting of *Tumulotriticites* cf. *T. tumidus* Wilde and species of *Triticites*: *T.* cf. *T. planus* Thompson and Thomas, *T.* cf. *T. myersi* Wilde and *T.* ex gr. *T. ohioensis* Thompson (Lucas et al., 2011). At the Kinney Brick Quarry, a conodont fauna in the Tinajas Member (a few meters above the fusulinid bed) is characterized by *Idiognathodus corrugatus* Gunnell and *I. cherrivalensi* Gunnell, which suggest an assignment to the *Idiognathodus confragus* Zone of the North America Midcontinent region (Dennis cyclothem; middle Missourian) (Lucas et al., 2011).

At Cedro Peak, from limestones of the Council Spring, Burrego, Story, Del Cuerto and Moya members, Vachard et al. (2013) documented a fusulinid assemblage that includes the following taxa: *Triticites turgidus* Dunbar and Henbest in Dunbar et al. and *T.* cf. *T. bungerensis* Kauffman and Roth; and *T.* sp. (also see Fig. 2). This assemblage was dated as middle Virgilian (VC2) by Vachard et al. (2012, 2013) because of the presence of two characteristic species of *Triticites*: *T. turgidus* and *T.* cf. *bungerensis* (Myers, 1988b; Wilde, 1990, 2006; Sanderson et al., 2001). This suggests the top of the Atrasado Formation at Cedro Peak is of middle Virgilian age (Fig. 1).

At Cedro Peak, Myers and McKay (1976) identified the fusulinids *Oketaella* sp. and *Triticites* sp. from horizons in the Burrego Member of the Atrasado Formation (Fig. 1). Myers (1988b, pl. 6) illustrated *Oketaella?* sp. from the lower horizon (USGS locality f10260) and (pl. 7) illustrated *Triticites* cf. *T. turgidus* Dunbar and Henbest from the

upper horizon (USGS locality 10261). He considered the sites to straddle his *Triticites bensonensis* and overlying *Triticites cullomensis* zones of early/middle Virgilian age.

Myers and McKay (1976) considered the stratigraphically highest bedrock at Cedro Peak to belong to the La Casa Member of the Wild Cow Formation. Yet, these rocks can be correlated to more complete sections nearby that indicate they are the mixed siliciclastic-carbonate strata of the Bursum Formation. In the southern Manzano Mountains, the Bursum Formation yields fusulinids of early Wolfcampian age, as it does to the south in Socorro and Sierra Counties (e.g., Myers, 1988b; Lucas and Krainer, 2004). However, at Cedro Peak and in other sections in the Manzanita Mountains, the Bursum Formation yields middle/late Virgilian fusulinids (this apparent north-south Bursum diachroneity was discussed by Lucas and Krainer, 2004).

Myers (1988b, Myers and McKay, 1976) recognized the Virgilian fusulinids in the Manzanita Mountains as the stratigraphically highest fusulinids in the section, occurring just below nonmarine siliciclastic red beds of the Abo Formation. We believe Myers used biostratigraphy, not lithostratigraphy, to assign the strata that produced them to his La Casa Member of the Wild Cow Formation, which is of Virgilian age, but does not lithologically resemble the strata in the Manzanita Mountains that yielded this stratigraphically highest fusulinid assemblage. Instead, we assign these Virgilian strata to the Bursum Formation because of their lithology and stratigraphic position (Fig. 1), not because of their age. Myers evidently was not aware that the Bursum Formation is time transgressive from north to south in the Manzano uplift.

Myers and McKay (1976) reported the fusulinids *Triticites* aff. *T. rhodesi* Needham, *Dunbarinella* sp. and *Ozawainella* sp. from the stratigraphically highest limestone beds at Cedro Peak, which they termed La Casa, but we term Bursum (Fig. 1). Myers (1988b, pl. 7) illustrated *Triticites* cf. *T. beedei* Dunbar and Condra, *Dunbarinella* sp. and *Ozawainella?* sp. from these strata (USGS localities f10262 and 10263). He assigned these fusulinids to his *Triticites beedei* subzone of late Virgilian age.

Thus, the ~ 340 m thick Pennsylvanian section at Cedro Peak spans Atokan-middle Virgilian time (Fig. 1). However, additional sampling is needed (and underway) to fill gaps and further refine the biostratigraphy of this regionally important Pennsylvanian section.

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ROAD LOG FROM ALBUQUERQUE TO THE CARBONIFEROUS-PERMIAN BOUNDARY SECTION AT CARRIZO ARROYO, VALENCIA COUNTY, NEW MEXICO

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SUMMARY

This trip is from Albuquerque to Carrizo Arroyo in the Lucero uplift of Valencia County, New Mexico. Carrizo Arroyo exposes an ~100 m thick section of Bursum Formation, strata of mixed marine-non-marine origin that yield a diverse and well known fossil record of everything from cockroaches to conodonts. This is one of the finest exposures of the Carboniferous-Permian boundary in New Mexico, and certainly one of the most thoroughly studied. Carrizo Arroyo is approximately 33 miles (53 km) as the crow flies southwest of downtown Albuquerque. Note that the final 5 miles (8 km) of the journey traverses poorly marked, deeply rutted trails and areas of deep sand, passable only in vehicles having high ground clearance and four-wheel drive.

Mileage

Description

0.0 Parking lot, east exit, New Mexico Museum of Natural History and Science. **Turn right** to proceed south on 18th Street. **0.1**

0.1 Traffic light at Mountain Road. **Turn right** to go west on Mountain Road. **0.3**

0.4 Traffic light at Rio Grande Boulevard. **Turn right** to go north on Rio Grande Boulevard. **0.3**

0.7 I-40 on ramp to right. **Enter I-40 and go two lanes to the left.** Sandia Mountains are straight ahead to east—a basement-cored, eastward tilted, fault-block uplift of late Cenozoic age that bounds the eastern margin of the Rio Grande rift here. The steep west face of the range is composed of reddish Precambrian crystalline rocks, capped on the skyline by Pennsylvanian sandstone and limestone. **2.1**

2.8 I-25 interchange, **go right to proceed south on I-25 and go to left lanes.** **3.5**

6.4 Sunport Blvd, Exit 221. The highway here is on late Cenozoic fill of the Albuquerque basin of the Rio Grande rift. These strata, broadly termed Santa Fe Group, locally yield fossil mammals and radioisotopically-dated pumice beds by which their age (here Plio-Pleistocene) is determined. **7.8**

14.2 Bridge over Rio Grande River. **1.3**

15.5 Note Pliocene (~3 Ma) basalt flow here adjacent to the highway above Santa Fe Group strata. **2.7**

18.2 Exit 209 to Isleta Pueblo. The Manzano Mountains form the eastern skyline here. Like the Sandia Mountains to the north, the Manzanos are a basement-cored, fault-block uplift along the eastern edge of the Rio Grande rift. **2.3**

20.5 Milemarker 208. At 2:00 on skyline, basalt-capped Mesa Lucero in distance. **4.0**

24.5 Exit 203 to Los Lunas. **Exit I-25 here.** **0.3**

24.8 Stop light at top of off ramp. **Turn right to proceed west on NM 6.** The prominent Isleta volcanic center just to the southwest is a compound volcano with a diameter of about 2 km that includes five basalt flows, the oldest of which is Pliocene in age, about 2.8 Ma. **3.7**

28.5 Green San Clemente sign. Crest hill. Pleistocene Cat Hills volcanic field to north on right (note the near perfect cinder cones). Mesa Lucero, capped by late Cenozoic basalts, is black on skyline directly ahead. **1.1**

29.6 Bridge over railroad tracks. This is the Belen cutoff of the old Santa Fe Railroad, which used Abo Pass at the southern end of the Manzano Mountains as a relatively good grade to traverse its major freight route from east to west. **9.2**

38.8 After another bridge, note road to left. **Turn left towards Los Lunas landfill (note green signs) and stop at railroad crossing.** **0.5**

39.3 Landfill entry point building. Continue straight on all-weather gravel road. **0.7**

40.0 “Inscription Rock” or “Phoenician Rock” or the “Commandment Stone,” also known as “The New Mexico Decalogue,” is an 80-ton block of basalt with Semitic inscriptions translated as the Ten Commandments. It is on the northeast corner of the basalt-capped hill to the left. The top of the hill also has many pictographs by Native Americans. First reported in the 1930s, the script is ostensibly similar to several Semitic languages from approximately 1000 BCE. Some have claimed that the stone is pre-Columbian in age, and therefore indicates early Semitic contact with the New World (a lost tribe of Israel or Phoenicians sailing up the Rio Grande?). We, however, must voice a healthy skepticism as to the stone’s authenticity. **3.0**

43.0 Crest of hill. Mesa Aparejo directly ahead is the type section of the Middle Pennsylvanian Gray Mesa Formation (Kelley and Wood, 1946; Krainer and Lucas, 2004). The deep gorge in the plateau directly to the right is Carrizo Arroyo, our destination. **0.9**

43.9 Road is entrenched at broad, left turning arc. At end of entrenched road **turn right onto unimproved trail. Engage four-wheel drive from here on.** **0.2**

44.1 Subtle dirt road (little more than a cattle path here) to right at blue stock tank. **Turn right to go north on two-track road (jeep trail), parallel to gravel road to landfill.** **0.2**

44.3 Road curves left to head northwest. **0.3**

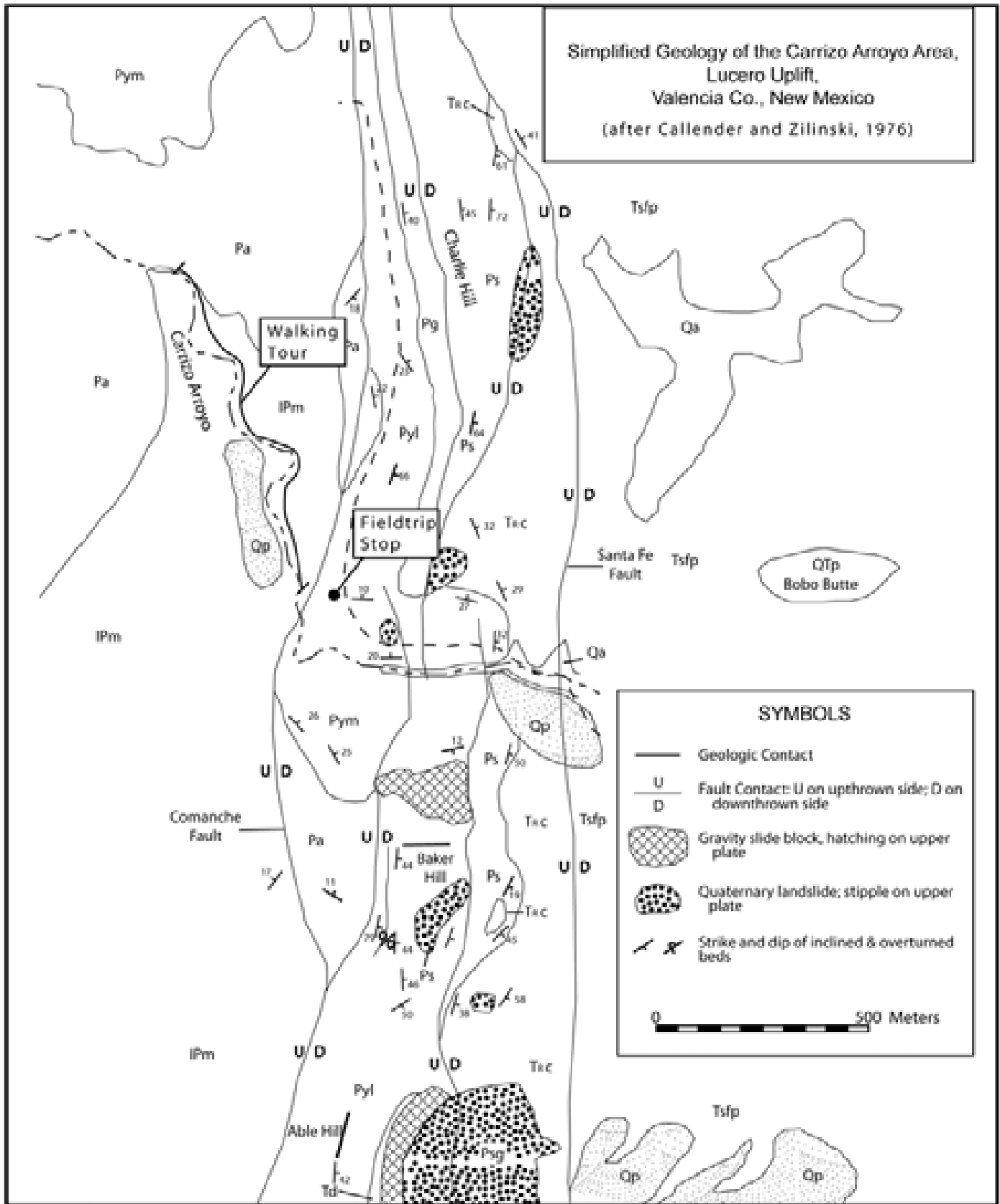


FIGURE 1. Simplified geologic map of the Carrizo Arroyo area (after Callender and Zilinski, 1976). **Acronyms (with revised stratigraphic assignments):** **IPm**, Atrasado and Bursum formations; **Pa**, Abo Formation; **Pym**, Yeso Group, DeChelly Sandstone; **Pyl**, Yeso Group, Los Vallos Formation; **Pg**, Glorieta Sandstone; **Ps**, San Andres Formation; **Trc/Km**, Triassic Moenkopi Formation and Chinle Group; **Td**, microdiorite intrusive; **Tsfp**, Santa Fe Group; **QTp**, travertine-cemented pediment gravels; **Qp**, pediment gravels; **Qt**, travertine; **Qa**, alluvium and colluvium.

- 44.6 Cross small, sandy arroyo; **definitely engage four wheel drive here. 0.4**
- 45.0 Road drops into broad, braided arroyo. Continue to left and stay left in arroyo. **0.1**
- 45.1 On left, leave arroyo to get on old two-track road that parallels arroyo. **1.0**
- 46.1 Road drops into arroyo. Go down arroyo to right. **0.1**
- 46.2 Cross main arroyo. Continue north. **0.1**
- 46.3 Road drops into arroyo. **Turn hard left up arroyo.** Drive up arroyo to southwest/west. Barbed wire fence on right. **0.2**
- 46.5 **Road to right. Leave arroyo to get on road. 0.3**
- 46.8 Arroyo embankment and dirt tank to right. Road curves left. **0.1**
- 46.9 **Fork in road, bear left. 0.1**
- 47.0 **Two track road to right. Turn right. 0.2**
- 47.2 Crest small rise. Good view ahead of basalt-capped Mesa Carrizo. Bright red slopes and hills in foreground are developed in Lower Permian Abo Formation. Light-colored, taller hills are developed in structurally-deformed Yeso, Glorieta and San Andres strata. Note Able (left), Baker (middle) and Charlie (right) hills. Callender and Zilinski (1976) show that these hills are situated within a complex structural zone between the Comanche and Santa Fe fault zones (Fig. 1), which will be the location of Stop 1, later this morning. The flows capping Mesa Carrizo are part of the Lucero volcanic field and have a Pliocene age of 4.3-3.3 Ma (Baldrige, 2004).
- The Lucero uplift is the western boundary of the Rio Grande rift and the eastern boundary of the Colorado Plateau. It is a west-tilted fault block that exposes gently dipping Pennsylvanian and Permian strata on its eastern face and dip slope (Callender and Zilinski, 1976). The eastern margin of the uplift between the two faults is a complexly faulted and folded terrain of steeply eastward dipping Pennsylvanian, Permian, Triassic, Mesozoic and upper Cenozoic strata. Of the faults exposed here, the Comanche fault is probably the oldest, with a Laramide (Late Cretaceous-Eocene) ancestry. The Santa Fe fault also has a Laramide ancestry and is closely related to the Comanche fault. The Santa Fe fault has been reactivated and overprinted by Rio Grande rift deformation. Northward, these faults merge with the Rio Puerco fault zone and finally become part of the Nacimiento fault 100 km north. The features exposed in these hills are a result of both Laramide (primarily right-lateral transpressional reverse-oblique strike slip, of Late Cretaceous-Eocene age) and rift related (primarily tensional, of late Cenozoic age) deformation.
- Eastward, a series of major extensional faults step downward into the rift, with the major stratigraphic displacement along the Santa Fe and Rio Puerco fault zones ranging from a few hundred meters to more than 3500 meters (Callender and Zilinski, 1976; Russell and Snelson, 1994). COCORP seismic reflection profiles show that the overall geometry of these faults is listric, dipping eastward, flattening deep in the earth's crust (Russell and Snelson, 1994). Oil and gas test drilling by Humble and Shell verify the presence of the faults and the thickness of the Santa Fe Group.
- The Lucero uplift was reactivated at approximately 4-5 Ma as the Popotosa basin filled in and the ancestral Rio Grande began to flow southward (Connell, 2004). **0.4**
- 47.6 Road passes through fence (gate). **0.1**
- 47.7 Crest small rise. "Bobo Butte" ahead exposes Pliocene-age strata of the Santa Fe Group. **0.3**
- 48.0 Road descends to cross arroyo. Note angular unconformity to left in Santa Fe Group strata and stone house to right (old movie set). Road crosses Carrizo Arroyo. **0.1**
- 48.1 Movie set on right. **0.2**
- 48.3 Note structurally-deformed Yeso/Glorieta strata to southwest (Fig. 1). **0.1**
- 48.4 Cross Carrizo Arroyo. Hill ahead has dark-colored Middle Triassic Moenkopi Formation strata (grayish red sandstones) dipping toward us above light-colored Permian San Andres Formation (mostly limestone) and brown crest of Glorieta Sandstone. **0.1**
- 48.5 Cross Carrizo Arroyo again. **0.1**
- 48.6 Crest hill. **0.1**
- 48.7 Cross Carrizo Arroyo. Water from Carrizo Spring is precipitating an incipient travertine. Spectacular hoodoo of Santa Fe Group sedimentary breccia on right. **0.1**
- 48.8 Road curves to left to follow border of Carrizo Arroyo. We have just crossed the rift-bounding fault here, as evidenced by the steeply-dipping strata, springs and rapid, structurally-produced change from late Cenozoic strata that are rift fill to Paleozoic-Mesozoic strata along the Colorado Plateau edge. The base of the Upper Triassic Chinle Group here is the Ojo Huelos Member, a complex calcareous paleosol (Tanner and Lucas, 2012). Also note underlying, dark reddish brown Middle Triassic Moenkopi Formation over light-colored Lower Permian San Andres Formation at Carrizo Spring. **0.1**
- 48.9 Road begins steep ascent into gorge at Carrizo Arroyo, passing from Triassic Moenkopi Formation into Lower Permian San Andres Formation. **0.2**
- 49.1 Crest hill. Good view into canyon formed by Carrizo Arroyo **STOP.** From here, on foot, we will proceed downhill to the contact of the Atrasado and Bursum (Red Tanks Member) formations. We will then walk up Carrizo Arroyo through the section of the Red Tanks Member to the base of the overlying Abo Formation (Fig. 2).
- At Carrizo Arroyo, an approximately 105-m-thick section of upper Paleozoic clastic and carbonate rocks (Fig. 2) yields extensive fossil assemblages of marine and nonmarine origin. Most of the section at Carrizo Arroyo belongs to the Red Tanks Member of the Bursum Formation, ~ 100 m thick and mostly variegated shale, mudstone and siltstone of nonmarine origin, intercalated with some beds of limestone and shale of marine origin. Red Tanks Member fossils include palynomorphs (Traverse and Ash, 1999; Utting et al., 2004), charophytes, plant megafossils (Tidwell and Ash, 1980, 2004; Ash and Tidwell, 1982, 1986; Tidwell et al., 1999; DiMichele et al., 2004; Knaus and Lucas, 2004), non-fusulinid foraminiferans (Krainer and Lucas, 2004), fusulinids (Wahlman and Kues, 2004), bryozoans, brachiopods, gastropods, bivalves and nautiloids (Kues, 1983, 1984, 2004), eurypterids (Kues and Kietzke, 1981), ostracods (Kietzke, 1983), syncarid crustaceans (Schram, 1984), conchostracans, insects and some other arthropods (Kukalova-Peck and Peck, 1976; Durden, 1984a, b, c; Rowland, 1997; Hannibal et al., 2004; Rasnitsyn et al., 2004; Schneider et al., 2004), echinoids, crinoids, conodonts (Orchard et al., 2004; Lucas et al., 2013), ichthyoliths (Johnson and Lucas, 2004) and bones of amphibians and rep-

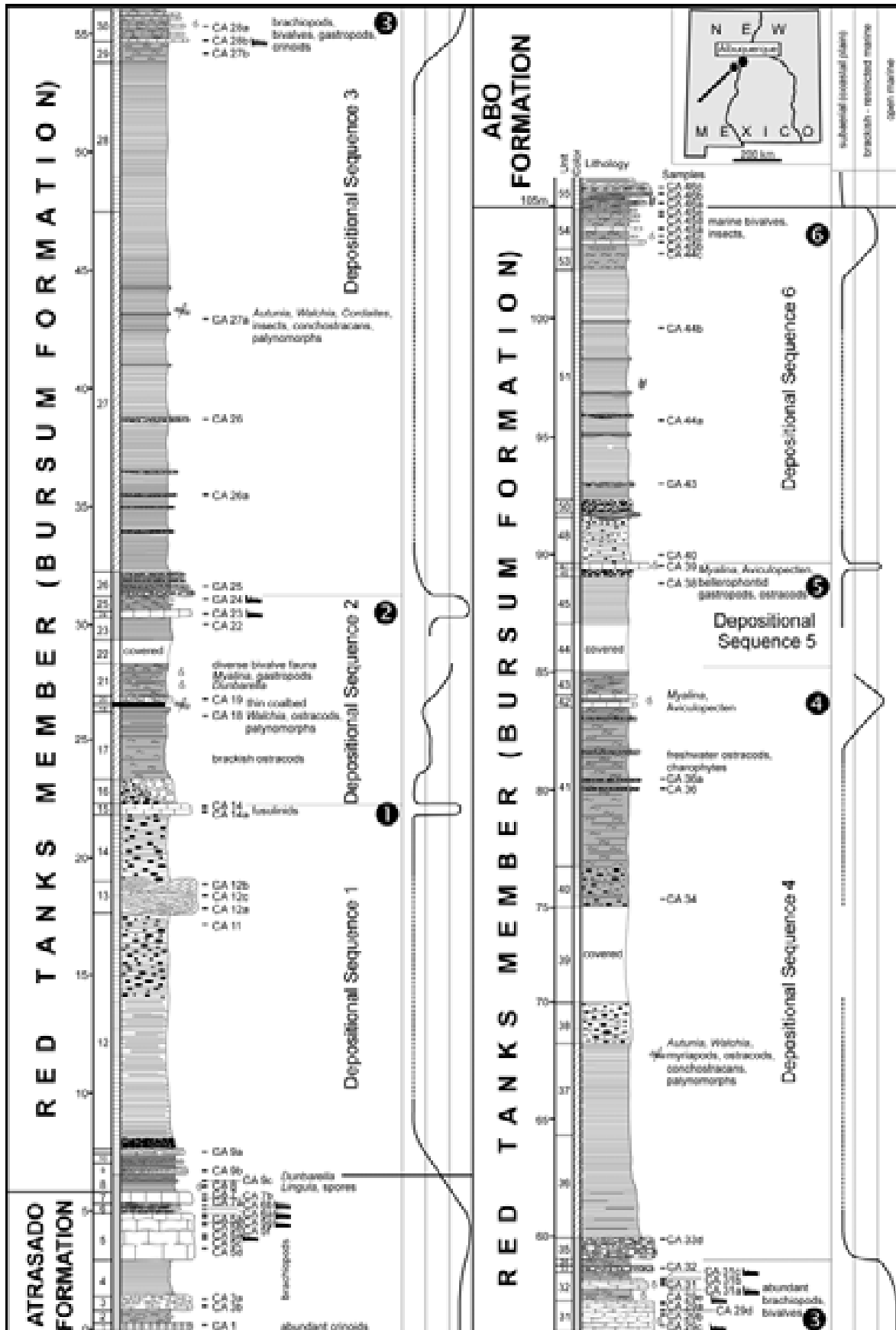


FIGURE 2. Measured stratigraphic section of the Red Tanks Member of the Bursum Formation at Carrizo Arroyo (from Krainer and Lucas, 2004). Symbols indicate beds that yielded conodonts, bivalves, gastropods, fossil plants and trace fossils.

tiles (Cook and Lucas, 1998; Harris and Lucas, 2001, 2003; Harris et al., 2003, 2004).

At stratigraphic levels 43 m and 68 m above the base of the section are Lagerstätten of plants, insects, crustaceans, eurypterids and other fossils that form unique Late Paleozoic nearshore arthropod assemblages. Most of the fossil groups from the Red Tanks Member have been used to support diverse placements of the Pennsylvanian-Permian boundary at Carrizo Arroyo (Fig. 3).

New conodont data presented by Lucas et al. (2013) include the presence of *Streptognathodus virgilicus* in the uppermost part of the Atrasado Formation, which constrains its age to the middle to upper part of the Virgilian (Late Pennsylvanian) and to a comparable position in the Gzhelian. The only biostratigraphically-significant conodont assemblage in the Red Tanks Member comes from a marine horizon near the middle of member, and is probably equivalent in age to the Midcontinent *Streptognathodus nevaensis* Zone, of early to middle Asselian (Early Permian) in age. Thus, a significant amount of latest Pennsylvanian to earliest Permian time apparently is absent at the Carrizo Arroyo section, most likely at a major disconformity at the top of the Atrasado Formation and smaller ones at the bases of depositional sequences in the lower part of the Red Tanks Member (Fig. 2). Conodont biostratigraphy therefore provides compelling evidence that Bursum Formation deposition was not simply driven by glacio-eustatic cyclicity but instead was subject to significant local tectonic control (Krainer and Lucas, 2009)

Highlights of the walking tour up the arroyo can be keyed to the measured section (Fig. 2):

1. Examine the boundary between the Atrasado and Bursum formations (Fig. 2, units 5-10); note the overturned crossbeds in basal sandstones of the Bursum (delta front?) above basal Bursum shales that contain a dwarfed (brackish water?) fauna of brachiopods, including *Lingula*. The conodonts from the uppermost limestones of the Atrasado Formation indicate a middle Virgilian age (see above).

2. Note paleosol with calcrete nodules and rhizoliths in floodplain overbank mudstones of the lower part of the Bursum

Formation (Fig. 2, units 12, 14).

3. The only known coal bed in the Bursum Formation crops out here, associated with a conifer-dominated flora and brackish water beds that contain myalinid bivalve packstones (Fig. 2, units 18-21).

4. Conglomerate beds and a clastic dike are evidence of syndepositional tectonism of the Bursum Formation (Fig. 2, units 26-27); see accompanying minipaper by Nelson et al.

5. In the shale near the middle of unit 27, the lower Lagerstätte interval has yielded a diverse flora, many insects and eurypterids, among other fossils. This was the only such fossiliferous interval known here before the 1990s, so most of what has been published on the paleoflora and paleoentomology of Carrizo Arroyo is based on fossils from this interval.

6. Limestone beds (Fig. 2, units 30-32), which yield abundant productoid brachiopods (and other macroinvertebrates), are the stratigraphically highest “normal marine” strata in the section. They also yield a conodont assemblage of early to middle Asselian age (see above).

7. The upper Lagerstätte interval (Fig. 2, near top of unit 37) was discovered in the 1990s and yields plants and insects, among other fossils, similar to those of the lower Lagerstätte. However, this upper Lagerstätte is not as extensive and has also not been as intensively collected or studied as the lower Lagerstätte interval.

8. Stratigraphically higher, thin limestone beds (Fig. 2, units 42 and 47) contain myalinid bivalves and little else. We have attempted to extract conodonts without success, probably because the facies is too shallow marine.

9. The base of the Abo Formation features an extensive, nearly monotaxial ichnofossil assemblage of *Palaeophycus striatus* that Lucas and Lerner (2004) interpreted as a low diversity example of the *Cruziana* ichnofacies indicative of a marginal marine estuary. It immediately overlies the last evidence of marine deposition in the Bursum Formation—a thin interval of shale and nodular limestone (Fig. 2, unit 54) that yields numerous steinkerns of the bivalve *Permophorus*. Again, attempts at conodont collection proved futile in this very shallow marine facies. **End of log.**

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At the Minas del Chupadero, northeast of Socorro, the Pennsylvanian-Permian transition strata are remarkably well exposed. The lower, light-colored limestone-shale interval in the foreground is the middle Virgilian Moya Member of the Atrasado Formation, and it is disconformably overlain by the lower Wolfcampian (Newwellian) Bursum Formation—the reddish purple slope capped by the first cuesta in the foreground. Beyond that cuesta, overlying, deep red strata are the Abo Formation; the Bursum-Abo contact approximates the Pennsylvanian-Permian boundary. In the distance, lighter-colored red beds (orange) of the Lower Permian Yeso Group overlie the Abo Formation.

CLASTIC DIKE IN THE LATE PALEOZOIC BURSUM FORMATION AT CARRIZO ARROYO, NEW MEXICO—A LIKELY PALEOSEISMIC SAND BLOW

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DESCRIPTION

A sandstone dike (Fig. 1) is exposed in the northwest bank of Carrizo Arroyo along the field trip route. The dike is composed of coarse, pebbly sandstone and conglomerate from unit 26 in the Bursum measured section of Krainer and Lucas (2004); it intrudes overlying silty mudstone and siltstone in the lower part of unit 27. Thus, the dike occurs about 33 m above the base of the Bursum, which is about 100 m thick at Carrizo Arroyo.

The dike strikes N 45° W and dips 45° to vertical, steepening downward (Fig. 1A). It is about 3 m high, narrowing upward from 75-90 cm at the concealed base to about 30 cm at the top. There are several narrow sills along the sides of the dike. Dike material is poorly sorted and lacks apparent internal structure. The upper end of the dike terminates sharply along a bedding plane in the mudstone. A thin layer (2-3 cm) of pebbly sandstone follows the same bedding plane, gradually pinching out away from the top of the dike.

ORIGIN

The dike evidently formed when sand and gravel from unit 26 was forced upward through fissures in unit 27. The absence of brittle fractures indicates the sediments were not lithified when this happened. No flow structures were observed. Although definite evidence is lacking, the thin layer of sand and gravel at the top of the dike may be material that was ejected onto the ground surface (or under shallow water) when the dike formed.

Clastic dikes have been reported from a wide range of sedimentary environments, and a variety of formative mechanisms have been proposed (Owen, 1987; Jolly and Lonergan, 2002; Tuttle and Hartleb, 2012). Some dikes result from water under pressure transporting sandy sediments through existing voids, zones of weakness, or fissures that formed as a result of lateral spreading. Others form when the overlying deposit breaks apart and founders into the underlying liquefied sediment. Artesian springs, non-seismic landslides, and sudden flooding or lowering of water levels may create clastic dikes (Obermeier et al., 1993; Li et al., 1996). However, there are no indications that any of these mechanisms applied at Carrizo Arroyo. Structures formed during floods are termed “sand boils.” Typically, they contain fine, well-sorted sand and show definite stratification (Li et al., 1996), unlike the dike at Carrizo Arroyo. Liquefaction also may occur due to sudden dewatering of rapidly deposited sediment (Lowe, 1975). However, the sedimentology and fossil content of Unit 27 and other Bursum mudstone points to relatively slow deposition in fresh-water ponds on a flat coastal plain (Krainer and Lucas, 2004).

Strong seismic shaking of water-saturated sediment may create “sand blows.” The great New Madrid, Missouri earthquakes of 1811-1812 produced thousands of sand blows across a large swatch of the Mississippi River alluvial plain. Eyewitnesses told of fissures “sufficiently wide to swallow horses and cattle” that ejected water and sediment “to the height of a tall tree” to the sound of “loud explosions” (Fuller, 1912; Penick, 1981). Surprisingly large objects were blown out of the earth. These included chunks of “coal” (more likely peat or lignite) weighing as much as 15 to 20 pounds (7 to 9 kg) (Penick, 1981) and the skull of an extinct (Pleistocene) musk ox (Fuller, 1912).

Seismic sand blows form where coherent silt or clay overlies water-saturated sand. When strongly agitated, wet sand becomes fluid “quick-sand” that erupts through fissures, voids, and other zones of weakness (such as roots and burrows) in the overlying sediment. Modern earthquakes have induced liquefaction, resulting in sand blows in a variety of coastal, estuarine, fluvial, lacustrine, and lagoonal settings (Youd and Perkins, 1987; Tuttle et al., 2002a). Depending on material properties, the threshold magnitude for seismic sand blows may range from M 4.5 up to about M 6.0 (Obermeier et al., 1991, 1993; Jolly and Lonergan, 2002; Tuttle et al., 2002b).

The clastic dike of Carrizo Arroyo lies just a few hundred meters west of the eastern boundary fault of the Lucero uplift (Kelley and Wood, 1946). This fault underwent movement during both the Laramide orogeny (Late Cretaceous-Eocene) and younger Cenozoic crustal extension (rifting) (Callender and Zilinski, 1976). Although earlier movement has not been demonstrated, this fault parallels a primary trend of Pennsylvanian-age Ancestral Rocky Mountain structures in the region (Woodward et al., 1999). It further parallels faults that developed during failed rifting approximately 800 Ma ago (Karlstrom et al., 1999). Note, however, that the evident fluidity of the sediment injected along the clastic dike documented here means that this injection must have occurred during the Late Pennsylvanian-Early Permian, during the interval of Bursum deposition, not more than 200 million years later when the Laramide and rift-related structures formed.

Thus, the clastic dike at Carrizo Arroyo has features consistent with sand blows produced by earthquake liquefaction. Several non-seismic modes of origin can be ruled out. Given that this feature lies close to a major fault zone that likely was active during Bursum deposition, it can be added to a list of evidence that suggests significant syndepositional tectonics took place during the interval of Bursum deposition, which straddles the Pennsylvanian-Permian boundary.

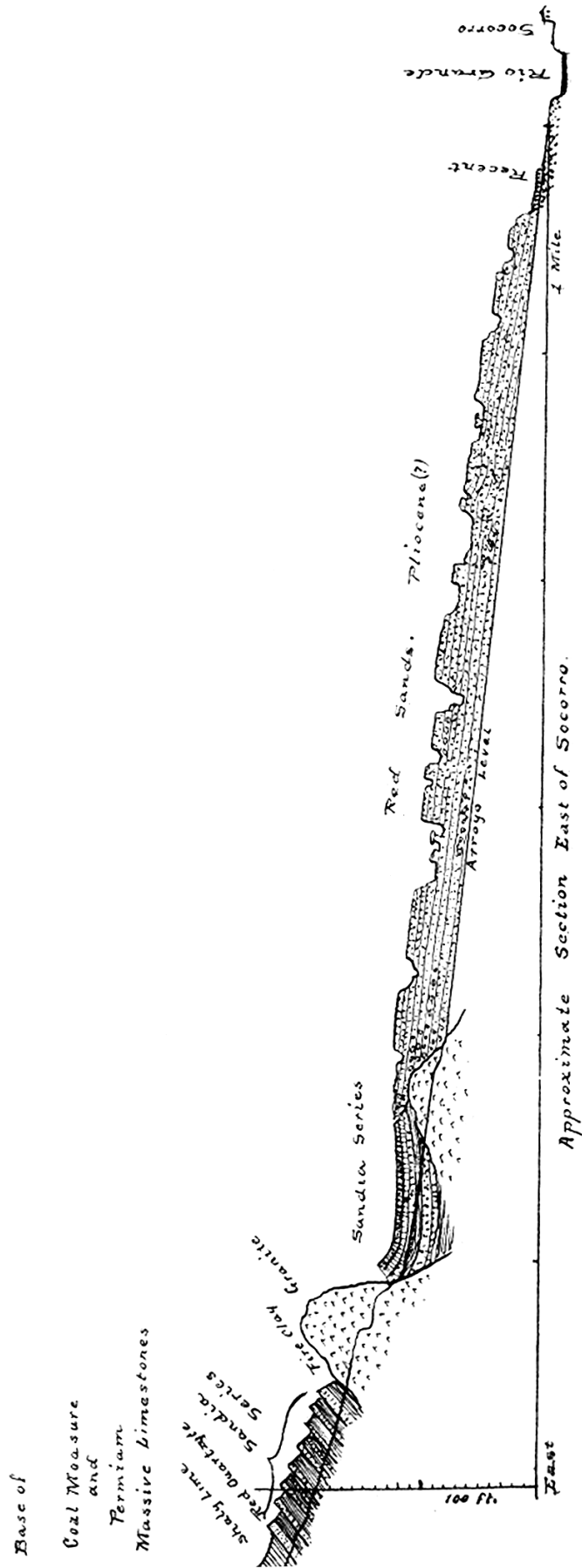
Coordinates of clastic dike: UTM 308,800 E, 3,849,835 N, grid zone 13, NAD 83; legal location approximately 1,900 feet from south line, 2,400 feet from west line, Sec. 6, T.6N., R.2W, Valencia County, on South Garcia SE 7.5' topographic map (1954).



FIGURE 1. **A**, View looking northwest approximately along strike of dike, which terminates on a bedding plane at approximately head level. **B**, Inclined hand indicates upper surface of dike termination. **C**, A thin layer of sand and gravel, indicated by arrows, follows bedding plane in mudstone at level where the dike terminates. This may be remnants of material erupted at the surface from a sand blow. **D**, View looking southwest, roughly perpendicular to northwest trend of the dike. The base of the dike is located at bottom of picture and narrows upwards, abruptly ending on a bedding plane located just below the end of the hammer handle.

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Herrick's geological cross section from Socorro (on right) eastward to the Cerros de Amado, published in 1904. The cross section is largely along or near the course of the Arroyo de la Presilla. Finally, Herrick believed that the Rio Grande Valley was a vast breached anticline with east- and west-dipping limbs of sedimentary rocks over a granitic core that was mostly eroded away but still evident at the bases of the Sandia, Manzano and Lemitar Mountains and the Sierra Ladrones. Indeed, Herrick's structural cross section from Socorro eastward to the Cerros de Amado shows "red sands Pliocene(?)" (now called Santa Fe Group) dipping towards the Rio Grande to the west, and lapping onto exposed granites draped with Carboniferous strata to the east; there is no indication of the major faults that are now known to bound the Socorro basin of the Rio Grande rift where the granite crops out.

SEDIMENTOLOGY, TAPHONOMY AND ORIGIN OF EARLY PERMIAN INSECT BEDS AT CARRIZO ARROYO, NEW MEXICO

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Coastal plain deposits of the Wolfcampian Red Tanks Member of the Bursum Formation, located in Carrizo Arroyo, southwest of Albuquerque, New Mexico, contain one of the most interesting Earliest Permian transitional marine-terrestrial fossil-lagerstaette of North America and darüberhinaus in Euramerica (Lucas et al., 2013, in press). High-frequency transgressions and regressions have formed a sequence of siliciclastic coastal deposits with intercalated marine carbonates and fluvial channels. Therefore, marine, shallow marine and alluvial environments with typical biota are set in a close lateral and vertical succession.

At Carrizo Arroyo, the Red Tanks Member reaches a thickness of about 100 m but thickens southward (Krainer and Lucas, 2004). The most common lithologies are greenish-gray and gray, rarely purple and red mud- and siltstones, which locally contain pedogenic carbonate nodules. Intercalated are some layers of sandstone and conglomerate, one coalbed and thin, gray beds of limestone. These limestone beds represent transgression events and were used to divide the profile into six depositional sequences (DS), each ending with a limestone bed in its top. About 81% of each sequence is made up by mudstones and siltstones with local carbonate nodules (Krainer and Lucas, 2004). Carbonate conglomerates comprise about 3%, sandstones about 4% and fossiliferous marine limestone beds and marly mudstone about 5% of the section (Krainer and Lucas, 2004).

From 2005 to 2009, JWS has documented and sampled the lithology and fossil content with cm-resolution of four very fossiliferous intervals, each between 3.5 to 5 m in thickness: the coal bed profile (units 17 to 21 after Krainer and Lucas, 2004) in DS 2 (completed by R. Werneburg in 2009), NMMNH locality 7724, in the middle of the lower half of unit 27 in DS 3, NMMNH locality 3437 in the middle of unit 27 in DS 3, NMMNH locality 3433 in DS 4, and units 53 and 54 at the top of depositional sequence 6 directly below the Abo base. Here, we present a preliminary review of the data and its interpretation.

The most common lithotypes in these fossiliferous intervals are (see above) greenish-gray and gray, more or less sandy mudstones and siltstones, silty sandstones and marls. The sediment is usually fine- to very fine-grained and matrix-rich. The matrix is made up of clay minerals and varying amounts of carbonate. Sometimes, bioturbation could be recognized. The common mineral assemblage is, in the order of decreasing percentage, carbonates (including carbonate clasts, bioclasts and carbonatic alterations), quartz, feldspars, which usually appear in brown color because of alteration, white mica and rarely chlorite. Additionally, a relatively high amount of opaque components, including opaque minerals as well as plant material, is characteristic. Intraclasts do occur. It is conspicuous that most non-carbonate minerals show a marginal or even area-wide alteration to carbonates, and sometimes only small parts of the former mineral were still visible. So it can be assumed that many carbonate grains are the result of such a process. Traces of azurite were observed in nearly all samples.

The sand-sized components are dominantly bioclasts of heavily abraded brachiopod and bivalve shells, bryozoans, and echinoderm sclerites and spines as well as very rare fragments of multichambered foraminiferans. Very common are smooth shelled ostracods as isolated single shells or well preserved complete, double-shelled specimens. Small pseudostheriid conchostracans and *Carbonicola*-like bivalves can be common together with insect and other arthropod remains in places on distinct bedding planes (see below taphonomy); rare are up to 3 cm long

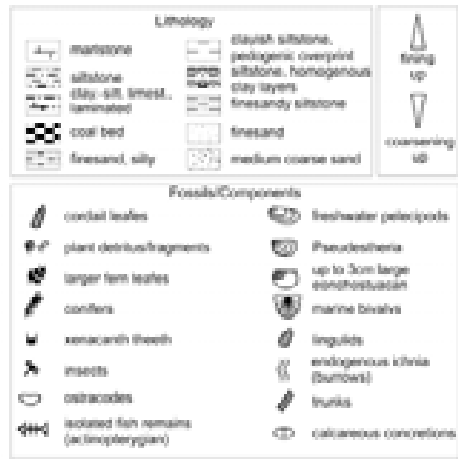
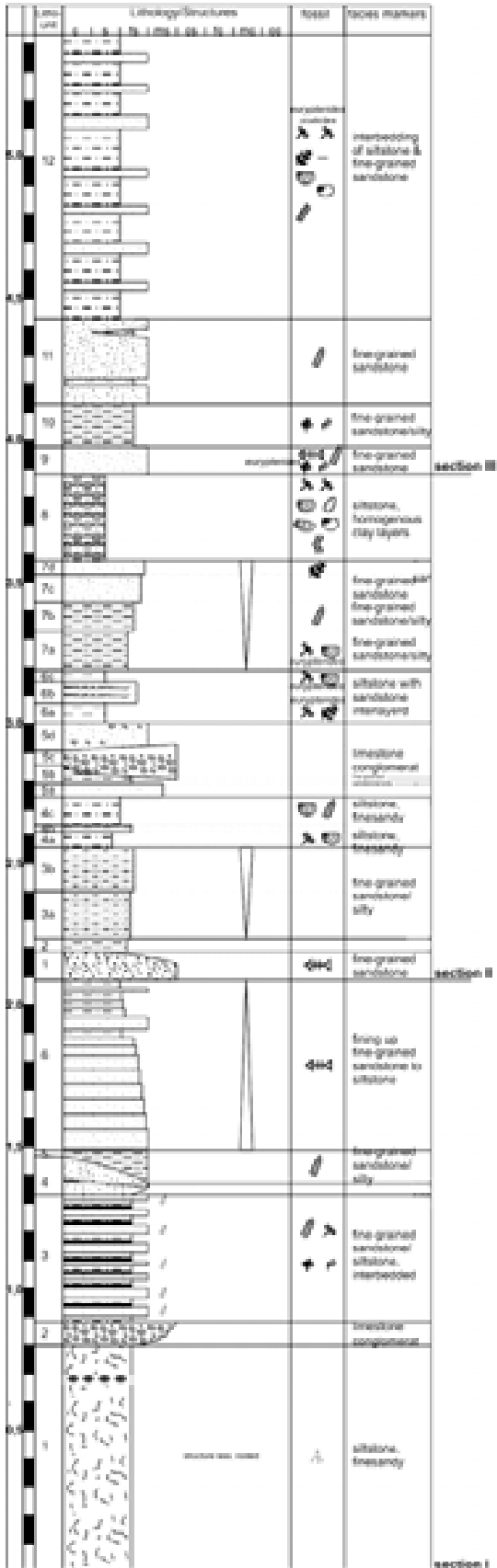
palaeolimnadiid conchostracans. Microconchids, attached to plant remains, especially to cordaite leaves, are not rare. Isolated vertebrate remains, such as scales and teeth of actinopterygians and acanthodian scales, are equally rare but mostly not corroded. Plant detritus or even coaly substance is present in nearly all horizons. Sand sized and larger siliciclastic components are in general rather rare.

Bedding is developed at the larger scale as indistinct dm to several dm thick horizontal beds; intercalated are rare, m-thick, trough-cross bedded channel fills. At the smaller scale, the widely horizontal bedsets consist of several cm to dm-thick and some meter to decameter wide shallow, stacked, and mostly indistinctly horizontal-bedded (mm- to cm-scale) channel bodies. Varying pedogenic overprint caused destruction of bedding up to a completely structureless and massive appearance. Calcic paleosoils of different maturity are indicated by mm- to cm-sized carbonate nodules in the greenish fine clastics, as well as by violet to red and purple massive mudstones and siltstones with up to several cm size carbonate nodules and some dm long and ca. 10 cm thick calcitic rhizoconcretions. Fine mm-scale rooting is observed in places; larger, cm thick subvertical roots are rare. Bedding planes often expose layers of larger plant detritus; common are dm-size fern leaves and cordaite leaves; the latter can form coaly layers up to 1 or 2 cm thick. Trunks are rare, and diameters range from cm to dm. In the coal bed profile in DS 2 cordaite leaves dominate, and the coal itself is formed mainly by these leaves. At NMMNH locality 3437 in the middle of unit 27 in DS 3 hygrophile to mesophile floral elements dominate. Higher up, in NMMNH locality 3433 in DS 4 the mesophile to xerophile conifers (“walchians”) are absolutely dominant. We only once observed distinct desiccation cracks. They occur in horizon 30 directly at the base of unit 31, the marine limestone in the top of DS 3, in the form of 40 to 50 cm deep and 10 to 20 cm wide vertical clastic dikes.

In DS 3, several thin conglomerate beds (~1 m thick), made up of limestone clasts with thin intercalated sandstone layers, are intercalated in the greenish-gray mudstones and siltstones. At NMMNH locality 3437, one of this beds could be traced from the documentation trench laterally to a several m high cliff (base not exposed). This cliff consists of stacked limestone conglomerate channels at the bottom and of pebbly sandstone at the top. It is interpreted as a fluvial main channel that pinched out laterally in the trench. The conglomerate is grain-supported; the subrounded components are poorly sorted granules and pebbles in a sandy matrix. Most of them are carbonates. The following types could be distinguished: biosparite and biomicrite, mudstones, clayey carbonates, carbonate silt and clay clasts and some micrites. The carbonates are mostly pedogenic and show a strong red coloring, either on their margins or in their centers. This may result from a former oxidization in another milieu and a following reduction after erosion and redeposition. In addition to the sedimentary clasts, large plant fragments also occur. The sandy matrix consists of mostly undulous quartz, typical brownish altered feldspars, plagioclase and white mica. The conglomerate is rich in fossil fragments, which are singly distributed in the sediment as well as included in the clasts.

Sedimentology and taphonomy of several of the insect beds in the above mentioned localities can be exemplified by unit 8 at NMMNH locality 3437, the classic insect bed, also called the “eurypterid horizon” of Carrizo Arroyo (Fig. 1). Unit 7 below consists of 39 cm of fine- and medium-grained sandstone. Directly at the boundary with unit 8 occurs

Section: Carrizo Arroyo 03-04/2005 & 09-10/2006



Section: coal bed profil 03-04/2005

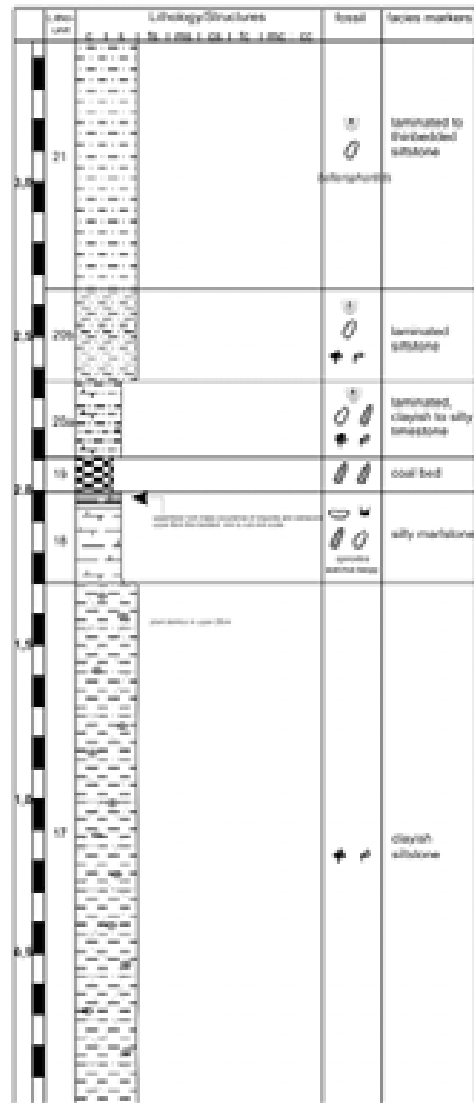


FIGURE 1. Stratigraphic profiles at the coal bed and at NMMNH locality 3437, the main insect and eurypterid beds of the Red Tanks Member of the Bursum Formation, Carrizo Arroyo, New Mexico.

a layer of large fern leaves that is followed by 1 cm of silty claystone. Unit 8 is made up of 30 cm of siltstone to claystone with cm-scale, indistinct horizontal bedding. About 10 cm from the base, 2 cm of fine sandstone rich in plant detritus is intercalated, exhibiting 0.5 cm deep load casts at the bottom as well as burrowing traces of bivalves. One single bedding plane in the middle of unit 8 that pinches out laterally over a strike laterally of about 3 m exhibited patchily arranged very fine plant detritus with common isolated and often fragmented insect wings. Besides several hundreds of insect wings, only one indistinctly preserved insect body with the articulated wings has been found. In abundance follow pseudetheriid conchostracans, relatively small *Carbonicola*-like bivalves and eurypterids. Single complete eurypterids occur as early juveniles of mm-size juveniles up to adults of dm length; the same size spectrum is covered by common eurypterid body fragments. Not rare are complete monurans (*Dasyleptus*); very rare are small enrolled diplopods.

The laterally restricted occurrence of insect wings together with tiny plant fragments point to very low-energy deposition in a laterally restricted water body. Co-occurring conchostracans form mass-occurrences. The bivalves are not fragmented, and in places their locomotion traces are observed. The aquatic, completely preserved monurans lived together with the conchostracans and small bivalves in this quiet water body. Eurypterid fragments are interpreted as moulting remains of the co-occurring complete specimens of all size classes. Plants and insects were obviously fragmented by transport. Euryhaline bivalves such as

Dunbarella and *Myalina* as well as the brackish marine lingulids, which are all together common in several beds at Carrizo Arroyo, e.g., in the coal bed profile, are missing in the insect beds. Most likely, the laterally restricted quiet water bodies that generated the insect beds were freshwater ponds.

Generally, the Red Tanks Member at Carrizo Arroyo has been interpreted as coastal plain deposits on a very shallow shelf during repeated transgressions and regressions (Krainer and Lucas, 2004). Transgression built up carbonate ramps that, during the following regressions, were subaerially exposed and eroded, delivering the carbonate detritus that built up most of the mudstone/siltstone sequences. Fluvial main channels transport the carbonate pebbles into the depositional area as bed load. The greenish-grayish mudstone-siltstone-sandstone sequences were deposited in a very low-gradient depositional environment as overbank deposits and channel fills of an anastomosing system of minor channels (lenticular bodies; horizontal bedded) besides the meandering main channels (trough-cross bedded). Beds with large plant leaves and fronds originate during falling flood. The insect beds were formed in ponds on the floodplain at the very end of a flooding event, when the finest material started to settle down in quiet water. These ponds must have existed for weeks, if they contain only conchostracans and monurans to complete their life cycle. If freshwater bivalves occur, which is not the case in all of these beds, a standing time of these ponds of several months could be assumed.

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Photograph of pioneering geologist Clarence Luther Herrick (1858-1904), who made substantial contributions to our knowledge of the geology of central New Mexico, published in articles between 1896 and 1904. These articles provide the first detailed descriptions of the regional geology, focusing on economic geology, stratigraphy, structure and paleontology. In two of his articles, Herrick provided the first detailed geology of the Cerros de Amado east of Socorro.

FIELD GUIDE TO THE CARBONIFEROUS-PERMIAN TRANSITION IN THE CERROS DE AMADO AND VICINITY, SOCORRO COUNTY, CENTRAL NEW MEXICO

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OVERVIEW

This field guide contains roadlogs for a three-day trip to the Cerros de Amado and vicinity east of Socorro, northern Socorro County, New Mexico to examine rocks and fossils of Pennsylvanian and Early Permian age. The first day's tour looks at the Middle Pennsylvanian record in this area, which encompasses the oldest Carboniferous rocks present locally. The second day tour focuses on the Upper Pennsylvanian record, and the third day on the Lower Permian record.

The three roadlogs begin at what we call the Bosquecito road junction just east of the Rio Grande at Escondida Lake (Fig. 1). We also provide a separate "entry log" (based on the roadlog in the 2009 NM Geological Society Guidebook) that starts at the Macey Center of New Mexico Tech in Socorro and proceeds to the Bosquecito road junction.

Entry Log

Mileage Description

- | | |
|--|--|
| <p>0.0 Parking lot entrance (electronic marquee) at Macey Center, New Mexico Tech Campus. Turn right. 0.1</p> | <p>3.2 Former Eagle-Pitcher battery plant on right. 0.2</p> |
| <p>0.1 Turn right onto Canyon Road. Tech Golf Course constructed on Holocene and Late Pleistocene piedmont slope. 0.3</p> | <p>3.4 Large roadcut in floodplain facies deposits of ancestral Rio Grande on left. These are fluvial and floodplain sands, silts and muds of Pliocene age. 0.1</p> |
| <p>0.4 Intersection with Buck Wolff Road, continue straight. NM Tech Police station on right. Buildings on right are New Mexico Bureau of Geology and Mineral Resources core storage facilities. Note Socorro Peak to west at about 7:00 with prominent "M" (for School of Mines) near summit. Socorro Peak is a west-tilted intrarift horst that exposes the northern topographic wall of the Socorro caldera. Most of the rocks that make up the peak are of Miocene age, capped at the peak by the Upper Miocene rhyolite of Socorro Peak. 0.2</p> | <p>3.5 Cross frontage road, New Mexico State Police station on right. Continue straight through underpass under I-25 and past frontage road on east side of interstate. This underpass was filled up to a meter deep with water when the levee failed in 2006. 0.1</p> |
| <p>0.6 Cross flood control ditch and take immediate right onto East Road. Note dissected alluvial fans here that are the piedmont deposits of Socorro Peak. 0.5</p> | <p>3.6 Sign for Scenic Byway at the village of Escondida, cross cattleguard. Escondida (Spanish for "hidden") is a defunct village, first known on the 1860 Census of the New Mexico Territory (Julyan, 1998). 0.1</p> |
| <p>1.1 Stop sign at New Mexico Tech Research Park, turn right and continue on East Road. IRIS/PASSCAL, a NSF Earth Science Research facility for geophysical studies is on left after turn. 0.2</p> | <p>3.7 Turn hard left and stay on pavement. 0.1</p> |
| <p>1.3 Soil profile covered by alluvial fan debris exposed on left in Mid to Late Pleistocene sediments. 0.2</p> | <p>3.8 The old Escondida School on left. 0.2</p> |
| <p>1.5 U. S. Geological Survey groundwater research lab on right. This facility is located on ancestral Rio Grande sands and gravels. This facility was used to study infiltration and evaporation rates in these sediments. 0.1</p> | <p>4.0 Pliocene floodplain deposits of the ancestral Rio Grande unconformably overlain by piedmont gravels of mid to late Pleistocene age in roadcuts on left. Good view to right of Rio Grande floodplain; note bosque (forest) beyond agricultural fields. The floodplain originally was covered with cottonwood bosque, and early Pleistocene plants found near the Albuquerque International Airport indicate an approximately 1.4 Ma minimum antiquity of the bosque. However, human intervention (for agricultural purposes) has generally cut the bosque back so that it only adjoins the Rio Grande along a relatively narrow strip. 0.4</p> |
| <p>1.6 Muddy floodplain deposits (mostly mudstones and siltstones) of the ancestral Rio Grande are unconformably overlain by</p> | <p>4.4 Excellent exposure of axial river deposits of the ancestral Rio Grande in large cut behind house on left. Note the trough cross</p> |

Late Pleistocene gravels in roadcuts, particularly to the left. **0.2**

1.8 Gate across road, often locked on weekends and after 8:00 p.m. in the evening. **0.1**

1.9 Socorro flood control berm visible to your right at 2:00. This berm diverts flash floods from the mountain range around the city. **0.2**

2.1 A Quaternary age fault can be seen at 10:30 as an east-facing fault scarp. Polvadera Peak at 10:00, Manzano Mountains at 12:00, Los Pinos Mountains at 1:30 and Joyita uplift (including the Cerros de Amado) at 2:00-4:30. **0.9**

3.0 Bridge over Nogal Canyon. Old sanitarium buildings against the hill to your left, are now occupied by a paintball park. Note breach in levee to left caused by recent flooding. Socorro wellhead at right. **0.2**

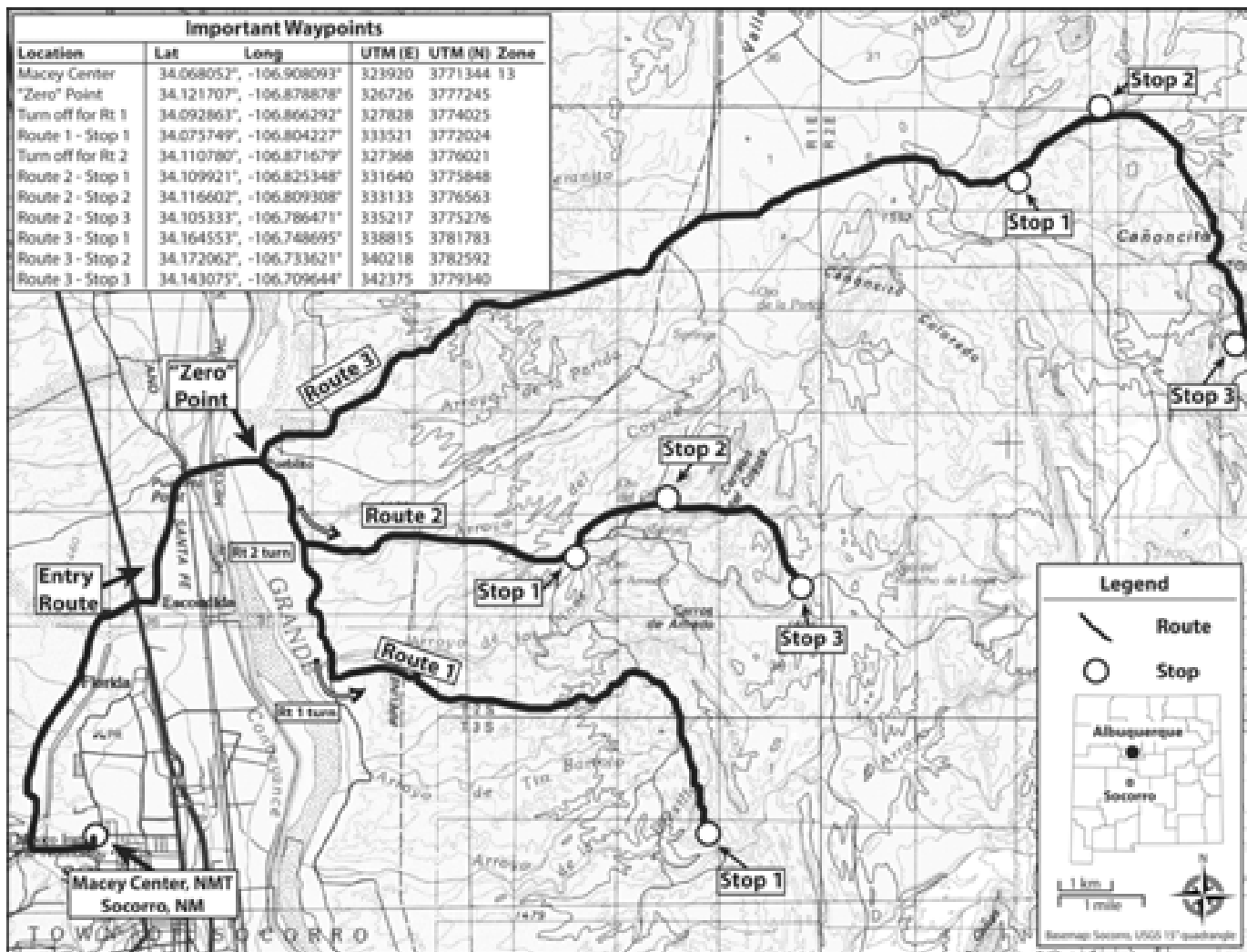


FIGURE 1. Route map for roadlogs to the Carboniferous-Permian transition in the Cerros de Amado and vicinity.

bedding. Across the Rio Grande, to the east, in Arroyo de la Parida, these strata yield fossil mammals that indicate a Pliocene (~2.7-3.2 Ma) age. Road follows Middle Rio Grande Conservancy District irrigation ditch, which sustains cultivation on the modern Rio Grande floodplain below and to your right. **0.7**

5.1 Intersection with Pueblito Road, **turn right** to Escondida Lake (and Park). The lake was originally an abandoned oxbow of the Rio Grande that has been modified into the modern lake that resembles a borrow pit. The lake is stocked and provides opportunities for fishing and picnicking for visitors. Cross Burlington-Santa Fe Railroad tracks. **0.2**

5.3 Intersection with Escondida Lake access road on left. Continue straight over low flow conveyance channel of the Rio Grande and main flood control levee. Entering Rio Grande bosque. **0.1**

5.4 Cross "Four Mile" bridge over the Rio Grande. Note groundwater monitor well field to your right. **0.2**

5.6 Leave Rio Grande bosque. **0.1**

5.7 Ascending low terrace into the little village of Pueblito. According to Julyan (1998), Pueblito was probably settled during the 1850s by the former residents of La Parida, a nearby

village that was lost to the Rio Grande. On the 1860 census it was called Pueblito de la Parida, and remains today as an unincorporated cluster of houses and small farms. **0.2**

5.9 Intersection with the Bosquecito Road, which runs down the east side of the Rio Grande, eventually intersecting U.S. 380 near San Antonio. The "Quebradas Backcountry Byway" branches off the Bosquecito road approximately 1 mile south of this point. **This is the zero point of the following three days of roadlogs.** Days 1 and 2 proceed south on this road, day 3 proceeds north.

FIRST DAY ROADLOG – SANDIA AND GRAY MESA FORMATIONS AT THE ARROYO DE LA PRESILLA

Summary

Today's trip is short, but traverses a very rugged road, to the head of the Arroyo de la Presilla, just south of the Cerros de Amado (Fig. 1). The stop is at one of the few places where the Proterozoic basement is exposed locally, overlain depositionally by the Middle Pennsylvanian Sandia Formation. The Sandia Formation here has yielded fossil plants, conodonts and selachians, part of the stop's focus. A walking tour from the stop allows close inspection of the excellent (and characteristic) exposures of the Sandia Formation, and of much of the overlying Middle Pennsylvanian Gray Mesa Formation.

Mileage Description

- 0.0 Bosquecito road junction; turn right and proceed south. 0.1**
- 0.1 Pavement ends. 0.8**
- 0.9 Quebradas road (the BLM's "Quebradas Backcountry Byway") to left is the route of the second day roadlog; continue straight. 0.5**
- 1.4 Cross cattleguard. 0.2**
- 1.6 Road forks (three roads); take the middle road and then make a hard right. 0.2**
- 1.8 Cross cattleguard; sign on right says we are entering the Eva Hilton Lewis Ranch, which is the property of New Mexico Tech (New Mexico Institute of Mining and Technology). 0.3**
- 2.1 Cross Arroyo de Los Pinos. 0.1**
- 2.2 Road to left, turn left. Put vehicle in four wheel drive. 0.7**
- 2.9 Cross pipeline; continue straight. 0.1**
- 3.0 Road to left; continue straight to climb hill to right. 0.3**
- 3.3 We have ascended a bajada surface (coalesced alluvial fans); go to left to pass through gate. 1.0**
- 4.3 The road dips through a saddle. The tan and gray gravels to the left belong to the Upper Cenozoic Santa Fe Group. 0.7**
- 5.0 Road forks; both forks lead to the same place, but the left fork is a better road. 0.2**
- 5.2 Crest of hill; the road crosses the rift-bounding fault about here. 0.2**
- 5.4 Outcrops to right are Pennsylvanian Sandia Formation. 0.1**
- 5.5 Road enters from left, continue to right. 0.1**
- 5.6 Road now on Pennsylvanian limestone. 0.2**
- 5.8 Road enters from left; stay on main road by bearing right. 0.1**
- 5.9 Another road enters from left; good view of step- and ledge-forming outcrops of Gray Mesa Formation ahead. 0.1**
- 6.0 The road is now entrenched in an arroyo; sandstones to right are part of the Sandia Formation. 0.2**
- 6.2 Pass through gate. 0.2**
- 6.4 Ridge on left is a linear, very localized outcrop of the Proterozoic granitic basement. Road to right, stay left to cross the arroyo (this is probably the worst segment of the road on today's trip). 0.3**
- 6.7 Crest of hill; outcrops all along road here are faulted strata of the Sandia Formation. 0.3**
- 7.0 Breached stock dam that was excavated in shale of the Sandia Formation; the road now climbs the hill to the left. 0.5**
- 7.5 STOP at saddle before road proceeds steeply down the hill. The hill (ridge) immediately to the west of us is a granite ridge draped with sandstone of the Sandia Formation.**

Here, at the "head" of the Arroyo de la Presilla we can examine three main features. Two are within 100 meters of the stopping point – the depositional contact of the Middle Pennsylvanian Sandia Formation on the Proterozoic basement and Herrick's (1904) lycopsid-dominated fossil plant site, his "coal-measure forest," in the Sandia Formation. The third

feature is the entire Sandia Formation section and most of the overlying Gray Mesa Formation. Beautifully exposed here, examination of these units requires a walk (roundtrip) of about 3 to 5 km over rugged terrain.

1. Sandia Formation. At Arroyo de la Presilla, the Atokan Sandia Formation is 162 m thick, rests on granitic Precambrian basement and consists of a cyclic succession of siliciclastics and carbonate, nonmarine and marine strata forming well developed transgressive cycles (Figs. 2-3). The lower 46 m are almost entirely composed of siliciclastic sediments with only one thin limestone bed. The next 46 m are composed mostly of siliciclastic sediments with several intercalated fossiliferous limestone horizons. The uppermost 70 m are dominantly siliciclastics with thin limestone interbeds in the upper part.

We recognize the following lithotypes in the Sandia Formation in the Arroyo de la Presilla section (Fig. 3): conglomerate (2.5% of the section) coarse sandstone (23.5%), fine-grained sandstone/coarse siltstone (11.2%), shale/fine siltstone (11.5%, covered 44.3%) and limestone (7%). The base of the formation is a 2-m-thick conglomerate bed, that grades upward into pebbly sandstone and sandstone; stratigraphically higher conglomerate beds are thinner, about 1 m thick. These conglomerates are quartz-rich, poorly to moderately sorted and relatively fine-grained, with a maximum grain size of about 3 cm; the grains are mostly subrounded, but angular to subangular near the base. At the base of the formation a thin lag with boulders up to 20 cm in diameter is developed. The conglomerate beds have erosive bases and are indistinctly to distinctly trough cross-bedded.

Coarse-grained sandstone is commonly trough cross-bedded, rarely displays planar cross-bedding and may be pebbly. The sandstone is quartz-rich, reddish and individual quartz grains are up to 1-2 cm in diameter. Sandstone intervals are up to 4.7 m thick, fining upward, and composed of multistoried channel fills. Individual sandstone beds contain fossil plant fragments, including stem fragments up to > 1 m long. The sandstone intervals display erosive bases, and rarely (unit 53) mudstone clasts (rip-up clasts) up to 10 cm are present. The sandstone is composed of abundant monocrystalline quartz, subordinate polycrystalline quartz, rare detrital feldspars, which are almost completely altered to clay minerals ("pseudomatrix"), very rare detrital muscovite and rock fragments of quartz and feldspar (granitic), rare chert grains and phyllic (metamorphic) rock fragments. Some opaques--rare grains of zircon, tourmaline, apatite and sphene--are present. A few clayey sedimentary rock fragments may also be present (AP 20). Sandstones (quartzarenite) are cemented by authigenic quartz overgrowths, with locally fine crystalline quartz cement in the pore space. The sandstone is stained red by very small hematite, finely dispersed in the quartz cement. Some clayey matrix (<5%) may be present. Feldspar content is less than 5 % throughout the succession. Locally small patches of coarse calcite cement replacing quartz or fine-crystalline carbonate cement occur.

In the upper part of the formation coarse blocky poikilotopic calcite cement replaces quartz and feldspar. Also, in the upper part mixed siliciclastic carbonate fossiliferous sandstone is present with thin micritic layers (or large rip-up clasts). This sandstone is medium- to coarse-grained, subangular to subrounded, poorly sorted and indistinctly laminated.

Most abundant is monocrystalline quartz, subordinate polycrystalline quartz, some micritic carbonate grains and few detrital mica. Fossils include shell fragments of gastropods and brachiopods, crinoid fragments, few brachiopod spines, rare ostracods, large micritic rip-up clasts (several cm long) containing few gastropods, bryozoans and echinoderms.



FIGURE 2. Photograph of the old “fire-clay” pit in the Sandia Formation at the Stop at the head of Arroyo de la Presilla.

The “fire clay” of Herrick (1904) in the lower part of the section is light gray, 0.5 m thick and contains plant fossils. With considerable effort, this material was mined during the late 19th and early 20th century and packed or hauled to Socorro for making bricks. Although “fire clay” usually denotes non-fissile claystone such as the underclay of a coal seam, this deposit consists of laminated silty shale to siltstone. It is underlain by dark gray, laminated, silty claystone containing *Lingula* and plant fossils. The overlying yellowish-brownish siltstone contains abundant impressions of *Lepidodendron* and associated wetland plants (Herrick, 1904; Darton, 1928; Lucas et al., 2003, 2009)

Thick siltstone/fine-grained sandstone intervals occur in the lower part of the Sandia Formation section, below and above the “fire clay.” The thickness of these intervals is from 1.2 to 4.7 m. Thinner siltstone/fine-grained sandstone layers are also developed on the top of conglomerate/sandstone units (up to 1.3 m thick) and rarely as thin (0.3 m) intercalations in shale. The most common lithofacies are horizontally laminated and ripple laminated (small-scale current ripples) siltstone to fine-grained sandstone. Small-scale trough cross-bedding is also observed. Rarely, fine-grained sandstone is bioturbated.

From 21 to 26 m above the base of the Sandia Formation section, greenish-brownish silty shale is poorly exposed with a thin micaceous sandy siltstone intercalated in the lower part

and a thin fossiliferous limestone in the upper part. The shale immediately below and particularly above the limestone contains abundant marine fossils such as crinoids, bryozoans, brachiopods and rugose corals. In the middle and upper part of the section shale intervals are mostly covered and up to 6.6 m thick in the middle part and up to 12.8 m thick in the upper part. Marine fossils such as crinoids and brachiopods occur in a 4-m-thick brownish shale in the middle part of the section (unit 46).

In the lower part of the section, only one thin limestone bed (10-20 cm thick) is poorly exposed, which is fossiliferous and contains brachiopods and bryozoans. The microfacies of this limestone is coarse-grained, poorly sorted bioclastic wackestone to packstone containing few bioclasts larger than 1 cm. The most abundant fossils are bryozoans and crinoids, subordinate are brachiopod shell fragments and brachiopod spines, echinoderm spines, gastropods and trilobite fragments, rare ostracods and smaller foraminifers (*Endothyra*). Many skeletons are encrusted by cyanobacteria, *Calcvirtella* and *Clara crusta*.

In the middle and upper part of the Sandia Formation, limestone intervals are 0.3-2.2 m thick, commonly brownish weathered, gray to dark gray, with bed thicknesses of 5-30 cm. Typically, the limestones are coarse-grained, sandy and fossiliferous with abundant fragments of crinoids,

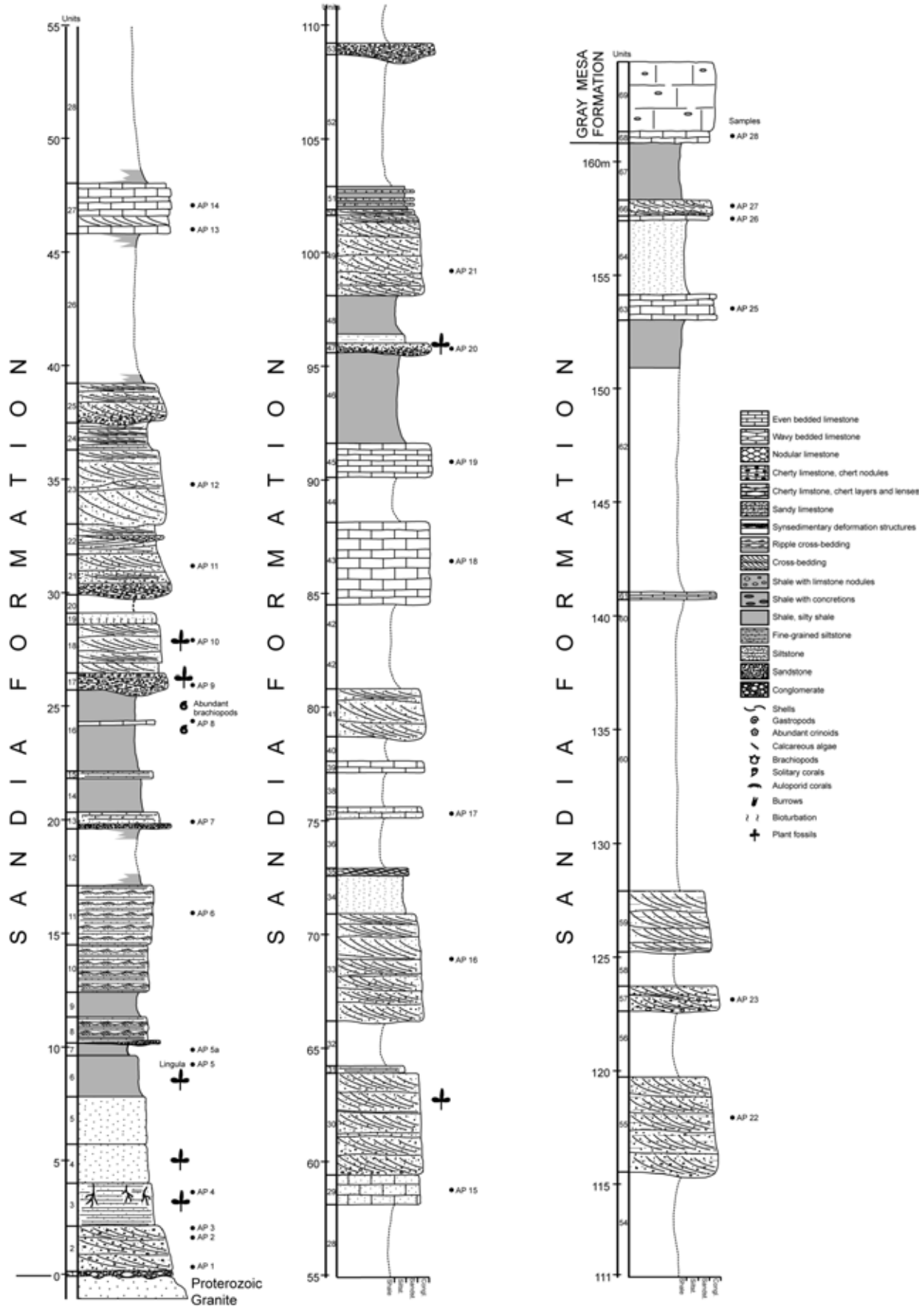


FIGURE 3. Measured section of the Sandia Formation at Arroyo de la Presilla. After Lucas et al. (2009).

brachiopods, bryozoans and (subordinately) solitary corals. Rarely the limestone displays cross-bedding (unit 27) or appears massive. The following microfacies types are observed: bioclastic wackestone to packstone and crinoidal wackestone to packstone-rudstone, and bioclastic floatstone to rudstone. A small Atokan conodont fauna of *Declinognathodus marginodosus*, *Idiognathoides sulcatus*, and *Idiognathodus incurvus?* was recovered from unit 27 (Lucas et al., 2009).

In the upper part of the section (unit 61), fusulinids and gastropods are present in a strongly altered, dark brown- to black-stained limestone bed. Quartz grains (0.1-0.3 mm, rarely up to 1 mm) are also present in amounts up to 10%. A few bioclasts are encrusted by *Calcvivertella*. The rock contains micritic matrix, and locally some calcite cement. The upper most thin limestone bed (unit 65) is a recrystallized, non-laminated, fine-grained dolomitic mudstone containing abundant spicules and rare smaller foraminifers.

Bioclastic wackestone to packstone is coarse-grained, commonly grading into floatstone to rudstone, poorly sorted, indistinctly laminated and contains few bioclasts (shell fragments) up to several cm in size. The abundant, strongly fragmented bioclasts include bryozoans, crinoids and brachiopod shell fragments, subordinate are trilobite fragments, ostracods and brachiopod spines. In the upper part of the section (unit 61) fusulinids and gastropods are present in a strongly altered, dark brown to black stained limestone bed. Quartz grains (0.1-0.3 mm, rarely up to 1 mm) are also present in amounts up to 10%. Few bioclasts are encrusted by *Calcvivertella*. The rock contains micritic matrix, and locally some calcite cement.

The crinoidal packstone is coarse-grained, moderately to poorly sorted, non-laminated to indistinctly laminated, with grain sizes mostly 0.5-2 mm, rarely up to > 1 cm (rudstone). This microfacies contains abundant crinoid ossicles, and detrital, angular to subangular quartz grains (monocrystalline, subordinately polycrystalline quartz mostly 0.3-1mm). Subordinate are bryozoans, brachiopod shell fragments and spines and rare trilobite fragments and ostracods. Few bryozoan fragments, rarely crinoids are encrusted by *Calcvivertella*. This type is well washed and calcite cemented. Quartz is present in various amounts, mostly < 5%; individual thin layers contain 20-70% quartz.

The uppermost thin limestone bed (unit 65) is a recrystallized, non-laminated, fine-grained dolomitic mudstone containing abundant spicules and rare smaller foraminifers.

Structurally, this site reveals two east-tilted fault blocks composed of Proterozoic granite and Sandia Formation. Both are west dipping, high-angle normal faults that probably developed during Cenozoic rifting. Mineralization and hydrothermal alteration along the faults prompted efforts at small-scale mining. The eastern fault is well exposed on the north side of the main arroyo, southeast of the parking place. **2. Herrick's "coal forest."** In 1904, Clarence Luther Herrick described a lycopsid flora (including three new species of *Lepidodendron*) from "fire clay" (shale) of Pennsylvanian age being mined for brick manufacturing here, at the head of Arroyo de la Presilla (Figs. 2, 4). Herrick's description of the locality was vague, and it had not been revisited in nearly a century, but in 2002 we relocated Herrick's locality.

The "fire clay" is a refractory gray to black shale in the lower part of the Sandia Formation that can be followed on strike through a series of fault blocks for more than 2 km. We suggest that the succession of Sandia Formation strata that encompass Herricks' plant locality represent fluvial deposits directly overlain by an estuarine deposit (lycopsid beds and

"fire clay"). Our collections of fossil plants from the lycopsid bed (Fig. 5) include *Lepidodendron aculeatum*, *Lepidostrobus*, possibly *Synchysidendron*, stigmarian roots and strap-like leaves of the lepidodendrids, *Sphenophyllum*, and neuropterid foliage. Because the type specimens of the species of *Lepidodendron* Herrick named were destroyed in a fire in 1910, we collected new specimens to serve as "topotypes" of the species. However, most of Herrick's species appear to be within the range of variability known from the single species *Lepidodendron aculeatum*. This lycopsid locality in the Sandia Formation is significant because it indicates that a typical wetland swamp flora existed in New Mexico during early tectonism of the Ancestral Rocky Mountain orogeny.

3. Gray Mesa Formation. Descend the steep draw southeast of the parking area to the main stream bed and proceed up stream (east) approximately 400 meters beyond exposures of the eastern fault having Precambrian granite in the footwall. At this point, turn left and follow a large side ravine on the north-northwest to view the Gray Mesa Formation along the moderately steep, west-facing hillside.

At the Arroyo de la Presilla, we measured most of the Gray Mesa Formation (Fig. 6). The measured section is 192.6 m thick; the lowermost 8 m represent the uppermost Sandia Formation. The Gray Mesa Formation can be divided into the Elephant Butte Member (95 m), Whiskey Canyon Member (25 m) and Garcia Member (94 m). The Elephant Butte Member consists of different types of limestone, covered shale intervals, two thin sandstone beds, two thin limestone conglomerate beds, and a prominent, 10-m-thick sandstone interval in the lower part (units 32-35), which displays an erosive base and begins with coarse, pebbly quartzitic sandstone that is indistinctly cross-bedded. In the coarse-grained lower part, about 3.5 m above the base, a pebbly horizon is present that contains a few crinoid and brachiopod fragments.

In the lower part of the Elephant Butte Member, below the prominent sandstone interval, a 0.6-m-thick cross-bedded quartzose sandstone bed and a 0.3-m-thick carbonate conglomerate bed with limestone nodules up to 3 cm in diameter is present. In the middle of the member, a 0.9-m-thick cross-bedded calcareous sandstone bed is present, and 20 m higher in the section a 0.8-m-thick intraformational limestone conglomerate with limestone clasts up to 3 cm in diameter is present. Covered intervals (0.3-6.7 m thick) most likely represent shale, which is rarely exposed and of gray color. The thickness of the limestone units ranges from individual beds 0.1 m thick to intervals of multiple limestone beds 6.4 m thick. We distinguish the following limestone types: (1) thin, wavy bedded limestone with bed thickness mostly 10-20 cm; (2) thick-bedded limestone, with bed thickness commonly 20-50 cm; (3) thick-bedded, coarse, crinoidal limestone; (4) massive to indistinctly bedded algal limestone, 0.9-1.8 m thick; and (5) wavy bedded to nodular cherty limestone, thin bedded (mostly 10-20 cm).

Fossils observed in the field are algae (particularly in the massive algal limestone facies), crinoids (abundant in the crinoidal limestone facies), brachiopods, bryozoans and solitary corals. Fusulinids are rare, occur in units 80 and 99 in the upper part of the member. At the top of unit 84, small colonies of *Syringopora* are present. In many limestone units fossils are silicified. It is interesting that *Chaetetes*, which is common in the lower part of the Gray Mesa type section in the Lucero uplift (Krainer and Lucas, 2004), is completely absent in the Elephant Butte Member at the Arroyo de la Presilla. Conodont faunas with representatives of the *Idiognathodus obliquus* group, *Neognathodus bothrops*, and *Diplognathodus coloradoensis*, indicating an early

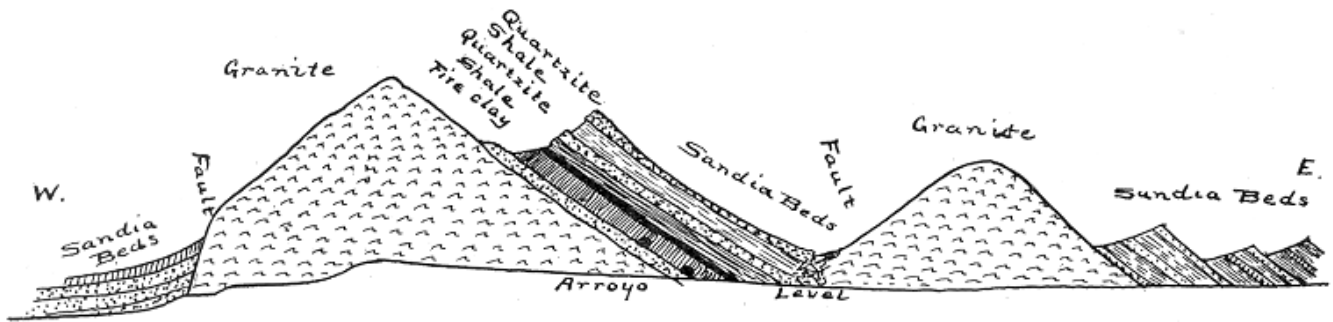


FIGURE 4. Herrick's (1904, fig. 1) geological cross section at the *Lepidodendron* locality. The plant-bearing bed is just below the "fire clay" above the Precambrian granite in the middle of the cross section.

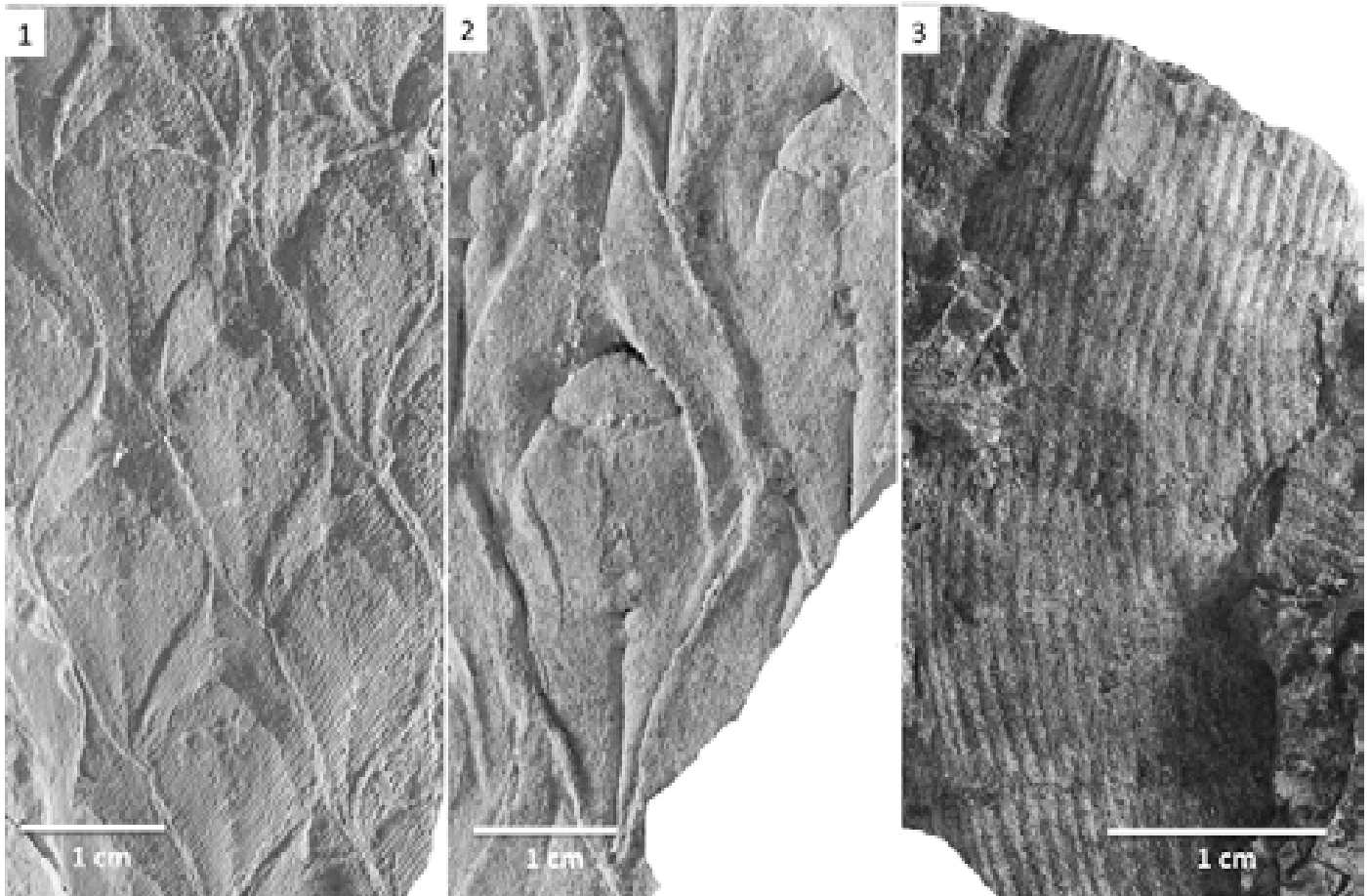


FIGURE 5. Plants from the Herrick "Coal Forest" site in the Smithsonian (USNM) collection. 1, *Lepidodendron aculeatum*, USNM specimen 536612, USNM locality 41896. 2, *Lepidodendron aculeatum* leaf cushion, USNM specimen 536614, USNM locality 41896. 3, *Calamites* stem, USNM specimen 536624, USNM locality 41896. Scale bars = 1 cm.

Desmoinesian age, occur in the the Elephant Butte Member at the Arroya de la Presilla (units 16-68; Lucas et al., 2009).

We recognize the Whiskey Canyon Member as an interval about 25 m thick of very cherty limestone beds, some with abundant rugose corals, near the middle of the section of the Gray Mesa Formation (Fig. 6). Thus, this member is composed of different types of bedded limestone and some covered (shale) intervals. The cherty limestone is composed of thin and wavy cherty limestone alternating with brownish cherty silty layers. It is fossiliferous with abundant crinoids, brachiopods, rugose corals, bryozoans and local fusulinid packstone. The member also contains thick-bedded algal and

crinoidal limestone, locally with minor chert, fusulinids and bryozoans. In the Whiskey Canyon Member (units 86-102), conodonts are uncommon, but members of the *Idiognathodus obliquus* group occur as do rare *Neognathodus asymmetricus*, a species indicating an early, but not earliest Desmoinesian age.

The Garcia Member (total about 94 m thick) is composed of different types of limestone, conglomerate, sandstone, shale and covered (shale) intervals. The most common limestone type is thick-bedded, locally cross-bedded crinoidal limestone. Individual units are 0.5 to 6.5 m thick. The crinoidal limestone contains some chert and locally also detrital quartz. Less abundant is thick-bedded, fossiliferous limestone containing algae,

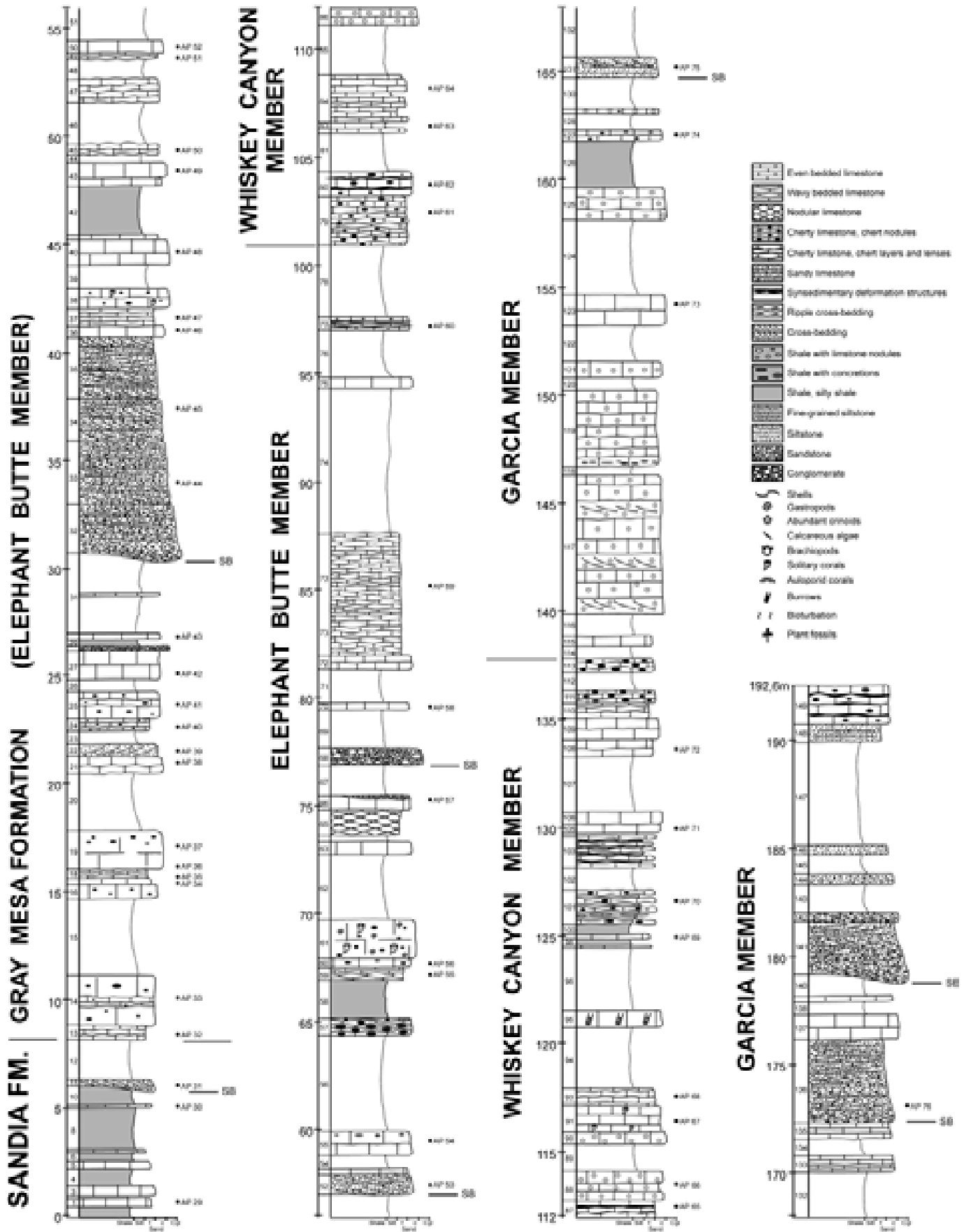


FIGURE 6. Measured section of most of the Gray Mesa Formation at Arroyo de la Presilla. After Lucas et al. (2009).

gastropods and some chert, and thin-bedded micritic fossiliferous limestone. Conglomerate/sandstone units are 0.4 to 2.5 m thick. Conglomerate beds are cross-bedded, and maximum clast size is about 2 cm, and they grade upward into cross-bedded sandstone. The sandstone is fine- to coarse-grained, arkosic, partly micaceous, partly pebbly, and displays trough cross-bedding, crude horizontal lamination or appears massive. The bases of the conglomerate and sandstone units are erosional, indicating sequence boundaries. The thicker conglomerate-sandstone units show a fining-upward trend, and are overlain by covered shale intervals and limestone on top, thus representing transgressive cycles. Compared to the quartzose conglomerate and sandstone of the Elephant Butte Member, the conglomerate and sandstone of the Garcia Member is arkosic and contains abundant granitic debris. One greenish-gray shale interval is exposed, the thickness is 2.1 m. Covered shale slopes are common and range in thickness from 0.4 to 4.8 m.

The lower part of the Garcia Member at this section yielded a small conodont fauna like that of the Whiskey Canyon Member. Starting at about 10 meters above of the base of the Garcia Member (units 122-133), a conodont fauna was obtained that included species similar to those from the middle Desmoinesian Verdigris cyclothem (upper Cherokee Group) in the Midcontinent and correlative strata in North America: *Idiogoniodus robustus?*, *I. iowaensis?*, and *Neognathodus medadulimus* (Lucas et al., 2009).

In the Cerros de Amado, we have found scrappy plant material in the sandstones and shales of the Garcia Member. The plants are mostly sticks, with occasional bits of foliage – none identifiable to date. The material is all allochthonous.

End of first day roadlog.

SECOND DAY ROADLOG – ATRASADO, BURSUM AND ABO FORMATIONS IN THE CERROS DE AMADO

Summary

Today's trip is short but mostly over an easily traversed, all-weather unpaved road. Its focus is on Middle-Upper Pennsylvanian strata and fossils of the Atrasado Formation, but also includes an opportunistic stop at an excellent section of the Lower Permian Abo Formation.

- 0.0 Bosquecito road junction; turn right and proceed south. 0.1**
- 0.1 Pavement ends. 0.8**
- 0.9 Turn left, to proceed east onto the “Quebradas Backcountry Byway” and cross cattleguard. 1.7**
- 2.6 On Quebradas Road, cross Arroyo del Coyote. Road is climbing hill, a bajada of Cenozoic alluvial deposits. 0.8**
- 3.4 Turn right at Quebradas Road sign, proceed about 0.1 mile to parking area.**
STOP 1. Overlook of Socorro, Rio Grande valley and distant mountain ranges (on a clear day). Ojo de Amado (spring) is along valley floor to southeast. View of geology south of spring is best in afternoon sun. Pennsylvanian bioherms are lined up here along the rift boundary fault. Proceed on foot down jeep trail to bottom of canyon, continue ahead (south) about 100 m. Here we will examine a superbly exposed section of much of the Middle-Late Pennsylvanian Atrasado Formation, overlain by a characteristic section (Fig. 7) of the early Wolfcampian Bursum Formation up to the base of the Abo Formation red beds:

Stop 1A. Rift-boundary fault. Here is the eastern major boundary fault zone of the Rio Grande rift. The zone juxtaposes upthrown Pennsylvanian sedimentary rocks on the east side of the arroyo with downthrown, steeply dipping, sheared Cenozoic volcanic rocks on the west side. The latter consist of volcanoclastic sandstone, conglomerate, debris-flow breccia, and minor mudstone of the lower part of the Spears Group, dated as middle to late Eocene (Cather and Colpitts, 2005). Overlying the Spears Group with an angular unconformity is horizontal gravel of the Santa Fe Group (Pliocene to Pleistocene).

Pennsylvanian rocks nearest to the boundary fault belong to the Amado Limestone Member of the Atrasado Formation (lower Missourian, Upper Pennsylvanian). Bedding is overturned, dipping 50 to 60° west. Eastward from this point, as we will shortly observe, bedding rotates through vertical to rapidly decreasing east dips with the strata right side up. The geometry of this overturned fold clearly is inconsistent with the sense and direction of Cenozoic down-to-the-west extensional faulting. The overturned fold reflects an early episode of east-verging (west side upthrown) compressional deformation, which undoubtedly took place during the Laramide orogeny (Late Cretaceous to Eocene). A series of smaller north-east-verging thrust faults and folds further represent Laramide deformation in this area (Cather and Colpitts, 2005).

This site also bears evidence of a third episode of structural deformation during the Ancestral Rocky Mountains (ARM) orogeny of Pennsylvanian time. A series of large algal bioherms (best viewed from the parking area in afternoon) in the Amado Limestone and older Gray Mesa Formation are closely aligned with the rift-boundary fault. These units rarely exhibit bioherms away from the fault. Moreover, a large bioherm occurs in the Council Spring Limestone Member close to the boundary fault a short distance southeast of where we are standing (Hambleton, 1962). Bioherms developed in relatively deep, quiet water as algae and other organisms built mounds, seeking to remain within the photic zone. Their alignment along a fault at several stratigraphic levels suggests recurrent episodes of subsidence along the fault. Walk up the arroyo toward Ojo de Amado, climbing a cow path on the north side to reach the top of the natural rock wall crossing the stream.

Stop 1B. Ojo de Amado overlook. We are standing atop vertically dipping limestone layers in the upper part of the Tinajas Member of the Atrasado Formation (Missourian, Upper Pennsylvanian). As we continue up the canyon, we will climb through the younger Council Spring, Burrego, Story, Del Cuerto and Moya members of the Atrasado (Missourian and Virgilian, Upper Pennsylvanian) (Fig. 7). Near the lowest limestone ledge is a sharp structural flexure, east of which the bedding dip rapidly declines to about 20°.

Proceed up the canyon, observing diverse rock types in the upper Atrasado Formation. The massive Moya Limestone has been deeply sculptured by flowing water, creating a small maze of potholes and widened crevices. Upon reaching the top of the Moya, climb the ledges on the north side of the canyon to reach the top of the Atrasado Formation.

At Ojo de Amado most of the Atrasado Formation (approximately 141 m) is well exposed, overlain by 24 m of Bursum Formation and Abo red beds (Fig. 7).

The section starts with the Tinajas Member, which is approximately 93 m thick and can be divided into three intervals:

- a) The lower 33 m are mainly composed of gray and brownish, and rarely reddish shale. Many intercalated limestone beds are mostly 0.1-0.3 m thick, and there is one

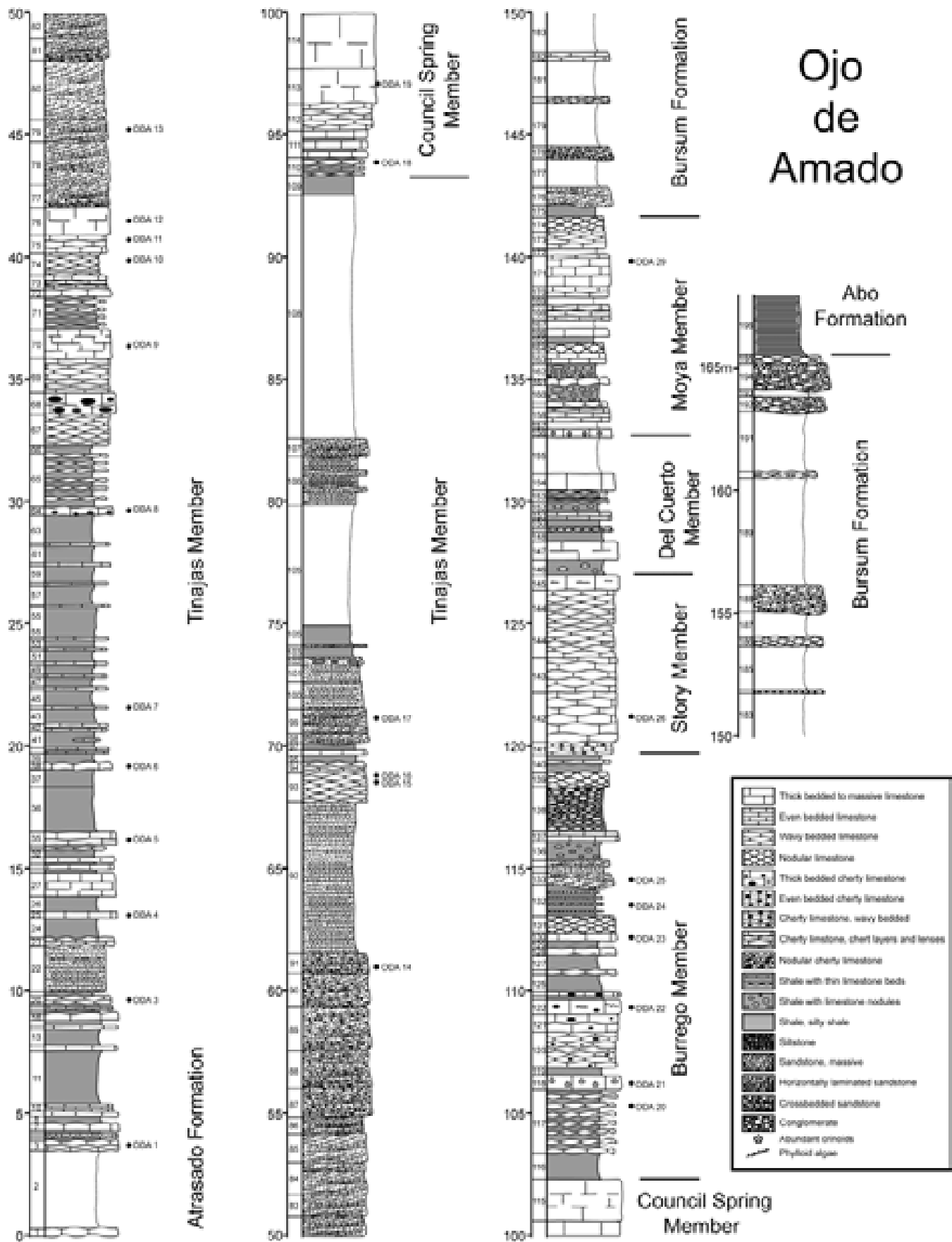


FIGURE 7. Measured stratigraphic section of much of the Atrasado Formation, the Bursum Formation and the base of the Abo Formation at Ojo de Amado.

thicker, bedded limestone interval (1 m). One limestone bed in the upper part contains chert nodules. Limestone is muddy and composed of mudstone, wackestone and rare floatstone. In the lower part, laminated calcareous siltstone (0.2 and 1.9 m thick) is intercalated.

b) The middle interval is approximately 10 m thick and dominantly limestone: wavy limestone beds alternate with thin, reddish-brown shale, nodular to wavy bedded limestone, thick bedded to massive cherty limestone and thick bedded to massive limestone. The microfacies is wackestone with a diverse fossil assemblage. Fusulinids occur in the upper part.

c) Above the middle limestone unit follows a 51-m-thick succession of mainly siliciclastic sediments with one thin limestone bed interval in the middle. Limestone is erosively overlain by 20 m of trough crossbedded, coarse-grained, partly pebbly sandstone (stacked sets 0.5 to 1 m thick), followed by pale green to gray, horizontally laminated silty shale and 2.1 m of thin, wavy to nodular bedded limestone (wackestone) containing fusulinids in the middle and upper part. This limestone interval is overlain by thin shale with a thin limestone bed intercalated. The next sequence starts with trough crossbedded sandstone that grades upward into horizontally laminated, fine-grained sandstone and greenish-gray shale with a thin sandstone bed intercalated, followed by a covered interval and 2.7 m of laminated siltstone with thin sandstone intercalated, and crossbedded sandstone. Above the sandstone is 10 m of cover, probably representing shale. The uppermost 0.8 m of the Tinajas Member is represented by greenish shale.

The Council Spring Member is approximately 9 m thick, starts with 0.7 m of thin wavy limestone beds (coarse crinoidal packstone/rudstone) and thin shale intercalations, followed by thicker bedded gray limestone (crinoidal wackestone/packstone), thin bedded gray limestone (wackestone) containing crinoid stem fragments up to 10 cm long, followed by massive gray limestone (mostly crinoidal wackestone/packstone) and thick-bedded limestone containing phylloid algae. The top of this limestone interval is probably a subaerial exposure surface.

The Borrego Member measures approximately 18 m and is composed of alternating limestone and shale. In the upper third thin massive carbonate sandstone beds and two crossbedded carbonate sandstone beds are intercalated. Shale is reddish and gray, and individual shale intervals are up to 1.1 m thick, mostly less than 20 cm. Limestone occurs as thin wavy beds (commonly 0.2-0.3 m thick), wavy-bedded cherty limestone, thick-bedded cherty limestone containing algae and fossiliferous nodular limestone. Common microfacies are wackestone, floatstone and rudstone with a diverse fossil assemblage. The most abundant fossil fragments are echinoderms, bryozoans and brachiopods. The crossbedded carbonate sandstone is composed of grainstone/packstone/rudstone containing strongly fragmented skeletons and many detrital quartz grains.

The Story Member is approximately 7 m thick and composed of light gray, wavy bedded limestone. The dominant microfacies is wackestone and floatstone (containing large skeletons of echinoderms, brachiopods and bryozoans). Individual limestone beds contain phylloid algae. The top is formed of a thicker limestone bed that also contains phylloid algae.

The Del Cuerto Member is ~5 m of alternating limestone, shale and thin covered (shale) intervals. Shale is pink, and individual shale units contain limestone nodules. Limestone occurs as thin beds (0.1-0.3 m) that are nodular or wavy, or as thicker limestone beds (0.6-0.8 m) that appear indistinctly

bedded to massive and contain gastropods, larger crinoid fragments and phylloid algae. Subaerial exposure surfaces are developed on top of two thicker limestone beds represented by thin, dark brecciated crusts and rhizoliths (?).

The Moya Member is thin, measures ~9 m and is composed of fossiliferous nodular limestone composed of limestone nodules and shale and wavy bedded limestone in the lower part, thicker bedded cherty limestone containing abundant phylloid algae (phylloid algal wackestone to floatstone) in the middle and wavy bedded and nodular limestone in the upper part.

The lower limestone beds of the Tinajas Member in the Ojo de Amado section (units 7-29) yielded a moderately abundant conodont fauna with common, but small, specimens of the middle Missourian/Kasimovian *Streptognathodus gracilis* Zone, mostly *S. gracilis* and *S. elegantulus* (Barrick et al., this volume). Higher in the Tinajas Member (unit 64), an extremely diverse and abundant *Idiognathodus eudoraensis* fauna is present. In addition to *I. eudoraensis*, ungrooved *Idiognathodus* species occur, as well as common *Streptognathodus firmus*, including specimens transitional to the older Eurasian species *S. isakovae*. A much reduced *eudoraensis* fauna continues up through unit 68. The middle of one limestone bed in the upper part of the Tinajas Member (unit 93), produced elements of *S. pawhuskaensis*, which ranges from the late Missourian into the Virgilian, and from the late Kasimovian into the Gzhelian.

A few small *Idiognathodus* elements, most of which can be assigned to *I. toretzianus* and *I. lobulatus*, were obtained from the lower part of the Council Spring Member (unit 111). The presence of the latter species suggests a early Virgilian and an earliest Gzhelian age for the Council Spring Member. In addition to a large number of *S. pawhuskaensis* elements, a few specimens of *S. ruzhencevi* and *S. vitali* were recovered from a unit 118 in the lower part of the Borrego Member; the latter species is the index for the early Virgilian/Gzhelian *vitali* Zone. At the base of the Story Member at the Ojo de Amado section (unit 141), specimens transitional from *S. vitali* to *S. virgilicus* appear, which indicate the *virgilicus* Zone. No diagnostic conodonts were obtained from the Del Cuerto Member. The *S. virgilicus* fauna continues into the lower part of the overlying Moya Member (unit 159), where morphotypes similar to the middle Virgilian species *S. holtensis* appear. No age-diagnostic conodonts were obtained from the upper part of the Moya Member nor the overlying Bursum Formation at the Ojo de Amado section.

There are plant fossils, mostly allochthonous in the Atrasado Formation here, depending on how far down the arroyo one progresses. The fossils are primarily allochthonous and occur in shales of the Middle Pennsylvanian/Desmoinesian Bartolo Member and the Late Pennsylvanian/Missourian Tinajas Member of the Atrasado Formation. Compositionally, the flora is dominated by forms found most commonly in seasonally dry climatic regimes, particularly conifers, *Sphenopteridium*, and *Charliea*, mixed with wetland elements, particularly neuropterid pteridosperms.

Stop 1C. Bursum and Abo outcrops. The Bursum Formation here is rather thin and composed of mostly covered intervals of probably red mudstone/siltstone (Fig. 7). In the lower part two coarse-grained, crossbedded sandstone beds are intercalated. Above three thin (0.1-0.3 m) nodular pedogenic limestone beds and one nodular marine limestone bed (0.4 m) containing crinoidal debris are exposed.

Above a thin covered interval a 1.1 m thick coarse-grained, crossbedded sandstone bed occurs that contains granitic detritus. The top of the Bursum Formation is formed by two

conglomerate beds. The lower conglomerate is 0.6 m thick, massive, poorly sorted, clast-supported and composed of abundant limestone clasts and sandy matrix. The upper conglomerate is 1.1 m thick, poorly sorted, clast-supported and contains abundant subangular grains of reworked granitic rocks with diameters up to 3 cm. This conglomerate is overlain by 0.3 m of nodular micritic limestone which forms the very top of the Bursum Formation. The Bursum (upper Virgilian to lower Wolfcampian, latest Pennsylvanian) consists of poorly exposed, gray and variegated non-fissile mudstone with a few layers of gray nodular limestone and dark gray weathering, coarse arkosic sandstone. Although the base of the Bursum is a regional unconformity, little relief is evident along the contact here. Near the top of the Bursum, large rounded quartz pebbles and fresh pink feldspar pebbles can be found. These probably were sourced from Precambrian granite in the Joyita uplift, about 20 km north of this point. The Joyita underwent tectonic uplift near the end of the Pennsylvanian, resulting in local truncation of the Pennsylvanian section with the Abo and Bursum Formations locally deposited directly on granite (Kottowski and Stewart, 1970; Krainer and Lucas, 2009). Overlying the Bursum with apparent conformity is the Abo Formation, which is composed of predominantly brick-red mudstone, siltstone, very fine sandstone, and intraformational conglomerate. We will not linger to observe the Abo here, because it is exposed much better at the next stop. **Return to the vehicles. After the STOP, continue to drive east on the Quebradas Road. 0.9**

- 4.3** Abo Formation outcrop along road, overlain by Yeso Group strata north of road. About here we cross the rift-bounding fault, which juxtaposes Permian red beds (Abo Formation-Yeso Group) on the east against Pliocene (Santa Fe Group) clastics. **0.4**
- 4.7** **Park on left side of road where narrow south-trending ravine crosses. STOP 2. Walk down the arroyo to the right to turn around and traverse Bursum-Abo-basal Yeso section here (Fig. 8).**

We are parked near the middle of the Abo Formation (middle-upper Wolfcampian, Lower Permian). This is one of the few complete, unfaulted sections of the Abo in the Socorro area, measuring ~213 m thick (Fig. 8). We will begin by walking down section, to the south, to observe the lower Abo, Bursum, and upper Atrasado formations. To return, we will walk north, up section through the upper Abo to the transition with the overlying Yeso Group. These same strata will be seen tomorrow on Day 3, Stop 1.

Walk down the small canyon to the south a distance of 500-600 m, observing the geology in passing. The canyon is fenced and posted by the Four Hills Ranch; permission should be obtained before visiting Stop 2A. The other highlights of this site, Stops 2B, 2C, 2D, and 2E, are on government land and may be freely visited, detouring around fenced private property.

Stop 2A. Lower Abo Formation. This little canyon affords continuous exposures of the lower half of the formation. Immediately south of the road, we descend through ledges of reddish brown coarse siltstone to very fine sandstone. Bedding is somewhat lenticular, frequently showing shallow channel geometry. Planar and wavy lamination, cross lamination, and small-scale crossbedding are prevalent. Some channels display low-angle accretionary bedding indicative of meandering streams. Interbedded mudstone is mottled and veined with light greenish to bluish gray and contains only rare carbonate nodules.

Downward the lithology changes to 80-90% mudstone that is dark reddish brown, very silty, and massive to blocky. Irregular small nodules, inclined stringers, and vertical cylinders (rhizoliths) of limestone are abundant. Near the base of the Abo, some layers of carbonate nodules are intergrown to form semi-continuous beds (calcrete ledges). Within the mudstone are layers and lenses of siltstone and very fine sandstone. Some such layers are tabular, whereas others show channel geometry. Some siltstone and sandstone beds are massive to blocky and show paleosol features, whereas others have planar to wavy lamination and small-scale cross-lamination along with simple horizontal burrows and mud cracks. Fossil conifers, dominantly *Walchia*, are abundant in the upper part of a sandstone ledge 58 m above the base of the Abo (near the midpoint of the canyon traverse) (Fig. 9).

Two types of conglomerate occur in the lower Abo. Confined to the lower 15 m is coarse, pebbly sandstone containing clasts of quartz and fresh, pink feldspar. As with similar conglomerate in the Bursum, the Joyita uplift is the probable source. Intraformational conglomerate composed of carbonate nodules and siltstone or mudstone clasts in a silty to sandy matrix also occurs in the lower Abo, filling small bowl- or scoop-shaped channels. **At mouth of canyon, follow the larger arroyo about 300 m northeast to a cut bank on the northwest side of the arroyo.**

Stop 2B. Bursum Formation. Here is a good exposure of the upper 10 m of the formation, which consists of intercalated mudstone and bedded to nodular mudstone. Features to observe include root traces near the top, arkosic limestone about 1.5 m below the top, and the thick paleosol in the lower part of the exposure. Large slickensides, limestone nodules and stringers, and rhizoliths are developed in the variegated mudstone. The Bursum is much thinner here than at Carrizo Arroyo and lacks the thick intervals of laminated, fossiliferous shale and siltstone that characterize the latter locality. Evidently, the Cerros de Amado area underwent much slower subsidence than Carrizo Arroyo during Bursum deposition. The alternating marine limestone and claystone paleosols seen here are reminiscent of Kansas cyclothems of the same age (West et al., 2010). **Walk approximately 400 m south-southwest, entering a small canyon cutting through the ridge.**

Stop 2C. Upper Atrasado Formation. Seen here is a northwest-dipping succession similar to that viewed this morning at Stop 1. The Bursum Formation is largely eroded off the dip slope; basal beds may be found along both sides of the canyon exit and higher on the slope northeast of the canyon, where underground copper mining has taken place. Within the canyon, upper Atrasado strata comprise alternating limestone units that form small cliffs and diverse clastic rocks that erode to slopes. Despite close proximity to other measured Atrasado sections, the authors of this field guide do not agree on the identity of the members. These rocks may comprise the Story (oldest), Del Cuerto, and Moya Members or, alternatively, the Council Springs, Burrego, Story, and lower Del Cuerto Members, the upper Del Cuerto and Moya having been eroded prior to Bursum deposition.

At the upper (east) end of the canyon, we cross a fault zone that brings Abo Formation into contact with Atrasado (southeast side down). This is normal faulting, presumably related to Cenozoic rifting. **Walk northeast, climbing part way up the ridge to the abandoned open-pit mine. Beware of open shafts and, as always, watch out for snakes!** **Stop 2D. Minas del Chupadero.** Small-scale mining of strata-bound copper took place here between 1958 and 1960,

yielding about 2000 tons (1800 kg) of ore that averaged 2% copper (Jaworski, 1973). Low-grade ore, mainly malachite, is disseminated in gray sandstone, siltstone, and silty shale. Although Jaworski identified the host rock as Moya, we consider it to be basal Bursum Formation. Fossil plants (mostly fragmentary) collected at this site include *Odontopteris* and other ferns, *Walchia*, *Cordaites*, and *Calamites*. They suggest wetlands surrounding a lake or other small body of water.

Return to the vehicles.

Stop 2E. Upper Abo Formation and basal Yeso Group. If time permits, walk up the small canyon north of the vehicles, keeping to the left where the canyon forks and continuing to the head of the ravine. Approximately 400 m north of Quebradas Road, hikers will reach a more gently sloping hillside, where the U.S. Bureau of Land Management formerly allowed citizens to quarry flagstone. This area marks the transition from the Abo Formation to the overlying Yeso Group (Leonardian; Lower Permian). Characterizing the Yeso is an absence of mudstone (common in the Abo) and a change in color and character of sandstone. Whereas Abo sandstone is almost uniformly brick red, sandstone of the Yeso may be white, light gray, yellow, light red to orange, and bluish to greenish gray. The sand is fine grained, well sorted, quartzose, and cemented by calcite. Straight-crested and ladderback ripple marks are common, whereas crossbedding and lateral accretion are not developed. Sandstone layers are typically tabular rather than channel form. Fossils of any kind are exceedingly rare. These changes may reflect an increasingly arid climate, lack of flowing streams, and an increase in sediment transport by the wind. Also note that tetrapod tracks are the most common fossil remains of the Abo Formation in this section (Figs. 8, 10). Footprints have been found at six discrete stratigraphic levels corresponding to bed 2A, 11, 17, 24, 32 and 34 (Fig. 8).

The NMMNH (New Mexico Museum of Natural History) collection houses about 100 specimens of fossil footprints from this site attributed to *Batrachichnus*, *Limnopus*, *Amphisauropus* (Fig. 10C), *Dimetropus*, *Tambachichnium* (Fig. 10E), cf. *Varanopus* (Fig. 10F), cf. *Hylodichnus* (Fig. 10G), and *Dromopus* (Fig. 10H). These tracks can be referred to small and large temnospondyls, seymouriamorphs, “pelycosaurs,” “captorhinomorphs,” and araeoscelids and are typical for the Abo red beds of central New Mexico (Lucas et al., 2013). *Batrachichnus*, *Limnopus* and *Dromopus* are remarkably abundant in the lower part of the section (bed 2A, 11, 12), whereas *Tambachichnium*, cf. *Varanopus* and cf. *Hylodichnus* are known only from the upper part of the section (bed 17, 32, 34). This distribution may reflect the phylogeny and ecology of Early Permian terrestrial tetrapods considering the radiation of “captorhinomorphs” during the late Early Permian (Voigt et al., 2009) and potential habitat preferences of certain trackmakers (Voigt and Lucas, 2012). Also, a taphonomic bias cannot be excluded.

In general, fossil-plant-bearing deposits are rare in the Bursum Formation. In the area of Quebradas Road most of the exposed terrestrial deposits of the Bursum are paleosols. However, in the area of the former Chupadera copper mine, sandstones are exposed that contain a scrappy flora of conifers, pteridosperms and tree ferns. Although not stratigraphically diagnostic, or even sufficiently preserved to characterize a Bursum assemblage, this flora does indicate that the mixture of wetland and seasonally-dry substrate plants continued in this region into the latest Pennsylvanian. A much better representation of Bursum plants is preserved to the north in the area that will be examined on Day 3 of this field trip.

Plant fossils occur throughout the Abo Formation in the Quebradas Road area (Fig. 9). The Abo Formation is characterized here, as elsewhere, by an exceptionally low diversity flora dominated almost exclusively by conifers of several kinds and the peltasperm *Supaia thinnfeldioides*. The next most commonly encountered kind of plant, *Autunia* (formerly *Callipteris*) *conferta* occurs much less commonly. Of the other 25-30 species known from the Abo Formation throughout New Mexico, nearly all are very rare, though they may appear as dominants at any given collecting site. Plant fossils are most commonly found in siltstone/fine-sandstone sheet deposits, most often in upper parts of these fining upward benches where, thin coarser-grained beds are clay draped; this often results in platy debris aprons on slopes below these beds. The plant fossils occur in association with invertebrate and vertebrate trackways, mudcracks, raindrop imprints and fine sedimentary structures.

After the STOP, continue east on the Quebradas Road. 0.3

5.0 After crossing small valley, road bends to right. Note tilted Yeso Group strata to left in fault zone that has aspects of transtensional or pull-apart. Cresting ridge, looking south at a mosaic of tilted blocks of Yeso on lower slopes with cliff of Atlasado behind. Major fault (down to NW) crosses mid slope. **0.6**

5.6 Concrete ford, crossing fault just mentioned. The trace fossil site reported by Minter and Lucas (2009) is to the left. This NMMNH locality 6708, an outstanding occurrence of trace fossils in the Joyita Hills east of Socorro. Minter and Lucas (2009) described a diverse invertebrate ichnofauna from this site including *Cruziana problematica* (Fig. 10I), *Diplichnites gouldi*, *Diplichnites* isp., *Monomorphichnus* isp., *Palaeophycus tubularis*, *Rusophycus carbonarius*, and *Striatichnium* cf. *S. natalis*. The assemblage is dominated by striated bilobate traces that probably have been produced by crustaceans. Previously referred to floodplain deposits of the upper Abo Formation (Minter and Lucas, 2009) the traces have been recently reinterpreted as shallow marine ichnia situated in the upper part of the Arroyo de Alamillo Formation, about 34 m below the base of the Los Vallos Formation (Lucas et al., 2013). **0.2**

5.8 Climbing toward gap in ridge. There is a fault along the bottom of the gully to the east (left). Mesa to east has Bartolo Member capped by Amado Limestone, whereas younger Tinajas and Council Spring crop out on hillside to west. The Tinajas black shale deposit is also exposed to the right. **0.4**

6.2 Crest of ridge. Loma de las Cañas on skyline capped by San Andres over Yeso and older units. In middle distance are west dipping upper Atlasado strata. To right is Tinajas Member capped by Council Spring Member. **0.2**

6.4 Bear right onto side road where Quebradas Road swings left, park. STOP 3.

The area south of the Quebrada Road preserves one of the best stratigraphic and paleofloral successions of the Desmoinesian-Missourian transition in central New Mexico. Numerous fossiliferous shales and sandstones crop out in Arroyo de los Pinos and its tributaries. The plant-bearing deposits represent a variety of depositional environments, nearly all of which are allochthonous deposited in nearshore brackish to marine conditions or, more rarely in terrestrial channel deposits. Among these settings are two autochthonous assemblages that contrast strongly in the environ-

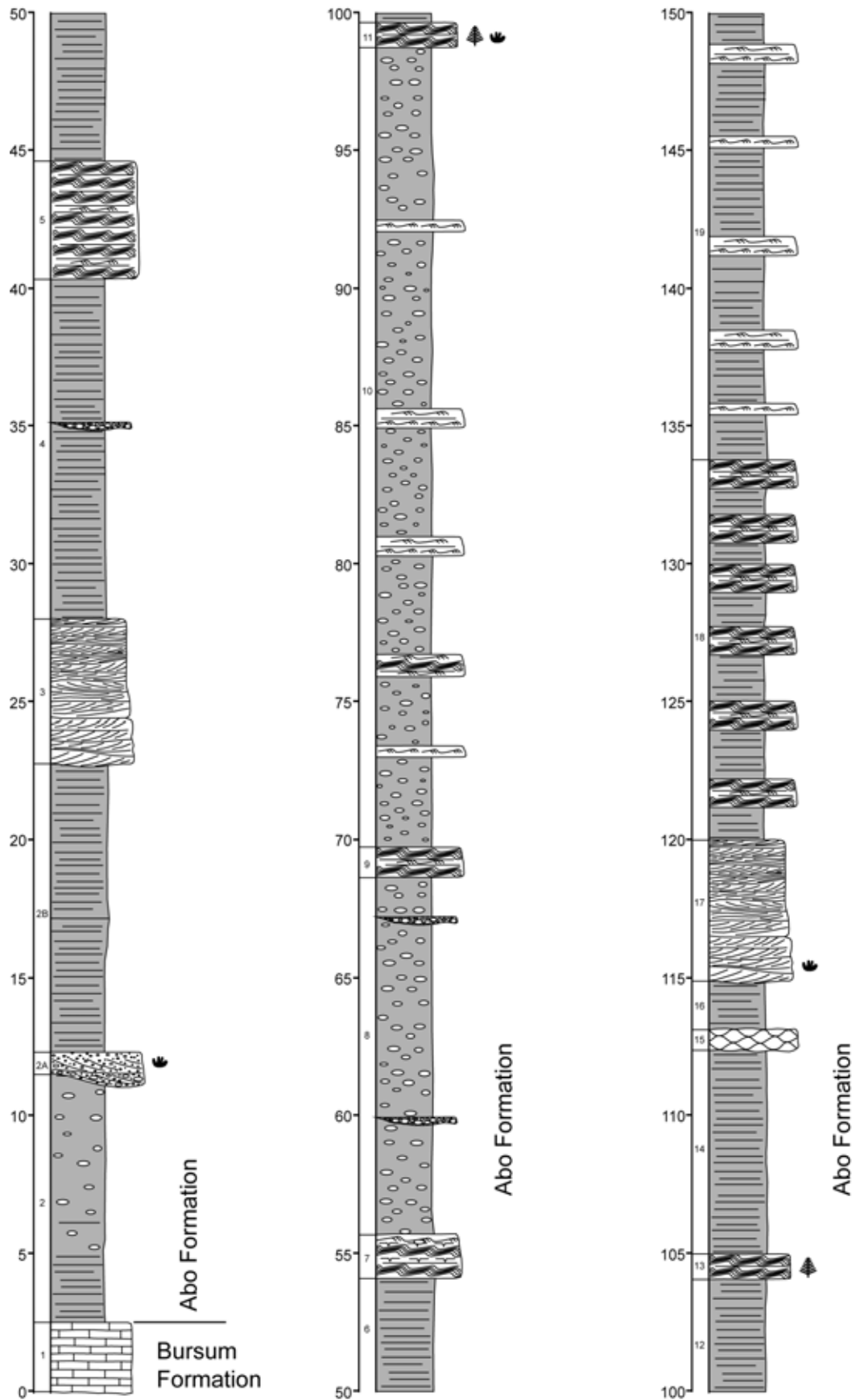


FIGURE 8. Measured stratigraphic section of the uppermost Bursum Formation, entire Abo Formation and lower part of the Yeso Group (lower part of Arroyo de Alamillo Formation) at Stop 2.

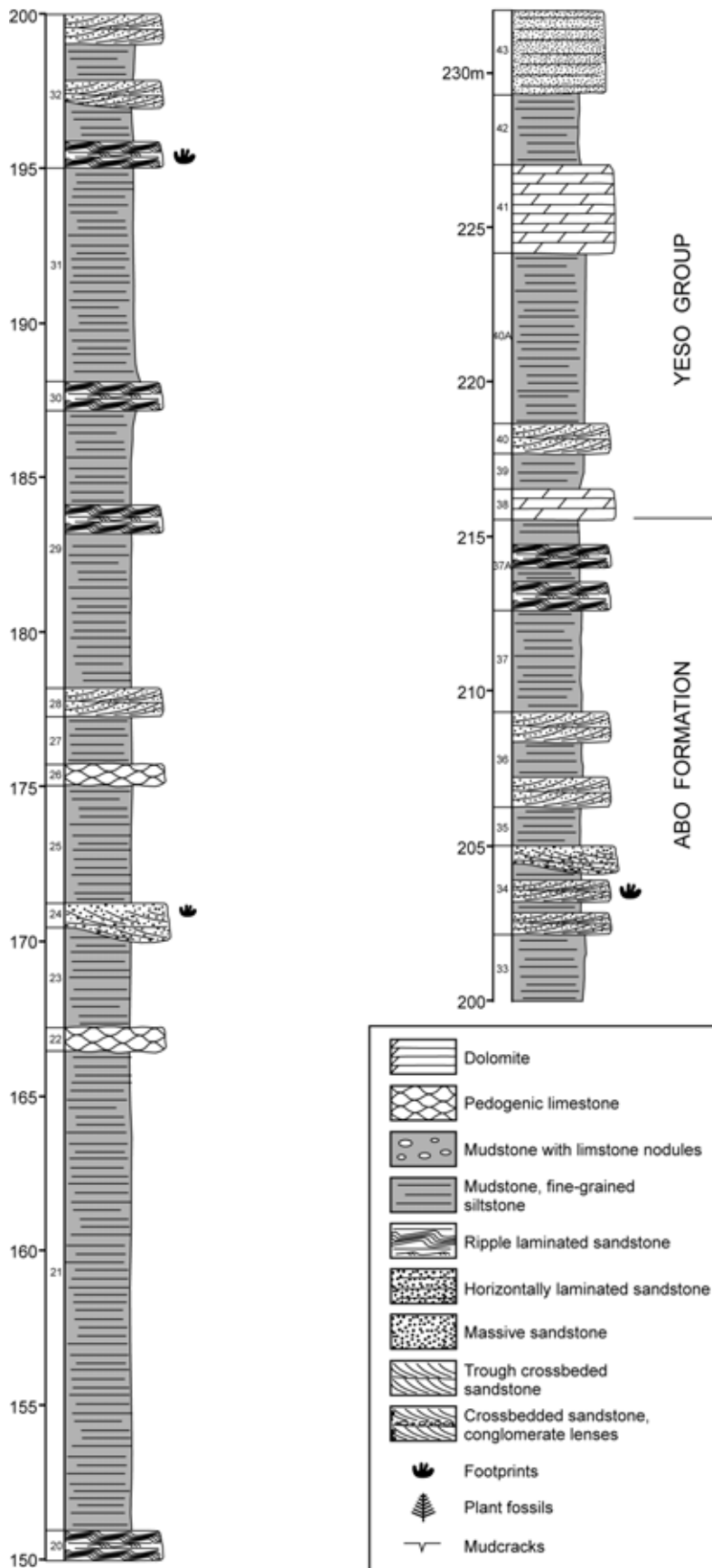


FIGURE 8. Continued. Measured stratigraphic section of the uppermost Bursum Formation, entire Abo Formation and lower part of the Yeso Group (lower part of Arroyo de Alamillo Formation) at Stop 2.

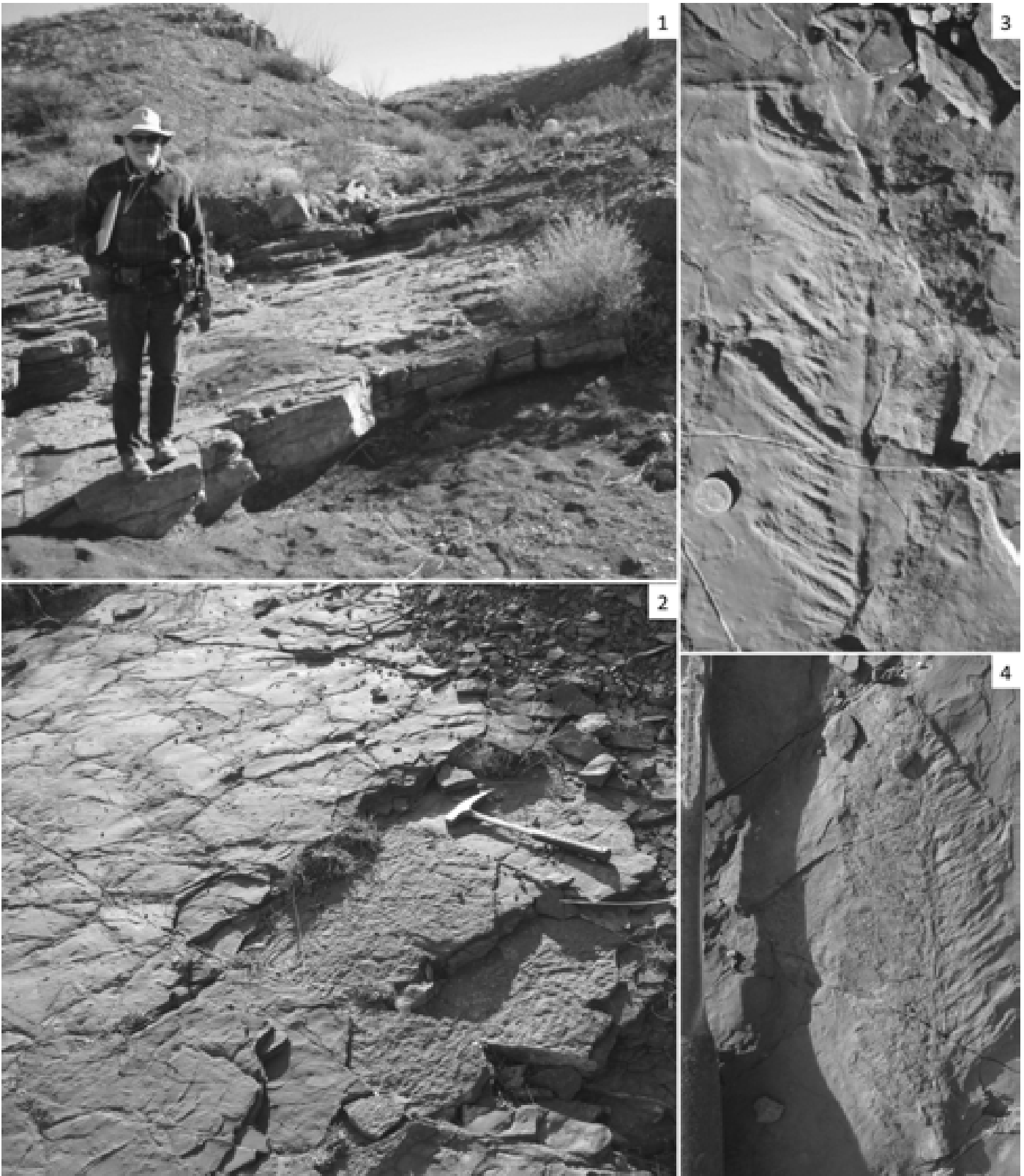


FIGURE 9. Abo Formation and flora in Quebradas Road Area. **1**, Outcrop of typical Abo Formation plant-bearing siltstone beds in ravine south of Quebradas Road. **2**, Plant-bearing siltstone beds south of Quebradas Road showing multiple bedding surfaces covered with conifer branch fragments. **3**, *Walchia* sp. branch fragment, field photograph. **4**, *Walchia* sp. branch fragment, field photograph.

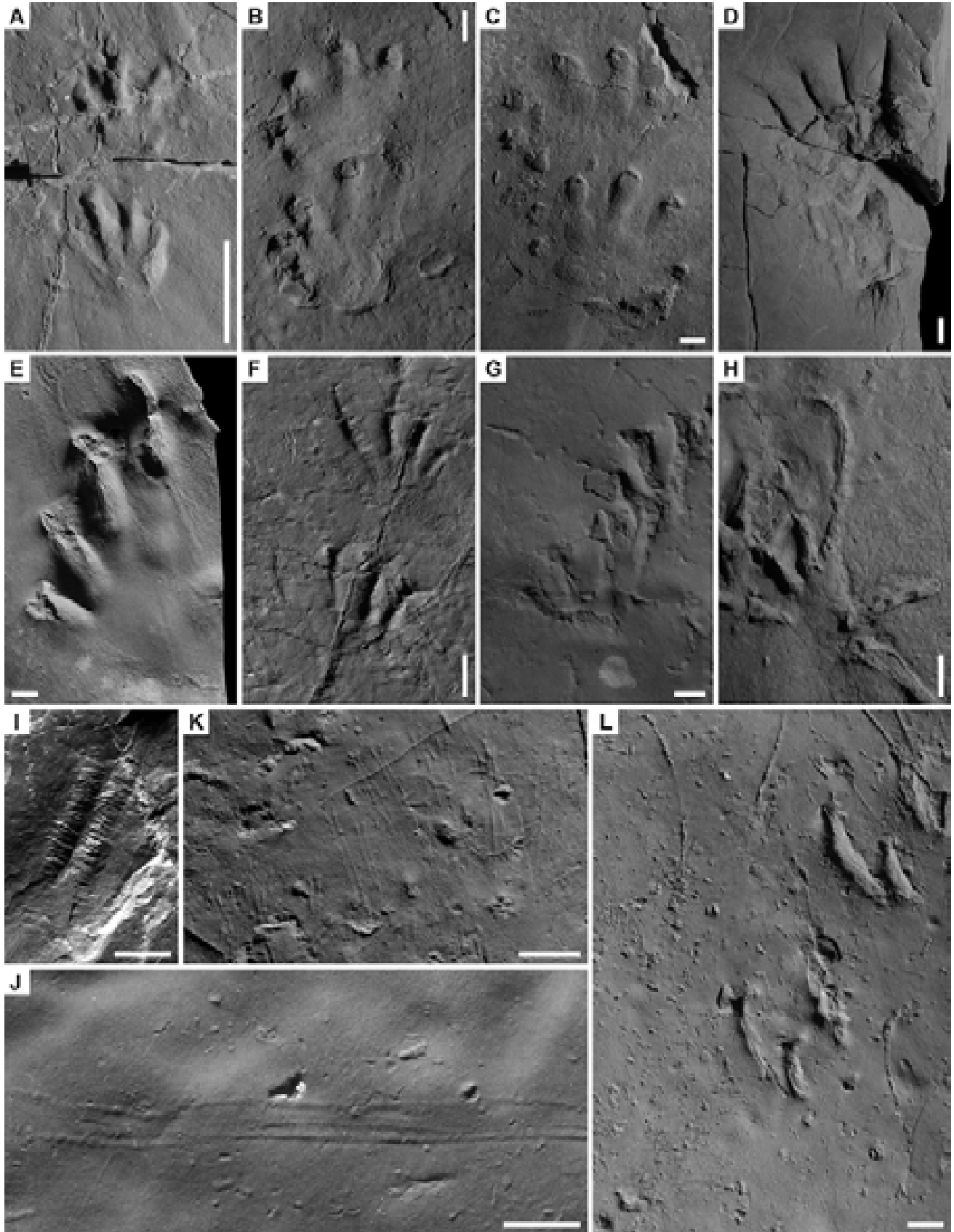


FIGURE 10. Invertebrate and vertebrate traces from the Abo (A-H) and Arroyo de Alamillo (I-L) formations in the Joyita Hills east of Socorro. A, *Batrachichnus salamandroides*. B, *Limnopus vagus*. C, *Amphisauropus kablikae*. D, *Dimetropus leisnerianus*. E, *Tambachichnium schmidti*. F, cf. *Varanopus* isp. G, cf. *Hyloidichnus bifurcates*. H, *Dromopus lacertoides*. I, *Cruziana problematica*. J, *Diplopodichnus biformis* with transition into *Diplichnites* isp. K, *Stiallia pillosa*. L, *Stiallia pillosa* and *Dromopus lacertoides*. Scale bars equal 1 cm.

mental settings: one, from the latest Desmoinesian is a coastal swamp deposit dominated by the seed fern *Neuropteris ovata*; the other, from the early Missourian is a sabkha deposit dominated by coniferous plants.

Fossil plants have been found from the lower part of the Atrasado Formation, in the Bartolo Member, beginning immediately above its contact with the Gray Mesa Formation, through the top of the overlying unit, the Tinajas Member, at its point of contact with the Council Springs Limestone (Figs. 11-15). This encompasses rocks of the uppermost Desmoinesian through the uppermost Missourian.

The Bartolo Member of the Atrasado Formation is of late Desmoinesian age. The following fossil-plant patterns are most notable in the Bartolo Member:

1. Basal Bartolo floras are rich in conifers and *Sphenopteridium*, both plants that have been linked to seasonally dry climatic conditions. Although allochthonous, these very fragmentary floras reveal that plants tolerant of seasonal moisture deficits were well established in the western parts of equatorial Pangea well back into the Pennsylvanian. In more central regions of Pangea, such as the coal basins of the eastern United States, such plants appear only as scraps, mainly during times of marine high-stand to regression (Falcon-Lang et al., 2009; Plotnick et al., 2009). Within the lower Bartolo flora are also highly fragmentary remains of wetland plants such as *Neuropteris*, calamitaleans, and pectopteroid tree ferns. This flora occurs intermixed with marine pectinoids and snails.

2. In the uppermost Bartolo, at three different sites, separated by more than a kilometer each, a flora is found that is dominated almost entirely by the pteridosperm *Neuropteris ovata*, a characteristic plant of wetland habitats. These

accumulations are autochthonous, containing abundant stems and large fragments of fronds, along with laminate foliage. They occur several meters below the Amado Limestone, suggesting that they may be low-stand deposits, which climate models (Cecil et al., 2003; Peyser and Poulsen, 2007; Horton et al., 2012) and sequence stratigraphic studies (Eros et al., 2012) have identified as the most likely times of very wet conditions throughout the Pennsylvanian equatorial region

The Tinajas Member of the Atrasado Formation is of Missourian age. It is lithologically complex and plants are preserved in a variety of different facies. The flora differs among these facies, but in general is much like that of the Bartolo Member, consisting of a mixture of wetland plants and those typical of more seasonally dry substrates. This mixture suggests that the background climate was seasonally dry and that wetland elements fringed coastal zones and riverine corridors under that same climate. The following plant patterns are most notable in the Tinajas Member of the Atrasado Formation:

1. Near the base of the Tinajas Member, a series of fossiliferous beds have been found that contain a conifer-dominated flora that appears to have grown in a coastal sabkha setting (Falcon-Lang et al., 2011). One of the most interesting exposures of this flora is a field of in situ, silicified stumps or tree bases of conifers or cordaitaleans rooted in a micritic limey mudstone, buried by dune deposits composed of calcium carbonate and gypsum grains (Fig. 13). The tree stumps may extend over a meter upward into these dune deposits, indicating tree burial during life or shortly after death. This deposit is widespread, also found north of the Quebradas Road, but is quite patchy in coverage, probably due to the rapidity

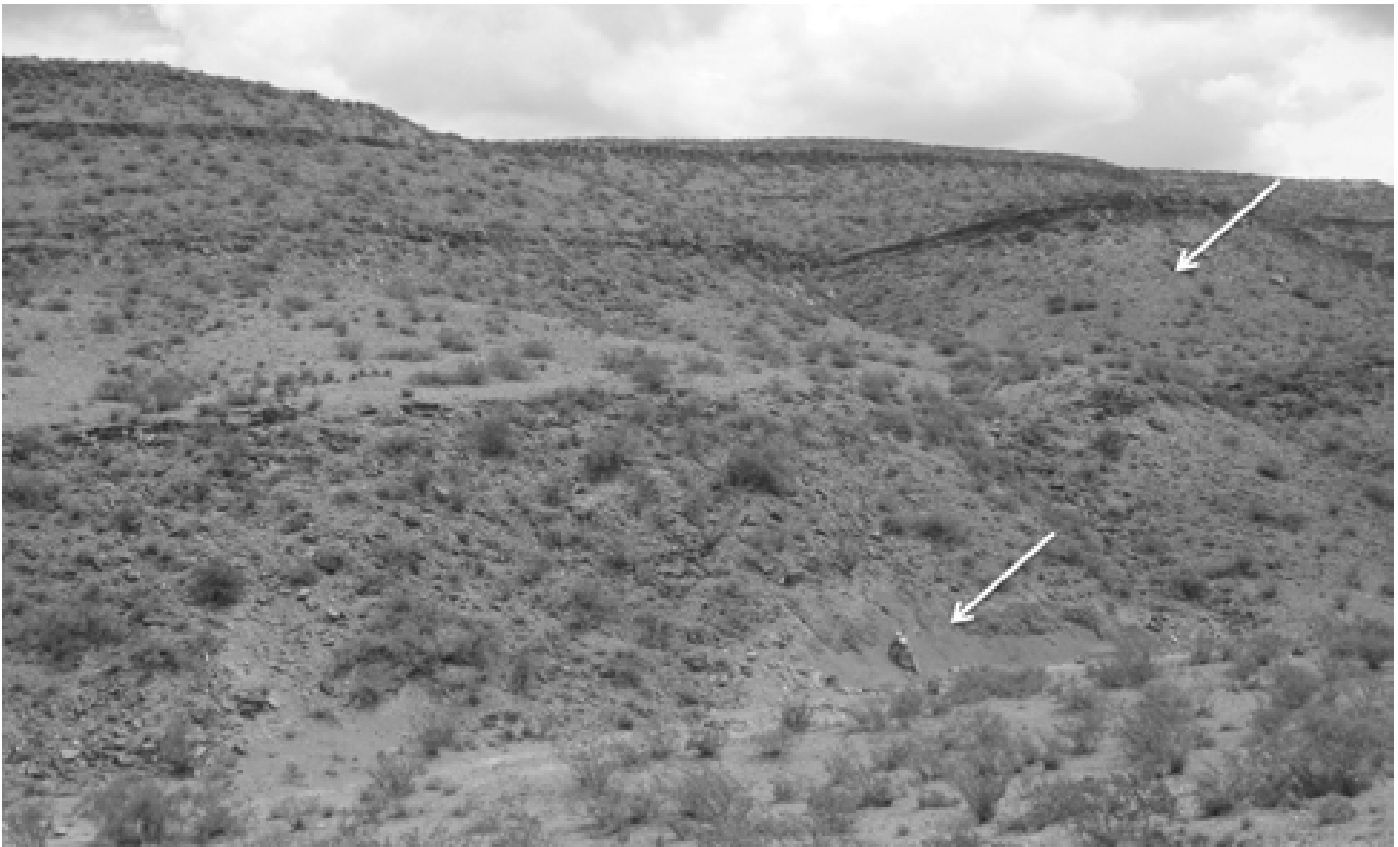


FIGURE 11. Bartolo Member, Fossiliferous Shales. This photograph illustrates two fossiliferous shale beds in the lower Bartolo Member of the Atrasado Formation (white arrows). The shales contain a sparse, allochthonous flora of mixed seasonally dry substrate and wetland elements.

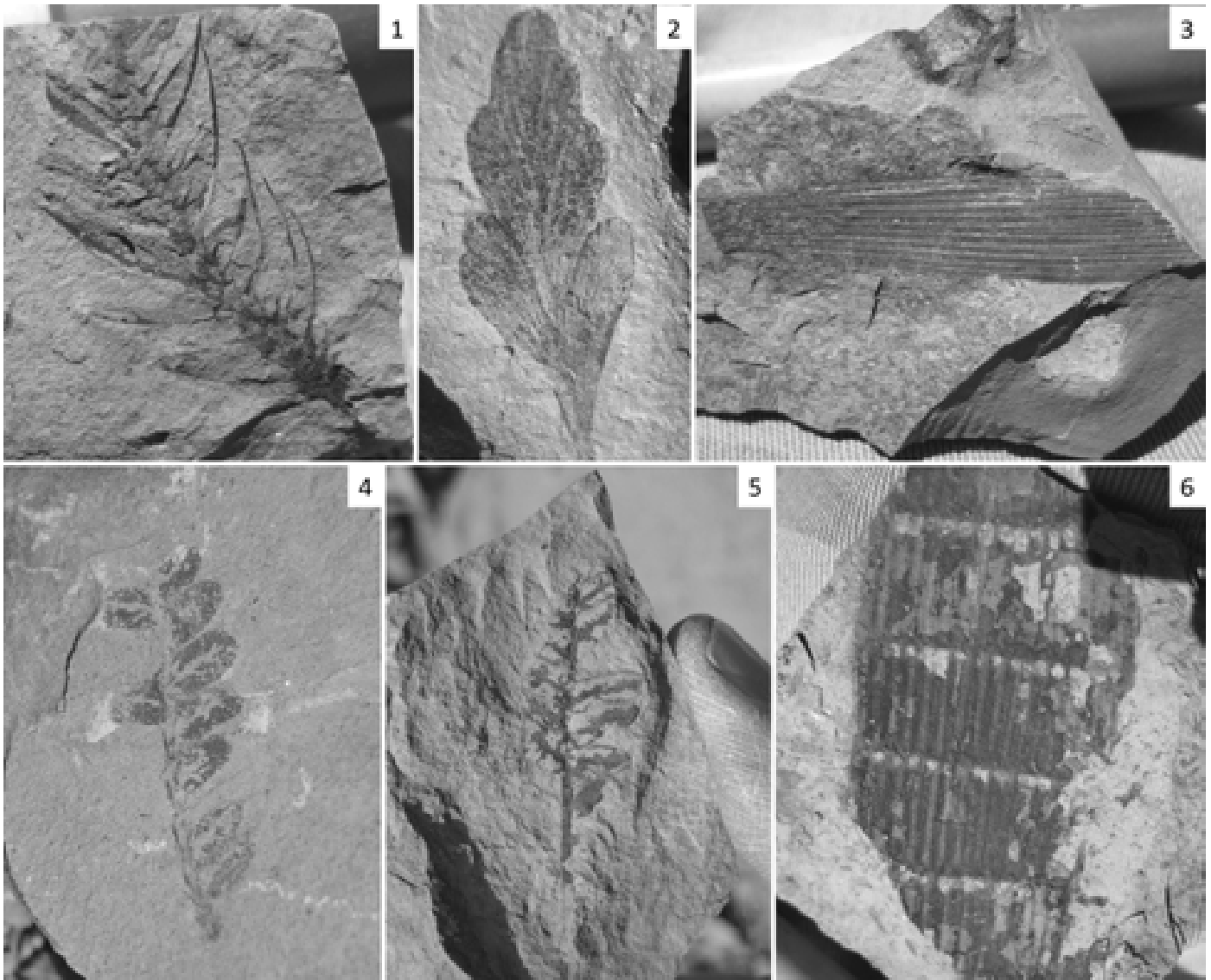


FIGURE 12. Bartolo Member, plant fossils from the lower part of the unit. 1, Walchian conifer, field photograph. 2, *Sphenopteridium* sp., field photograph. 3, Unidentified parallel veined leaf lamina, cf. *Plagiozamites* sp., field photograph. 4, Probable neuropterid pteridosperm foliage, field photograph. 5, Pecopteroid fern foliage, field photograph. 6, *Calamites* stem, field photograph.

with which it erodes. Tree stumps are larger in the more northerly exposures.

2. Above the sabkha deposit, in the middle Tinajas Member, a dark, organic-rich shale, containing distinctive “discus”-shaped siderite nodules, crops out throughout the area. This shale contains a sparse allochthonous flora of conifers, *Sphenopteridium*, the noeggerathialean *Charliea*, *Neuropteris ovata*, calamitalean foliage and tree fern foliage. Thus, again, the mixture of seasonally dry and wetland taxa is found, which characterizes both the Bartolo and Tinajas members.

3. A series of dark shales and olive shales crops out near the top of the Tinajas Member. The best exposed and most abundantly fossiliferous of these is the olive shale that crops out along Arroyo de los Pinos. This shale contains an allochthonous flora that, locally, is abundantly fossiliferous, and contains no evidence of brackish or marine conditions during deposition. The plants are scrappy but identifiable and, again, are the typically mixed seasonally dry and wetland

floras, although wetland species are more common in this deposit than in the other plant-bearing Tinajas units.

4. Throughout the Tinajas Member, at various levels, there are silicified fossil logs that probably floated out into the near-shore environments that represent most of the Atrasado Formation in this area. Analyses of these woods have shown them to be primarily cordaitalean/coniferous (Tidwell et al., 2000).

The Atrasado Formation flora is significant because it occurs throughout a section that crosses the Desmoinesian-Missourian boundary. Across this boundary the flora shows little change, the same basic elements occurring both below and above the boundary, in particular conifers, the pteridosperm *Sphenopteridium* and various wetland pteridosperms and ferns. A few new, and significant taxa, such as *Charliea*, an element of the seasonally dry flora, appear in the Missourian. This pattern of little change across this boundary stands in marked contrast to patterns well documented in more easterly coal basins of the U.S., where a major floristic turnover occurred at this time. Wetlands, in particular, the history and composition

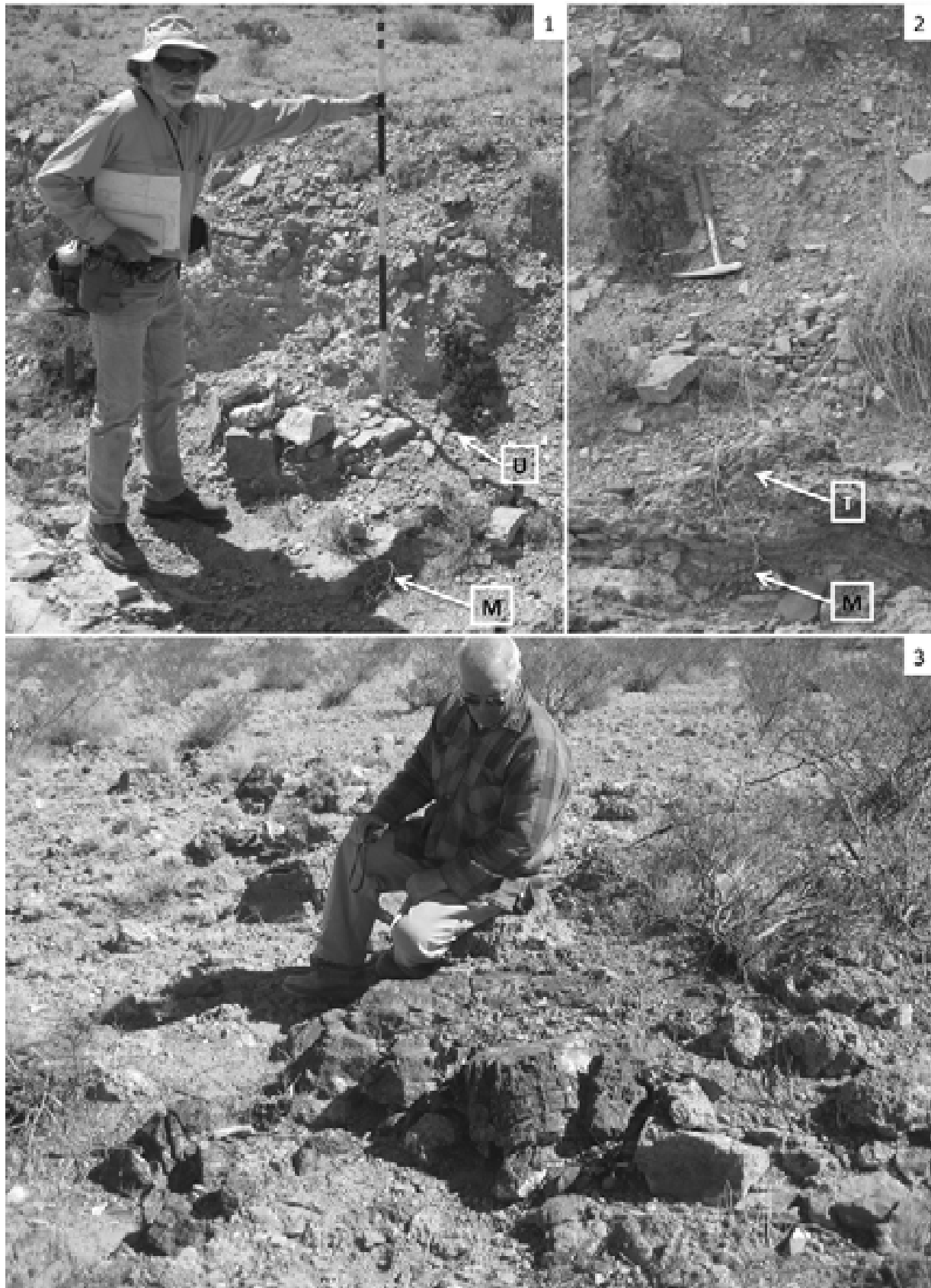


FIGURE 13. Tinajas Member, gypsum deposit with buried trees. **1**, Upright tree (U at arrow) surrounded by carbonate and gypsum bed. Gypsum deposit rests on a micritic limestone (M at arrow) and tufa bed (see 2). Jacob's Staff ruled in feet. **2**, Same upright stump as in 1, at hammer. Gypsum rests on a tufa (T at arrow) above a micritic limestone (M at arrow). Roots of trees are found in the micritic limestone. **3**, Large tree stump with roots extending down into the micritic limestone. Blocks of tufa can be seen scattered on the ground around the tree to the right.

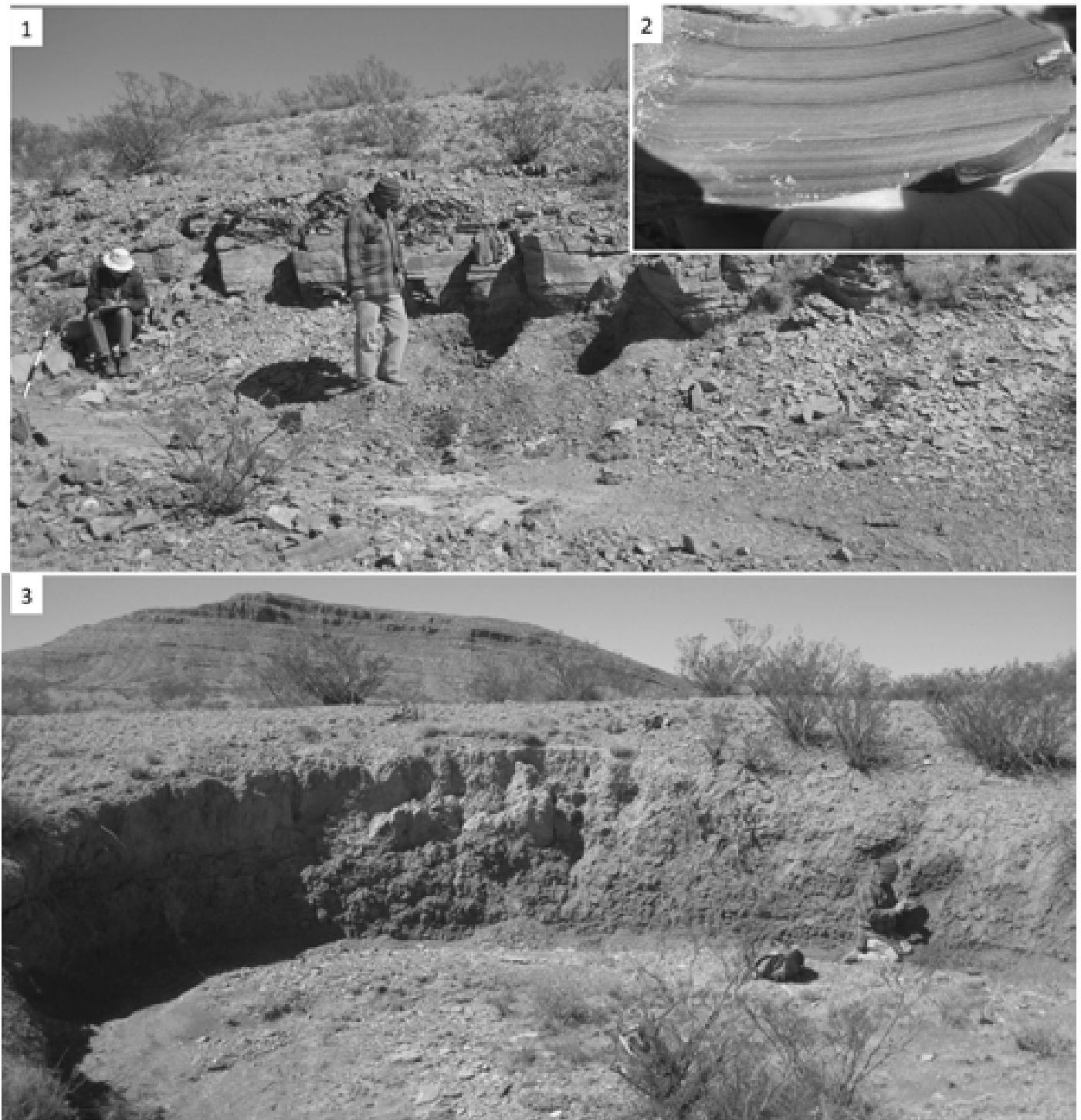


FIGURE 14. Middle Tinajas Member black, fossiliferous “disc” shale. This shale is widely traceable due to the distinctive “disc”-shaped siderite nodules it contains. It also contains a sparse but diverse, allochthonous flora and brackish water invertebrates. **1**, Outcrop exposure of shale. The unit is generally covered by talus. **2**, Detail of fine lamination of shale, indicative of accumulation in a quiet water setting, possibly lagoonal. **3**, Outcrop exposure of shale in small tributary of Arroyo de los Pinos.



FIGURE 15. Tinajas Member, upper units and fossil log. **1**, Uppermost Tinajas Member, Atrasado Formation exposure on Arroyo de los Pinos. Arrow points to highly fossiliferous gray shale (person for scale, just above point of arrow). Council Spring Limestone at top of hill marks base of Virgilian. **2**, Exposure of fossiliferous gray (GSh at arrow) and organic, black (BSh at arrow) shales in Arroyo de los Pinos. The black shale contains both flora and fauna, which have been described by Lerner et al. (2009). **3**, Silicified fossil log from the Tinajas Member of the Atrasado Formation. Such logs crop out at several levels in shales of this unit. Tape measure ruled in inches and centimeters.

of which is much better known than that of seasonally dry assemblages, underwent a major vegetational turnover, with many lineages disappearing or being greatly reduced in importance, particularly lycopsid trees, and others rising greatly in importance, particularly tree ferns (Phillips et al., 1974). The turnover in wetlands appears to have taken place during one glacial-interglacial cycle and to have encompassed as much as a 2/3 species turnover (DiMichele and Phillips, 1996).

Proceed on foot down side road, then cross wash to south about 300 m south of parking place.

Stop 3A. Silicified wood in Atrasado Formation.

Numerous masses of dark gray silicified wood can be found on the gently sloping pediment and adjacent gullies south of the main arroyo. Both stumps in growth position and prone logs have been observed in bedrock, which belongs to the lower part of the Tinajas Member of the Atrasado Formation (lower Missourian). Similar occurrences, including one described by Tidwell et al. (2000), have been encountered in the surrounding area, all from the same or closely similar

stratigraphic horizons. Surprisingly, fossil wood is encased in gypsum, itself a rare rock type in the Atrasado. For trees to grow on a sabkha seems to be a contradiction, yet close inspection reveals that the trees were rooted in limestone or shale and afterwards covered by gypsum. Identity of the wood is uncertain; it may belong to cordaitaleans or (more likely) to conifers. Falcon-Lang et al. (2011), who investigated this site, measured a tree density of about 100 individuals per hectare, signifying “open woodland with large canopy gaps.” Setting of the fossils and presence of growth rings point to seasonal aridity in a coastal sabkha setting. Trees may have grown during an episode of increased rainfall or in sites where a perched water table was present (Falcon-Lang et al., 2011). **Proceed down the main arroyo approximately 600 m in a westerly direction, pausing at a cut bank that exposes shale and Stop 3B.**

Stop 3B. Lacustrine shale outcrop. Lerner et al. (2009) described the paleontology of this site in detail (Fig. 16). The cut bank exposes Tinajas Member strata slightly younger than

those observed at the previous stop. The upper Tinajas Member, capped by light gray ledges of the Council Spring Limestone Member, is well displayed on the hillside immediately to the north. This is, in fact, the type locality of the Tinajas Member (Lucas et al., 2009).

The lower 5.6 m of the exposed Tinajas Member, units 1-5 below the black shale, are composed of greenish shale and micaceous siltstone to fine-grained sandstone. Silty shale contains fossil plants (seed ferns). Intercalated thin, nodular limestone beds are composed of fine-grained breccia containing subangular to rounded recrystallized micritic intraclasts floating in micritic matrix. The mudstone contains peloids and rare ostracods. Siltstone immediately below the black shale contains plant debris on bedding planes.

The siltstone is overlain by a 5.2 to 7 m thick interval of black shale which contains a few small carbonate nodules throughout the succession. The black shale is fissile in the lower part and laminated in the upper part and contains abundant conchostracans.

X-ray analysis of the black shale yielded calcite, quartz, kaolinite, ferroan phlogopite and clinochlor. In the shale abundant tubular structures occur within distinct horizons. Individual laminae are less than 0.1 mm thick and lense out laterally, locally the laminae may also be slightly folded.

The black shale is overlain by brownish-gray and greenish shale with a thin (3-5 cm) wavy limestone bed intercalated. The shale contains abundant *Dunbarella*. The thin limestone bed is composed of bioclastic wackestone containing few larger bioclasts. The most abundant fossils are brachiopod shell fragments and spines, subordinate crinoid fragments, ostracods and few smaller foraminifers. A few bioclasts are encrusted by cyanobacteria and *Palaeonubecularia*. The *Dunbarella*-bearing shale is overlain by two limestone beds (0.4 and 0.3 m thick), which are separated by a thin shale interval.

The thin fissile “paper shale” near the top of this unit contains abundant, densely packed, decalcified, complete valves of *Dunbarella*. The abnormally high density of the shells in this shale indicates that this may represent a mass-kill event, possibly from a sudden influx of marine water into the brackish environment in which these bivalves lived. The overlying thin bed of gray shale and profuse irregular encrusting algal growths containing a fauna which is dominated by the brachiopods *Derbyia* and *Crurithyris* marks the transition into marine conditions. Above follow 2 m of interbedded thin gray limestone and shale which contain a diverse, stenohaline fauna composed of abundant brachiopods *Hustedia* and *Hystriculina* and subordinate other brachiopods (*Juresania*, *Cancrinella*, *Rhipidomella*, *Composita*, *Punctospirifer*, *Beecheria*, *Cleiothyridina*, *Crurithyris*). The presence of the brachiopod *Chonetinella felmingi*, which is a widespread Missourian index species, indicates early to middle Missourian age. Limestone contains solitary rugose corals, fenestrate bryozoans and crinoid fragments, rarely molluscs and trilobites. The thin, sparsely fossiliferous gray limestone unit on top which contains a fauna essentially limited to solitary rugose corals, *Crurithyris* and crinoid debris, possibly represents a slightly hypersaline lagoon.

The lower limestone bed consists of bioclastic wackestone and locally of bindstone. The wackestone contains brachiopod shells and spines, crinoids, subordinate bryozoans, ostracods, rare smaller foraminifers (*Globivalvulina*, *Hemigordius*, *Planoendothyra*, *Syzranella*, *Tetrataxis*) and trilobite fragments. Locally abundant encrusting organisms are present (cyanobacteria and *Palaeonubecularia* transitional to

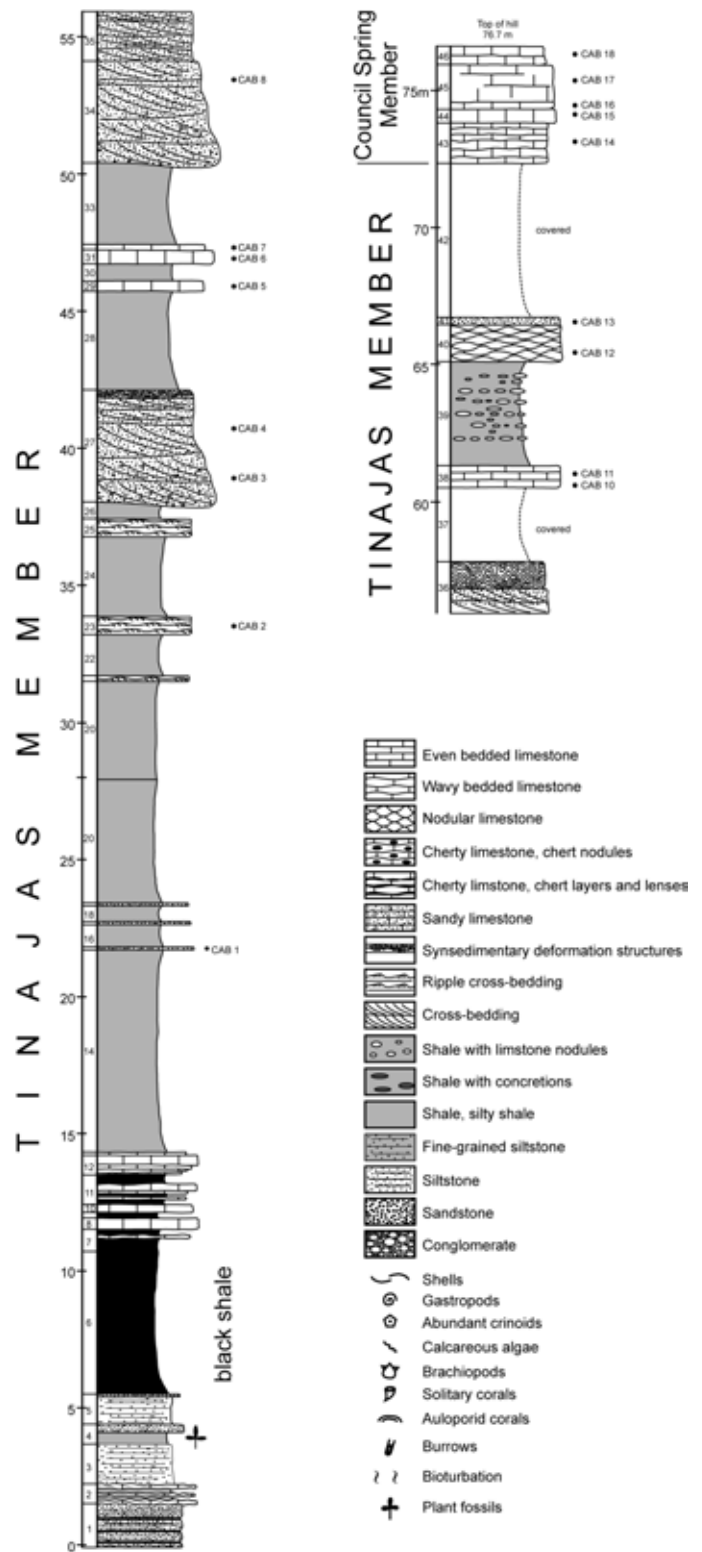


FIGURE 16. Measured stratigraphic section at the black-shale locality in the Tinajas Member (from Lucas et al., 2009).

Tubiphytes) forming patches of bindstone. The upper limestone bed is composed of coarse-grained crinoidal wackestone containing abundant crinoid stem fragments (mostly up to 2 mm, rarely up to 5 mm), subordinate brachiopod shells and spines and a few bryozoans, ostracods, rare trilobite fragments and very rare smaller foraminifers.

The limestone beds are overlain by a 1.1 m thick shale interval. Intercalated in the shale are abundant thin limestone lenses in the lower part and a 0.2 m thick limestone bed in the middle. The shale contains abundant brachiopods (*Dunbarella*).

The limestone is composed of bioclastic wackestone containing fragments of crinoids, brachiopods and bryozoans in varying amounts. Less abundant are small gastropods, ostracods, smaller foraminifers, bivalves and rare trilobite fragments. A few bioclasts are encrusted by cyanobacteria and *Palaenubecularia*. This shale interval is overlain by dark gray, micritic, bedded limestone containing few solitary corals. Microfacies of this limestone is bioclastic and intraclast wackestone. The bioclastic wackestone contains few large and abundant small skeletons which float in micritic matrix. Common skeletons are crinoids, brachiopod shells and bryozoans subordinate are gastropods, ostracods, brachiopod spines, smaller foraminifers (*Calcitornella*, *Globivalvulina*, glomospiroid *Miliolata*, *Hemigordius*, *Spireitlina*, *Syzrania*), corals and trilobite fragments. Locally abundant spicules are present. The intraclast wackestone contains abundant recrystallized intraclasts up to 10 mm, mostly up to 1 mm in diameter and skeletons of brachiopods, gastropods, ostracods, bryozoans, brachiopod spines and rare calcivertellid foraminifers.

The limestone is overlain by a thick succession of greenish, partly brownish shale to fine-grained siltstone, partly sandy in the upper part with ripple laminations and few 0.1 to 0.2 m thick fine-grained sandstone beds intercalated. The lowermost sandstone bed contains fossil plants (*Calamites*). In the upper part two fine-grained sandstone intervals with ripple lamination are intercalated, each 0.6 m thick.

Proceed downstream approximately 400 m, where shale and siltstone are exposed in the steep north bank.

Stop 3C. Plant-bearing siltstone exposures. The bank and hillside on the north side of the large arroyo reveals olive-gray silty shale and siltstone of the Tinajas Member, slightly younger than the black shale of the previous stop. Here and elsewhere in the vicinity, this unit has yielded a modest assemblage of fossil plants, together with uncommon bivalves and rare insect remains. At this exposure, the rock becomes finer-grained upward from stream level, reversing to upward coarsening in transition to the sandstone that forms broken ledges 12 to 15 m above the arroyo. Upward fining is atypical for a near-shore, bay-fill succession, such as this appears to be. Perhaps, this deposit developed in response to nearby tectonic uplift onshore or to offshore subsidence that outstripped sediment accumulation.

Stop 3D. Bartolo Member plant locality. A few hundred meters to the east, is an excellent example of an autochthonous accumulation of plants in the Bartolo Member (Fig. 17).

In this area, latest Desmoinesian conodonts occur in thin limestone beds of the upper part of the Bartolo Member and the basal bed of the overlying Amado Limestone Member of the Atrasado Formation at the Cerros de Amado A section (Fig. 17; Barrick et al., this volume). The Bartolo fauna is dominated by elements of *Adetognathus* and *Hindeodus*, but *Idiognathodus swadei*, *I. expansus*, *Neognathodus roundyi*, *N. expansus*, and *Swadelina* species occur. The latest Desmoinesian species *I.*

expansus and *Swadelina nodocarinata* also occur in the basal limestone of the Amado Limestone Member. In the middle and upper parts of the Amado Limestone Member, a small number of early Missourian species of *Idiognathodus* occur in addition to the more common *Adetognathus* and *Hindeodus* elements: *I. swadei*, *I. harkeyi*, *I. eccentricus*, and in the upper part, *I. gemiformis*, *I. corrugatus*, and *I. turbatus*. These species indicate an age range extending from the basal Missourian *I. eccentricus* Zone up as high as possibly the *I. cancellosus* Zone. The base of the international Kasimovian Stage, which has been tentatively placed at the first occurrence of the *I. turbatus*, lies within the Amado Limestone Member.

The next highest conodont fauna occurs in the middle part of the Tinajas Member, in limestone beds that overlie the lacustrine black shale in the Cerros de Amado B section. This fauna contains abundant elements of *Idiognathodus eudoraensis* and *Streptognathodus firmus*, the co-occurrence of which is indicative of the middle Missourian *eudoraensis* Zone. Samples from higher in the Tinajas yielded small collections of *S. pawhuskaensis*, a species that ranges from the late Missourian into the early Virgilian.

The Council Spring Limestone Member at this section yielded sparse faunas with specimens of *Idiognathodus toretzianus* and *I. lobulatus*, the latter species of which occurs in Virgilian strata in North America, and above the base of the international Gzhelian Stage. In the overlying Borrego Member, P₁ elements of *Streptognathodus ruzhencevi* and *S. vitali* appear, which indicate the middle Virgilian *S. vitali* Zone.

Return to vehicles. End of Day 2 roadlog.

THIRD-DAY ROAD LOG – FROM SOCORRO TO GALLINA WELL VIA THE ARROYO DE LA PARIDA AND THE CAÑONCITO DE LA UVA

Summary

The third day's trip crosses part of the Joyita Hills uplift to focus on Permian stratigraphy and sedimentation and on the locally complex structural deformation of the Permian (and adjacent) strata, which could be the result of various tectonic events, ranging from the late Paleozoic Ancestral Rocky Mountain orogeny through the Cretaceous-Eocene Laramide orogeny to the late Cenozoic Rio Grande rift. There is also a component of large-scale, low-angle to horizontal detachment faulting, for which a variety of explanations have been proposed. These structures are particularly well developed in the gypsiferous portions of the Yeso Group.

From the Bosquecito road junction, the route continues north across the Arroyo de la Parida, one of the principal drainages on the west flank of the Joyita Hills uplift. The route then heads east, across the margin of the Rio Grande rift, into an extensive outcrop belt of Permian sedimentary rocks that are only mildly deformed, mostly by normal faulting. Stops 1 and 2, close to each other, include the type section of the Yeso Group. These stops allow us to observe the entire Permian section exposed in central New Mexico, more than 400 m thick here. We thus begin in the uppermost Pennsylvanian Bursum Formation and continue through the Lower Permian Abo Formation, Yeso Group (Arroyo de Alamillo and Los Vallos formations), Glorieta Sandstone and San Andres Formation. These rocks record the last pulses of the Ancestral Rocky Mountain orogeny as well as the final collapse of the late Paleozoic Gondwana ice sheets, and our examination of these strata focuses on how and what they record of these events. At Stop 3, we examine the Abo Formation at the Gallina Well vertebrate fossil site and the Bursum Formation nearby.

0.0 Bosquecito road junction; turn left and proceed north. 0.2

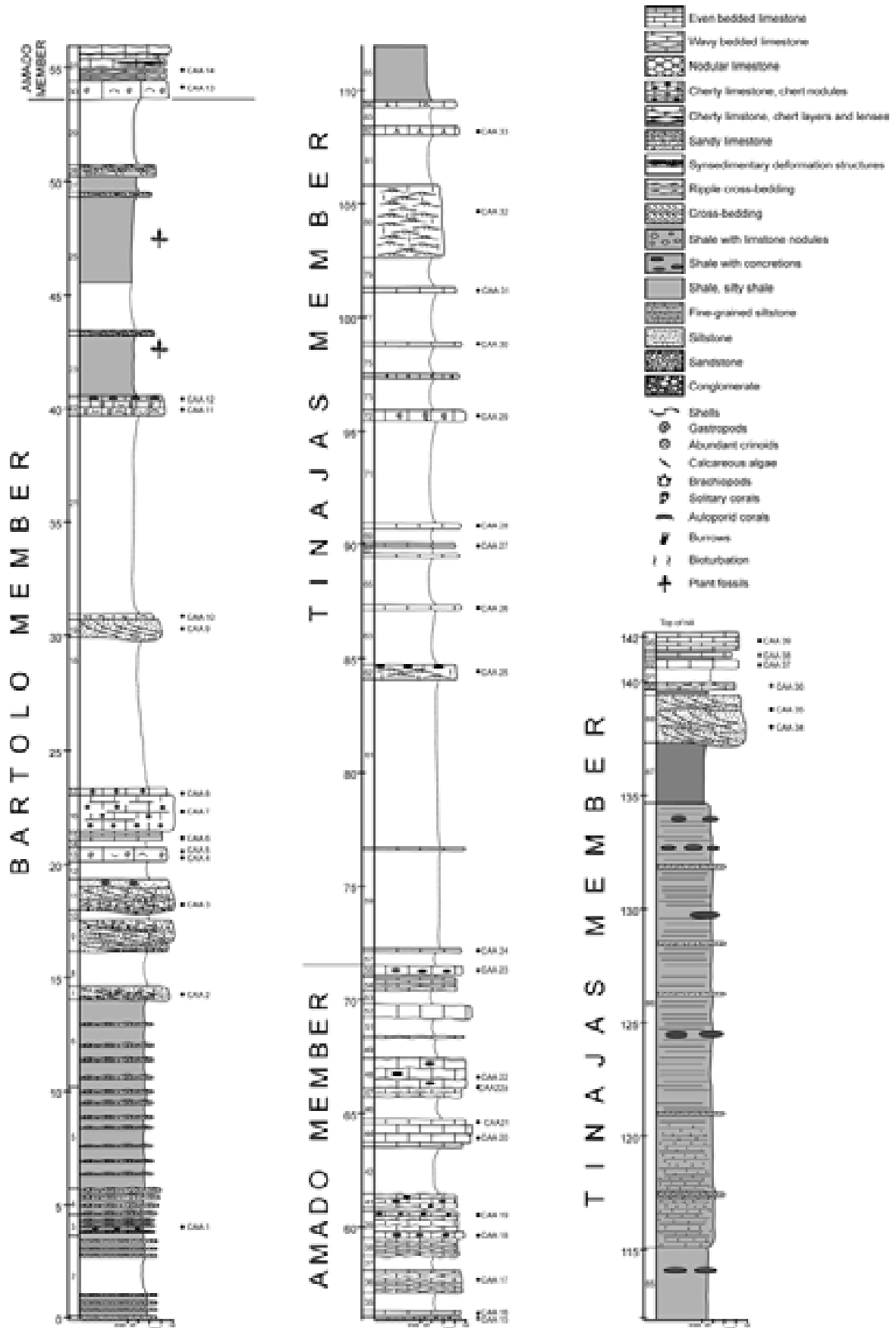


FIGURE 17. Measured section of part of the Atrasado Formation at the Cerros de Amado A section of Lucas et al. (2009). Note the plant localities in the upper part of the Bartolo Member.

- 0.2** Cattleguard, road forks, **bear right**. Johnson Hill is straight ahead. Note lower slopes are composed of tan and gray beds of the axial river facies of the ancestral Rio Grande. These beds are covered by reddish piedmont gravels. **0.7**
- 0.9** Crossing the Arroyo de la Parida. The Santa Fe Group forms red cliffs to the right. A fault that juxtaposes piedmont sediments against axial river facies of the ancestral Rio Grande is visible near the west end of the cliffs. **0.1**
- 1.0** Intersection with Johnson Mesa Road, **stay right**. **0.5**
- 1.5** Ascending the “real” Johnson Hill, road forks, **keep right**. Climbing Johnson Hill, our route approximates part of the Ocean-to-Ocean Highway, the first transcontinental automobile route in the United States, which opened in 1910 (Smith et al., 1983). **0.4**
- 1.9** To left, note redbeds of piedmont facies of Sierra Ladrones Formation of Santa Fe Group (mostly sedimentary conglomerates, breccias and sandstones). Color of these beds is imparted by clasts of Lower Permian Abo Formation and Yeso Group to the east, which have been stripped and transported by late Cenozoic alluvial fans to become the piedmont facies of the Sierra Ladrones Formation. **0.2**
- 2.1** Good exposure of piedmont facies in roadcuts on left. Note the abundant tabular clasts of Abo and Yeso sandstone and of Pennsylvanian limestone. East to west transport direction is indicated by west-facing clast imbrications. **0.6**
- 2.7** Crest of Johnson Hill, road to left, bear right. Road is still very narrow with blind curves and wash-outs. Roadcuts are still in piedmont red-bed facies of Sierra Ladrones Formation. Proceed cautiously. **0.5**
- 3.2** Road curves to left. Cerros de los Coyotes at 12:00 are a northeast-dipping hogback of Laramide origin. Looking to right, note excellent exposures of piedmont facies of Sierra Ladrones Formation in cut banks along Arroyo de la Parida. **0.9**
- 4.1** Crest hill. Road on uppermost piedmont surface of the Sierra Ladrones Formation that grades toward the Rio Grande. At 9:00-9:30, are west dipping hogbacks of Oligocene ash flow tuffs derived from the Socorro caldera complex. The skyline to the east consists of Paleozoic sedimentary rocks. Also in view are Mesa del Yeso at 11:30, type section of the Yeso Group; Sierra de la Cruz at 12:00, and La Cebolla at 1:00. We are driving eastward on the very eastern edge of the Socorro basin of the Rio Grande rift, about to cross the rift-bounding fault, which is buried by alluvium here. **0.5**
- 4.6** Cattleguard. **0.5**
- 5.1** Begin descent into the Valle del Ojo de la Parida. Cenozoic ash-flow tuff units at 11:00, Manzano Mountains at 12:00 with gasoline pipeline station in foreground, Los Pinos Mountains at 12:30-1:00, Mesa del Yeso at 1:00, Sierra de la Cruz at 2:00, and La Cebolla, a prominence of Sierra Larga, at 2:30. **0.3**
- 5.4** Driving on alluvium of Valle del Ojo de la Parida, road to pump station on left, continue straight. Low hill next to pump station is composed of middle Cenozoic volcanic Spears Formation. **0.3**
- 5.7** Fork in road, **bear right**. The road to left leads to the Palo Duro Canyon area of the Sevilleta National Wildlife Refuge. **0.6**
- 6.3** Road ascends hill of Los Vallos Formation of the Yeso Group (note characteristic salmon-colored siltstones/fine sandstones in roadcuts on left). **0.1**
- 6.4** Water tanks on hill at right. Good view of Mesa del Yeso at 12:00. **0.6**
- 7.0** Road sign on left. When the sign is flipped open, the road is closed for missile testing; as on rare occasions, missiles have escaped the bounds of the White Sands Missile Range to the south. Unnamed hills straight ahead are composed of Yeso Group strata. Mesa del Yeso at 10:30 is a west-dipping homocline. **0.2**
- 7.2** Intersection with road to right. Continue straight. **0.7**
- 7.9** Note Los Vallos Formation outcrops on both sides of road. **0.1**
- 8.0** Cattleguard. The Lower Permian Glorieta Sandstone forms cliffs that are overlain by limestone and dolomite of the Lower Permian San Andres Formation capping the hill at 11:30. Redbeds dipping northerly at 10:00 are lower part of Yeso Group (Arroyo de Alamillo Formation of Lucas et al., 2005). **0.6**
- 8.6** Crossing normal fault that juxtaposes Yeso Group (west side, hanging wall) against Abo Formation. **0.3**
- 8.9** Road curves right, pull to right for **STOP 1**. Faulted hill to the north is Yeso strata capped by Glorieta-San Andres. Sierra de la Cruz at 12:00 composed of Yeso Group. We are stopping very close to the base of the Yeso Group here, which is the base of the Arroyo de Alamillo Formation of Lucas et al. (2005). **We will walk from here approximately 1.5 km to the Bursum Formation outcrops exposed along the Cañoncito de la Uva to the south** (Fig. 18). We will then walk back through a complete, though faulted, Abo section (Fig. 19). Points of interest include:
1. The Bursum Formation is characteristically mixed marine and nonmarine strata in this area (Fig. 20) above the entirely marine upper part of the Atrasado Formation and below the entirely nonmarine Abo Formation. About 25 m thick here, conglomerates in the upper part of the Bursum Formation yield a pelycosaur-dominated vertebrate fossil assemblage. Finer grained strata yield a flora that includes cordaitaleans, conifers and peltasperms (Fig. 21).
 2. The Abo Formation is relatively thin here, about 100 m thick (Figs. 19, 22), and its two members are readily recognized. Thus, the lower 42 m of the Abo are mudstone-dominated slopes with some trough-crossbedded sandstone and conglomerate beds, the Scholle Member of Lucas et al. (2005b). The overlying 58 m of the Abo contain numerous beds of ripple- and climbing-ripple laminated sandstone characteristic of the Cañon de Espinosa Member. This change in fluvial architecture within the Abo Formation can be explained by a stalling of subsidence or a change to a drier climate.
 3. In the lower part of the Cañon de Espinosa Member we examine a tracksite with a conspicuous *Limnopos* trackway. It is NMMNH locality 5811 that predominantly yields footprints of *Batrachichnus* (Fig. 10A), *Limnopos* (Fig. 10B), *Dimetropus* (Fig. 10C), and *Dromopus*. These tracks can be referred to small and large temnospondyl, “pelycosaurian”-grade, and araeoscelid trackmakers. *Batrachichnus*, *Limnopos*, *Dimetropus*, and *Dromopus* are typical elements of Early Permian coastal plain tetrapod ichnofaunas (Lucas et al., 2011; Voigt et al., 2013) suggesting that this part of central New Mexico was a nearshore lowland environment during deposition of the Cañon de Espinosa Member.
 4. As elsewhere in central New Mexico, Abo floras of both the Scholle and Cañon de Espinosa members in the Cañoncito de la Uva area (Fig. 22) are overwhelmingly dominated by conifers (primarily *Walchia*, *Otovicia* and *Culmitzschia*) and

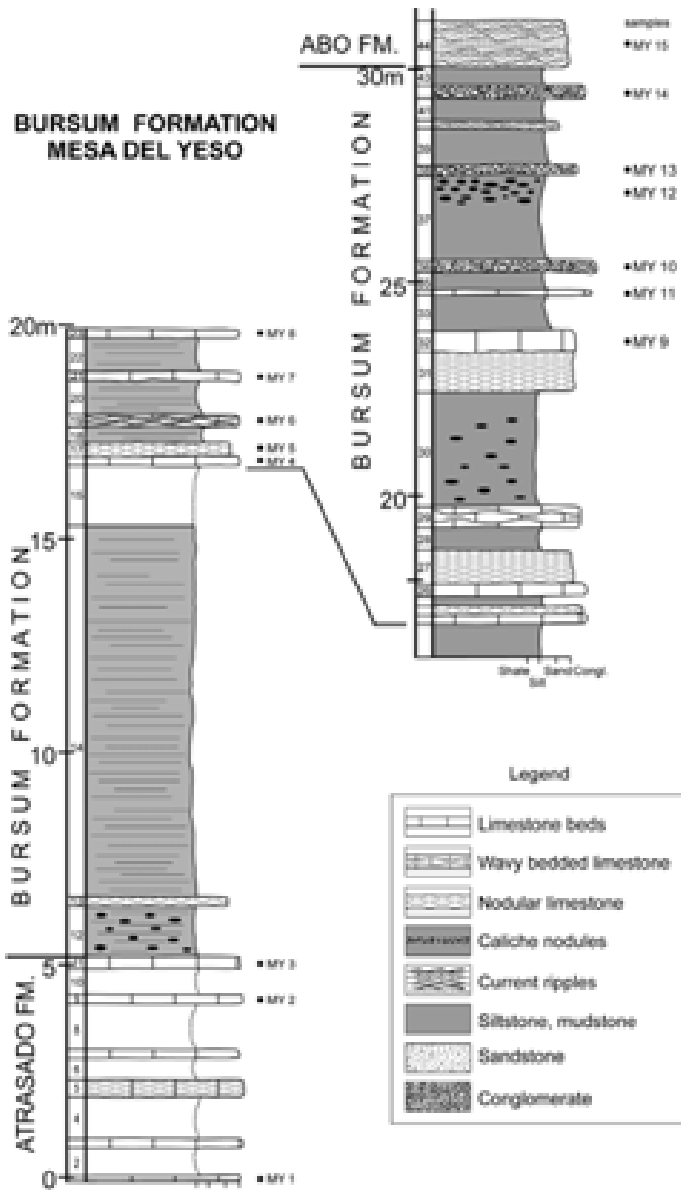


FIGURE 18. Measured section of the Bursum Formation at Cañoncito de la Uva.

the peltasperm *Supaia thinnfeldioides*. Smaller numbers of occurrences have been noted of *Autunia (Calipteris) conferta*. In the Scholle Member, plants are often found disposed across bedding planes or in troughs, indicating transport and deposition in active, high energy channels. In both members, and almost exclusively so in the Cañon de Espinoso Member, the plants are found most commonly in the upper portions of sheet sandstone-siltstone tabular units, where clay drapes and laminae are interbedded with coarser-grained siltstones. The plants often occur on surfaces that also contain trackways of vertebrates and invertebrates, mudcracks and other sedimentary features. The Abo Formation flora from Cañoncito de la Uva has been described in detail and illustrated by Hunt (1983) in one of the first full descriptions of an Abo flora, also placed in a larger context.

5. The Abo-Yeso contact (Fig. 23) is gradational (conformable) but can be picked out here as a distinct change in color and lithology from red-bed mudstone/siltstone and ripple-laminated/crossbedded sandstone of the Abo to mostly greenish

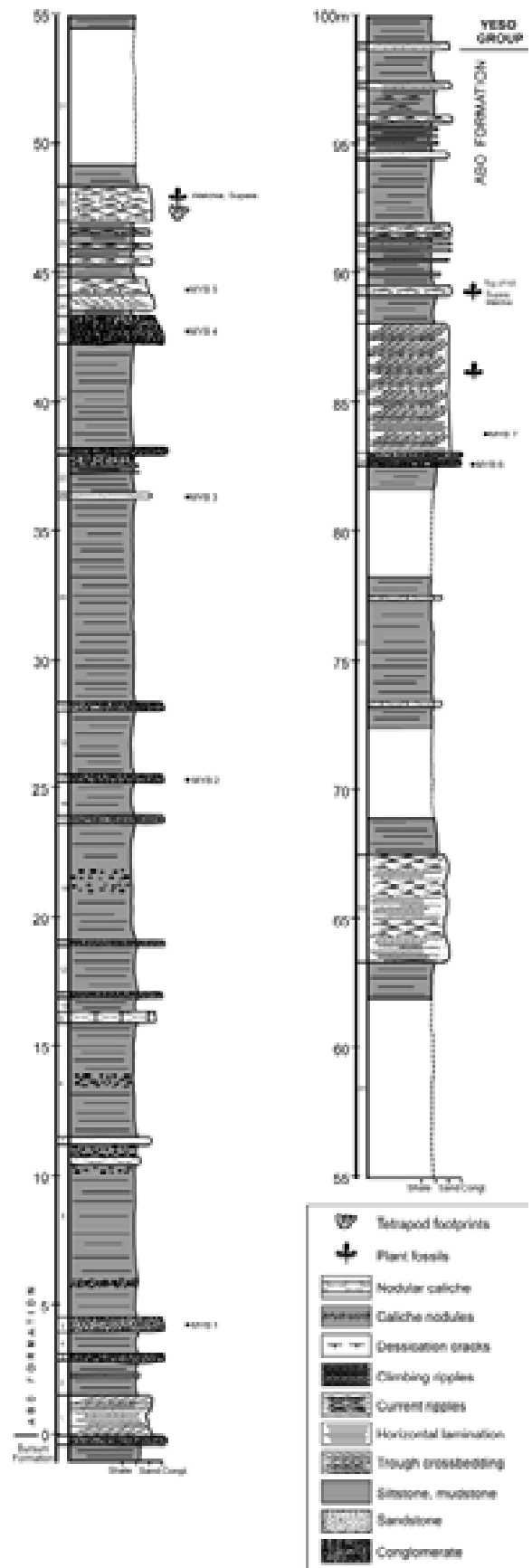


FIGURE 19. Measured section of the Abo Formation just north of Cañoncito de la Uva.

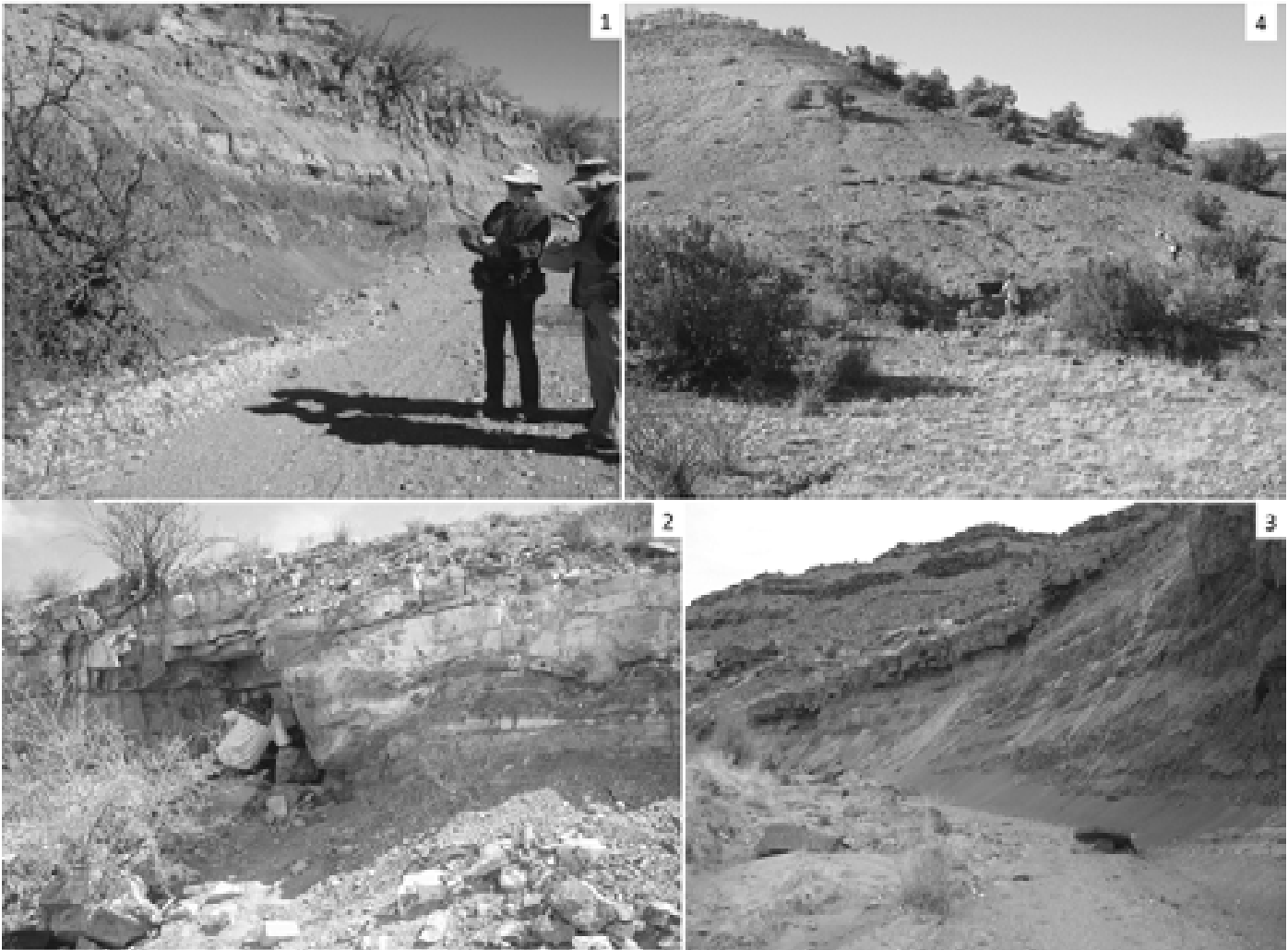


FIGURE 20. Bursum Formation outcrops. **1**, Typical Bursum Formation exposure. Terrestrial beds are represented primarily by paleosols, such as the one that crops out at the bottom of the exposure. Cerros de Amado area. **2**, Bursum Formation, Minas de Chupadera. Plant fossils come from the more finely bedded gray sandstones at the base of the exposure. **3**, Bursum Formation-Abo Formation contact, Carrizo Arroyo. The sharp contact shown in this image is typical of Bursum-Abo contact throughout the area of its exposure, including in the Socorro region. **4**, Fossiliferous Bursum Formation exposure. Thinly bedded to finely laminated gray siltstone and claystone, probable lagoonal environment. Plants at multiple levels, generally large and well preserved, indicated little transport prior to burial. Canoñico de la Uva.

silty mudstone and siltstone with thin intercalations of sandy limestone and siltstone beds with abundant halite pseudomorphs at the Yeso base.

6. The halite pseudomorphs, siltstones, thin limestones and sandstones of the Arroyo de Alamillo Formation at the Yeso Group base are arid coastal plain deposits that formed seaward of the De Chelly erg of northwestern New Mexico and northeastern Arizona (Fig. 23).

After stop continue east on road through lower Yeso strata. 0.1

9.0 Road almost washed out to right, be careful. **0.6**

9.6 Road crosses arroyo, a tributary of Cañoncito de la Uva. **STOP 2 (pull off to left)**. Approximate location of Abo-Yeso contact. Proceed up arroyo to view Yeso-Glorieta-San Andres section. To examine the type Yeso Group section (Fig. 23), we will walk through this outstanding type section, which was first described by Needham and Bates (1943) and later described in detail by Lucas et al. (2005). Along our traverse contemplate the obvious cycles and possible sequence boundaries present in the Yeso. Clearly, Yeso deposition here took

place during a time of relatively little tectonism on the coastal plain northwest of the West Texas Permian basin (Fig. 24). In the Permian basin, Ross and Ross (1988) identified three major Leonardian-age transgressions that should be evident in the Yeso. Probably these equal the base of the Arroyo de Alamillo Formation, the base of the Los Vallos Formation and the thick limestone with oncoids near the top of the Torres Member.

The Yeso Group comprises two formations, the upper of which is divided into three members (Fig. 23). The lower formation, Arroyo del Alamillo, is about 107 m thick. The lower part of the formation is composed of massive to laminated mudstone and siltstone alternating with thin, tabular sandstone beds in a variety of colors. The upper part contains thicker and more numerous sandstone layers that are more uniformly reddish brown and exhibit larger scale bedforms, such as crossbedding. This upper portion can be mistaken for the Abo Formation when encountered in faulted areas.

The Los Vallos Formation, 225 m thick in the type section, contains the Torres Member (oldest), the Cañas Gypsum member, and the Joyita Member. The Torres, 156 thick, is

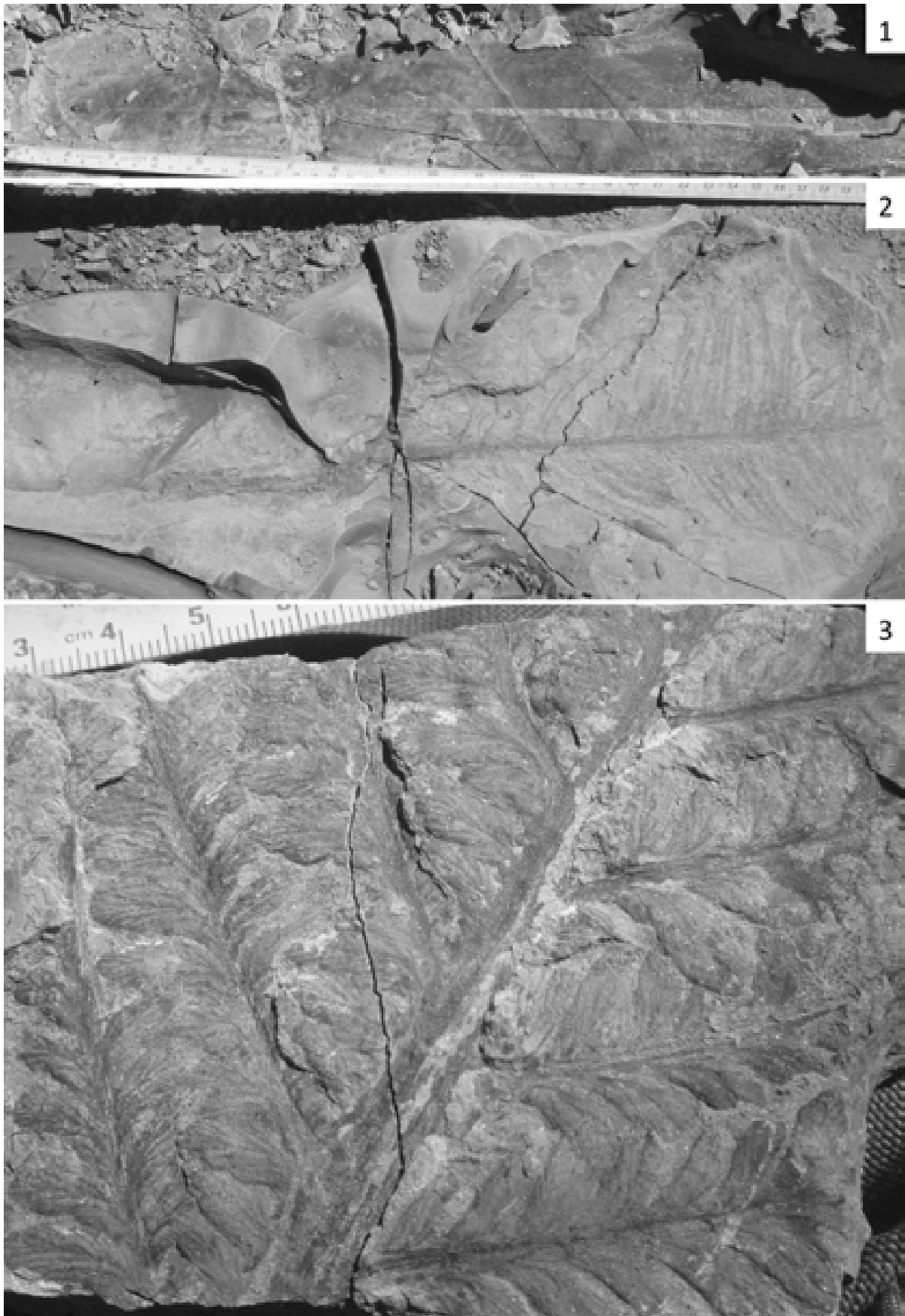


FIGURE 21. Bursum Formation plant fossils from Canoñcito de la Uva. 1, *Cordaites cordaitalean* foliage showing tip of a large leaf. Field photograph. 2, *Walchia* conifer branch, field photograph. 3, *Rhachiphyllum*, a callipterid peltasperm, field photograph. Metric scales ruled in mm.

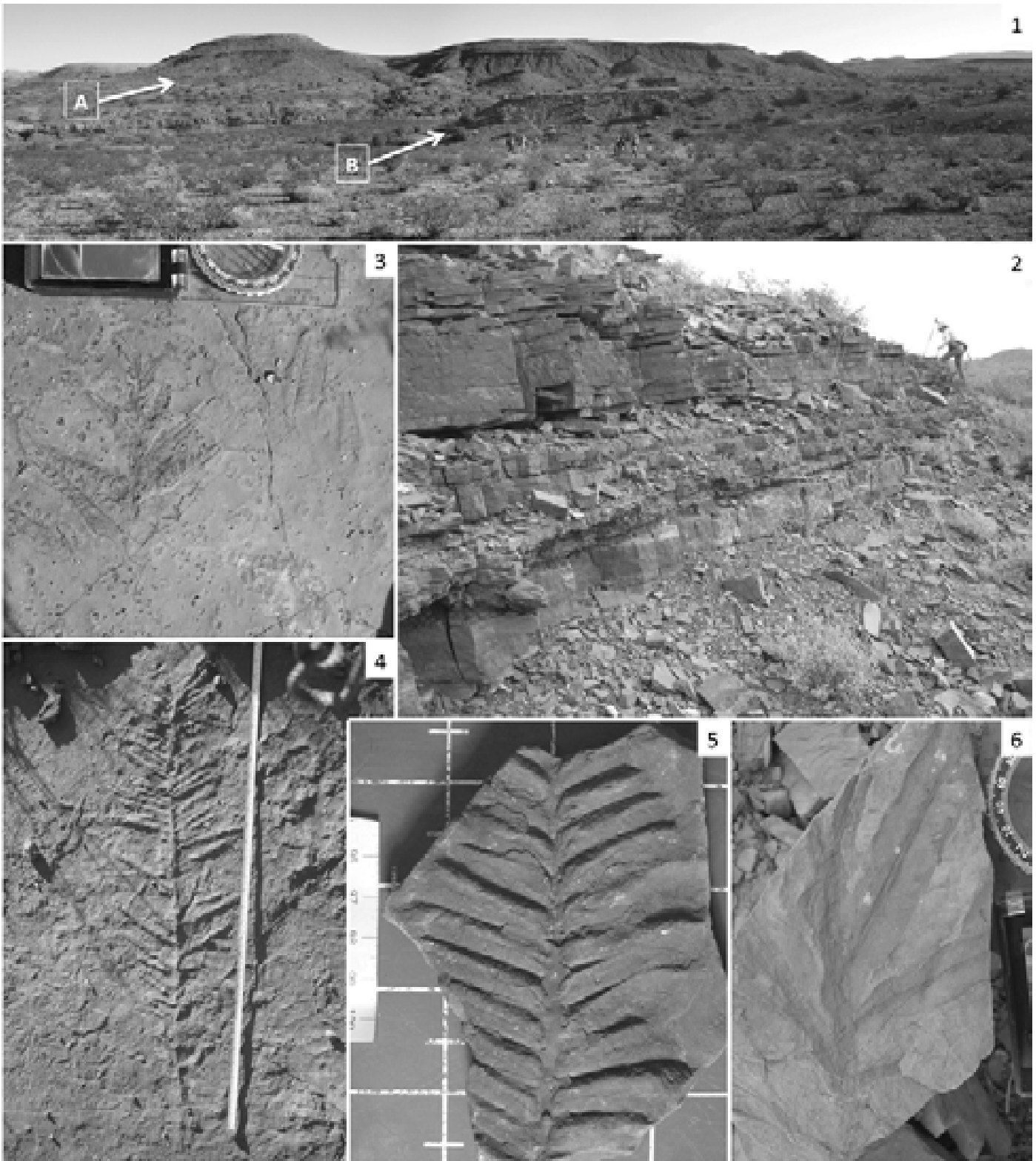


FIGURE 22. Abo Formation outcrops and flora in Cañoncito de la Uva. **1**, Bursum (B at arrow) and Abo (A at arrow) formation exposures in Cañoncito de la Uva looking south from the road. **2**, Abo Formation sheet-like siltstone-sandstone bed showing platy bedding in upper, fossiliferous portion. **3**, *Autunia conferta*, USNM specimen 543955. **4**, *Walchia* sp., entire branch, field photograph. **5**, *Supaia thinnfeldioides*, NMMNH specimen P-42811. **6**, *Supaia anomala*, field photograph.

composed of cyclically alternating layers of siltstone, bedded gypsum, and limestone or dolomite. Much of the siltstone is massive and may represent ancient loess, or wind-blown dust. Gypsum can be evenly laminated, wavy laminated, or nodular. The limestone and dolomite layers, which erode to ledges, are mostly carbonate mudstone and peloidal, oncoidal, and oolitic wackestone yielding a sparse fauna of ostracods, bivalves, and rare foraminifera. Overlying the Torres is the Cañas Gypsum, which is 66 m thick and composed of massive to bedded gypsum having interbeds of siltstone and dolomite. This material erodes to a “moonscape” that is practically devoid of vegetation and highly subject to karst development. The Joyita Member, at the top of the Yeso Group, is 12.3 m of reddish to greenish gray siltstone and sandstone that lacks gypsum and forms a reddish slope beneath the light gray cliffs of Glorieta Sandstone.

The more agile participants in the field trip will scramble up to the lower layers of the Glorieta Sandstone, the light-colored, crossbedded sandstone that forms prominent cliffs (Fig. 25). The Glorieta is largely an eolian deposit that is correlative with the Coconino Sandstone of the Grand Canyon, but toward the southeast, Glorieta sand underwent considerable reworking in shallow marine settings (Milner, 1978). Forming the top of the high knob is the youngest Permian formation in this area of New Mexico, the San Andres Formation. The San Andres is late Leonardian to early Guadalupian age and is composed largely of marine limestone and dolomite, with interbeds of gypsum, anhydrite, and sandstone in some localities. Access to the San Andres here is difficult and will not be attempted by the field party.

The outcrop area of the Arroyo de Alamillo Formation in this area, near Tomas Baca Well, also includes the stratigraphically highest record of terrestrial fossils of the Yeso Group in central New Mexico. Plant fossils, root traces, and invertebrate and vertebrate traces have been recently found at two sites (NMMNH localities 8715 and 8716) north of Tomas Baca Well in pale reddish to greenish-gray beds ~45–50 m above the base of the Arroyo de Alamillo Formation (Lucas et al., 2013). Plants are known from an *in situ* callipterid pteridosperm and various conifer remains. Abundant root traces indicate that plants must have been locally common during deposition of the Arroyo de Alamillo Formation. Invertebrate traces are remarkably diverse in siliciclastic strata of NMMNH locality 8715 and tentatively attributed to *Diplichnites*, *Diplopodichnus* (Fig. 10J), *Gordia*, *Helminthoidichnites*, *Scoyenia*, *Stiallia* (Fig. 10K, L), *Striatichnium*, and *Treptichnus*. According to this record aquatic to semiaquatic arthropods must have been an important element of the invertebrate fauna. The tetrapod ichnofauna consists of *Batrachichnus*, *Dromopus* (Fig. 10L) and undetermined “captorhinomorph” tracks. Taking the lenticular geometry of the fossil-bearing beds, the predominance of mud and the abundance of rain drop imprints and mudcracks into account, the strata most likely represent deposits of temporary pools.

Plant fossils are very rare in the Yeso Group, and where they occur, the depositional conditions appear very similar to those under which Abo Formation plants were deposited. Also, similarly to the Abo, Yeso floras are dominantly conifers and *Supaia*. At present, too few exposures are known to permit a definitive characterization of the Yeso flora and, thereby, to understand if there were any significant floristic compositional changes that accompanied the environmental transition from Abo to Yeso deposition.

After stop, continue east on all-weather road. 0.2

- 9.8** Prominent knob of Glorieta Sandstone at 10:30 on Yeso beds. Sierra de la Cruz at 2:00. Road is in lower Yeso strata. **0.3**
- 10.1** Cross tributary of Cañoncito de la Uva. Note lower Yeso strata exposed in arroyo banks. **0.2**
- 10.3** Cattleguard, Tomas Baca Well; road in Abo Formation. John Nelson’s retirement home (not paid for yet!) on left. Note excellent view of Mesa del Yeso at 7:00. Arroyo crossing here is in lower Yeso strata. **0.2**
- 10.5** Cross arroyo; outcrops to left are very close to the Abo-Yeso contact. These are uppermost Abo red beds with rhizoliths in the siltstones and trough-crossbedded fluvial sandstone bodies. **0.5**
- 11.0** Cross arroyo (another tributary of Cañoncito de la Uva) floored by Abo red beds. Notice change in dip in the Abo Formation rocks here, east of the road dip east, west dip west. Apparently we are on a small anticlinal flexure. **0.2**
- 11.2** Crossed fault, rocks to right of road are interbedded siliclastic limestones and red beds of the Bursum Formation. **0.2**
- 11.4** Cross arroyo; note top of Bursum limestone to right; road is at Bursum-Abo contact here: dip slope on top of Bursum to right and Abo red beds to left. **0.4**
- 11.8** Cattleguard and brass cap for NE1/4 NE1/4 sec. 10 T02S R02E. The road here is on the upper part of the Bursum Formation. **0.1**
- 11.9** Small prospect on right. Many small prospects are found in the Bursum Formation in this area. Localized redbed copper deposits are common, together with small concentrations of uranium mineralization, though nothing large enough to be of economic significance. Road ascends hill along dip of limestone unit in upper part of the Bursum Formation. **0.2**
- 12.1** Crest hill, road now descends dip slope of upper Bursum strata. **0.2**
- 12.3** Cross arroyo, a tributary of the Cañoncito de la Uva. Outcrops near the road have been mapped as Bursum Formation. The road parallels a swarm of normal faults striking north-northwest (Colpitts, 1986). **0.1**
- 12.4** Road junction, keep left. Note larger mining prospect straight ahead. Road to right goes to Del Cuerto Ranch headquarters, 7 mi. Road to left (our route) to Creel Headquarters (18 mi.). **0.1**
- 12.5** Cross Cañoncito de la Uva. Gallina Well (windmill) to left. Note Bursum outcrop to left of and downstream of Gallina Well. **0.1**
- 12.6** Road(s) to left to Gallina Well, **continue straight on main road. 0.2**
- 12.8** Road enters arroyo floor and is driving in upper Bursum strata; note thick Abo section up hills to left. **0.1**
- 12.9** Road follows strike valley with some local faulting (note numerous slickensides in Abo rocks to right) along Bursum-lower Abo stratigraphic interval. **0.2**
- 13.1** Road climbs out of arroyo. Note low, conical hill of Abo Formation at 12:00; this is the Gallina Well vertebrate fossil locality. **STOP here.**

The Gallina Well vertebrate fossil locality. The Gallina Well locality is a Lower Permian vertebrate body and ichnofossil site located approximately 20 km northeast of Socorro, New Mexico in the Joyita uplift. The locality is situated

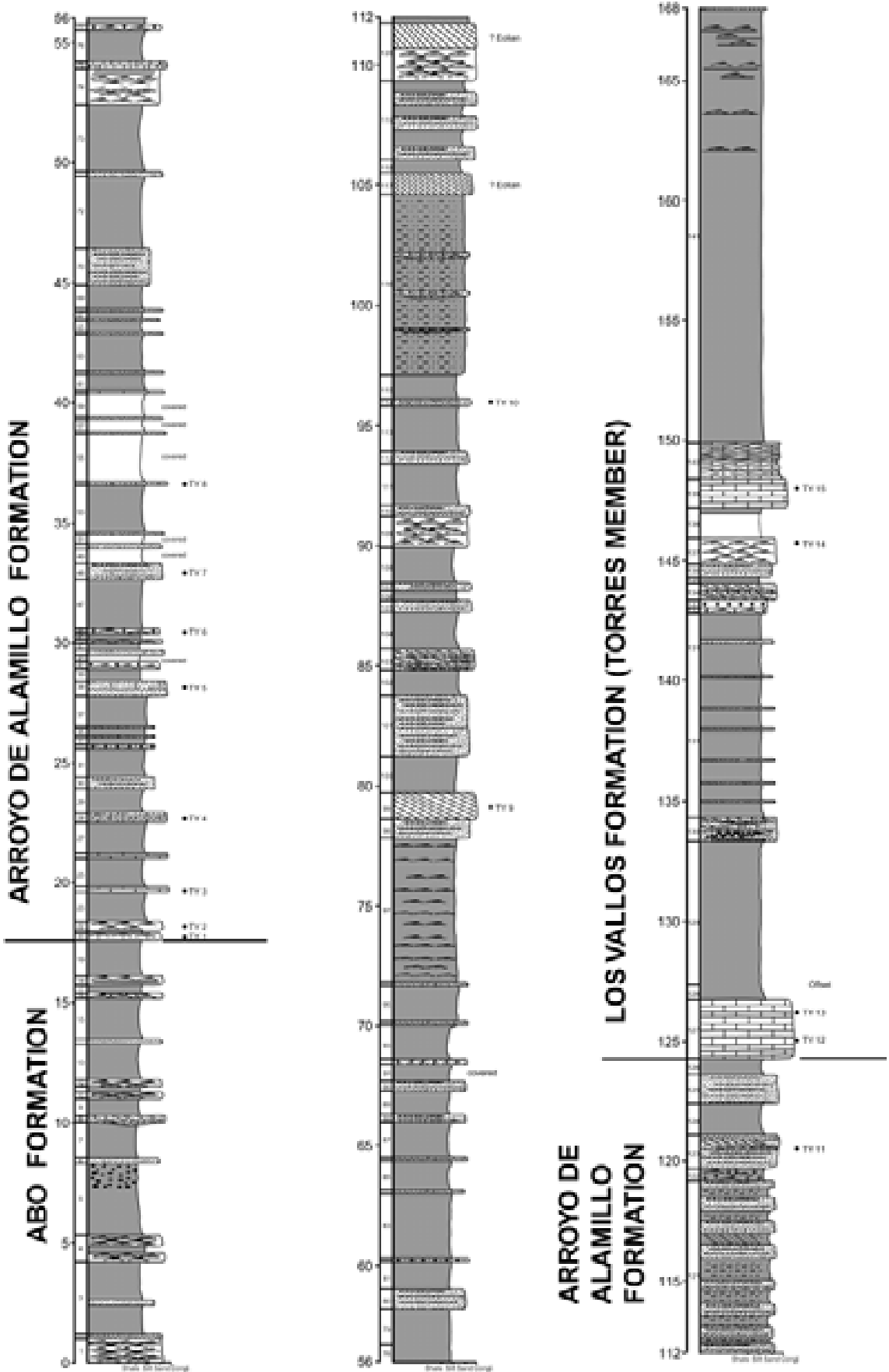


FIGURE 23. Type section of the Yeso Group (from Lucas et al., 2005).

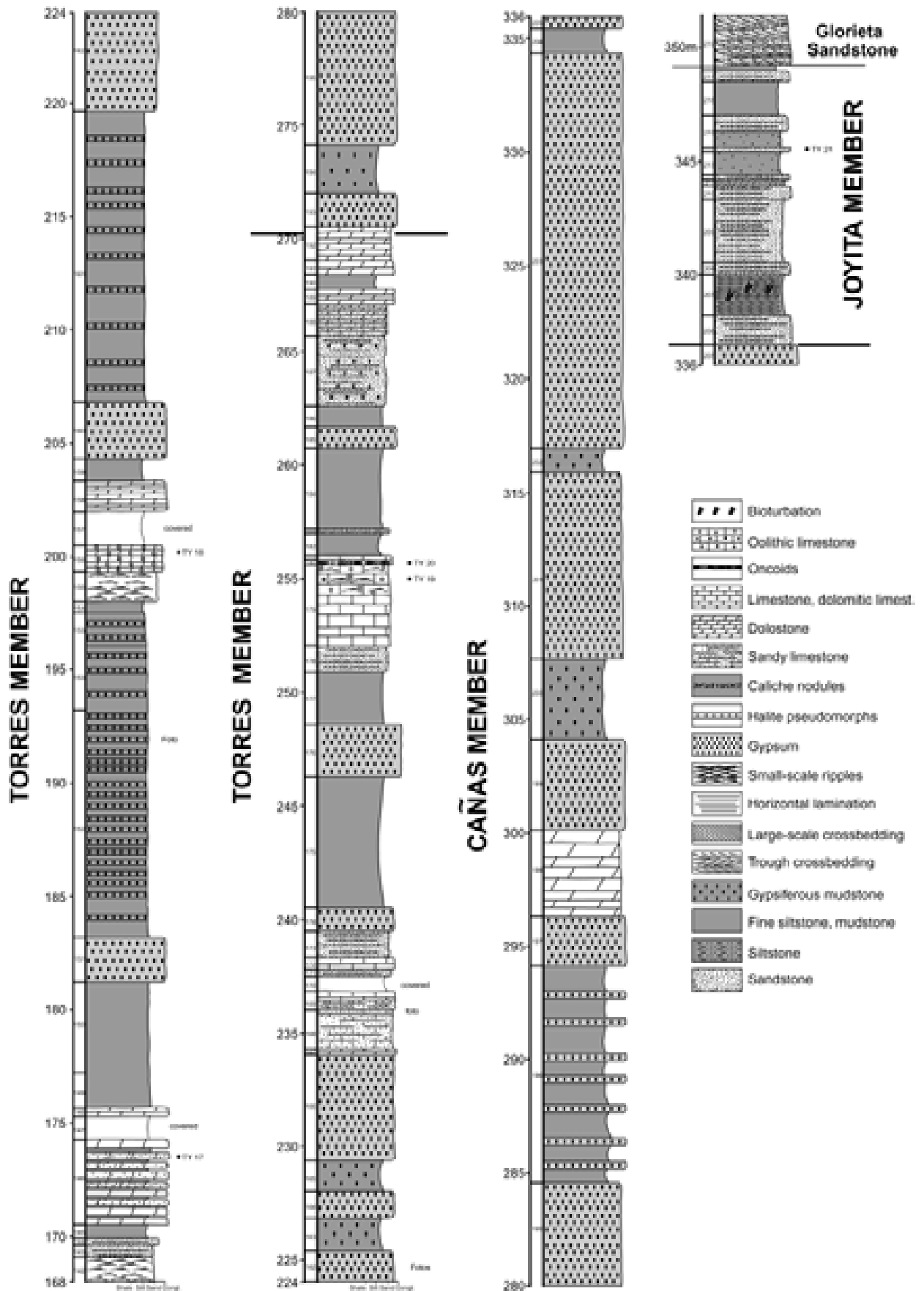


FIGURE 23. Continued. Type section of the Yeso Group (from Lucas et al., 2005).

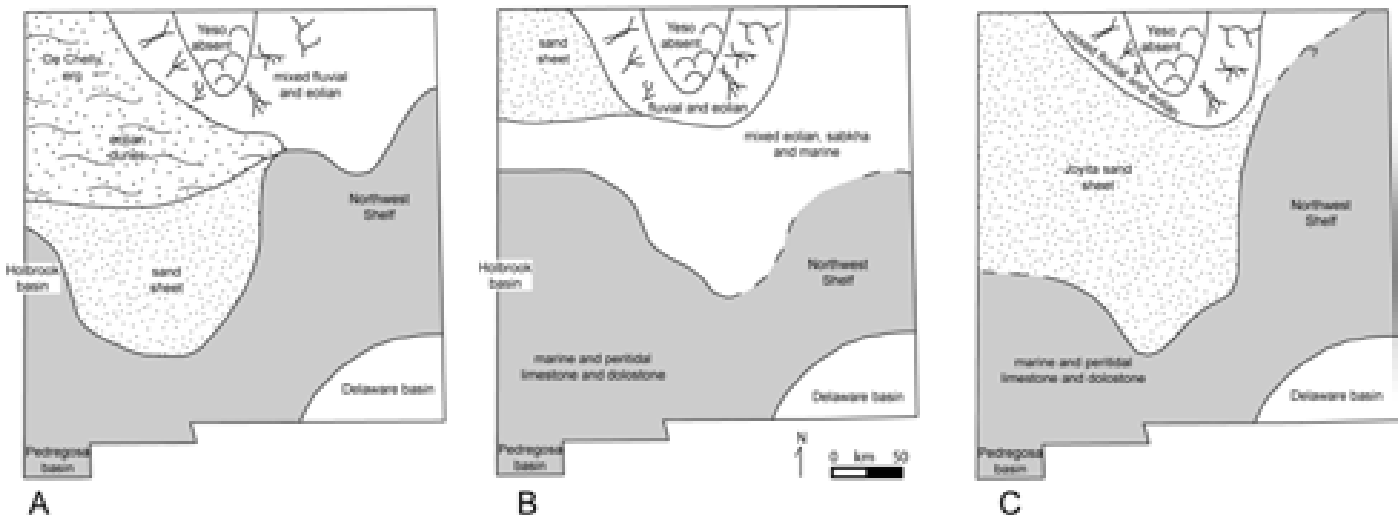


FIGURE 24. Yezo paleogeography (after Mack and Dinterman, 2002; Lucas et al., 2013). A, Early, B, middle and C, late Yezo deposition.

stratigraphically low in the Scholle Member of the Abo Formation and is Coyotean in age. The fossiliferous beds are reddish-brown, fluviially-deposited, calcrete-pebble conglomerates and mudrock. Vertebrate body fossils from the site include paleoniscoid fish; the temnospondyl amphibians *Eryops* sp., *Trimerorhachis* sp., *Platyhystrix* sp., and *Zatrachys* sp.; a skull fragment of the lepospondyl *Diplocaulus* sp.; posterania of the diadectid *Diadectes* sp.; a captorhinid skull and postcranial skeleton; and specimens of the eupelycosaur *Ophiacodon* sp., *Sphenacodon* sp. and *Dimetrodon* sp. The coprolite ichno-assemblage includes *Dakryonocopros arroyoensis*, *Alococopros triassicus*, *Heteropolacopros texaniensis* and amorphous coprolites. The Gallina Well locality yields the most diverse and extensive vertebrate body fossil and coprolite assemblage of Early Permian age known from southern New Mexico. Its basic composition differs little from the pelycosaur-dominated assemblages found to the north, indicating some uniformity of the Coyotean vertebrate fauna across New Mexico.

Not far to the northwest, we can examine an unusually thick section of the Bursum Formation.

Bursum Formation. The Bursum Formation here is 120 m thick and assigned to the Red Tanks Member (Fig. 26). The succession rests on nodular to wavy bedded fossiliferous limestone of the Atrasado Formation and is overlain by nonmarine red beds of the Abo Formation. The lower 45 m are composed of predominantly greenish gray and reddish mudstone-siltstone, some covered (?shale) intervals and intercalated conglomerate and sandstone beds. Three conglomerate beds are intercalated in the lower part between 6 and 14 m above the base (0.2-1 m thick). Conglomerate beds are poorly sorted, clast supported and composed of carbonate clasts with diameters up to 20 cm in the lowermost bed. Intercalated sandstone beds are up to 0.8 m thick. Thin sandstone beds (0.1-0.3 m) are massive, horizontally laminated or display ripples. Thicker sandstone beds display trough crossbedding and lense out laterally (channel fills). In the lower part one thin (0.3 m), dark gray micritic limestone bed is intercalated in shale about 35 m above the base.

The middle part of the Bursum Formation (45-65 m) is composed of greenish-gray and gray shale, some covered (?shale) intervals and many thin (mostly 5-20 cm, rarely up to 50 cm) micritic limestone beds (mudstone to wackestone). Some beds contain fossils such as brachiopods, gastropods,

crinoidal debris and bryozoans (packstone). Two thin crossbedded sandstone beds are intercalated (0.2 and 0.4 m thick). At the top of this interval nodular limestone (0.6 m) is intercalated which contains brachiopods and bryozoans. A fault is observed about 50 m above the base, but does not appear to remove much of the section. Between approximately 65 and 80 m above the base several coarse-grained sandstone and conglomerate beds and one nodular calcrete horizon are intercalated in red mudstone which locally contains abundant small pedogenic limestone nodules. The conglomerate beds are up to 2.1 m thick, poorly sorted and composed of carbonate clasts and sandy arkosic matrix. The conglomerate beds are partly crossbedded, the clasts are subangular to subrounded; grain-size is mostly < 5 cm, rarely up to 20 cm. The upper part of the section (80-120 m) is composed of red and subordinately greenish-gray mudstone with intercalated carbonate conglomerate beds (0.3-0.6 m thick), crossbedded arkosic sandstone beds (0.4-1 m thick) and few fossiliferous, partly sandy limestone beds (3-30 cm), a nodular limestone (1.2 m) in the lower part and nodular calcrete horizons (0.6-2.8 m). Red mudstone locally contains small calcrete nodules. The uppermost conglomerate lacks arkosic material and contains bone fragments. A thin limestone bed in the lower part, approximately 90 m above the base contains crinoidal debris, bryozoans and brachiopods. The limestone bed approximately 104 m above the base is characterized by the occurrence of abundant echinoid spines, some of them up to 5 cm long. The thin limestone beds which are intercalated in pink shale near the top contain *Dunbarella* and represent the last marine horizon.

The lower part of the section is dominantly nonmarine, the middle part shallow marine and the upper part again dominantly nonmarine with few thin marine horizons represented by fossiliferous thin limestone beds.

The terrestrial deposits of the Bursum are primarily paleosols and plant fossils usually have been obliterated by pedogenesis. However, in the Canoñcito de la Uva area there are a number of plant-bearing deposits have been found in gray shales and occasionally sandy siltstones. These deposits occur in close association with syndepositional faulting and appear to represent lakes that were preserved by the creation of accommodation space allowing shorter-term burial. The floras of the Bursum are, for the most part, dominated by conifer branches and foliage cordaitalean foliage, and callipterids with other plants represented locally. This is the first unit in

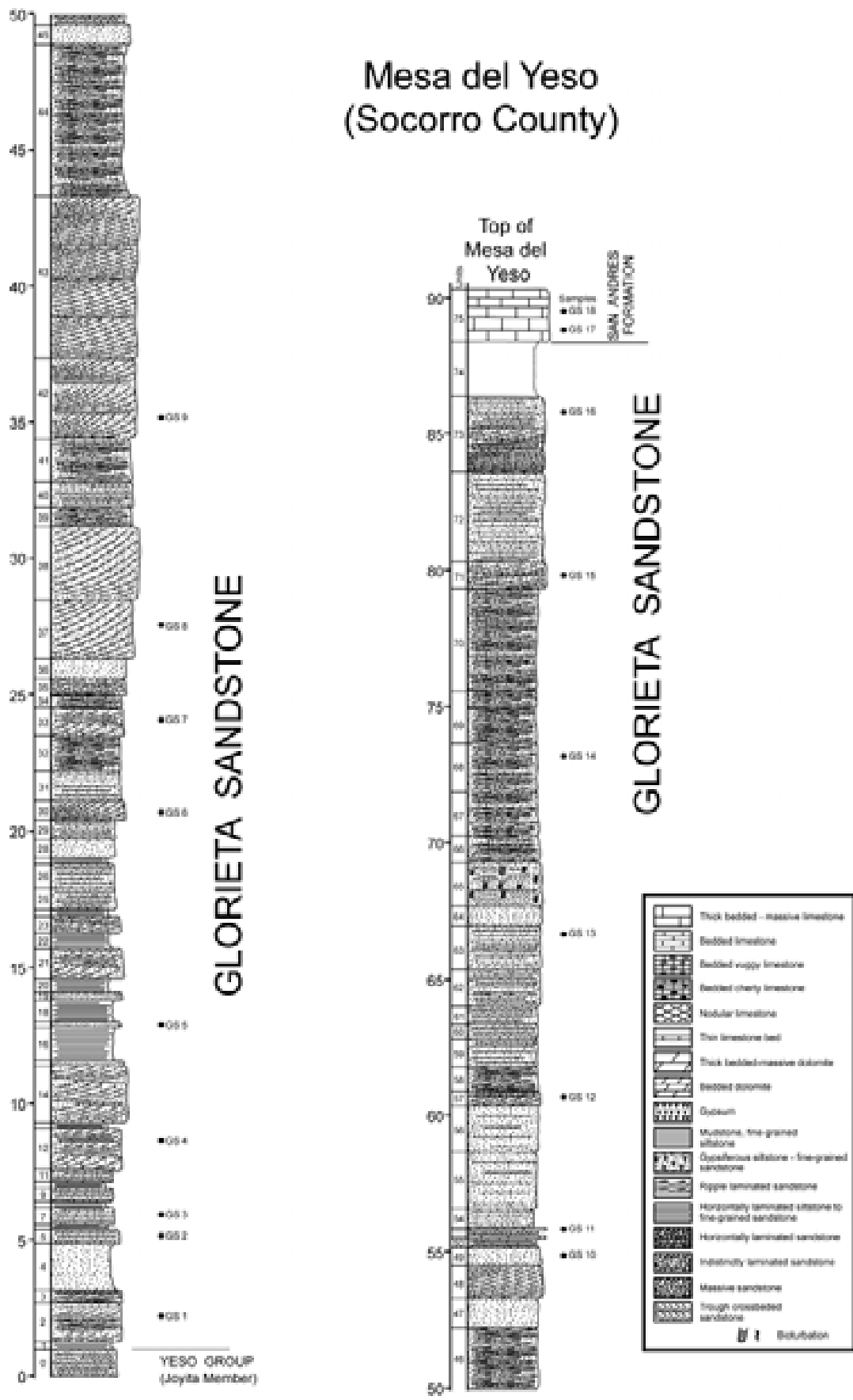


FIGURE 25. Measured stratigraphic section of the Glorieta Sandstone at Mesa del Yeso.

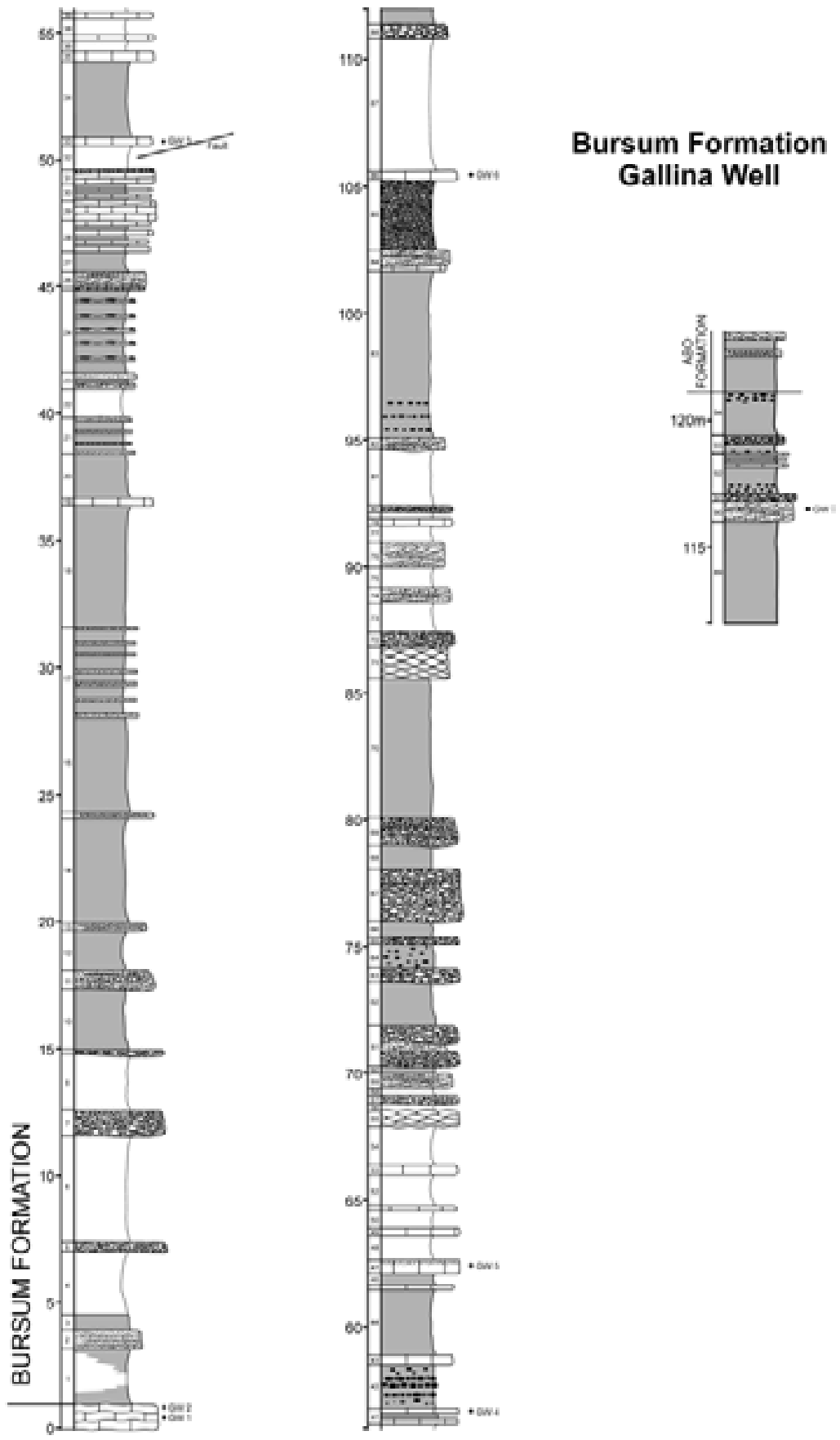


FIGURE 26. Bursum Formation section near Gallina Well.

the regional succession in which callipterid peltasperms are quantitatively important parts of the flora. The overall plant assemblage is indicative of seasonally dry climatic conditions.

The wetland elements, typically found in the terrestrial facies of the Atrasado Formation, are mostly lacking or very rare in the Bursum Formation.

End of third-day road log.

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THE PENNSYLVANIAN SANDIA FORMATION IN NORTHERN AND CENTRAL NEW MEXICO

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Abstract—Across much of northern and central New Mexico, the Sandia Formation forms the lower part of the Pennsylvanian succession and represents early synorogenic deposits associated with the initiation of the Ancestral Rocky Mountains (ARM) orogeny. Named in 1900, the Sandia Formation rapidly became part of the New Mexico lithostratigraphic lexicon. It has been identified or mapped as a mixed siliciclastic and carbonate succession, particularly characterized by thick beds of quartz-rich sandstone, conglomeratic sandstone and conglomerate. The high amount of siliciclastic sediment in the formation was derived from tectonically active local ARM uplifts composed of Proterozoic basement rocks. Thickness of the Sandia Formation in central New Mexico is 7-162 m, with the thinner sections proximal to ARM highlands. The Sandia Formation at most places rests on Proterozoic basement with an angular unconformity but locally overlies older, Mississippian or Lower Pennsylvanian sedimentary rocks. At most places the dominantly or entirely siliciclastic sediments of the Sandia Formation are conformably overlain by thicker limestone/cherty limestone units of the Gray Mesa Formation. To the north, sandstone of the Sandia Formation is commonly subarkose, whereas to the south most sandstones are quartzarenites. Limestone microfacies of the Sandia Formation are various types of mudstone, wackestone to floatstone and grainstone to rudstone containing a diverse fossil assemblage. The depositional environment of siliciclastic sediments ranges from fluvial to fluvio-deltaic, coastal swamp and brackish, to coarse-grained high-energy nearshore (storm dominated shelf) and fine-grained middle-outer shelf deposits. Limestone containing a low-diversity fossil assemblage indicates deposition in a restricted, shallow marine shelf environment. Limestone types with a diverse fossil assemblage accumulated in open, normal marine shallow shelf settings of low to high energy.

The Sandia Formation yields abundant fossils ranging from those of nonmarine (plants and amphibians) to fully marine (especially algae, foraminifers including fusulinids, brachiopods, crinoids, bryozoans and sharks teeth) organisms. Fusulinids and limited conodont records indicate the Sandia Formation is mostly of Atokan age, but locally the lower part of the formation is late Morrowan. The Sandia Formation was deposited as part of the extensive Absarokan marine transgression and is characterized by distinct lateral changes in thickness and facies as a result of the ARM deformation. Cycles are present within the dominantly marine successions of the Sandia Formation, but cannot be traced laterally over long distances, so we assume that the formation of the cycles was caused mainly by tectonic movements of the ARM, although some glacio-eustatic influence cannot be ruled out.

INTRODUCTION

Across much of northern and central New Mexico, the Sandia Formation forms the lower part of the Pennsylvanian succession and represents early synorogenic deposits associated with the initiation of the Ancestral Rocky Mountains (ARM) orogeny (e.g., Armstrong et al., 1979). Named by Herrick (1900a), the Sandia Formation was also one of the first lithostratigraphic names introduced in New Mexico. Nevertheless, no type section of the unit was specified by Herrick (1900a), and no subsequent worker proposed a lectostratotype section until Krainer et al. (2011) defined and characterized the Sandia Formation. Here, we present an overview of the Sandia Formation in northern and central New Mexico (principally in Sandoval, Bernalillo, Valencia, Torrance, Socorro and Sierra Counties; Fig. 1).

PREVIOUS STUDIES

Herrick (1900a) and Herrick and Bendrat (1900) described the “Sandia series” as a succession of shale, sandstone and conglomerate with intercalations of sandy limestone, resting on Precambrian granitic basement and present in the Sandia, Manzano and San Andres Mountains. The name refers to the Sandia Mountains, which form the skyline directly east of Albuquerque. Herrick (1900a, p. 114) described strata above the “Sandia series” as dark conchoidal limestone and shales. Above that limestone is a sandstone/conglomerate unit that Herrick (1900a, p.

115) named the Coyote Sandstone. In subsequent articles, Herrick (1900b; Herrick and Bendrat, 1900) made it clear that he considered the “Sandia series” to be about 150 ft (46 m) thick and consist of shale, sandstone, conglomerate and a few beds of sandy limestone.

According to Keyes (1904), the Sandia Formation is about 300 ft (90 m) thick and consists of limestone, which underlies the Madera limestone and overlies the Lake Valley limestone. Gordon (1907) described the Sandia Formation as a 500 to 700 ft (152 to 213 m) thick succession of alternating beds of blue and black shale, limestone, sandstone and conglomerate, which underlie the Madera limestone.

The Sandia Formation rapidly became part of the New Mexico lithostratigraphic lexicon and has been identified or mapped across much of northern and central New Mexico (see Wilmarth, 1938; Keroher, 1966; Kues et al., 1982; also see Kottlowski, 1963, Armstrong et al., 1979, and Baltz and Myers (1999) for a review of the regional lithostratigraphy of the Sandia Formation in northern and central New Mexico). Read et al. (1944; also see Wilpolt et al., 1946) divided the Sandia Formation into two members, a lower limestone member and an upper clastic member. However, Armstrong (1955) recognized that the lower limestone member is of Mississippian age and renamed it the Arroyo Peñasco Formation (later raised to group status). In the Jemez Mountains of northern New Mexico, the lower part of the Sandia Formation was separated out as the Osha Canyon Formation (DuChene, 1974; DuChene et al., 1977; Krainer and Lucas, 2005).

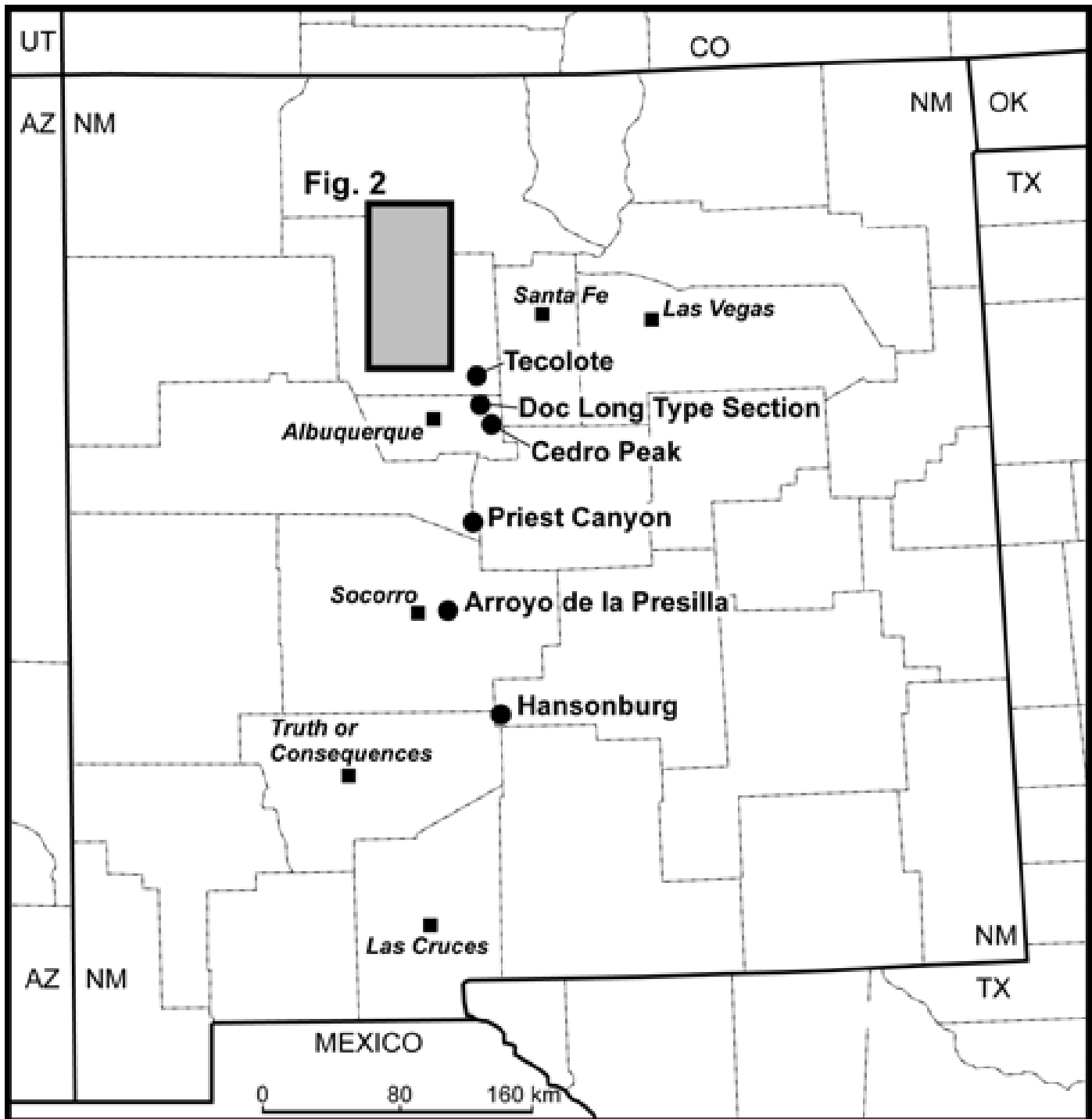


FIGURE 1. Map showing locations of the studied sections of the Sandia Formation and location of more detailed map in Figure 2.

Thompson (1942), in his review of Pennsylvanian strata in New Mexico, argued that the name Sandia Formation should be restricted to the lower 127 ft (39 m) of coarse-grained sandstone, conglomerate, and variegated shale at the base of the Pennsylvanian section in the Sandia Mountains. South of the Sandia Mountains, Thompson (1942) stated that the Sandia sandstone and arenaceous beds change laterally into the lower limestone interval of his Elephant Butte Formation.

In the Sandia Mountains and vicinity, a series of Master's theses (Toomey, 1953; Szabo, 1953; Reynolds, 1954; Catacosinos, 1962; Phillips, 1964) undertaken at the University of New Mexico provided detailed data on the Sandia Formation around its type area (see Kelley

and Northrop, 1975, for a published overview of the Sandia Formation in the Sandia Mountains and vicinity, based primarily on these theses).

Kues (2001) summarized the development of the concept of the Sandia Formation in the Sandia Mountains and neighbouring areas. To the south, in Socorro County, Darton (1928) gave a brief description of the Arroyo de la Presilla section, which he mentions as a completely exposed Pennsylvanian section (see also Thompson, 1942; Wilpolt et al., 1946; Read and Wood, 1947; Baltz and Bachman, 1956).

The type locality of the Sandia Formation is on the southern end of the Sandia Mountains (Fig. 1). Although for about one century no type section had been designated for the Sandia Formation, the unit had

long been recognized and mapped in central and northern New Mexico as the basal, clastic-dominated interval of the Pennsylvanian section. Krainer and Lucas (2005) designated a lectostratotype section (principal reference section) of the Sandia Formation at Doc Long Campground in the Sandia Mountains. This well-exposed, fossiliferous and readily accessible section well represents Herrick's and subsequent workers' concept of the Sandia Formation and was described in detail by Krainer et al. (2011).

LITHOSTRATIGRAPHY

Introduction

Here we review the lithology, petrography, microfacies, lateral variation, paleontology and age of the Sandia Formation based primarily on outcrops that we have studied in the Jemez and Nacimiento Mountains of Sandoval County, the Sandia and Manzanita Mountains of Bernalillo County, the southern Manzano Mountains in Torrance County, the Oscura Mountains of Socorro County and the Cerros de Amado of Socorro County (Figs. 1-2).

Lithology

The Sandia is a formation composed of interbedded sandstone, shale, and marine limestone. Overall, the clastic rocks decrease in grain size upward, but abrupt lateral changes in thickness and lithologic succession are the rule.

Sandia Mountains

Toomey (1953) studied the Sandia Formation in the northern Sandia Mountains, which is just north of its type area. He described the Sandia Formation as a primarily clastic unit, consisting of white to gray-buff and green quartzite, conglomeratic sandstone, crossbedded sandstone and fossiliferous calcareous sandstone, gray to black fossiliferous shale, red conglomeratic shale and highly fossiliferous arenaceous limestone. Szabo (1953) studied the Sandia Formation at Tijeras Canyon, Manzanita Mountains (just south of its type area), where it is a 115 ft (35 m) thick unit composed of alternating shale, sandstone, argillaceous sandstone and limestone. Phillips (1964) noted that the Sandia Formation consists of sandstone, shale, conglomerate and minor limestone, and that the sections are difficult to correlate due to rapid lateral facies change.

At the Sandia type section, the Sandia Formation rests with an angular unconformity and erosional paleorelief on Precambrian granitic rocks (Fig. 10A). The lower part consists of alternating shale, sandstone, pebbly sandstone, sandy limestone and limestone (Fig. 10B). Shale is dark gray to black and locally contains carbonate concretions and plant fossils. Individual shale units are up to 3.3 m thick. Sandstone and pebbly sandstones occur as thin beds and horizons up to 4.2 m thick. Thin sandstone beds are massive or display ripple lamination, and thicker sandstone units commonly display trough cross-bedding and, rarely, swaley topography. The bases of sandstone units are erosive, and the units display an upward-fining trend. Some sandstone horizons also contain carbonate material and a few fossil fragments. Pebbly sandstones contain pebbles up to 4 cm in diameter and shale intraclasts that are up to 10 cm long.

Sandy limestone horizons are 0.2-2 m thick and mostly appear massive without internal bedding. Rarely, thicker limestone intervals are bedded with a bed thickness of 20-40 cm. Sandy limestones are either mixed siliciclastic-carbonate sandstone or rudstone containing quartz grains. Rarely, crinoidal packstone is present.

With a sharp contact, pebbly sandstone of the upper part of the Sandia Formation overlies cherty limestone of the middle part of the formation. In the upper part, crossbedded, coarse-grained and pebbly sandstone is the most common lithology. Sandstone units are up to 10.4 m thick. Shale is not exposed, but probably represented by the covered intervals, which are 0.5-2.7 m thick. Individual sandstone units show a distinct upward fining trend. All sandstones display large-scale trough

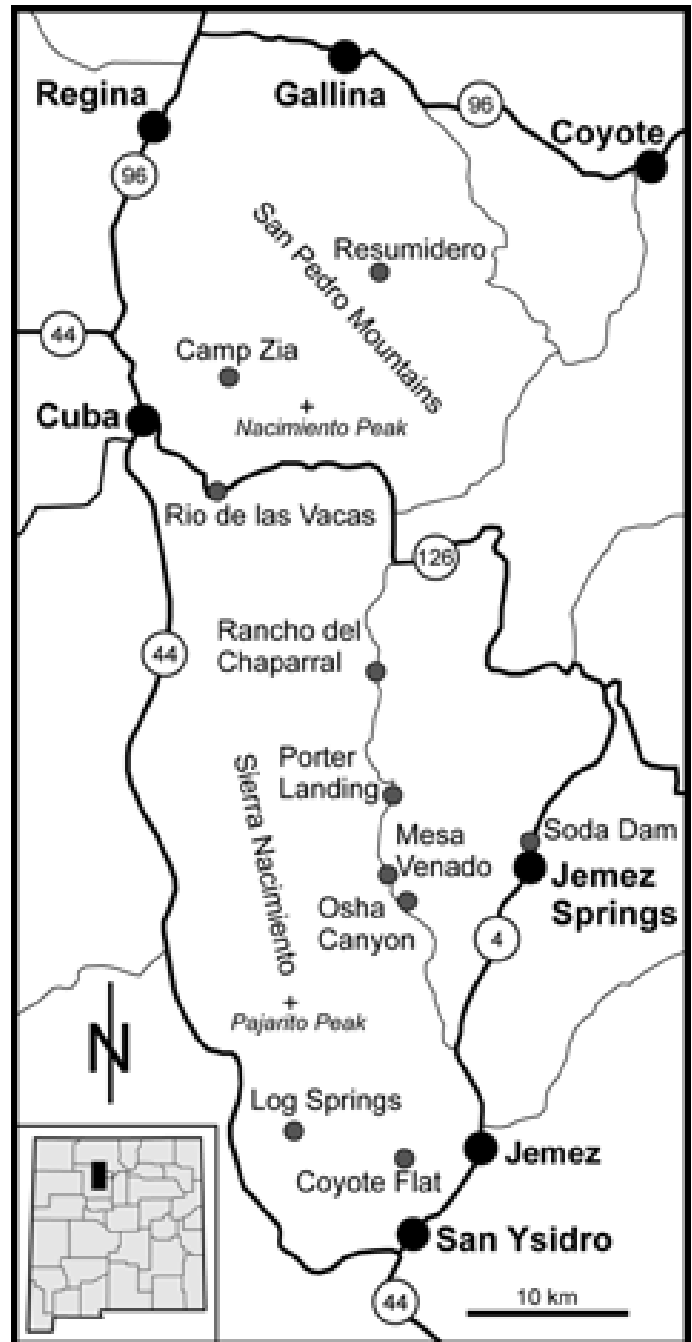


FIGURE 2. Map showing locations of the studied sections of the Sandia Formation in the San Pedro Mountains-Sierra Nacimiento-Jemez area.

cross-bedding. Intercalated are three thin (0.3-0.5 m), fossiliferous, partly sandy limestone horizons. The uppermost sandstone is sharply overlain by gray, bedded cherty limestone of the Gray Mesa Formation (Krainer et al. 2011).

At Tecolote near the northern end of the Sandia Mountains, the Sandia Formation thins to 24 m, rests on Mississippian limestone and is overlain by the Gray Mesa Formation (Figs. 1, 8). The Sandia Formation is entirely siliciclastic and composed of fining-upward cycles. The lowermost cycle is 8 m thick and starts with a thin conglomerate that fills up a paleorelief and contains angular clasts up to 5 cm in diameter. The basal conglomerate is overlain by trough crossbedded, coarse-grained, pebbly sandstone, grading into poorly exposed, finer grained sandstone and finally into thin sandstone beds that contain marine fossils, overlain by shale containing plant fossils and fine-grained sandstone partly display-

ing ripple lamination. The next cycle is thin (1 m), has an erosive base, starts with crossbedded pebbly sandstone and grades upward into fine-grained sandstone and shale. The shale is overlain by 6.2 m of crossbedded sandstone, partly pebbly, and with thin conglomerate beds containing clasts with diameters up to 2 cm. Above, follows dark gray to black siltstone-shale. The shale is thinly laminated, and siltstone beds are up to a few cm thick. The upper part of this shale and siltstone contains plant fossils, brachiopods and crinoid fragments. The top of the Sandia Formation is composed of 0.9 m thick, coarse-grained, partly pebbly sandstone displaying trough crossbedding overlain by wavy bedded, fossiliferous gray limestone of the Gray Mesa Formation containing abundant brachiopods.

Sierra Nacimiento

The Sandia Formation in the Sierra Nacimiento, north-central New Mexico (Figs. 1 and 2), is characterized by distinct lateral changes in thickness and facies as a result of the ARM deformation. In the Sierra Nacimiento, the Sandia Formation is thinner than at the type section (124 m). The thickest sections (72 m) are exposed on the eastern side of the Sierra Nacimiento north of Jemez Springs (Soda Dam; Figs. 1, 2, 6) where the Sandia Formation rests on Precambrian gneiss and is composed of shale with intercalated sandstone and fossiliferous limestone containing the fusulinids *Millerella*, *Eostaffella* and *Fusulinella*.

North of Guadalupe Box (at Osha Canyon; Figs. 1, 2, 6) the Sandia Formation overlies the Lower Pennsylvanian (Morrowan) Osha Canyon Formation, is 32 m thick and composed of several fluvial fining-upward cycles and a thin marine horizon at the top. The lowermost cycle is 1.8 m thick and starts with 0.3 m of fine-grained, poorly sorted conglomerate containing clasts up to 3 cm in diameter, overlain by trough crossbedded quartzose sandstone and gray micaceous siltstone–mudstone. The next cycle is 4.8 m thick, has an erosive base and starts with trough crossbedded, coarse-grained, pebbly sandstone, grading into medium-grained sandstone displaying trough crossbedding, horizontal lamination and (locally) soft sediment deformation structures. The overlying cycle is 6.9 m thick and composed of coarse-grained, pebbly, trough crossbedded sandstone, fining upward and grading into crossbedded sandstone with soft sediment deformation structures, overlain by greenish siltstone to fine-grained sandstone locally displaying current ripples. The top of the cycle is composed of gray micaceous shale containing *Lingula* and plant debris, overlain by greenish-gray siltstone (1.1 m).

The uppermost cycle is 16 m thick, starts with coarse-grained sandstone displaying large-scale trough crossbedding erosively cut into the underlying siltstone and overlain by greenish siltstone and shale with several intercalated fine-grained sandstone intervals up to 1.2 m thick. The sandstone is massive or displays horizontal lamination, and rare ripple lamination. Bioturbation is also observed. Root structures are present in the lowermost sandstone horizon. The uppermost 0.9 m of the Sandia Formation are composed of gray, fossiliferous arenaceous limestone (packstone: Plate 1-8) containing crinoidal debris, overlain by nodular cherty limestone of the Gray Mesa Formation.

At Mesa Venado (section E, Figs. 1, 2, 5), the Sandia Formation rests on Precambrian granitic basement and is approximately 27 m thick. The granite is overlain by 1 m of strongly weathered granitic grus, followed by approximately 6 m of brownish, yellowish and dark gray shale (0.1-0.4 m) and intercalated, coarse-grained, partly pebbly quartzose massive sandstone (0.2-0.7 m). Above a covered interval, 2.3 m of trough-crossbedded, very coarse-grained, pebbly sandstone is exposed. Above a fault at which probably a few meters of the section are missing, brownish and black shale with small carbonate nodules is exposed, overlain by 1.4 m of arenitic wavy limestone beds containing brachiopods and separated by thin shale intercalations, followed by poorly exposed shale, a thin (0.1 m) limestone bed, greenish-gray shale with abundant limestone nodules and fossils (brachiopods, crinoids), overlain by 6.8 m of poorly exposed brownish-gray and yellowish shale. The overlying basal Gray

Mesa Formation is a 0.3 m thick arenitic limestone bed and 0.5 m of nodular limestone containing brachiopods.

At Porter Landing (Figs. 1, 2, 5), a few km farther north, the exposed thickness of the Sandia Formation is 17 m. The Sandia Formation is entirely siliciclastic. The contact with the basement is not exposed, although the covered part above the granitic basement is thin. The Sandia Formation is composed of fine- to medium-grained sandstone. Individual sandstone beds are up to 30 cm thick, commonly display small-scale trough crossbedding (Fig. 10F), subordinately horizontal lamination and ripple lamination. Intercalated is brownish, micaceous siltstone to silty mudstone. Sandstone locally contains abundant fine-grained plant debris. The Sandia Formation is overlain by limestone of the Gray Mesa Formation.

At Rancho del Chaparral (Figs. 1, 2, 4), the thickness of the Sandia Formation varies. At section A the exposed Sandia is 8 m thick, the contact with the basement is concealed. The section is composed of poorly exposed greenish-brownish shale with an intercalated coarse-grained sandstone bed (0.7 m). The sandstone is crossbedded, subarkosic in composition, and it contains pebbles up to 2 cm, plant fragments and amphibian bone fragments (Rinehart et al., 2011). At section B the exposed thickness is approximately 8 m, and the contacts with underlying basement rocks and the overlying Gray Mesa Formation are not exposed. The Sandia Formation at section B is composed of greenish micaceous shale and siltstone, and intercalated massive to indistinctly horizontally laminated, mostly coarse-grained, partly pebbly sandstone. At the top coarse-grained, arkosic sandstone (1.1 m) is exposed. At section C the Sandia Formation is thinner (approximately 6 m): the granitic basement is overlain by poorly exposed granitic grus, 2 m of quartzose pebbly sandstone, and 2.8 m of cover (probably siltstone-shale) followed by coarse, crinoidal limestone of the Gray Mesa Formation.

At Rio de las Vacas southeast of Cuba (Figs. 1-3) we assign the basal 63 m of the Pennsylvanian succession resting on the basement to the Sandia Formation: Granitic gneiss of the Proterozoic basement is overlain by sandstone. The lower 24 m are dominantly siliciclastic strata composed of reddish mudstone-siltstone, micaceous sandstone, pebbly sandstone and fine-grained conglomerate. The sandstone is partly crossbedded. Mudstone-siltstone intervals are up to 4.9 m thick, and sandstone and conglomerate units are up to 2.8 m. In the lower part, 0.5-1 m thick limestone units are intercalated. The limestone is recrystallized, indistinctly bedded and nodular. The overlying 39 m are composed of pebbly sandstone to fine-grained conglomerate (up to 1.5 m thick), thin bedded gray nodular limestone containing brachiopods (5.3 and 1.3 m thick), a thin limestone bed (0.2 m) and covered intervals up to 7.3 m thick. This succession is overlain by thin bedded and nodular gray limestone containing crinoidal debris and brachiopods, forming the base of the Gray Mesa Formation.

The northernmost Sandia Formation outcrops we examined are at Resumidero in the San Pedro Mountains (Figs. 1-3) where the Sandia Formation is approximately 13 m thick and overlain by the Gray Mesa Formation. The Sandia Formation is composed of 3.6 m of poorly exposed granitic grus, 0.7 m of coarse-grained, quartzose pebbly sandstone, followed by a covered interval that probably represents mudstone-siltstone, overlain by indistinctly bedded cherty limestone of the Gray Mesa Formation.

Locally, near the western and southern margin of the Sierra Nacimiento, the Sandia Formation is absent. At Log Springs near the southern end of the Sierra Nacimiento (Figs. 1, 2, 7) the Log Springs Formation is unconformably overlain by thin Osha Canyon Formation (5 m) and Gray Mesa Formation, indicating phases of uplift during Late Mississippian and Early Pennsylvanian time. At Coyote Flat west of Jemez (Figs. 1, 2, 7) the basement, which is composed of coarse-grained reddish granite with a paleorelief of up to several meters, is overlain by coarse-grained limestone (fossiliferous packstone containing abundant crinoid fragments and subordinately brachiopods and bryozoans) of the Guadalupe Box Formation, an equivalent of the Atrasado Formation

Rio de las Vacas

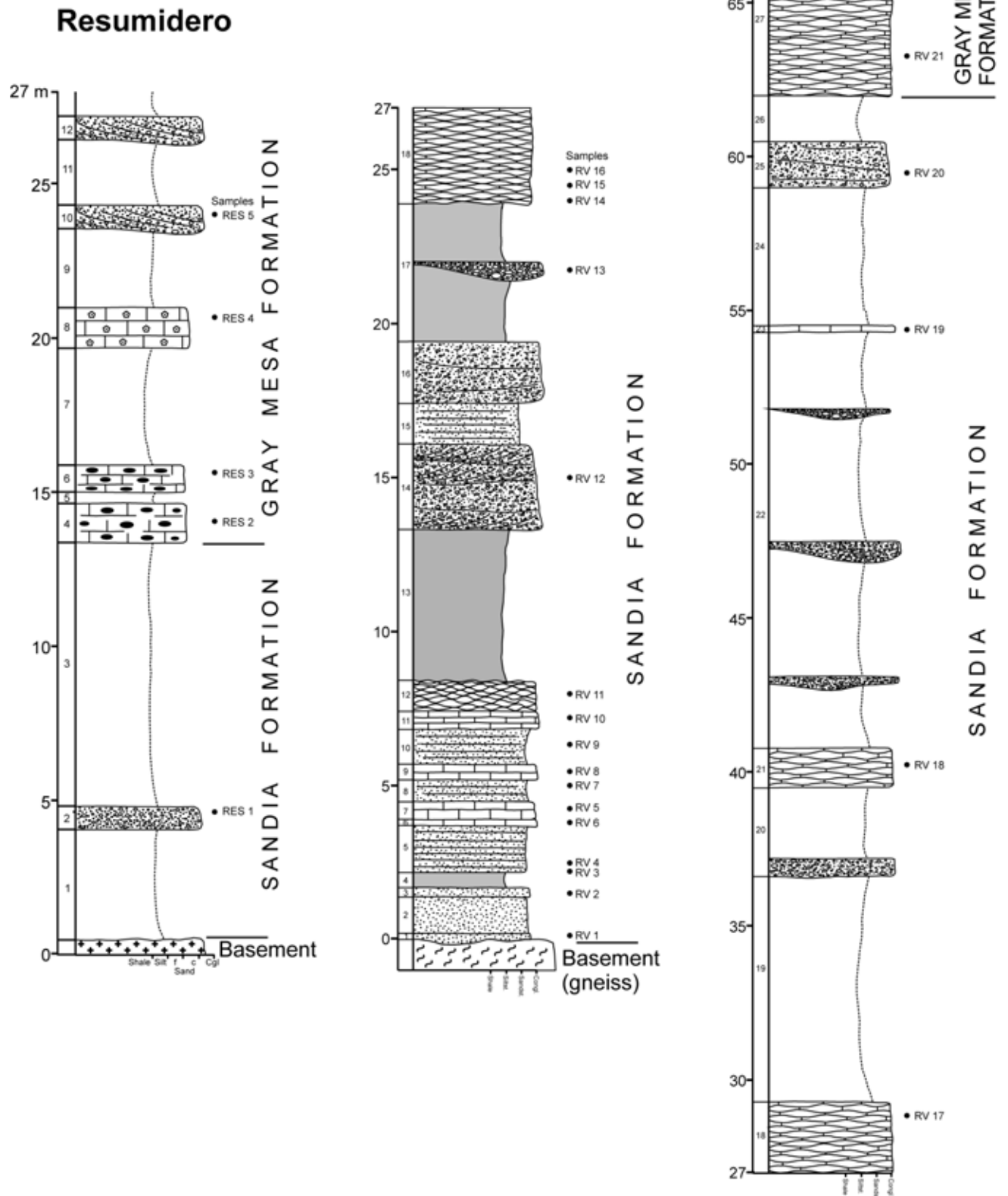


FIGURE 3. Measured sections of the Sandia Formation at Resumidero (San Pedro Mountain) and Rio de las Vacas (Sierra Nacimiento). For location see Figures 1 and 2, legend see Figure 6.

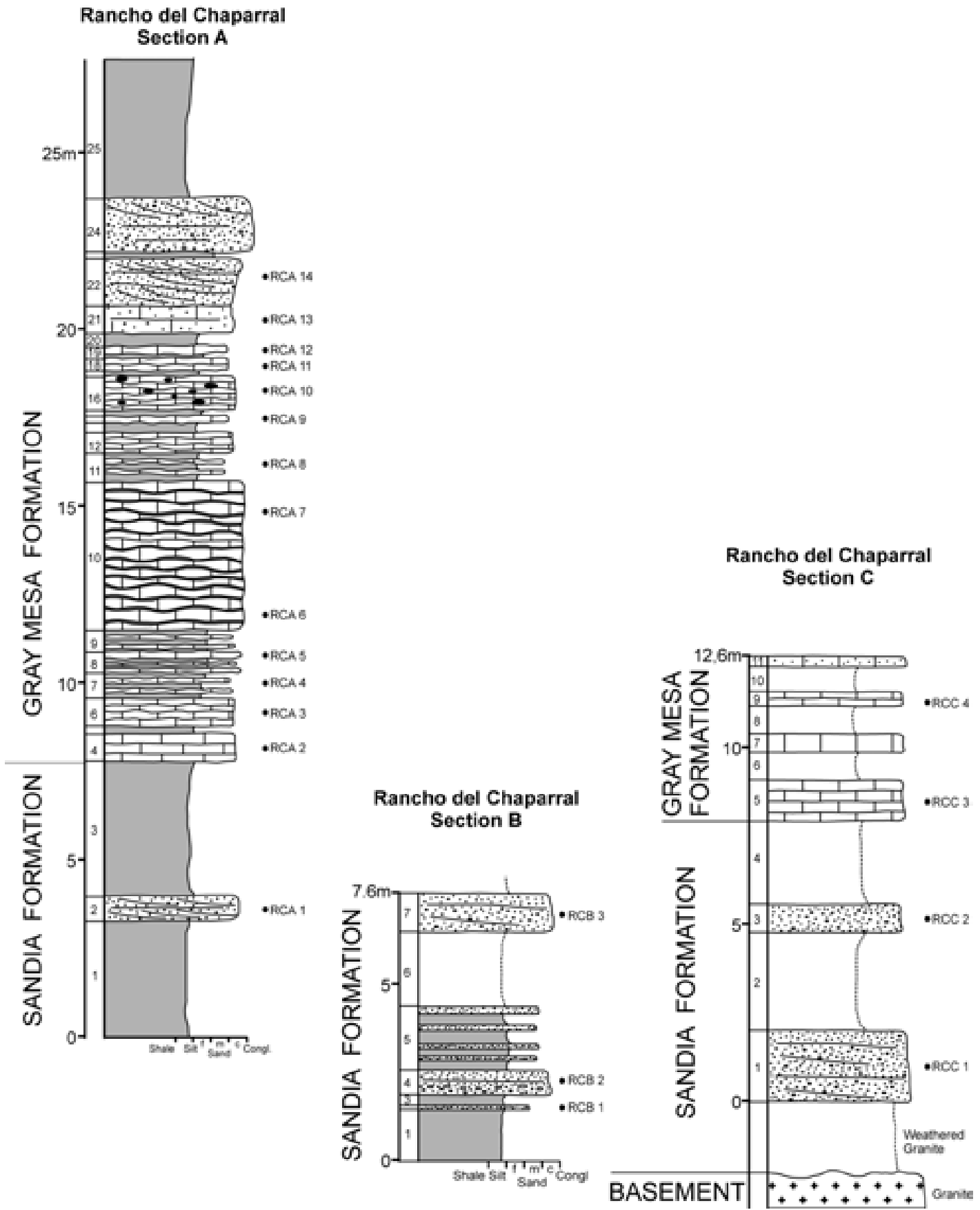


FIGURE 4. Measured sections of the Sandia Formation at Rancho del Chaparral (Sierra Nacimiento). For location see Figures 1 and 2, legend see Figure 6.

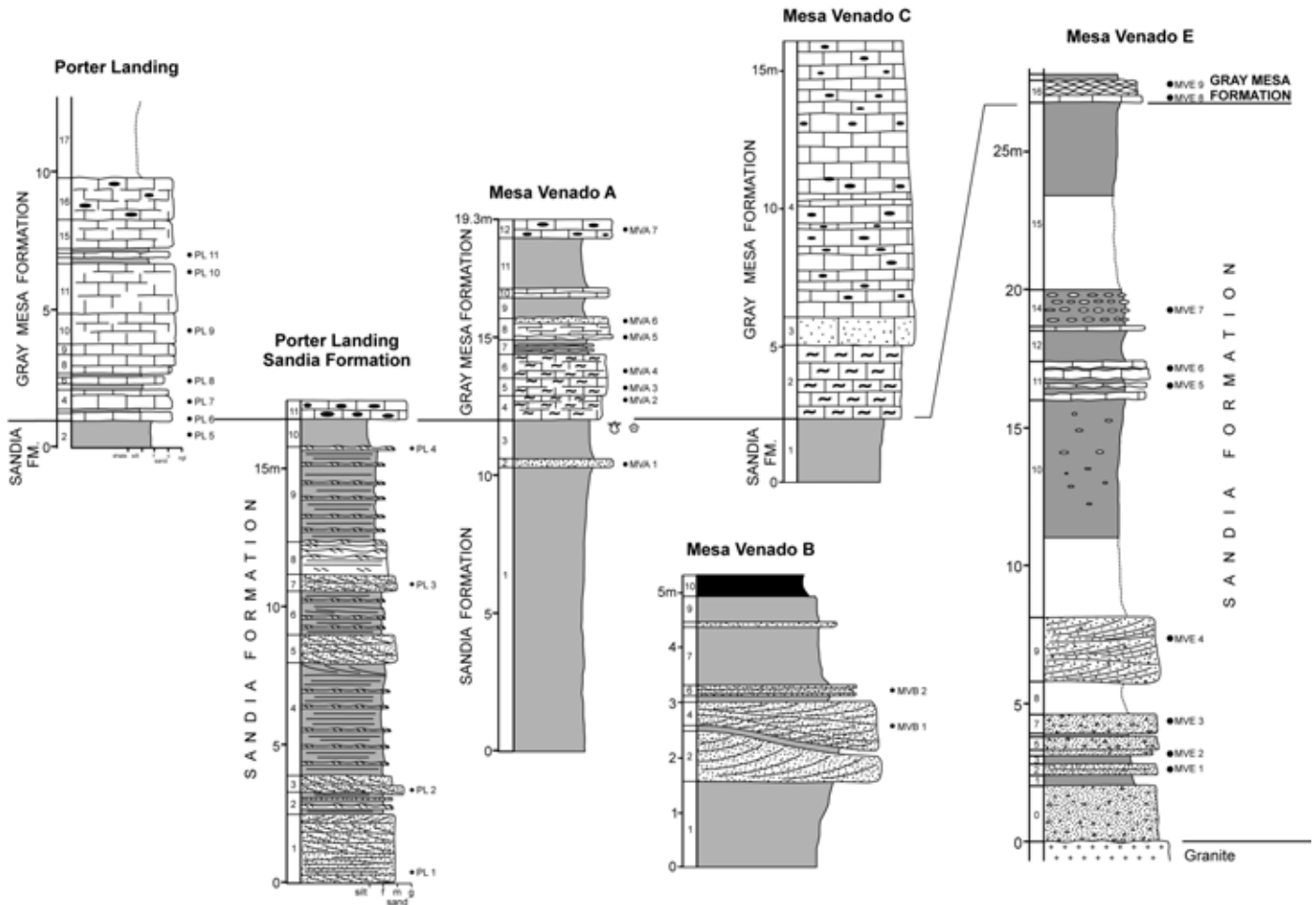


FIGURE 5. Measured sections of the Sandia Formation at Porter Landing and Mesa Venado (Sierra Nacimiento). For location see Figures 1 and 2, legend see Figure 6.

(late Desmoinesian-middle Virgilian; Krainer et al., 2005). In the area of Camp Zia northwest of Cuba (Figs. 1, 2, 7) the Precambrian basement is directly overlain by red beds of the Lower Permian Abo Formation, indicating that the Peñasco uplift existed as a positive high during the entire Pennsylvanian.

Manzanita Mountains

At Cedro Peak in the Manzanita Mountains (Fig. 1) the Sandia Formation is 13.6 m thick. The underlying Proterozoic basement rock is composed of folded phyllitic schist. The base of the Sandia Formation is 0.9 m thick massive greenish sandstone with small quartz pebbles up to 1-2 cm in diameter in its upper part. This sandstone overlies an erosional relief on the basement rocks. Above this basal sandstone follows a partly covered interval of gray shale overlain by 2.2 m thick interval of red mudstone to siltstone with small pebbles up to 1 cm floating in it. The uppermost unit is again a partly covered interval of probably gray shale, overlain by cherty limestone, which forms the base of the Gray Mesa Formation (section Y in Vachard et al., 2012, 2013).

Manzano Mountains

Southward, at Priest Canyon in the southern Manzano Mountains (Figs. 1, 8) the Sandia Formation is approximately 70 m thick, although the upper part is poorly exposed. Granite of the basement is overlain by 2 m of granitic grus and 0.4 m of coarse-grained, pebbly sandstone. The next 32 m are composed of several sandstone units, a few limestone intervals and covered intervals. The sandstone is mostly coarse-

grained, pebbly, and locally conglomeratic at the base. Sandstone units are 0.3-1.5 m thick, and commonly trough crossbedded, but horizontal lamination is also observed. Rarely, sandstone appears massive. The first limestone interval is exposed 19 m above the base of the formation. This interval is 0.7 m thick and contains fusulinids, *Syringopora* and *Chaetetes*. A thin limestone bed (0.2 m) is exposed 31 m above the base. The upper 38 m of the Sandia Formation are mostly covered (probably shale-siltstone), with one exposed limestone interval, which is 1.3 m thick and composed of three limestone beds. The basal limestone bed contains small quartz grains and rare chert nodules. The overlying bed is cherty limestone, contains silicified brachiopods and is bioturbated near the top. The upper limestone bed is thin and fossiliferous.

Socorro County

Northeast of Socorro, in the Joyita Hills, the Proterozoic basement is overlain by thin, dominantly siliciclastic Bursum Formation documenting the existence of a local ARM uplift and the absence of the Sandia Formation (Krainer and Lucas 2009, fig. 3).

Approximately 15 km farther south, at Arroyo de la Presilla east of Socorro (Fig. 1, 9), the Sandia Formation is 162 m thick, rests on granitic Precambrian basement (Fig. 10D) and consists of a cyclic succession of siliciclastic and carbonate, nonmarine and marine strata forming well developed transgressive cycles (Lucas et al., 2009a, b). The lower 46 m are almost entirely composed of siliciclastic sediments with only one thin limestone bed intercalated (Fig. 10C). The next 46 m are composed of mostly siliciclastic sediments with several intercalated fossilif-

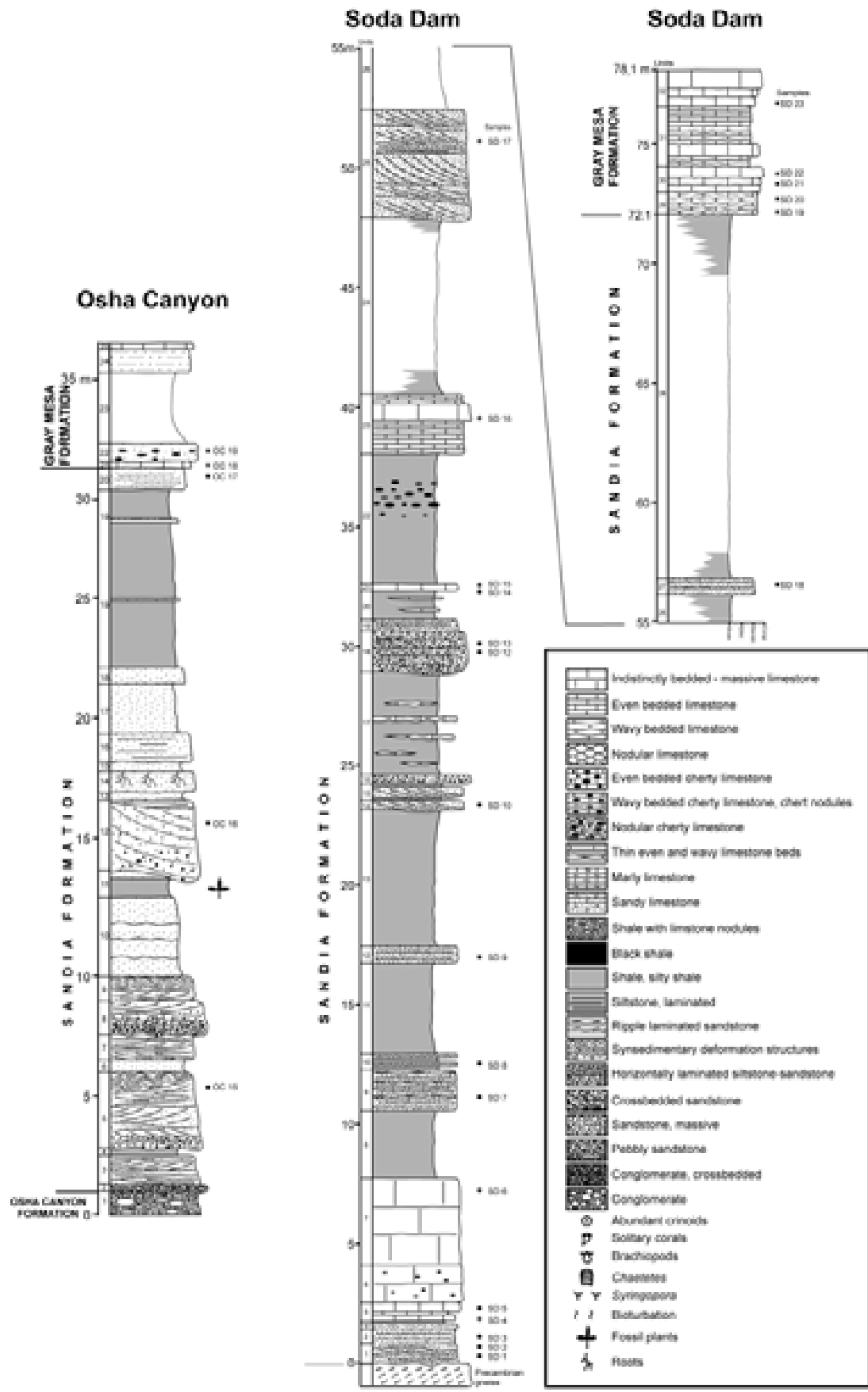


FIGURE 6. Measured sections of the Sandia Formation at Osha Canyon (Sierra Nacimiento) and Soda Dam (north of Jemez Springs). For location see Figures 1 and 2.

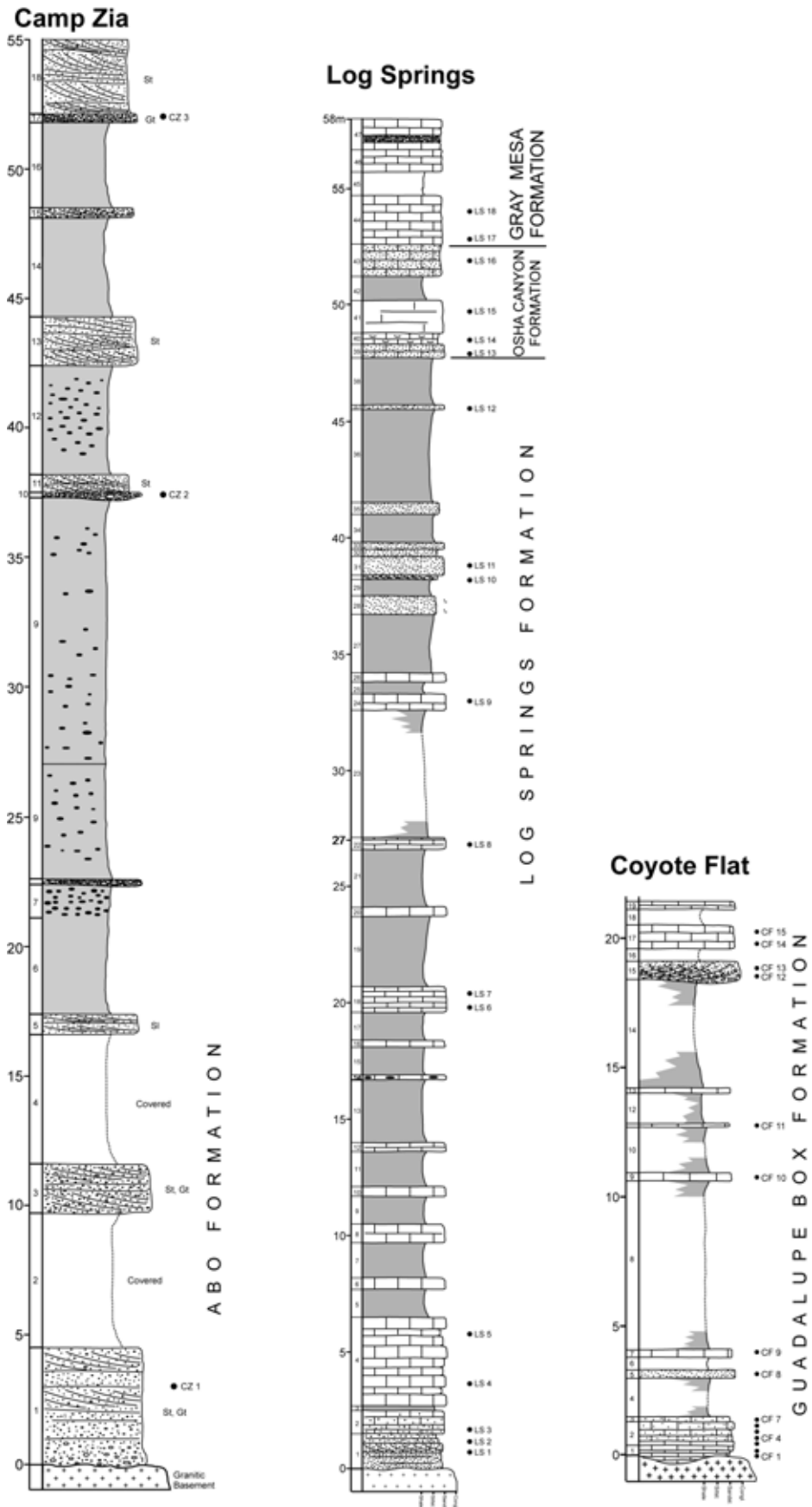


FIGURE 7. Measured sections of the Sandia Formation at Camp Zia, Log Springs and Coyote Flat (Sierra Nacimiento). For location see Figures 1 and 2, legend see Figure 6.

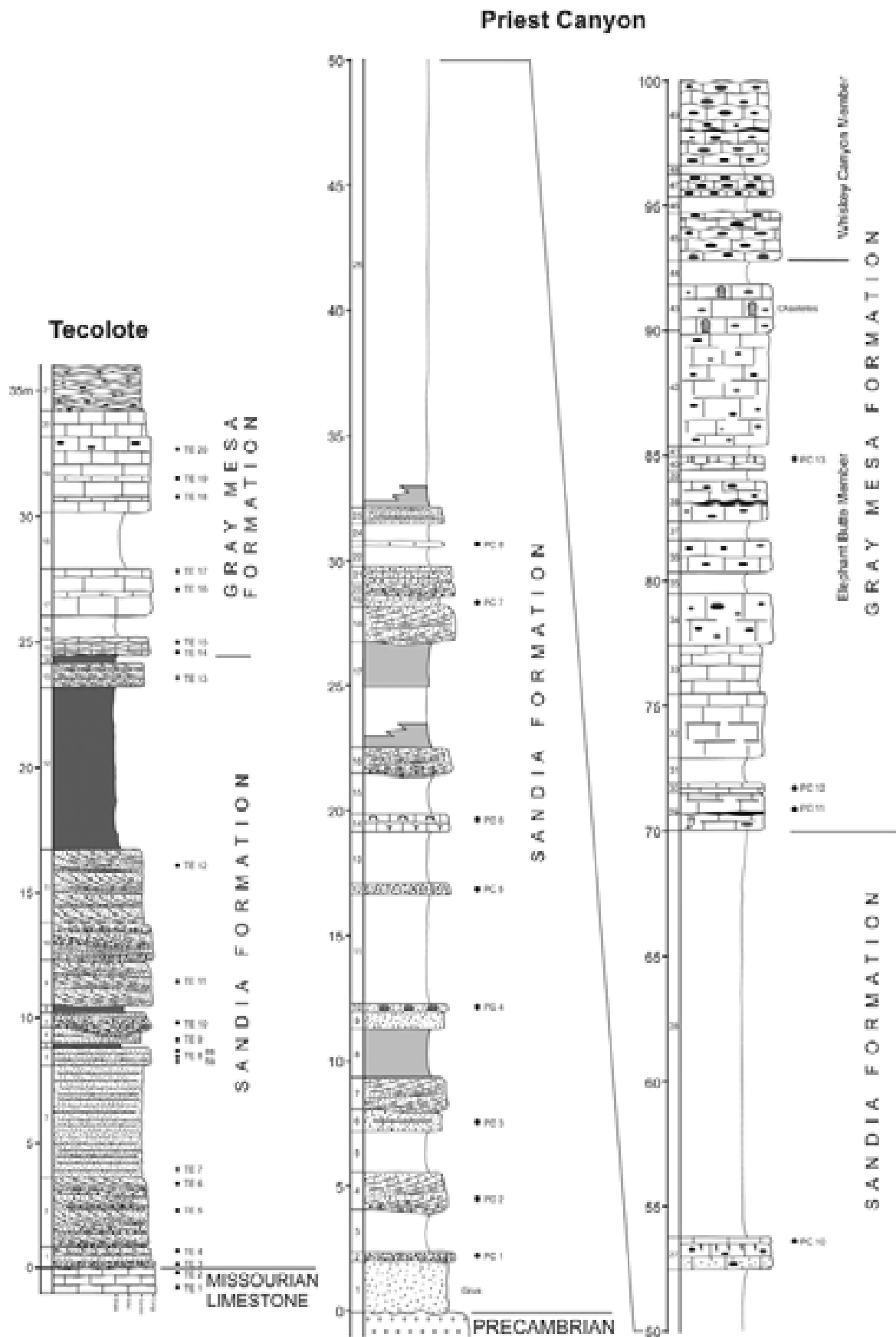


FIGURE 8. Measured sections of the Sandia Formation at Tecolote (northern end of the Sandia Mountains) and Priest Canyon (southern Manzano Mountains). For location see Figure 1, legend see Figure 6.

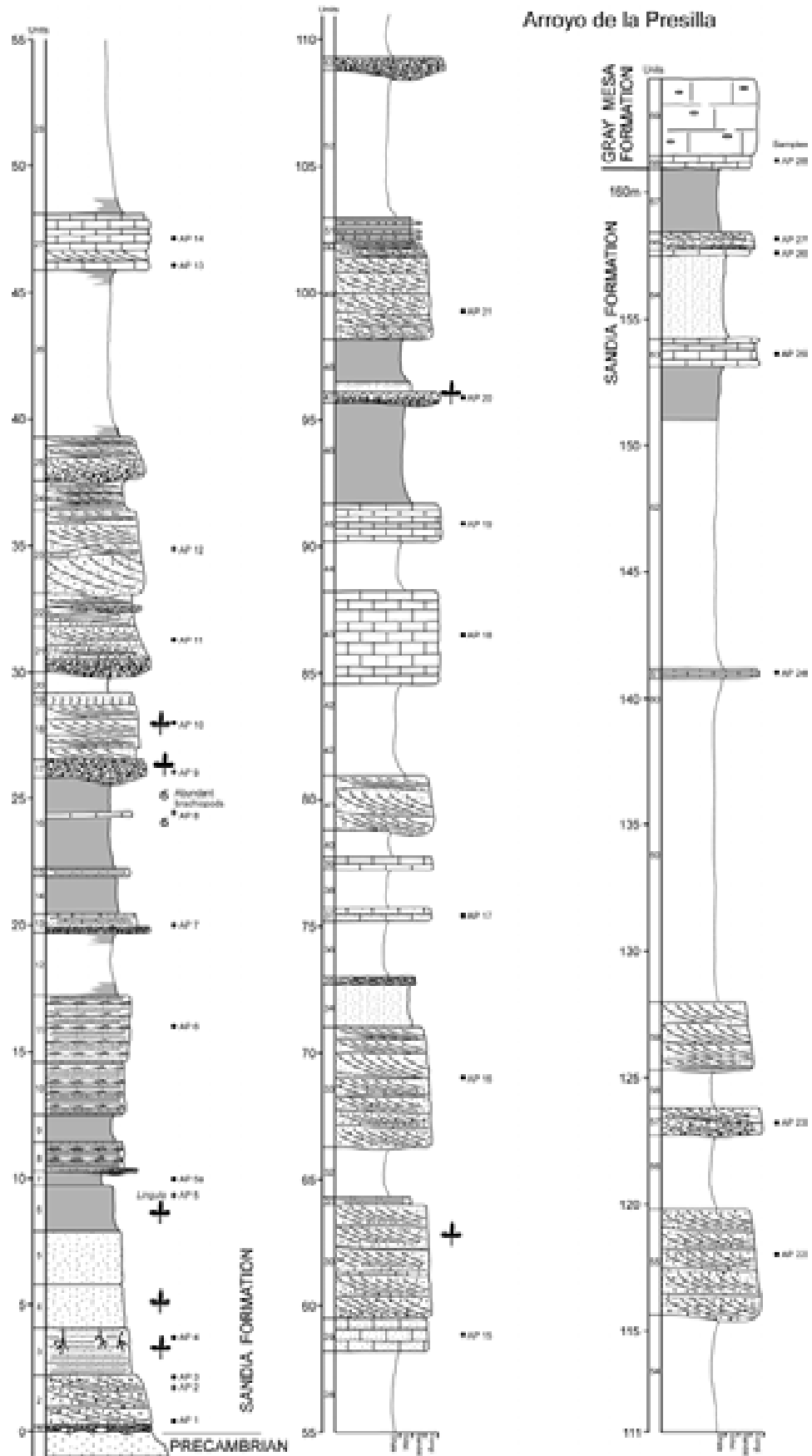


FIGURE 9. Measured section of the Sandia Formation at Arroyo de la Presilla in the Cerros de Amado east of Socorro. For location see Figure 1, legend see Figure 6.

erous limestone horizons. The uppermost 70 m are dominantly siliciclastics with thin limestone intervals in the upper part.

The following lithotypes are recognized: (a) Conglomerate (2.5%); (b) Coarse sandstone (23.5%); (c) Fine-grained sandstone - coarse-grained siltstone (11.2%); (d) Shale- fine siltstone (11.5%, covered 44.3%); and (e) Limestone (7%).

Conglomerate is 2 m thick at the base, up to about 1 m higher in the section, grading into pebbly sandstone and sandstone. It is fine-grained with a maximum grain size around 3 cm; grains are mostly subrounded, and angular to subangular near the base. At the base a thin lag with granite boulders up to 20 cm in diameter is developed (Fig. 10D). The conglomerate is quartz-rich, and poorly to moderately sorted. Conglomerate units display an erosive base and are indistinctly to distinctly trough cross-bedded (Gt). Conglomerate units show normal grading (conglomerate at the base, grading into pebbly sandstone, then coarse and finally fine sandstone).

Coarse-grained sandstone is commonly trough cross-bedded (St), rarely displays planar cross-bedding (Sp) and may be pebbly. Sandstone is quartz-rich, reddish, and individual quartz grains are up to 1-2 cm in diameter. Sandstone intervals are up to 4.7 m thick, fining upward, and composed of multistorey channel fills. Individual sandstone beds contain fossil plant fragments, including stem fragments > 1 m long. Sandstone intervals display an erosive base (type 1 sequence boundary). Rarely (unit 53), mudstone clasts (rip-up clasts) up to 10 cm in diameter are present.

Thicker siltstone to fine-grained sandstone intervals occur in the lower part of the section, below and above the "fire clay" of Herrick (1904), which was being mined for brick manufacturing. The thickness of these intervals is from 1.2 to 4.7 m. Thinner siltstone/fine-grained sandstone layers are also developed on the top of conglomerate-sandstone units (up to 1.3 m thick) and rarely as thin (0.3 m) intercalations in shale.

The most common lithofacies are horizontally laminated (Sh) and ripple laminated (small-scale current ripples; Sr) siltstone to fine-grained sandstone. Small-scale trough cross-bedding (St) is also observed. Rarely, bioturbated fine-grained sandstone is present.

The thicker siltstone to fine-grained sandstone unit above the fire clay is composed of alternating, horizontally and ripple-laminated siltstone and fine-grained sandstone, overlying thin channel-fill sandstone with an erosive base. Small-scale trough cross-bedding (small channel fills) is observed on top of the conglomerate-sandstone units. The lowermost siltstone to fine-grained sandstone contains fossil plants and root structures. The overlying yellowish-brownish siltstone contains abundant impressions of *Lepidodendron* and other plants (Lucas et al., 2009b).

The "fire clay" in the lower part of the section is light gray claystone, 0.5 m thick and contains plant fossils. It is underlain by dark gray, laminated, silty clay containing *Lingula* and plant fossils. The silty clay intercalated in the thicker horizontally and ripple-laminated siltstone to fine-grained sandstone interval above the fire clay is gray and laminated. From 21 to 26 m above the base greenish-brownish silty shale is poorly exposed with a thin micaceous sandy siltstone intercalated in the lower part and a thin fossiliferous limestone in the upper part. The shale immediately below and particularly above the limestone contains abundant marine fossils such as crinoids, bryozoans, brachiopods and rugose corals. In the middle and upper part of the section, shale intervals are mostly covered and up to 6.6 m thick in the middle part and up to 12.8 m thick in the upper part. Marine fossils such as crinoids and brachiopods occur in a 4 m thick brownish shale in the middle part of the section (unit 46).

In the lower part of the section only one thin limestone bed (10-20 cm thick) is poorly exposed. This limestone bed is fossiliferous and contains brachiopods and bryozoans.

In the middle and upper part limestone intervals are 0.3-2.2 m thick, commonly brownish weathered, gray to dark gray, 5-30 cm bedded. Typically the limestones are coarse-grained, sandy and fossiliferous with abundant fragments of crinoids, brachiopods, bryozoans and sub-

ordinately solitary corals (Fig. 10E). Rarely the limestone displays cross-bedding (unit 27) or appears massive. Several types of microfacies are observed: wackestone, packstone, floatstone and rudstone.

The uppermost thin limestone bed (unit 65) is a recrystallized, non-laminated, fine-grained dolomitic mudstone.

At Hansonburg in the northern Oscura Mountains of Socorro County, the southernmost area examined during this study, the lower 12.5 m of the Pennsylvanian section are represented by the Sandia Formation, which overlies red granite and a 0.8 m thick layer of granitic gneiss (Lucas and Krainer, 2009). The lower 8.5 m of the Sandia Formation consist of brown, sandy shale with three intercalated thin, sandy, bioclastic limestone beds in the middle and one in the upper part. The upper 4 m are composed of thin-bedded nodular limestone (0.8-1.1 m), thin sandy bioclastic limestone beds, shale with limestone nodules, gray shale and a thin covered interval. The Sandia Formation seems to be entirely marine and shows a transgressive sequence with shale and sandy limestone beds in the lower part and limestone with less siliciclastic material and less shale intercalations in the upper part. The boundary with the overlying Gray Mesa Formation is drawn at the base of the first cherty limestone interval.

Sandstone Petrography

Sandstone of the Sandia Formation is mostly classified as quartzarenite and subarkose, rarely as lithic arenite. Monocrystalline quartz is by far the most abundant grain type; subordinately polycrystalline quartz and detrital feldspars are present. Rock fragments and other grain types are rare to absent. Sandstone is commonly cemented by quartz, which occurs as authigenic overgrowths, and by calcite cement. Matrix is locally present in small amounts, but mostly absent.

In the Sierra Nacimiento (e.g., Guadalupe Box, Porter Landing) sandstone of the Sandia Formation is commonly moderately sorted subarkose with dominantly subangular to subrounded grains (sandstone classification after Pettijohn et al., 1987). The sandstone is composed of abundant monocrystalline quartz, subordinate polycrystalline quartz, detrital feldspars (dominantly untwinned potassium feldspar, some microcline and perthitic feldspars, rarely plagioclase), rare detrital mica (muscovite, subordinate biotite), rare granitic rock fragments and very rare fine-grained metamorphic rock fragments. It contains pseudomatrix (sensu Dickinson, 1970), most likely derived from the alteration of detrital feldspars. Detrital quartz grains commonly display authigenic overgrowths, and locally the pore space is filled with microcrystalline quartz cement.

At Rio de las Vacas the lowermost sandstone is calcite-cemented quartz arenite with a few detrital feldspars and very rare granitic and metamorphic rock fragments. Higher in the section sandstone is subarkose containing many detrital feldspars (mostly potassium feldspars) and a few granitic rock fragments. The sandstone is cemented by coarse blocky calcite or contains brownish-stained matrix.

At Soda Dam, sandstone near the base of the Sandia Formation is carbonate-cemented subarkose (Plate 1-3), whereas higher in the section sandstone is quartz arenite that is moderately to well sorted and contains subrounded to rounded grains (Plate 1-1). The sandstone is composed mainly of monocrystalline quartz, a few polycrystalline quartz and rare chert grains. Detrital micas are very rare; detrital feldspars and rock fragments are almost absent. The sandstone is well cemented by authigenic quartz overgrowths; locally some carbonate cement is present. Also, locally some pseudomatrix derived from in situ alteration of detrital feldspars is present. Higher in the section (SD 7-18) is quartz arenite with very few detrital mica and rare detrital feldspars. Quartz arenite is well cemented by authigenic quartz overgrowths, and locally by carbonate cement.

Moving to the Sandia Mountains at Tecolote, sandstone in the lower part of the Sandia Formation is quartz arenite that contains pseudomatrix, some primary clayey matrix, and rare authigenic overgrowths. In the upper part of the section sandstone contains a few

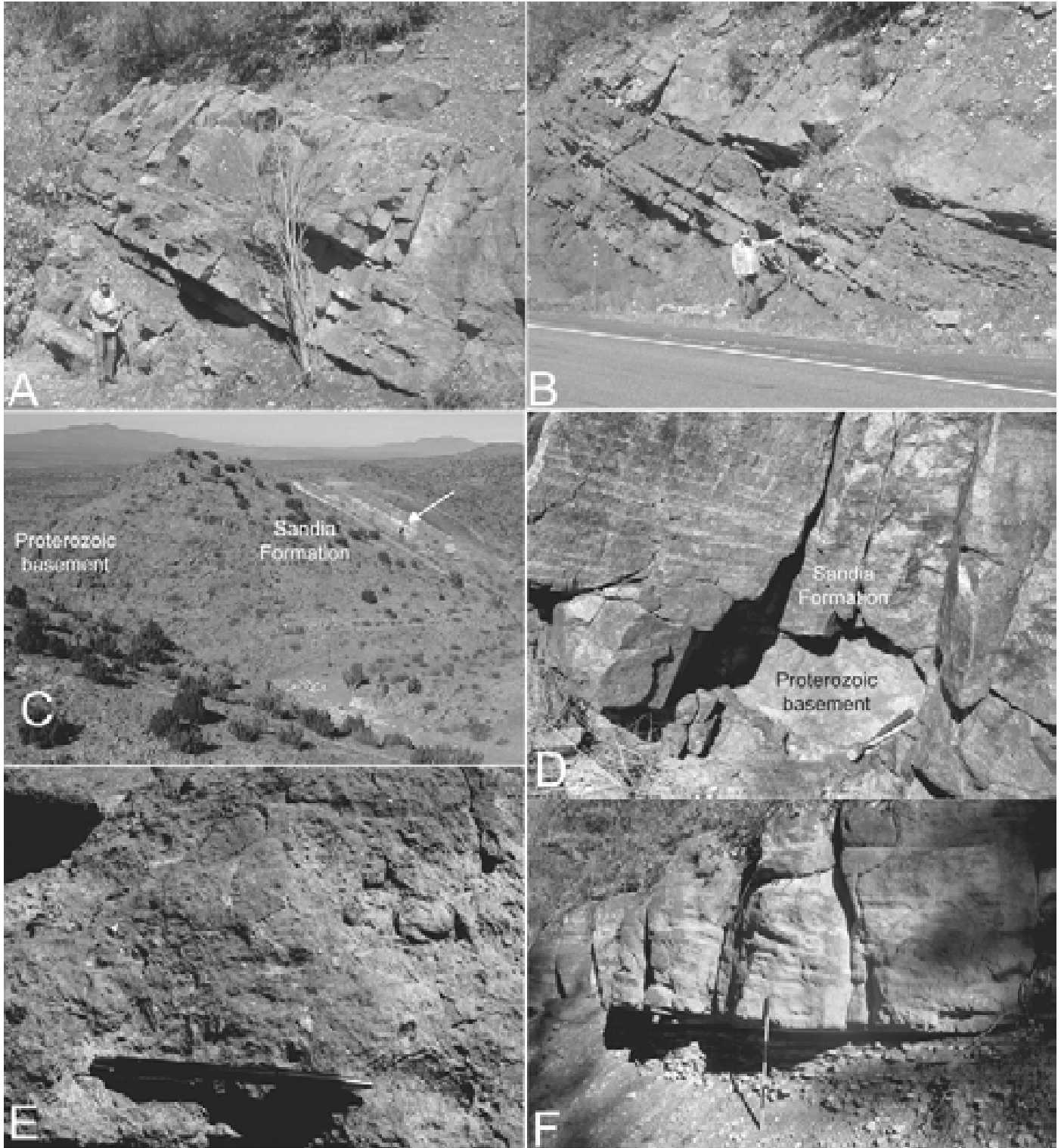


FIGURE 10. Photographs of selected outcrops of the Sandia Formation. **A-B**, Lectostratotype section at Doc Long in Sandia Mountains, Bernalillo County; **A**, Base of the section---base of the measuring stick is at the contact between Proterozoic basement (below) and Sandia Formation (above); **B**, Interbedded shale, sandy limestone and quartz-rich sandstone in lower part of Sandia Formation. **C-D**, Lower part of Arroyo de la Presilla section of the Sandia Formation in Socorro County; **C**, Overview of base of section showing Proterozoic granite depositionally overlain by basal sandstone sheet of Sandia Formation; arrow indicates adit in “fire clay;” strata below the adit yield a lycopsid-dominated flora; **D**, Detail of scoured contact of base of Sandia Formation on granitic basement. **E**, Characteristic Sandia Formation conglomeratic limestone with sedimentary rip-up clasts and bioclastic debris, Arroyo de la Presilla section. **F**, Crossbedded subarkosic sandstone bed of Sandia Formation at Porter Landing section.

detrital feldspars, many altered rock fragments (sublitharenite; Plate 1-2, 1-5), pseudomatrix and common authigenic overgrowths on detrital quartz grains. In the middle of the section, coarse-grained, poorly sorted mixed siliciclastic-carbonate sandstone and rudstone containing small amounts of siliciclastic grains are present. Both contain echinoderms (mostly crinoid fragments) brachiopods, rare bryozoans, ostracods, brachiopod spines, smaller foraminifers and trilobites (Plate 2-8).

At the type section, sandstone in the lower part of the Sandia Formation is quartz arenite. In the middle and upper parts of the formation, some detrital feldspars are present (subarkose); for details see Krainer et al. (2011).

To the south at Priest Canyon, subarkosic sandstone contains a few granitic rock fragments, some matrix and rare quartz overgrowths. Quartz arenite cemented by quartz overgrowths and microcrystalline quartz cement is also present. Mixed siliciclastic-carbonate sandstone contains abundant quartz grains, common detrital feldspars, some granitic rock fragments, a few carbonate grains and fossil fragments (echinoderms, brachiopods, rare bryozoans) embedded in micritic matrix.

There are no significant changes in composition and texture throughout the sandstones of the Sandia Formation at Arroyo de la Presilla (Lucas et al., 2009b).

Sandstone is composed of abundant monocrystalline quartz, subordinate polycrystalline quartz, rare detrital feldspars, which are almost completely altered to clay minerals ("pseudomatrix"), very rare detrital muscovite and rock fragments of quartz and feldspar (granitic), rare chert grains and phyllitic (metamorphic) rock fragments. The sandstone is classified as quartz arenite and is cemented by authigenic quartz overgrowths, and, locally, by fine crystalline quartz cement in pore space (Plate 1-4). Some clayey matrix (<5%) may be present. Feldspar content is < 5% throughout the succession (quartz arenite). Locally, small patches of coarse calcite cement replacing quartz or fine-crystalline carbonate cement occur. In the upper part of the Sandia Formation, mixed siliciclastic-carbonate fossiliferous sandstone is present that is medium- to coarse-grained, subangular to subrounded and poorly sorted. Most abundant is monocrystalline quartz, subordinate polycrystalline quartz, some micritic carbonate grains and a few detrital micas. Fossils include shell fragments of gastropods and brachiopods, crinoid fragments, a few brachiopod spines, rare ostracods, and large micritic rip-up clasts (several cm long) containing a few gastropods, bryozoans and echinoderms.

Carbonate Microfacies

Limestone of the Sandia Formation is composed of various microfacies types, including rare mudstone, abundant wackestone (various types) that may grade into packstone and floatstone, and grainstone to rudstone. Many limestone samples contain siliciclastic grains; mixed siliciclastic-carbonate fossiliferous sandstone is also present. Except for mudstone, the microfacies types contain a diverse fossil assemblage. In grainstone and rudstone the fossils are strongly fragmented.

In the Sierra Nacimiento, limestones are intercalated in the Sandia Formation at our measured sections at Rio de las Vacas, Mesa Venado E and Soda Dam. At Rio de las Vacas, limestone intervals are 0.2-5.3 m thick. In the lower part, limestone is indistinctly bedded, gray and strongly recrystallized. In the middle part of the section two limestone intervals (5.3 and 1.3 m thick) are developed, composed of gray, thin, wavy bedded to nodular fossiliferous limestone containing crinoidal debris and brachiopods. Microfacies are phylloid algal wackestone, bioclastic mudstone and wackestone to floatstone containing a diverse fossil assemblage. In the upper part, a thin limestone bed is intercalated. The basal part of the overlying Gray Mesa Formation is composed of nodular gray limestone (bioclastic wackestone to floatstone).

At Soda Dam, a 6-m-thick limestone succession is intercalated in the lower part of the Sandia Formation that is composed of strongly recrystallized, bedded limestone near the base, grading into indistinctly bedded to massive cherty limestone. In the middle of the section, thin

nodular limestone beds and lenses composed of mudstone with few bioclasts are intercalated in shale. In the middle a coarse-grained arenaceous limestone composed of grainstone (Plate 1-6) is developed. A 0.3-m-thick limestone bed contains fusulinids (fusulinid wackestone – Plate 2-2), overlain by grainstone-packstone (Plate 2-6) and rudstone containing a diverse fossil assemblage (Plate 2-7). The uppermost limestone unit within the Sandia Formation is 2.5 m thick and composed of wavy bedded marly limestone with one intercalated limestone bed composed of peloidal mudstone with a few fossil fragments.

At the lectostratotype section of the Sandia Formation, sandy limestone horizons are 0.2-2 m thick and mostly appear massive. Sandy limestone composed of mixed siliciclastic-carbonate sandstone and rudstone contains few quartz grains. Rarely, crinoidal packstone is present. Limestone intervals are 0.2-2 m thick, gray and fossiliferous. Thicker limestone intervals are commonly thin, even to wavy bedded. The most common microfacies types are wackestone and packstone, which may grade into floatstone and rudstone (see details in Krainer et al., 2011). The middle part of the lectostratotype section of the Sandia Formation is composed of dm-bedded gray limestone, which is locally bioturbated and in the uppermost 3.9 m contains abundant chert nodules and a thin layer with abundant bryozoans. Limestone is composed of rudstone, floatstone and wackestone

At Priest Canyon, a few thin limestone horizons are intercalated in the Sandia Formation that are 0.2-1.3 m thick. The lowermost limestone is 0.7 m thick, with the lower limestone bed containing fusulinids (fusulinid wackestone, Plate 1-1) and *Syringopora*, and the upper bed containing *Chaetetes*. A thin limestone bed in the middle of the formation is composed of grainstone/packstone (Plate 2-3) containing a diverse fossil assemblage. The uppermost limestone interval is 1.3 m thick, and medium to thick bedded with quartz grains in the lower part. The limestone contains chert nodules and silicified brachiopods.

In the lower part of the Arroyo de Presilla section, only one thin limestone bed (10-20 cm thick) is poorly exposed. This limestone bed is fossiliferous and contains brachiopods and bryozoans. The microfacies of this limestone is coarse-grained, poorly sorted bioclastic floatstone to rudstone containing a few bioclasts > 1cm (Plate 2-5). The most abundant fossils are bryozoans and crinoids; subordinate are brachiopod shell fragments and brachiopod spines, echinoderm spines, gastropods and trilobite fragments, rare ostracods and smaller foraminifers (*Endothyra*). Many skeletons are encrusted by cyanobacteria, *Calcivertella* and *Clara crusta*. The rock contains some micritic matrix; in densely packed parts calcite cement is present. The rock is recrystallized.

The bioclastic floatstone to rudstone in the middle part is coarse-grained, poorly sorted and contains a few bioclasts (shell fragments) up to several cm in diameter. Abundant bioclasts (strongly fragmented) are bryozoans, crinoids and brachiopod shell fragments; subordinate are trilobite fragments, ostracods, and brachiopod spines. In the upper part of the section (unit 61), fusulinids and gastropods are present in a strongly altered, dark brown to black-stained limestone bed. Quartz grains (0.1-0.3 mm, rarely up to 1 mm) are also present in amounts up to 10%. A few bioclasts are encrusted by *Calcivertella*. The rock contains micritic matrix, and locally some calcite cement.

Crinoidal packstone (Plate 2-4) is coarse-grained, moderately to poorly sorted, and has a grain size of mostly 0.5-2 mm, rarely > 1 cm. This microfacies contains abundant isolated crinoid ossicles, and detrital, angular to subangular quartz grains (monocrystalline, subordinately polycrystalline quartz mostly 0.3-1mm; mixed siliciclastic carbonate sandstone, Plate 1-7). Subordinate are bryozoans, brachiopod shell fragments and spines and rare trilobite fragments and ostracods. A few bryozoan fragments and rare crinoids are encrusted by *Calcivertella*. This type is well washed and calcite cemented. Quartz is present in various amounts, mostly < 5%; individual thin layers contain 20- 70% quartz. The uppermost dolomitic mudstone contains abundant spicules and rare smaller foraminifers.

Regional Variation

The Sandia Formation is characterized by marked changes in thickness and lithology over short distances, which is well documented in the Sierra Nacimiento, the Sandia Mountains and east of Socorro. Lithology ranges from entirely siliciclastic, to dominantly siliciclastic with intercalated limestone.

In the Sierra Nacimiento, the Sandia Formation is locally absent like at Camp Zia east of Cuba where the Abo Formation rests on Proterozoic basement or at Coyote Flat where the Guadalupe Box Formation rests on basement. At Rancho de Chaparral, locally the Sandia Formation is absent and the Gray Mesa Formation directly rests on basement. Locally thin Sandia Formation (7.6+ m) rests on the Proterozoic basement rocks, overlain by Gray Mesa Formation. At Log Springs the Gray Mesa Formation rests on the Osha Canyon Formation, and the Sandia Formation is absent. At Guadalupe Box the Sandia Formation is 31.2 m thick, completely siliciclastic, and rests on the Osha Canyon Formation and is overlain by the Gray Mesa Formation. At Resumidero and Porter Landing the Sandia Formation is thin, entirely siliciclastic and rests on Proterozoic basement: Resumidero 12.9 m, Porter Landing 16.8 m+. At Mesa Venado, section E of the Sandia Formation is 27 m thick. The lower part is entirely siliciclastic and nonmarine, the upper part is composed of marine shale and intercalated fossiliferous limestone. At Rio de las Vacas, the lowermost ~ 63 m of dominantly siliciclastic sediments resting on Proterozoic granitic gneiss are assigned to the Sandia Formation (though there is no biostratigraphic evidence to support this), overlain by Gay Mesa Formation.

At Soda Dam north of Jemez Springs the Sandia Formation is 72 m thick, rests on Proterozoic gneiss and is overlain by the Gray Mesa Formation. Siliciclastic sediments comprise 87% and limestone 13% of the section.

The thickness of the Sandia Formation in the northern Sandia Mountains varies markedly due to the paleorelief of the basement, and Toomey (1953) reported thicknesses of 49 to 86 ft (15 to 26 m). At Tecolote near the northern end of the Sandia Mountains, the Sandia Formation rests on Mississippian limestone, is 24.4 m thick and entirely siliciclastic. At Montezuma Mountain on the northern flank of the Sandia uplift, the Sandia Formation is 159 ft (48 m) thick according to Reynolds (1954). Catacosinos (1962) stated that the Sandia Formation in the Sandia Mountains ranges from 175 ft (53 m) at the Kiwanis Cabin section to about 300 ft (91 m) at the Thumb Peak section.

At the lectostratotype section (Doc Long) the Sandia Formation measures 124 m, rests on Proterozoic basement and consists of 77% siliciclastic sediments and 23% limestone. The Sandia Formation is only 13.6 m thick and entirely siliciclastic at Cedro Peak in the Manzanita Mountains just about 15 km SSE of the type section.

At Priest Canyon in the southern Manzano Mountains the Sandia Formation again rests on Proterozoic basement, is 70 m thick and mostly siliciclastic with only about 4% limestone.

The thickest Sandia section we studied is exposed at Arroyo de la Presilla east of Socorro, measuring 162 m. Here, the Sandia Formation overlies Proterozoic granite and consists dominantly of siliciclastic sediments and about 7% limestone. At Hansonburg in the southern Oscura Mountains the Sandia Formation thins to 12.5 m, rests on basement and contains about 16% limestone.

Towards the south (Fra Cristobal, Mud Springs Mountains, Caballo Mountains, San Andres Mountains) the Sandia Formation interfingers with and is replaced laterally by the Red House Formation, which is dominantly carbonate with minor amounts of siliciclastic sediments (Lucas et al., 2012).

Contacts

Lower Contact

The Sandia Formation at most places rests on Proterozoic basement with an angular unconformity. Locally, the Sandia Formation over-

lies older sedimentary rocks such as the Osha Canyon Formation (Guadalupe Box) or thin Mississippian limestones (Tecolote).

Upper Contact

At most places the dominantly or entirely siliciclastic sediments of the Sandia Formation are conformably overlain by thicker limestone/cherty limestone units of the Gray Mesa Formation. The contact is a sharp lithologic boundary. Locally, like at Porter Landing in the Nacimiento Mountains, the Sandia Formation is overlain by the Guadalupe Box Formation (?).

PALEONTOLOGY AND AGE

The fossil assemblage of the lectostratotype section of the Sandia Formation is diverse and contains representatives of most late Paleozoic marine fossil groups (Krainer et al., 2011). Fragments of echinoderms (crinoids), brachiopods and bryozoans are by far the most abundant fossils observed in thin section throughout the section. Brachiopods include impunctate and punctate shells and spines. Bryozoans include trepostome, fistuliporid and fenestrate forms. Completely recrystallized and commonly fragmented bivalves and gastropods are subordinate.

From the type section Krainer et al. (2011) recognized smaller foraminifers that are present although quite rare in all limestone samples: *Bradyina* (Plate 3-5), *Calcitornella* (Plate 3-8), *Calcivertella* (Plate 3-8, 3-11), *Climacamina* (Plate 3-12), *Cornuspira multivoluta*, *Earlandia* ex gr. *elegans*, *Endothyra*, *Endothyranella*, *Eotuberitina reitlingerae*, *Globivalvulina* ex gr. *G. bulloides* (Plate 3-8), *Glomospira*, *Palaeonubecularia*, *Planoendothyra aljutovica*, *Spireitlina conspecta*, *Tetrataxis* ex gr. *T. paraconica*, *Tetrataxis* ex gr. *T. acuta*, *Trepeilopsis* and *Turrispiroides*. Fusulinids are very rare at the type section and include *Eostaffella pinguis*, *Profusulinella* (*P. cf. P. millecreekensis*, *P. spicata* and *P. cf. P. copiosa*) (Plate 3-1, 3-2, 3-3, 3-4), *Pseudostaffella* and *Schubertellina*. Trilobite fragments are present in limestone throughout the type section.

Calcisponges (sphinctozoan) are locally common in limestone of the middle part of the Sandia Formation, and rare in the lower part. Corals are very rare. Sponge spicules and very rare sponge fragments occur. The spicules are derived from siliceous sponges, but are all calcified. Ancestors of *Tubiphytes* are very rare in the middle part of the formation.

Ostracods are common. *Spirorbis* (calcareous worm) is very rare. Algae are very rare. At the type section the red alga *Lysvaella* or *Hortonella*, a possible coralline alga, is present; in another sample a stracheiid alga was observed. A thin limestone bed in the upper part of the type section contains *Komia* (ungdarellacean alga).

Commonly, crinoids and other skeletal grains are encrusted by laminar cyanobacteria, locally forming oncoids. Rarely, skeletons are encrusted by tubular *Girvanella*. Phosphatic fossil fragments, probably derived from fishes, occur in few samples at the type section.

Among macrofossils are scarce brachiopods, bryozoans, and plant fossils.

From dark shales, Toomey (1953) described fossil plant fragments of *Neuropteris scheuchzeri*, *N. tenuifolia*, *Cordaites* sp. and *Cardiocarpon* sp., and brachiopods from limestones indicating an Atokan to early Desmoinesian age. *Eostaffella pinguis*, which occurs in the lowermost part of the Sandia Formation at its type section, indicates an early Atokan or late Morrowan age (Krainer et al., 2011). The occurrence of *Profusulinella* in the higher part of the Sandia Formation and in the basal Gray Mesa Formation indicates that most of the Sandia Formation and also the basal Gray Mesa Formation are of Atokan age (Krainer et al., 2011).

At Tecolote a diverse marine fossil assemblage is present in mixed siliciclastic-carbonate sandstone in the middle of the section.

At Soda Dam limestones contain a similarly diverse fossil assemblage including mainly fragments of echinoderms, bryozoans and bra-

chiopods; subordinately gastropods, foraminifers, ostracods and trilobite fragments occur. Limestone in the middle of the Sandia Formation contains *Fusulinella* (Plate 3-1 – 3-4), *Millerella* and *Eostaffella*, indicating an Atokan age. Additionally, *Climacammina*, *Endothyra*, *Schubertellina* and *Tetrataxis* are present. *Komia* (Plate 3-9) and *Paraepimastopora* were also recognized.

At Guadalupe Box the packstone on top of the succession contains abundant echinoderm fragments, subordinately skeletons of brachiopods, bryozoans, foraminifers, and ostracods.

At Rio de las Vacas intercalated limestones contain echinoderm fragments, bryozoans, brachiopods, gastropods, ostracods, foraminifers (*Bradyina*, *Calcitornella*, *Climacammina*, *Globalvulina*, *Syzrania*), rare bivalves, trilobites and phylloid algae.

At Priest Canyon intercalated limestones contain a diverse fossil assemblage composed of echinoderms, brachiopods, bryozoans, gastropods, fusulinids (*Fusulinella*, *Pseudonovella*: Plate 3-6, 3-7), smaller foraminifers, ostracods, phylloid algae, *Chaetetes* and *Syringopora*.

At Arroyo de la Presilla, Lucas et al. (2009b) reported a few *Lingula* and a flora from the lower part of a 4 m thick interval of gray and black shale, siltstone and fine-grained sandstone in the lower part of the Sandia Formation. The flora includes *Lepidodendron aculeatum*, *Lepidostrobus*, probable *Synchisdendron*, stigmarian roots, strap-like leaves of the lepidodendrids, *Sphenophyllum* and neuropterid foliage indicating a tropical wetland swamp flora.

From coarse-grained fossiliferous limestones at Arroyo de la Presilla, which are developed on top of transgressive cycles, Ivanov et al. (2009) described fish remains of typical marine chondrichthyans (predators and durophagous fishes) including *Stethacanthus* sp., *Bythiacanthus* sp., *Petalodus* cf. *P. acuminatus*, *Peripristis* cf. *P. semicircularis*, a bradyodont and palaeonisciformes.

At Rancho del Chaparral an incomplete amphibian rib was collected from a pebbly sandstone bed approximately 4 m below the top of the Sandia Formation (Rinehart et al. 2011). This is the oldest tetrapod fossil known from New Mexico.

The most useful fossils for age determination of the Sandia Formation are fusulinids. The occurrence of *Eostaffella pinguis* in the lowermost part of the type section indicates a late Morrowan or early Atokan age (Krainer et al., 2011). The presence of *Profusulinella* in the upper part of the type section and in the basal part of the overlying Gray Mesa Formation demonstrates that most of the Sandia Formation and the basal part of the Gray Mesa Formation are of Atokan age (Krainer et al., 2011).

The occurrence of *Eostaffella*, *Fusulinella* and *Millerella* at Soda Dam, and the occurrence of *Profusulinella* at Priest Canyon also point to an Atokan age of the Sandia Formation.

DEPOSITIONAL ENVIRONMENTS

Sediments of the Sandia Formation were deposited in various depositional environments ranging from nonmarine (fluvial) to coastal to shallow marine siliciclastic and to carbonate inner-to-outer shelf settings.

The lectostratotype section of the Sandia Formation is almost entirely marine, which is indicated by its fossils (details in Krainer et al., 2011). In the lower part of the type section, which consists of irregularly alternating shale, sandstone, pebbly sandstone, sandy limestone and limestone, some poorly developed transgressive cycles are recognizable.

Shale is interpreted to have been deposited from suspension on the middle to outer shelf below wave base under low-energy conditions. Intercalated thin, mixed siliciclastic-carbonate sandstone and sandy limestone, frequently containing fragments of marine fossils, indicate deposition in a shallow marine, high-energy nearshore depositional environment with high siliciclastic influx (storm deposits). Crossbedded, partly pebbly sandstone units represent high-energy deposits of the shoreface. The black shale containing fossil plants (unit 23) probably represents deposits of a coastal swamp environment.

Limestone units composed of packstone-rudstone are high-en-

ergy shelf deposits that were deposited above wave base, whereas wackestone-floatstone indicates low-energy depositional environments below wave base. Some transgressive (deepening upward) cycles are developed. The middle part of the Sandia Formation is entirely composed of carbonate sediments with little siliciclastic influx in the lowermost part. Limestone is composed of floatstone to wackestone with intercalated rudstone-packstone, all containing a diverse fossil assemblage indicating deposition in a low- to high-energy shelf environment with normal salinity. The cherty limestone in the upper part indicates deposition in a deeper shelf environment during a relative sea-level highstand. The basal crossbedded pebbly sandstone of the upper part of the Sandia Formation, which contains abundant marine fossils, indicates deposition in a high-energy, nearshore depositional environment. The overlying crossbedded, fining-upward sandstone most probably was also deposited in a high-energy, storm-dominated nearshore environment during transgression. The transgression culminated in the deposition of a thin limestone interval. The next cycle again starts with partly pebbly sandstone, overlain by a covered (shale) interval and a thin limestone on top. Sandstone units of the upper part of the type section are interpreted as high-energy shoreface deposits of a storm-dominated shelf. In contrast to other sections, the Sandia Formation of the lectostratotype section, although dominated by siliciclastic sediments, is almost entirely marine (Krainer et al., 2011).

The irregular stacking pattern of the different lithofacies, particularly in the lower part of the lectostratotype section, indicates that deposition was strongly influenced by syndepositional tectonic movements. Intercalated coarse clastic intervals are probably the result of uplift of the hinterland, causing strong clastic influx. During periods of tectonic quiescence and high sea-level, dark shales and fossiliferous limestones were deposited.

The Sandia Formation of central New Mexico is characterized by rapid lateral facies changes as well as changes in thickness. This is best observed in the San Pedro Mountains-Sierra Nacimiento-Jemez area, where the thickness ranges from zero (Camp Zia, Log Springs, Coyote Flat) to approximately 72 m (Soda Dam). The facies are almost entirely siliciclastic (Resumidero, Porter Landing, Rancho de Chaparral, Osha Canyon) or dominantly siliciclastic with intercalations of marine limestone (Mesa Venado, Rio de las Vacas, Soda Dam).

Siliciclastic facies include fine-grained conglomerate and coarse-grained pebbly sandstone commonly displaying trough crossbedding and channel geometry, different types of sandstone (massive, horizontally laminated, ripple laminated and trough crossbedded) and siltstone-shale intervals which are up to several m thick and locally poorly exposed or covered. Limestone intervals occur as individual thin beds or up to several m thick intervals, commonly intercalated in shale.

At Osha Canyon the Sandia Formation is composed of several fluvial fining-upward cycles, starting with conglomerate at the base (lag deposit), grading into trough-crossbedded sandstone representing channel fill deposits and into siltstone-shale at the top interpreted as overbank fines. The presence of *Lingula* in a shale interval approximately 13 m above the base indicates a short period of brackish environment. The uppermost arenaceous limestone bed, which contains abundant crinoidal debris, is interpreted to be formed during marine flooding.

The entirely siliciclastic succession at Porter Landing is developed in a different facies, characterized by thin sandstone beds with small-scale trough crossbedding, intercalated in fine-grained sandstone to siltstone with abundant ripple lamination and horizontal lamination, probably representing a distal fluvial or fluvio-deltaic system. A similar facies is developed at Rancho de Chaparral.

At Mesa Venado, which is geographically very close to Osha Canyon, the lower part of the Sandia Formation is interpreted as nonmarine regolith, fluvial channels and overbank fines, grading into marine shale with intercalated thin fossiliferous limestone beds.

The Rio de las Vacas section is mostly covered (?shale) with rarely exposed shale. Intercalated are fine-grained arkosic conglomerates

DISCUSSION

and coarse-grained pebbly sandstones which are either massive or display trough-crossbedding and channel geometry. Conglomerate and coarse-grained sandstone are poorly sorted and rounded, and interpreted as sheetflood deposits and fluvial channel fill sediments that developed on alluvial plains. Intercalated limestones that contain a low-diversity fossil assemblage indicate periodic floodings during which shallow, restricted marine conditions prevailed.

At Soda Dam the Sandia Formation is also dominantly siliciclastic with subordinately intercalated limestone intervals. But, in contrast to Rio de las Vacas, conglomerate is missing; sandstone (quartzarenite) is better rounded and sorted and almost completely lacks detrital feldspars and granitic rock fragments. Sandstone approximately 1 m above the basement contains echinoderm fragments, indicating a shallow marine origin. Also, the mixed siliciclastic-carbonate sandstone (grainstone to packstone) 30 m above the base contains a diverse assemblage of strongly fragmented fossils, indicating a shallow-marine, high-energy depositional environment. The crossbedded sandstone interval approximately 50 m above the base is composed of moderately sorted, subrounded quartzarenite, probably deposited in a high-energy nearshore marine environment (similar to the type section). Microfacies types (mudstone, wackestone, packstone, grainstone, rudstone) contain a diverse fossil assemblage, indicating deposition in a shallow, open marine setting of low to high water turbulence.

Near Tecolote at the northern end of the Sandia Mountains, the Sandia Formation is only 24 m thick, rests on Mississippian limestone and is entirely siliciclastic and mostly nonmarine. A thin coarse-grained mixed siliciclastic-carbonate sandstone containing a diverse marine fossil assemblage intercalated about 9 m above the base indicates a marine flooding event. Marine environment is also indicated by brachiopods and crinoid fragments in black shale near the top of the Sandia Formation, approximately 23 m above the base.

At Cedro Peak the entire Sandia Formation is siliciclastic and probably of nonmarine origin. In the southern Manzano Mountains at Priest Canyon, the Sandia Formation overlies Precambrian basement and is 70 m thick. The succession is dominantly siliciclastic and nonmarine in the lower part with intercalated thin marine limestone intervals. The upper part is mostly covered.

At Arroyo de la Presilla in the Cerros de Amado (Socorro County), the Sandia Formation rests on Precambrian granite, is 162 m thick and consists of a cyclic succession of siliciclastic and carbonate, nonmarine and marine lithologies (Lucas and Krainer, 2009; Lucas et al., 2009a). Textural and structural properties of the sandstones and conglomerates at the bases of the cycles, including erosional bases, trough cross-bedding, poor sorting and rounding, and fossil plants, indicate a fluvial origin (multistoried channel fills). The siliciclastic material is derived mostly from granitic, and subordinately from metamorphic source rocks. Poor sorting and rounding of the larger clasts indicate short distances of transport. The small amount of detrital feldspars, which are almost completely diagenetically altered to clay minerals, indicates that most of the feldspars and other unstable grains underwent intense chemical weathering.

The cross-bedded, mixed siliciclastic-carbonate sandstone in the upper part of the section, which contains abundant fossil fragments (unit 57), formed in a shallow marine, high energy nearshore environment. Siltstone to fine-grained sandstone and shale is of nonmarine and deltaic to shallow marine origin.

Crinoidal packstone, locally crossbedded, formed in a higher energy environment such as wave-agitated skeletal banks. The spicular wackestone, characterized by a low diversity benthic fauna, is interpreted to have been deposited in a low-energy environment below the fair weather wave base, probably also below the storm wave base. The Sandia Formation is composed of several mixed siliciclastic-carbonate cycles starting with fluvial channel fill sediments, grading into shale and finally into coarse-grained limestone. Fossiliferous limestone formed in a shallow, high-energy open marine environment.

The dominantly siliciclastic succession of the Sandia Formation was deposited as a result of the onset of the ARM deformation. The ARM formed as a result of collision between the northern part of Gondwana and Laurussia (including North America), which caused internal deformation of the southwestern part of North America and the formation of sedimentary basins and uplifts. The ARM was not a single tectonic event but a series of tectonic pulses that started during the latest Mississippian and continued until the Early Permian.

Intraplate stress created by transpression along the Wichita megashear resulted in the formation of basins and uplifts of the ARM (e.g., Kluth and Coney, 1981; Kluth 1986, 1998; Goetz and Dickerson, 1985; Pindell, 1985; Ross and Ross, 1985; Budnik, 1986; Algeo, 1992; Dickerson, 2003). Another hypothesis explains the ARM by subduction along the southwest margin of Laurussia, which is indicated by a volcanic arc in east-central Mexico. It is assumed that the subducted slab dipped at a very low angle beneath Laurussia, resulting in intraplate stress within the overriding plate associated with deformation of the crust (ARM) (Ye et al., 1996, 1998). According to Dickinson and Lawton (2003), the ARM deformation was induced by intracontinental stress associated with continued subduction of western parts of Laurentia after more eastern parts had locked against Gondwana.

Strong tectonic activity during deposition of the Sandia Formation is well documented in northern New Mexico where the Taos trough started to develop and subside along the active Picuris-Pecos fault and was filled with a thick succession of siliciclastic deposits during the Atokan. From the active San Luis and Uncompahgre uplifts, siliciclastic sediments of the Sandia Formation were deposited to the west and southwest, and Flechado Formation to the east into the Taos trough (Casey, 1980a, b; Soegaard, 1990; Kues and Giles, 2004). Another example in New Mexico is the Perro graben, a sub-basin of the Estancia basin that was filled with up to 1700 m of sediments of Atokan age (Broadhead 1997), indicating strong tectonic activity along the graben-bounding fault and subsidence of the basin. The Peñasco uplift was also active during the Atokan, as is documented by the local absence of the Sandia Formation (e.g., Camp Zia: Abo red beds on Proterozoic basement, Log Springs: Gray Mesa on Osha Canyon and Coyote Flat: Guadalupe Box Formation on basement) (see also Woodward, 1987, 1996).

During the late Morrowan-Atokan, an extensive marine transgression occurred across large parts of central New Mexico, which is referred to as the Absarokan transgression. The high amount of siliciclastic sediment was derived from uplifts composed of Proterozoic basement rocks, such as the San Luis, Peñasco, Pederal, Zuni and the local Joyita uplifts.

The Lower-Middle Pennsylvanian (?late Morrowan-early Atokan) Sandia Formation in the San Pedro Mountains-Sierra Nacimiento-Jemez area, north-central New Mexico, is characterized by distinct lateral changes in thickness and facies as a result of the ARM deformation. In this area, the Sandia Formation is thinner than at the type section in the Sandia Mountains near Albuquerque (124 m). The thickest sections (72 m) are exposed on the eastern side of the Sierra Nacimiento north of Jemez Springs (Soda Dam). Towards the west and north the Sandia Formation thins and is locally entirely siliciclastic (Mesa Venado, Porter Landing, Rancho del Chaparral). North of Guadalupe Box, the Sandia Formation overlies the Osha Canyon Formation, is 32 m thick and composed of several fluvial, fining-upward cycles and a thin marine horizon at the top. The northernmost outcrops are at Resumidero east of San Pedro Peak where the Sandia Formation is poorly exposed, approximately 13 m thick and overlain by Gray Mesa Formation. Locally, near the western and southern margin of the Sierra Nacimiento the Sandia Formation is absent. At Log Springs near the southern end of the Sierra Nacimiento the Log Springs Formation is unconformably overlain by thin Osha Canyon Formation and Gray Mesa Formation, indicating phases of uplift during Late Mississippian and Early Permian time (Peñasco uplift).

At Coyote Flat west of Jemez and at Rio de las Vacas, the Sandia

Formation is absent and the late Desmoinesian-middle Virgilian Guadalupe Box Formation rests directly on Precambrian basement. In the area of Camp Zia northwest of Cuba, the Lower Permian Abo Formation rests directly on basement, indicating that the Peñasco uplift existed there as a positive high during the entire Pennsylvanian. Thus, thickness and facies changes and the distribution of the Sandia Formation in the Sierra Nacimiento can be attributed to ARM tectonic movements of the Peñasco uplift.

Also, from the type section towards the north (Tecolote), the Sandia Formation thins to 24 m, and towards the south (Cedro Peak) to 13.6 m. The Sandia Formation is thicker (70 m) in the southern Manzano Mountains (Priest Canyon). The thickest section in central New Mexico is exposed at Arroyo de la Presilla in the Cerros de Amado east of Socorro, measuring 162 m (Lucas et al., 2009a). At this section, approximately 25% of the strata are conglomerate and coarse-grained sandstone, indicating the presence of a local uplift (Joyita uplift), where the Sandia is locally absent. Farther southeast, at Hansonburg in the southern Oscura Mountains, the Sandia is only 12.5 m thick.

CONCLUSIONS

Fusulinids indicate that sedimentation of the Sandia Formation probably started during the late Morrowan, which means that during that time uplift and erosion of the Pedernal uplift had begun, as already noted by Ye et al. (1996). This event, which caused the first transgression in the adjacent basins, marks the beginning of the ARM deformation in central New Mexico. The onset of uplift and basin subsidence of the ARM varied from basin to basin and started as early as during the latest Mississippian, culminated during the Pennsylvanian and ceased near the end of the Early Permian (Kluth and Coney, 1981; Kluth, 1986; Ye et al., 1996, 1998; Dickinson and Lawton, 2003; Nelson and Lucas, 2011). In central New Mexico (San Mateo, Lucero and Estancia basins), ARM deformation started during the latest Morrowan/Atokan (Kues and Giles, 2004). Farther north, during the Morrowan, the Taos trough started to develop adjacent to the Uncompahgre uplift along the Picuris-Pecos

fault and was filled by a thick succession of siliciclastic sediments of the Sandia Formation (Soegaard, 1990). According to Casey (1980), a wide-spread marine transgression occurred during Morrowan time in north-central New Mexico. Kues and Giles (2004) note that during Atokan time the sea transgressed from central New Mexico through the Cabezon seaway across northwest New Mexico into the Paradox Basin in the Four Corners region. Sediments of the Sandia Formation at the lectostratotype section were deposited on the eastern margin of the northern shelf of the Orogrande Basin, west of the Pedernal uplift. Distinct lateral and vertical changes in facies and thickness indicate that sedimentation was strongly affected by tectonic movements (Krainer et al., 2011).

Isbell et al. (2003) concluded that cyclothems that formed during Glacial I (Late Devonian to earliest Carboniferous) and II (Namurian to earliest Westphalian) of the Gondwana glaciation are not of glacio-eustatic origin. According to Rygel et al. (2008) the latest Mississippian to earliest Pennsylvanian period was characterized by widespread glaciations on the southern hemisphere, and sedimentary successions of this age commonly show evidence of glacio-eustatic sea-level fluctuations of as much as 40-100 m (see also Fielding et al., 2008).

Cycles are present within the dominantly marine successions of the Sandia Formation, particularly at the type section and Arroyo de la Presilla. As these cycles cannot be traced laterally over long distances we assume that the formation of the cycles is caused mainly by tectonic movements of the ARM, although some glacio-eustatic influence cannot be ruled out.

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PLATE CAPTIONS

PLATE 1. Thin-section photographs of sandstone and limestone microfacies of the Sandia Formation. **1**, Well-sorted quartzarenite composed of abundant monocrystalline quartz, a few polycrystalline quartz and rare detrital feldspar grains. Grains are well cemented by quartz in the form of authigenic overgrowths on detrital quartz. Sample SD 17 (Soda Dam), polarized light, width of photograph is 1.2 mm. **2**, Sublitharenite composed of abundant mono- and polycrystalline quartz grains, a few granitic rock fragments and rare detrital feldspars. Some detrital quartz grains display authigenic overgrowths. Sample TE 8 (Tecolote), polarized light, width of photograph is 3.2 mm. **3**, Subarkose composed of detrital quartz and a few feldspar grains cemented by coarse blocky calcite. Sample SD 1 (Soda Dam), polarized light, width of photograph is 3.2 mm. **4**, Quartzarenite composed of detrital quartz grains and rare feldspars (altered to pseudomatrix), cemented by coarse calcite which randomly replaces detrital quartz grains. Sample AP 22 (Arroyo de la Presilla), polarized light, width of photograph is 1.2 mm. **5**, Fine-grained sandstone containing mono- and polycrystalline quartz, rare feldspar grains, micas (muscovite) and matrix (partly pseudomatrix derived from the alteration of unstable lithic fragments). Sample TE 9 (Tecolote), polarized light, width of photograph is 1.2 mm. **6**, Grainstone composed of strongly fragmented fossils including echinoderms, bryozoans, fusulinids, and a few detrital quartz grains, cemented by coarse blocky calcite. Sample SD 12 (Soda Dam), polarized light, width of photograph is 3.2 mm. **7**, Mixed siliciclastic-carbonate sandstone containing many detrital quartz grains, echinoderm fragments, a few bryozoans and other skeletons. Sample AP 15 (Arroyo de la Presilla), plane light, width of photograph is 6.3 mm. **8**, Packstone composed of abundant echinoderm fragments and subordinate bryozoans, fusulinids (*Fusulinella* sp.) and other skeletons. Rarely detrital quartz grains are present. Sample OC 17 (Osha Canyon), plane light, width of photograph is 6.3 mm.

PLATE 2. Thin section photographs of typical microfacies observed in limestone of the Sandia Formation. All photographs under plane light, width of all photographs is 6.3 mm. **1**, Wackestone containing fusulinids, echinoderm fragments, ostracods and many indeterminate and recrystallized skeletons. Sample PC 6 (Priest Canyon). **2**, Fusulinid wackestone-packstone containing abundant, mostly complete tests of fusulinids (*Profusulinella*) and a few other skeletons. Sample SD 14 (Soda Dam). **3**, Packstone containing abundant echinoderm fragments, subordinately brachiopods, bryozoans, rare trilobite fragments and a few detrital quartz grains. Sample PC 10 (Priest Canyon). **4**, Crinoidal packstone composed of abundant crinoid fragments, many detrital quartz grains and rare bryozoans. Sample AP 13 (Arroyo de la Presilla). **5**, Floatstone composed of large skeletons of bryozoans, echinoderms, trilobites and smaller skeletons. Sample AP 8 (Arroyo de la Presilla). **6**, Grainstone-packstone composed mainly of echinoderm and bryozoan fragments and rare smaller foraminifers (*Tetrataxis* sp., top center). The macrofossils are strongly fragmented. Sample SD 15 (Soda Dam). **7**, Rudstone composed of abundant echinoderm and bryozoan fragments that are poorly sorted and strongly fragmented. Sample SD 15 (Soda Dam). **8**, Rudstone containing large skeletons of brachiopods, bryozoans (e.g., *Prismapora* sp., right center), echinoderms, many other smaller skeletons, and detrital quartz grains. Sample TE 8b (Tecolote).

PLATE 3. Thin section photographs of some typical fossils from limestone of the Sandia Formation. **1-4**, *Fusulinella* spp. **1**, sample OC 17 (Osha Canyon), **2**, sample PC 6 (Priest Canyon), **3-4**, sample SD 14 (Soda Dam). **5**, *Bradyina* sp. sample RV 17 (Rio de las Vacas). **6-7**, *Pseudonovella* sp. **1**, sample PC 6 (Priest Canyon). **8**, *Calcivertella* sp. (top, left), *Calcitornella* sp. (middle, right) and *Globivalvulina* sp. sample RV 15 (Rio de las Vacas). **9**, *Komia* sp. sample SD 14 (Soda Dam). **10**, *Tetrataxis* sp., sample DL 34, Doc Long (type section). **11**, *Calcivertella*, sample PC 6 (Priest Canyon). **12**, *Climacammina*, sample PC 6 (Priest Canyon).

PLATE 1

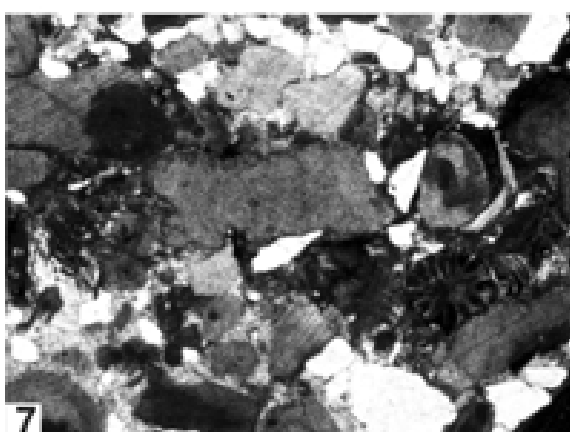
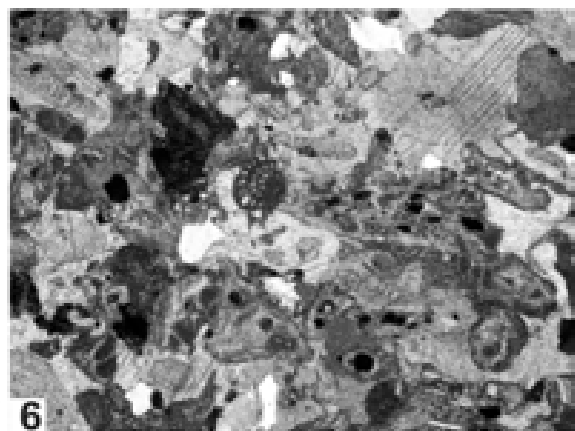
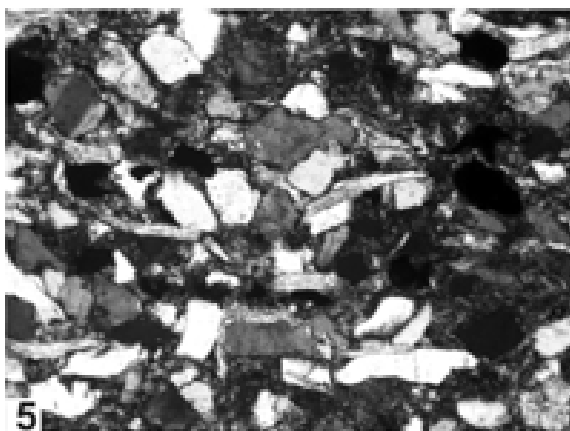
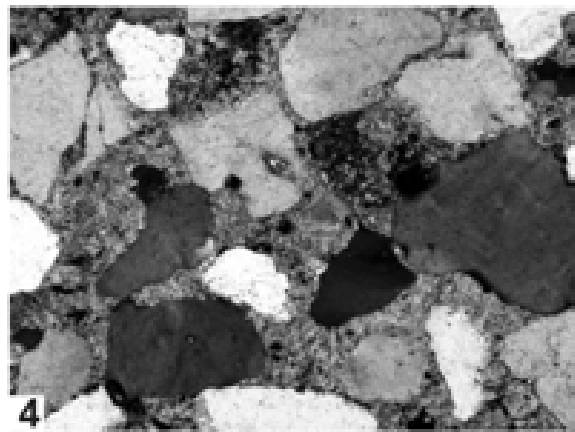
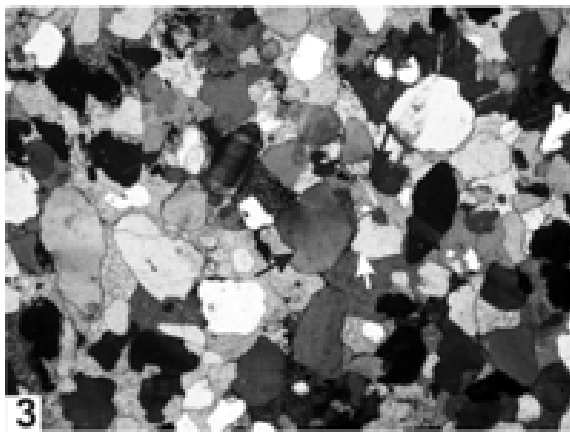
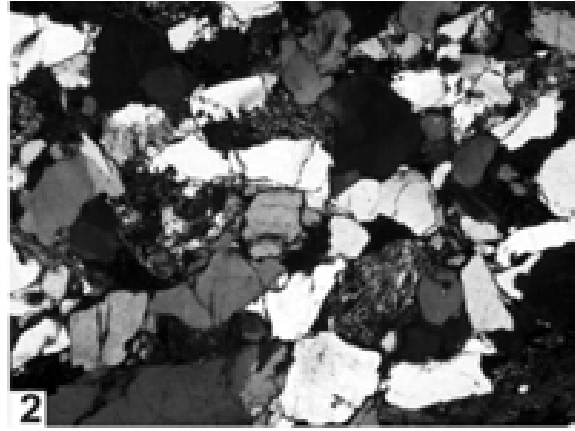
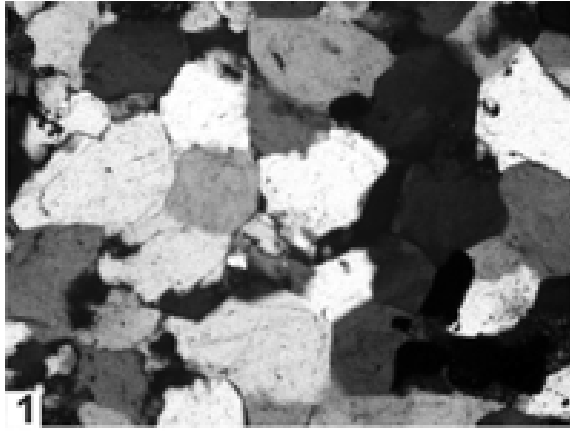


PLATE 2

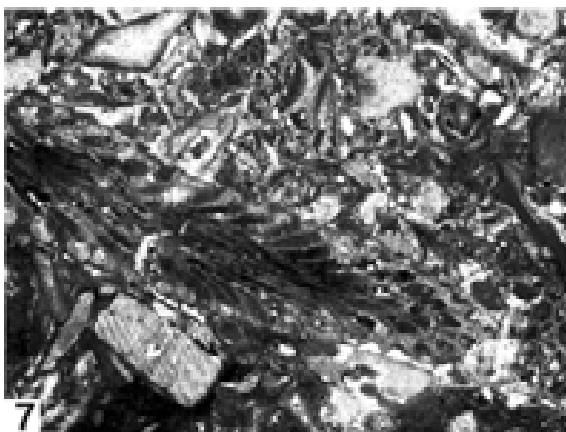
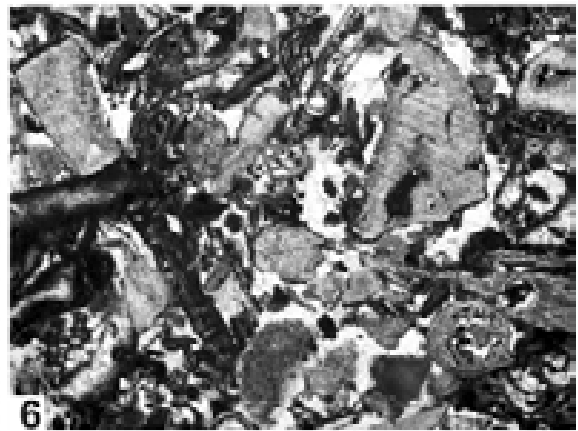
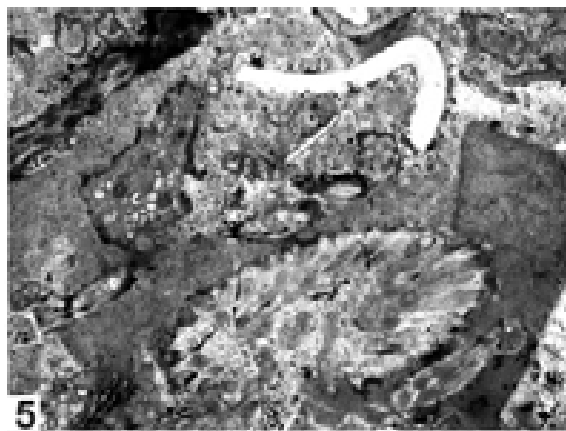
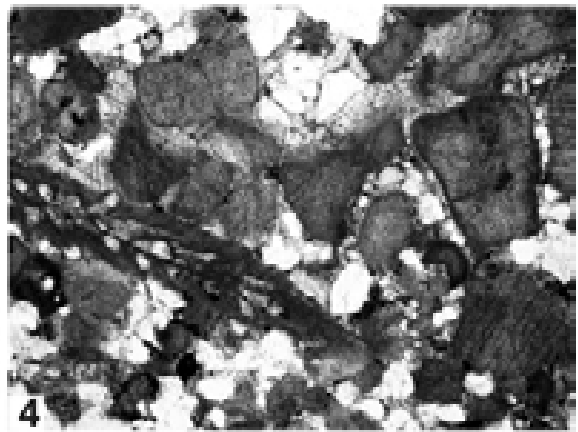
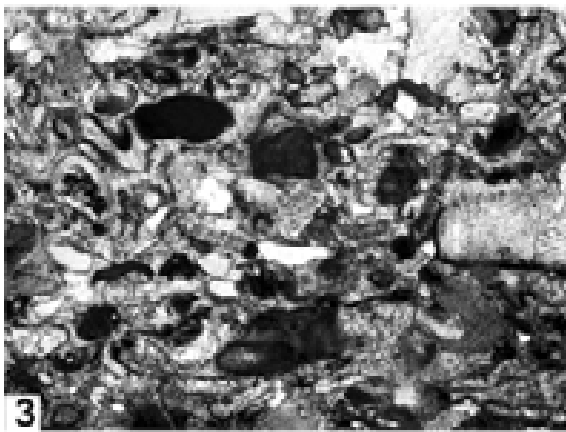
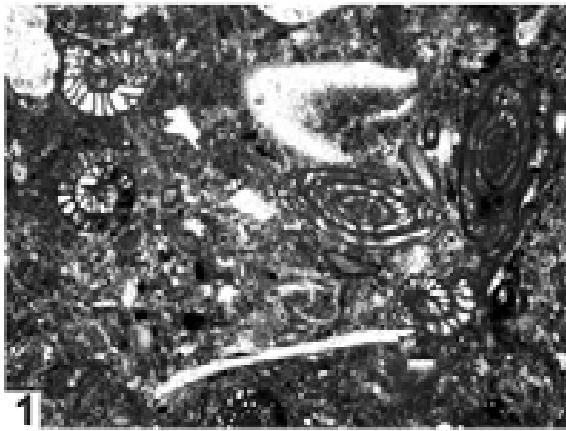
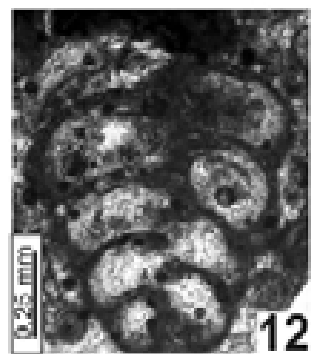
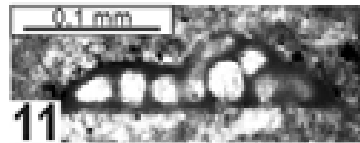
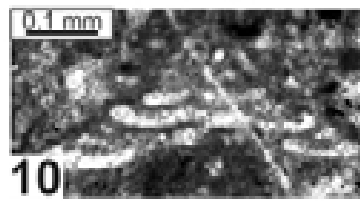
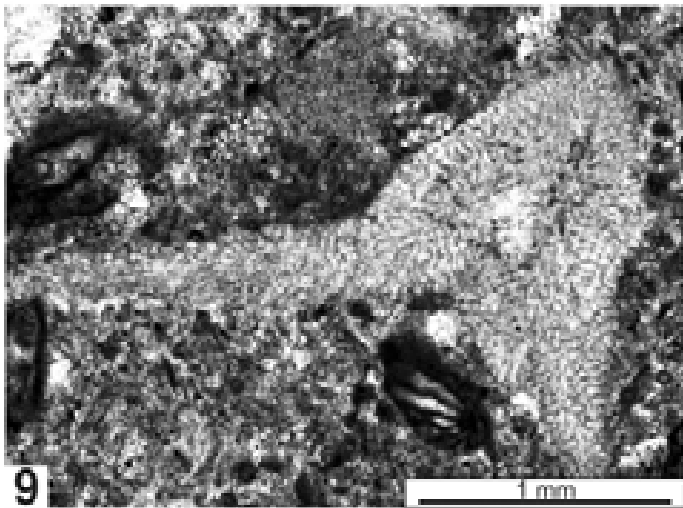
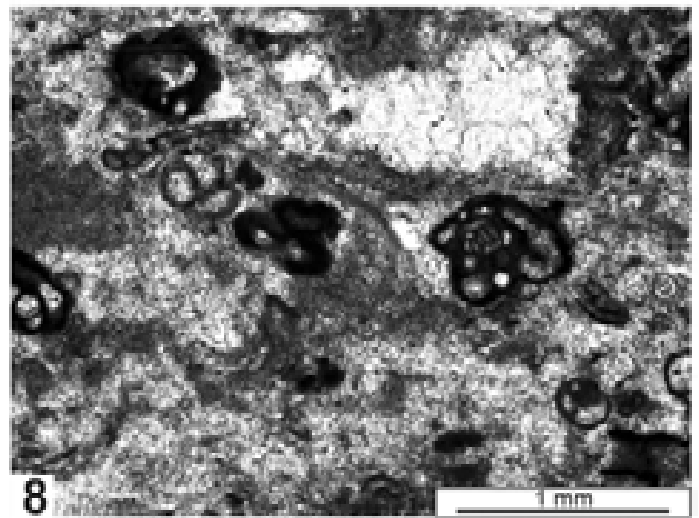
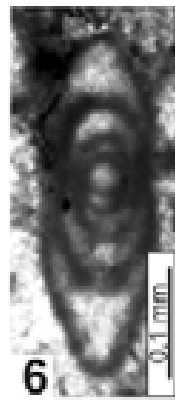
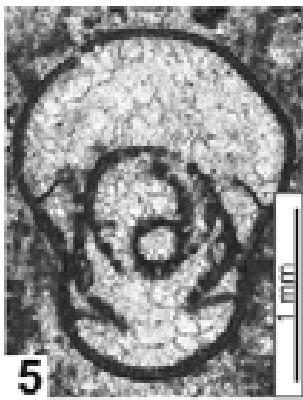
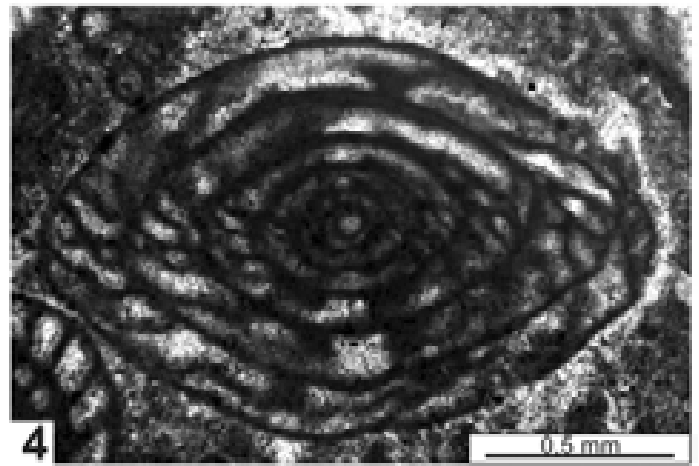
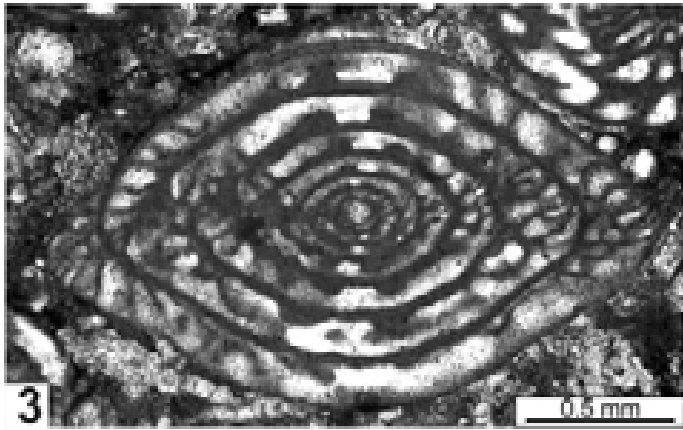
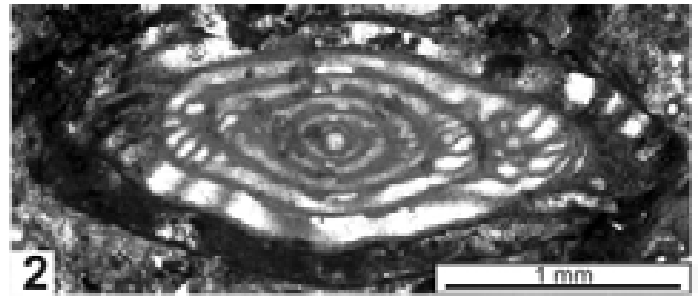
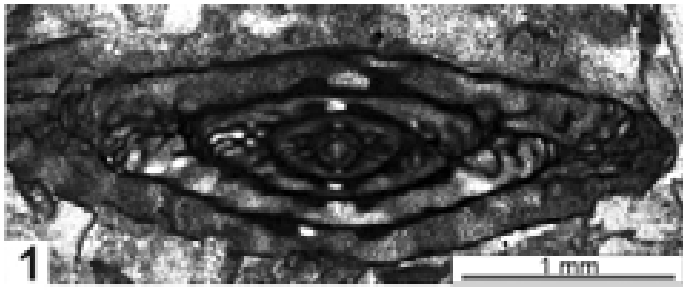


PLATE 3



THE GRAY MESA FORMATION (MIDDLE PENNSYLVANIAN) IN NEW MEXICO

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Abstract—The Gray Mesa Formation is a distinctive interval of ledge- and cliff-forming, cherty limestone in the middle of the Pennsylvanian succession of New Mexico. This unit contrasts with generally slope-forming shale, sandstone, and limestone above and below. Outcrops extend southward from the Nacimiento and southern Sangre de Cristo Mountains, near the latitude of Santa Fe, to the Robledo and northern Organ Mountains near Las Cruces. The Gray Mesa Formation is largely of Desmoinesian age, but both its upper and lower contacts cross time lines and vary from locally unconformable to gradational and arbitrary. Limestone textures are diverse, but lime mudstone and fossiliferous wackestone and packstone are most prevalent. Interbeds of shale, siltstone, and sandstone were sourced from nearby Ancestral Rocky Mountain uplifts. Depositional cycles on the scale of 2 to 10 m record repeated eustatic fluctuations of water depth from below storm-wave base to peritidal, frequently capped by subaerial exposure. Except on the eastern margin of the Uncompahgre uplift and along the southern Pedernal uplift, Ancestral Rocky Mountain uplifts were relatively inactive during Gray Mesa deposition. Reduced tectonic activity meant reduced influx of terrestrial clastics, favoring limestone deposition. Paleoclimate was seasonal wet-dry, fluctuating from subhumid to semi-arid. New Mexico during the late Middle Pennsylvanian was markedly less arid than Colorado and Utah, where thick successions of evaporite rocks accumulated. Wind-blown dust from deserts to the north may be the source of the silica in the prolific chert characteristic of many limestone beds in the Gray Mesa Formation.

INTRODUCTION

Through a large area of New Mexico, Pennsylvanian rocks may be divided into three parts. These are (1) a lower unit of interbedded sandstone, shale, and limestone of Morrowan and Atokan age, (2) a middle unit composed dominantly of ledge- and cliff-forming cherty limestone, mostly of Desmoinesian age, and (3) an upper unit of interbedded shale, mudstone, siltstone, arkosic sandstone, and limestone, mostly of Missourian and Virgilian age. Since the 1940s, geologists have applied a variety of names to the three divisions. In fact, there are almost as many naming schemes as geologists who have studied these rocks. However, as Kues (2001) pointed out, good stratigraphic practice calls for employing the formal names that have the oldest priority of usage. Accordingly, we use the name Gray Mesa Formation for the resistant, cherty Middle Pennsylvanian limestone-dominated lithosome that is the subject of this article.

This article focuses on the area of the field trips in central New Mexico, in Bernalillo, Valencia, and Socorro Counties. However, to provide context we discuss the entire extent of the Gray Mesa Formation, together with its relationship to time-equivalent rocks of different character.

GEOLOGIC SETTING

The Pennsylvanian Period records the assembly of Pangaea through the collision of North America with African and South American tectonic plates. This activity culminated in the Ouachita-Marathon orogeny, to the southeast (present directions) of New Mexico.

During the Mississippian, New Mexico was part of a stable, shallow carbonate shelf on the western North American craton. Maximum transgression took place during the Osagean. Mississippian rocks today are mostly confined to southern New Mexico, in the Delaware, southern Orogrande, and Pedregosa basins. Scattered outliers and small basins exist in the northern half of the state. Original extent probably was much greater; widespread erosion took place during Late Mississippian and Early Pennsylvanian time.

As the Ouachita orogeny got under way, widespread tectonism enveloped the western and central North American craton. This episode is known as the Ancestral Rocky Mountains (ARM) orogeny. Beginning during the Early Pennsylvanian and continuing into Early Permian time, the ARM partitioned New Mexico into a series of uplifts that supplied sediment to adjacent basins (Fig. 1). The Delaware and Pedregosa basins continued to subside. The Orogrande basin extended northward, linking with the Paradox basin in southeastern Utah. ARM faults are mostly high angle, involving Precambrian basement. They have a variety of map orientations, but most of those along the Rio Grande valley trend north. Some are believed to have large elements of strike slip (Nelson and Lucas, 2011).

Two subsequent later tectonic episodes deformed Pennsylvanian rocks, frequently reactivating ARM structures. The Laramide orogeny of Late Cretaceous and early Cenozoic time produced the present Rocky Mountains. Like the ARM, this encompassed primarily high-angle reverse and strike-slip faulting that involved crystalline basement. Beginning during the Oligocene and continuing to the present day, east-west crustal extension has taken place over a large area of the southwestern United States and adjacent Mexico. This action created the Rio Grande rift, running north-south along the length of the state. Widespread intrusive and volcanic activity accompanied rifting.

Plate tectonic reconstruction shows New Mexico lying just north of the equator during Pennsylvanian time, with the continent rotated so that the newly formed Appalachian and Ouachita Mountains lay just south of and ran parallel to the equator (Ron Blakey, Northern Arizona University, <http://cpgeosystems.com>, accessed 9/7/12).

HISTORY OF STUDY

Summary of Geologic Study

Given that New Mexico did not attain statehood until 1912, the history of geologic investigations here is relatively brief. Kues (2012) provides excellent accounts of pioneering geologists in New Mexico and the conditions under which they labored. Most early workers were at-

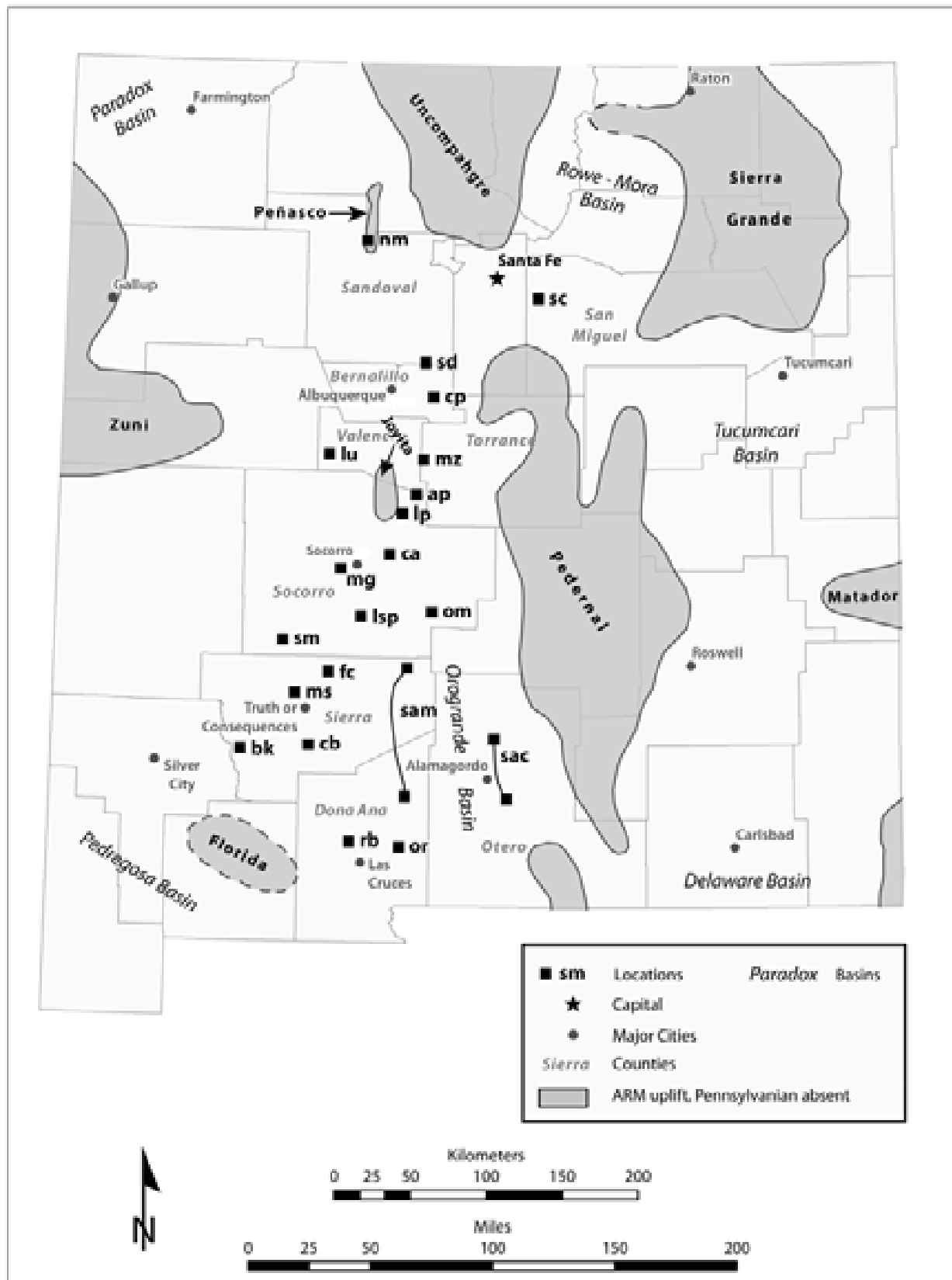


FIGURE 1. Map of New Mexico, showing basins, uplifts, and areas of Pennsylvanian outcrops. **Key to localities:** ap, Abo Pass; bk, Black Range; ca, Cerros de Amado; cb, Caballo Mountains; cp, Cedro Peak; fc, Fra Cristobal Mountains; lp, Los Pinos Mountains; lsp, Little San Pascual Mountains; lu, Lucero uplift; mg, Magdalena Mountains; ms, Mud Springs Mountains; mz, Manzano Mountains; nm, Sierra Nacimiento; om, Oscura Mountains; or, Organ Mountains; rb, Robledo Mountains; sam, San Andres Mountains; sac, Sacramento Mountains; sc, Sangre de Cristo Mountains; sd, Sandia Mountains; sm, San Mateo Mountains.

tached to the U.S. Geological Survey (USGS) or to the University of New Mexico and the New Mexico School of Mines (now New Mexico Institute of Mining and Technology), both established in 1889. Among the most notable were Nelson H. Darton, Clarence L. Herrick, Charles Rollin Keyes, and Willis T. Lee. These men and many others carried out both regional reconnaissance mapping and detailed studies of mines, mineral deposits, and fossil occurrences. Early regional studies established the presence of Pennsylvanian strata and a bare outline of their internal stratigraphy. Gordon (1907) first established a three-part division, comprising Sandia Formation (oldest), Madera Limestone, and Manzano Group (originally the Manzano “series” of Herrick, 1900) (Fig. 2). However, much of Gordon’s Manzano Group is now known to be Permian.

Motivated by the wartime search for oil and gas, the USGS commissioned extensive medium-scale geologic mapping in New Mexico during the 1940s. These geologists divided the Madera Limestone into members (Fig. 2). Four of the mapping teams--Northrop et al. (1946), Wilpolt et al. (1946), Wood and Northrop (1946), and Wilpolt and Wanek (1951)--informally broke out a “lower gray limestone member” and an “upper arkosic limestone member.” Read and Wood (1947) endorsed the USGS usage in their summary of Pennsylvanian stratigraphy of northern New Mexico. However, working on the Lucero uplift (Fig. 2), Kelley and Wood (1946) formally named three members: Gray Mesa (oldest), Atrasado, and Red Tanks.

After World War II, the New Mexico Bureau of Geology and Mineral Resources in Socorro published a large number of detailed geologic maps and reports that cover most of the Pennsylvanian outcrop areas in the state. Particularly important are the synthetic works of Kottlowski (1960, 1963). These have been supplemented by continued efforts by the USGS, together with faculty and graduate students from universities in New Mexico and elsewhere. Since 1950, the New Mexico Geological Society holds annual field conferences; their guidebooks contain a wealth of information on the Pennsylvanian strata of New Mexico.

The present authors re-examined the Gray Mesa type section (Krainer and Lucas, 2004) and described the formation in the Oscura Mountains (Lucas and Krainer, 2009), the hills east of Socorro (Lucas et al., 2009), and in the Manzanita Mountains (Lucas et al., 2011). Additional stratigraphic and paleontologic investigations of the Gray Mesa and correlative units, some as yet unpublished, have been carried out by us in the Sandia, Manzano, Fra Cristobal, Mud Springs, Caballo, and Robledo Mountains, and in several other areas of New Mexico.

Impediments to Geologic Study

There are many limitations to understanding Pennsylvanian geology in New Mexico. Outcrops lie mostly along the margins of the Rio Grande rift, which has a basin-and-range topography (Fig. 1). Separating the ranges are broad alluvial valleys, where Paleozoic rocks lie hundreds or thousands of meters below the surface. Several mountain groups are on military bases or wildlife reserves, where access for scientific study is seldom granted. Others are in remote, roadless areas where physical access is difficult. The Gray Mesa Formation commonly forms cliffs that are inaccessible by normal means. Despite magnificent exposures, Gray Mesa sections are incomplete except in a few deep canyons, highway cuts, and railroad cuts. Non-resistant lithologies such as shale, mudstone, and nodular limestone generally are covered.

Economic incentive to study Pennsylvanian rocks in New Mexico is lacking. The Pennsylvanian is far better known (for example) in Illinois, which has few outcrops but a century and a half of coal mining combined with tens of thousands of oil test holes. Deep drilling in New Mexico is largely confined to the Delaware basin in the southeast and the San Juan Basin in the northwest. For the most part, records from the widely scattered wildcat holes in the Rio Grande rift have not been systematically studied. Many of these holes were drilled before the advent of modern logging techniques.


Virgilian	Gordon 1907 Rio Grande Valley	Manzano Group	Madera	Limestone	Sandia Formation
Misouevian	Spencer & Paige (1933) Silver City area	Syrena Formation	Ornaldo Formation		
Desmoinesian	Thompson 1943 Central NM	Red Tanks Member Atrasado Member Gray Mesa Member	Madera (Limestone)	Derry Series	
	Kelley & Wood (1946) Lucero Uplift	Red Tanks Member Atrasado Member Gray Mesa Member	Madera (Limestone)	Sandia Formation	
	Northrop et al. (1946), Wilpolt et al. (1946), Wood & Northrop (1946), Wilpolt & Wanek (1951)	upper arkosic limestone member lower gray limestone member	Madera (Limestone)	Sandia Formation	
	Kelley & Silver (1952) Caballo Mts.	Bar B Formation	Nakaye Formation	Red House Formation	
	Wengert & Matheny (1958) Paradox Basin	Honaker Trail Formation	Paradox Formation Pinkerton Trail Formation Molas Formation		
	Stiles (1959), Pray (1961)	Holder Formation Beeman Formation		Cobbler Formation	
	Zeller (1966) Pedregosa Basin	Horquilla Formation			
	Bachman & Harbor (1976) Southern San Andres Mts.	Parlier Sep Formation	Lead Camp Formation	Sandia Formation	
	Harbour (1972) Franklin Mts.	Upper Member Bishop Cap Member Bingo Member	Magdalena Formation	La Tuna Member	
	Myers (1973) Manzano Mts.	La Caca Member Pee Dee Member Vol Le Mire Member	Wild Cow Formation	Los Moyos Limestone	Sandia Formation
	Bull & Myers (1999) Southern Sangre de Cristo	Alambres Formation	Madera Group	Powell Formation	Sandia Formation
	This Report Rio Grande Rift	Burnian Formation	Atrasado Formation (North) & Bar B Formation (South)	Gray Mesa Formation	Sandia Formation (North) & Red House Formation (South)

FIGURE 2. Chart showing development of Pennsylvanian stratigraphic nomenclature in New Mexico. Abbreviations as in Figure 1.

STRATIGRAPHIC NOMENCLATURE

Prior to the 1940s, geologists generally considered New Mexico's Pennsylvanian rocks in reconnaissance fashion. At most, they recognized a lower Sandia Formation and an upper Madera Formation or Limestone, together comprising the Magdalena Group. For further discussion of early classification, see Kues (2001, 2012).

Thompson (1942) attempted the first detailed subdivision of these rocks (Fig. 2). He named 15 formations and 8 groups, which were correlated to Midcontinent series using fusulinids. Most subsequent workers, however, were unable to map Thompson's formations and rejected his classification. A few authors applied Thompson's formations. For example, Rejas (1965) used them in large-scale thesis mapping in the hills east of Socorro. As Kues (2001) recommended, some of Thompson's formations are distinctive and useful as member-level units (Lucas and Krainer, 2009; Lucas et al., 2009).

In their investigation of the Lucero uplift, Kelley and Wood (1946) divided the Madera Formation into Gray Mesa (oldest), Atrasado, and Red Tanks members (Fig. 2). The name Gray Mesa refers to a landform that is labeled Mesa Aparejo on current maps. Although Kelley and Wood published graphic columns, they did not describe a type section.

The name Gray Mesa found little favor for several decades. Most geologists continued to use informal names such as "lower gray limestone member of the Madera Formation." Meanwhile, the formal names Nakaye Formation (Kelley and Silver, 1952), Lead Camp Limestone (Bachman and Myers, 1969) and Los Moyos Formation (Myers, 1973) were introduced for essentially the same rocks. The multiplicity of names impeded understanding (Fig. 2).

Reviewing Pennsylvanian nomenclature, Kues (2001) recommended that the Gray Mesa be elevated from a member to a formation in the Madera Group. He further argued that Gray Mesa, being the oldest formal name, takes precedence over some of the names mentioned above. Acting on these suggestions, Krainer and Lucas (2004) described a lectostratotype (principal reference section) for the Gray Mesa Formation at Mesa Aparejo (Fig. 3). Lucas and Krainer (2009), Lucas et al. (2009) and Lucas et al. (2011) extended the Gray Mesa Formation into the Oscura Mountains, the hills east of Socorro, and the Manzano and Manzanita Mountains, respectively.

The Gray Mesa Formation is further extended geographically in this report, using the following definition.

The Gray Mesa Formation is the unit of fossiliferous, cherty, ledge-and cliff-forming limestone in the middle part of the Pennsylvanian System across a large area of central New Mexico. Siliciclastic rocks are absent or a minor, inconspicuous component. The Gray Mesa overlies the Sandia Formation (north) and Red House Formation (south), and it underlies the Atrasado Formation (north) and the Bar B Formation (south). Age of the Gray Mesa is largely, but not exclusively Desmoinesian.

LITHOSTRATIGRAPHY

Extent and Distribution

The Gray Mesa Formation can be identified in most of the mountain ranges of central New Mexico from the Nacimiento Mountains on the north to the Caballo and San Andres Mountains on the south (Figs. 1-2). The eastern limit extends from the southern Sangre de Cristo Range south through the Sandia, Manzano, Oscura, San Andres, and northern Organ Mountains. The western limit, vaguely defined, swings through the San Juan Basin and passes west of the Lucero uplift, Magdalena Mountains, and Mud Springs Mountains.

North and Northeast

The northern limit of the Gray Mesa Formation is in the southern Sangre de Cristo Range, where the name Porvenir Formation is applied to

Desmoinesian strata (Fig. 2). Baltz and Myers (1999) mapped three facies of the Porvenir. The southern facies is mostly limestone, and could be called Gray Mesa, although the proportion of sandstone is higher here than it is farther south. Northward, the limestone grades into a sandstone-shale-limestone facies that contains increasingly coarse clastics northward. Toward the northwest, the limestone largely gives way to dark gray shale. These changes reflect proximity to the tectonically active eastern Uncompahgre uplift and to the Taos trough, which was sinking rapidly during Middle Pennsylvanian time.

East of the Sandia Mountains, the Gray Mesa wedges out against the flanks of the Sierra Grande and Pedernal uplifts, both active elements of the ARM orogeny (Fig. 1). Farther east in the subsurface, the Tucumcari basin, another tectonically active feature, contains a thick and varied Pennsylvanian succession in which the name Gray Mesa is not applicable (Broadhead and King, 1988; Broadhead, 2001).

East

The Gray Mesa Formation is well developed in the mountains and hills east of the Rio Grande between the latitudes of Albuquerque and Socorro. From north to south, these uplands include the Manzanita, Manzano, and Los Pinos Mountains, the Joyita Hills, and mostly unnamed hills around Cerros de Amado, east of Socorro. In the northern three mountain ranges, the strata that Myers (1973) and subsequent authors called Los Moyos Limestone (Fig. 2) are readily assigned to the Gray Mesa. A representative section, measured near the southern end of the Manzano Mountains on a branch of Priest Canyon, is 190 m thick (Fig. 4). The Gray Mesa at Priest Canyon can be further divided into members, as discussed in a later section of this article.

The Pennsylvanian succession in the Joyita Hills is anomalously thin because the hills coincide with a small but actively rising ARM uplift. The entire Middle Pennsylvanian limestone is less than 30 m thick, and is truncated with angular unconformity at the base of the Upper Pennsylvanian Bursum Formation at the southern end of the hills (Kottlowski and Stewart, 1970).

Although considerably faulted, the hills east of Socorro present excellent exposures of the Gray Mesa, and a composite section has been assembled (Lucas et al., 2009). As at Priest Canyon and several other sites, the formation is divisible into members here. Algal bioherms are locally developed near the top of the Gray Mesa (Fig. 5).

Southeast and South

Typical Gray Mesa Formation is developed in the northern Oscura Mountains (Lucas and Krainer, 2009). Resistant, cherty limestone of latest Atokan and Desmoinesian age continues into the southern Oscuras and northern San Andres Mountains. These rocks overlie Sandia Formation and underlie a shaly Upper Pennsylvanian succession (Bachman, 1968; Bachman and Harbour, 1970).

Although Kottlowski et al. (1956) did not divide the Pennsylvanian into formations in the San Andres Mountains, their graphic columns show cliff-forming cherty limestone, largely of Desmoinesian age. Below is a slope-forming succession of mixed lithology that fits either Sandia or Red House Formation. Above is the Panther Seep Formation, a thick succession of dark shale, siltstone, fine sandstone, micritic shaly limestone, and gypsum deposited along the keel of the rapidly subsiding, narrow, north-trending Orogrande basin. Thus, the Desmoinesian limestone can readily be called Gray Mesa. Thickness is 150 to 200 m thick in most of the range, thinning to less than 100 m near the southern end.

Bachman and Myers (1969) introduced the name Lead Camp Limestone for "mostly massive cliff-forming limestone" in the southern San Andres Mountains (Fig. 2). Fusulinids indicate Morrowan to early Missourian age. When the Lead Camp type section is drafted (Fig. 6), a subdivision becomes evident. The lower 80 m includes more than 60% covered intervals (shale?), with interbeds of mostly crinoidal or bioclastic limestone having little chert, along with sandstone and pebbly sandstone mostly in the lower part. This lithology matches the Red House Forma-

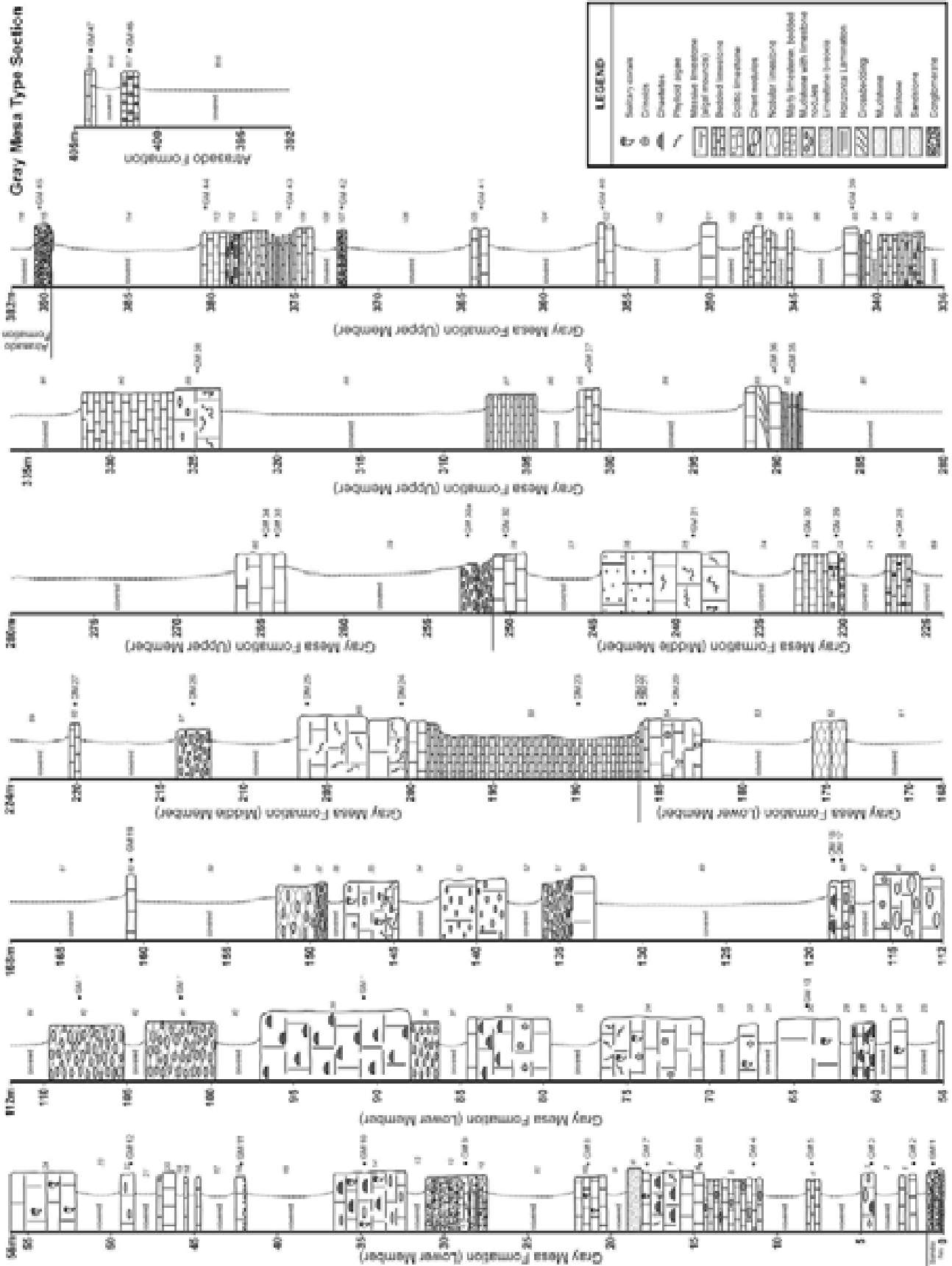


FIGURE 3. Graphic column of type section of Gray Mesa Formation (from Krainer and Lucas, 2004).

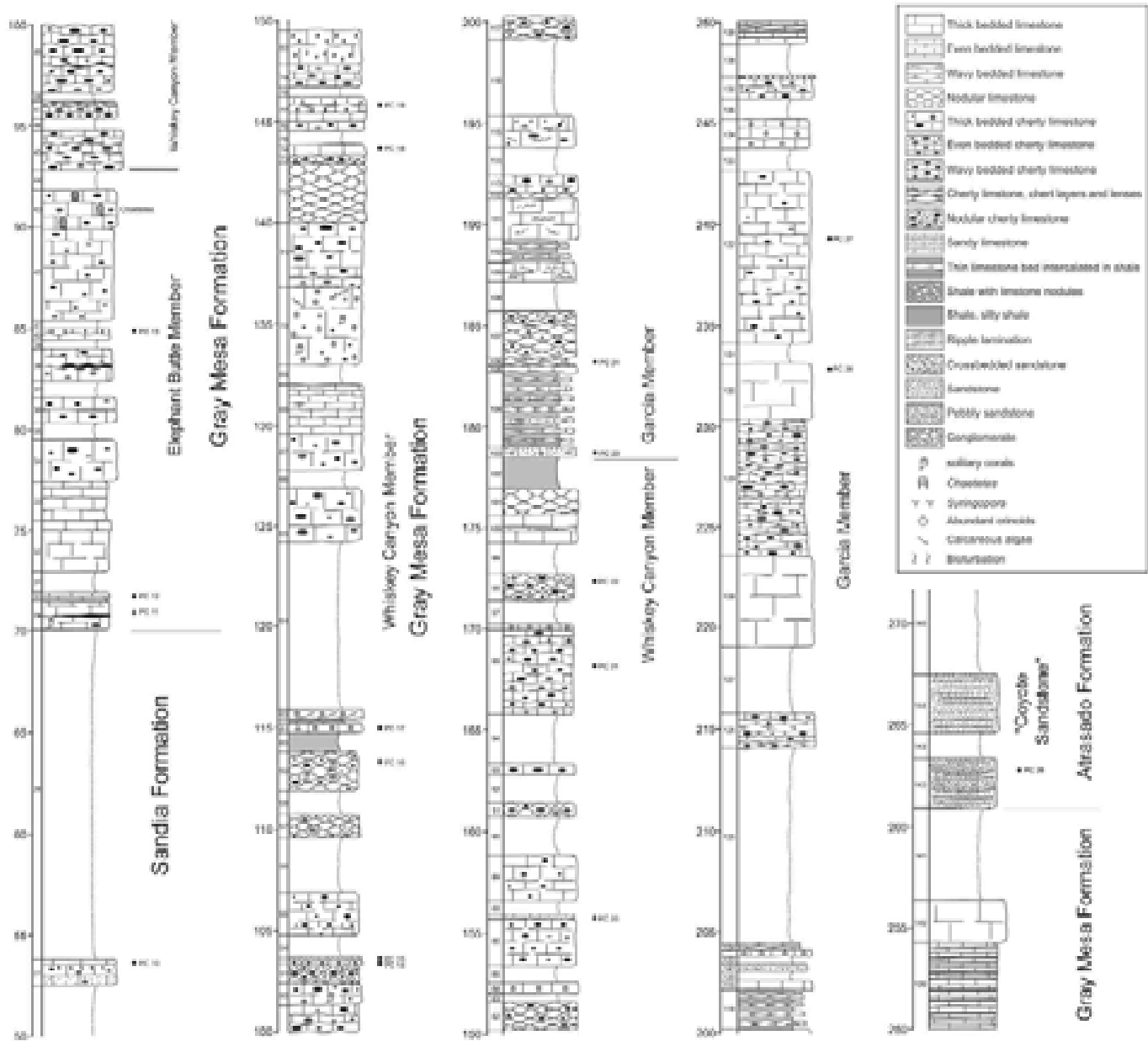


FIGURE 4. Graphic column of Gray Mesa Formation from section measured in a branch of Priest Canyon in the southern Manzano Mountains. Scale is in meters.

tion in mountain ranges to the west. The remainder of the Lead Camp, 180 m of ledge- and cliff-forming, cherty limestone, is typical Gray Mesa Formation.

Sections by Seager (1981) indicate that Red House and Gray Mesa can be split out of the Lead Camp Limestone in the northern Organ Mountains (Fig. 7). A facies change, however, takes place in the southern Organ Mountains. The Morrowan (?) and Atokan changes to cliff-forming, cherty limestone (La Tuna Formation), while the Desmoinesian takes on more shale and non-resistant, marly limestone (Berino Formation). The La Tuna and Berino continue southward into the Franklin Mountains near El Paso, Texas (Nelson, 1940; Harbour, 1972). The southern limit of the Gray Mesa Formation thus may be placed in the central Organ Mountains.

In the Sacramento Mountains, the Morrowan and Atokan comprise 60 to 150 m of interbedded gray to black shale, quartzose sandstone, and dark-colored, impure cherty limestone. Overlying Desmoinesian and lower Missourian rocks undergo dramatic facies changes. In some

places these rocks consist of cliff-forming, biohermal limestone (Bug Scuffle) that is largely micritic, sparsely cherty, and low in clay content. Within a few km the limestone intergrades bed-for-bed with shale and sandstone (Pray, 1961; Benne, 1975). This area lay along the active hinge line between the Pedernal uplift and the Orogrande basin (Algeo, 1996; Algeo et al., 1991). The name Gray Mesa has not been used here, and we do not propose to introduce it.

Southwest

Kelley and Silver (1952) introduced the name Nakaye Formation for cliff-forming, cherty Middle Pennsylvanian limestone in the Caballo Mountains (Figs. 2, 8). The Nakaye contrasts with the slope-forming, shaly Red House Formation below and the Bar B Formation above. The same nomenclature has been extended into the Fra Cristobal Mountains (Cserna, 1956; McCleary, 1960) and the Mud Springs Mountains (Maxwell and Oakman, 1990). However, the Nakaye is identical to the Gray Mesa in all key aspects, and the name Gray Mesa should be used be-



FIGURE 5. Photograph near Cerros de Amado, east of Socorro, showing shaly upper portion of Gray Mesa (Garcia Member) capped by a massive algal bioherm.

cause it has priority (Lucas et al., 2012).

The southwesternmost place where the Gray Mesa may be recognizable is on the north face of Robledo Peak, north of Las Cruces. This site affords a nearly complete Pennsylvanian section, which has been zoned using fusulinids but not divided into formations (Kottlowski 1960, Seager et al. 2008). A section measured by the authors is summarized in Table 1.

Although the Gray Mesa interval is recognizable here, formally extending this unit into the Robledo Mountains seems inadvisable. The overall succession is more closely allied to the Horquilla Formation, as exposed in far southwestern New Mexico and southeastern Arizona (Kottlowski 1960, p. 67; Zeller, 1966; Lucas et al., 2012).

Pennsylvanian rocks of the Santa Rita-Silver City area are assigned to Oswald (older) and Syrena formations (Spencer and Paige, 1935). Both consist of massive to bedded and nodular limestone having interbeds of shale and local bodies of sandstone. Fusulinids indicate the Oswald ranges from Morrowan to early Missourian, whereas the Syrena is Missourian and Virgilian (Jones et al., 1967; LeMone et al., 1974). The Gray Mesa Formation cannot be identified here. The overall succession resembles that on Robledo Mountain, with perhaps a larger proportion of shale.

Pennsylvanian rocks are absent or greatly thinned on the Florida uplift, yet the extent of this feature is poorly constrained. Permian rocks directly overlie Mississippian in the Florida Mountains, southeast of

Deming. However, Kottlowski (1960) and Corbitt and Woodward (1970) suggest the Pennsylvanian has been faulted out of the section. Without fault complications the Pennsylvanian thins from about 55 to 12 m southward at Cookes Peak (north of Deming). Fusulinids near the base of the section are lower Desmoinesian, and they are Missourian near the top. Kottlowski (1960, p. 65) suggested that the Florida uplift of the ARM was either a “low landmass” or “scattered islands amid shallow seas.” We favor the latter interpretation.

West

Widely scattered Pennsylvanian outcrops occur in west-central New Mexico. Most sections are incomplete and complicated by faulting, igneous intrusion, and contact metamorphism. For example, Kuellmer (1954) described a Pennsylvanian section at Kingston, about midway between Silver City and the Caballo Mountains. At the base is a 15 m interval of interbedded shale, shaly limestone, and sandstone that is similar to Red House lithologies. This is overlain by 17 m of massive, cherty limestone that yields Desmoinesian fusulinids. Younger strata, about 110 m thick, include various types of limestone together with siltstone and shale, from which age-specific fossils were not reported.

Sierra Cuchillo is a small mountain range northwest of Truth or Consequences. The Pennsylvanian here includes about 45 m of Sandia Formation overlain by “Madera Limestone,” about 380 m thick and

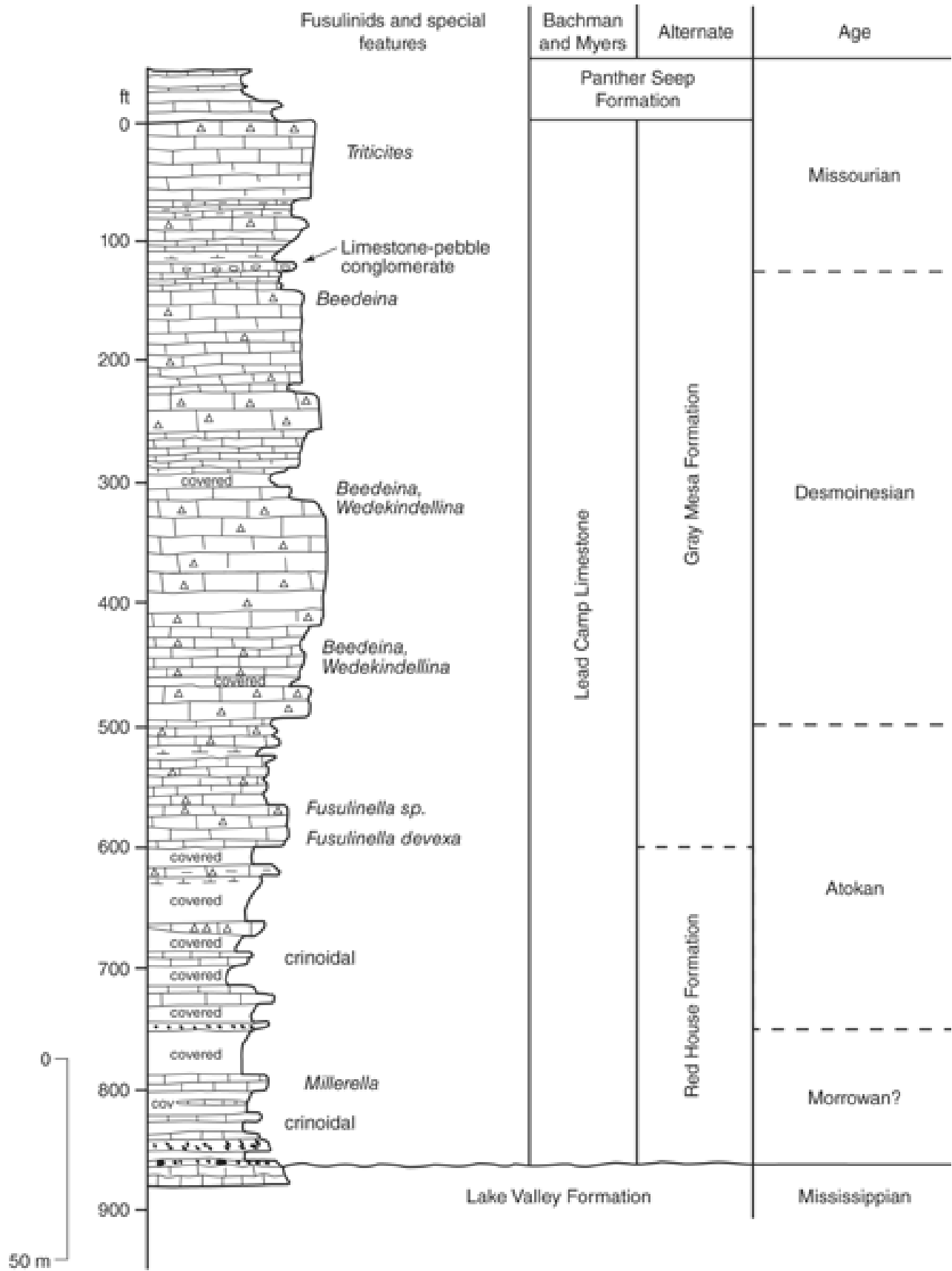


FIGURE 6. Type section of the Lead Camp Limestone, southern San Andres Mountains, based on written description by Bachman and Myers (1968). The Lead Camp appears readily divisible into lower slope-forming interval of mixed lithology (Red House) and upper ledge-and-cliff-making cherty limestone (Gray Mesa).

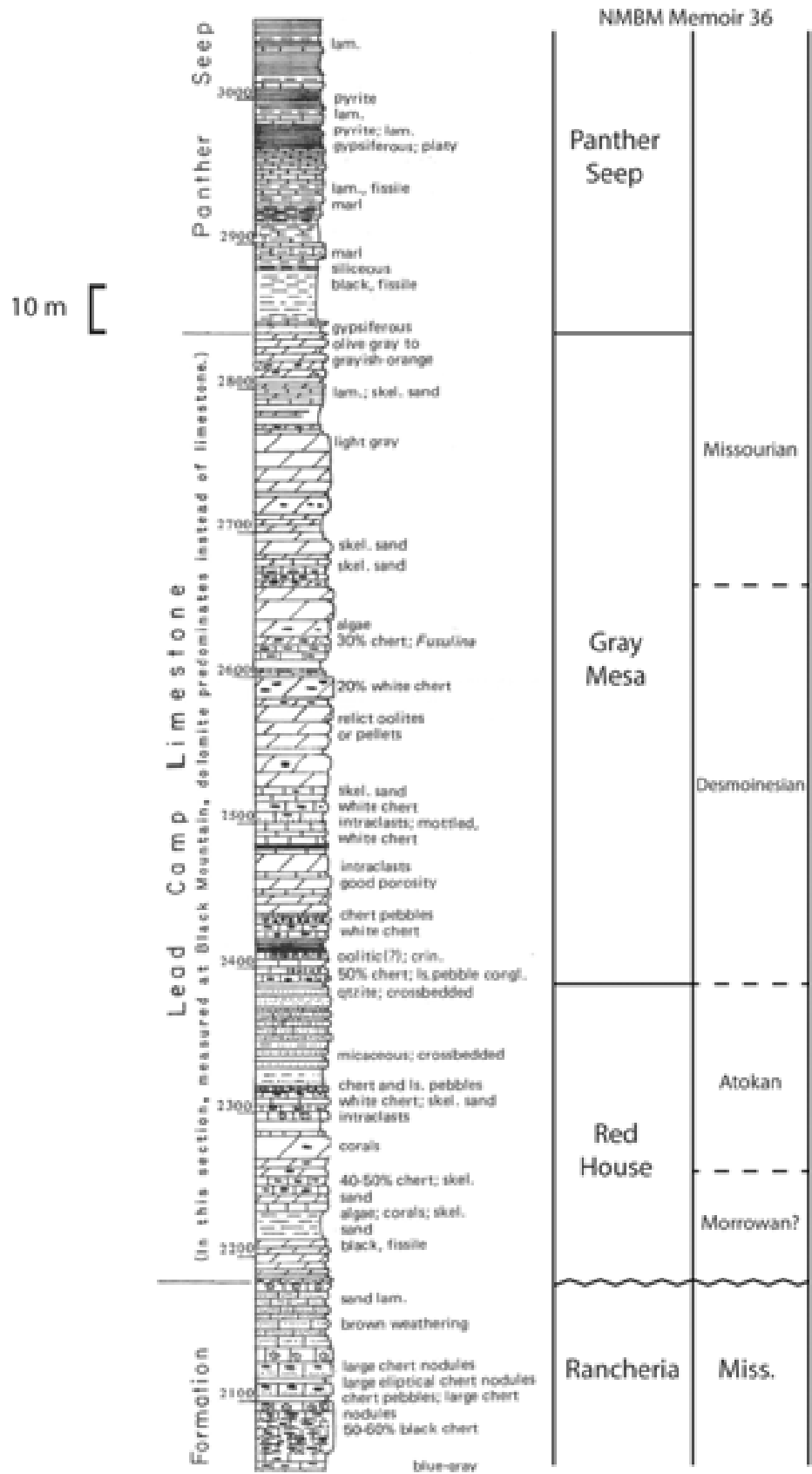


FIGURE 7. Pennsylvanian columnar section from northern Organ Mountains, by Seager (1981). Again, the name Lead Camp appears redundant, as Red House and Gray Mesa formations are readily recognizable. Original scale is in feet; metric bar scale added.

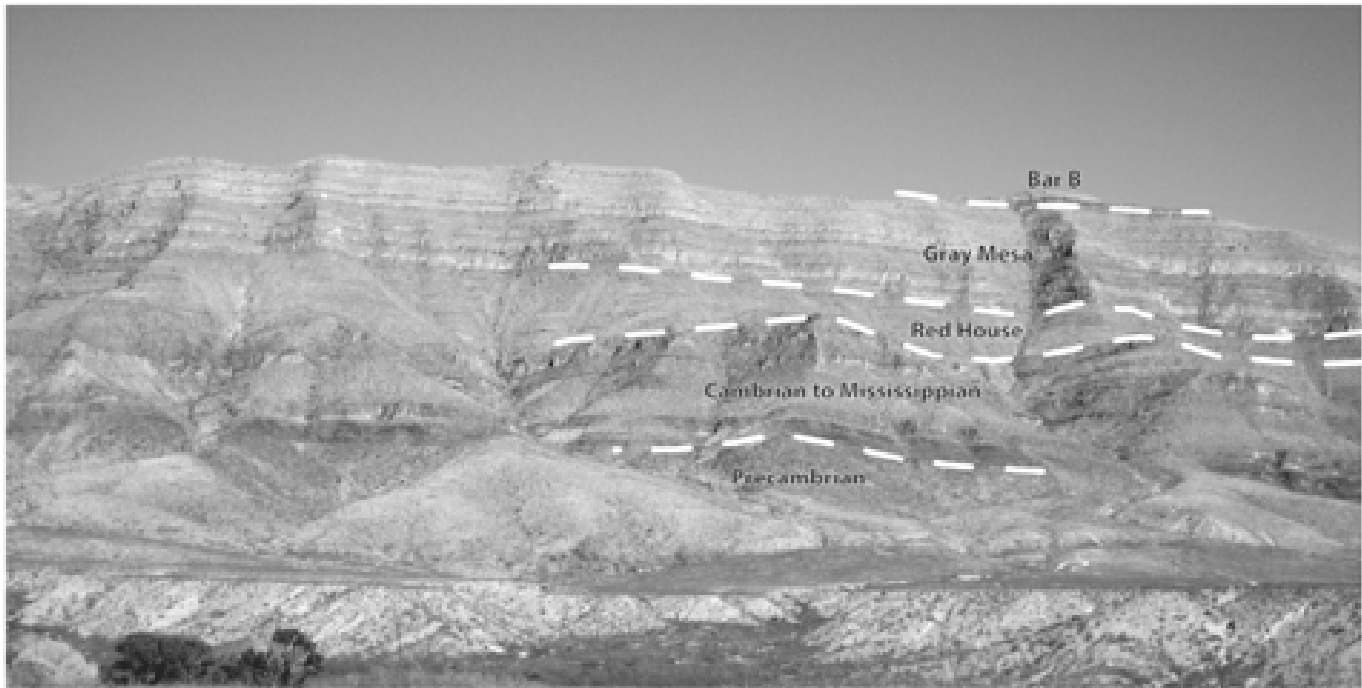


FIGURE 8. Photograph of western escarpment of Caballo Mountains in the area of South Ridge, where Kelley and Silver (1952) described type section of Nakaye (now Gray Mesa) Formation.

largely cherty limestone; with lenses of limestone conglomerate and a few sandstone interbeds near the top (Jahns, 1955). Without biostratigraphic control or a detailed section, defining the Gray Mesa here is difficult.

More intriguing is the section in the San Mateo Mountains (southwestern Socorro County) reported by Kottlowski (1960). After carefully mapping and measuring the faulted section and gathering fusulinid control, Kottlowski concluded that the Sandia Formation is at least 76 m thick and the Desmoinesian portion, largely limestone, may be as thick as 550 m. Missourian rocks total about 170 m thick with the top eroded, and consist largely of limestone, similar to the Desmoinesian. Kottlowski suggested that these rocks occupied a small, deep basin separate from the Orogrande and Lucero basins.

Volcanic rocks dominate the high plateaus of Catron County. A few outcrops and deep wells reveal Pennsylvanian rocks, including limestone having Desmoinesian fusulinids. These rocks pinch out northward as they lap onto the Zuni uplift, an ARM feature where Pennsylvanian rocks are virtually absent. The ancestral Zuni uplift occupies most of Cibola County, southwestern McKinley, and northwestern Catron Counties (Kottlowski, 1959, 1960).

Northwest

The outcrops of Gray Mesa farthest to the northwest are in the Nacimiento Mountains. Exposed here is part of the Peñasco uplift, an active ARM element where Pennsylvanian rocks were eroded or never deposited. On the flanks of the uplift the Gray Mesa pinches out, as do the underlying Sandia Formation and overlying Atrasado Formation (Wood and Northrop, 1946; Woodward, 1987, 1996; Krainer et al., 2005).

Extent of the Gray Mesa Formation in the subsurface of the San Juan Basin is largely a matter of definition. A broad carbonate shelf or platform separates Nacimiento Mountain outcrops from the Paradox basin. Within the latter, the Paradox Formation of cyclically interbedded carbonate and evaporite rocks is the time equivalent of the Gray Mesa. Evaporites barely reach the northwest corner of New Mexico. As Huffman and Condon (1993, p. 14) wrote, "Southeast of the evaporite facies of the Paradox, equivalent rocks are composed of mixed carbonate beds and

shale marker beds. The Paradox sequence of cyclically bedded deposits has a distinctive geophysical-log response that was traced as far as possible to the eastern and southern parts of the [San Juan] basin. The term Madera Limestone is used where the cyclic beds could no longer be recognized."

Thickness

Thickness patterns of the Gray Mesa involve much uncertainty. Control points are widely scattered and difficult to verify (Fig. 9). Many are imprecisely located. Measurements of the same section by different geologists (including ourselves) may differ by 20%. Faults and other structural complications are pervasive in many mountain ranges.

Pennsylvanian rocks are thin, patchy and/or absent on ARM uplifts, including the Uncompahgre, Sierra Grande, Pederal, and Zuni uplifts together with the smaller Peñasco, Joyita, and Florida structures. In most cases, the manner in which the Gray Mesa wedges out against these features is poorly documented.

The greatest reported thickness of the Gray Mesa is as much as 550 m in the San Mateo Mountains (Table 2). Several faults interrupt the section (Kottlowski 1959, 1960). Siemers (1983) discussed this remote section, but whether he personally visited it is not clear. The San Mateo section provides the sole basis for inferring a small Pennsylvanian sub-basin (San Mateo basin) in southern Socorro County.

The Lucero basin or sub-basin appears better established, based on Gray Mesa thicknesses that approach 400 m. This is double the thickness reported from mountain ranges to the east. The greatest thickness in southern New Mexico is in the Fra Cristobal Mountains, where the Gray Mesa exceeds 300 m.

The overall pattern in much of the region is one of narrow, elongate north-trending fault-bounded basins and uplifts. Broadhead (2001) described three Pennsylvanian "elevator basins on the eastern side of the Rio Grande valley," only one of which has been drilled. The term "elevator" refers to small basins having large vertical displacements, in some cases direction of throw reversing during the Pennsylvanian. The true pattern of Gray Mesa thickness is probably far more complicated than shown on the isopach map (Fig. 9).

TABLE 1. Section on north face of Robledo Peak measured by present authors.

Thickness	Lithology	Age	Name
27 m	Massly covered (probably shale) with thin limestone layers	Atokan	Honquilla (Red House equivalent)
55 m	Cherty, bioclastic limestone, thin to thick bedded, forms ledges	Lower and middle Desmoinesian	Honquilla (Gray Mesa equivalent)
25 m	Shale with thin bedded and nodular limestone layers, conglomerate near middle	Upper Desmoinesian and lower Missourian	Honquilla
193 m	Dominantly thick bedded to massive limestone, numerous algal bioherms	Upper Missourian to lower Wolfcampian	Honquilla

TABLE 2. Thickness and age of Gray Mesa Formation.

Area	Thickness (m)	Age	Sources
Nacimiento and Jemez Mts.	0 to 100	Late Atokan and early Desmoinesian	Wood and Northrop (1948), Read and Wood (1947)
Southern Sangre de Cristo	180 to 500, thicker north	Lower and middle Desmoinesian	Baird and Myers (1999)
Sandia Mountains	137 to 150	Desmoinesian	Kelley and Northrop (1973)
Cedro Peak	124	Latest Desmoinesian to middle Virgilian	Lucas et al. (2009), Vachard et al. (2012, 2013)
Manzano, Manzanita, and Los Pinos Mts.	180 to 250	Early Desmoinesian to earliest Missourian (?)	Read and Wood (1947), Myers (1973 and 1977), Myers and McKay (1974 and 1976), Myers et al. (1981 and 1986), S.G. Lucas and K. Krainer, unpublished data
Lucero uplift	180 to 369	Early, but not earliest Desmoinesian to late Desmoinesian	Kelley and Wood (1948), Wenger (1959), Martin (1971), Krainer and Lucas (2004)
Southern Ladron Mts.	185	Early to late Desmoinesian	Kottowski (1959)
Joyita uplift and hills east of Socorro	0 to 290	Late Atokan to earliest Missourian (varies locally)	Wipolt and Wanek (1951), Kottowski and Stewart (1970), Cather and Colpits (2005), Lucas et al. (2009)
Magdalena Mts.	172	Desmoinesian	Read and Wood (1947)
Little San Pascual Mts.	136	Desmoinesian	Siemers (1983)
Oscura Mts.	82 to 162	Late Atokan to early Missourian, sources differ	Bachman (1946), Lucas and Krainer (2009)
San Andres Mts.	110 to 200	More or less entire Desmoinesian	Kottowski et al. (1956), Seager (1981)
San Mateo Mts.	275 to 550	Desmoinesian	Read and Wood (1947), Kottowski (1960)
Sierra Cuchillo	380?	Uncertain	Johns (1955)
Fra Cristobal Mts.	210 to 330	Not determined	Caena (1956), McCleary (1982), Lucas et al. (2012)
Mud Springs Mts.	134 to 215	Early and middle Desmoinesian	Thompson (1942), Maxwell and Oakman (1995), Lucas et al. (2012)
Caballo Mts.	150 to 200	Early and middle Desmoinesian	Kelley and Silver (1952), Seager and Mack (2003), Lucas et al. (2012)
Robledo Mountains	55	Early and middle Desmoinesian	S.G. Lucas and K. Krainer, unpublished data

Lithology

By definition, the Gray Mesa is a unit composed primarily of limestone, in contrast to formations above and below, which contain higher proportions of siliciclastic rocks. In most places, 70% or more of the thickness of the Gray Mesa is limestone. Under the arid climate of New Mexico, the Gray Mesa consists of stepped cliffs punctuated by occasional ledges (Figs. 8, 10, 11, 12). As the name implies, colors (fresh and weathered) are largely shades of gray. Often described as “massive,” the Gray Mesa seldom is truly so. On close inspection, even bold cliffs commonly show thin bedding, the layers separated by clay or stylolitic partings.

Cherty limestone of the Gray Mesa Formation is massive to indistinctly bedded, medium to thick bedded, wavy bedded and nodular. Non-cherty limestone commonly is thin and wavy bedded. Fossils observed on outcrop include echinoderm fragments, brachiopods, fusulinids, bryozoans, calcareous algae, solitary corals and *Syringopora*, and *Zoophycos* (trace fossil). The demosponge *Chaetetes* (Fig. 13) is common near the base of the formation.

A wide variety of limestone textures are represented. Lime mudstone and skeletal wackestone, packstone and floatstone are most preva-

lent, skeletal and crinoidal grainstone-packstone are subordinate, and rudstone and foraminiferal grainstone to packstone are rare. Grains are mostly fossil fragments, including fragments of echinoderms (mostly crinoids) and bryozoans (most abundant) along with brachiopods, mollusks, and corals. Thin sections also reveal small foraminifera (*Bradyina*, *Calcivertella*, *Climacammina*, *Cornuspira*, *Earlandia*, *Endothyra*, *Endothyranella*, *Eolasiiodiscus*, *Eotuberitina*, *Globivalvulina*, *Glomospira*, *Hemigordius*, *Nodosinelloides* (?), *Palaeonubecularia*, *Polytaxis*, *Spireitlina*, *Syzrania*, *Tetrataxis*, *Tuberitina*) and fusulinids, ostracods, brachiopod spines, sponge spicules, calcareous algae, and algaespongia (mostly phylloid algae, *Ivanovia*, *Eugonophyllum*, rare *Komia*, *Anthracooporellopsis*, *Asphaltina*, *Epimastopora*, *Efluegelia*, *Fourstonella*; see Vachard et al. 2012, 2013), rare trilobite fragments, *Tubiphytes*, echinoid spines and calcisponges. Locally spiculite is present, composed of abundant unoriented calcified sponge spicules. Peloids and intraclasts are common, but oolitic limestone is uncommon. Some layers are sandy, grading to calcareous sandstone. Nodular, conglomeratic, brecciated, and intensively burrowed textures are present. Dolomite is a minor lithology, largely microgranular and weathers orange or yellow.

Chert is generally conspicuous, making up 10% to more than 50% of beds. Colors are mostly gray, brown, and black. Chert occurs as “isolated microscopic patches of chalcidony” (Siemers 1983, p. 152) together with conspicuous lenses, nodules, ribbon bands, and intricate irregular masses (Fig. 14). Fossils are commonly silicified; weathered outcrop surfaces are rough and jagged due to dispersed silica.

Most Gray Mesa sections include interbeds of shale, siltstone, and sandstone; conglomerate and non-fissile mudstone are rare. Shale and siltstone are mostly gray, greenish gray, and olive gray, calcareous and moderately fissile. Being non-resistant to erosion, shale and siltstone tend to be poorly exposed and are commonly covered except in deep ravines and artificial cuts. Sandstone is variable in grain size, sorting and rounding and composition. Most of it is quartz arenite and subarkose containing abundant quartz grains, detrital potassium feldspars, and a few granitic rock fragments. Sandstone beds commonly are lenticular or channel-form and range up to 10 m thick in the Cerros de Amado of Socorro County (Lucas et al., 2009).

Members

The Elephant Butte, Whiskey Canyon, and Garcia formations of Thompson (1942), together, comprise the Gray Mesa Formation in their type area, the Mud Springs Mountains of Sierra County. Revised to members of the Gray Mesa Formation, these units can be identified in some, but by no means all Gray Mesa exposures.

Basically, the medial Whiskey Canyon Member is almost entirely limestone, which has high silica (chert) content and erodes to a bold escarpment (Fig. 15). The Elephant Butte and Garcia members contain higher proportions of siliciclastic rocks and of thin-bedded or nodular limestone and hence, are less resistant to erosion than the Whiskey Canyon (Fig. 5). The relationship of the Whiskey Canyon to the Gray Mesa thus mirrors the relationship of the Gray Mesa to the Pennsylvanian as a whole.

Closest to the Mud Springs Mountains, the three members of the Gray Mesa are not evident in many of the outcrops in the Fra Cristobal or Caballo Mountains (Lucas et al., 2012). They also are not apparent in the Oscura Mountains (Lucas and Krainer, 2009). However, the members are readily identified east of Socorro (Rejas, 1965; Lucas et al., 2009) and appear to persist as far north as Cedro Peak in the Manzanita Mountains east of Albuquerque (S.G. Lucas and K. Krainer, unpublished data). Contacts of the members are not sharply defined and probably do not mark time lines.

Cyclicality

Authors working in widely separated areas of New Mexico have described cyclicality in the Gray Mesa Formation and correlative strata. Patterns of cyclicality vary from one place to another, and authors differ on their inferences about the driving mechanism.

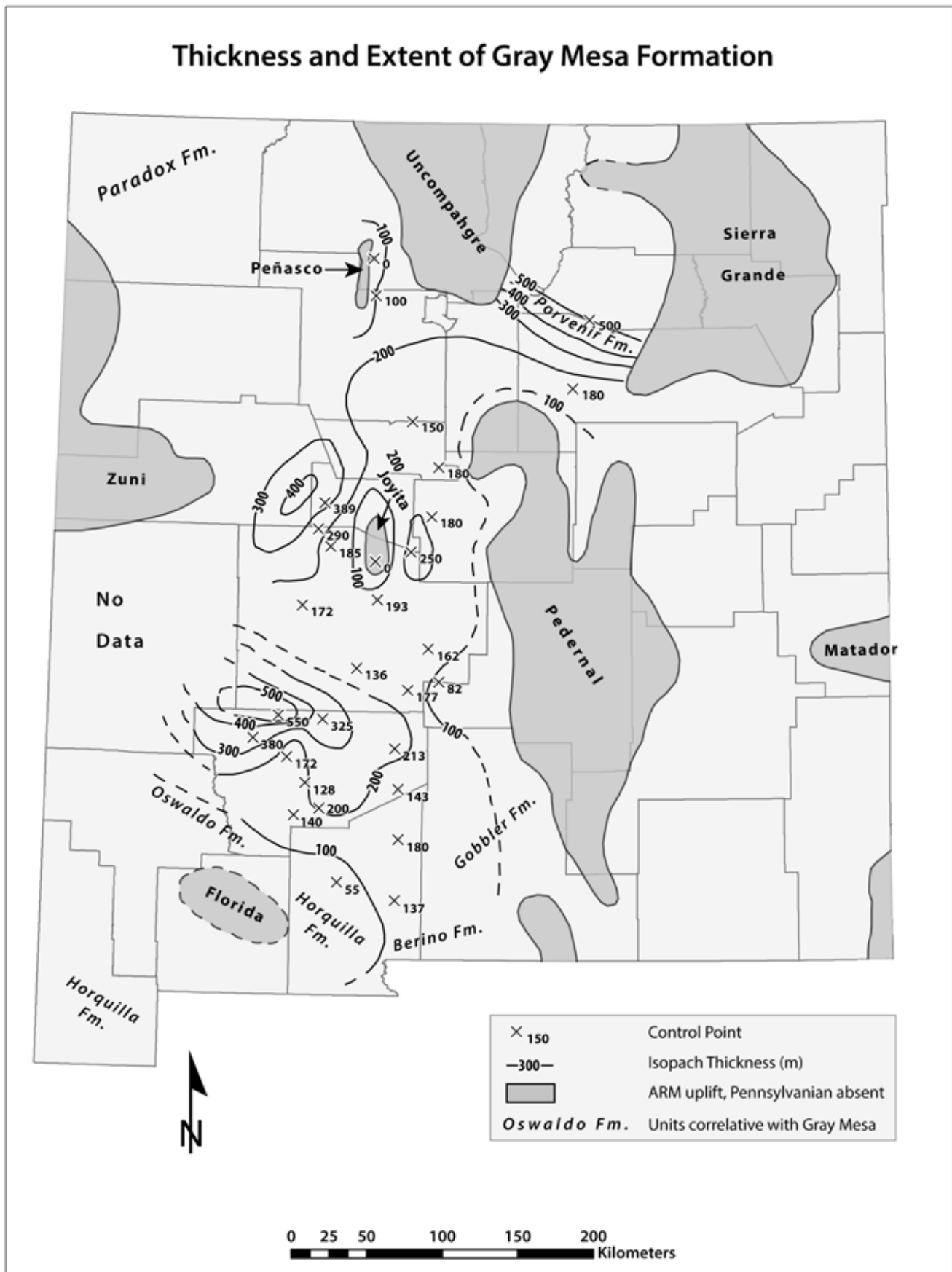


FIGURE 9. Map of New Mexico showing extent and thickness of the Gray Mesa Formation.



FIGURE 10. Photograph on the western flank of South Ridge in the Caballo Mountains, illustrating contrast between shaly, slope-forming Red House and overlying stepped cliffs of Gray Mesa limestone. At this locality the Red House Formation measured 93 m thick (Lucas et al., 2012).

Near the Gray Mesa type locality, Scott and Elrick (2004) identified about 77 shoaling-upward cycles (average thickness 3.7 m). They interpreted two types of cycles: (1) entirely regressive and (2) thin basal transgressive unit overlain by thicker regressive unit. Features indicating subaerial exposure top many cycles. In turn, meter-scale cycles appear to be stacked into $4\frac{1}{2}$ larger sequences. Lacking tight control on the duration of Gray Mesa cycles, Scott and Elrick (p. 41) noncommittally concluded “that high-frequency glacio-eustatic sea-level fluctuations and associated climate changes and/or long-term changes in accommodation space brought about by Ancestral Rocky Mountain deformation influenced marine sedimentation in the Lucero basin during the Middle Pennsylvanian.” However, these authors favored eustasy over tectonism because exposing the tops of carbonate cycles would require recurrent up-and-down movements that do not comport with modern tectonic models.

Small-scale cycles are evident in many outcrops of the Gray Mesa Formation. Examples in sections measured by two of us (WJN and SDE) are from highway cuts near Cedro Peak, about 25 km east of Albuquerque (Fig. 16). The left-hand column contains six cycles that appear to record shoaling through time. Each begins with shale at the base, changing upward to lime mudstone, wackestone, and packstone; and capped in some cases by subaerial exposure features such as algal, brecciated limestone. The right-hand column contains four cycles that show the opposite trend, grading from packstone or grainstone at the base to wavy-

bedded lime mudstone at the top. These cycles seem to indicate decreasing depositional energy and deepening through time. To correlate such cycles even locally is difficult because of lack of continuous exposures and numerous faults. Regional correlation would require precise biostratigraphic control, which is not available.

Wiberg (1993) and Wiberg and Smith (1994) correlated sections of the “lower Madera Limestone” in the Sandia Mountains with the lower part of the Desmoinesian Los Moyos Limestone of Myers (= Gray Mesa Formation) at Cedro Peak and provided a cycle-to-cycle correlation. According to these authors, the “lower Madera Limestone” in the Sandia Mountains is characterized by 4th-order transgressive-regressive cycles, which they correlated with those in other regions, concluding that cycle formation was mainly controlled by eustatic sea-level changes and less by tectonic movements of the Ancestral Rocky Mountain deformation.

However, two of us (SGL and KK) have re-examined some of the sections in the work of Wiberg (1993) and Wiberg and Smith (1994), and are unable to recognize many of the cycles they posit. Indeed, their key section in the Sandia Mountains, at Tejano Canyon, is a series of fault blocks (not the homoclinal section Wiberg and Smith present), mostly of Sandia Formation, and contains little Gray Mesa Formation. Moreover, there is no age control of their measured sections, as they did not identify any fusulinids; and it is unclear how Wiberg (1993) and Wiberg and Smith (1993) correlated the top of their sections to the Cedro Peak section of Myers and McKay (1976).

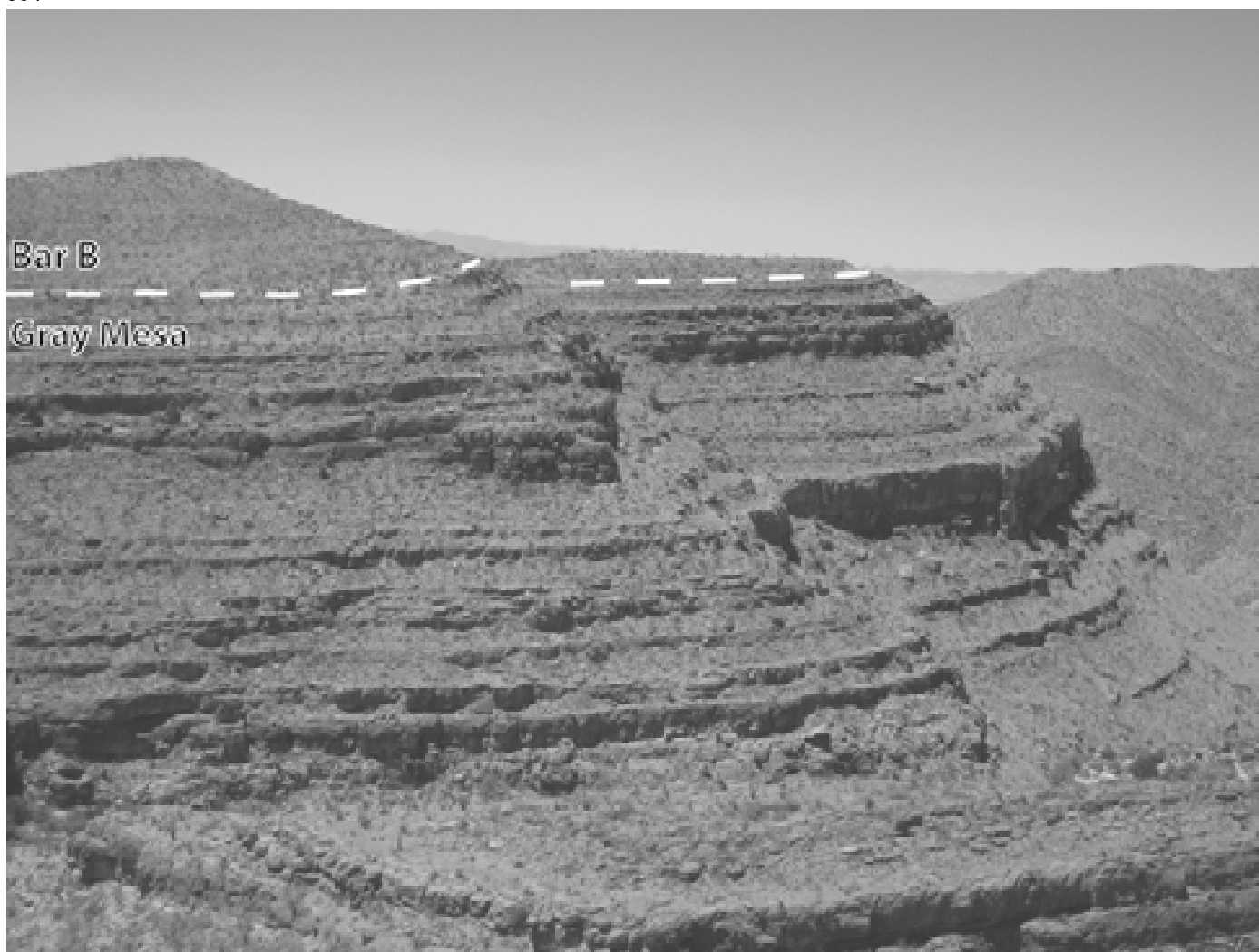


FIGURE 11. Photograph in Hellion Canyon, Fra Cristobal Range, showing contrast between stepped cliffs of the Gray Mesa and rounded slopes of the overlying Bar B Formation.

Baltz and Myers (1999) described cycles that appear similar to those of Wiberg and Smith in the Porvenir Formation (Gray Mesa equivalent) of the southern Sangre de Cristo Mountains. Without embarking on a detailed analysis of cyclicity, Baltz and Myers favored tectonics as the primary driving mechanism.

In the Orogrande basin of southern New Mexico, Algeo et al. (1991) and Algeo (1996) described two levels of cyclicity in Desmoinesian limestone. Smaller cycles, generally 3 to 10 m thick, were termed parasequences. Shale or nodular, marly limestone at the base of each parasequence passes upward to cherty, burrowed wackestone topped by fossiliferous packstone or grainstone. Oxidized sandy crusts, fossil lags, breccias, and karst features indicate subaerial exposure at the tops of parasequences. Larger bundles are termed sequences, each commonly containing 4 to 6 parasequences and showing an overall upward decrease of shale and marl and an increase in coarser, “cleaner” limestone. The authors correlated sequences and some parasequences more than 100 km across the basin, from margins to basin center. They favored glacial eustasy as the primary process controlling cyclicity.

Also in the Orogrande basin, Soreghan (1994a, b) described the same sort of cycles in Missourian and Virgilian strata. Using fusulinid control, she correlated cycles between the Orogrande and the Pedregosa basin in southeastern Arizona.

Raatz et al. (1994) described transgressive-regressive cycles in late Desmoinesian and Missourian rocks of the Sacramento Mountains.

Cycles are mostly 10 to 15 m thick and have a thin basal transgressive unit (commonly just a lag zone) overlain by thick, upward-shoaling regressive strata. Water depth fluctuated repeatedly from below storm wave base to supratidal exposure.

Connolly and Stanton (1992) identified about 50 shoaling-upward cycles in the Horquilla Limestone of southeastern Arizona. Their measured sections span the entire Gray Mesa time interval, from Atokan through lower Missourian. Cycles typically contain (1) lower calcareous siltstone, (2) middle nodular to wavy-bedded, shaly lime mudstone and wackestone, and (3) upper thick-bedded to massive wackestone, packstone, and grainstone commonly showing peritidal and pedogenic features at the top. Using fusulinid biostratigraphy, Connolly and Stanton correlated Horquilla cycles on a one-for-one basis with those of southeastern Kansas, thereby acknowledging glacial eustasy as the primary driver. However, it is important to note that the Horquilla was deposited in a depositional basin separate from that of the Gray Mesa Formation.

Cyclicity of Middle and Upper Pennsylvanian rocks in the Paradox basin has long been appreciated (Hite and Cater, 1972; DeVoto, 1980; Hite and Buckner, 1981; Reid and Berghorn, 1981). Using conodont biostratigraphy, cycles in the Paradox basin have been matched with those in the Midcontinent of the United States (Ritter et al., 2002).

Within the Gray Mesa Formation at several locations (including Cedro Peak and Priest Canyon in the Manzano Mountains, Bruton Canyon in the northern Oscura Mountains, Cerros de Amado, Fra



FIGURE 12. Photograph of “Amphitheater Canyon” on the steep western face of the Fra Cristobal Range, where nearly the entire Pennsylvanian succession is displayed.

Cristobal, Caballo Mountains), an alternation of distinct lithofacies is observed, and shallowing-upward sequences are partly developed, particularly within the Whiskey Canyon Member (shale-wackestone-crinoidal packstone). But, we did not observe that the entire Gray Mesa Formation is composed of well-developed shallowing-upward cycles as described by Wiberg (1993) and Wiberg and Smith (1994) from the Sandia Mountains and by Scott and Elrick (2004) from the Lucero uplift.

Cherty limestone is very common at sections in the Manzano Mountains, northern Oscura Mountains and Whiskey Canyon, particularly within the Whiskey Canyon Member. It seems to be almost absent in the sections studied by Wiberg (1993) and Wiberg and Smith (1994) and is also less common in the Cerros de Amado section. On the other hand, subaerial exposure surfaces, grainstone and coarse siliciclastic sediments (subarkosic sandstone) are common in the sections in the Sandia Mountains, but rare to absent at many other sections. *Chaetetes* is common at the type section of the Gray Mesa Formation in the Lucero uplift and rare to absent in other sections.

This indicates that the depositional environment at the sections in the Manzano Mountains, northern Oscura Mountains and Mud Springs Mountains (Whiskey Canyon), for example, was deeper compared to the Sandia Mountains and Lucero uplift: open marine shelf mostly below wave base, but still within the photic zone, probably some tens of meters deep.

Shallowing-upward cycles cannot be correlated between the indi-

vidual sections and to other regions (partly due to lack of detailed biostratigraphic age control), indicating that tectonics may have influenced deposition of the Gray Mesa Formation more strongly than suggested by Wiberg (1993), Wiberg and Smith (1994) and Scott and Elrick (2004).

Summarizing these findings, Middle Pennsylvanian cyclicality appears to be more prominent and regular in southern New Mexico than in the central and northern parts of the state. The southern area lay farther from terrestrial sediment sources, and for the most part may have been more stable tectonically. This allowed a stronger and clearer eustatic signal in the south, as in the Midcontinent and Illinois basins. Ongoing ARM tectonism and frequent influxes of terrestrial detritus often masked eustatic cycles in central and northern New Mexico. What is also clear is the need for more detailed age control within the Gray Mesa Formation, and more rigorous and consistent analysis of cyclicality across a larger part of its outcrop belt to answer questions about cycle duration and causation.

Regional Variation

As outlined in an earlier section, the Gray Mesa Formation and its enclosing strata change character near the defined lateral limits of the formation. Working clockwise from the north, these changes may be summarized as follows.

To the north in the Sangre de Cristo Mountains, the Gray Mesa intertongues with shale, siltstone, sandstone, and conglomerate that were



FIGURE 13. Photograph of silicified *Chaetetes*, the demosponge that is fairly common and conspicuous near the base of the Gray Mesa. This specimen is in Whiskey Canyon.

shed from the rising Uncompahgre uplift. On the east, the Gray Mesa either wedges out or is truncated by erosion along the flanks of the Sierra Grande and Pedernal uplifts. The nature of this termination is poorly known except in the Sacramento Mountains, where rapid facies changes across short distances are evident between Desmoinesian clastic and carbonate rocks (Gobbler Formation).

In the Organ Mountains east of Las Cruces, the Gray Mesa gradually loses its identity as the underlying strata change to all limestone while the Desmoinesian section takes on increasing clastic layers. Toward the southwest, the entire Pennsylvanian section becomes marine carbonate rocks with only a few thin shale and siltstone layers. The westward extent of the Gray Mesa Formation is poorly known due to volcanic cover and the scarcity of drilling records, but Pennsylvanian rocks wedge out or are eroded along the flanks of the Zuni uplift (Krainer et al., 2003). In the San Juan Basin, the Gray Mesa might be extended nearly to the northwest corner of New Mexico, where it intergrades with mixed evaporite and carbonate rocks of the Paradox Formation.

Increased amounts of sandstone and shale may be present in the lower and upper parts of the Gray Mesa Formation in northern to central Socorro County. In places, the lower and upper Desmoinesian strata reportedly become dominantly clastic, arguably restricting Gray Mesa lithology to the middle Desmoinesian (Whiskey Canyon Member). These relationships are suggested by earlier studies in the southern Los Pinos Mountains, together with the southern Sierra Ladrones and to a lesser extent, Cerros de Amado east of Socorro (Wilpolt et al., 1946; Read and Wood, 1947; Kottlowski, 1959, 1960; Myers et al., 1986; Lucas et al., 2009). The area of increased clastics seems to center around the Joyita Hills, an active ARM uplift during the Pennsylvanian. However, our own recent investigations and those of Bruce Allen (written communication, 2013) point to inconsistencies in some of the earlier

studies, which have yet to be fully resolved. Also, the lower Desmoinesian in the Joyita Hills is dominantly limestone, whereas the upper Desmoinesian is absent due to unconformity (Kottlowski and Stewart 1970). Most of the clastics may have come from the Zuni uplift to the west or the Pedernal uplift to the east (Kottlowski 1959 and 1960, Kottlowski and Stewart 1970). However, pinning down the source is difficult due to a paucity of control points.

Contacts

Lower and upper contacts of the Gray Mesa Formation vary in character and cross time lines. Unconformable in some places, these contacts elsewhere vary from sharp and conformable to gradational, intertonguing, and arbitrary.

Lower contact: The lower contact can be unconformable where the Gray Mesa laps onto ARM uplifts. For example, the Gray Mesa locally rests directly on Precambrian basement around the flanks of the Peñasco uplift in the Nacimiento Mountains (Wood and Northrop 1946; Woodward, 1987, 1996). Similar relationships occur on the southern flank of the Zuni uplift (Kottlowski, 1960), and the northeastern flank of the Florida uplift at Cookes Peak (Jicha, 1954). Such relationships probably signify structures that were actively rising during Gray Mesa deposition.

In a few places away from active ARM structures, the Gray Mesa Formation disconformably overlies older Pennsylvanian rocks. For example, at Ash Canyon in the southern San Andres Mountains, “Desmoinesian beds overlie Derryan [Atokan] strata with a relatively pronounced unconformity” (Kottlowski et al., 1956, p. 37). Basal Desmoinesian strata here include coarse, cross-laminated sandstone and sandy limestone. Also, fusulinid evidence suggests that upper Desmoinesian strata are absent. The Gray Mesa section at Ash Canyon is markedly thin compared to localities north and west.



FIGURE 14. Photograph showing typical intricate network of scraggly chert on weathered surface of Gray Mesa limestone, again from Whiskey Canyon.

Elsewhere, as Siemers (1983) observed, picking the lower contact of the Gray Mesa Formation can be somewhat arbitrary. Thick limestone layers occur in upper Sandia and Red House Formation sections, sandstone and shale layers in the lower Gray Mesa. Generally, we place the contact at the base of the succession that is dominantly limestone and in particular where the limestone is cherty. This contact may be sharp and planar, irregular, or gradational from calcareous shale to shaly limestone.

Upper contact: The upper contact of the Gray Mesa Formation can be conformable and gradational, a minor disconformity, or a major unconformity.

On the Joyita uplift north of Socorro, Missourian rocks rest unconformably on lower to middle Desmoinesian strata. Upper Desmoinesian layers are absent. The Bursum Formation (Virgilian to Wolfcampian) truncates Gray Mesa and other Pennsylvanian units with an angular unconformity (Kottlowski and Stewart 1970).

Wengerd (1959) and Krainer and Lucas (2004) indicate the Gray Mesa-Atrasado contact is disconformable in the Lucero uplift. The Bartolo

Member of the Atrasado is clearly evident on Wengerd's detailed graphic column from Monte de Belen (Mesa Sarca). Dominantly shale, the Bartolo contains a thin basal sandstone or conglomerate overlying Gray Mesa limestone.

An erosive channel is present on the north side of Whiskey Canyon in the Mud Springs Mountains at what may be considered the top of the Gray Mesa Formation. Thompson (1942) showed this feature in the upper part of his Bolander Group, near the top of the Desmoinesian Series. The channel has about 8 m of local relief and contains cobble-sized limestone clasts, but when traced north of the canyon, the contact appears to become conformable. Whether this is a regional unconformity related to sea-level drop or merely a local feature, such as a tidal channel, has not been determined.

In the Manzanita, Manzano, and Los Pinos Mountains, basal Atrasado sandstone (the "Coyote sandstone" of Herrick) commonly overlies Gray Mesa limestone with a sharp contact. This surface, however, has not been reported to be unconformable (Myers 1973, Myers et al. 1986), but we see it as a clear sequence boundary that represents at least a short hiatus.



FIGURE 15. Photograph looking eastward in Whiskey Canyon, with shaly limestone of the Elephant Butte Member poorly exposed in the foreground, overlain by stepped cliffs of the Whiskey Canyon Member where the gorge narrows.

AGE

Fusulinids have provided the primary means of dating the Gray Mesa Formation since the 1940s. Their biostratigraphy is well established, providing correlation to Midcontinent stages and substages. Many authors have published fusulinid data on Pennsylvanian rocks of New Mexico. Conodonts provide potentially greater resolution for zonation and also are better established than fusulinids for global correlation in the Late Paleozoic. To date, Lucas et al. (2009, 2012; also see Barrick et al., 2012) are the only authors who have published conodont data relevant to the age of the Gray Mesa Formation.

Not surprisingly, fusulinid and conodont data confirm that the upper and lower formational contacts of the Gray Mesa cross time lines. Basal beds range in age from late Atokan to early Desmoinesian; whereas the uppermost Gray Mesa varies from early to late Desmoinesian, although most records indicate late Desmoinesian age (Table 2).

The Gray Mesa Formation is entirely Desmoinesian in most of Socorro and Valencia Counties. In the Lucero uplift, the upper part of the Sandia Formation contains Desmoinesian fusulinids. These continue throughout the Gray Mesa and into the Bartolo Member of the Atrasado Formation, the highest occurrence being 50 to 60 m above the base of the Bartolo (Wengerd, 1959; Martin, 1971). Conodont data from east of Socorro likewise place the Desmoinesian-Missourian boundary in the upper part of the Bartolo Member. The base of the Desmoinesian is poorly constrained in this area, but is close to the base of the Gray Mesa Formation (Lucas et al., 2009).

A graphic column by Kottlowski (1959) shows fusulinid control for the section in the southern Sierra Ladrone, Socorro County. Although Kottlowski did not identify formations, the Sandia is evidently the interval of shale, sandstone, and thin limestone that contains Derryan [Atokan] and early Desmoinesian fusulinids. Above this is 185 m of mostly limestone, entirely Desmoinesian. The lowest occurrence of Missourian fusulinids is about 45 m above the base of what appears to be the Bartolo Member of the Atrasado Formation. In the Joyita Hills of Socorro County, the lower part of the Gray Mesa Formation contains upper Atokan fusulinids, whereas the upper Gray Mesa is lower Desmoinesian (Kottlowski and Stewart, 1970).

In the Cerros de Amado east of Socorro, Rejas (1965), Lucas and Estep (2000) and Lucas et al. (2009) noted the occurrence of Desmoinesian fusulinids in all three members of the Gray Mesa Formation. However, they provided little in the way of specific information, other than to note

the occurrence of *Beedeina* species in the upper part of the formation.

Myers (1973) reported that the Los Moyos [= Gray Mesa] Formation in the Manzano and Manzanita Mountains ranges from basal Desmoinesian to early Missourian, based on fusulinids. However, a close examination of his work suggests no evidence for a Missourian age for any part of the Gray Mesa Formation. Instead, Missourian fusulinids do not appear in the section until well above the base of the Atrasado Formation (Lucas et al., 2011).

Abundant fusulinid control in the southern Sangre de Cristo Mountains indicates that the entire Porvenir Formation is of Desmoinesian age. The Atokan-Desmoinesian boundary closely corresponds with the Sandia-Porvenir contact. Upper Desmoinesian fusulinids occur in the upper Porvenir and extend into the lower part of the overlying Alamitos Formation (Baltz and Myers 1999).

Without naming fossils, Wood and Northrop (1946) stated that the Sandia Formation is Morrow and Lampasas (Atokan) age in the Nacimiento Mountains, whereas the “upper arkosic member of the Madera Limestone” [Atrasado Formation] is early Desmoinesian through Virgilian age in the same area. This leaves the Gray Mesa to span part of Atokan and early Desmoinesian time. Note that the Morrowan strata referred to by Wood and Northrop (1946) are now assigned to the Osha Canyon Formation.

In the southern Oscura Mountains, Bachman and Harbour (1970) reported the Atokan guide fossil *Fusulinella* in the lower 15 m of the “lower member of the Madera” [Gray Mesa]. The youngest part of the Gray Mesa contained *Wedekindellina*, a genus regarded as early to middle Desmoinesian. The Missourian fusulinid *Triticites ohioensis* near the base of the overlying “upper member” [Atrasado] implies “a local faunal hiatus” at the Gray Mesa-Atrasado contact (Bachman and Harbour 1970, p. 22).

DEPOSITIONAL ENVIRONMENTS

General Environments

Lithology and the abundant, diverse fauna signify that Gray Mesa sediments accumulated almost entirely in marine settings. The abundance of filter feeders points to well oxygenated water, low turbidity, and near normal salinity. As discussed above in the section on lithologic cyclicity, water depth fluctuated repeatedly throughout Gray Mesa deposition. Many cycles record a range from deep water (deeper shelf, below storm wave base) to peritidal conditions. Features indicative of subaerial

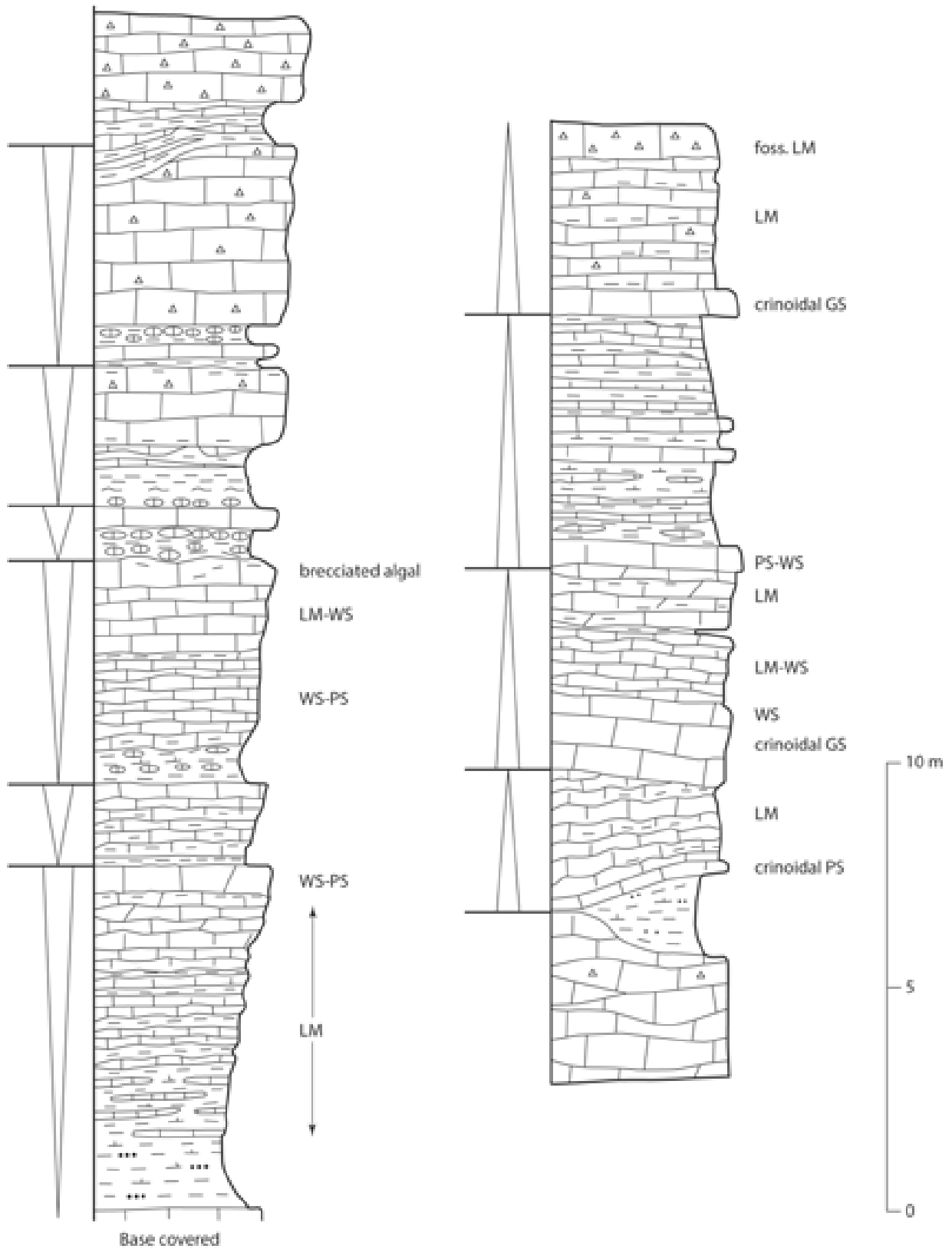


FIGURE 16. Graphic column of Gray Mesa sections measured by S.D. Elrick and W.J. Nelson along New Mexico Highway 337 near Cedro Peak in the Manzanita Mountains, illustrating depositional cycles. In the left-hand column, cycles generally begin with shale at the base and proceed upward through lime mudstone to wackestone and packstone, having in some cases brecciation suggesting subaerial exposure at the top. Right-hand column shows a different style of cycles; transgressive crinoidal packstone or grainstone at base changes upward to thinly bedded, shaly micritic limestone evidently deposited in deeper water. The right-hand column suggests more rapid subsidence. Due to numerous faults, correlation of the two columns is uncertain.

exposure occur at the tops of many cycles (Algeo et al., 1991, Wiberg and Smith, 1994; Soreghan, 1994a; Algeo, 1996; Scott and Elrick, 2004;). The general lack of siliciclastic layers in the Gray Mesa reflects reduced ARM tectonic activity, as compared to earlier and later in the Pennsylvanian. The Uncompahgre uplift was the most active structure in New Mexico during Gray Mesa deposition. Most of the detritus eroded from this uplift ended up in the nearby, rapidly sinking Taos trough. Abrupt facies changes in the Sacramento Mountains indicate some activity along the Pedernal front in that area. Otherwise, ARM uplifts contributed modest amounts of detritus to Gray Mesa seas. Pinning down where individual clastic wedges originated is difficult.

Chert is a conspicuous component of the Gray Mesa, but its origin has received little attention. As proposed by Cecil (2004), eolian dust is a likely source of at least some silica. Also, the presence of sponge spicules, locally enriched to form spiculite, indicate that siliceous sponges were the direct source of at least some of the silica. The ultimate source might have been wind-blown dust. Under an arid climate, the coastal plains and uplands surrounding the Paradox basin and Central Colorado trough probably generated large quantities of wind-blown silt.

Paleoclimate

Algeo (1996) concluded that climate in the Orogrande basin was warm and humid because Desmoinesian rocks lack dolomite, evaporites,

caliche, and desiccation features. Also, he (p. 360) described evidence for “strong meteoric diagenesis of subaerially exposed surfaces.” These conditions contrast with the Paradox basin and Central Colorado trough, where rocks of the same age include thick successions of anhydrite, halite, and other evaporites. Algeo cited plate reconstructions placing southern New Mexico close to the equator during the Middle Pennsylvanian, whereas Colorado and Utah lay at 10° to 20° north latitude, in the “horse latitudes” where many of Earth’s deserts are found today.

Soreghan (1994b) discussed numerous indicators that Late Pennsylvanian climate fluctuated in the Orogrande basin. Glacial episodes were linked with marine lowstands and semi-arid climate, interglacials with highstands and subhumid conditions. Seasonal wet-dry climate prevailed throughout. She attributed these changes to climate belts shifting north and south, further influenced by rain shadows from newly uplifted mountains.

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THE ATRASADO AND BAR B FORMATIONS (MIDDLE-UPPER PENNSYLVANIAN) IN CENTRAL AND SOUTHERN NEW MEXICO

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Abstract—The Atrasado and Bar B formations are broadly correlative units of interbedded carbonate and clastic rocks that crop out in the mountain ranges bordering New Mexico's Rio Grande Valley. The Atrasado extends from the Sandia Mountains of Bernalillo County on the north to the Mud Springs Mountains of Sierra County on the south, whereas the Bar B Formation occurs in the Fra Cristobal and Caballo mountains of Sierra County. Both units generally erode to rounded slopes and small ledges, in contrast to bold escarpments of the underlying, limestone-dominated Gray Mesa Formation. Characterizing the Atrasado are beds of arkosic sandstone and conglomerate, together with shale, mudstone, siltstone, limestone, and dolomite. Some of the carbonate units are widely traceable, allowing the Atrasado to be divided into members. The Bar B lacks sandstone and consists largely of micritic, shaly limestone and dolomite. Some fossiliferous and algal limestone layers produce discontinuous ledges, but the Bar B has not been subdivided. The lower contact is generally conformable, whereas the upper is commonly disconformable and locally angular. Fusulinids and conodonts indicate an age range from middle Desmoinesian to late Virgilian, but both contacts cross time lines. The Atrasado and Bar B record largely marine sedimentation ranging from below wave base to peritidal with low to moderate wave and current energy. Fluvial, lacustrine, and coastal-plain environments also are represented. Many outcrops display cyclicity, yet local tectonic activity commonly prevailed over eustatic controls.

INTRODUCTION

As outlined in related articles in this volume, Pennsylvanian rocks (excluding the Bursum Formation) may be divided into three formational units across large areas of New Mexico. The lower unit, of interbedded clastic and carbonate rocks, includes the Sandia and Red House formations. The middle unit, dominantly cherty limestone, is the Gray Mesa Formation. This chapter is concerned with the upper unit, which consists of limestone interbedded with clastic rocks that range from claystone to conglomerate. These rocks have been termed the Atrasado and Bar B formations, as they crop out in various locations across New Mexico (Fig. 1).

GEOLOGIC SETTING

Tectonic Setting

The Ancestral Rocky Mountains orogeny, which commenced during latest Mississippian or Early Pennsylvanian time, continued throughout the entire Pennsylvanian. The same general pattern of basins and uplifts discussed in the previous chapter on the Gray Mesa Formation (Nelson et al., 2013) prevailed during deposition of the Atrasado and Bar B formations (also see Nelson and Lucas, 2011). The most significant change was that subsidence accelerated markedly in the southern part of the Orogrande basin during the Late Pennsylvanian. Consequently, sediments deposited here differ substantially from those found farther north, and are called by a different name: Panther Seep Formation. Discussion of the Panther Seep, however, is beyond the scope of this article.

Conditions of Study

Outcrop study is the basis for almost everything known about Upper Pennsylvanian rocks in New Mexico. Outcrops are confined to isolated mountain ranges, products of Cenozoic crustal extension that produced the current basin-and-range topography of the Rio Grande Valley and adjacent valleys of southern New Mexico. Under the regional

arid to semi-arid climate, limestone and sandstone erode to ledges and small cliffs, whereas finer-grained clastic rocks erode to slopes and valleys. The Atrasado and Bar B formations commonly occur on dip slopes above escarpments of the Gray Mesa Formation. Good exposures of the less resistant lithologies are confined to a few deep canyons and artificial cuts.

HISTORY OF STUDY

Most of the geologists who wrote about Upper Pennsylvanian rocks in New Mexico considered the Pennsylvanian as a whole. The history of study of the Atrasado and Bar B formations therefore closely parallels that of the older Gray Mesa Formation (Nelson et al., 2013) and Sandia Formation (Krainer and Lucas, 2013). To avoid repetition, the reader is referred to these earlier chapters in this volume.

Many different naming schemes have been applied to Pennsylvanian rocks in New Mexico (Fig 2). Most geologists have assigned the Upper Pennsylvanian mixed clastic-carbonate succession to a single formation or member of a formation. Thompson (1942), in contrast, assigned these rocks to four groups and seven formations (Fig. 2). Although many later workers ignored or disparaged Thompson's nomenclature, we have found that a number of his units are useful at the rank of member (also see Lucas and Krainer, 2009; Lucas et al., 2009, 2012a, b). For reasons well discussed by Kues (2001) and by Lucas et al. (2009, 2011, 2012a), the mixed clastic-carbonate unit that is between the Gray Mesa and Bursum formations across much of central and southern New Mexico is termed either Atrasado Formation (most of the area) or Bar B Formation (in parts of Sierra County on the south).

LITHOSTRATIGRAPHY

General Overview

Prior to 1946, the rocks covered by this article were informally treated as subdivisions of larger formations or groups. In 1946, Vincent C. Kelley and Gordon H. Wood named the Atrasado Member of the

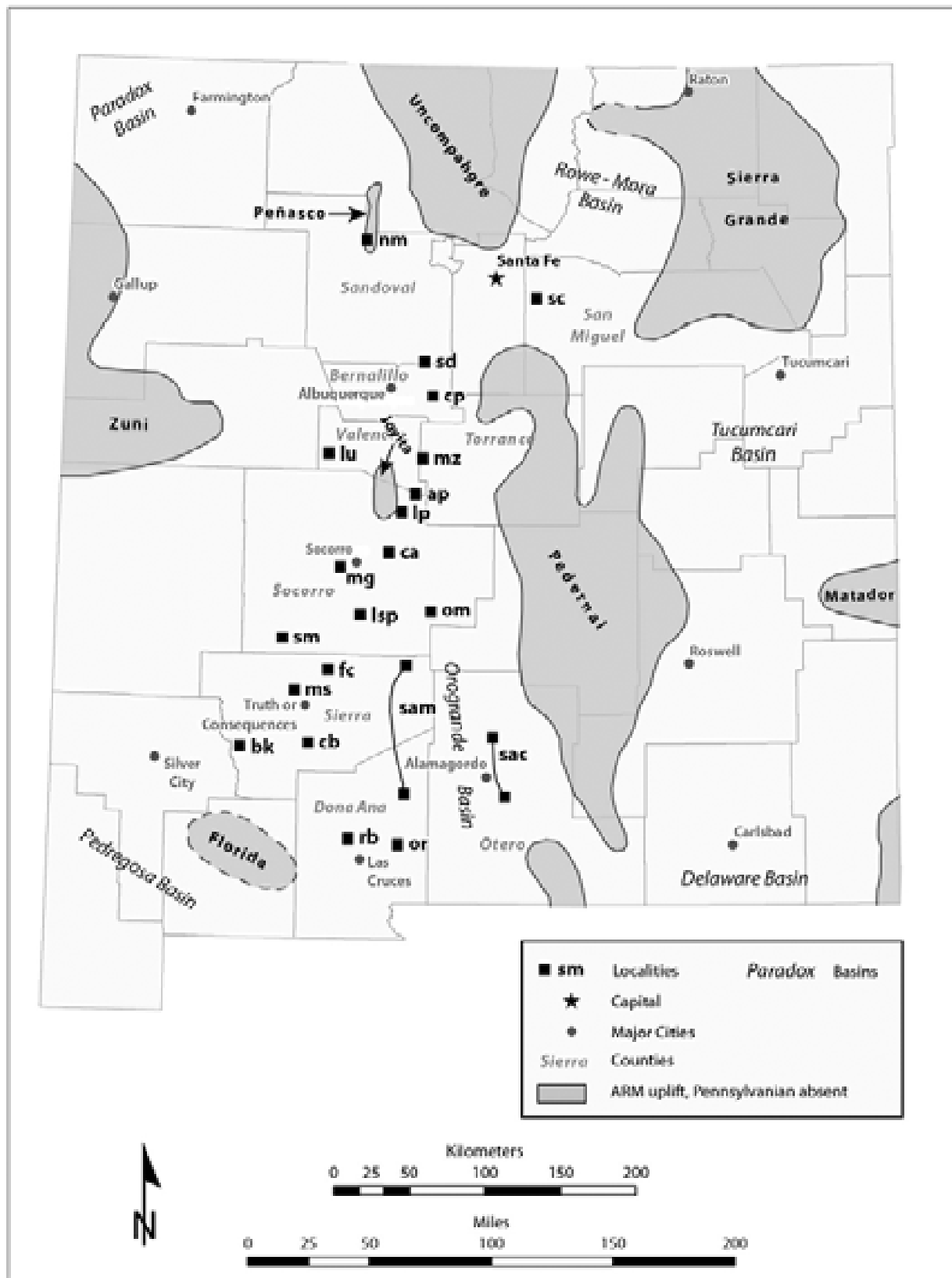


FIGURE 1. Map of New Mexico, showing Pennsylvania uplifts and localities discussed in this paper. **Key to localities:** **ap**, Abo Pass; **bk**, Black Range; **ca**, Cerros de Amado; **cb**, Caballo Range; **cp**, Cedro Peak; **fc**, Fra Cristobal Range; **lp**, Los Pinos Mountains; **lsp**, Little San Pasqual Mountains; **lu**, Lucero uplift; **mg**, Magdalena Mountains; **ms**, Mud Springs Mountains; **mz**, Manzano Mountains; **nm**, Nacimiento Mountains; **om**, Organ Mountains; **rb**, Robledo Mountains; **sac**, Sacramento Mountains; **sam**, San Andres Mountains; **sc**, Sangre de Cristo Mountains; **sd**, Sandia Mountains; **sm**, San Mateo Mountains.

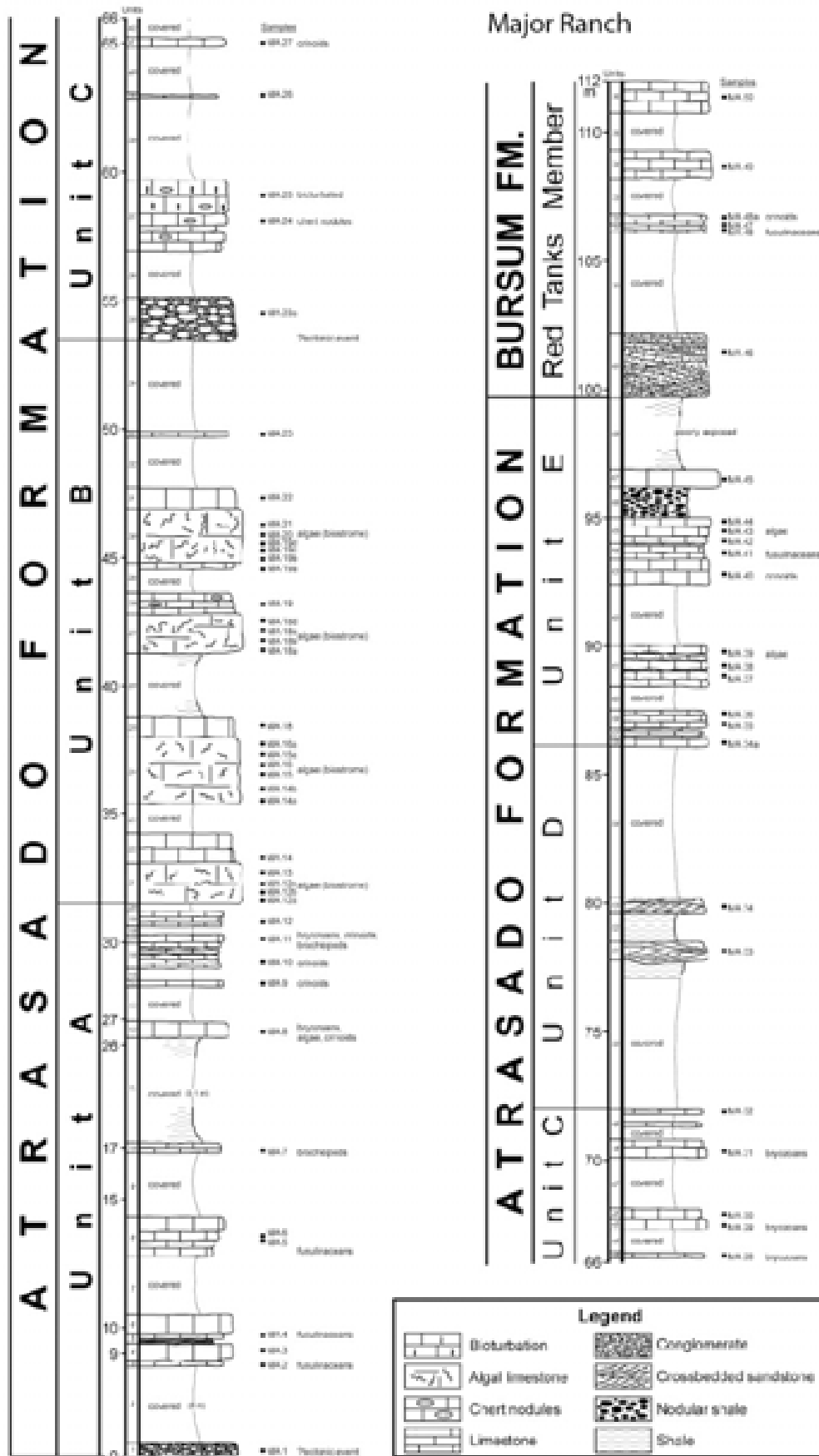


FIGURE 3. Graphic column of the type section of the Atrasado Formation from Major Ranch, on the Lucero uplift, Valencia County.

ingly, Lucas et al. measured a principal reference section of the Bar B Formation in nearby Caballo Canyon (NE $\frac{1}{4}$, Sec. 11, T15S, R4E), where the formation is 72 m thick (Fig. 4).

Extent and Thickness

Several names have been applied to rocks that are here considered to belong to the Atrasado and Bar B formations (Fig. 2). In the following paragraphs, Upper Pennsylvanian rocks are considered first from north to south on the west side of the Rio Grande, and then from north to south on the east side of the river. Thickness data are summarized (Table 1) with reference to the map (Fig. 1).

The Atrasado Formation crops out in mountain ranges bordering the Rio Grande Valley from Bernalillo County on the north to the Mud Springs Mountains in Sierra County on the south. The Bar B Formation, as presently understood, is confined to the Fra Cristobal and Caballo Ranges of Sierra County. Thickness patterns are difficult to ascertain because of a paucity of reliable, accurately located control points. In areas such as the Caballo Mountains, different authors have reported thicknesses that differ by a factor of two to three. Some of this variation may be real and related to tectonic activity during Late Pennsylvanian sedimentation. The Atrasado is entirely absent at the southern end of the Joyita Hills of Socorro County, whereas Virgilian strata are missing in part of the southern Caballo Mountains.

West of Rio Grande

Beginning in the Nacimiento Mountains on the northwest, Woodward (1987) used the old name “arkosic member of the Madera Formation” for the rocks of interest. Huffman and Condon (1993) used the same term in their discussion of the adjacent southeastern San Juan basin, in the subsurface. Woodward provided no thickness estimate, although he indicated that the entire Madera Group ranges upward to about 530 m. According to Woodward, the arkosic member contains abundant fresh, pink feldspar up to the size of small pebbles 2.5 cm across. In the upper part of the formation, the content of arkose increases at the expense of limestone. Krainer et al. (2005) named the Upper Pennsylvanian stratigraphic interval in the Jemez and Nacimiento Mountains of Sandoval County the Guadalupe Box Formation.

Southward through the Atrasado type area on the Lucero uplift, Kottlowski (1960) reported the Atrasado to be 232 m thick on the north side of Ladron Peak. South of the peak, the Pennsylvanian section thickens substantially. Allowing for faults and some leeway in the choice of contacts on Kottlowski’s graphic section, the Atrasado may be 465 m thick here. However, as noted above, the type section of the Atrasado Formation, at Major Ranch well north of Ladron Peak, is much thinner, only about 100 m thick.

In the Magdalena Mountains of Socorro County, the Pennsylvanian has been chopped up by faults, while soil and vegetation conceal many details. Kottlowski (1960) stated that bedded and massive limestone of the lower part of the Madera Formation is about 180 m thick near the ghost town of Kelley, while the total Madera may be 300 to 450 m thick. If the limestone near Kelley is assumed to be Gray Mesa, the Atrasado may be only 120 to 270 m thick; but most of the range of values is due to observational uncertainty.

The next Pennsylvanian outcrops southward are in another area we have not visited, the southeastern foothills of the San Mateo Mountains in southwestern Socorro County. Kottlowski’s (1960) graphic column indicates a total Pennsylvanian thickness of about 790 m. Although Kottlowski zoned these rocks to some degree using fusulinids, there is no lithologic basis to split out the Atrasado or Bar B Formation here. Faults and igneous intrusions add a further measure of uncertainty.

One of the finest Pennsylvanian sections in New Mexico is at Whiskey Canyon in the Mud Springs Mountains, northwestern Sierra County. Thompson (1942) named several formations here and Lucas et al. (2012a) logged the entire succession. The latter authors described the

Atrasado Formation as being about 187 m thick, and they divided the formation into six members. On the basis of overall lithology, the name Atrasado is chosen here over Bar B, which was previously used by Maxwell and Oakman (1990) and by Kues (2001).

Near Kingston in the Black Range to the southwest, Kuellmer (1954) described a Pennsylvanian succession but did not designate formations. Based on his description, the basal 15 m of shale with interbeds of limestone and sandstone is a good fit for the Red House Formation. The next 105 m consists of alternating ledges of massive, cherty limestone and recessive intervals of thinner bedded, shaly limestone and calcareous shale, which fits the description of the Gray Mesa Formation. Topping the Pennsylvanian is a ledge- and slope-forming interval 78 to 86 m thick that is a good fit for the Bar B Formation. Lithologies include nodular, shaly to silty limestone that commonly weathers yellowish brown; shale that is gray to yellow and contains nodules of fossiliferous limestone; and conglomeratic lenses of rounded limestone pebbles in a carbonate matrix.

Continuing southward, the next reasonably complete Pennsylvanian section is in the northern Robledo Mountains of Doña Ana County. These rocks have never been divided into formations. Measuring this section, we found that the Upper Pennsylvanian is nothing like typical Bar B or Atrasado. It is dominantly limestone, punctuated by numerous cliffs of massive, algal rock. The overall aspect is much closer to the Horquilla Formation of southwestern New Mexico than to the Bar B or Atrasado.

East of Rio Grande

Upper Pennsylvanian rocks in the southern Sangre de Cristo Mountains have been assigned to the Alamitos Formation (Miller et al., 1963; Baltz and Myers, 1999). These are strata deposited in the Taos trough (Rowe-Mora basin of some authors), which is a different depositional basin than the one(s) in which the Atrasado and Bar B formations were deposited. In general, the Alamitos is composed of coarse, arkosic sandstone and conglomerate, with lesser proportions of siltstone, shale, and limestone (Baltz and Myers, 1999). Although the Atrasado contains a similar mix of lithologies, that name has never been used in the Sangre de Cristo Mountains, and we do not propose to introduce it.

In the Sandia Mountains east of Albuquerque, the name Atrasado appears appropriate for the “arkosic limestone member of the Madera Formation,” as described by Read et al. (1944) and Kelley and Northrop (1975). Thickness is reported to be in the range of 238 to 267 m. Numerous faults, the wide extent of outcrops down the dip slope, and extensive soil and vegetation cover impede development of a detailed section, although work on this is underway.

Myers (1973) introduced the name Wild Cow Formation for Upper Pennsylvanian rocks in the Manzano and Manzanita Mountains. However, the Wild Cow is essentially the same lithology as the Atrasado Formation, which has priority of naming (e.g., Kues, 2001). The best exposures of the Atrasado Formation here occur on Cedro Peak in the Manzanita Mountains (Fig. 5) (Vachard et al., 2012, 2013), and at Priest Canyon (Fig. 6) and Abo Pass near the southern end of the Manzano Mountains. Thickness is 190 to 220 m in most of this area, but 272 m of Atrasado has been measured at Priest Canyon. As will be outlined below, eight named members can be identified in these sections.

An abnormally thin Upper Pennsylvanian succession occurs in the Joyita Hills (also known as Los Cañoncitos), about 25 km north of Socorro. The Atrasado Formation here is truncated with an angular unconformity beneath the Bursum Formation. The thickest Atrasado section is less than 100 m, and the Bursum lies directly on Precambrian granite at the south end of the range. Clearly, this area underwent tectonic uplift during Late Pennsylvanian time (Kottlowski and Stewart, 1970).

Pennsylvanian rocks are well displayed in the hills east of Socorro (Fig. 7). The Atrasado measures about 290 m thick in this area. Eight members have been identified, of which three were named here (Lucas et al., 2009).

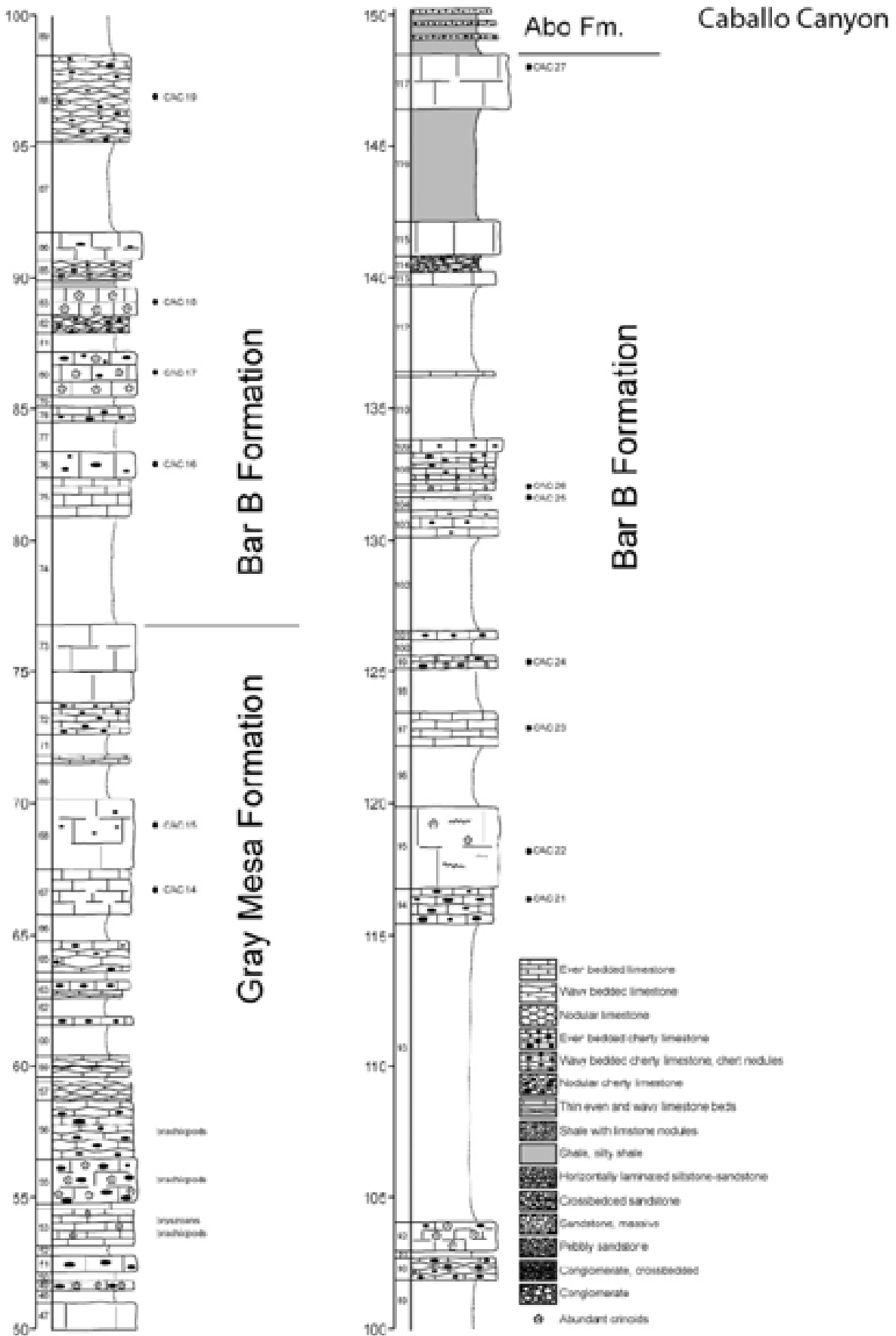


FIGURE 4. Graphic column of the type section of the Bar B Formation from Caballo Canyon in the Caballo Mountains, Sierra County.

TABLE 1. Previous usage, thickness range, and age of Atrasado and Bar B formations in various localities. **Bar B Formation (bold face)** denotes areas where this name is presently used.

Locality	Name Previously Used	Thickness (meters)	Age Range, if known
Nacimiento Mts.	Arkosic member of Madera Fm.	No accurate data	Desmoinesian through Virgilian
Lucero Mesa	Atrasado Member of Madera Fm.	110-275	Late Desmoinesian through Virgilian
Sierra Ladrones	Atrasado Member of Madera Fm.	232-465, thicker S.	Missourian and Virgilian
Magdalena Mts.	Madera Formation, upper part	120 to 270?	Missourian and Virgilian
Whiskey Canyon	Bar B Formation	187a	Late Desmoinesian through Virgilian
Kingston area, Black Range	Magdalena Formation, undivided	78 to 86	Not determined
Sandia Mts.	Arkosic limestone member of Madera Fm.	238-267	Not determined
Manzanita and Manzano Mts.	Wild Cow Formation	190 to 272	Late Desmoinesian through Virgilian
Joyita Hills	Undivided	0 to 100a	Missourian
Socorro area	Arkosic limestone member of Madera Fm.	290a	Late Desmoinesian through middle Virgilian
Northern Oscura Mts.	Arkosic limestone member of Madera Formation	120 to 250	Early Missourian through middle Virgilian
Southern Oscura Mts.	Upper member of Madera Formation	250 to 480	Missourian through very early Wolfcampian(?)
San Andres Mts.	Panther Seep (in part)	444 to 728, thicker S.	Late Desmoinesian through Virgilian
Fra Cristobal Mts.	Bar B Formation	81 to 107	Middle Desmoinesian through Virgilian
Caballo Mountains	Bar B Formation	72 to 96, possibly greater	Middle Desmoinesian through Virgilian, but Virgilian locally absent

Southward, near Bruton Canyon in the northern Oscura Mountains, the Atrasado Formation is about 121 m thick (Lucas and Krainer, 2009). The upper part of the formation matches well with that in the Socorro area, but the lower Atrasado is much thinner in the northern Oscuras and contains much more limestone than shale (Fig. 8). Bachman (1968) reported that the “upper member of the Madera Formation” rapidly thickened southward, reaching 480 m in the southern part of the Oscuras. A substantial increase in the proportion of shale and sandstone accompanies the southward thickening and at some point these strata should be referred to the relatively thick and shaley Panther Seep Formation, best exposed farther south in the San Andres Mountains. Bachman’s “upper Madera” clearly is in part Atrasado, but based on his detailed graphic columns, division into members is problematic. Bachman stated that most of the limestone beds lacked continuity.

In the Caballo Mountains farther southwest, Kelley and Silver (1952) reported that the Bar B Formation is 103 m thick at the type section, increasing in thickness to the south and east (no figures given). However, as noted above, the Bar B type section is actually Gray Mesa Formation. Nearby at Bar B Draw, the exposed thickness of the Bar B Formation is 63 m. The succession is composed of alternating nodular, fossiliferous limestone that is partly cherty, with gray fossiliferous limestone beds up to 1 m thick and limestone intervals up to 3 m thick having thin to medium, even to wavy bedding. Separating limestone layers are covered or poorly exposed intervals as thick as 6.5 m, some of which contain reddish to purple shale having limestone nodules up to 10 cm in diameter. Limestone containing the fusulinid *Triticites ventricosus* is in the upper part of the section.

Elsewhere in the Caballo Mountains, Seager and Mack (2003) reported thicknesses of 93 m in the Derry Hills, 215 m in the McLeod Hills, and 280 m on the eastern flank of the Caballo Mountains. Soreghan

(1992, 1994) quoted a thickness of 260 m in the southern Caballos. While we have not visited all of these localities, the Bar B measures 72 m thick at Caballo Canyon in the central part of the range (Fig. 4), and 96 m at Green Canyon in the southern part (Lucas et al., 2012b). Part, but not all of the extra thickness reported by prior authors represents the overlying Bursum Formation, which we have logged separately from the Bar B.

Cserna (1956) and McCleary (1960) reported Bar B thicknesses of 81 and 79 m, respectively, in the Fra Cristobal Mountains. These figures compare with 107 m at a section in Hellion Canyon on the western flank of the range (Fig. 9; Lucas et al., 2012b). As in the Caballo Mountains, the Bar B Formation here generally erodes to rounded slopes, from which discontinuous ledges of more massive limestone protrude (Fig. 10). A large part of the succession is composed of limestone in thin, tabular to wavy beds separated by layers of calcareous shale (Fig. 11).

In the San Andres Mountains, Kottowski et al. (1956) partitioned Pennsylvanian rocks into Derry, Des Moines, Missouri, and Virgil Series on the basis of fusulinids. For the most part, these authors did not apply formation names. They named one new formation, the Panther Seep Formation of Virgilian age. They stated that the Panther Seep and Bar B formations are dissimilar, although they did not explain the differences. It appears that the following are the chief distinctions:

1. The Panther Seep is much thicker than the Bar B or Atrasado. From 444 m at Rhodes Canyon in the northern San Andres Mountains, the Panther Seep thickens to 728 m at Ash Canyon, in the southern part of the range.

2. The Panther Seep contains a larger portion of silty shale, siltstone, and fine-grained sandstone than the Bar B or Atrasado.

3. Gypsum, which is rare in the Atrasado and Bar B, is an important constituent of the Panther Seep.

The Panther Seep Formation also has been mapped in the Organ Mountains, south of the San Andres Range in Doña Ana County (Bachman and Myers, 1969; Seager, 1981). Thicknesses here are estimated at 600 to more than 800 m. Deposition of the relatively thick and clastic Panther Seep Formation took place during rapid subsidence along the subsiding keel of the Orogrande basin (e.g., Schoderbeck, 1994).

A lateral transition between the Atrasado, Bar B, and Panther Seep formations presumably exists somewhere in the southern Oscura or northern San Andres Mountains (see above). Unfortunately, this area is part of the White Sands Missile Range, where scientific access is currently severely restricted.

Upper Pennsylvanian rocks of the Sacramento Mountains to the east have been classified in the Beeman (older) and Holder formations. Lithologies include gray to red shale and mudstone, sandstone, conglomerate, and various types of limestone (Otté, 1959; Pray, 1961). Although similarities to the Atrasado are evident, we do not propose to change the long-established terminology here.

Members

Lucas and Krainer (2009) and Lucas et al. (2009) divided the Atrasado Formation of central New Mexico into members. Among these, the upper five originally were named as formations by Thompson (1942) based on exposures in the northern Oscura Mountains. The Bartolo and Amado members have been formalized after original use by Rejas (1965) in a mapping thesis in the Cerros de Amado area, east of Socorro. Lucas et al. (2009) newly named the Tinajas Member, also based on outcrops near Cerros de Amado. These members have been recognized as far north as Cedro Peak in the Manzanita Mountains (Lucas et al., 2011), and with less confidence on the Lucero uplift to the west. Several members are readily identified in a column (Fig. 12) from Mesa Sarca in the Lucero uplift (Wengerd, 1959, 1975), but not in our column from the type Atrasado at Major Ranch nearby (Fig. 3). Several of them also have been identified at Whiskey Canyon in the Mud Springs Mountains of Sierra County (Lucas et al., 2012a).

Thompson’s Adobe Formation also has been retained as the Adobe

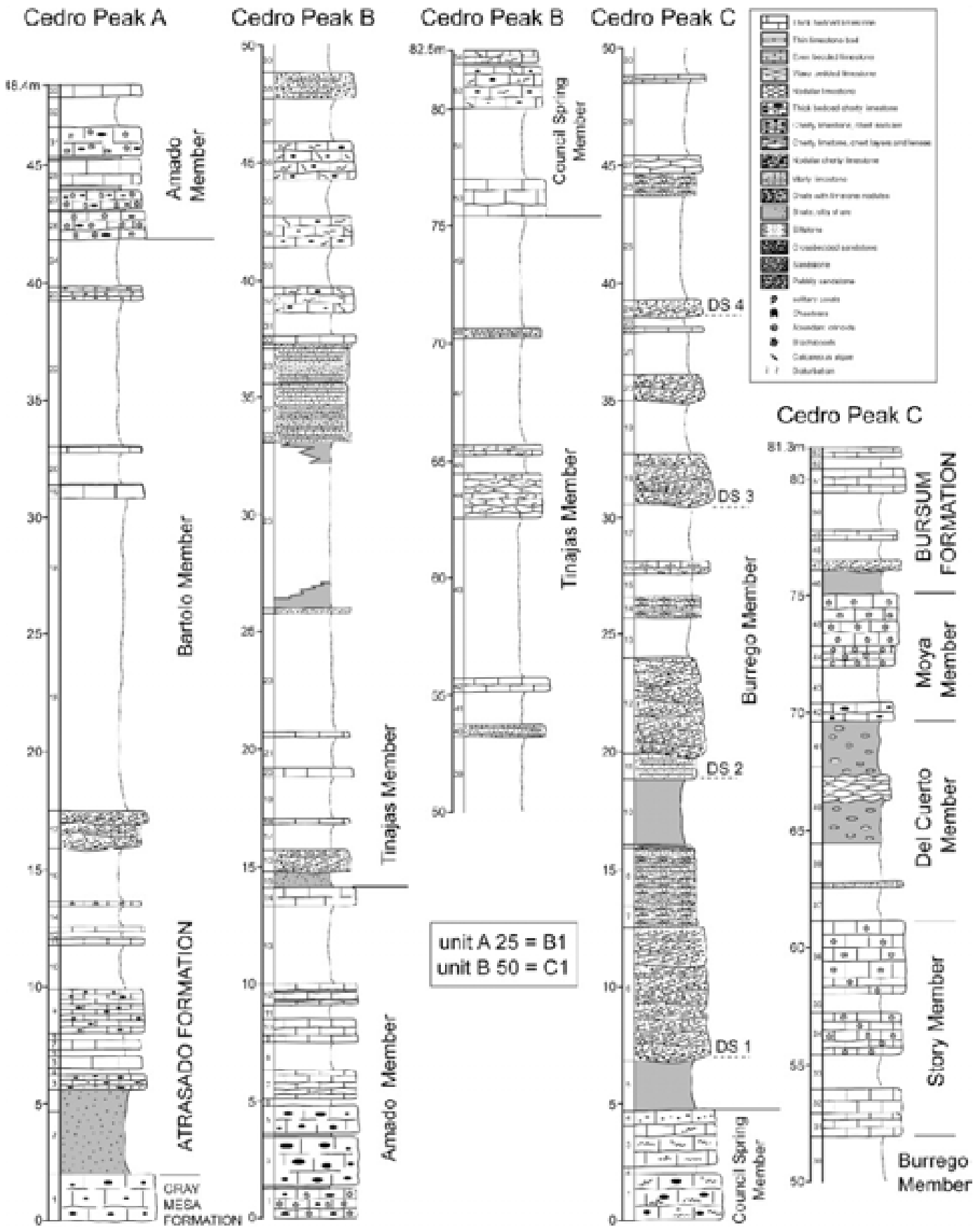


FIGURE 5. Graphic column of the Atrasado Formation from Cedro Peak, Bernalillo County.

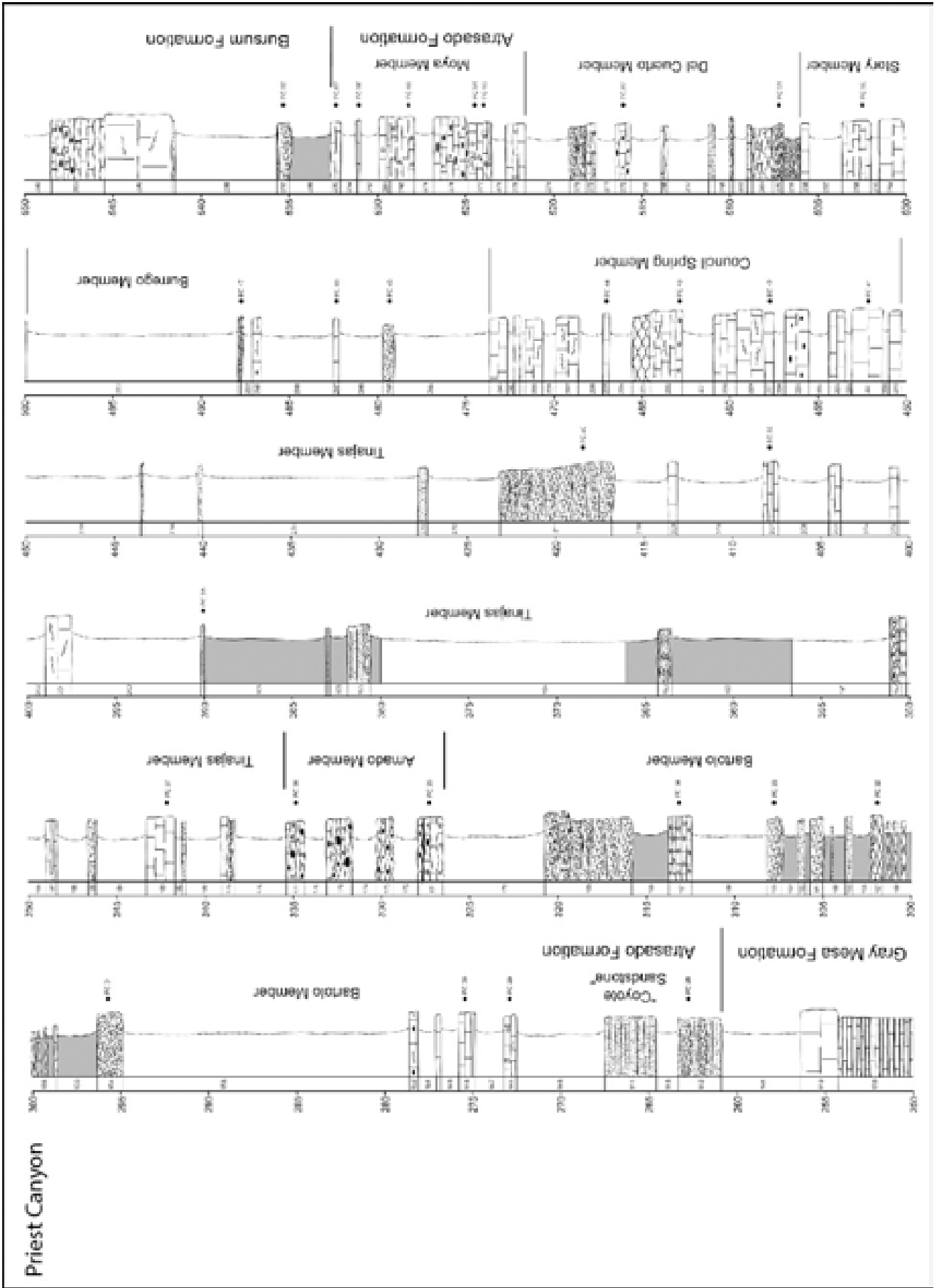


FIGURE 6. Graphic column of the Atrasado Formation from Priest Canyon in the Manzano Mountains, Torrance County.

Cerros de Amado

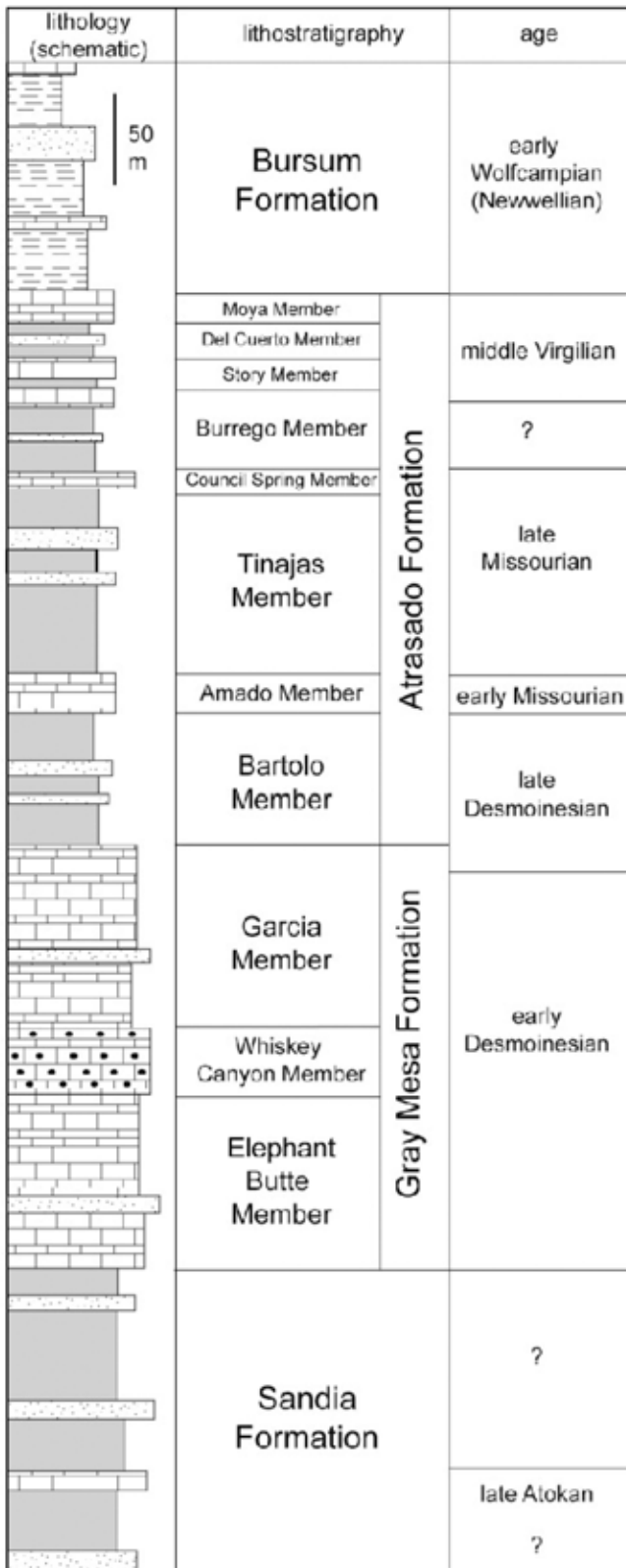


FIGURE 7. Simplified Pennsylvanian stratigraphic column from Cerros de Amado, Socorro County, illustrating division into members and biostratigraphic age assignments.

Oscura Mountains

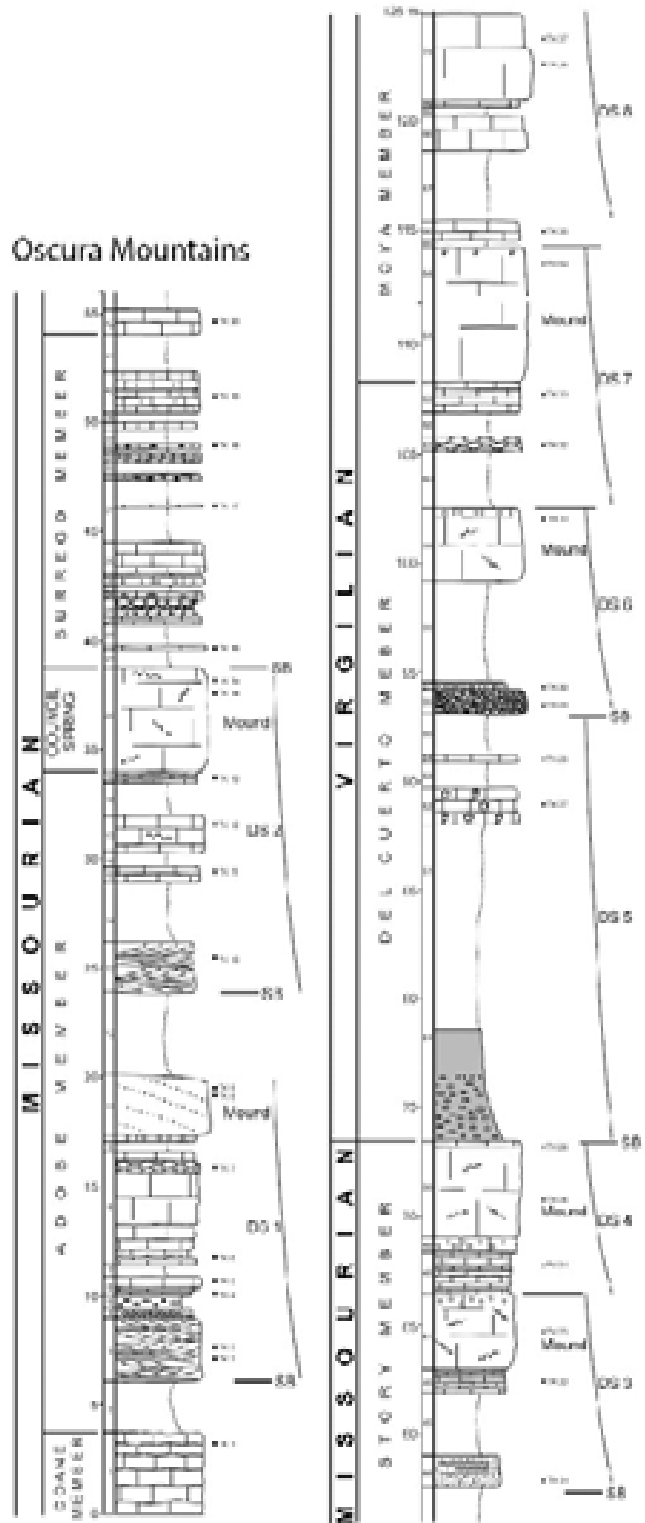


FIGURE 8. Graphic column of the Atrasado Formation from Bruton Canyon in the northern Oscura Mountains, Socorro County.

Hellion Canyon

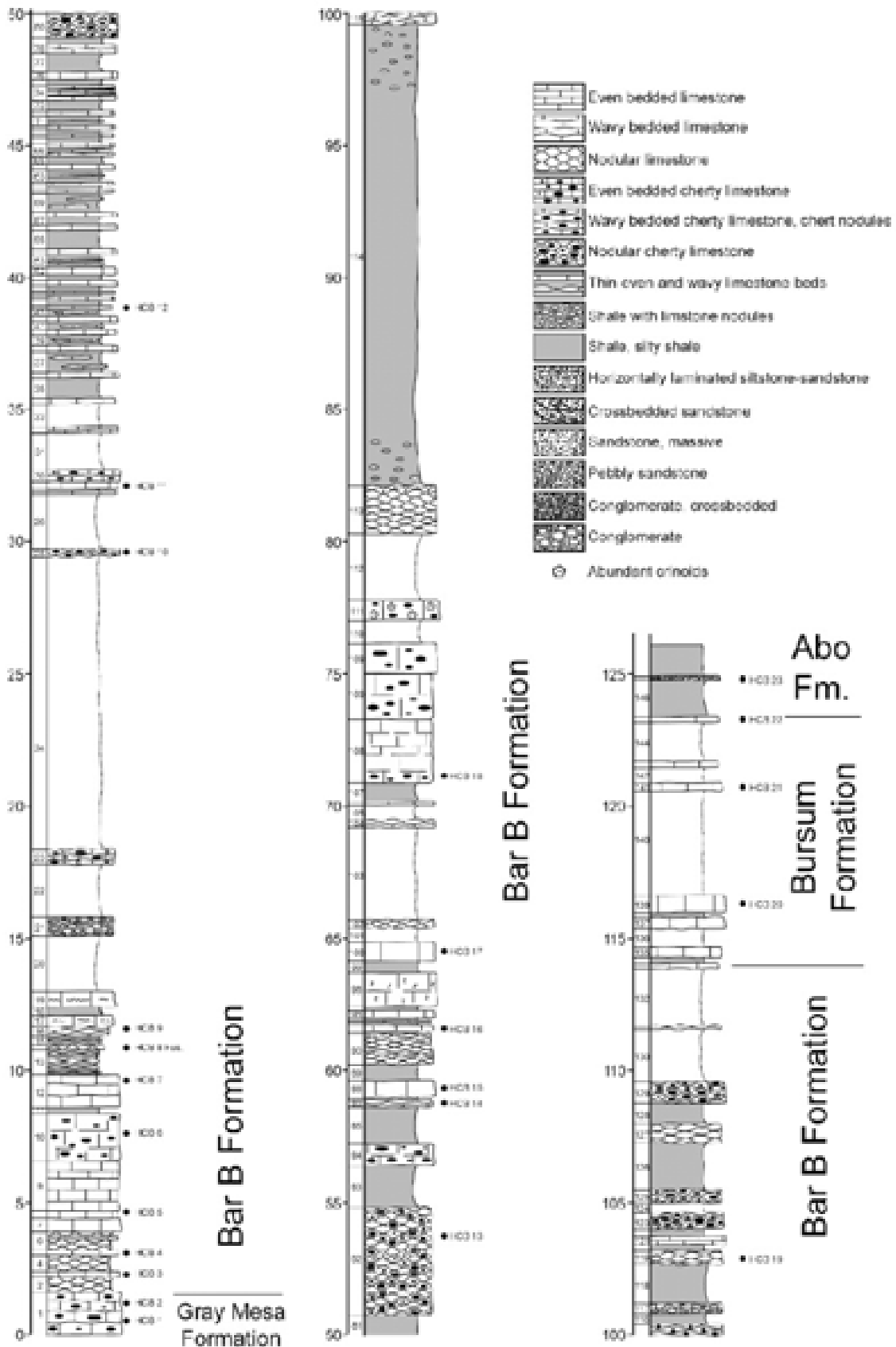


FIGURE 9. Graphic column of the Bar B Formation from Hellion Canyon locality "B" in the Fra Cristobal Mountains, Sierra County.



FIGURE 10. Photograph looking south across a branch of Hellion Canyon, illustrating the Bar B and enclosing units.

Member (Lucas and Krainer, 2009), although this unit is not known to occur outside its type area in the Oscura Mountains (Fig. 8). The Adobe is a relatively thin unit of interbedded carbonate and clastic rocks, equivalent to the combined Bartolo, Amado, and Tinajas Members in other areas. Thompson's Coane Formation, below the Adobe, is considered to be part of the Gray Mesa Formation, but is not distinctive enough to be differentiated as a member (Lucas and Krainer, 2009). The Bolander Group of Thompson has been abandoned (Lucas et al, 2012a).

Four of the members-- Amado, Council Spring, Story, and Moya--are dominantly thick-bedded to massive limestone that erodes to bold ledges and cliffs. These bracket the Bartolo, Tinajas, Burrego, and Del Cuerto members, which exhibit diverse, laterally variable rock types (including significant intervals of clastic rocks) and generally erode to slopes and valleys (Figs. 13, 15). As such, these members can be readily mapped on the ground and from aerial images, and should be distinctive in various types of well logs. In a region of lower topographic relief and less structural complexity, such units might be classified as formations.

Columnar sections (Figs. 5, 6, 7, and 8) exhibit characteristics of the members of the Atrasado Formation at several localities. The chart below (Table 2) summarizes characteristics of these units.

Petrography and Microfacies

Limestone of the Atrasado Formation is dominated by microfacies with a muddy texture, wackestone being the most abundant microfacies type. Varieties include skeletal wackestone, foraminiferal wackestone,



FIGURE 11. Photograph of lower part of Bar B Formation from the small draw shown near the center of Fig. 10, showing the common lithology of interbedded shale and limestone. Staff is 5 feet (1.5 m) long.

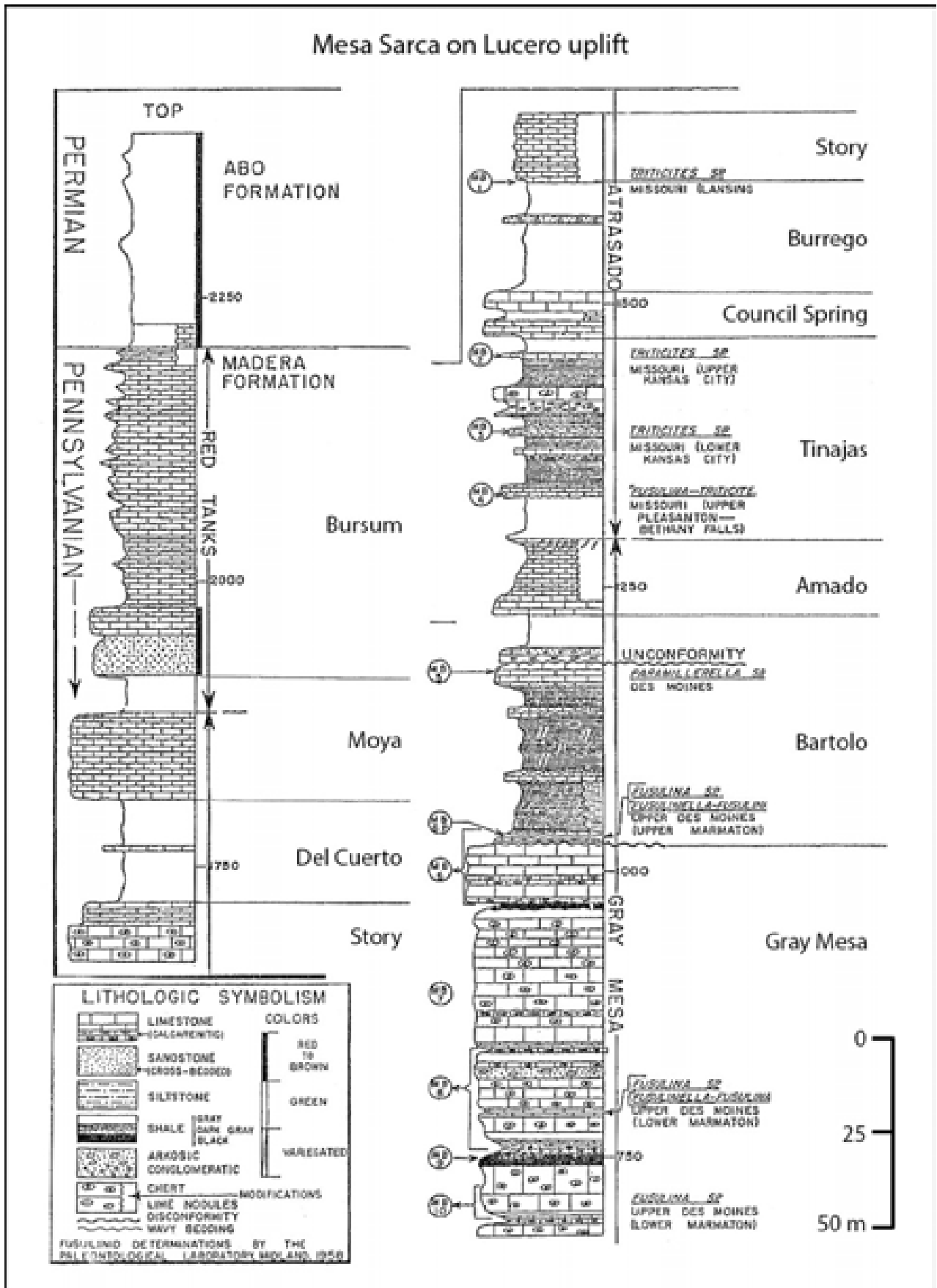


FIGURE 12. Column from Wengert (1959, 1975) on Mesa Sarca (Lucero uplift, Valencia County) modified to show probable division into named members.



FIGURE 13. Photograph of type section of the Bartolo Member of the Atrasado Formation near Cerros de Amado, Socorro County. The member is largely non-resistant shale and siltstone, having occasional sandstone and limestone layers that form small ledges. Capping the skyline is the Amado Limestone Member, with a massive algal bioherm (arrow).

fusulinid wackestone, crinoidal wackestone, algal wackestone, and rare peloidal wackestone. Subordinate are carbonate mudstones (such as skeletal mudstone, peloidal mudstone, and rare ostracodal mudstone); algal and bryozoan floatstone; and skeletal and crinoidal packstone. Grainstone is less common and is largely bioclastic, rarely peloidal. Most microfacies types contain a diverse fossil assemblage.

Phylloid algal wackestone, floatstone, and bafflestone are common in the Council Spring, Story and Moya members, locally forming algal mounds (Fig. 15).

Sandstones of the Atrasado Formation are classified as arkose to subarkose. Monocrystalline quartz is abundant; while different types of polycrystalline quartz, including coarse-grained magmatic, fine-grained metamorphic, and schistose metamorphic varieties are less common. The sandstone contains abundant detrital feldspar, including slightly altered potassium feldspar and mostly untwinned and perthitic plagioclase. Some feldspar grains are almost completely altered to clay minerals, partly replaced by carbonate. Other grain types include granitic rock fragments, very rare fine-grained phyllitic rock fragments of quartz and mica, and rare detrital muscovite and myrmekitic feldspars. Bluish tourmaline, apatite, zircon and opaque minerals occur as accessory grains. Sandstone is cemented by quartz that occurs as authigenic overgrowths, and by coarse, blocky carbonate cement.

Dominant microfacies of limestone of the Bar B Formation are

skeletal mudstone and bioclastic wackestone, together with subordinate packstone and rare grainstone (Lucas et al. 2012a).

Fossils

Most microfacies types contain a diverse fossil assemblage, which is similar to that of the Gray Mesa Formation (see Krainer and Lucas, 2004; Lucas and Krainer, 2009; Lucas et al., 2009, 2011, 2012a; Vachard et al., 2012, 2013). The most common fossils are skeletons of echinoderms (mostly crinoid stem fragments), brachiopods, bivalves and bryozoans (fistuliporid, trepostome and fenestrate species). Locally fusulinids (fusulinid wackestone) and smaller foraminifers (mainly *Calcivertella*) are abundant (foraminiferal wackestone). Smaller foraminifers are present in most microfacies types and include species of *Bradyina*, *Calcivertella*, *Calcitornella*, *Climacammina*, *Cornuspira*, *Deckerella*, *Earlandia*, *Eotuberitina*, *Endothyra*, *Globivalvulina*, *Glomospira*, *Hedraites*, *Hemigordius*, *Nankinella*, *Palaeonubecularia*, *Planoendothyra*, *Polytaxis*, *Pseudojanischewskina*, *Reitlingerina*, *Schubertellina*, *Spireitlina*, *Syzranella*, *Syzrania*, *Tetrataxis*, and *Tuberitina*.

Fusulinids are dominantly species of *Triticites*, rarely *Nankinella* and *Schubertella*.

Ostracods are present in all studied samples. Calcareous algae are mostly recrystallized fragments of phylloid algae, rare *Epimastopora*, *Paraepimastopora*, *Eugonophyllum*, *Ivanovia*, *Komia*, and *Efluegelia*.



FIGURE 14. Photograph of the type section of the Tinajas Member of the Atrasado Formation near Cerros de Amado. The lower part of the Tinajas is out of view, behind the photographer. The black, lacustrine shale unit of Lerner et al. (2009) is at the base of the ridge, the lowest ledge in view. Capping the hill is the Council Spring Limestone Member. The hill is about 50 m high.

Rare are fragments of small gastropods, trilobites, brachiopod spines, echinoid spines, calcified spicules, fragments of calcisponges and *Tubiphytes*, and the alga *Archaeolithoporella lamellosum* (locally encrusting other skeletons). Locally fecal pellets occur. In the Cerros de Amado conodonts were recovered from limestones of the Atrasado Formation (Lucas et al., 2009). In the Joyita Hills, *Chaetetes* and *Syringopora* and plant fossils (*Neuropteris*) occur in the Bartolo Member, *Dunbarella* in black shale and *Calamites* in sandstone of the Tinajas Member, and solitary corals in the Moya Member (Lucas et al., 2009). In the Whiskey Canyon section, solitary corals occur in the Adobe Member (Lucas et al., 2012a).

Of interest is the “Tinajas locality” in the Cerros de Amado east of Socorro, where lacustrine deposits (black shale) within the Tinajas member contain a diverse fossil assemblage of plants, crustaceans, insects, a cnidarian, mollusks, fishes, abundant conchostracans and freshwater ostracods. The overlying gray shale is rich in *Dunbarella* (Lerner et al. 2009). The Kinney Brick Quarry Lagerstätte, a world famous locality for Late Pennsylvanian fossil plants, invertebrates and vertebrates, is also located within the Tinajas Member (see Lucas et al., 2011 and references therein).

The fossil assemblage of the Bar B is less diverse compared to that of the Atrasado Formation. Common fossils are echinoderm fragments (mainly derived from crinoids), brachiopods, ostracods, gastropods, bryo-

zoans, smaller foraminiferans (*Bradyina*, *Calcitornella*, *Diplosphaerina*, *Endothyra*, *Eostaffella*, *Globivalvulina*, *Hemigordius*, *Syzrania*), fusulinids (*Triticites*) and locally recrystallized phylloid algae (e.g., Lucas et al., 2012a).

Contacts

The contact between the Atrasado/Bar B and underlying Gray Mesa Formation was discussed in the previous chapter, so that material will not be repeated here.

At most localities, the Bursum Formation (see following chapter) overlies the Atrasado and Bar B formations. For mapping purposes, the contact is generally picked at the top of the highest marine limestone overlain by arkosic conglomerate or pastel variegated, non-fissile mudstone typical of the Bursum. In some cases, an arbitrary choice presents itself, because the upper Atrasado contains variegated mudstone and the Bursum contains layers of marine limestone. Various authors have reported the contact to be gradational, transitional, or erosive and unconformable.

An angular unconformity separates the Bursum Formation from older rocks on the Joyita uplift in northern Socorro County. The Bursum truncates the entire older Pennsylvanian succession, resting directly on Precambrian granitic rock at the southern end of the uplift (Kottlowski and Stewart, 1970; Krainer and Lucas, 2009). Southward in the Cerros de



FIGURE 15. Photograph of an algal bioherm in the Council Spring Member, the subject of an article by Hambleton (1962). Height of the cliff is 7 to 8 m.

Amado area, the Bursum variably rests with erosive contact on the Moya, Del Cuerto, and Story members of the Atrasado. There is, however, no angular discordance of bedding in this area, and relief observed in individual outcrops is slight. Conodont biostratigraphy indicates that part of the Missourian and all of the Virgilian is missing between the Bar B and Bursum formations at Green Canyon in the southern Caballo Mountains of Sierra County (Barrick et al., 2012).

The Abo Formation directly overlies the Atrasado or Bar B at a few localities. One such site is Bar B Draw on the eastern flank of the Caballo Mountains. Within the Caballos, the Bursum varies in thickness from 0 to 80 m (Lucas et al., 2012b). A direct Abo to Bar B contact presumably is unconformable, although erosion is not necessarily obvious in the field.

In summary, the Bursum Formation reflects the onset of increased tectonic activity in several areas of New Mexico. As a result, sedimentation across the Atrasado/Bar B and Bursum contact was interrupted in some areas, but was more or less continuous in others.

Age

Fusulinid biostratigraphy has been applied to Pennsylvanian rocks in New Mexico since the 1940s (e.g., Thompson, 1942). Recently, conodont biostratigraphy also has been applied. These tools indicate that the Atrasado and Bar B Formations range in age from middle Desmoinesian, at the oldest, through late Virgilian, at the youngest. Ages recorded at various localities differ because of diachronous lateral facies shifts and local hiatuses in deposition.

The lower part of the Atrasado and Bar B formations are Desmoinesian age at most localities where good biostratigraphic control is available. Specifically, at Cerros de Amado (Fig. 7), the Desmoinesian-Missourian boundary lies in the upper part of the Bartolo Member (Lucas et al., 2009). The lower Atrasado also is Desmoinesian northward at Mesa Sarca (Fig. 12; Wengerd, 1959, 1975) and southward at Whiskey Canyon (Lucas et al., 2012a). The lower Bar B Formation at Green Canyon in the Caballo Mountains yielded conodonts associated with the middle Desmoinesian Verdigris cycle of the American Midcontinent (Barrick et al., 2012), an age assignment also supported by fusulinid biostratigraphy (Lawton et al., 2002).

An exception has been found in the northern Oscura Mountains, where the upper part of the Gray Mesa Formation contains early Missourian fusulinids (Thompson, 1942; Lucas and Krainer, 2009). The lower part of the Atrasado here is greatly thinned and dominantly limestone rather than the usual shale. These findings suggest that this area underwent less subsidence than normal during this time, while being remote from sources of terrigenous clastics. As a result, carbonate precipitation prevailed, creating a lateral facies shift.

The upper Atrasado is early to mid Virgilian in the northern Oscura Mountains (Thompson, 1942; Lucas and Krainer, 2009), Cerros de Amado area (Lucas et al., 2009), and the Manzano and Manzanita Mountains (Myers and McKay, 1976; Lucas et al., 2011). As noted in the discussion of Contacts, Virgilian strata are missing in the southern Caballo Mountains (Lucas et al., 2012a). The only report of possible Wolfcampian age is from the Mockingbird Gap area of southeastern Socorro County

TABLE 2. Characteristics of members of the Atrasado Formation, based on sections from Cedro Peak to Oscura Mountains.

Member Name	Thickness (m)	Description
Moya Limestone	0 to 16	Largely light gray, thick-bedded to massive algal limestone that forms cliffs; subordinate wavy-bedded and nodular limestone. Widely eroded at the base of the Bursum Formation.
Del Cuerno	25 to 33	Bedded and nodular, fossiliferous limestone; variegated mudstone, gray to green fissile shale and siltstone, fine to coarse arkosic sandstone, and conglomerate or rounded chert and quartz pebbles. Much lateral variation. Erodes to slopes and small ledges.
Story Limestone	4 to 18	Cliff-forming, light gray, massive algal limestone that is siliceous but not cherty. As originally defined, the Story included shale and sandstone underlying the cliff-forming limestone. For the sake of consistency and ease of mapping, these strata should be transferred to the Barrego Member.
Barrego	16 to 49	Bedded and nodular, fossiliferous limestone exhibits a variety of textures. Gray to reddish brown, arkosic sandstone becomes more prevalent northward. Red to gray mudstone with limestone nodules and gray to green, fissile shale and sandstone also are present. Rapid lateral changes in thickness and lithology are common. This unit erodes to slopes or a series of low ledges.
Council Spring Limestone	4 to 23; normally 5 to 10	Cliff-forming limestone much like the Story Member. Algal biostromes are commonly developed. Hamblenton (1962) described examples.
Tinajas	50 to 115	Largely shale and siltstone ranging from gray to greenish and olive-gray to black, partly calcareous, both marine fossils and plant fossils common. Limestone is mostly dark gray, micritic, and dolomitic. Fine to coarse, crossbedded arkose is common in the upper part.
Amado Limestone	3 to 20; 10 m is typical	Medium gray, cherry lime mudstone and wackestone in wavy beds, thin interbeds of gray, calcareous shale. Erodes to stepped ledges rather than a sheer cliff, except where algal bioherms are developed.
Bartolo	50 to 70	Thick intervals of slope-forming, olive- and greenish-gray silty shale and siltstone, punctuated by ledges of various types of limestone, arkosic sandstone, and conglomerate.
Adobe	30+	Gray fossiliferous limestone with algal bioherms, reddish-brown, crossbedded sandstone; poorly exposed shale and mudstone. Known only in northern Oscura Mountains.

(Bachman, 1968). This author cited Lloyd G. Henbest of the U.S. Geological Survey, who concluded that fusulinids collected 45 m below the base of the Bursum Formation could be very late Virgilian or very early Wolfcampian age. Because the international Pennsylvanian-Permian boundary subsequently was moved upward, these rocks are likely Pennsylvanian according to current definition.

DEPOSITIONAL ENVIRONMENTS

Diverse rock types of the Atrasado and Bar B formations relate to a variety of depositional settings. These are most easily assessed for carbonate rocks, virtually all of which are marine. Sparsely fossiliferous and non-fossiliferous lime mudstone and microgranular dolomite record quiet water conditions, either below wave base or (especially for dolomite, which suggests elevated salinity) in protected bays or lagoons. The common occurrence of diversely fossiliferous wackestone and packstone is evidence for slightly to moderately agitated, shallow marine shelves having near normal salinity. Scarce are well-sorted, cross-bedded or laminated grainstones and packstones of high-energy, shallow-water settings such as shoals, bars, and beaches. Algal biostromes, such as those constituting much of the Council Spring, Story, and Moya members, evidently are products of quiet water within the photic zone.

Coarse, poorly sorted arkose and conglomerate contain plentiful fresh, pink feldspar, indicating proximity to granitic source terranes and minimal weathering. These rocks are generally calcareous and grade to sandy marine limestone, implying energetic, near-shore marine conditions. Fluvial deposits are not common, but include some fluvio-deltaic strata in the Manzanita Mountains (Lorenz et al., 1992).

The laminated shale and siltstone that make up much of the Bar B and Atrasado probably represent near-shore, shallow to moderately deep settings such as bays. Thick shale units tend to have sparse and restricted faunas of bivalves, inarticulate brachiopods, and ostracods, suggesting rapid deposition in brackish water. Land plant fragments are common, together with occasional silicified logs. Other shale deposits are calcareous and contain normal marine faunas, indicating open shelf conditions. A black shale interval within the Tinajas Member near the Cerros de Amado in Socorro County contains a fossil flora and fauna consistent with lacustrine deposition (Lerner et al., 2009).

The only extensive, real nonmarine strata are the non-fissile, variegated mudstone intervals that occur mainly in the upper part of the Atrasado and Bar B. Their mottled appearance and plentiful carbonate nodules indicate soil formation under alternating wet-dry, oxidizing and reducing environments. Other types of paleosols, rooted zones, and features of subaerial exposure are seldom encountered in these formations.

Cyclicality

Much has been written about apparent cyclic sedimentation in Pennsylvanian rocks of New Mexico. Wiberg (1993), Wiberg and Smith (1994) and Scott and Elrick (2004) described cycles in the Gray Mesa Formation of the Albuquerque area. Algeo et al. (1991) and Algeo (1996) did the same for rocks of similar age in the Sacramento Mountains. Raatz et al. (1994) defined “cyclothems” in Missourian strata of the Sacramento Mountains. Soreghan (1992, 1994a, b) analyzed apparent cyclicality in the Bar B and correlative formations of southern New Mexico and southeastern Arizona.

Repetitive transgressive and regressive cycles are common in the Atrasado and Bar B formations. Describing these features at an outcrop is one thing, but assessing their significance is another. Few are the long, continuous sections where multiple shale and mudstone units are clearly visible. Isolated outcrop patterns, complex structure, and the difficulty of obtaining precise, detailed biostratigraphic information all impede correlation of cycles from one place to another. The situation in New Mexico contrasts with that in the Illinois and Midcontinent basins, where tens of thousands of borehole logs make up for the scarcity of outcrops.

Cyclothems comparable to those of the Midcontinent have not been observed in New Mexico. Seldom observed are the paleosols and incised valleys that mark sequence boundaries in the Midcontinent (e.g., Nelson et al., 2002). Such features mark episodes of marine withdrawal, when the land surface was weathered and eroded. Moreover, we have evidence for many hiatuses in the Pennsylvanian rock record of New Mexico, although these are seldom evident on the outcrop. For example, eight depositional cycles have been identified in the entire Atrasado Formation of the northern Oscura Mountains, whereas the time-equivalent succession in Kansas contains 74 such cycles (Lucas and Kainer, 2009). This observation suggests that the Oscura Mountains area lay relatively higher on the shelf than the Midcontinent, and thus contains only major cycles, lacking nearly all the cycles of lesser scale that are found only on the middle to lower part of the Midcontinent shelf (Heckel, 2008).

The Atrasado and Bar B contain plentiful evidence for tectonic activity concurrent with sedimentation. Among the lines of evidence are rapid lateral changes of thickness and lithofacies, hiatuses in the section, and frequent intercalation of arkose and conglomerate with marine shale and limestone. Apparently, tectonic activity severely masks and complicates the record of glacial eustasy.

Paleoclimate

Not a great deal can be inferred about climate from a dominantly marine succession. The local development of sabkha facies (Tinajas Member, near Cerros de Amado) demonstrates episodes of pronounced aridity, while growth rings in associated fossil wood indicate seasonal wet-dry conditions (Falcon-Lang et al., 2011). Bedded gypsum in Pan-

ther Seep strata indicates dry episodes in the southern Orogrande basin. Soreghan (1994b) inferred a seasonal, wet-dry climate in the Orogrande basin, with wetter and drier episodes linked to glacial cycles.

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In the Cerros de Amado, northeast of Socorro, the Upper Pennsylvanian (late Missourian) Council Spring Member of the Atrasado Formation is a light-colored (white/light gray) algal limestone that forms an excellent stratigraphic marker bed, capping local topography. Here, it overlies the slope-forming, mostly clastic Tinajas Member. The black shale at the base of the outcrop shown here is a lake bed in the Tinajas Member that yields numerous fossils of plants, crustaceans and fishes.

THE PENNSYLVANIAN-PERMIAN BURSUM FORMATION IN CENTRAL NEW MEXICO

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Abstract—In central New Mexico, the Bursum Formation (late Virgilian-early Wolfcampian) represents the transitional facies between the underlying, dominantly shallow marine Pennsylvanian carbonates (Gray Mesa, Atrasado and Bar B formations) and overlying nonmarine red beds of the Lower Permian (Wolfcampian) Abo Formation. The Bursum Formation is up to 330 m thick, but generally less than 100 m thick, and composed of interbedded nonmarine red beds and shallow marine shale, sandstone, conglomerate and limestone. It is characterized by strong regional and local variation in thickness and facies, which is best expressed by recognizing four members of the formation: (1) the Oso Ridge Member represents the proximal facies, which is thin and composed of mainly nonmarine red beds, laterally grading into the (2) Red Tanks Member, which is up to approximately 120 m thick and composed of alternating nonmarine red beds and marine shale, sandstone, conglomerate, marl and limestone, which locally form well-developed cycles; basinward (towards the southeast) this facies grades into the (3) Bruton Member, which is of similar thickness to the Red Tanks Member but dominantly marine with minor intercalations of fine-grained nonmarine red beds and locally developed cycles; and (4) the Laborcita Member represents the transitional facies of nonmarine and marine siliciclastic sediments and shallow marine limestone in the northern Sacramento Mountains, which accumulated on the narrow eastern shelf of the Orogrande basin.

The sharp lithologic contact between underlying strata of different ages and also between overlying red beds indicates the presence of a major unconformity at the base of the Bursum Formation and, locally, also at the top. There is also evidence for the existence of minor disconformities within the Bursum Formation, particularly at the base of the depositional sequences in the lower part. Cycles that are locally well developed in the Red Tanks and Bruton members, in the more basinward facies of the Bursum Formation, may have been influenced by glacio-eustatic sea-level fluctuations related to the Gondwana glaciation. However, regional synsedimentary tectonic movements of the Ancestral Rocky Mountain (ARM) orogeny strongly influenced Bursum sedimentation, resulting in conspicuous lateral variations in lithofacies and thickness. The widespread Bursum Formation thus represents significant tectonic activity of the ARM orogeny during the late Virgilian-early Wolfcampian.

INTRODUCTION

In central New Mexico (Figs. 1-2), the Bursum Formation represents the transitional facies between the underlying, dominantly shallow marine Upper Pennsylvanian carbonate deposits of the Atrasado Formation (and equivalents) and the overlying Lower Permian red beds of the Abo Formation. Due to regional and local variations in facies, Lucas and Krainer (2004) proposed to distinguish four members of the Bursum Formation (Fig. 3): (1) the thin and dominantly nonmarine Oso Ridge Member exposed in the Zuni Mountains and Joyita Hills; (2) the mixed siliciclastic-carbonate, marine and nonmarine, locally cyclic succession of the Red Tanks Member, present in the Lucero uplift, Sandia Mountains, Manzano Mountains, Los Pinos Mountains and locally in the Joyita Hills; (3) the dominantly marine Bruton Member, present in the Joyita Hills, Cooke's Range, Oscura Mountains, northern San Andres Mountains and Caballo Mountains; and (4) the mixed siliciclastic-carbonate, dominantly marine Laborcita Member deposited on the eastern shelf of the Orogrande Basin, exposed in the Sacramento Mountains. Here, we review these members of the Bursum Formation to summarize their lithostratigraphy, sedimentology and significance for understanding regional deposition in central New Mexico across the Pennsylvanian-Permian boundary.

PREVIOUS STUDIES

Across much of New Mexico (Figs. 1-2), the lithosome transitional between the marine Pennsylvanian and nonmarine Permian is a mappable lithologic unit deserving the rank of formation. Several

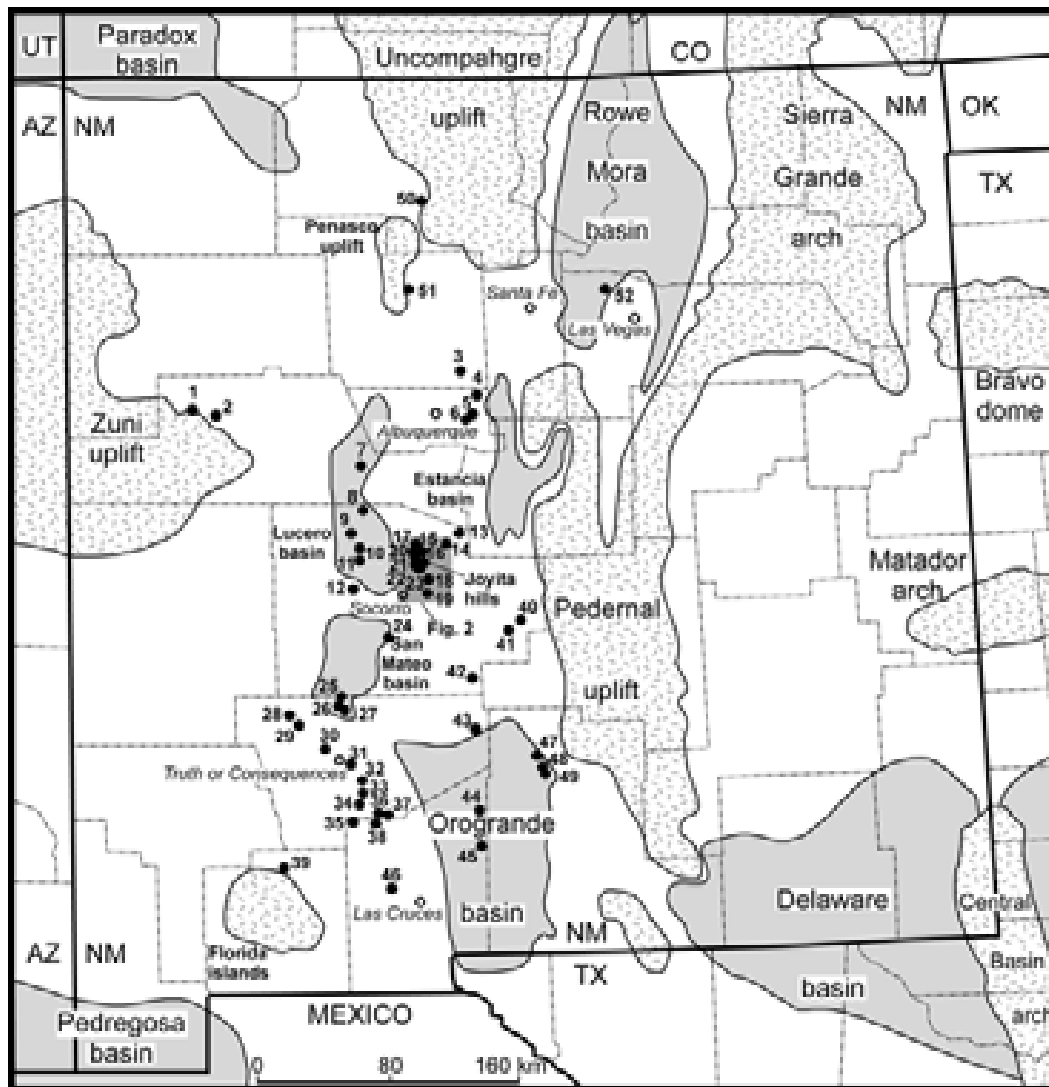
lithostratigraphic names have been applied to these strata: Bruton Formation (Thompson, 1942), Bursum Formation (Wilpolt et al., 1946), Aqua Torres Formation (Stark and Dapples, 1946), Red Tanks Member of Madera Limestone (Kelley and Wood, 1946), Laborcita Formation (Otté, 1959) and Oso Ridge Member of Abo Formation (Armstrong et al., 1994). Over the last 15 years, we have restudied the type sections of all these units, as well as many other correlative sections across New Mexico (Figs. 1-2) to conclude that only one, formation-rank name is necessary for the transitional Pennsylvanian-Permian lithosome. Our work has been published in a variety of articles devoted mostly to local studies of the Bursum lithosome, many of which are referred to here (see especially Lucas and Krainer, 2004 and Krainer and Lucas, 2009), that cite the diverse literature on Bursum geology and paleontology.

LITHOSTRATIGRAPHY

Oso Ridge Member

The Oso Ridge Member is exposed in the Zuni Mountains (west-central New Mexico) and locally in the Joyita Hills (Figs. 1-2). The best outcrops in the Zuni Mountains are located at La Jara Springs, Sawyer and Oso Ridge (Fig. 4). The succession is up to ~ 12 m thick, rests on Proterozoic basement and is overlain by red beds of the Abo Formation (Krainer et al., 2003a; details in Krainer et al., 2003b).

In the Joyita Hills the Oso Ridge Member is 10.5 (Joyita Hills A) and 9 m thick (Joyita Hills B) (Krainer and Lucas, 2009). Section A rests on Precambrian basement rocks, and section B is on the Pennsylvanian Gray Mesa Formation.



- | | |
|------------------------------------|--|
| 1 Sawyer/Zuni Mountains | 27 Red Gap/Fra Cristobal |
| 2 La Jara Springs/Zuni Mountains | 28 Iron Mountain/Cuchillo Mts. |
| 3 Placitas | 29 Deep Well/Cuchillo Mts. |
| 4 Sandia Park | 30 Whisky Canyon/Mud Springs Mts. |
| 5 Los Pinos | 31 Truth or Consequences |
| 6 Tijeras | 32 Red Canyon/Caballo Mountains |
| 7 Carrizo Arroyo | 33 Green Canyon/Caballo Mountain |
| 8 Red Tanks Arroyo | 34 Bar B Draw/Caballo Mountains |
| 9 Coyote Draw | 35 Derry Hills |
| 10 Southern Ladron Mts. | 36 McLeod B |
| 11 San Lorenzo Canyon A + B | 37 McLeod Draw |
| 12 Kelly Mine /Magdalena | 38 McLeod Hills |
| 13 Abo Pass | 39 Fryingpan Canyon/Cookes Range |
| 14 Aqua Torres/Los Pinos Mts. | 40 Bruton Canyon/Oscura Mountains |
| 15 Joyita Hills A, B | 41 Bursum Type Section |
| 16 Sierra de la Cruz/Joyita Hills | 42 Mockingbird Gap |
| 17 Arroyo Tinajas/Joyita Hills | 43 Rhodes Canyon/San Andres Mountains |
| 18 Loma de Las Canas /Joyita Hills | 44 Hembrillo Canyon/San Andres Mountains |
| 19 Cerro del Viboro/Joyita Hills | 45 Ash Canyon/San Andres Mountains |
| 20 Gallina Well | 46 Robledo Mountains |
| 21 Mesa del Yeso | 47 Scorpion Mound |
| 22 Minas de Chupadera | 48 Laborita Canyon |
| 23 Ojo de Amado | 49 Fresnal Canyon |
| 24 Little San Pascual Mountains | 50 El Cobre Canyon |
| 25 Fra Cristobal North | 51 Jemez Monument |
| 26 Hellion Canyon /Fra Cristobal | 52 Gallinas Creek |

FIGURE 1. Virgilian-Wolfcampian paleogeographic map of New Mexico showing the location of the studied sections of the Bursum Formation and equivalent successions.

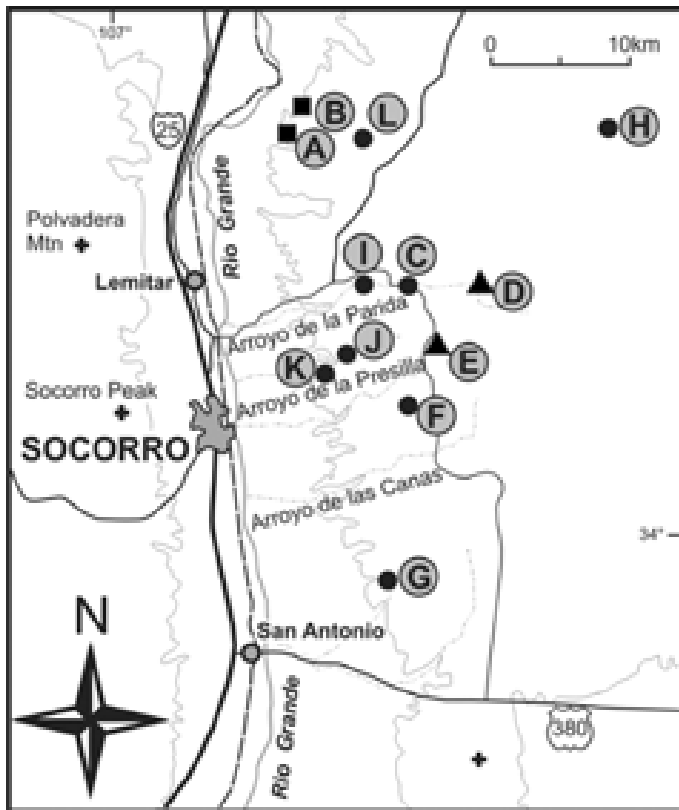


FIGURE 2. Detailed map showing location of the studied sections in the area east of Socorro (see Fig. 1). **Squares:** Oso Ridge Member, **circles:** Red Tanks Member and **triangles:** Bruton Member. **A,** Joyita Hills section A. **B,** Joyita Hills section B. **C,** Mesa del Yeso. **D,** Sierra de la Cruz. **E,** Arroyo Tinajas. **F,** Loma de las Cañas. **G,** Cerro del Viboro. **H,** Agua Torres. **I,** Gallina Well. **J,** Minas de Chupadera. **K,** Ojo de Amado. **L,** Cibola Canyon.

Lithology

The Oso Ridge Member in the Zuni Mountains is composed of two depositional sequences, each comprising red beds that are overlain by thin limestone intervals (Fig. 4). The red beds include poorly sorted, clast-supported conglomerate and crossbedded sandstone, fine-grained sandstone, siltstone and mudstone. Limestone intervals are 0.3-1.7 m thick, wavy to nodular bedded, partly indistinctly bedded and contain abundant siliciclastic grains (Krainer et al., 2003b).

In the Joyita Hills (Joyita Hills A and B: Krainer and Lucas, 2009), the Oso Ridge Member is composed of red beds (conglomerate, sandstone and shale) with intercalated pedogenic limestone at section B and thin limestone beds at section A.

In the upper part of section A, coarse-grained arkosic sandstone is intercalated that contains large fragments of echinoderms, bryozoans and brachiopods.

Petrography

Conglomerate of the Oso Ridge Member in the Zuni Mountains contains abundant quartz grains; sandstone is poorly sorted and composed of angular to subangular grains of quartz, feldspar, granitic and volcanic rock fragments. Detrital grains are cemented by calcite (Krainer et al., 2003b).

Microfacies

Limestone of the Oso Ridge Member in the Zuni Mountains is composed of several microfacies such as bioclastic grainstone to rudstone, bioclastic wackestone and packstone, bioclastic mudstone, peloidal packstone and rare bindstone (thin stromatolitic layer). The fossil as-

semblage is of low to moderate diversity. Siliciclastic grains are present in most limestone samples, locally forming fossiliferous, mixed siliciclastic-carbonate sandstone (Krainer et al., 2003b).

Limestone of the Oso Ridge Member in the Joyita Hills is composed of rudstone containing abundant fragments of bryozoans, echinoderms and brachiopods (Krainer and Lucas, 2009).

Red Tanks Member

The type section of the Red Tanks Member is exposed at Carrizo Arroyo in the Lucero uplift of Valencia County (Fig. 1) (Kelley and Wood, 1946). There, the Red Tanks Member is 98 m thick and is mostly green and red shale, mudstone and siltstone with minor intercalated beds of limestone, conglomerate and sandstone (Fig. 5). The section has been studied in detail by Krainer and Lucas (2004). In the Lucero uplift the thickness of the Red Tanks Member increases from Carrizo Arroyo towards the south to 120.5 m at Coyote Draw and then thins from there to the south towards the Ladron Mountains (Lucas and Krainer, 2004).

At Sandia Park, the Red Tanks Member is approximately 125 m thick, and at Los Pinos, a 90 m thick, complete section of the Red Tanks Member is well exposed along the road. At both locations the Red Tanks Member rests on bedded limestone of the Atrasado Formation and is overlain by sandy red beds of the Abo Formation (Fig. 6).

East of Socorro the Red Tanks Member is up to 120 m thick and locally displays a well-developed cyclic succession (details in Krainer and Lucas, 2009).

The only section in the Fra Cristobal Mountains assigned to the Red Tanks Member is Red Gap A, where the Bursum Formation consists of greenish-gray and reddish shale with abundant limestone nodules in the upper part. Intercalated are a thin carbonate conglomerate and sandstone in the lower part and several nodular pedogenic limestone intervals (0.2-0.4 m) in the upper part.

In the Caballo Mountains, the Bursum Formation is assigned to the Red Tanks Member, which is exposed at several localities. The Red Tanks Member rests on the Bar B Formation and is overlain by red beds of the Abo Formation. Thickness varies considerably from zero (Abo red beds directly on Bar B Formation: Bar B Draw) to 8 m (Green Canyon), 18 m (McLeod Hills), 25 m (Red Canyon), 45 m (McLeod Draw) and 80 m (McLeod B) (Lucas et al., 2012). In the McLeod Hills the Bursum Formation is up to 107 m thick according to Singleton (1990).

Lithology

At the type section the main lithotypes of the Red Tanks Member are mudstone-siltstone, marly shale and marl, sandstone, conglomerate and limestone (Fig. 5).

Mudstone-siltstone, marly shale and marl units are up to several m thick, gray, greenish, purple, red, and locally contain pedogenic limestone nodules. Greenish-gray mudstone-siltstone contains several horizons with a rich fossil flora and fauna.

Sandstone and siltstone units are up to 1.7 m thick, crossbedded and ripple laminated, and locally bioturbated. Conglomerate beds are up to 1 m thick, poorly sorted and composed mainly of carbonate clasts with diameters up to 20 cm, mostly <5cm. Limestone occurs as thin limestone beds or thicker bedded limestone units up to 3.8 m thick, and subordinately as pedogenic limestone horizons (Krainer et al., 2001; Lucas and Krainer, 2002; Lucas et al., 2004; Krainer and Lucas, 2004).

Farther south in the Lucero uplift (Red Tanks Arroyo and Coyote Draw) limestone is more abundant and contains a higher diversity fossil assemblage in the upper part compared to Carrizo Arroyo (Lucas and Krainer, 2004). A cyclic pattern similar to that at Carrizo Arroyo is not observed, and a correlation of individual limestone horizons and coarse clastic intervals is not possible (Lucas and Krainer, 2004).

At the Sandia Park and Los Pinos sections the dominant lithofacies is mudstone-siltstone, in which thin conglomerate beds, sandstone units up to 7 m thick and a few thin limestone beds are intercalated (Fig. 6). Mudstone-siltstone is micaceous, mostly red, subordinately greenish-

AGE	ZUNI MOUNTAINS	CENTRAL NM	SOUTHERN NM	SACRAMENTO MOUNTAINS
WOLFCAMPIAN	Abo Formation	Abo Formation	Abo Formation & Hueco Group	Abo Formation & Hueco Group
		Bruton Member	Bruton Member	Laborcita Member
VIRGILIAN	Oso Ridge Member	Red Tanks Member	Pennsylvanian marine strata	Pennsylvanian marine strata
		Pennsylvanian marine strata		

BURSUM FM. →

FIGURE 3. Regional stratigraphic nomenclature of the Bursum Formation (from Lucas and Krainer, 2004).

gray, laminated and non-fissile. Small (mostly 1-2 cm) carbonate nodules are interspersed in the mudstone-siltstone, particularly in the middle part where they occur in distinct horizons.

At Los Pinos two thin, fine-grained conglomerate beds are intercalated which are composed of carbonate clasts and 0.3 m thick in the lower part and 0.1 m thick in the upper part. At Los Pinos sandstone is a common lithofacies and occurs as 0.3-7 m thick intervals. The thicker sandstone intervals form fining-upward sequences starting with coarse, partly pebbly sandstone that erosively overlies mudstone-siltstone, displays large-scale trough crossbedding and grades upward into fine-grained sandstone with small-scale trough or planar crossbedding, and finally horizontally laminated sandstone and (rarely) ripple-laminated sandstone on top.

East of Socorro, the Bursum Formation is composed of alternating shale, sandstone, conglomerate and limestone of the same lithotypes as described from the type section. The Oso Ridge Member, which is exposed in the Joyita Hills as a thin and dominantly nonmarine succession of red beds, represents the proximal facies (to the source rocks) of the Bursum Formation, and towards the east and southeast grades into the thicker Red Tanks Member, which is exposed in a number of sections that have been studied in detail by Krainer and Lucas (2009): Mesa del Yeso, Gallina Well, Minas de Chupadera, Ojo de Amado, Loma de las Cañas, Cerro del Viboro and Aqua (Agua de) Torres. The sections recently described from Cibola Canyon east of the Joyita Hills by Allen et al. (2013) and at Abo Pass (Krainer et al., 2009) are also assigned to the Red Tanks Member. Farther to the east-southeast the Red Tanks Member grades into the dominantly marine Bruton Member represented by the Sierra de la Cruz and Arroyo Tinajas sections.

The thin Red Tanks Member at Green Canyon in the Caballo Mountains is mainly composed of red mudstone with a few thin intercalated limestone beds and a polymict limestone conglomerate near the top, which is overlain by a thin pedogenic carbonate bed and red beds of the Abo Formation.

In the McLeod Hills the Red Tanks Member begins with a carbonate conglomerate at the base, which is up to 1.1 m thick. Conglomerate and sandstone beds are also intercalated higher in the section (0.2-1.7 m thick). Locally (McLeod B), carbonate clasts with diameters up to 30 cm occur.

In general, the Red Tanks Member is composed of alternating conglomerate, sandstone and limestone intervals, separated by covered intervals (most probably representing mudstone) and mudstone. Limestone occurs as individual limestone beds or as bedded or nodular limestone intervals up to 4.3 m thick.

Petrography

At the type section the most abundant grain type of sandstone is quartz; detrital feldspars are also common, and other grain types are rare (subarkose-arkose). Conglomerate beds in the Carrizo Arroyo section are poorly sorted, clast-supported and composed of carbonate clasts and subordinate reworked mudstone and siltstone clasts. Conglomerate beds contain abundant fossil fragments, are cemented by calcite and contain small amounts of matrix (Krainer and Lucas, 2004). At Red Tanks Arroyo and Coyote Draw, sandstone is composed of quartz grains with very rare feldspar grains (quartzarenite). Conglomerate beds are composed of various types of carbonate clasts (Lucas and Krainer, 2004).

In the Sandia and Manzanita Mountains, sandstone is arkosic in composition, and conglomerate is composed of mostly micritic carbonate clasts; some fossil fragments are present.

East of Socorro three types of conglomerate (with regard to petrographic composition) are recognized in the Red Tanks Member: siliciclastic, mixed siliciclastic and carbonate. Sandstones are classified as subarkose to arkose and contain noticeable amounts of granitic rock fragments (see details in Krainer and Lucas, 2009).

In the McLeod Hills, conglomerate beds are clast-supported, poorly sorted and composed of rounded carbonate clasts and subordinate chert clasts. Commonly grain size is less than 3 cm, but locally clasts with diameters up to 30 cm occur. A few conglomerate and sandstone beds are well washed, cemented by calcite and contain abundant fragments of echinoderms, brachiopods, bryozoans and rare fusulinids.

Microfacies

The dominant microfacies of limestone of the lower three sequences of the Red Tanks Member at Carrizo Arroyo are bioclastic wackestones containing a diverse fossil assemblage (Krainer and Lucas, 2004). However, typical microfacies of the limestones that cap sequences 4, 5 and 6 are ostracodal wackestones (sequence 5) and bioclastic mudstones and wackestones with a low-diversity fossil assemblage (Fig. 5), indicating a restricted marine environment (Krainer and Lucas, 2004). Limestones in the Red Tanks Arroyo and Coyote Draw sections are mainly composed of bioclastic wackestone, subordinately of bioclastic mudstone, peloidal mudstone, ostracodal wackestone to mudstone and rare grainstone to packstone (Lucas and Krainer, 2004).

Limestone at Sandia Park, Los Pinos and Tijeras is dominantly wackestone to packstone, and subordinately packstone to rudstone with a diverse fossil assemblage. At Los Pinos two thin limestone beds in the uppermost part of the section are composed of wackestone to floatstone with a low diversity biota (Lucas et al., 2013b)

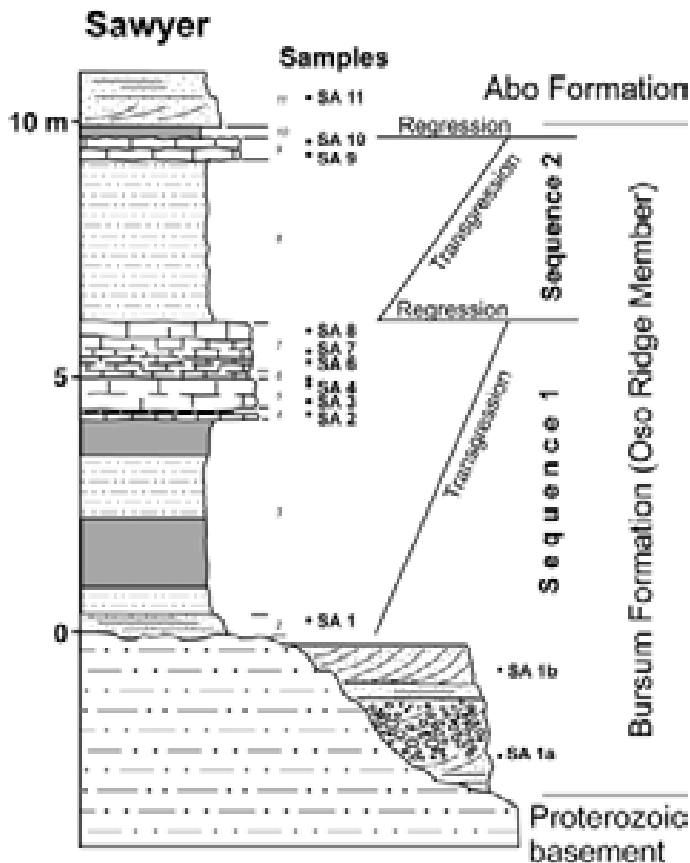


FIGURE 4. Measured section of the Oso Ridge Member in the Zuni Mountains at Sawyer (from Krainer et al., 2003b; for legend see Fig. 11).

East of Socorro the most common microfacies types of limestone are bioclastic mudstone and wackestone to packstone; rudstone and grainstone are rare (Krainer and Lucas, 2009).

Limestone in the southern Caballo Mountains-McLeod Hills is mainly bioclastic mudstone and wackestone.

Bruton Member

The type section of the Bruton Member of the Bursum Formation is at Bruton Canyon in the northern Oscura Mountains of Socorro County (type section of the Bruton Formation of Thompson, 1942), where it overlies indistinctly bedded to massive algal limestone (Moya Member) of the Atrasado Formation and is overlain by fine-grained red beds of the Abo Formation (Figs. 1, 7). A 5-m-thick Tertiary intrusive is intercalated in the uppermost limestone. At the type section the Bruton Member is approximately 115 m thick. The Bursum type section measures 85 m, and, due to its facies, is also assigned to the Bruton Member (Lucas et al., 2000a, b, 2002).

The Bruton Member is approximately 80 m thick at Sierra de la Cruz and 53 m at Arroyo Tinajas. At Fryingpan Canyon in the Cooke's Range of Luna County, a 25 m thick succession of interbedded shale, limestone-cobble conglomerate and marine limestone, which rests on Desmoinesian Gray Mesa Formation, and is overlain by red beds of the Abo Formation, can be assigned to the Bruton Member of the Bursum Formation (Fig. 8).

At Red Gap in the Fra Cristobal Mountains of Sierra County, above the Bar B Formation is a relatively thin (24 m thick) section of Bursum Formation (Fig. 9). Towards the north, the Bursum Formation thins to 10-16 m near Hellion Canyon and to 7.4 m near the northern end of the mountain range (Fig. 9), where the succession starts with a red mudstone slope (1.2 m), followed by a gray carbonate conglomerate (0.5

m), overlain by an interval composed of abundant gray, partly fossiliferous limestone nodules embedded in red mudstone (1.6 m), a nodular gray limestone bed (0.8 m), and purple red mudstone containing fossiliferous gray limestone nodules (1.5 m). The top is composed of bedded gray, recrystallized dolomitic limestone (1.8 m), overlain by red beds of the Abo Formation.

Lithology

At the type section the Bruton Member is composed of siltstone-mudstone comprising about 75% of the section and limestone comprising about 21% of the succession. Sandstone is very rare, and conglomerate is absent. Limestone occurs as thin beds (0.2 m) and thicker, bedded limestone intervals up to 6 m thick (Lucas et al., 2002).

The main lithologies at Sierra de la Cruz and Arroyo Tinajas are shale (56% at Arroyo Tinajas and 50% at Sierra de la Cruz), sandstone (1.5 and 2.5%), conglomerate (1.5 and 2.5%) limestone (20 and 15%) and rare pedogenic limestone (1% at Sierra de la Cruz).

Limestone occurs as even and wavy beds and wavy and nodular intervals up to 1.5 m thick. The Sierra de la Cruz section is dominantly marine (nearshore and restricted to open shallow marine), but a very rare nonmarine facies is developed near the base (Krainer and Lucas, 2009). At Arroyo Tinajas, the Bruton Member is nonmarine near the base grading into restricted and open, shallow marine facies in the upper part, with intercalated nonmarine facies (Krainer and Lucas, 2009).

At Fryingpan Canyon, the Bruton Member is composed of reddish mudstone containing abundant pedogenic limestone nodules, and intercalated limestone conglomerate and limestone (Fig. 8). Thin limestone beds containing abundant crinoidal debris are intercalated in the lower part, and fossiliferous nodular limestone in the middle and upper part (Krainer et al., 2003c).

At Red Gap B the Bursum Formation is composed of mostly covered (shale) slopes and red shale, locally containing small pedogenic carbonate nodules (Fig. 9). In the lower part two bedded fossiliferous gray limestone intervals are intercalated that contain crinoidal debris, brachiopods and gastropods. In the middle part several gray micritic limestone beds (0.1-0.5 m thick) are exposed. Nodular limestone beds, probably of pedogenic origin, occur near the base and in the upper part. Stromatolitic limestone forms the top of the Bursum Formation.

Near Hellion Canyon the Bursum Formation is 10 m thick (section B, Fig. 9) and consists of individual micritic limestone beds (0.2-0.4 m thick) partly containing ostracods, and rare limestone beds containing abundant crinoid fragments (crinoidal wackestone-floatstone). Limestone beds are separated by covered (shale) intervals. At section C (Fig. 9) the Bursum Formation is 15 m thick and composed of individual limestone beds and bedded limestone intervals up to 2 m thick, separated by covered (shale) intervals.

Petrography

Conglomerate at Sierra de la Cruz is massive, clast-supported and composed of abundant carbonate clasts and a few marine fossils. At Arroyo Tinajas mixed siliciclastic-carbonate conglomerate is intercalated. Sandstone in the lower part of both sections is mixed siliciclastic-carbonate in composition and contains abundant marine fossils. In the upper part arkosic sandstone is intercalated (Krainer and Lucas, 2009).

In the Joyita Hills sections A and B, conglomerate is poorly sorted and contains abundant clasts derived from the granitic basement. Sandstone is arkosic in composition (Krainer and Lucas, 2009).

Conglomerate at Fryingpan Canyon is clast-supported, moderately to poorly sorted, and the clasts are composed of different types of carbonate rocks. Rarely, fossil fragments (bryozoans, shell debris) are present.

One thin conglomerate bed is developed at Red Gap A that erosively overlies limestone. It is a polymict limestone conglomerate up to 0.8 m thick, containing clasts with diameters up to 5 cm, mostly <2 cm. Above the conglomerate a thin sandstone bed is intercalated.

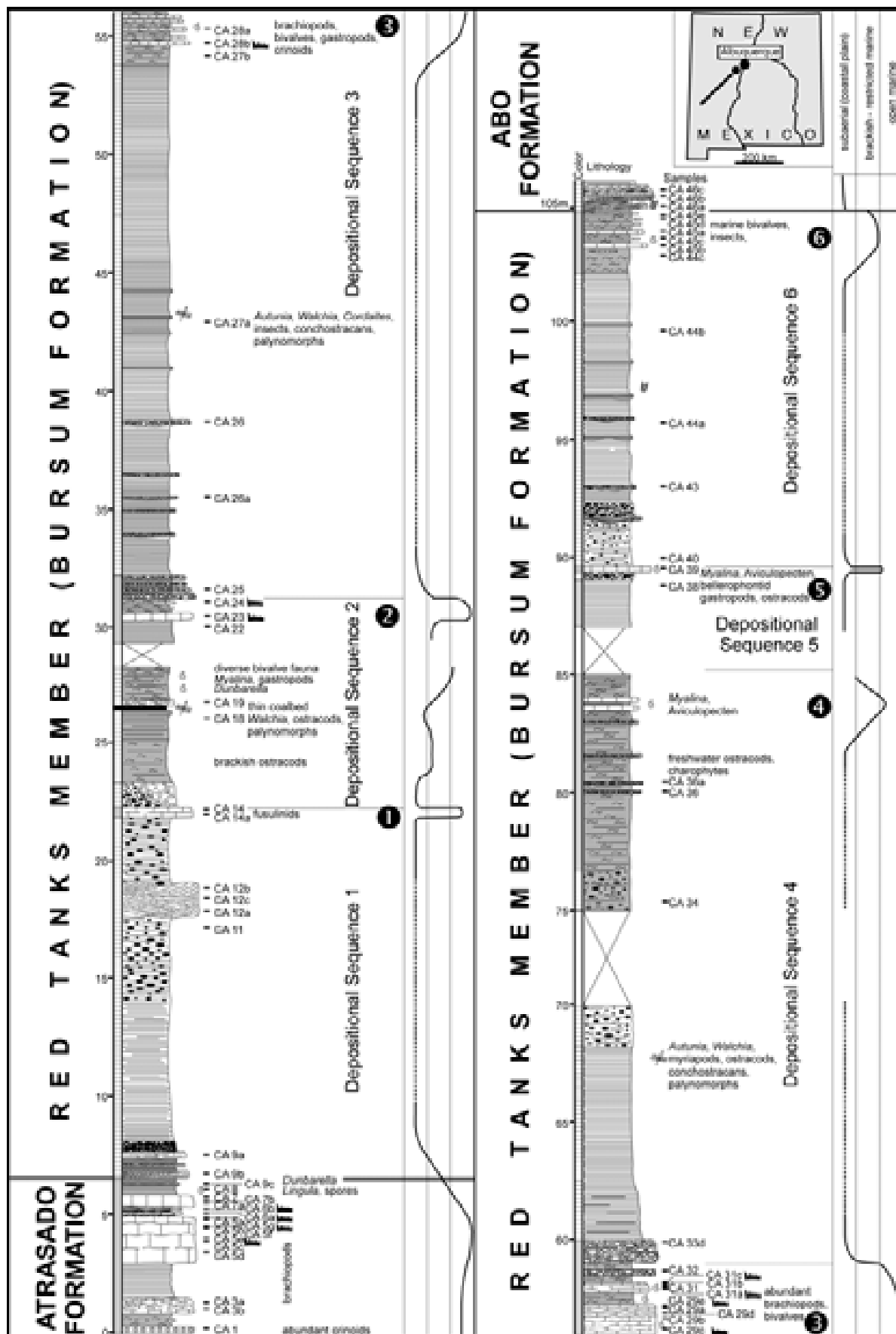


FIGURE 5. Type section of the Red Tanks Member at Carrizo Arroyo (from Krainer and Lucas, 2004; for legend see Fig. 11).

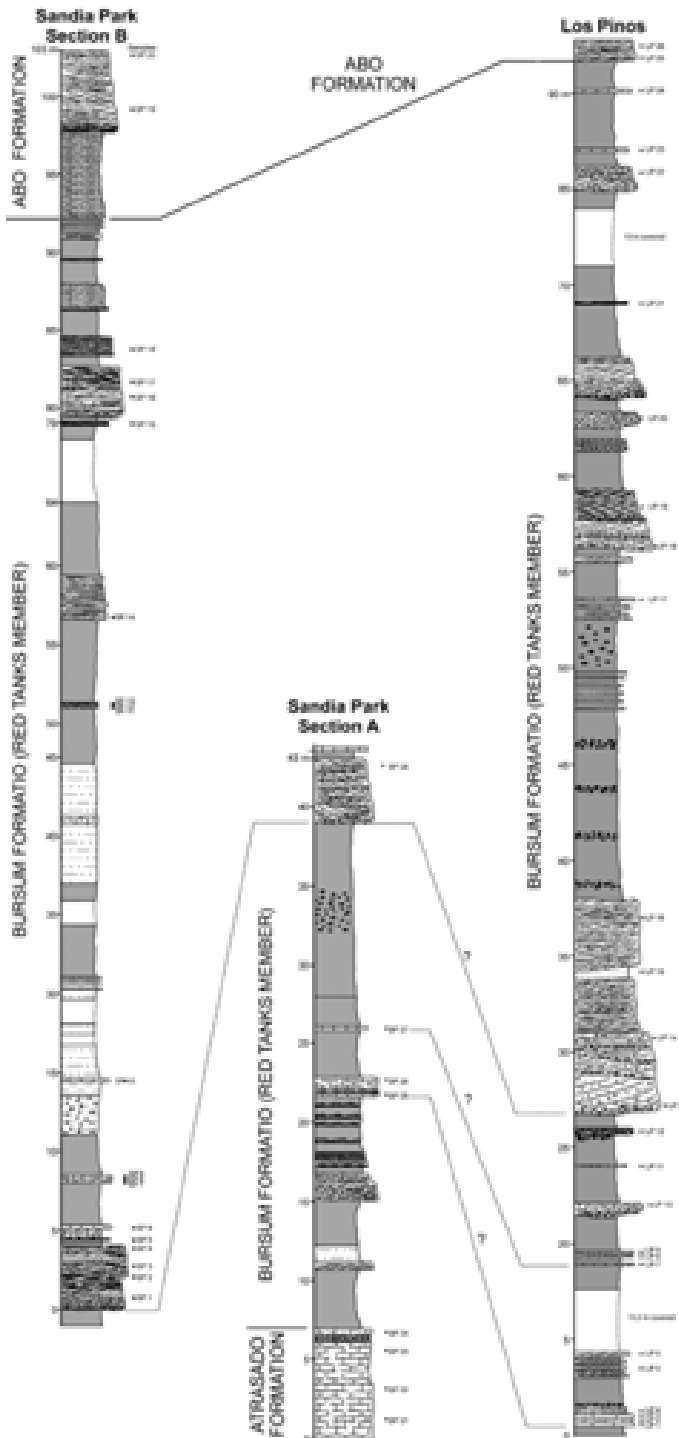


FIGURE 6. Measured sections of the Red Tanks Member at Sandia Park and Los Pinos (for legend see Fig. 11).

Microfacies

The most common microfacies at Bruton Canyon is bioclastic wackestone to packstone; fusulinid wackestone, bioclastic packstone and rudstone (storm layer) are subordinate. Most limestone samples contain a diverse fossil assemblage (Lucas et al., 2002).

The main microfacies types at Sierra de la Cruz and Arroyo Tinajas are bioclastic mudstone, wackestone and packstone, and rarely grainstone occurs. Most limestone beds contain a diverse fossil assemblage; subordinatedly, limestone beds (bioclastic mudstone) with a low-diversity biota occur (Krainer and Lucas, 2009).

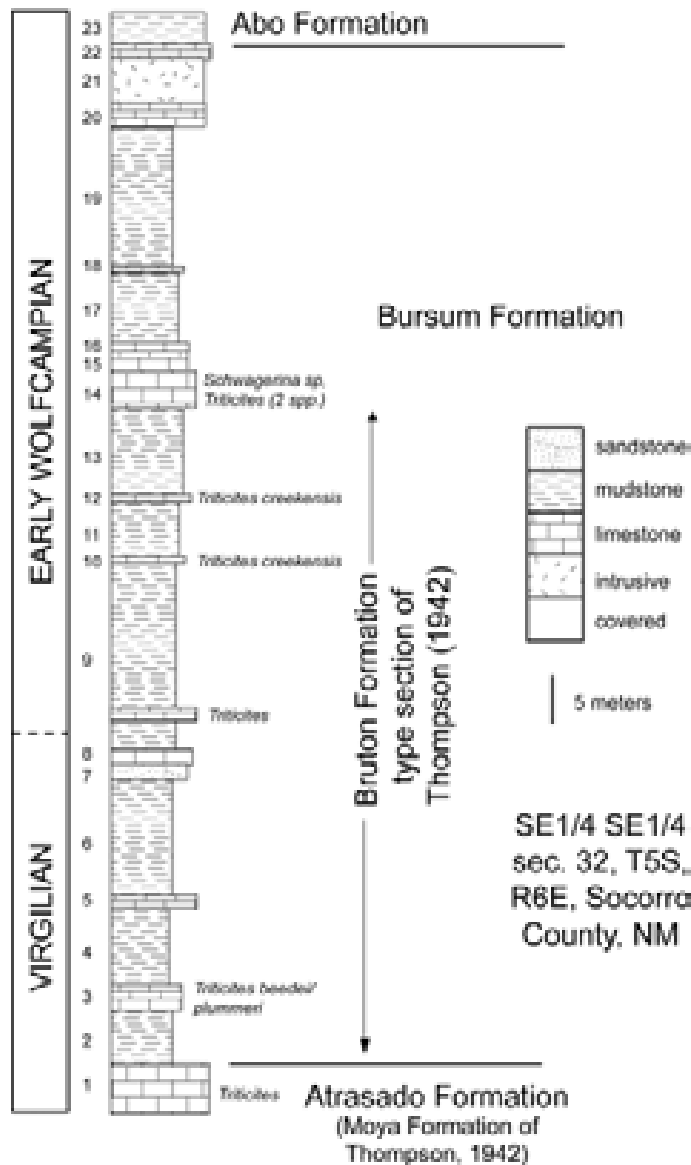


FIGURE 7. Measured section of the type section of the Bruton Member (from Lucas et al., 2002).

Limestone at Fryingpan Canyon is composed of bioclastic mudstone and wackestone, rare grainstone, packstone and rudstone. In the upper part, bindstone (stromatolite) is present (Krainer et al., 2003c).

Limestone in the Fra Cristobal Mountains is mainly composed of bioclastic mudstone and wackestone, and rarely floatstone and bindstone (stromatolite).

Laborcita Member

Otté (1959) coined the term Laborcita Formation for sedimentary rocks composed largely of gray and red mudstone, gray limestone, sandstone and conglomerate. The Laborcita Member is well known for its algal mounds (e.g. “Scorpion Mound”). Lucas and Krainer (2004) identified the Laborcita as a member of the Bursum Formation. The Laborcita Member is exposed primarily 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 (1959). At Scorpion Mound the Laborcita Member is 268 m thick, and the base is not exposed (Fig. 10). The boundary between the Laborcita Member and overlying Abo Formation is drawn at the top of the highest marine limestone of the Laborcita Member.

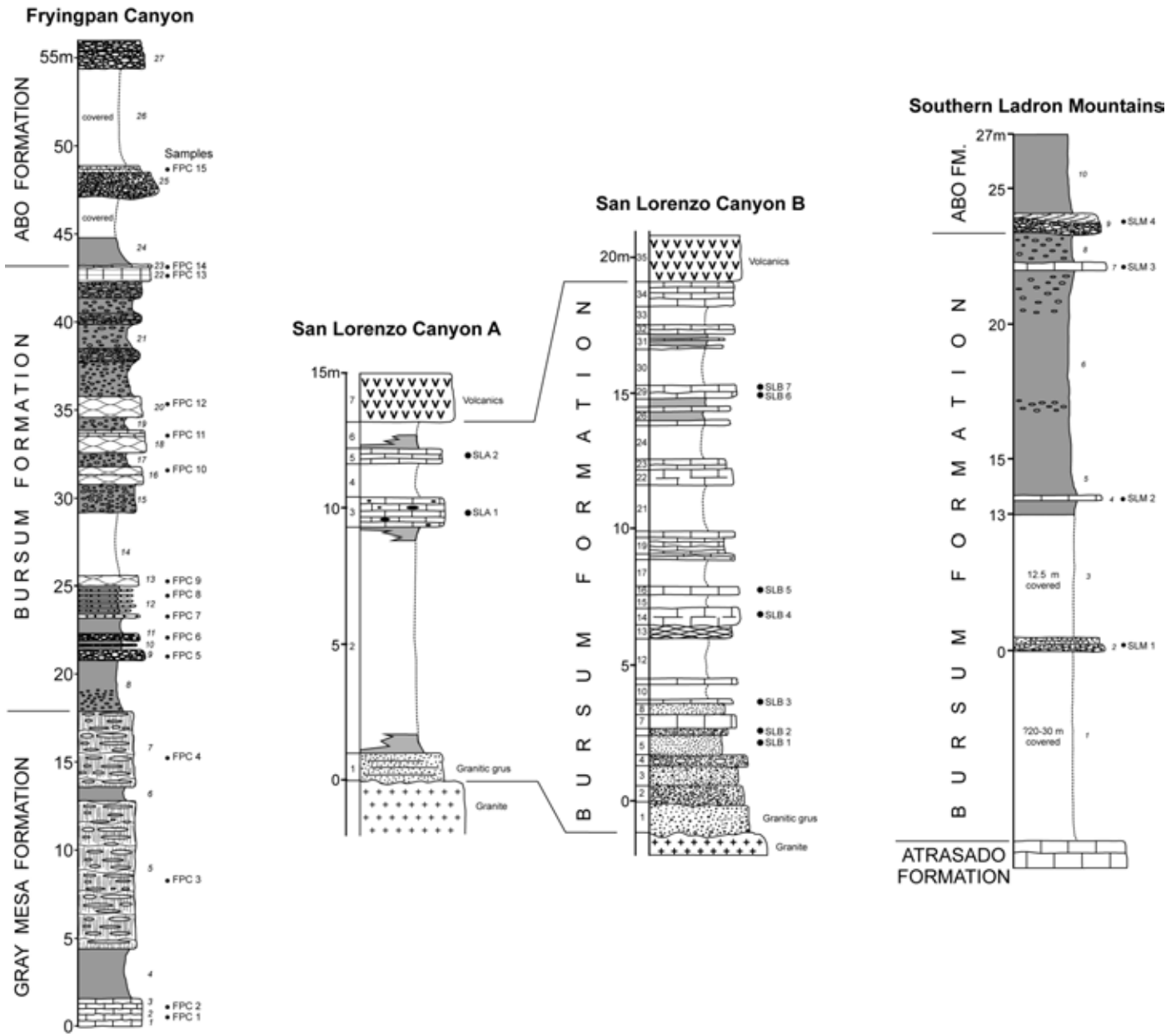


FIGURE 8. Measured sections of the Bursum Formation in the southern Ladoron Mountains, San Lorenzo Canyon and Fryingpan Canyon (for legend see Fig. 11).

Lithology

The different lithotypes of the Laborcita Member are arranged to form alternating cycles of limestone and siliciclastic sedimentary rocks. Otté (1959) pointed out that these lithologies change abruptly laterally and do not extend over long distances.

An ideal cycle begins with a conglomerate, 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.

In the field, the following lithotypes of limestone are recognized:

1. Thin-bedded limestone with even to wavy bedding composed of different types of wackestones and packstones; grainstones are rare.
2. Thin-bedded limestone with wavy to nodular bedding. The most frequent microfacies type is bioclastic wackestone; subordinate are

bioclastic packstone/grainstone and fusulinid wackestone.

3. Thick-bedded limestone composed of bioclastic wackestone/grainstone and algal wackestone, subordinate fusulinid wackestone, ooid wackestone and ooid/oncoid grainstone.

4. Massive algal mound facies composed of phylloid algal floatstone and subordinate other microfacies types.

5. Rare, thin pedogenic carbonate beds that occur in silty shale.

At Scorpion Mound (Fig. 10), the lower 90 m of the Laborcita Member is entirely siliciclastic and composed of greenish (and rare reddish) siltstone with intercalated sandstone beds (massive, horizontally laminated, ripple laminated) and crossbedded sandstone. Intercalated conglomerate units are up to 2.1 m thick and contain clasts with diameters up to 30 cm, mostly quartzite and rhyolite. The first thin limestone bed occurs approximately 90 m above the base. The base of the phylloid algal mound complex (“Scorpion Mound”) is about 135 m above the base; the mound is up to 11 m thick and overlain by a succession of predominantly siltstone with intercalated sandstone beds, rare conglom-

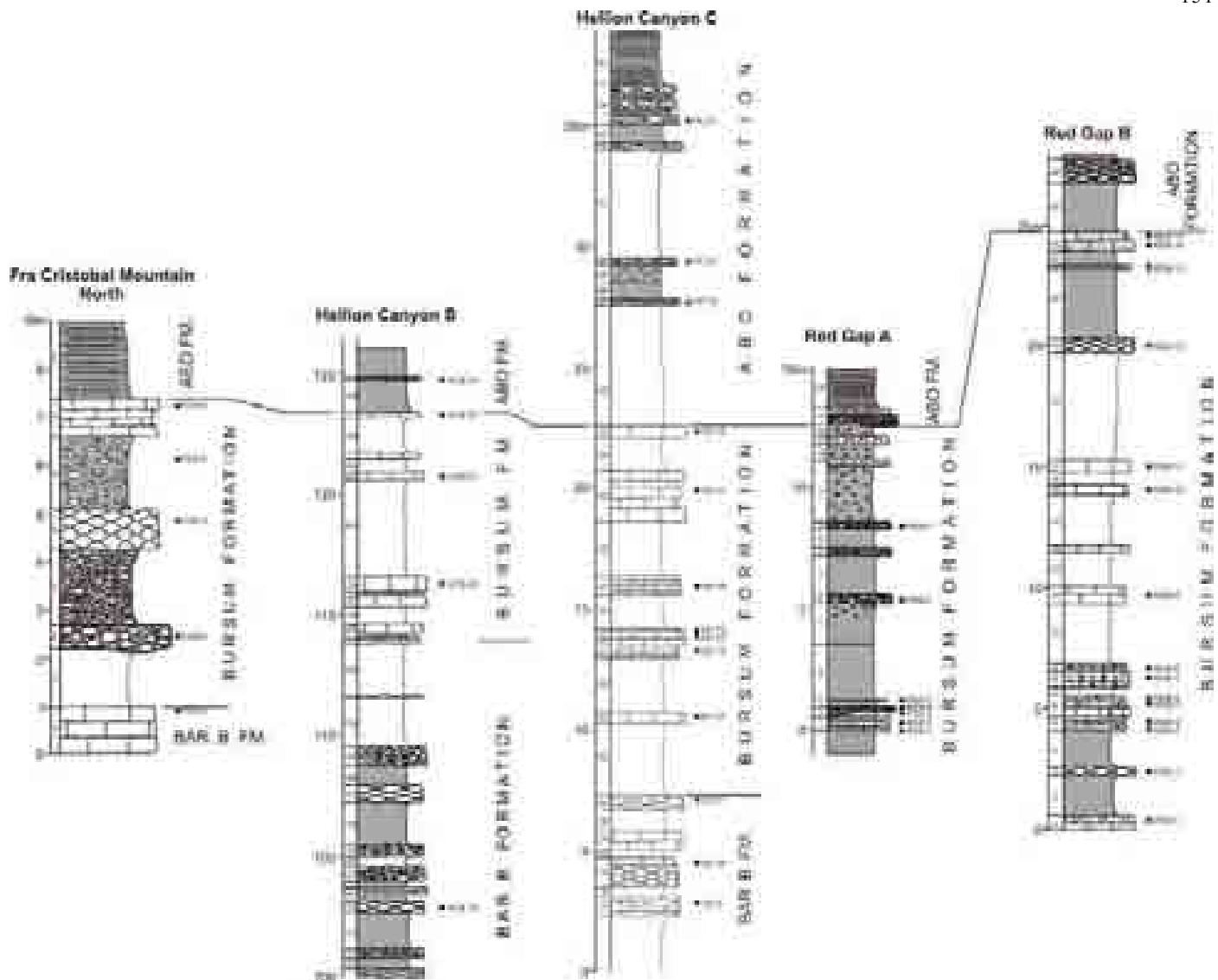


FIGURE 9. Measured sections of the Bursum Formation in the Fra Cristobal Mountains (for legend see Fig. 11).

erate and several limestone horizons. Algal limestone forms the top of the section and is overlain by red beds of the Abo Formation.

The mound complex extends from Scorpion Mound, which is located 1.5 km NE of Tularosa, 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 lithologies within the mound facies: (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.

Microfacies

Limestone of the Laborcita Member 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, phylloid algal wackestone, foraminiferal wackestone, oolitic wackestone, and oncolithic wackestone. Wackestone locally grades into packstone (mostly bioclastic packstone, rare peloidal packstone) and

algal wackestone 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, oolitic grainstone and peloid/oncoid grainstone (Krainer et al., 2007). From outcrops north of Laborcita Canyon, Fly (1985) described foreset-bedded sandy grainstones, blue-green algal dismicrites, digitate algal stromatolites and phylloid algal bindstones.

Petrography

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, too.

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 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.

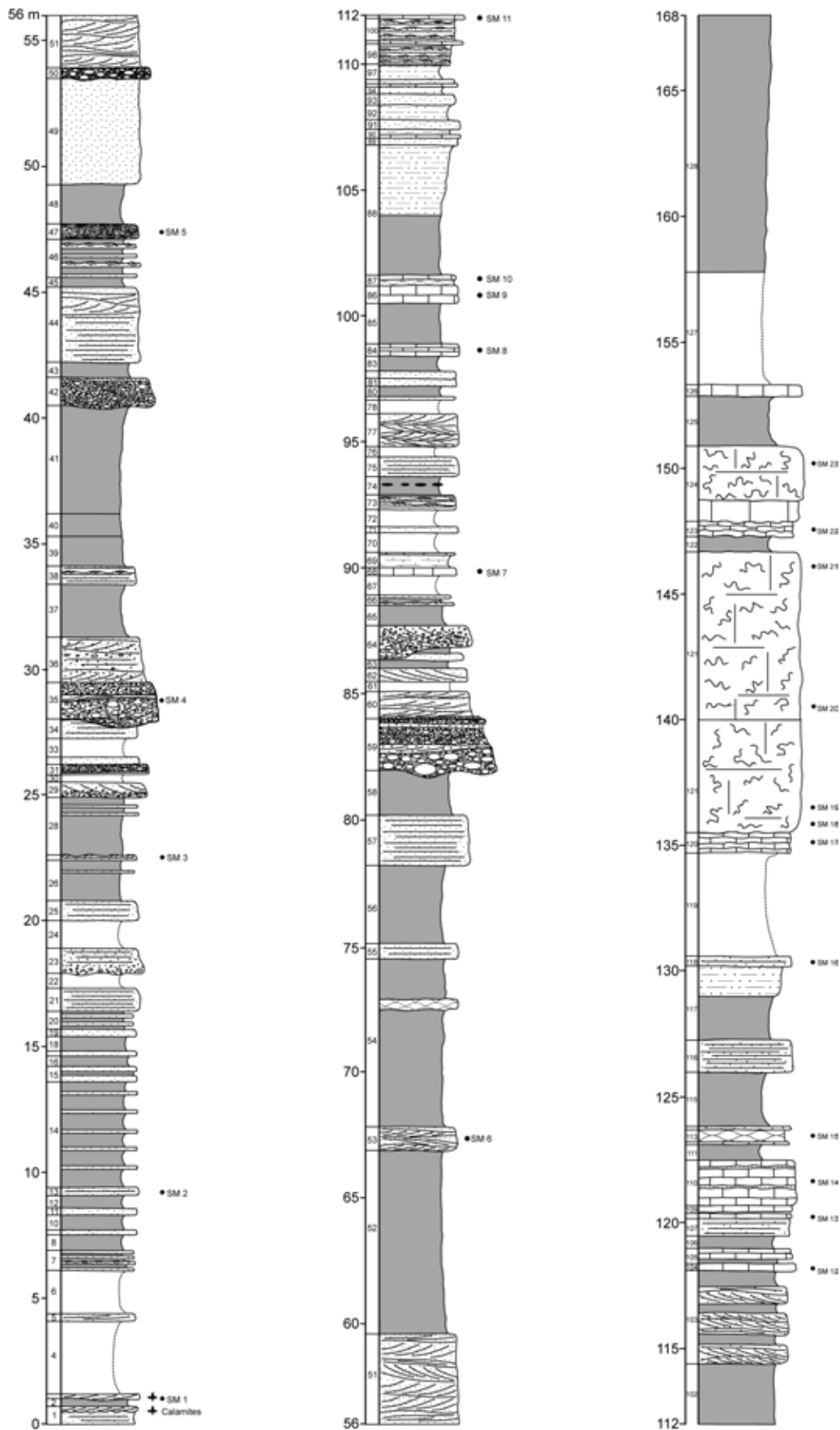


FIGURE 10. Measured section of the Laborcita Formation at Scorpion Mound northeast of Tularosa (for legend see Fig. 11).

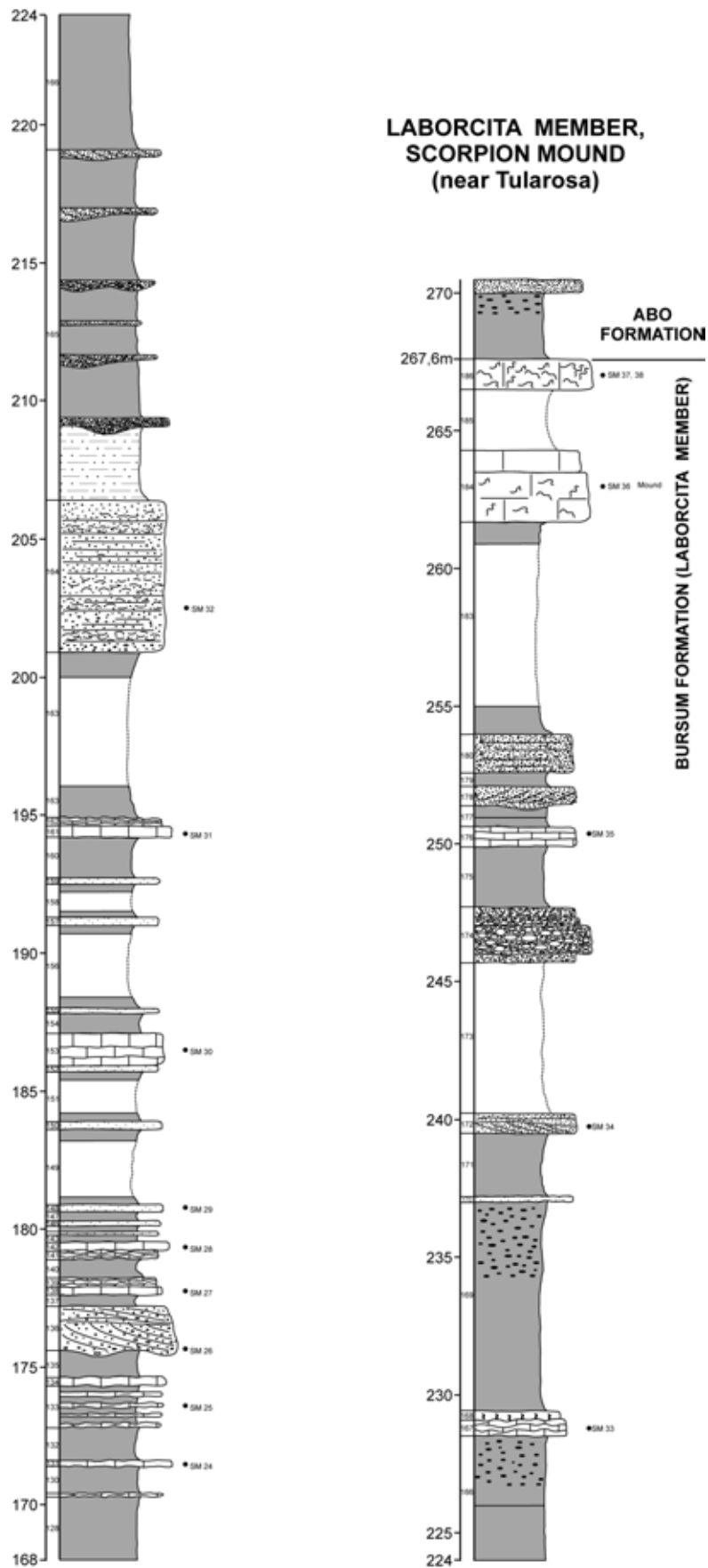


FIGURE 10. Continued. Measured section of the Laborcita Formation at Scorpion Mound northeast of Tularosa (for legend see Fig. 11).

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. Sandstones are frequently cemented by calcite replacing quartz, feldspar and chert grains (Krainer et al., 2007).

Sandstone of the Laborcita Member 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, contains rounded clasts of quartzite and rhyolite, rare carbonate clasts and sandy matrix.

REGIONAL VARIATION

The Bursum Formation, which represents the transitional facies between the dominantly shallow marine, carbonate Atrasado Formation and equivalents (Pennsylvanian) and the nonmarine red beds of the Abo Formation (Lower Permian), is characterized by pronounced lateral variations in thickness and facies, expressed by the different members (Krainer and Lucas, 2003).

The proximal facies, which is exposed close to the source rocks of the ARM uplifts (e.g., Zuni uplift, Joyita uplift), is represented by a thin succession composed of dominantly nonmarine red beds termed Oso Ridge Member. The maximum thickness is 12 m. So far, the Oso Ridge Member is known only from the Zuni Mountains and Joyita Hills (Figs. 3-4).

The proximal facies of the Oso Ridge Member grades basinward into the more distal facies of the Red Tanks Member, which is thicker and composed of alternating nonmarine red beds (conglomerate, sandstone, mudstone) alternating with shallow marine mudstone and limestone. Locally, the different lithotypes form well-developed cycles. Most Bursum sections in central New Mexico (Sandia Mountains, Lucero uplift, southern Ladron, Cuchillo Mountains, Cooke's Range, most sections east of Socorro, Caballo Mountains) due to their facies are assigned to the Red Tanks Member. Thickness reaches 124 m at Sandia Park and 120 m in the Lucero uplift and east of Socorro. West of the Rio Grande and south of the Lucero uplift (southern Ladron Mountains, San Lorenzo Canyon, Cuchillo Mountains, Cooke's Range; Fig. 8) the Bursum sections are less than 30 m thick. In the Caballo Mountains thickness varies strongly, ranging from zero to 107 m. Farther to the south, in the Orogrande Basin, the upper part of the Panther Seep Formation and lowermost part of the Hueco Group are equivalents of the Bursum strata.

Basinward, the Red Tanks Member grades into the distal Bruton Member, which is rather thick and dominantly shallow marine with minor amounts of nonmarine red beds. Particularly coarse-grained clastic sediments (conglomerate and sandstone) are rare. Locally, the Bruton Member displays a cyclic pattern. Typical Bruton Member sections are exposed in the Oscura Mountains, northern San Andres Mountains, east of Socorro (Sierra de la Cruz, Arroyo Tinajas) and in the Fra Cristobal Mountains. The maximum thickness of the Bruton Member is 80-85 m in the Oscura Mountains, northern San Andres Mountains and east of Socorro, and less than 30 m in the Fra Cristobal Mountains.

On the narrow eastern shelf of the Orogrande Basin the Bursum Formation is represented by a thick succession of mostly shallow marine, mixed siliciclastic-carbonate deposits termed Laborcita Member, exposed in the northern Sacramento Mountains.

In northern New Mexico (Fig. 1), in the Jemez Mountains, the Lower Permian Abo Formation is underlain by the Guadalupe Box Formation, which is dated as Late Desmoinesian to early-middle Virgilian (Krainer et al., 2005), indicating a gap (late Virgilian hiatus) between the Guadalupe Box Formation and Abo Formation.

Farther north (El Cobre Canyon) the Pennsylvanian-Permian transition is represented by nonmarine red beds of the Cutler Group (El Cobre Canyon Formation and Arroyo del Agua Formation) (Lucas et al., 2010; Krainer and Lucas, 2010).

In the Rowe-Mora basin ("Taos trough") of northern New Mexico, the Late Pennsylvanian to late Virgilian-Wolfcampian succession is represented by the Alamitos Formation, which is composed of well-developed transgressive-regressive cycles of nonmarine siliciclastics and shallow marine mudstone and limestone (Krainer et al., 2004). The facies is similar to that of the Red Tanks Member, although coarse siliciclastic sediments are more abundant in the Alamitos Formation.

CONTACTS

Lower Contact

With a sharp lithologic contact, the Bursum Formation rests on rocks of different age: on Proterozoic basement rocks (Zuni Mountains – Fig. 4, Joyita Hills, section A), Middle Pennsylvanian Gray Mesa Formation (Joyita Hills section B, Fryingpan Canyon in the Cooke's Range – Fig. 8), Upper Pennsylvanian Atrasado Formation (e.g. Sandia Mountains – Fig. 6, Lucero uplift – Fig. 5, east of Socorro, Oscura Mountains) and Middle-Upper Pennsylvanian Bar B Formation (Fra Cristobal – Fig. 9, and Caballo Mountains). Locally the Bursum Formation is absent, and the Abo Formation directly rests on the Atrasado Formation (e.g., Kelly Mine, Little San Pascual Mountains) or on the Bar B Formation (Bar B Draw) (Krainer et al., 2003c; Krainer and Lucas, 2009; Lucas et al., 2012).

At Carrizo Arroyo conodonts from the uppermost Atrasado Formation and Red Tanks Member indicate that a significant amount of latest Pennsylvanian to earliest Permian time is not represented by sediment, most likely at a major disconformity at the top of the Atrasado Formation. Additionally, there are probably smaller disconformities at the bases of depositional sequences in the lower part of the Red Tanks Member (Lucas et al., 2013a).

East of Socorro the basal conglomerate of the Red Tanks Member locally rests on pre-Virgilian strata with an erosional contact. This conglomerate contains reworked limestone clasts that indicate that locally parts of the underlying Pennsylvanian carbonate rocks were reworked prior to deposition of the Red Tanks Member. The age of the uppermost Atrasado Formation in most sections is middle Virgilian, indicating that the unconformity at the base of the Bursum Formation represents a hiatus that is equivalent to at least the late Virgilian (Krainer and Lucas, 2009).

A similar situation is observed in the Caballo Mountains, where the Red Tanks Member shows strong variations in thickness and is locally absent. The basal conglomerate of the Red Tanks Member in the McLeod Hills sections is composed of abundant limestone clasts derived from the underlying Pennsylvanian carbonate rocks. The upper Bar B Formation yielded conodonts of early Missourian age (Lucas et al. 2012), indicating strong evidence of a substantial hiatus in the Caballo Mountains at the base of the Red Tanks Member.

The presence of Bursum strata directly above Desmoinesian Gray Mesa limestones in the Cooke's Range suggests a profound hiatus at least equivalent to Missourian and much (or all) of Virgilian time. Indeed, the Gray Mesa-Bursum contact in Fryingpan Canyon is marked by a 3-m thick pedogenic calcrete at the Bursum base directly overlying cherty marine limestone at the Gray Mesa top, indicative of a significant unconformity. The presence of a Late Pennsylvanian unconformity in the southern Cooke's Range is further evidence of a regional tectonic event(s) in the Orogrande basin during Missourian time.

Upper Contact

At all studied sections the Bursum Formation is overlain by nonmarine red beds of the Abo Formation. East of Socorro, locally a basal conglomerate of the Abo Formation erosively overlies the Bursum For-

PALEONTOLOGY AND AGE

Oso Ridge Member

Limestone of the Oso Ridge Member in the Zuni Mountains contains abundant fragments of brachiopods and bryozoans; subordinate are echinoderm fragments, gastropods, ostracods, rare smaller foraminiferans (*Syzrania*) and cyanobacteria that locally form bindstone (stromatolite). Armstrong et al. (1994) reported brachiopods, bivalves, scaphopods, gastropods, nautiloids, crinoids, echinoderm spines and conodonts.

According to Krainer et al. (2003b), brachiopods of the Oso Ridge Member are mostly represented by *Composita subtilita* (Hall); other species determined include *Derbya*, *Neospirifer*, *Punctospirifer kentuckyensis* (Shumard), *Parajuresania nebrascensis* (Owen), *Linoproductus prattenianus* (Norwood and Pratten), *Antiquatonia*, bivalve steinkerns of *Wilkingia terminale* (Hall), *Schizodus*, *Solemya trapzoids* Meek, the pectinid bivalves *Aviculopecten* and *Polidevcia*, myalinid bivalves of the species *Septimyalina burmai* Newell. Other fossils are the gastropod *Naticoposis*, several types of bryozoans, fish teeth and spines. The conodonts "*Ellisonia*" *conflexa* (Ellison), *Adetognathus lautus* (Gunnell) and *Streptognathodus elongatus* Gunnell indicate a late Virgilian to earliest Wolfcampian age. Two of the brachiopods and bivalves also range into the Wolfcampian.

The only fossils observed in thin limestone (rudstone) and coarse-grained arkosic sandstone of the Oso Ridge Member in the Joyita Hills A and B sections are fragments of echinoderms (crinoid stem fragments), bryozoans and brachiopods (Krainer and Lucas, 2009).

Red Tanks Member

The Red Tanks Member at Carrizo Arroyo contains by far the richest and most diverse assemblage of nonmarine and marine fossils of the Bursum Formation (Lucas and Zeigler, 2004). Common fossils observed in limestone are echinoderm fragments (mostly crinoids), brachiopods, brachiopod spines, bivalves, gastropods, bryozoans, ostracods, smaller foraminifers, locally fusulinids and spicules.

Fusulinids from the uppermost Atrasado Formation indicate a late Virgilian age based on the occurrence of *Triticites* cf. *T. fresnalensis* and *T. imperialis*.

A limestone bed of cycle 2 of the Red Tanks Member contains *Triticites whetstonensis* and *T. cf. bensonensis* and a primitive *Leptotriticites* related to *L. eotexana*.

Limestone of the Red Tanks Member on top of depositional sequences 1, 2 and 3 (Fig. 5) contain a foraminiferal assemblage similar to that of the topmost Atrasado Formation: calcivertellids, *Climacamina*, *Diplosphaerina*, *Earlandia* ex gr. *elegans*, *Globivalvulina bulloides*, *Globivalvulina* ex gr. *moderata*, *Globivalvulina* sp., *Glomospira*, *Nodosinelloides potievskayae*, *Nodosinelloides* sp., *Palaeonubecularia*, *Spireitlina conspecta*, *Syzrania bella*, *Syzrania* sp. and *Tuberitina*. The wackestone on top of depositional sequence 2 contains abundant calcivertellid foraminiferans (Krainer and Lucas, 2004). Limestone on top of depositional sequence 4 has a low-diversity foraminiferal assemblage containing *Earlandia* and *Palaeonubecularia*. The uppermost limestone beds on top of depositional sequence 6 contain a low diversity foraminiferal assemblage with rare *Calcitornella* and *Syzrania* (Krainer and Lucas, 2004). These genera of non-fusulinid foraminiferans present in limestones of the Red Tanks Member are stratigraphically long ranging, and do not support precise placement of the Pennsylvanian-Permian boundary in the Carrizo Arroyo section.

Limestones of sequences 1, 2 and 3 yield conodonts (Orchard et al., 2004; Lucas et al., 2013a). New conodont data (Lucas et al., 2013a) include the presence of *Streptognathodus virgilicus* in the uppermost part of the Atrasado Formation, which constrains its age to the middle to upper part of the Virgilian. The only biostratigraphically significant conodont assemblage in the Red Tanks Member occurs in a limestone hori-



FIGURE 11. Legend for the sections shown in Figs. 4-10.

mation. Limestone clasts of the conglomerate indicate that substantial parts of the Bursum Formation were eroded prior to deposition of the red beds of the Abo Formation (Krainer and Lucas, 2009). At many locations where the Bursum Formation is thin or absent, carbonate conglomerates occur at the base or in the lowermost part of the Abo Formation. These conglomerates also contain limestone clasts, which are most likely derived from limestone of the underlying Bursum Formation, indicating that there is a major unconformity between the Bursum Formation and overlying Abo Formation (Krainer and Lucas, 2009).

zon near the middle of member, and is probably equivalent in age to the Midcontinent *Streptognathodus nevaensis* Zone, indicating an early to middle Asselian age.

Two assemblages of palynomorphs were reported from the Red Tanks Member at Carrizo Arroyo by Traverse and Ash (1999). Utting et al. (2004) identified palynomorphs from 10 stratigraphic levels within the Red Tanks Member.

Plant megafossils are found at three stratigraphic levels (Ash and Tidwell, 1982, 1986; Tidwell et al., 1999; Tidwell and Ash, 2004; Knaus and Lucas, 2004; DiMichele et al., 2004).

Ostracods are present in the Red Tanks Member but have not been fully published (Kietzke 1983). Extensive assemblages of conchostracans occur in two nonmarine horizons but are mostly unstudied (Schneider et al., 2004). Durden (1984) reported an insect fauna from the Red Tanks Member at Carrizo Arroyo. Insects were also described by Schneider et al. (2004) and Rasnitsyn et al. (2004). Kues (1983, 1984, 2002, 2004) documented more than 120 species of brachiopods and mollusks from the uppermost Atrasado Formation and Red Tanks Member.

The only vertebrate fossils identified from the Red Tanks Member at Carrizo Arroyo are teeth, denticles and scales of fishes (Harris et al., 2004; Johnson and Lucas, 2004). Vertebrate fossils from the Red Tanks Member of the Lucero uplift were reported by Harris and Lucas (2001, 2003) and Harris et al. (2003).

In the Lucero uplift, at Red Tanks Arroyo and Coyote Draw, as well as in the Sandia Mountains and east of Socorro, limestones contain a fossil assemblage similar to that from Carrizo Arroyo. Age-diagnostic fossils such as fusulinids are rare (Lucas and Krainer, 2004). East of Socorro, in the lower part of the sections Loma de las Cañas and Minas de Chupadera, and in the upper part of Cerro del Viboro, fusulinids (*Triticites*) occur. Of interest is the presence of *Triticites* in the lower part and *Triticites* and *Schwagerina* in the middle of the Red Tanks Member at Cibola Canyon (Allen et al., 2013), indicating a late Virgilian to early Wolfcampian age.

Bruton Member

At the type section of the Bursum Formation, Lucas et al. (2000a,b, 2002) recognized two fusulinid zones. Intercalated limestones from the lower part of the section contain *Triticites* sp., *T. creekensis* and *Leptotriticites fivensis*. The upper part is characterized by the occurrence of *Triticites creekensis*, *Leptotriticites glenensis*, *Schwagerina grandensis* and *S. campensis*. According to Lucas et al. (2000a,b, 2002) the fusulinid fauna indicates an Early Wolfcampian age.

In thin section limestone of the Bruton Member at Bruton Canyon contains a diverse fossil assemblage including bivalves, brachiopods, bryozoans, gastropods, echinoderms, fusulinids, ostracods, smaller foraminiferans (*Bradyina*, *Calcitornella*, *Calcivertella*, *Climacammina*, *Diplosphaerina*, *Globivalvulina*, *Hemigordius*, *Nodosinelloides*, *Pseudovidalina*, *Spireitlina*, *Syzrania*, *Tetrataxis*, *Tuberitina*), rare trilobites, *Tubiphytes*, algae (recrystallized phylloid algae, *Eugonophyllum*, *Epimastopora*, *Efluegelia johnsoni*), and cyanobacteria that encrust skeletal grains and locally form oncoids.

The lowermost limestone of the Bruton Member at Bruton Canyon (Fig. 7) contains the fusulinids *Triticites beedei* and *T. plummeri*, indicating a Virgilian age. Limestone in the upper part contains *Triticites creekensis*, *Triticites* sp. and *Schwagerina* sp., pointing to an early Wolfcampian age (Lucas et al., 2002).

From the Bursum type section Kues et al. (2000) briefly described an invertebrate fauna. In the Fra Cristobal Mountains the limestones of the Bruton Member contain brachiopods, crinoids, gastropods, and locally abundant ostracods. We did not observe age-diagnostic fossils such as fusulinids.

At Fryingpan Canyon (Fig. 8) limestone of the Bruton Member contains echinoderm fragments (mostly derived from crinoids), brachiopods, bryozoans, ostracods, small gastropods, rare smaller foraminifers,

phylloid algae and thin layers of cyanobacteria (stromatolite). Age-diagnostic fossils are absent. The underlying Gray Mesa Formation contains the Desmoinesian fusulinaceans *Beedeina* ex. gr. *euryteines* (Thompson) and *Wedekindellina* ex. gr. *euthysepta* (Henbest) (Krainer et al., 2003c).

Laborcita Member

Most limestone samples of the Laborcita Member contain a diverse fossil assemblage, including skeletons of calcareous algae (phylloid algae, mostly *Eugonophyllum*, subordinate dasycladacean algae *Epimastopora*, *Anthracoportella* and other algae), bivalves, brachiopods, bryozoans, echinoderms, fusulinids, gastropods, ostracods, smaller foraminiferans, rare trilobites, and *Tubiphytes* (Krainer et al., 2003d). In the mound facies the algal thalli are completely recrystallized and are assigned to the genus *Anchicodium*. Locally skeletons are encrusted by tubular foraminiferans and cyanobacteria, partly forming oncoids.

The lower part of the Laborcita Member at the type locality was considered to be of late Virgilian age by Otté (1959). He drew the base of the Wolfcampian at the base of the first limestone bed containing the fusulinid *Schwagerina* (probably *Thompsonites*), which occurs 26 m above the base. According to Steiner and Williams (1968), the Laborcita Formation is wholly Wolfcampian in age due to the occurrence of the fusulinids *Dunbarella americana*, *Schwagerina campensis*, *S. emaciata*, *Triticites creekensis* and *T. ventricosus*. Non-fusulinid foraminiferans indicate an early Wolfcampian age (Krainer et al., 2003d).

Rasbury et al. (1998) determined the Pennsylvanian-Permian boundary at 302 ± 2.4 Ma by U-Pb-dating of paleosols from the underlying Holder Formation and the Carboniferous-Permian boundary at 301 ± 2 Ma.

DEPOSITIONAL ENVIRONMENT

In general, the Red Tanks and Bruton Members are composed of alternating marine and nonmarine sediments that locally form well-developed cycles. Thickness and facies show strong vertical and lateral variations. The marine facies is represented by limestone and marine shale and rare marine sandstone and conglomerate. Nonmarine facies are shale, sandstone and conglomerate, and rare pedogenic limestone. The most common microfacies are bioclastic wackestone with a diverse fossil assemblage, and mudstone with a low diversity fossil assemblage. Less common are grainstone, packstone, rudstone and floatstone. Limestone was deposited in a dominantly low-energy, open marine to restricted shallow shelf environment. Most mudstone to siltstone is purple, red, gray, greenish-gray, and greenish, and locally contains a rich flora and fauna. Most of the mudstones, particularly those containing pedogenic carbonate nodules, were deposited in a nonmarine coastal plain environment. Marine mudstones are of greenish and gray color and contain marine fossils. Sandstone and conglomerate intervals may extend laterally over more than 100 m and display a sheet-like external geometry. Sheet-like sandstones and conglomerates are interpreted as deposits of broad, shallow channels of a braided stream system. Thin, massive carbonate conglomerates probably record turbulent debris flows. Thin sandstone sheets are sheet splays; thin sandstone lenses may represent crevasse channels. Locally, sandstones and conglomerates contain marine fossils, indicating deposition in a nearshore, high-energy shallow marine setting. Locally, nodular and massive pedogenic limestone horizons are intercalated.

Oso Ridge Member

In the Zuni Mountains conglomerate and crossbedded sandstone of the Oso Ridge Member are interpreted as fluvial channel-fill deposits; massive to horizontally laminated fine-grained sandstone, siltstone and mudstone are seen as sheetflood deposits, intercalated with thin channel-fill deposits. Microfacies of the limestone intervals indicate deposition in a shallow restricted to open marine shelf environment with siliciclastic influx. Also, the macrofossil assemblage indicates an offshore marine environment of normal salinity (Krainer et al., 2003b).

In the Joyita Hills the Oso Ridge Member is entirely nonmarine at section B and dominantly nonmarine in section A, with thin intercalated limestone composed of rudstone and coarse fossiliferous arkosic sandstone indicating a high-energy, shallow marine depositional environment (storm layers) (Krainer and Lucas, 2009).

Red Tanks Member

The stratigraphic architecture of the Red Tanks Member lithotypes at Carrizo Arroyo has been interpreted to indicate the presence of six depositional sequences (Krainer and Lucas, 2004). The base of each sequence is a bed of conglomerate or sandstone that is sharply incised into underlying mudstone-siltstone. Each sequence displays a fining-upward trend into mudstone-dominated floodplain or lacustrine strata. A marine limestone caps each sequence, and these limestones identify six marine flooding events. The sequences are mostly composed of mudstone/siltstone beds, some of which contain abundant calcrite nodules and other evidence of pedogenesis. A thin coal bed in the middle of depositional sequence 2 is underlain by fossiliferous siltstone (plants, ostracods, fish bones) and overlain by marly mudstone containing marginal marine molluscs and plant debris. Carbonate conglomerates at the bases of depositional sequences 3 and 4 contain abundant fossil fragments and probably represent upper shoreface deposits. Thin conglomerate beds in depositional sequence 4 are interpreted as small channel fills. Sandstones are present at the base of sequence 1 (shoreface deposits), in the upper part of sequence 1 (fluvial channel fills) and in sequence 6 (thin fluvial channel-fill deposits). The top of each depositional sequence is thin, fossiliferous gray limestone beds or interbedded gray mudstone and limestone.

According to Krainer and Lucas (2004), the Red Tanks Member sequences indicate that the coastal plain environment represented by mudstones/siltstones was repeatedly inundated by short term transgressive events during which fossiliferous, shallow marine limestones were deposited. Thus, limestones accumulated during relative highstands of sea level. Although glacio-eustatic fluctuations of sea level may be the source of at least some of these transgressive events, the Carrizo Arroyo section was deposited in the ARM orogenic belt, so regional tectonism was also an important force that drove local sedimentation (Krainer and Lucas, 2009).

Bruton Member

In general, the depositional environment of the Bruton Member is similar to that of the Red Tanks Member, ranging from nonmarine coastal plain with intercalated fluvial channel-fill deposits, to nearshore environments and restricted to open shallow marine shelf settings (Lucas et al., 2002).

Coarse clastic sediments (conglomerate and sandstone), particularly of nonmarine origin, are rare to absent in the Bruton Member. Nonmarine mudstone-siltstone, locally containing pedogenic limestone nodules, is less common in the Bruton Member compared to the Red Tanks Member. Shallow marine mudstone and limestone is more abundant in the Bruton Member. Locally, a cyclic pattern is observed as, for example, in the Sierra de la Cruz section east of Socorro. These differences suggest that the Bruton Member is a more distal facies compared to the Red Tanks Member and was deposited in the more basinal part of the Bursum depositional system.

Laborcita Member

The Laborcita Member, 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

Pederal uplift to the east during Early Permian time (Otté, 1959; Delgado, 1977; Fly, 1986; Krainer et al., 2003d, 2007). Particularly well studied are the algal reefs (mounds) in the upper part of the Laborcita Member (Otté and Parks, 1963; Cys and Mazullo, 1977; Parks, 1977; 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 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.

DISCUSSION

The dominant depositional basin across much of New Mexico during the Pennsylvanian-Permian transition was the Orogrande basin of the southern and central parts of the state. Sedimentary rocks that encompass that transition have been given several names: Bruton, Bursum, Aqua Torres, Red Tanks, Laborcita and Oso Ridge formations or members. In other areas, the transitional strata have either been assigned to the Abo Formation or to the Madera (Magdalena) Group/Formation without a separate name. A single formation name, Bursum Formation, best describes these strata, which constitute a mappable lithostratigraphic unit throughout their outcrop area. The Bursum Formation is a sedimentary succession up to 330 m thick and composed of interbedded siliciclastic red beds and marine limestone and shale. The Bursum Formation differs from underlying Pennsylvanian strata by its substantial content of red-bed shale and mudstone and some beds of limestone-pebble conglomerate and trough-crossbedded sandstone. Unlike the overlying Abo Formation, the Bursum contains beds of marine limestone and calcareous shale.

Thus, the Bursum represents a transitional facies between the dominantly shallow marine carbonate facies of the underlying Pennsylvanian Atrasado and Bar B formations and the continental red-bed facies of the overlying Abo Formation. Sedimentation of the different members of the Bursum Formation was strongly influenced by synsedimentary tectonic movements of the ARM orogeny, resulting in conspicuous lateral variations in lithofacies and thickness. In the keel of the Orogrande basin (southern Oscura Mountains-northern San Andres Mountains), the Bursum grades laterally into the upper part of the Panther Seep Formation and the lower part of the overlying Hueco Group.

Regional and local variation in Bursum lithofacies is best expressed by recognizing four members of the formation depending on their position related to the source rocks of ARM uplifts: (1) Oso Ridge Member, a thin (< 12 m) unit containing much reworked local Proterozoic basement in the Zuni Mountains and Joyita Hills, representing a very proximal facies; (2) Red Tanks Member, a moderately thick (<100 m) unit dominated by nonmarine shale and mudstone present in the Lucero uplift, Sandia, Manzano and Los Pinos Mountains and locally east of Socorro, representing a more distal facies; (3) Bruton Member, a moderately thick (< 85 m), dominantly marine lithofacies present east of Socorro, Cooke's Range and the Oscura, northern San Andres, Fra Cristobal and Caballo Mountains, representing the distal or basinal facies; and (4) Laborcita Member, a thick (< 330 m) unit dominated by nonmarine red-beds, basement-cobble conglomerates, shallow marine siliciclastics, bedded limestone and algal bioherms in the Sacramento Mountains.

Conodont biostratigraphy provides compelling evidence that Bursum Formation deposition was not simply driven by glacio-eustatic cyclicity but instead was subject to significant tectonic control (Lucas et al., 2013a). The sharp lithologic contact between underlying strata of various ages (Proterozoic basement, Gray Mesa Formation, Atrasado Formation and Bar B Formation) indicates a distinct unconformity caused by a tectonic pulse of the ancestral Rocky Mountain (ARM) deformation. There is also evidence for a regional unconformity at the base of the

overlying Abo Formation. This unconformity resulted from another tectonic pulse of ARM deformation and reactivation of basement uplifts that caused increased siliciclastic influx and deposition of nonmarine red beds of the Abo Formation across large parts of central New Mexico (Krainer and Lucas, 2002, 2004, 2009; Lucas and Krainer, 2003a, b).

Cycles, which are locally well developed within the Red Tanks and Bruton Member, i.e., in the more basinward facies of the Bursum Formation, may also have been influenced by glacio-eustatic sea-level fluctuations related to the Gondwana glaciation. However, conodont data from the Red Tanks Member at Carrizo Arroyo indicate the existence of unconformities at the base of depositional sequences in the lower part of the succession. These unconformities suggest syndepositional tectonic activity, indicating that a significant driving force of deposition

of the Bursum Formation was regional tectonic activity related to the ARM (see discussion by Krainer and Lucas, 2004, 2009).

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THE LOWER PERMIAN ABO FORMATION IN CENTRAL NEW MEXICO

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Abstract—In central New Mexico, the Lower Permian (middle-upper Wolfcampian) Abo Formation generally overlies the Upper Pennsylvanian (Newwellian) Bursum Formation disconformably and is conformably overlain by the Lower Permian (Leonardian) Yeso Group. The Abo Formation is 100-309 m thick and can be divided into two members: (1) lower, Scholle Member, 40-140 m thick, mostly mudstone; and (2) upper, Cañon de Espinosa Member, 58-256 m thick, characterized by numerous sheet-like beds of sandstone. Across central New Mexico, the Abo Formation is a relatively uniform unit that is almost entirely mudstone and sandstone; minor lithologies are shale, siltstone, calcrete and intraformational conglomerate.

Abo Formation vertebrate fossils are of Coyotean age (late Virgilian-late Wolfcampian), and regional correlations indicate that the Abo Formation in central New Mexico is of middle-late Wolfcampian age. Trace fossils from the Abo Formation are an ichnoassemblage of rhizoliths, arthropod locomotion and feeding traces and tetrapod footprints of the *Scoyenia* ichnofacies. Fossil plants from the Abo Formation are mostly conifers and belong to two paleofloras: (1) red siltstone assemblages that are of low diversity at any given sampling site, and almost always dominated by either conifers or the peltasperm *Supaia*; (2) green shale/siltstone assemblages, locally more diverse, and still dominated by conifers, but with a significant component of cordaitaleans, callipterid peltasperms, and a wide variety of wetland plants. Invertebrate body fossils are rare and are only known in a carbonaceous shale in the northern Sandia Mountains (conchostracans) and the green shale, estuarine facies of the Abo Formation in the Caballo Mountains, which yields gastropods and diverse bivalves, including euryhaline pectins and myalinids. Vertebrate body fossils are mostly localized in strata of the Scholle Member and are a pelycosaur-dominated assemblage that includes lungfishes, palaeoniscoids, temnospondyl and lepospondyl amphibians, and diadectomorphs.

The Abo Formation is composed of various conglomerate, sandstone, nodular limestone and mudstone lithofacies that can be combined into three principal architectural elements: (1) sandstone sheets formed by amalgamated channels or by relatively unchannelized flow; (2) sandstone lenses and bodies that represent fluvial channels; and (3) siltstone/mudstone with pedogenic limestone that represents deposits of floodplains. Abo deposition took place on an extensive alluvial plain in which well-defined, bedload river channels within extensive muddy floodplains were succeeded by sandstone sheets formed by low sinuosity river deposits subject to episodic avulsion and sheetflooding. This change in stratigraphic architecture can be attributed to tectonic changes in which falling base level (relatively rapid subsidence) during deposition of the lower Abo was followed by episodically stable base level (slower subsidence) during deposition of the upper Abo.

INTRODUCTION

In central New Mexico (Fig. 1), siliciclastic red beds of the Lower Permian Abo Formation are a conspicuous lithostratigraphic unit in the upper Paleozoic section. Well exposed in the mountain ranges and other uplifted areas that border the Rio Grande rift (Fig. 1), study of Abo Formation outcrops during the last two decades has resulted in a much more refined understanding of their stratigraphy, paleontology, microfacies, correlation and depositional environments (e.g., Mack et al., 1991, 1995, 2003; Berman, 1993; Lucas et al., 2005a, 2012 a, b; DiMichele et al., 2007; Krainer and Lucas, 2010). Here, we review our current understanding of the Abo Formation in central New Mexico, which refers primarily to Sandoval, Bernalillo, Valencia, Tarrant, Socorro and Sierra Counties (Fig. 1).

HISTORY OF STUDY

Stratigraphic study of the Abo Formation began with Lee (1909), who divided the Manzano Group of Herrick (1900) into the Abo, Yeso

and San Andreas (sic) formations (Fig. 2). Lee (1909) named the “Abo sandstone” for “Abo Canyon” at the southern end of the Manzano Mountains (just north of Abo Pass), describing it as “coarse-grained sandstone, dark red to purple in color and usually conglomeratic at the base, with a subordinate amount of shale....” (p. 12). Lee (1909) included marine limestone in the base of the Abo sandstone that eventually became part of the underlying Bursum Formation. He also included a sandstone-dominated interval in the top of the Abo that was later included in the overlying Yeso Formation (Fig. 2).

Darton (1928) used Abo sandstone as had Lee (1909), including some strata later termed Bursum in the basal Abo, and some strata later termed Yeso in the upper Abo (Fig. 2). Recognition of the Abo Formation across central New Mexico owes much to Darton’s (1928) work.

Needham and Bates (1943) described type sections of the Abo and Yeso formations of Lee, and thereby modified somewhat his concepts of those units. Of note, Needham and Bates (1943) excluded marine limestones from the basal Abo Formation, a decision followed by all subsequent workers (Fig. 2).



FIGURE 1. Map of New Mexico, showing locations of outcrops of the Abo Formation discussed in the text.

The longstanding concept of the Abo and Yeso formations in central New Mexico comes from four maps published between 1946 and 1951 by the United States Geological Survey as part of its Oil and Gas Mapping Programs carried out during World War Two. These included maps of the Jemez Pueblo and Sierra Nacimiento in Sandoval County (Wood and Northrop, 1946), the Lucero uplift in Valencia County (Kelley and Wood, 1946), the Joyita Hills, Los Piños Mountains and northern Chupadera Mesa of Socorro County (Wilpolt, et al., 1946) and the Cerrillos del Coyote, northern Jornada del Muerto and southern Chupadera Mesa in Socorro County (Wilpolt and Wanek, 1951). These workers followed Needham and Bates (1943) in their placement of the base of the Abo Formation. Wilpolt et al. (1946) named the Bursum Formation to encompass the mixed marine-clastic interval immediately below the Abo Formation, whereas Kelley and Wood (1946) named that interval the Red Tanks Member of the Madera limestone (Fig. 2). Most significantly, Wilpolt et al. (1946) removed the upper sandstone interval from the Abo Formation, and called it the Meseta Blanca sandstone member of the Yeso Formation (a name proposed by Wood and Northrop, 1946).

Baars' (1962) important synthesis of Lower Permian stratigraphy on the Colorado Plateau used Abo Formation in the same sense as the published maps of 1946. However, he pointed out that the type section of the Meseta Blanca Member of the Yeso Formation near Jemez Pueblo is simply the same unit that Gregory (1917) had named the De Chelly Sandstone near Chinle, Arizona. Therefore, Baars (1962) abandoned the name Meseta Blanca and replaced it with De Chelly Sandstone, which he removed from the Yeso Formation.

The recommendations of Baars (1962), however, were not heeded for more than 40 years by geologists working in New Mexico, who continued to use the Abo-Yeso stratigraphy of the 1946 published maps (e.g., Hatchell et al., 1982; Cook et al., 1998; Dinterman, 2001; Mack and Dinterman, 2002; Seager and Mack, 2003; Kues and Giles, 2004). Nevertheless, our own work (Lucas et al., 1999, 2005, 2009, 2012a, b; Lucas and Krainer, 2004; Lucas and Zeigler, 2004; DiMichele et al., 2007)

convinced us that some modifications of the Abo-Yeso stratigraphy were justified to produce a lithostratigraphic nomenclature that better reflects current understanding of the Lower Permian lithostratigraphy of central New Mexico. These modifications included recognizing two formal members within the Abo Formation, adopting Baars (1962) suggestion to abandon the name Meseta Blanca Member of Yeso Formation and replace it with DeChelly Sandstone, raising the Yeso Formation to group status and naming the siltstone-dominated facies of the lower Yeso south of the DeChelly erg the Arroyo de Alamillo Formation (Fig. 2).

STRATIGRAPHIC NOMENCLATURE

From the Abo type section in Valencia County southward into Socorro County, Lucas et al. (2005a) divided the Abo Formation into two members, a lower mudstone-dominated unit, the Scholle Member, and an upper unit with many sandstone sheets, the Cañon de Espinosa Member. The same subdivision of the Abo Formation is evident in Abo outcrops in the southern Jemez Mountains of Sandoval County (Lucas et al., 2012b), on the northern end of the Sandia uplift in Bernalillo County (Lucas et al., 1999), in the Lucero uplift of Valencia County (Lucas and Zeigler, 2004), in the Cerros de Amado-Joyita Hills area of Socorro County (Lucas et al., 2009), in the northern Oscura Mountains of Socorro County (DiMichele et al., 2007) and in the Fra Cristobal and Caballo Mountains of Sierra County (Lucas et al., 2012a). Therefore, we apply the member-level terminology of Lucas et al. (2005a) to the Abo Formation across central New Mexico (Fig. 3).

LITHOSTRATIGRAPHY

Contacts

Across central New Mexico, the Abo Formation generally overlies sedimentary rocks of Pennsylvanian age. Mostly, these are strata of the Virgilian-early Wolfcampian Bursum Formation, which are a succession of mixed carbonate and siliciclastic rocks of marine and nonmarine origin (e.g., Lucas and Krainer, 2004; Krainer and Lucas, 2009). Typically, the highest marine limestone bed underneath Abo red-bed siliciclastic strata is chosen as the Abo-Bursum contact. A similar criterion is used to distinguish the basal contact of the Abo Formation where it overlies older, Pennsylvanian marine rocks. In a few places that were positive areas (uplifts) of the Ancestral Rocky Mountain orogeny, the Abo Formation rests directly on Proterozoic basement. In central New Mexico, these are very localized areas in the Nacimiento uplift at the southern end of the Jemez Mountains (Woodward, 1987) and in the Joyita Hills of Socorro County (Kottlowski and Stewart, 1970).

Across central New Mexico, strata of the Lower Permian Yeso Group conformably overlie the Abo Formation. At most outcrops, there is either an interbedding of Abo and Yeso lithotypes over a stratigraphic interval of 10 m or less, or the base of the Yeso Group is picked at the lowest, non-arkosic, fine-grained sandstone bed or the lowest dolomite bed above Abo mudrock or arkosic sandstone (Fig. 4).

Thickness

The Abo ranges in thickness from about 100 to 300 m across central New Mexico (Table 1). The Scholle Member is 40 to 115 m thick, and the Cañon de Espinosa Member is 60 to 256 m thick. The Abo is thickest to the south, but there are no consistent thickness trends geographically. Instead, we suspect Abo thickness reflects the position of Ancestral Rocky Mountain uplifts, such as the relatively thin Abo section in the Cerros de Amado, just south of the Wolfcampian Joyita uplift.

Lithology

Scholle Member

The Scholle Member of the Abo Formation is mudstone dominated. Its sandstones are coarse-grained and conglomeratic and are trough-

Lee (1909)	Darton (1928)	Needham & Bates (1943)	Kelley & Wood (1946)	Wood & Northrop (1946)	Wilpolt et al. (1946)	Baars (1962)	this paper
San Andres limestone		San Andres Formation	San Andres Formation	San Andres Formation	San Andres Formation	San Andres Formation	San Andres Formation
		Glorieta sandstone	Glorieta sandstone mbr.	Glorieta sandstone mbr.	Glorieta sandstone mbr.	Glorieta Sandstone	Glorieta Sandstone
		Joyita Member			Joyita Member		Joyita Member
		Cañas Member			Cañas Member		Cañas Member
		unnamed member			Tomas Member		Tomas Member
Yeso Formation	Chupadera Formation	Yeso Formation	Yeso Formation	Yeso Formation	Yeso Formation	Yeso Formation	Yeso Group
			Los Valles Member	San Ysidro Member			Los Valles Formation
							San Ysidro Formation
			Meseta Blanca sandstone member	Meseta Blanca sandstone member	Meseta Blanca sandstone member	De Chelly Sandstone	Arroyo de Alamillo Formation
							De Chelly ls.
Abo sandstone	Abo sandstone	Abo Formation	Abo Formation	Abo Formation	Abo Formation	Abo Formation	Abo Formation
							Cañon de Espinoso Member
							Scholle Member
		unnamed basal Permian limestone	Red Tanks Member of Madera limestone	Madera limestone	Bursum Formation	Bursum Formation	Bursum Formation

FIGURE 2. Development of lithostratigraphic nomenclature of the Abo Formation and adjacent strata in central New Mexico.

crossbedded, channelform deposits (Fig. 4). As a good example, the Scholle Member is well exposed at its type section at Abo Pass (Fig. 4), where it is ~ 140 m thick and consists of thick mudstone slopes (87% of the measured section) broken by thin ledges of trough-crossbedded sandstone and conglomerate (11% of the section), and minor calcrete ledges (2% of the section).

Cañon de Espinoso Member

The Cañon de Espinoso Member, though it may also be mudstone dominated, has a significant component of siltstone, and its sandstones are sheet-like bodies with prominent ripple and climbing-ripple bedforms. At its type section at Abo Pass (Fig. 4), the Cañon de Espinoso Member is ~ 170 m thick and consists mostly of slope-forming covered intervals of mudstone (70% of the section), but includes significant siltstone beds (9% of the section) and many thin ledges of climbing-ripple laminated sandstone (21% of the section).

Lithofacies

Within the Abo Formation, Krainer and Lucas (2010) and Lucas et al. (2012a, b) described 17 lithofacies types following the classification and lithofacies codes of Miall (1978, 1981, 1985, 1996, 2010). All lithofacies types are listed in Table 2.

These Abo lithofacies can be used to distinguish three architectural elements: (1) sandstone sheets, (2) intercalated thin sandstone beds and lenses, and (3) siltstone-mudstone. Sandstone sheets are the most characteristic facies assemblage in the Abo Formation and correspond to the architectural element CH (channel) and SB (sandy bedforms) of Miall (1996). Sandstone sheets are up to several m thick and form distinct, resistant ledges that can be traced laterally over long distances. The base is commonly erosive; locally, channels are cut into the underlying fine-grained sediments with a relief of up to about 3 m.

Two types of sandstone sheets are observed. Sandstone sheets composed dominantly of lithofacies St, minor Sl, Sh, Sm and rare Sr correspond to architectural element CH. Intercalated thin conglomerate layers and lenses within stacked cosets of multistory channel-fill succes-

sions represent conglomerate lags at the base of individual channels. These sandstone sheets were deposited in broad, shallow channels of a low sinuosity river system (Krainer and Lucas, 2010).

The other type of sandstone sheet is dominated by lithofacies Sr, associated with Sh, Sm and rare Sl, representing the architectural element SB. Individual sandstone beds are 0.1-5 m thick. Different types of ripple lamination are observed, such as isolated asymmetric current ripples or thin layers of current ripples that occur within fine-grained sediment. In thicker sandstones, climbing ripples are common. Both type A (erosional-stoss) and type B (depositional-stoss) climbing ripples are observed. Draped lamination is very rare. Climbing ripples may grade upwards into horizontal lamination. Mudcracks are common within the ripple-laminated sandstone units.

The flow conditions under which different types of ripple drift cross lamination are formed are well known from flume experiments (e.g., Jopling and Walker, 1968; Allen, 1973, 1985; Banks and Collinson, 1975; Ashley et al., 1982). Different types of climbing ripples are attributed to fluctuations in current velocity, variations in grain-size and the concentration of suspended sediment (Jopling and Walker, 1968). Climbing ripple sequences 10-20 cm thick are deposited within a few tens of hours. Draped lamination results from continuous fallout of sediment from suspension after ripple migration ceases or almost ceases. Ripple drift cross lamination is very common in glaciofluvial and glaciolacustrine sediments (Jopling and Walker, 1968; Gustavson et al., 1975), but is also known from other depositional environments.

In ripple-laminated sandstones of the Abo Formation, variations in height between the climbing ripple sets as well as fluctuations in the climbing angle are observed. These patterns result from fluctuations in flow velocity or fluctuations in sediment transport volume (Rubin, 1987). The sandstone sheets of architectural element SB were deposited in very broad, shallow channels of a braided stream system during periods of high influx of fine sand. They are essentially sheetflood deposits in which random fluctuations in flow velocity and depositional rate caused compound cross-bedding.

Sandstone beds and lenses are another prominent architectural element in the Abo Formation in central New Mexico. Thin intercalated

Chama Basin		Jemez Springs		Cerro de Amado		Caballo Mountains		Robledo Mountains		AGE	
Yeso Group (De Chelly Sandstone)		Yeso Group (De Chelly Sandstone)		Yeso Group (Arroyo de Alamillo Formation)		Yeso Group (Arroyo de Alamillo Formation)		Apache Dam Formation		Leonardian	
Cutler Group	Arroyo del Agua Formation	Abo Formation	Cañon de Espinoso Member	Abo Formation	Cañon de Espinoso Member	Abo Formation	Cañon de Espinoso Member	Hueco Group	Robledo Mountains Formation	late (Lenoxian)	Wolfcampian
	El Cobre Canyon Formation		Scholle Member		Scholle Member		Scholle Member		Community Pit Formation		
				Bursum Formation	Bursum Formation	Horquilla Formation			early (Newwellian)		
			Guadalupe Box Formation			Atrasado Formation			Virgilian		

FIGURE 3. Correlation chart of Abo Formation and related stratigraphic units from north to south in New Mexico.

beds and lenses of coarse-grained siltstone to fine-grained sandstone are 0.1 to 0.5 m, rarely up to 1 m thick. They occur as single sandstone beds or as stacked units. Typical lithofacies are Sr, Sh, St and Sm, forming tabular or lense-shaped sandstone bodies. Krainer and Lucas (2010) interpreted the tabular sandstone beds as sheet splay deposits formed by sheet-like, non-channelized flow from a crevasse channel onto the floodplain, thus representing architectural element CS (crevasse splay) of the overbank environment (Miall, 2010). Sandstone lenses are interpreted as minor channel fills representing feeder channels (crevasse channels, architectural element CR of Miall, 2010) of the sheet splays. Such intercalated sandstone beds and lenses of limited lateral extent are characteristic elements of the overbank environment.

Siltstone-mudstone is the third principal architectural element of the Abo Formation. Indeed, this lithology is the dominant lithofacies of the Abo Formation across central New Mexico, occurring as units up to tens of m thick that extend laterally over large distances. This facies is interpreted to represent floodplain deposits (architectural element FF – floodplain fines according to Miall) of the overbank environment resulting mainly from overbank sheet flow and from deposition from suspension during waning flood. Absence of lamination partly results from bioturbation. Intercalated limestone beds are interpreted as pedogenic limestones, indicating relatively long periods of soil formation under dry conditions (Mack and James, 1986; Mack et al., 1991).

Sandstone Petrography

To the north in central New Mexico (Jemez Mountains, Abo Pass), sandstone of the Abo Formation is classified as arkose (Fig. 5A, B, D, E, F) and lithic arenite (Fig. 5 G-H), subordinately as subarkose (Fig. 5C). The dominant grain type is monocrystalline quartz, and subordinate is polycrystalline quartz. Detrital feldspars, mostly potassium feldspars, including many microcline are common. Coarse-grained sandstone contains many granitic rock fragments; metamorphic rock fragments are

rare. Individual sandstone layers contain sedimentary rock fragments, mainly carbonate grains (Fig. 5G-H), and subordinately reworked mudstone-siltstone grains are present. Sandstone contains small amounts of matrix (Fig. 5B). Quartz cement is present in many sandstones and occurs as authigenic overgrowths on detrital quartz grains (Fig. 5C-D). Most sandstones are cemented by coarse blocky calcite (Fig. 5A, C; see details in Krainer and Lucas, 2010).

In central New Mexico, towards the south (Fra Cristobal Mountains, Caballo Mountains) grain size decreases and sandstone is fine-grained. The decrease in grain size causes changes in composition: the amount of monocrystalline quartz is higher, whereas polycrystalline quartz and detrital feldspars are less abundant, and granitic and metamorphic rock fragments are almost absent (Lucas et al. 2012a). Due to the composition, sandstone is subarkose.

Clearly, the main source rock of the Abo Formation sandstones was granitic in composition; subordinately metamorphic rocks were reworked. Sedimentary rock fragments in the Abo Formation are mainly derived from intrabasinal reworking of pedogenic carbonate horizons (caliche).

Distribution

Our description of the Abo Formation relies primarily on seven outcrop areas in central New Mexico (Fig. 1). Clearly, the Abo Formation was deposited across all of central New Mexico, because it is always present on outcrop or in the subsurface unless it was removed by post-Early Permian erosion. Here, we briefly describe the key Abo outcrop areas in central New Mexico.

Jemez Mountains

The Abo Formation crops out extensively in the southern Jemez Mountains, particularly along the canyons formed by the Jemez River and its tributaries (Wood and Northrop, 1946; Osburn et al., 2002; Kelley

Abo Type Section

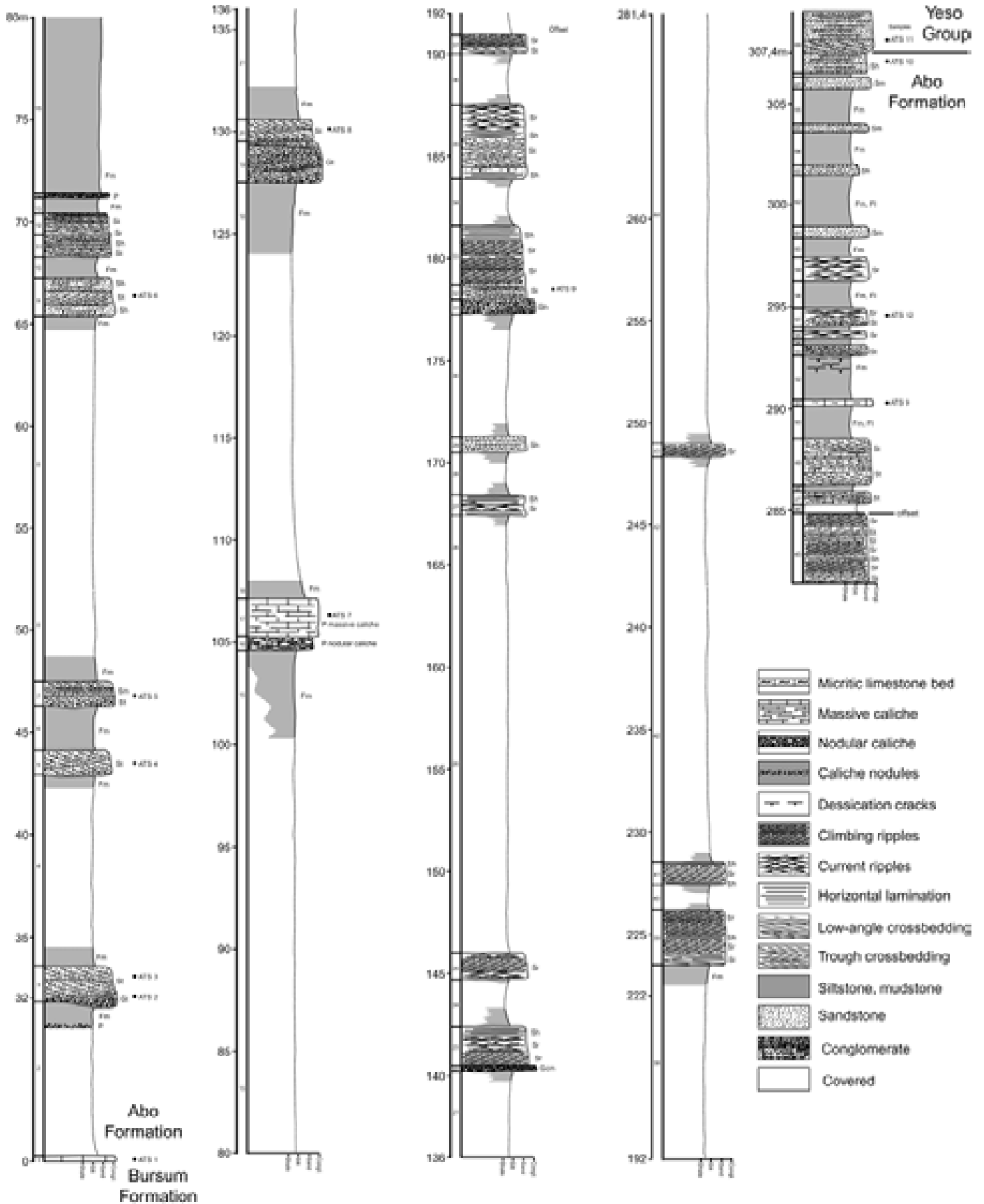


FIGURE 4. Type section of the Abo Formation at Abo Pass (from Lucas et al., 2005a).

TABLE 1. Thickness (in meters) of Abo Formation and members of Abo Formation at selected outcrop belts in central New Mexico.

Outcrop area	Scholle Member	Cañon de Espinoso Member	Total Abo thickness
Jemez Mountains	90-115	60-75	150-190
Sandia Mountains	40-50	60-70	100-120
Abo Pass	140	170	310
Lucero uplift	77	73	150
Cerros de Amado	42	58	100
Oscuro Mountains	77	79	156
Fra Cristobal Mountains	40	254	294
Caballo Mountains	83-75	203-256	278-309

et al., 2003). Here, it is as much as 190 m thick and readily divided into the Scholle and Cañon de Espinoso members. The Abo Formation in this area also yields extensive assemblages of vertebrate fossils and a diverse paleoflora from one locality (Lucas et al., 2012b).

Sandia Mountains

An ~ 120-m thick section of the Abo Formation crops out on the northern end of the Sandia uplift, near Placitas (Kelley and Northrop, 1975; Lucas et al., 1999). In the Scholle Member here, a thin lens of carbonaceous shale yields conchostracans and a paleoflora (Knaus and Lucas, 2005; Martens and Lucas, 2005), and other beds yield a sparse assemblage of bones and pelycosaur footprints (Lucas et al., 1999).

Abo Pass

The type section of the Abo Formation and its two named members is located in Abo Pass, along the boundary of Valencia and Torrance Counties at the southern end of the Manzano Mountains. Here an ~ 190 m thick section of Abo Formation is exposed (Lee, 1909; Hatchell et al., 1982; Lucas et al., 2005a; Oviatt, 2010, 2011) (Fig. 4). Extensive fossil footprint assemblages, typically walchian-dominated plant assemblages and less prolific bone assemblages are present in the Abo here (Berman, 1993).

Lucero uplift

In the vicinity of Carrizo Mesa in the Lucero uplift of Valencia County, an ~ 150-m thick section of the Abo Formation crops out (Kelley and Wood, 1946; Lucas and Zeigler, 2004). These strata yield important footprint assemblages (Lucas et al., 2004).

Cerros de Amado-Joyita Hills

East of Socorro, a relatively thin section of Abo Formation (~ 100 m thick) crops out on numerous fault blocks in the complex structural zone extending from the Joyita Hills through the Cerros de Amado. Both members of the Abo are readily recognized here and yield important fossil assemblages of traces, plants and vertebrates (e.g., Hunt, 1983; DiMichele et al., 2007; Lucas et al., 2009; Spielmann et al., 2009)

Northern Oscura Mountains

Near Bingham in Socorro County, an extensive outcrop belt of Abo Formation is present between the Pennsylvanian strata on the northern dip slope of the Oscura Mountains and the Yeso Group strata that form the flanks of Chupadera Mesa to the north (Wilpolt et al., 1946). Here, the Abo Formation is about 150 m thick and readily divided into the Scholle and Cañon de Espinoso members. Strata of the Cañon de

Espinoso Member yield important fossil assemblages of footprints and plants (DiMichele et al., 2007; Lucas and Spielmann, 2009).

Fra Cristobal and Caballo Mountains

Strata of the Abo Formation exposed in the Fra Cristobal and Caballo Mountains of Sierra County can be assigned to the Scholle and overlying Cañon de Espinoso members. At about 300 m thick, these are the thickest Abo sections in central New Mexico and yield important assemblages of trace fossils, fossil plants and tetrapod bones (Lucas et al., 2012a). To the south of the Caballo Mountains, the Abo Formation disappears into the marine strata of the Lower Permian Hueco Group (Fig. 3). The shoreline of the Hueco seaway was for a time as far north as the Caballo Mountains, as evidenced by marine and estuarine facies in the lower part of the Abo Formation in the Derry and McLeod Hills (Lucas et al., 2012a).

PALEONTOLOGY AND AGE

Introduction

Trace fossils, fossil plants, invertebrate body fossils (mostly bivalves and conchostracans) and vertebrate bones and teeth have been recovered from the Abo Formation in central New Mexico. Here, we briefly review these fossil records.

Trace Fossils

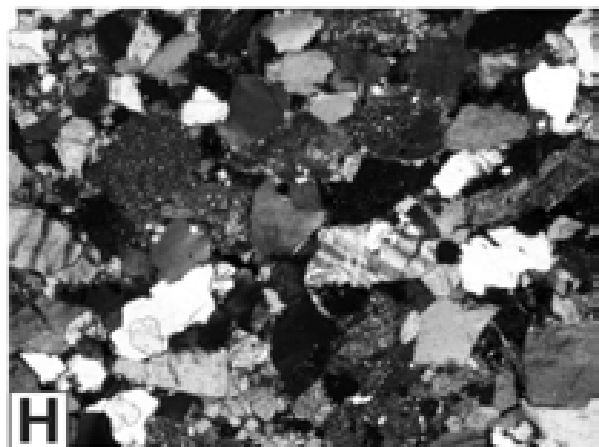
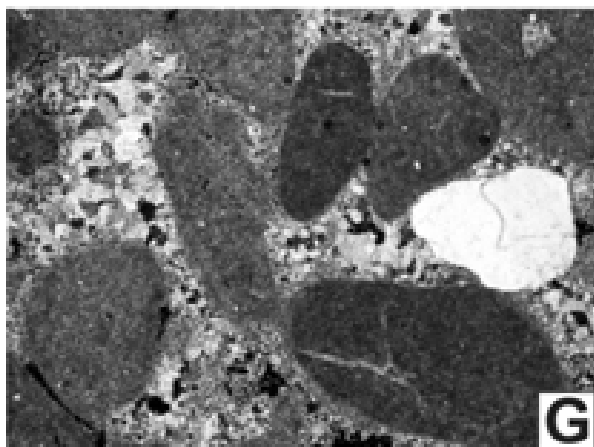
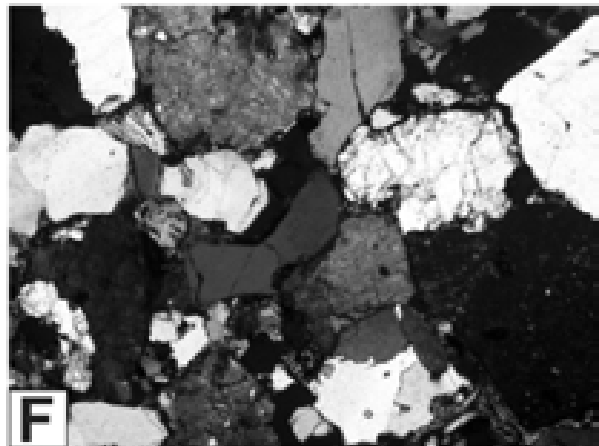
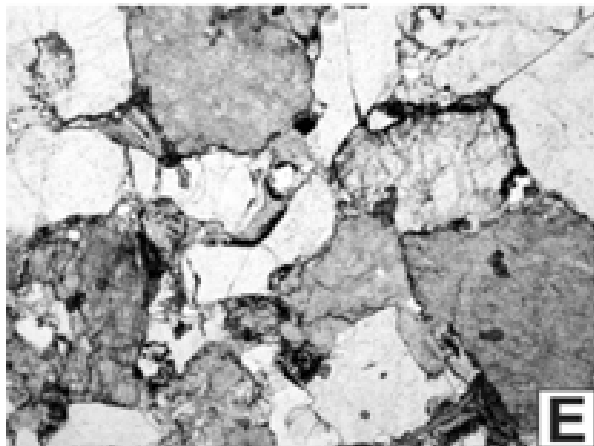
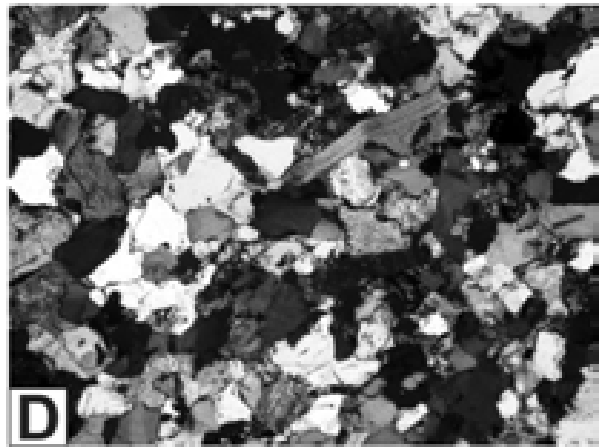
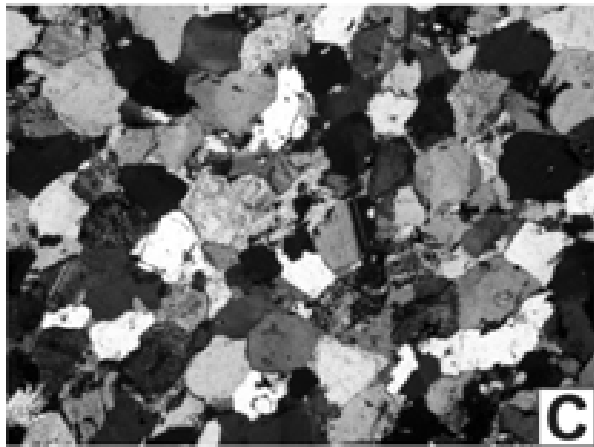
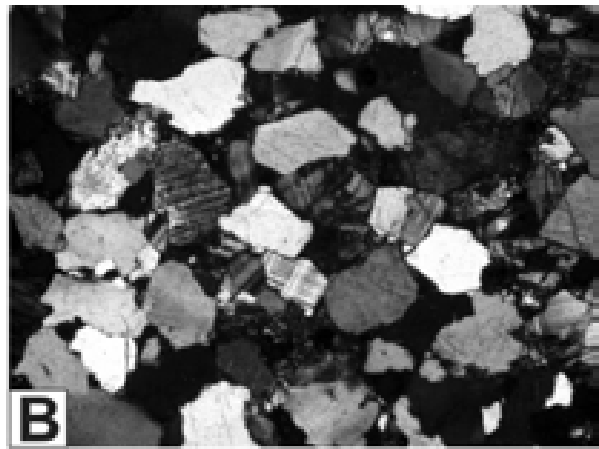
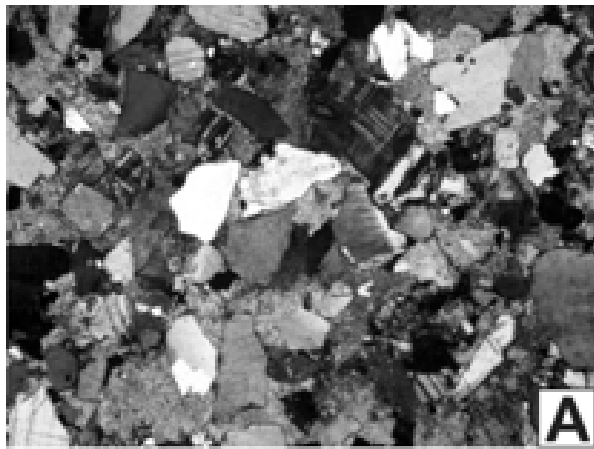
Traces are among the most common fossil remains in the red beds of the Abo Formation in central New Mexico. The collection of the New Mexico Museum of Natural History (NMMNH) houses about one thousand catalogued specimens of ichnofossils from more than 80 localities in the Abo Formation of the Sandia Mountains (15 specimens/6 localities), Abo Pass (105/8), Lucero uplift (44/10), Socorro area (Cerros de Amado and Joyita Hills, 267/20), northern Oscura Mountains (74/15), Fra Cristobal Mountains (99/3), and Caballo Mountains (366/22). Although Early Permian trace fossils from this part of the state have long been known (Degenhardt, 1840; Branson and Branson, 1946), significant work on them was not done before the 1990s (Lucas and Heckert, 1995). Since then, trace fossils from the Abo Formation of central New Mexico claim an almost continuously growing interest with regard to studies in ichnotaxonomy, biostratigraphy and paleoenvironmental analysis (e. g., Hunt and Lucas, 1998; Lucas et al., 1999, 2001, 2004, 2005a, b, 2009, 2012a; Lucas and Lerner, 2004; Hunt et al., 2005a-d; Minter et al., 2008; Lucas and Spielmann, 2009).

The Abo Formation in central New Mexico contains an ichnoassemblage of rhizoliths, invertebrate traces, and tetrapod tracks and coprolites (Fig. 6). Traces occur on clayey mud-draper within reddish-brown, horizontally laminated, flaser-bedded or small-scale trough cross-bedded siltstone to very fine-grained sandstone. They are commonly accompanied by inorganic sedimentary surface structures such as tool marks, raindrop impressions, mud-cracks, and wrinkle marks. Root traces (Fig. 6A) are the most abundant biogenic structures and are preserved on approximately one third of the collected specimens. The record of fossil root traces is dominated by reddish-brown, mud-filled cylindrical tubes that measure less than 2 mm in cross section and run parallel to the bedding plane or dissect it at angles up to 90 degrees. Horizontal root traces may exhibit orthogonal dichotomous branching. Unfortunately, to date there is no reliable organ relationship between root traces and macrophytic remains from Abo red beds.

The Abo Formation in central New Mexico yields a diverse assemblage of moderately common invertebrate traces. Characteristic ichnotaxa include *Augerinoichnus helicoidales* (Fig. 6B), *Lithographus hieroglyphicus* (Fig. 6C), *Sphaerapus larvalis* (Fig. 6D), *Stiaria intermedia* (Fig. 6E), and *Tonganoxichnus robledoensis* (Fig. 6F), which are interpreted as locomotion and feeding traces of vermiform animals as well as pterygote and apterygote insects (Minter et al., 2008; Minter and Braddy,

TABLE 2. Lithofacies types of the Abo Formation (modified from Krainer and Lucas, 2010).

Facies code	Facies	Sedimentary structures	Interpretation
Gcm	clast-supported, poorly sorted conglomerate, sandy matrix, abundant reworked limestone clasts, subord. extraformational clasts and mudstone clasts	crude stratification, erosional base, commonly grades into St	lag deposits
Gt	poorly sorted conglomerate, sandy matrix, abundant reworked limestone clasts, also extraformational clasts	trough crossbedding	channel fill
Se	pebbly sandstone, abundant rip-up clasts, rare	crude stratification, erosional base	scour fill
St-l	coarse, pebbly sandstone	large-scale trough crossbedding (> 1 m)	lateral accretion
St-m	coarse to fine-grained sandstone	medium-scale trough crossbedding (> 50 cm)	lower flow regime, 3d-dunes
St-s	medium to fine-grained sandstone	small-scale trough crossbedding (< 50 cm)	lower flow regime, 3d-dunes
Sp	medium to fine-grained sandstone	planar crossbedding	bars, sand waves
Sl	coarse to medium-grained sandst.	low-angle crossbedding	upper flow regime
Sh	mostly fine- to medium-grained sandstone	horizontal lamination	?upper flow regime
Sm	coarse- to fine-grained sandstone	massive, no bedding	sediment gravity flow deposit
Sr	fine-grained sandstone	asymmetric ripples, ripple cross-lamination (< 5 cm)	lower flow regime, current ripples
Sb	medium- to fine-grained sandst.	bioturbation structures	bioturbated
Pc	micritic limestone	nodular	paleocaliche
Fsm	mudstone - siltstone	massive	suspension deposit
Fl	mudstone - siltstone	laminated	overbank, waning flood deposit
Fm	mudstone - siltstone	massive, desiccation cracks	overbank deposit
Fr	mudstone - siltstone	massive, root structures	root horizon



2009; Lucas et al., 2013b). Other recorded invertebrate traces from the Abo red beds of the study area are *Arborichnus*, *Cochlichnus*, *Diplichnites*, *Diplopodichnus*, *Kouphichnium*, *Monomorphichnus*, *Palaeophycus*, *Protovirgularia*, *Scoyenia*, *Skolithos*, *Stiallia* and *Striatichnium* (Lucas and Lerner, 2004; Minter and Braddy, 2009; Lucas et al., 2012a; SV pers. obs.). Mack et al. (1995) identified invertebrate burrows from Abo red beds of the Caballo Mountains as *Arenicolites* that, however, belong to *Scoyenia* and *Skolithos* (Lucas et al., 2012a). The diverse arthropod trace fossil assemblage described by Minter and Lucas (2009) from the Abo Formation in the Joyita Hills actually comes from strata of the Arroyo de Alamillo Formation of the Yeso Group based on our recent reevaluation of its stratigraphic position (Lucas et al., 2013a).

Tetrapod tracks from the Abo Formation of central New Mexico are relatively diverse and remarkably abundant (Hunt et al., 1995, 2005a, d; Lucas et al., 1995, 1999, 2001, 2004, 2005a, b, 2009, 2012a; Lucas and Spielmann, 2009). According to a comprehensive revision (SV, unpublished) they can be assigned to *Batrachichnus salamandroides* (Fig. 6G), *Limnopus vagus*, *Amphisauropus kablikae* (Fig. 6H), *Ichniotherium cottae*, *Dimetropus leisnerianus*, *Tambachichnium schmidti*, *Varanopus* sp., *Hyloidichnus bifurcatus* (Fig. 6I), *Erpetopus willistoni*, and *Dromopus lacertoides* (Fig. 6J). More than 80% of all finds belong to *B. salamandroides* and *D. lacertoides*, which most likely represent tracks of small temnospondyls and araucoscelid diapsids (Lucas et al., 2005a, b; Haubold and Lucas, 2003; Voigt, 2005). *Limnopus*, *Amphisauropus*, *Ichniotherium*, *Dimetropus*, *Tambachichnium*, *Varanopus*, *Hyloidichnus*, and *Erpetopus* are supposed tracks of large temnospondyls (Haubold, 1996), seymouriamorphs (Haubold, 2000; Lucas et al., 2001; Voigt, 2005), diadectomorphs (Voigt et al., 2007), “pelycosaurian-grade” synapsids (Haubold, 2000; Voigt, 2005), and captorhinomorphs (Haubold, 2000; Voigt et al., 2009) and are relatively rare. Vertebrate coprolites are well documented from one Abo Formation locality in Socorro County (Cantrell et al., 2012).

Though the trace fossils of the Abo Formation in central New Mexico exclusively represent assemblages of the *Scoyenia* ichnofacies, their lateral and vertical distribution is not necessarily uniform. Invertebrate traces are most common and diverse in the south and near the base and top of the formation (Lucas et al., 2004, 2005a, b, 2012a, 2013b). Seymouriamorph and diadectomorph footprints (*Amphisauropus*, *Ichniotherium*) are most common or even restricted to the northern and central part of the study area (Hunt and Lucas, 1998; Lucas et al., 2001, 2009; Hunt et al., 2005b, d; Voigt and Lucas, 2012). These observations may reflect paleogeography and paleoenvironment as invertebrate traces are apparently most diverse and abundant in coastal plains and tidal flats (Minter and Braddy, 2009; Lucas et al., 2012a; Voigt et al., 2013) and, diadectomorph tracks especially appear to indicate inland paleoecosystems (Hunt and Lucas, 1998; Hunt et al., 2005b; Voigt and Lucas, 2012).

The ichnofauna of the Abo Formation in central New Mexico is of Early Permian age based mainly on some of the tetrapod ichnotaxa. *Amphisauropus* and *Ichniotherium* co-occur exclusively in Cisuralian deposits (Haubold, 2000; Voigt, 2005). This age is confirmed by the

relative abundance of “pelycosaurian”-grade tetrapod tracks (*Dimetropus*, *Tambachichnium*). A significant change in the tetrapod ichnofauna has recently been recorded for the uppermost part (0-6 m below the top) of the Abo Formation, where supposed captorhinomorph footprints (*Varanopus*, *Hyloidichnus*, *Erpetopus*) are dominant and even more common than the otherwise ubiquitous *Batrachichnus* and *Dromopus* (Lucas et al., 2013a; pers. obs. SV). The first appearance of the invertebrate ichnotaxon *Sphaerapus* in the upper third of the Abo Formation also may be stratigraphically significant; this appearance may reflect an evolutionary innovation in the ecology of beetle larvae (Lucas et al., 2013b).

Fossil Plants

Fossil plants (particularly conifer impressions) were observed at many Abo Formation outcrops in central New Mexico, particularly in red sandstones/siltstones (Figs. 7-8). The collections can be divided broadly into two groups, based on both facies and floristic content.

The most common of these facies consists of thinly bedded red sandstones/siltstones. Such deposits frequently contain both plant debris and trackways of vertebrates and invertebrates. In the Caballo Mountains, as elsewhere in the outcrop area of the red sandstone/siltstone facies of the Abo, these fossiliferous beds tend to be sheet sandstones tens to hundreds of meters in areal extent and of varying thickness, from a meter to several meters. They generally fine upward, with the fossiliferous intervals in the upper portions, where the plant remains typically occur on thin claystone drapes or interbeds between thin layers of siltstone or sandstone (Lucas et al., 2005b,c). At many sites in the red siltstone facies, plants occur often in association with vertebrate and invertebrate trackways (e.g., Lucas et al., 2005b, c). In such settings, the enclosed paleoflora appears to be parautochthonous to allochthonous.

The flora of the Abo Formation red beds has been collected for many years by field geologists, and large collections are extant in both the New Mexico Museum of Natural History and Science, and the National Museum of Natural History. However, few papers have been published characterizing the flora, primarily because it is preserved mainly as casts and molds, thus not particularly suitable for studies of plant anatomy and morphology, at least with traditionally applied techniques. Nonetheless, preservation is satisfactory for systematic analyses of the flora (Hunt, 1983; Lucas et al., 2009; DiMichele et al., 2013). Our study, of 172 sample sites, indicates that the flora has the following composition:

1. Conifers: present in 82% of the sites, the most common being *Walchia piniformis* (Figs. 7.1 and 7.2)
2. Supaoid peltasperms: present in 28% of the sites, consisting of two species, *Supaia thinfeldioides* (25% of the sites) (Fig. 7.3) and *Supaia anomala* (6% of the sites).
3. Callipterid peltasperms: present in 12% of the sites, primarily represented by *Autunia conferta* (Fig. 7.1).

Rare elements include a *Glossopteris*-like plant from the Zuni Mountains in northern New Mexico, the possible cycadophyte *Taeniopteris*, *Dicranophyllum*, a coniferophyte (Figure 6.2), *Cordaites*, the foliage of often very large, conifer-like woody plants, and scattered occurrences of the tree fern *Pecopteris*, and the sphenopsids *Calamites*

FIGURE 5 (facing page). Abo Sandstone petrography. **A**, Arkosic sandstone composed of mono- and polycrystalline quartz, detrital feldspars and rare granitic rock fragments. The detrital grains are cemented by calcite. Sample JR 17, Jemez River, polarized light, width of photograph is 6.3 mm. **B**, Arkosic sandstone composed of detrital quartz grains, many detrital feldspars, rare granitic rock fragments and some matrix. Calcite cement is absent. Sample GI 23, Gilman, polarized light, width of photograph is 6.3 mm. **C**, Subarkose containing quartz grains, some detrital feldspars and rare granitic rock fragments. Grains are cemented by quartz which occurs as overgrowths on detrital grains and coarse calcite cement. Sample ATS 3, type section at Abo Pass, polarized light, width of photograph is 3.2 mm. **D**, Arkosic sandstone containing detrital quartz and feldspar grains, rare granitic rock fragments and mica. Grains are cemented by quartz in form of authigenic overgrowths. Sample ATS 4, type section at Abo Pass, polarized light, width of photograph is 3.2 mm. **E-F**, Coarse-grained arkosic sandstone containing abundant detrital potassium feldspars which due to slight alteration appear as brownish grains under plane light (**F**). The sandstone contains some matrix but lacks calcite cement. Sample ATS 5, type section at Abo Pass, **E** under polarized light, width of photographs is 3.2 mm. **G**, Coarse-grained sandstone (lithic arenite) containing well rounded sedimentary rock fragments (micritic carbonate grains) and rare detrital quartz, cemented by calcite. Sample GI 17, Gilman, polarized light, width of photograph is 6.3 mm. **H**, Sandstone composed of detrital quartz, feldspar, sedimentary rock fragments (fine-grained carbonate grains) and rare granitic rock fragments, cemented by calcite. Sample GI 15, Gilman, polarized light, width of photograph is 6.3 mm.

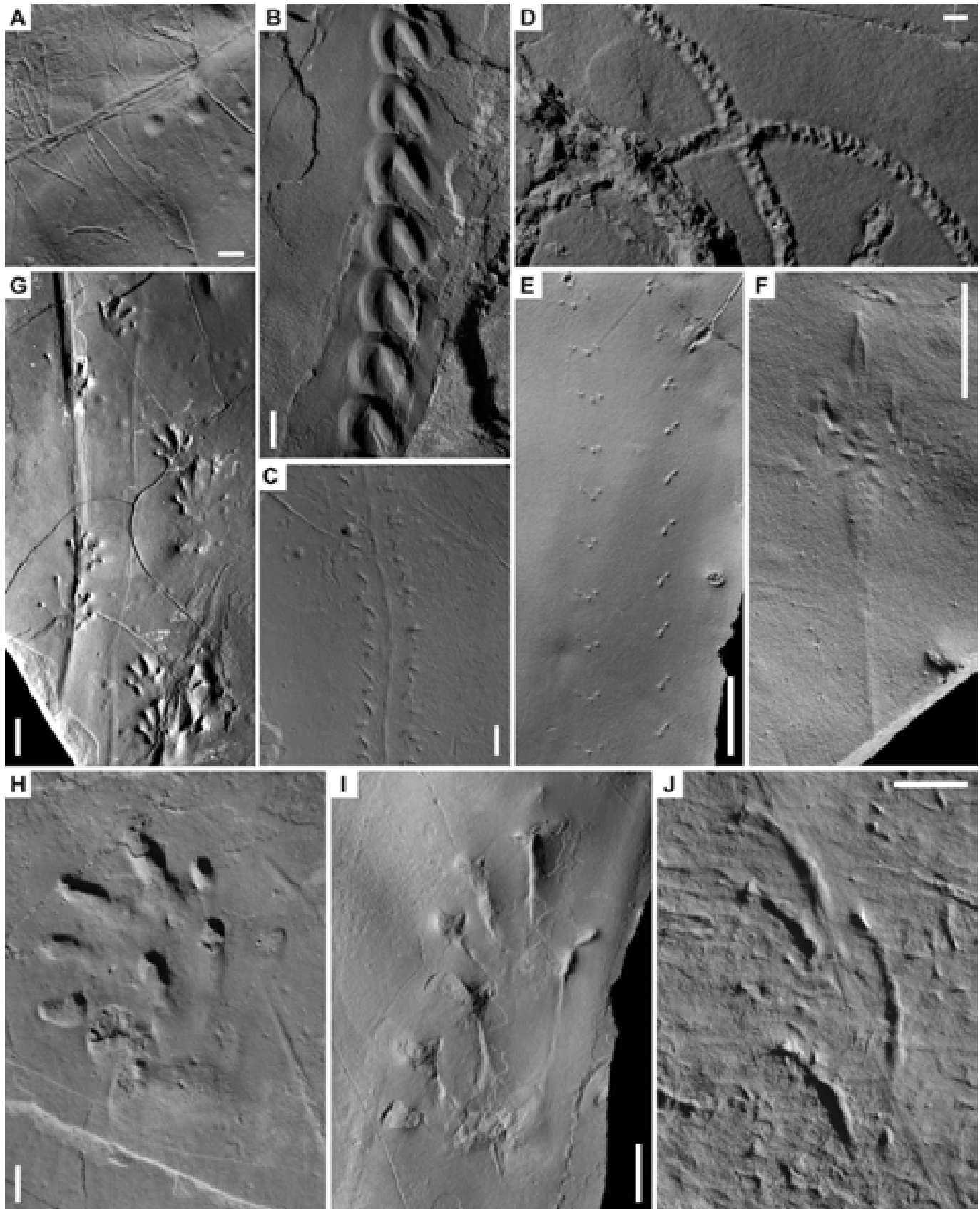


FIGURE 6. Selected trace fossils from the Abo Formation in central New Mexico. **A**, Root traces, NMMNH P-66088. **B**, *Augerinoichnus helicoidales*, NMMNH P-25990. **C**, *Lithographus hieroglyphicus*, NMMNH P-66598. **D**, *Sphaerapus larvalis*, NMMNH P-66550. **E**, *Stiaria intermedia*, NMMNH P-45778. **F**, *Tonganoxichnus robledoensis*, NMMNH P-65064. **G**, *Batrachichnus salamandroides*, NMMNH P-40901. **H**, *Amphisauropus kablikae*, NMMNH P-31338. **I**, *Hyloidichnus bifurcatus*, NMMNH P-65088. **J**, *Dromopus lacertoides*, NMMNH P-65114. Scale bars equal 10 mm.

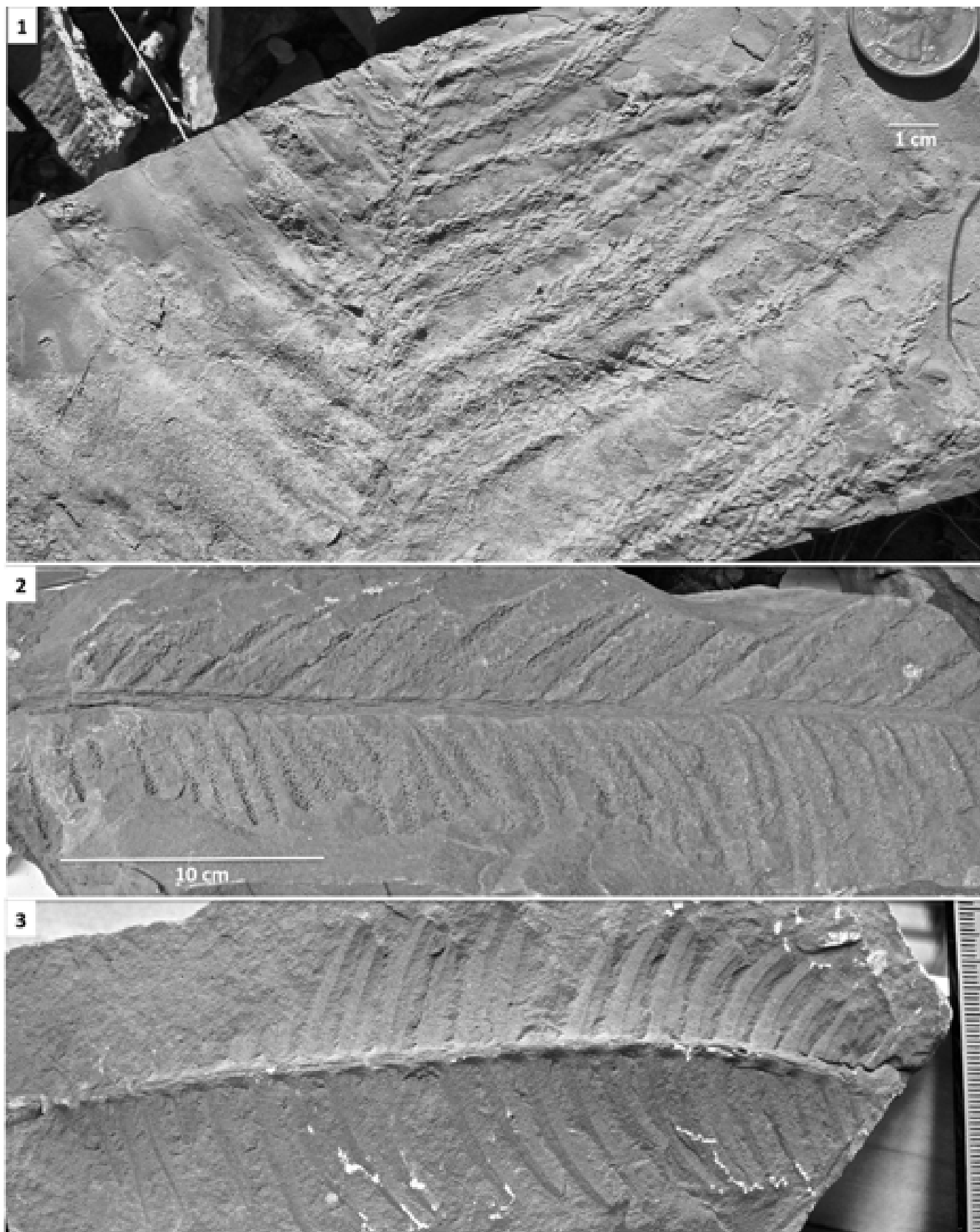


FIGURE 7. Principal Abo plants, preserved as molds or sediment casts of molds. Clearly the plant parts were quite stiff, perhaps dried out, at the time of deposition. **1**, *Walchia piniformis*, portion of a large lateral branch bearing the typically needle-like, upwardly curving leaves. Note that part of the branch is buried in a fine claystone. Field photograph, USNM Specimen 558250, USNM locality 43446. **2**, *Walchia piniformis*, nearly intact lateral branch mold. USNM Specimen 558251, USNM locality 43557. **3**, *Supaia thinfeldtioides*, one-half of a small forked frond showing the long, typically pointed, pinnules. USNM specimen 558252, USGS locality 8977. All scales in centimeters.

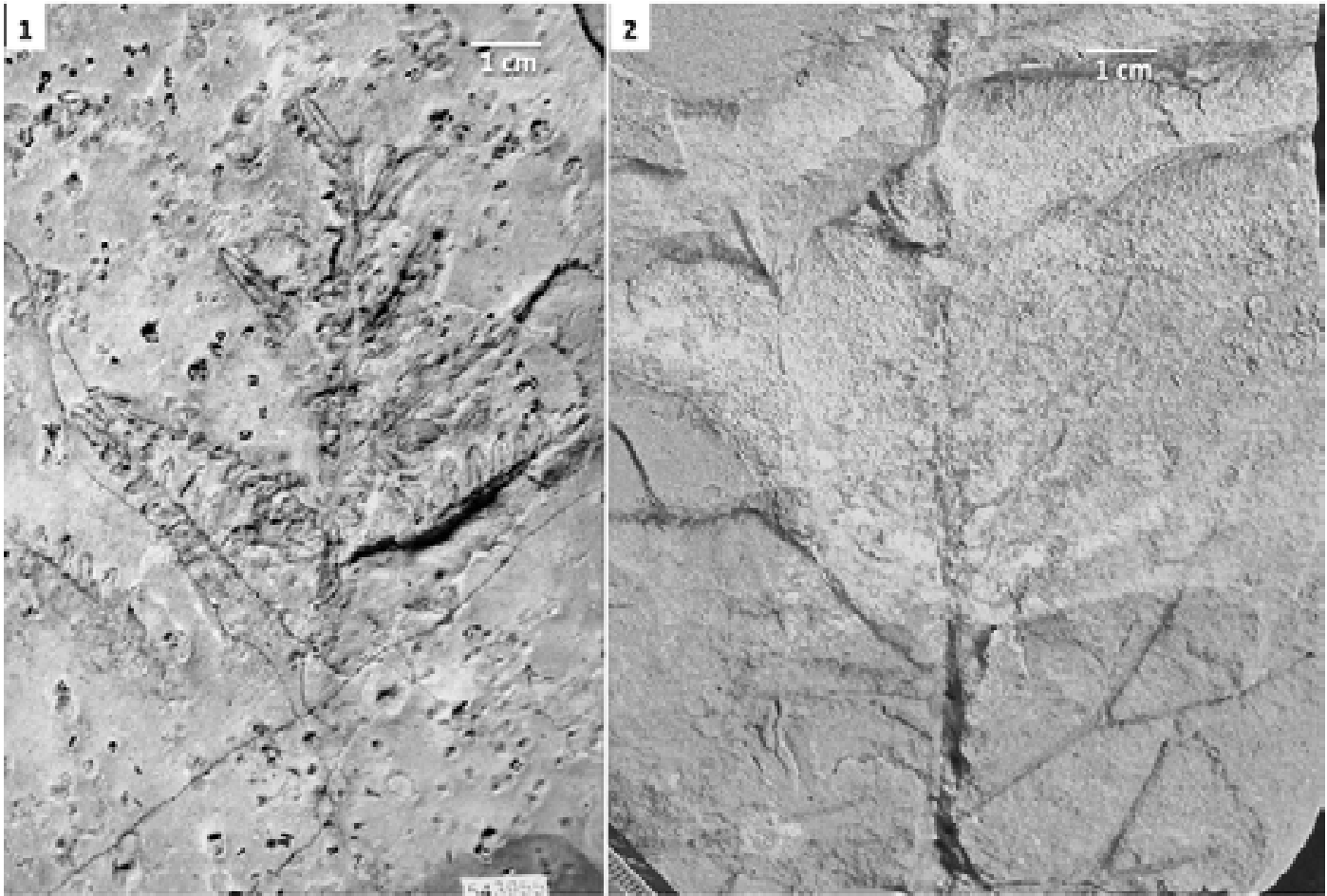


FIGURE 8. Minor Abo plants. Preserved as molds. 1, *Autunia conferta*, a callipterid peltasperm. This species was, for many years, considered to be a diagnostic fossil for the Permian. USNM Specimen 543955, USNM locality 42251. 2, *Dicranophyllum*, a coniferopsid. Note the forked leaves at the lower right of the photograph. USNM Specimen 558253, USNM locality 43614. Scales in centimeters.

and *Sphenophyllum*, the pteridosperm *Neuropteris*, and fern foliage *Sphenopteris*. With the exception of these last four taxa, which indicate substrates with high water tables for much of the year, the floral composition is typical of environments with seasonal moisture deficits. Furthermore, it is most common to find either conifers or supaioids as the dominant plant, and only rarely to find them in mixtures, though such mixtures do occur. DiMichele et al. (2012) suggest *Supaia thinnfeldioides* to be a colonizer of stream margins, in areas with high levels of flood disturbance.

Much less common, and found mainly in the lowermost portion of the Abo, are greenish gray to light gray shales/siltstones (the green/gray shale facies discussed above), most often with poorly-developed bedding, and fragmentary, often jumbled, variably dense plant material that appears allochthonous. Such deposits contain occasional evidence of rooting and may be mottled, as if overprinted by pedogenesis. On outcrop, most of these deposits are lenticular and appear to be channel fills. Some have conglomeratic layers at the base. The green shale/siltstones appear to be quite distinct depositionally from the red sandstones/siltstones, representing shallow channel bodies in coastal plain settings, with fresh to brackish salinities. The plants are intimately associated with invertebrates (see below), including pectinaceans and myalinids, other pelecypods and microchonchids, and also with fragmentary remains of fish.

The gray-bed paleoflora is superficially similar to that of the red beds in being dominated by conifers. Analysis of 30 sampling sites indicates that conifers occur at 73%, the most commonly occurring being, as in the red beds, *Walchia piniformis*. However, the flora at any given gray-

bed site tends to be more diverse than those in the red beds, the gray-bed sites on-average yielding about 5 taxa per excavation. These accessory taxa include the cordaitalean foliage, *Cordaites* at 40% of the sampling sites, a diversity of callipterid peltasperms at 37% and a variety of medullosan pteridosperms at nearly 30% of the sites (including, among others, *Neuropteris*, *Neurodopteris*, *Odopteris*, and *Alethopteris*). Other notable elements, because they are so much more widely occurring than in the red beds, include the tree-fern foliage *Pecopteris* (20%), *Taeniopteris* spp. (13%), the wetland lycopsid *Sigillaria brardii* (13%), and the peculiar lyginopterid pteridosperm *Sphenopteridium manzanitanum* (10%), a dominant element in the late Missourian flora of the Kinney Brick Company Quarry (Lucas et al., 2011c).

In summary, Abo paleofloras suggest conifer-dominated landscapes, containing a number of different species and genera of conifers. However, the red and gray facies present quite different pictures of this landscape and may represent different species-pools, with overlap only among the most abundant of those species. Most of the subdominant plants in the red beds are indicators of seasonality with only small local areas of wetlands on the landscape. The Abo red beds flora almost completely lacks ground cover or understory elements, which are expected to be greatly under-represented in the fossil record (Scheiing, 1980), due to their low growth habit or shelter from winds. In contrast, the gray beds capture a large component of wetland species, suggesting wetland corridors, perhaps fringing the channels in which these floras are preserved, surrounded closely by much drier interfluves, probably under a seasonal climate. Such deposits also include some ground cover elements.

The differences in composition and diversity between the red and gray facies emphasize the peculiarity of the extremely low diversity in the much more widely occurring and often densely fossiliferous Abo Formation red-bed plant deposits. These compositional and diversity differences are not easily attributable to some kind of taphonomic filtering. Such an interpretation might suggest that the red beds preserve only the most decay resistant of plant remains. However, the gray floras also have such forms as cordaitalean foliage, one of the most transport and decay resistant elements in late Paleozoic, equatorial vegetation. Furthermore, rare bits of callipterids, tree ferns and calamitaleans indicate that these plants were somewhere present on the red-beds landscape, yet they all remain almost absent from red-beds deposits. In addition, the red beds often preserve plant remains in dense mats, including large, nearly intact branches and fronds, often found crosscutting bedding planes. The former tend to occur in the finest claystone facies, suggesting deposition from suspension, thus not a sedimentary setting that would have subjected more delicate kinds of foliage to differential destruction during transport. Finally, the dominant elements in the red beds, conifers and supaioid peltasperms, are structurally quite distinct and likely would have had very different transport properties, both from viewpoints of survival and likelihood of being carried by streams; the rare co-occurrence of these two forms, particularly as dominant elements, further suggests that taphonomic biases have not created the distinctive red-beds flora.

Invertebrate Paleontology

Invertebrate body fossils are present in the Abo Formation at only a few outcrops—notably in the northern Sandia Mountains and in the Caballo Mountains.

In the northern Sandia Mountains near Placitas, the Abo Formation includes an unusual lens of carbonaceous shale in the Scholle Member (Lucas et al., 1999). Martens and Lucas (2005) documented conchostracans from this shale.

In the southern Caballo Mountains, the calcareous shells of microconchid gastropods and at least three different kinds of bivalves are present in the greenish-gray shale and siltstone beds of the lower Abo Formation, which Lucas et al. (2012a) identified as an estuarine facies. Pectinacean bivalves, some of which can be assigned to *Dunbarella*, and myalinid bivalves, are particularly common. Pectinaceans are marine to brackish water bivalves, as are myalinids. Thus, their presence indicates marine or brackish waters in the estuarine facies.

Fossil Vertebrates

Fossil bones and teeth are common locally in the Abo Formation in outcrops in Socorro and Sandoval Counties and in the correlative strata of the Cutler Group in Rio Arriba County, New Mexico (e.g., Berman, 1993; Lucas et al., 2010, 2012b). Of the seven primary Abo Formation outcrop areas listed above under “Distribution,” those designated as the Sandia, the northern Oscura Mountains and the Lucero uplift have not yielded identifiable skeletal remains of vertebrates. A survey of the vertebrate fossils from the other four outcrop areas was first reported by Berman (1993), who used, in part, different designations, with some having a far lesser aerial extent. Jemez Mountains, Abo Pass, Cerros de Amado-Joyita Hills, and Fra-Cristobal and Caballo mountains outcrop areas proposed here were referred to by Berman (1993) as the Jemez Springs, Los Pinos Mountains, Socorro, and Caballo Mountains (limited to Caballo Mountains) localities, respectively. It should be noted that in 1993 Berman also reported vertebrate fossils from several sites in the Lucero uplift area that he believed to be from the Abo Formation, but that are now considered to be from the underlying Upper Pennsylvanian (upper Virgilian) Red Tanks Member of the Bursum Formation (Harris et al., 2004). The Red Tanks terrestrial vertebrates, with the exception of the addition of the basal synapsids, a caseid and *Edaphosaurus*, and a bolosaurid reptile and the absence of the amphibians *Zatrachys*, *Platyhystrix*, and *Diplocaulus* and the basal synapsid

Ophiacodon (Harris et al., 2004), otherwise duplicate the faunal list recorded here for the Abo Formation (see below).

In 1993, Berman presented taxonomic lists of vertebrates from each of the four vertebrate-bearing Abo Formation outcrop areas listed above. All have been revisited several times by joint or single collecting parties of the New Mexico Museum of Natural History and Science (NMMNHS) and the Carnegie Museum of Natural History (CM), which has resulted not only in extensive new collections, but also discovery of new fossiliferous sites. Collections are housed not only at both institutions, but also the University of New Mexico, Albuquerque.

The Abo Formation vertebrate taxa are reported at generic or higher levels (Table 3) for each of the four, vertebrate-producing, outcrop areas mentioned above. The fragmentary condition of the vast majority of the specimens prevents confident species assignment. The Abo Formation vertebrate assemblage is summarized from the following publications: Romer (1937); Romer and Price (1940); Langston (1952, 1953); Vaughn (1969); Olson and Vaughn (1970); Berman (1976, 1977, 1979, 1993); Berman and Reisz (1980, 1986); Harris et al. (2005); Madalena et al. (2007); Spielmann et al. (2009); Lucas et al. (2009a, b, 2012b); and Cantrell et al. (2011).

Although most of the taxa have been identified on the basis of fragments or single elements, many are represented by partial, articulated specimens, including skulls and postcrania, some of which have received detailed descriptions: (1) from the Jemez Mountains area a new species of *Sphenacodon*, *S. ferocior* Romer, 1937, based on a skull and anterior vertebrae (Romer, 1937; Romer and Price, 1940; Spielmann et al., 2010); a partial vertebral column of an embolomere, possibly *Archeria*; a new species of *Dimetrodon*, *D. occidentalis* Berman, 1977, based on elements of the mandible and a partially articulated string of mid-dorsal vertebrae (Berman, 1977) (Fig. 9A); a new species of *Trimerorhachis*, *T. sandovalensis* Berman and Reisz, 1980, based on a nearly complete skull and a large, articulated portion of the postcranium (Berman and Reisz, 1980) (Fig. 9B); and a partial skull of *Sphenacodon ferox* (Lucas et al., 2012b); (2) from the Abo Pass area a new species of the lungfish *Gnathorhiza*, *G. bothrotreta*, Berman, 1976, based on complete skulls of individuals preserved in aestivation burrows (Berman, 1976, 1979) (Fig. 9D-E); (3) from a specific, highly fossiliferous site within the Cerros de Amado-Joyita Hills area, designated as the Gallina Well locality (Berman, 1993), closely associated, disarticulated elements of the skull and postcranium of an indeterminate captorhinomorph reptile (Berman and Reisz, 1986) and a partial skull of *Diplocaulus* (Harris et al., 2005) (Fig. 9C), and from the same outcrop area dorsal vertebrae and neural spine fragments of a specimen of *Dimetrodon* (Lucas et al., 2009).

Clearly, the above distributions suggest that most of the vertebrates, particularly the highly terrestrial tetrapods (*Platyhystrix*, *Trimerorhachis*, *Diadectes*, and basal synapsids), were widely distributed throughout the Abo Formation, whereas others (xenacanth, paleoniscoid, *Gnathorhiza*, *Eryops*, *Zatrachys*, embolomere, and lepospondyls) were undoubtedly restricted to localized freshwater pond/lake or stream environments. Among the identified taxa, only *Gnathorhiza* and perhaps also *Trimerorhachis* and *Diplocaulus* provide an important insight into climatic conditions during the deposition of the Abo Formation. Most notable in this respect are the *Gnathorhiza bothrotreta* specimens from the Abo Pass outcrop area that are preserved in aestivation burrow casts within a freshwater limestone believed to be a small pond or lake deposit. Outside of the Lower Permian of Texas and Oklahoma, this represents the only known occurrence of aestivation burrows containing *Gnathorhiza* still preserved in place. A single, eroded-free burrow containing skeletal remains of *Gnathorhiza* was reported from the Lower Permian El Cobre Canyon Formation of El Cobre Canyon, New Mexico (Berman, 1993) and the Eskridge Shale of Nebraska (Huttenlocker et al., 2005). Preservation of *Gnathorhiza* in an aestivating life phase clearly indicates a prevailing climate punctuated by periods of severe seasonal drought. Interestingly, in the same deposit are small skulls of *Trimerorhachis* and articulated strings of vertebrae associated with partial or disarticulated skulls of *Diplocaulus*. Olson (1958, 1977) sus-

TABLE 3. Fossil vertebrate taxa from the Abo Formation in central New Mexico.

	Jemez Mountains	Abo Pass	Cerros de Amado- Joyita Hills	Fra Cristobal and Caballo Mountains
Elasmobranch fish				
Xenacanth				
Probably <i>Orthocanthus</i>	X			X
Paleoniscoid fish			X	
Dipnoan (Lungfish)				
<i>Gnathorhiza</i>	X	X		
Amphibians				
Temnospondyls				
<i>Eryops</i>	X		X	
<i>Zatrachys</i>	X		X	
<i>Platyhystrix</i>	X	X	X	
A dissorophid		X		
<i>Trimerorhachis</i>	X	X	X	X
Embolomere, ? <i>Archeria</i>	X			
Lepospondyl				
<i>Diplocaulus</i>	X	X	X	X
Diadectomorph				
<i>Diadectes</i>	X		X	
Reptiles				
Captorhinomorph				
Captorhinid			X	
Basal Synapsids				
<i>Ophiacodon</i>	X	X	X	
<i>Sphenacodon</i>	X		X	
Sphenacodontid indet.	X			
<i>Dimetrodon</i>	X		X	X

pected that both forms may have been capable of aestivation, at least during early stages of life; he contended that they were clearly capable of surviving severe drying conditions, even as adults.

Using terrestrial vertebrates as index fossils, Lucas (2005, 2006) has divided the latest Pennsylvanian (Virgilian) and Permian into 10 time intervals referred to as land-vertebrate faunachrons, abbreviated LVF. Unfortunately, vertebrates from the Abo Formation do not provide a precise basis for biochronological correlations other than a Coyotean LVF of Virgilian-Wolfcampian age. Abo vertebrates consistently occur in the lower portion of the formation (Scholle Member).

AGE OF THE ABO FORMATION

The age of the Abo Formation has long and correctly been perceived to be Early Permian. As stated above, the Bursum Formation, which is the geologically youngest unit beneath the Abo is of late Virgilian-early Wolfcampian (Newwellian = "Bursumian") age (Lucas and Krainer, 2004). These are Late Pennsylvanian strata using the conodont-defined base of the Permian, which places the Pennsylvanian-Permian boundary within the Wolfcampian, close to the early-middle Wolfcampian boundary (e.g., Lucas et al., 2002). Correlation of the base of the Abo Formation to the Powwow Conglomerate of the Hueco Group in West Texas also suggests that the age of the base of the Abo Formation is close to the early-middle Wolfcampian boundary (Lucas et al., 2011a, b). Therefore, assigning a middle Wolfcampian (Nealian) age to the base of the Abo Formation can be supported by regional stratigraphic relationships. And, as noted above, some of the tetrapod footprint ichnotaxa known from the Abo Formation are restricted to Lower Permian strata elsewhere.

The upper age limit of the Abo Formation is less certainly known. Traditionally, geologists working in New Mexico have equated the Abo-Yeso contact with the Wolfcampian-Leonardian boundary. Red-bed facies of the Abo Formation that intertongue with the upper part of the Hueco Group in southern New Mexico are interbedded with marine rocks of late Wolfcampian age (e.g., Lucas et al., 2011b). On face value

this could indicate that the Abo is no younger than late Wolfcampian, but it is not clear that the red-bed intertongues in the Hueco Group represent the uppermost Abo Formation. This leaves open the possibility that the uppermost Abo is of early Leonardian age.

Unfortunately, fossils from the Abo Formation in central New Mexico contribute little to further resolving its precise age. Most of the ichnofossils, plant fossils and invertebrate body fossils are characteristic Late Pennsylvanian-Early Permian taxa that provide no precise age determinations within that time interval. The tetrapod body fossils are characteristic of the Coyotean land-vertebrate faunachron, which spans part of the Virgilian through much of the Wolfcampian (Lucas, 2006). Thus, regional correlation of the Abo Formation as of middle-late Wolfcampian age is not contradicted by the available fossil evidence. Also, within current temporal resolution, equating the Abo-Yeso contact to the Wolfcampian-Leonardian boundary is a good approximation.

DEPOSITIONAL ENVIRONMENTS

The overall, regional deposition of the Abo Formation is well understood as having taken place on a low-gradient alluvial plain in which rivers flowed primarily to the south toward the shoreline of the Hueco seaway in southern New Mexico (Kues and Giles, 2004). Thus, Seager and Mack (2003, p. 29) well stated that "the Abo fluvial system was characterized by widely dispersed silt-bed or very fine sand-bed rivers, broad, well-oxidized floodplains, and evidence in both the channel and floodplain strata for marked seasonality of paleoclimate" and that deposition was "strongly influenced by the overall dry and megamonsoonal climate postulated for Pangea in Early Permian time." The Abo Formation thus well represents the concept of "wet red beds" deposited during the late Paleozoic across much of tropical Pangea (e.g., Schneider et al., 2010).

Abo mudstones with calcrete paleosols represent extensive muddy floodplains subjected to seasonal aridity. Lenticular intraformational conglomerate beds and crossbedded sandstone bodies are channel deposits

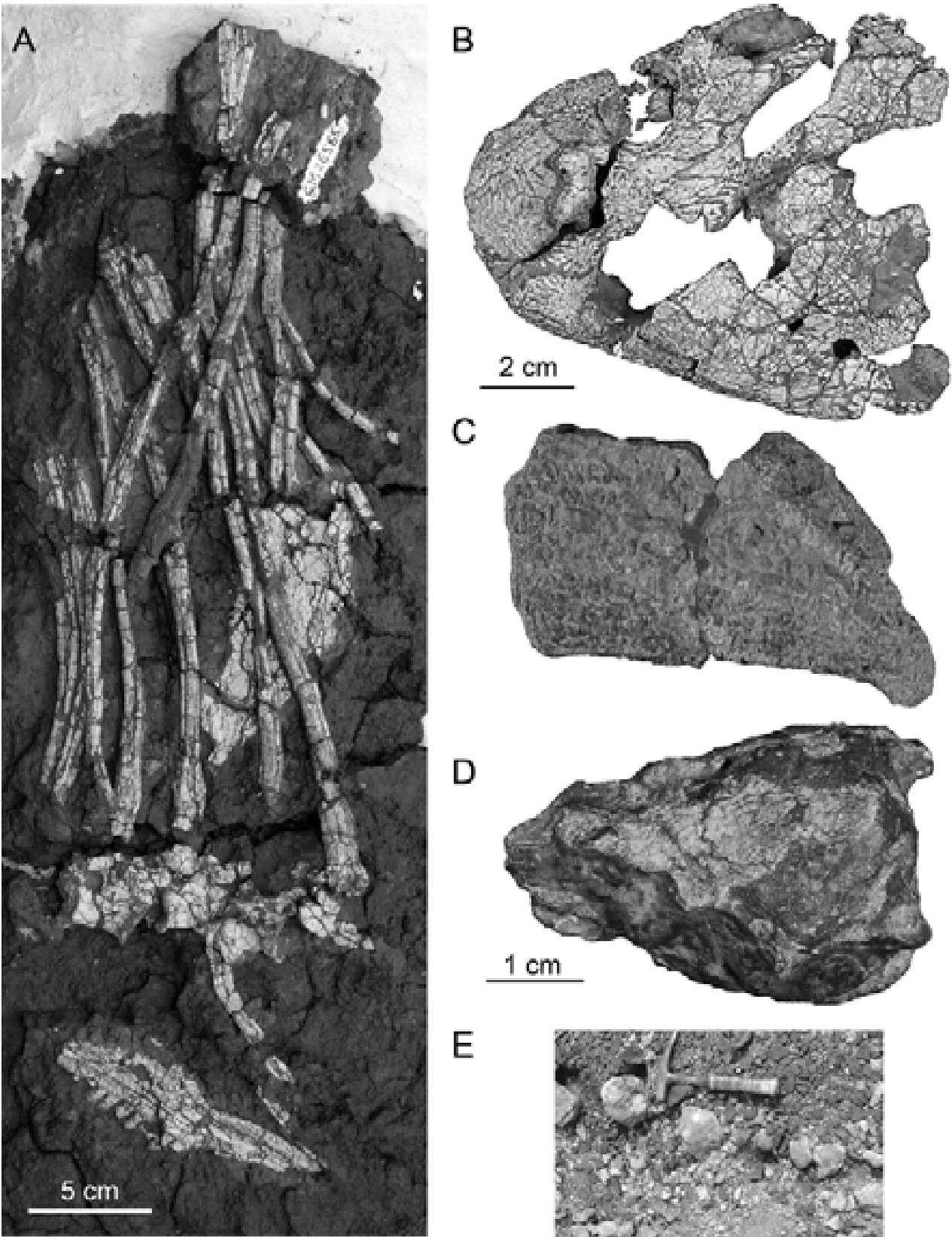


FIGURE 9. Selected vertebrate fossils from the Abo Formation, New Mexico. **A**, *Dimetrodon occidentalis*, holotype, CM 26565, large portion of dorsal sail elements and left mandible (angular, anterior coronoid and dentary) from the Jemez Mountains. **B**, *Trimerorhachis sandovalensis*, holotype, CM 38025, greater portion of skull in dorsal view from the Jemez Mountains. **C**, Right posterior portion of *Diplocaulus* skull, CM 38011, in dorsal view from the Gallina Well locality of the Cerros de Amado-Joyita Hills area. **D-E**, *Gnathorhiza bothrotreta*, referred specimen CM 30741, skull in dorsal view (anterior to left), and in place aestivation burrows casts containing in some *G. bothrotreta* preserved in a freshwater lake/pond deposit limestone in the Abo Pass area, respectively.

that accumulated during times of relatively rapid base-level fall. These are strata of the Scholle Member. Sheet sandstones characteristic of the Cañon de Espinosa Member indicate times of base level stability or relatively slow fall when Abo channels avulsed and spread across the floodplain to form thin, laterally extensive tabular sandstone bodies (Blakey and Gubitosa, 1984).

Some facies change is observed within the Abo Formation along a south-to-north transect in central New Mexico. Thus, conglomerate is very rare in Abo sections to the south, in Sierra County, comprising less than 1% of the total section, and consist entirely of reworked carbonate clasts (pedogenic carbonates). In addition, although the amount of sandstone is similar between Sierra County and the Abo type section in Torrance County, the average grain size is strikingly smaller in Sierra County outcrops (mostly < 0.2 mm) than in the Abo type section (average grain sizes > 0.5 mm). The small grain size is responsible for the different petrographic composition, particularly for the smaller amount of detrital feldspars, polycrystalline quartz grains and almost total absence of granitic and metamorphic rock fragments.

Bivalves from the greenish shale intervals in the Abo red beds in the Caballo Mountains indicate a brackish to marine environment. Lucas et al. (2012a) interpreted these shales as estuarine deposits based on their stratigraphic architecture and the association of terrestrial plants with marine bivalves. Intercalated conglomerates probably represent tidal-fluvial channel fills, and the fine-grained sandstone intercalations are distributary mouth bar deposits. These brackish to marine intercalations indicate that the Hueco transgression during the Wolfcampian reached as far north as the southern Caballo Mountains.

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At the top of the Abo type section, north of Abo Pass, brown sandstone at the base of the Lower Permian Yeso Group (base of the Arroyo de Alamillo Formation) at the top of the roadcut overlie repetitively-bedded sandstones, siltstones and mudstones of the upper member (Cañon de Espinoso Member) of the Lower Permian Abo Formation. Thin, laterally extensive sandstone beds of the Abo mostly represent unchannelized flow (sheetfloods), and Abo mudrocks contain numerous calcrete nodules and rhizoliths indicative of paleosol formation.

THE LOWER PERMIAN YESO GROUP IN CENTRAL NEW MEXICO

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Abstract—Strata of the Lower Permian Yeso Group exposed in central New Mexico (Sandoval, Bernalillo, Valencia, Torrance, Socorro and Sierra Counties) can be assigned to the De Chelly, San Ysidro, Arroyo de Alamillo and Los Vallos formations. The De Chelly Sandstone is the lower part of the Yeso Group in the Jemez Mountains, Lucero uplift and northern Sandia uplift. It is overlain by the San Ysidro Formation in the Jemez Mountains and northern part of the Sandia uplift. In the southern Jemez Mountains the De Chelly is approximately 80 m thick and the San Ysidro Formation is up to 135 m thick. To the south, from Abo Pass to the Caballo Mountains, the lower part of the Yeso Group is the Arroyo de Alamillo Formation, overlain by the Los Vallos Formation. The Arroyo de Alamillo Formation is ~40-90 m thick and consists primarily of red-bed siltstone and sandstone, which is very fine grained, silty, often gypsiferous and mostly thinly laminated or ripple laminated. Siltstone (mostly massive) is much less common, and there are a few beds of dolomite and gypsum. In Valencia, Torrance, Socorro and Sierra Counties, the overlying Los Vallos Formation is up to 240 m thick and can be divided into the Torres Member (~130-150 m of interbedded dolomite, gypsiferous siltstone, gypsum and siltstone to fine-grained sandstone), Cañas Member (~20-60 m of gypsum with lesser beds of gypsiferous siltstone and dolomite) and the Joyita Member (~30-65 m of red-bed siltstone and fine-grained sandstone) overlain by the Lower Permian Glorieta Sandstone. Six, dolomite-dominated intervals of the Torres Member can be readily correlated from the Yeso Group type section in Socorro County to the Fra Cristobal Mountains. Correlation farther south, into the Caballo Mountains is more difficult, due to facies and thickness changes within the Los Vallos Formation and stratigraphic relief of the unconformity at the base of the overlying San Andres Formation.

Marine strata of the Yeso Group have a sparse brachiopod-dominated macrofauna but a locally abundant microbiota of foraminifers and calcareous algae. Red-bed siliciclastics of the De Chelly and Arroyo de Alamillo formations yield plant fossils (callipterids and conifers) as well as diverse trace fossils. The trace fossils include arthropod burrows and walking and resting traces characteristic of the *Scoyenia* ichnofacies and one marginal marine ichnoassemblage of the *Cruziana* ichnofacies. Tetrapod tracks are of small temnospondyls, captorhinomorphs and araeoscelid diapsids. None of these Yeso Group fossils are age diagnostic, so it is assigned an early-middle Leonardian age based on its stratigraphic position and regional stratigraphic relationships. Deposition of the Arroyo de Alamillo Formation took place by eolian and fluvial processes on an arid coastal plain during a time of regional low sea level. Deposition of the Torres and Cañas members of the Los Vallos Formation was in cyclically shallow marine and sabkha environments. Joyita Member deposition again records eolian and fluvial processes during a time of regionally low sea level.

INTRODUCTION

In central New Mexico, the Lower Permian Yeso Group is well exposed in the mountain ranges and other uplifted areas that border the Rio Grande rift (Fig. 1). Study of these outcrops during the last decade has resulted in a much more refined understanding of their stratigraphy, paleontology, microfacies, correlation and depositional environments (e.g., Mack and Dinterman, 2002; Lucas et al., 2005; Lucas and Krainer, 2012). Here, we review our current understanding of the Yeso Group in central New Mexico, which refers primarily to Sandoval, Bernalillo, Valencia, Torrance, Socorro and Sierra Counties (Fig. 1).

HISTORY OF STUDY

Stratigraphic study of the Yeso lithosome began with Lee (1909), who divided the earlier defined Manzano Group of Herrick (1900) into the Abo, Yeso and San Andreas (sic) formations (Fig. 2). Lee's (1909, p. 12) Yeso Formation derived its name from Mesa del Yeso near Socorro, New Mexico, and he described it there as "1,000 to 2,000 feet [~310-620 m] of sandstone, shale, earthy limestone, and gypsum." He also included a sandstone-dominated interval in the top of the Abo Formation

that was later recognized as the lower part of the overlying Yeso Formation (see below) (Fig. 2). Furthermore, Lee's Yeso Formation also included strata later recognized as a separate formation by Keyes (1915), the Glorieta Sandstone.

Darton (1928) was unable to separate the Yeso and San Andres formations regionally, so he combined them in his "Chupadera Formation," a stratigraphic concept abandoned in the 1940s (Lucas, 2009). Needham and Bates (1943) described type sections of the Abo and Yeso formations of Lee, and thereby modified somewhat his concepts of those units and proposed and named some formal subdivisions of the Yeso Formation. Thus, Needham and Bates (1943) excluded the Glorieta Sandstone from the Yeso, and named two members of the upper part of the Yeso Formation, the Cañas [gypsum] member and the Joyita [sandstone] member (Fig. 2).

Until 2005, the concept of the Yeso Formation used in central New Mexico came primarily from four maps published between 1946 and 1951 by the United States Geological Survey (USGS) as part of mapping carried out during World War II. These were maps of the Jemez Pueblo and Sierra Nacimiento in Sandoval County (Wood and Northrop, 1946), the Lucero uplift of Valencia County (Kelley and Wood, 1946),

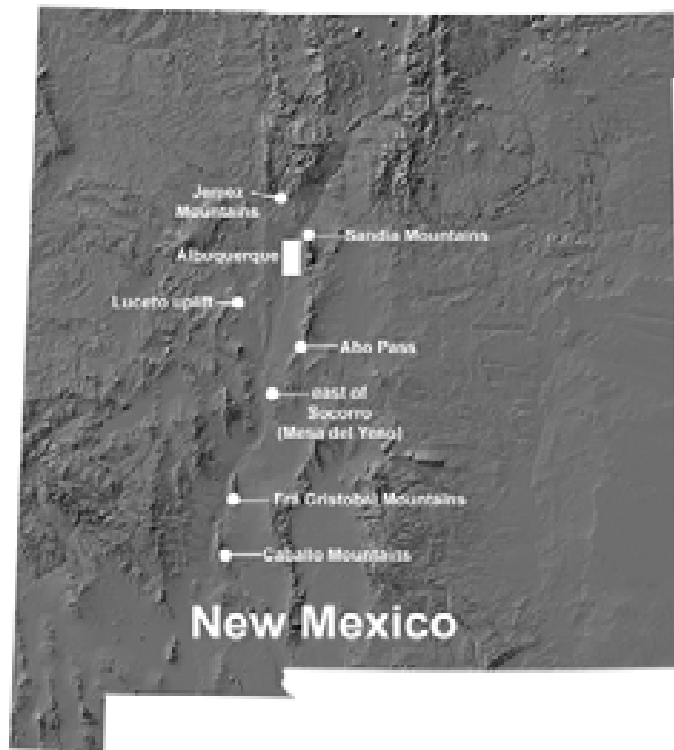


FIGURE 1. Map of New Mexico showing locations of major Yeso Group outcrops discussed in the text.

the Joyita Hills, the Los Piños Mountains and northern Chupadera Mesa of Socorro County (Wilpolt, et al., 1946) and the Cerrillos del Coyote, northern Jornada del Muerto and southern Chupadera Mesa of Socorro and Torrance Counties (Wilpolt and Wanek, 1951). Most significantly, these workers removed the upper sandstone interval from the Abo Formation, and called it the Meseta Blanca sandstone member of the Yeso Formation (name proposed by Wood and Northrop, 1946). They also proposed three member names for all or most of the remaining Yeso Formation: San Ysidro, Los Vallos and Torres members (Fig. 2).

Baars' (1962) important synthesis of Lower Permian stratigraphy on the Colorado Plateau used Abo Formation in the same sense as the USGS maps of 1946 cited earlier. However, he concluded that the type section of the Meseta Blanca Member of the Yeso Formation near Jemez Pueblo is simply the same unit that Gregory (1917) had named the De Chelly Sandstone near Chinle, Arizona. Therefore, Baars (1962) abandoned the name Meseta Blanca and replaced it with De Chelly Sandstone, which he removed from the Yeso Formation. He thus advocated abandoning member names such as San Ysidro and Los Vallos because they were synonyms of his redefined Yeso Formation, though Baars (1962) did concede that in the area of the type Yeso section it could be divided into multiple members.

Baars (1962) recommendations, however, were not heeded by subsequent workers in New Mexico, who, for more than 50 years, continued to use the Yeso stratigraphic nomenclature of the 1946 published maps (e.g., Hatchell et al., 1982; Mack and Suguio, 1991; Dinterman, 2001; Mack and Dinterman, 2002; Seager and Mack, 2003; Kues and Giles, 2004). Nevertheless, our own work (Lucas et al., 1999; Lucas and Krainer, 2012; Lucas and Zeigler, 2004; Lucas et al., 2005) convinced us that some modifications were justified to produce a lithostratigraphic nomenclature that better reflects current understanding of the lithostratigraphy of the Yeso Group interval in central New Mexico. Therefore, Lucas et al. (2005) elevated Yeso to group status consisting of three formation-rank units and several members. This lithostratigraphic nomenclature is used here.

STRATIGRAPHIC NOMENCLATURE

The Yeso Formation of traditional usage consists of various members, some of which are relatively thick (250-300 m) and lithologically distinctive units that have been routinely mapped by various workers at reasonable scales (including 1:24,000). Because of this, we concluded that the Yeso members merit formation rank, so we raised the Yeso to group rank (Lucas et al., 2005). The stratigraphic code (NACSN, 2005, p. 1569 [article 28]) states that a group rank unit is designed to "express the natural relationship of associated formations," and the Yeso Group meets this requirement. In other words, it unites an array of mappable formations that consist of siliciclastic, carbonate and evaporite sedimentary rocks that share a genesis in eolian, wadi, sabkha and arid coastal plain settings that lends them a lithostratigraphic integrity that long justified their inclusion in a single formation. However, the thickness, lithologic distinctiveness and great areal extent of the Yeso subdivisions (traditional members), and the fact that some of them can be further subdivided, warrant raising Yeso to group rank. Broadly correlative or homotaxial units of similar thickness and extent are also group-rank units, such as the Clear Fork Group of Texas, the Supai Group of Arizona and the Cutler Group of the Four Corners.

A further change that was long overdue was to adopt Baars (1962) recognition that the Meseta Blanca Member of the Yeso Formation of Wood and Northrop (1946) at its type section at Jemez Pueblo is the same unit as the De Chelly Sandstone of Gregory (1917) (also see Huffman and Condon, 1993). The name Meseta Blanca thus needed to be abandoned, and we included the De Chelly Sandstone in New Mexico as a formation-rank unit in the Yeso Group (Lucas et al., 2005). In Arizona, the De Chelly Sandstone is a formation in the Cutler Group (e.g., Peirce, 1989), a change in group inclusion regionally that is consistent with usage endorsed by the NACSN (2005).

The De Chelly Sandstone interval of the lower part of the Yeso Group changes facies substantially from north to south. Thus, at Jemez Springs (type section of the Meseta Blanca Member), the De Chelly is 70-80 m thick and composed primarily of sandstone with large scale crossbeds. This facies persists to Placitas in Bernalillo County (Lucas et al., 1999) but rapidly changes there to a facies of tabular- and ripple-bedded sandstone and siltstone with minor dolomite and gypsum beds (unpublished data). This tabular bedded and finer-grained facies persists south through Bernalillo, Valencia, Torrance and Socorro Counties to the southern Caballo Mountains in Sierra County (Lucas and Krainer, 2012). It is the lower part of the Yeso type section east of Socorro (Lucas et al., 2005). Lucas et al. (2005) named it the Arroyo de Alamillo Formation of the Yeso Group (Fig. 2).

In the Lucero uplift of Valencia County, the type section of the Los Vallos Member of the Yeso Formation of Kelley and Wood (1946) includes strata equivalent to the Torres, Cañas and Joyita members of the Yeso in the Joyita Hills and Cerrillos del Coyote of Socorro County (Lucas and Zeigler, 2004). The simplest nomenclatural solution, employed by Lucas et al. (2005), was to raise Los Vallos to formation rank, and recognize the Torres, Cañas and Joyita as member-rank units (Fig. 2). From Placitas northward, the upper part of the Yeso Group is the San Ysidro Formation, sandstone-dominated strata homotaxial with the Los Vallos Formation to the south.

LITHOSTRATIGRAPHY

Contacts

The basal contact of the Yeso Group in central New Mexico is generally conformable on underlying siliciclastic red beds of the Abo Formation. At most outcrops, there is either an interbedding of Abo and Yeso lithotypes over a stratigraphic interval of 10 m or less, or the base of the Yeso Group is picked at the lowest non-arkosic, fine-grained sandstone bed or the lowest dolomite bed above Abo mudrock or arkosic sandstone. Rarely, Yeso strata rest unconformably on rocks older than

Lee (1909)	Carlton (1928)	Needham & Bates (1943)	Kelley & Wood (1946)	Wood & Northrop (1946)	Valpoit et al. (1946)	Baars (1962)	Lucas et al. (2005)
San Andres limestone		San Andres Formation	San Andres Formation	San Andres Formation	San Andres Formation	San Andres Formation	San Andres Formation
		Glorieta sandstone	Glorieta sandstone mbr.	Glorieta sandstone mbr.	Glorieta sandstone mbr.	Glorieta Sandstone	Glorieta Sandstone
Yeso Formation	Chupadera Formation	Joyita Member	Yeso Formation	Yeso Formation	Joyita Member	Yeso Formation	Joyita Member
		Cañas Member			Cañas Member		Cañas Member
		unnamed member			Torres Member		Torres Member
			Los Valles Member	San Ysidro Member			Yeso Group
			Meseta Blanca sandstone member	Meseta Blanca sandstone member	Meseta Blanca sandstone member	De Chelly Sandstone	Arroyo de Alamillo Formation
Abo sandstone	Abo sandstone	Abo Formation	Abo Formation	Abo Formation	Abo Formation	Abo Formation	Abo Formation
							De Chelly ss.

FIGURE 2. Development of lithostratigraphic nomenclature of the Yeso Group.

the Abo Formation. A striking example, documented by Kottowski (1985), is near Clines Corners in Torrance County, where Yeso strata rest directly (and with evident nonconformity) on Proterozoic basement.

The contact of the overlying Lower Permian Glorieta Sandstone on the Yeso Group is generally described as gradational and intertonguing (e.g., Milner, 1978). However, at some locations in southern New Mexico the Lower Permian San Andres Formation rests directly on Yeso Group strata (no Glorieta strata are present) with evident unconformity (e.g., Kottowski et al., 1956; Lucas and Krainer, 2012).

Thickness

Table 1 summarizes thickness data from the key Yeso Group outcrop belts identified in Figure 1. In general, total Yeso thickness in central New Mexico is 200-300 m. The lower, clastic unit of the Yeso Group (De Chelly and Arroyo de Alamillo formations) is about 60-80 m of that thickness. Yeso thickness generally increases to the south and particularly to the southeast, where Yeso Group thickness locally exceeds 1200 m (Mack and Dinterman, 2002).

Lithology

In central New Mexico, Yeso Group strata are mostly mature sandstones and siltstones. Carbonate beds (usually dolomite or dolomicrite) are a relatively minor component of Yeso thickness, but form prominent marker beds. Gypsum is a prominent constituent of some intervals of the Yeso Group, and many of the siliciclastic beds are gypsiferous. In general, the lower part of Yeso Group is siliciclastic, whereas the upper part is mixed siliciclastic, evaporite and carbonate (Figs. 3-5).

De Chelly Sandstone

The De Chelly Sandstone in northern New Mexico (Figs. 3, 6A-B) is mostly yellowish brown to pale reddish brown sandstone. Typically crossbedded, it also includes some laminar, ripple laminar and pedoturbated beds. The lower half of the type section of the “Meseta Blanca Member” near Jemez Springs has beds with climbing ripples as well (Fig. 3). Very striking, though, are the largescale trough crossbeds (eolian foresets) in the upper half of the formation (Fig. 6B).

At Jemez Pueblo, the De Chelly Sandstone is approximately 80 m thick (Lucas et al., 2005). The lower half is composed of red siltstone,

TABLE 1. Thickness (in m) of Yeso Group and its subdivisions at outcrop areas in central New Mexico identified in Figure 1.

Unit/location	DeChelly Sandstone	San Ysidro Formation	Arroyo de Alamillo Formation	Los Valles Formation (Torres Member)	Los Valles Formation (Cañas Member)	Los Valles Formation (Joyita Member)	Total Yeso Group
Armez Mts.	82	131					213
Placitas	60	113					173
Lacero uplift	70			143	60	63	348
Abo Pass			60	230*			290
Type section			107	136	66	12	341
Fra Cristóbal Mountains			81	134	38	42	315

partly with mudcracks and thin interbeds of ripple-laminated, fine grained sandstone and of fine-grained sandstone displaying horizontal lamination and ripple lamination, local mudcracks, sandstone with small-scale cross-bedding (small channel fills), rare thin sandstone beds with eolian crossbedding and one fine-grained sandstone horizon with rhizoliths in the upper part. Lithofacies of the lower part indicate deposition on an alluvial plain with sheetflood, small fluvial channel and fine-grained flood-plain deposits as well as rare, small fossil dunes.

The upper half (upper 40 m) of the De Chelly Sandstone is composed of several thick sandstone beds with large-scale eolian cross-bedding (sets up to 4.5 m thick), and intercalated fine-grained sandstone with horizontal lamination and poorly-preserved ripple lamination. Also, in the upper part massive sandstone with rhizoliths (fossil dykaka) is present. One thin calcrete horizon is intercalated. Eolian foresets dip towards the S-SW, indicating constant wind directions during deposition. The upper half of the De Chelly Sandstone represents fossil dune deposits with some intercalated interdunal sediments.

Waterlaid sandstone beds of the De Chelly Sandstone are composed of angular to subangular grains, dominantly of monocrystalline quartz, but also many polycrystalline quartz grains, detrital feldspars (potassium feldspars, mostly microcline) and granitic rock fragments are present. Sedimentary rock fragments (reworked pedogenic carbonate clasts) are rare. This type of sandstone is commonly poorly sorted and cemented by calcite (Fig. 7A).

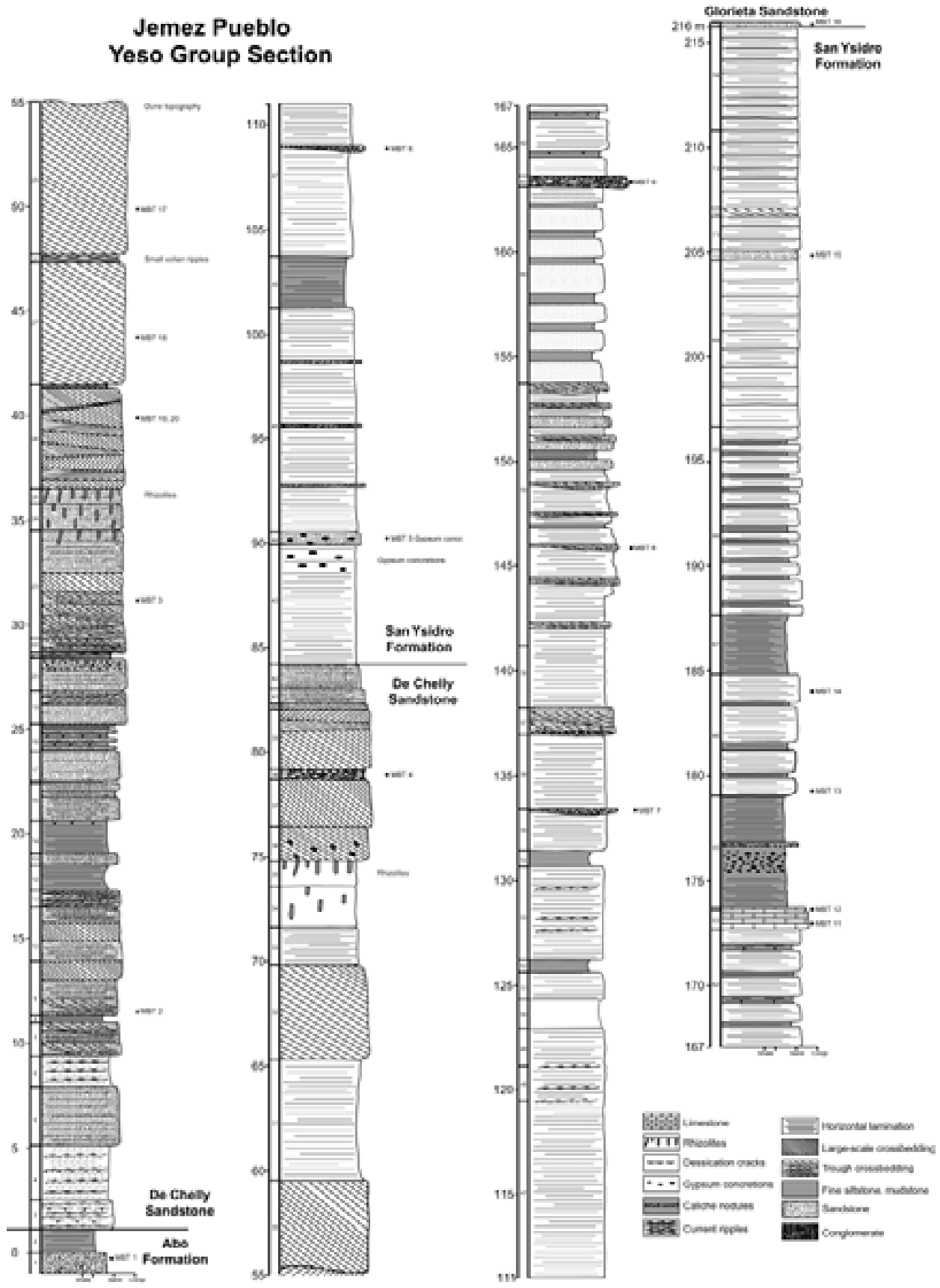


FIGURE 3. Type sections of the Yeso Group units named in the Jemez Mountains of Sandoval County. This is the type section of the Meseta Blanca Member (= De Chelly Sandstone) and San Ysidro Member of the Yeso Formation of Wood and Northrop (1946). Measured around NE ¼ NW ¼ sec. 3, T16N, R2E (from Lucas et al., 2005).

Eolian sandstone beds of the De Chelly Sandstone are well sorted, and the grains are mostly subrounded to rounded. The dominant grain type is monocrystalline quartz; polycrystalline quartz and detrital feldspars are rare, and granitic rock fragments are very rare to absent. The sandstone is cemented by quartz, which occurs as authigenic overgrowths.

Arroyo de Alamillo Formation

In central and southern New Mexico, the Arroyo de Alamillo Formation is mostly reddish brown, yellowish brown and gray sandstone and siltstone. In general, sandstones of the Arroyo de Alamillo Formation are very fine grained, silty, often gypsiferous and mostly thinly laminated and ripple laminated (Figs. 4-5, 6C). Only a few sandstone beds are cross bedded. Siltstones are also often gypsiferous and sometimes display thin laminations or ripple laminations. However, most siltstone beds are massive and blocky. A very characteristic feature of the Arroyo de Alamillo Formation is halite pseudomorphs that are most commonly present in beds of dolomitic siltstone stratigraphically low in the formation (Fig. 6F). A few beds of dolomite and gypsum are also present in the Arroyo de Alamillo Formation.

At the type section (Fig. 4) the Arroyo de Alamillo Formation is 107 m thick and can be divided into a lower, finer-grained and upper, coarser-grained part (Lucas et al., 2005). The lower part is composed of dominantly greenish and red silty mudstone to siltstone with intercalated thin beds of horizontally and ripple-laminated fine-grained sandstone, sandy limestone beds, and thin beds of carbonate siltstone to fine-grained sandstone displaying ripples and halite pseudomorphs. In the upper part of the Arroyo de Alamillo Formation, fine-grained sandstone with horizontal and ripple lamination, trough-crossbedded sandstone (channel fills) and sandstone with eolian foresets are intercalated. Sandstone intervals are up to 2.3 m thick. The sandstone is composed of abundant monocrystalline quartz grains, many detrital feldspars (mostly potassium feldspars, many of them are more or less altered), a few polycrystalline quartz grains and rare mica. In particular, eolian sandstones are well sorted and composed of rounded grains. The sandstone is cemented by quartz that occurs as authigenic overgrowths, and by calcite cement (Fig. 7C). The depositional environment of these strata was a coastal plain with periodic coastal sabkha conditions and rare eolian dunes.

In the Fra Cristobal and Caballo Mountains the Arroyo de Alamillo Formation ranges in thickness from approximately 42 to 87 m (Lucas and Krainer, 2012). The Arroyo de Alamillo Formation section at Massacre Gap in the Fra Cristobal Mountains is 81 m thick and almost completely exposed, with only a few covered intervals (Fig. 5). The succession is composed of various types of mudstone-siltstone, rarely containing halite pseudomorphs, with intercalated coarse-grained siltstone to fine-grained sandstone beds displaying ripple lamination, climbing ripples, trough crossbedding, partly with small ripples on top, crossbedding with small ripples on the foresets, horizontal lamination, and rare synsedimentary deformation structures. Greenish gypsiferous siltstone to fine-grained sandstone is intercalated, and, in the lower part, several carbonate beds 1-20 cm thick are present. Siltstone to fine-grained sandstone is moderately to well sorted and composed of abundant monocrystalline quartz grains, subordinate detrital feldspars, a few micas and very rare polycrystalline quartz. The rock is cemented by dolomite, randomly replacing quartz and feldspar. Subordinately, some quartz cement is present, which occurs as authigenic overgrowths on detrital quartz grains (Fig. 8A).

In the Caballo Mountains, at Broken House Tank, the Arroyo de Alamillo Formation is 74 m thick and overlain by 21.5 m of the lower part of the Torres Member of the Los Vallos Formation (Lucas and Krainer, 2012). The following lithofacies are observed within the Arroyo de Alamillo Formation: (1) red and greenish mudstone to fine-grained siltstone, commonly thinly laminated, also massive; (2) horizontally and ripple-laminated red siltstone to fine-grained sandstone; (3) crossbedded

fine-grained sandstone with erosional contacts at the bases of beds; (4) massive to indistinctly horizontally laminated sandstone; (5) crossbedded, fine-grained sandstone; (6) rare bioturbated siltstone to fine-grained sandstone; (7) gypsiferous siltstone to fine-grained sandstone, partly horizontally laminated; and (8) red gypsiferous siltstone to fine-grained sandstone containing clay rip-ups (resediment clasts). The Arroyo de Alamillo Formation in the McLeod Hills of the southern Caballo Mountains is composed of alternating greenish mudstone/shale, massive to indistinctly laminated siltstone to fine-grained sandstone, laminated siltstone to fine-grained sandstone, dolomitic siltstone to silty dolomite and gypsiferous siltstone to fine-grained sandstone.

The Arroyo de Alamillo Formation at Broken House Tank and Massacre Gap is interpreted to represent dominantly nonmarine, eolian sand sheet deposits with minor fluvial sediments and thin intervals of fine-grained sediments resulting from short periods of marine flooding with partly evaporitic conditions. In the McLeod Hills, the Arroyo de Alamillo Formation contains less eolian sediments, lacks fluvial deposits and contains a higher portion of sabkha and nearshore sediments compared to Broken House Tank and Massacre Gap (see Lucas and Krainer, 2012 for details).

San Ysidro Formation

In northern New Mexico, the San Ysidro Formation is mostly horizontally laminated, fine grained gypsiferous sandstone (Fig. 7B) and siltstone. A few crossbedded intervals are present, as are some thin beds of intraformational conglomerate. Interbedded sandstone and siltstone beds give a definite impression of cyclical sedimentation. Rare, thin limestone beds are present locally.

Thus, at its type section (Fig. 3), the San Ysidro Formation is 131 m thick and composed mostly of fine-grained sandstone and siltstone, either massive or displaying horizontal lamination. Less common is ripple lamination. Intercalated are thin sandstone beds with small-scale trough crossbedding (channel fills) and one thin sedimentary breccia bed. In the upper part of the section a 1-m-thick interval of thin-bedded limestone is intercalated. Red siltstone in the upper part contains abundant pedogenic carbonate nodules.

Los Vallos Formation (Torres Member)

The principal lithologies of the Torres Member are dolomite, gypsiferous siltstone, gypsum and siltstone to fine-grained sandstone. The dolomite is gray to dark gray, and the following types can be distinguished: indistinctly bedded dolomite, indistinctly bedded vuggy dolomite, thin-bedded (5-20 cm) dolomite with local halite casts, thin-bedded (5-20 cm) dolomite alternating with thin gypsum beds (10 cm), thicker-bedded (20-40 cm) dolomite, rarely containing fossil fragments (brachiopods, crinoids), and intercalated rare, thin (5-10 cm) coquina layers containing abundant gastropods and brachiopods. Within the Torres Member succession four to six dolomite intervals are present (Figs. 4-5) and can usually be traced as marker beds over great distances.

At the Yeso type section, the base of the Torres Member is drawn at the base of a distinctive, 2.5-m-thick dolomitic limestone ledge (Fig. 4). The Torres Member here shows a well-developed cyclic pattern of six transgressive-regressive cycles, each 15 to 35 m thick. In the lower two cycles, gypsum is absent, but the four cycles of the upper part contain considerable amounts of gypsum. Each cycle starts with a few-m-thick carbonate horizon overlain by siliciclastic sediments (mostly siltstone, subordinately fine-grained sandstone); the lower two cycles show a coarsening-upward trend and in their upper part contain laminated and nodular gypsum beds. Within the Torres Member siltstone predominates, constituting 60.7% of the section, followed by carbonate rocks (16.1%), gypsum (14%) and sandstone (9.2%).

Carbonate intervals are 1.2 to 4.2 m thick, gray, thin-bedded to indistinctly bedded and composed of dolomitic limestone to dolostone, more or less recrystallized and including peloidal wackestone, oolitic

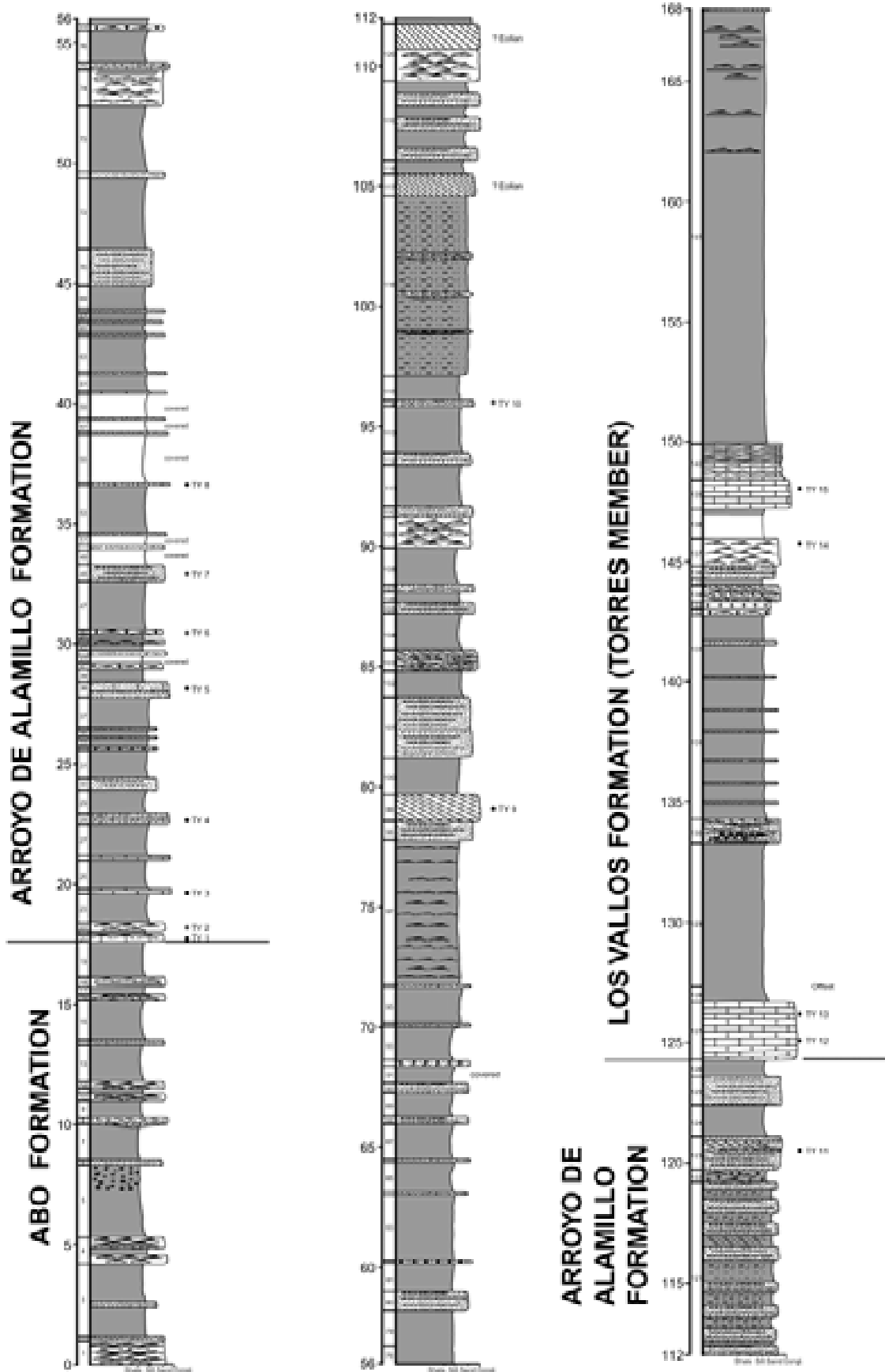


FIGURE 4. Type section of the Yeso Group at Mesa del Yeso in Socorro County. Measured in sec. 5, T2S, R2E and sec. 33 T1S, R2E (from Lucas et al., 2005).

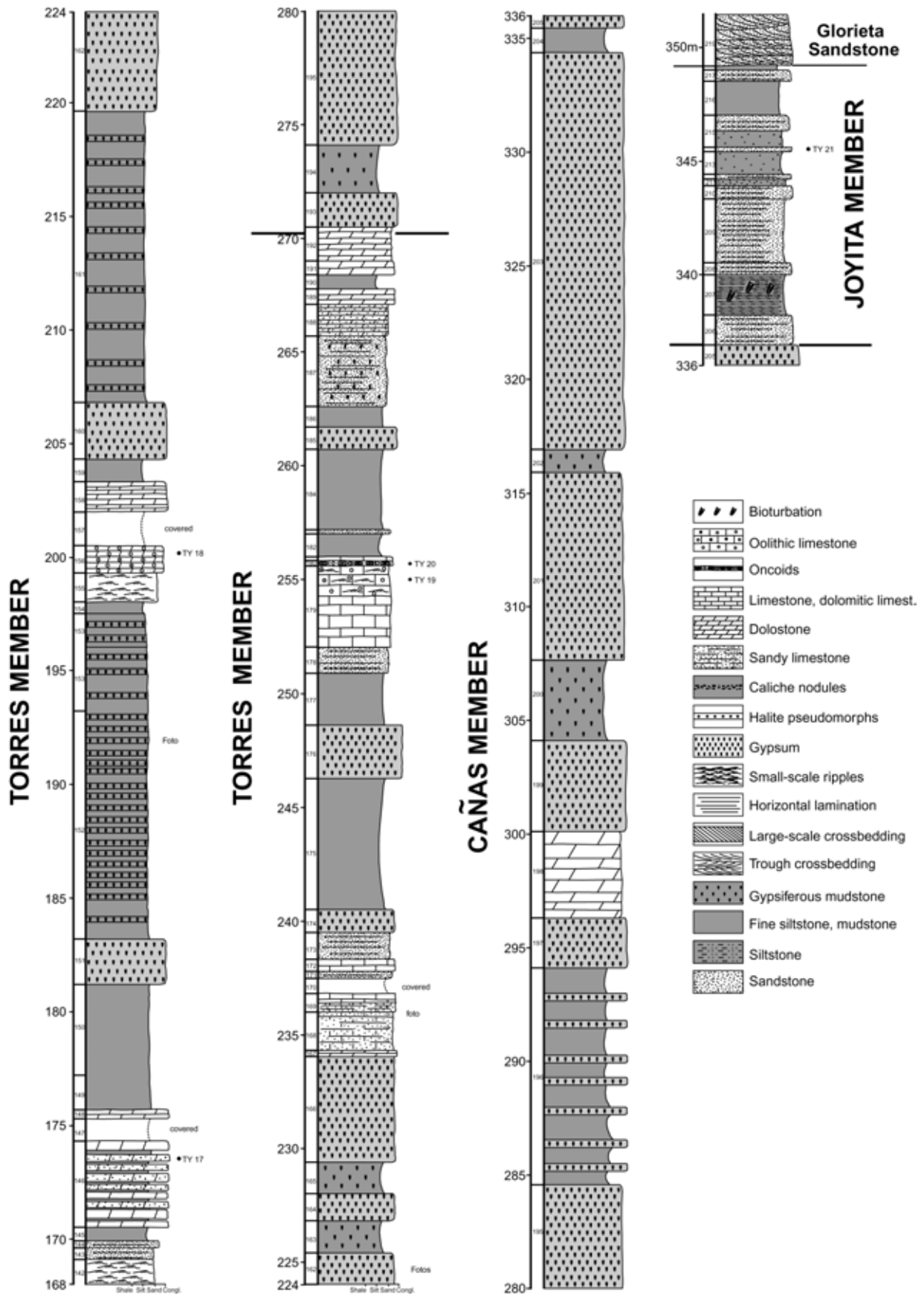


FIGURE 4. Continued. Type section of the Yeso Group at Mesa del Yeso in Socorro County. Measured in sec. 5, T2S, R2E and sec. 33 T1S, R2E (from Lucas et al., 2005).

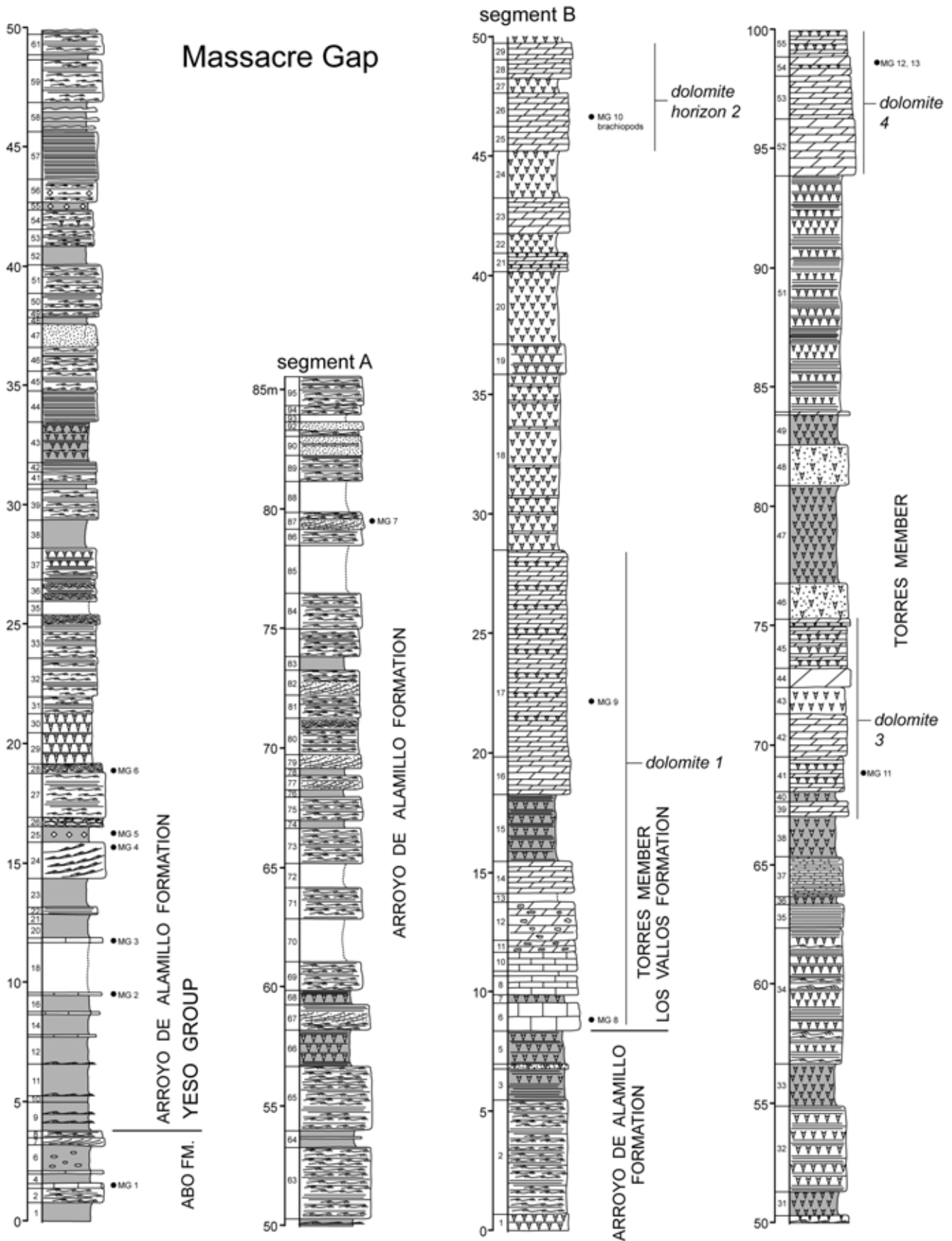


FIGURE 5. Measured stratigraphic section of the Yeso Group at Massacre Gap in the southern Fra Cristobal Mountains, Sierra County. Note that the section is measured in two segments, A and B, and that the top of A correlates approximately to the base of B.

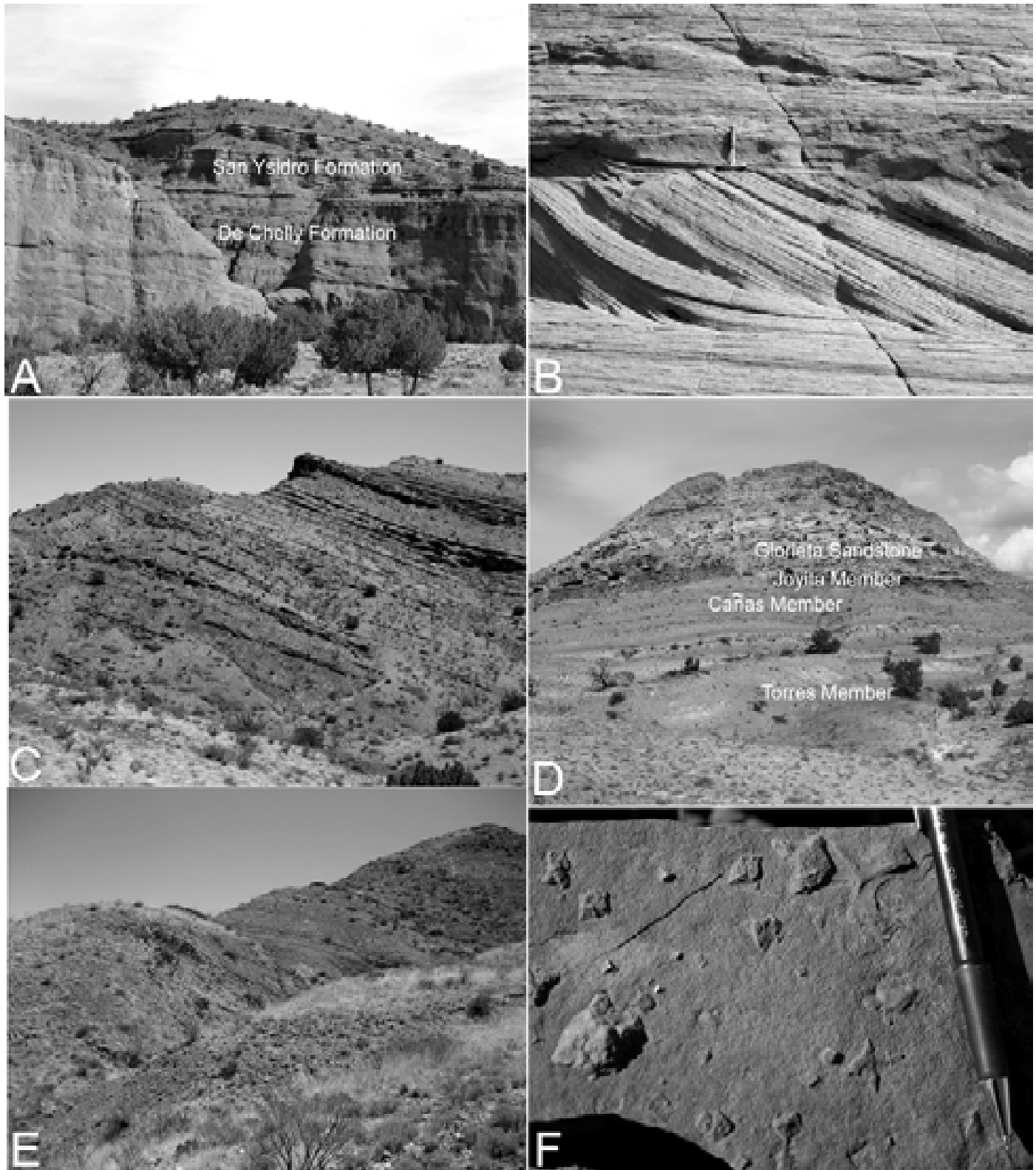


FIGURE 6. Photographs of selected Yeso Group outcrops in central New Mexico. **A-B**, Overview of section in Jemez Mountains (Fig. 3) showing De Chelly Sandstone overlain by San Ysidro Formation. **B**, Closeup of eolian crossbedded sandstone of De Chelly Sandstone at Jemez Mountains section. **C**, Overview of Arroyo de Alamillo Formation in the Loma de las Cañas east of Socorro, Socorro County. **D**, Upper part of type section of Yeso Group (Fig. 4) at Mesa del Yeso east of Socorro. **E**, Characteristic interbedded siltstone, dolomite and gypsum of Torres Member of Los Vallos Formation, Massacre Peak section, Fra Cristobal Mountains, Sierra County. **F**, Halite pseudomorphs in siltstone bed of Arroyo de Alamillo Formation at the Massacre Peak section.

wackestone (Fig. 7D), oncologic grainstone (Fig. 7E) to rudstone, and mudstone containing small quartz grains. The fossil assemblage in these strata is of low diversity and includes recrystallized shell fragments derived mainly from bivalves, ostracods and a few smaller foraminifers, indicating a shallow, restricted marine environment. The sandstone is fine-grained and displays horizontal lamination, trough crossbedding, ripple lamination and rare syndimentary deformation structures (Lucas et al., 2005).

In the southern Fra Cristobal Mountains (Massacre Gap; Fig. 5) and the McLeod Hills of the Caballo Mountains, the Torres Member is 134 m thick and is also a cyclic succession composed of dolomite, gypsiferous siltstone, gypsum and rare siltstone and fine-grained sandstone. Dolomite intervals are thicker (4.5–20 m) than at the type section, and the microfacies include peloidal grainstone to packstone, bioclastic grainstone, bioclastic wackestone (Fig. 8D, F), mudstone, rudstone (Fig. 8C, E) and oncoidal grainstone and rudstone (Fig. 8B). The fossil assemblage shows a higher diversity than at the type section (bivalve and brachiopod shells, gastropods, echinoderms, foraminifers, ostracods and locally abundant calcareous algae). Thin coquina layers (rudstone) containing abundant bivalve and brachiopod shells and a few echinoderms, ostracods, smaller foraminifers (Fig. 8G–I) and calcareous algae (Fig. 8H) are intercalated. Gypsiferous siltstone and gypsum is also more abundant than at the type section. Dolomite formed in a shallow marine, restricted environment, periodically under more normal, open marine conditions as is indicated by a relatively diverse fossil assemblage. Coquina layers are interpreted as storm layers. Gypsiferous siltstone and gypsum are deposits of a coastal sabkha environment (see details in Lucas and Krainer, 2012).

Los Vallos Formation (Cañas Member)

The term Cañas Member refers to the uppermost thick, gypsum-dominated interval of the Yeso Group section (Figs. 4–5). In central and southern New Mexico, the Cañas Member is 58 m (Massacre Gap, southern Fra Cristobal Mountains) to 68 m (Lucero uplift; type section) thick and composed of dominantly terra cotta-colored, rarely greenish, siltstone intercalated with gypsum and thin gypsum layers that make up several thick successions of gypsum. Locally, one or more relatively thin beds of dolomite are intercalated with the gypsum beds. At Mesa del Yeso, a 3.8 m thick dolomite interval is developed in the middle of the member (Lucas et al., 2005). At Massacre Gap, the dolomite interval in the middle of the member is much thicker (10 m), and additionally three thin dolomite horizons (1.2–1.4 m thick) occur at the top of the member (Lucas and Krainer, 2012).

Most of the gypsum beds of the Torres and Cañas members are poorly exposed. Where outcrop conditions are good, three types of fine crystalline gypsum can be distinguished: (1) “wavy” laminated (folded) gypsum composed of alternating dark and light gray gypsum and thin, dark gray and reddish mudstone laminae; (2) indistinctly laminated gypsum with even lamination; and (3) nodular gypsum with light gray gypsum nodules up to a few cm in diameter surrounded by dark gray mudstone.

Los Vallos Formation (Joyita Member)

Across central and southern New Mexico, the uppermost unit of the Los Vallos Formation is red-bed siliciclastics of the Joyita Member. Thickness ranges from 12.3 m at Mesa del Yeso to 65 m in the Lucero uplift. Most of the Joyita Member is massive to horizontally and ripple laminated siltstone to fine-grained sandstone. Some beds of massive red siltstone to fine-grained sandstone are also present. Gypsum is absent. Siltstone and sandstone is dominantly red, although greenish-gray and brownish colors are also present. Composition of the sandstones (e.g., Fig. 7F) is similar to the sandstones of the Arroyo de Amarillo Formation and Torres Member (Lucas et al., 2005; Lucas and Krainer, 2012).

Distribution

The Yeso Group is found across most of New Mexico. It is particularly well exposed in almost all the mountain ranges and other uplifted areas of the central part of the state. To the southeast, Yeso strata are laterally equivalent to the Bone Spring Formation, marine strata deposited in the Permian basin. Here is a brief synopsis of the main outcrop areas we have studied in central New Mexico (Fig. 1).

Jemez Mountains

Yeso Group strata are well exposed in the southern Jemez Mountains, particularly along the canyons formed by the Jemez River and its tributaries (Wood and Northrop, 1946; Osburn et al., 2002; Kelley et al., 2003). Here, the Yeso Group is ~ 150 to 220 m thick and can be divided into the lower, De Chelly Sandstone and upper, San Ysidro Formation (Fig. 3). Compared to the sections farther south (Mesa del Yeso, Massacre Gap, McLeod Hills), the Yeso Group in the southern Jemez Mountains is thinner and almost entirely nonmarine siliciclastic red beds with a high amount of eolian deposits in the lower part.

Sandia Mountains

On the northern end of the Sandia Mountains of Bernalillo County, near Placitas, an ~ 175-m thick section of the Yeso Group crops out (Kelley and Northrop, 1975; Lucas et al., 1999). The lower 60 m are the De Chelly Sandstone, overlain by a 115-m thick section of the San Ysidro Formation. Unpublished data recently gathered by us indicate a significant facies change where the De Chelly Sandstone laterally grades southward into the Arroyo de Alamillo Formation in this area (T13N, R5E).

Lucero uplift

In the vicinity of Carrizo Mesa in the Lucero uplift of Valencia County, an ~ 350-m thick section of the Yeso Group crops out (Kelley and Wood, 1946; Lucas and Zeigler, 2004). The lower part of this unit is 70 m of De Chelly Sandstone that is overlain by an ~ 278-m thick section of Los Vallos Formation. The Los Vallos can be divided into the Torres (145 m), Cañas (68 m) and Joyita (65 m) members.

Abo Pass

At Abo Pass, along the boundary of Valencia and Torrance Counties at the southern end of the Manzano Mountains, an ~ 290 m thick section of Yeso Group is exposed (Bates et al., 1947; Oviatt, 2010, 2011). The section here is very similar to the type section east of Socorro. Thus, the section consists of ~ 60 m of Arroyo de Alamillo Formation, which overlies the type section of the Abo Formation in the vicinity of the Abo Ruins (Lucas et al., 2005). Most of the Yeso Group is the 230 m thick Los Vallos Formation, divisible into the Torres, Cañas and Joyita members.

Mesa del Yeso

The Yeso Group at the type section (Figs. 4, 6D) in the Sierra de la Cruz area east of Socorro is ~ 341 m thick and was long divided into the basal Meseta Blanca Member, the Torres Member, the Cañas Gypsum Member and the Joyita Sandstone Member (e.g., Needham and Bates, 1943). Lucas et al. (2005) divided it into the Arroyo Alamillo Formation (107 m) overlain by the Los Vallos Formation consisting of the (in ascending order) Torres (156 m), Cañas (66 m) and Joyita (12 m) members. Besides the type section, there are extensive outcrops of the Yeso Group in the region between the Joyita Hills and Cerros de Amado, east of Socorro, one of the largest outcrop belts in the state.

Fra Cristobal and Caballo Mountains

Strata of the Lower Permian Yeso Group exposed in the Fra Cristobal and Caballo Mountains of Sierra County can be assigned to the

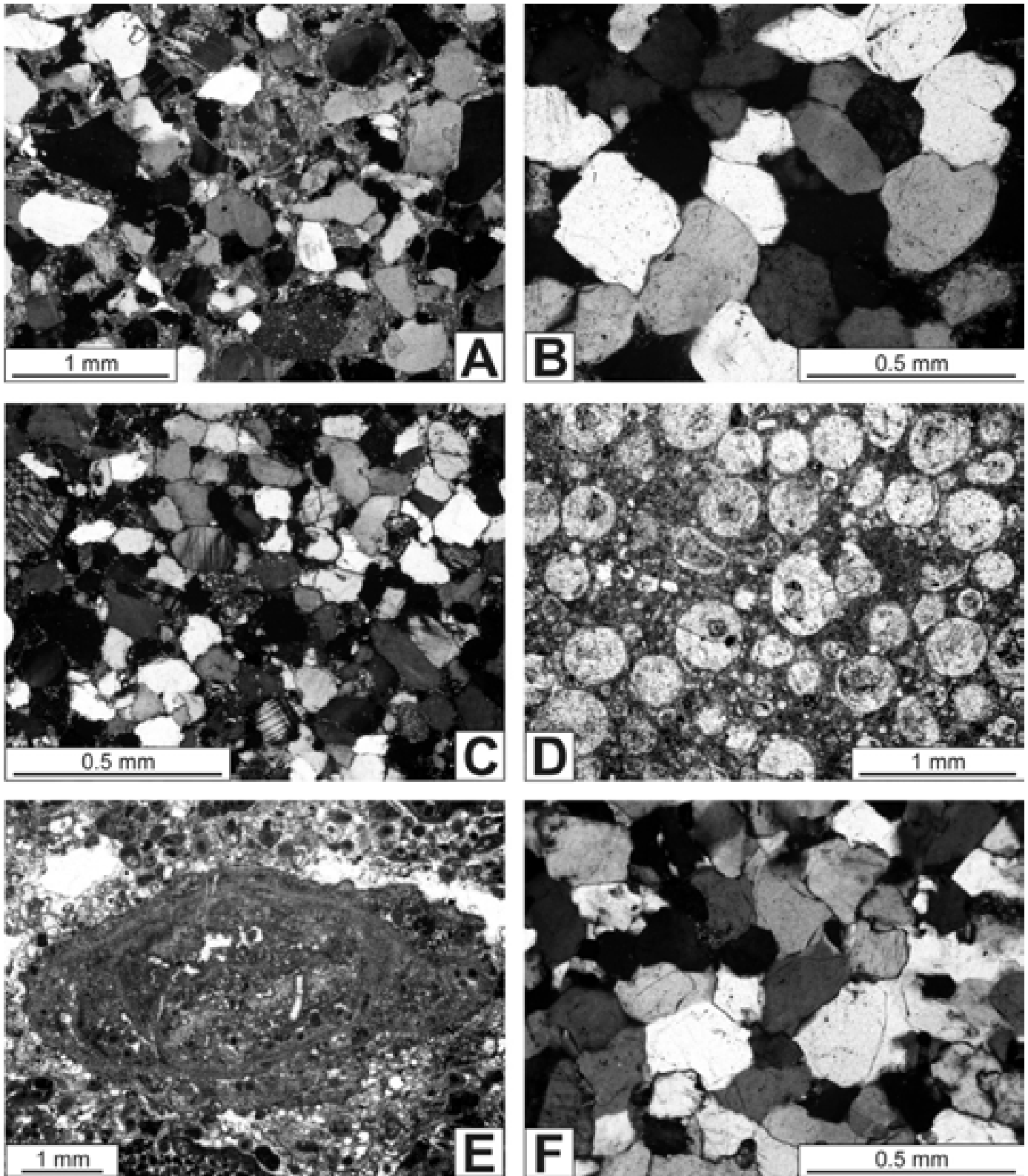


FIGURE 7. Thin sections of selected Yeso Group strata in central New Mexico. **A**, Calcite-cemented arkosic sandstone, composed of mono- and polycrystalline quartz grains, detrital feldspars, granitic rock fragments and rare metamorphic rock fragments. Sample MBT 2, De Chelly Sandstone, Jemez Mountains. **B**, Well-sorted quartzarenite composed of rounded grains of monocrystalline quartz displaying very visible authigenic overgrowths. Sample MBT 15, San Ysidro Formation, Jemez Mountains. **C**, Well-sorted, calcite-cemented sandstone composed of abundant detrital quartz grains and a few detrital feldspars (including microcline). Sample TY 9, Arroyo de Alamillo Formation, Yeso type section. **D**, Recrystallized oolitic wackestone, sample TY 18, Torres Member, Yeso type section. **E**, Oncoïd embedded in grainstone composed of abundant ooids and peloids. Sample TY 20, Torres Member, Yeso type section. **F**, Well-sorted sandstone (quartzarenite) composed mostly of monocrystalline quartz grains and quartz cement that occurs as authigenic overgrowths on detrital quartz grains. Sample TY 21, Joyita Member, Yeso type section.

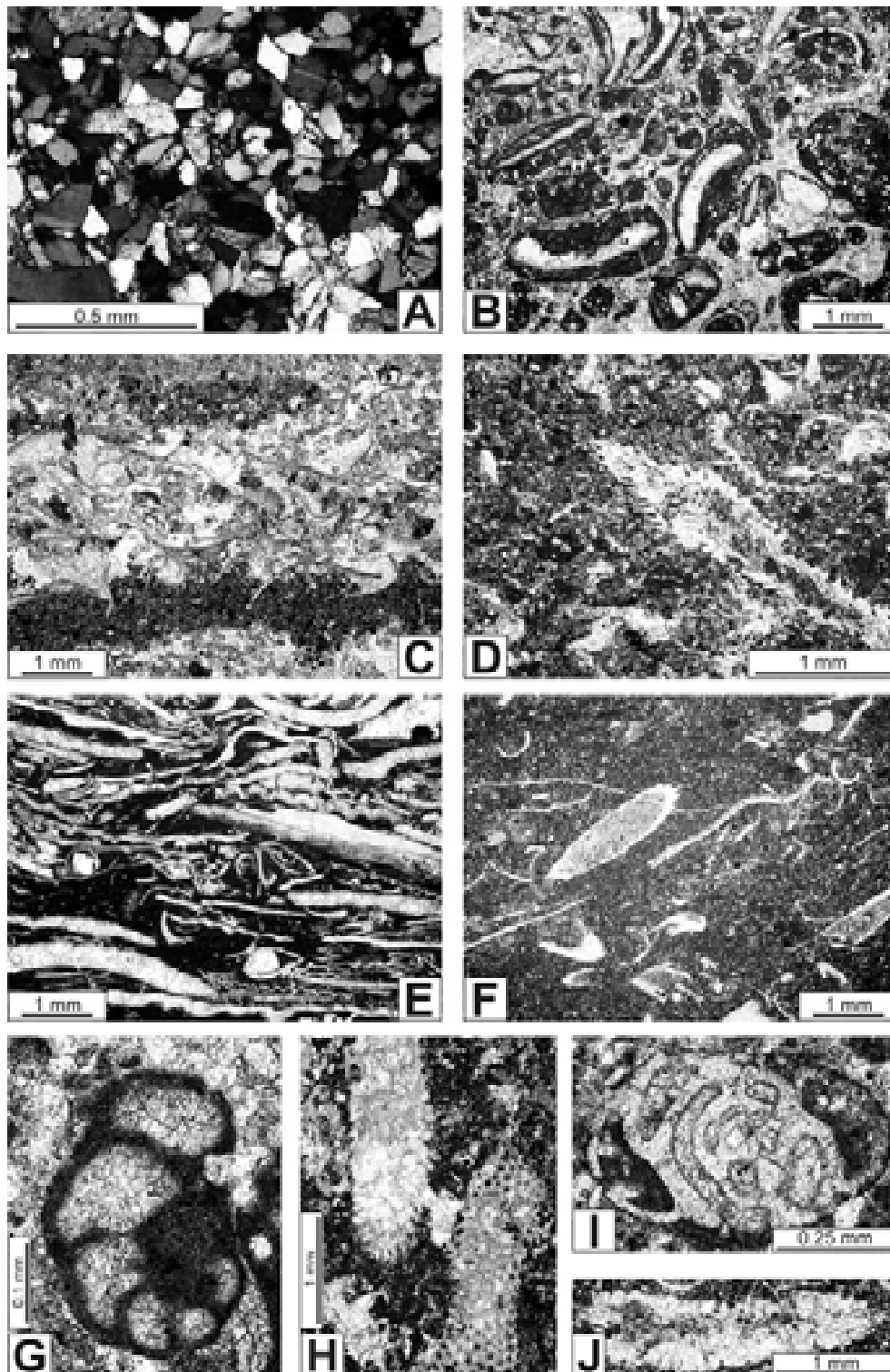


FIGURE 8. Thin sections of selected Yeso Group strata in central New Mexico. **A**, Fine-grained arkosic sandstone composed of detrital quartz grains and many detrital feldspar grains. The sandstone is cemented by authigenic quartz overgrowths and calcite cement. Sample MG 7, Arroyo de Alamillo Formation, Massacre Gap. **B**, Oncoidal rudstone containing abundant oncoids with diameters up to 6 mm, coated grains and skeletal grains embedded in microsparitic matrix with a few small quartz grains. Sample MG 10, Torres Member, Massacre Gap. **C**, Thin rudstone layer composed of abundant, small recrystallized shell fragments and carbonate cement. Sample MG 11, Torres Member, Massacre Gap. **D**, Bioclastic wackestone composed of recrystallized skeletons (including dasycladacean algae) embedded in peloidal micrite. Sample MG 12, Torres Member, Massacre Gap. **E**, Rudstone (coquina layer) composed of abundant brachiopod and bivalve shells aligned parallel to the bedding, and micritic matrix. Sample MG 13, Torres Member, Massacre Gap. **F**, Bioclastic wackestone composed of various skeletal grains, including bivalves and echinoderms embedded in micritic matrix. Sample MLY 1, Torres Member, McLeod Hills. **G**, *Globivalvulina* cf. *G. mosquensis*, a common foraminifer in the limestone facies of the Torres Member. Sample MG 10, Massacre Gap. **H**, **J**, *Gymnocodium*, a common dasycladacean algae in the limestone facies of the Torres Member, Massacre Gap (sample MG 12). **I**, The foraminifer *Orthovertellopsis* from an oncoidal rudstone of the Torres Member at Massacre Gap (sample MG 10).

Arroyo de Alamillo and overlying Los Vallos formations (Lucas and Krainer, 2012). The Arroyo de Alamillo Formation is 42-87 m thick and consists primarily of red-bed sandstone, which is very fine grained, silty, partly gypsiferous and mostly thinly laminated or ripple laminated. Siltstone (mostly massive) is much less common, and there are a few beds of dolomite and gypsum in the Arroyo de Alamillo Formation. At Massacre Gap in the Fra Cristobal Mountains (Fig. 5), the overlying Los Vallos Formation is ~ 234 m thick and can be divided into the Torres Member (~ 134 m of interbedded dolomite, gypsiferous siltstone, gypsum and siltstone to fine-grained sandstone), Cañas Member (~ 58 m of gypsum with some beds of gypsiferous siltstone and dolomite) and the Joyita Member (~ 42 m of red-bed siltstone and fine-grained sandstone) overlain by the Lower Permian Glorieta Sandstone.

A generally similar but much covered section of the Los Vallos Formation is exposed in the McLeod Hills of the southern Caballo Mountains and directly overlain by the San Andres Formation (the Glorieta Sandstone is not present in the Caballo Mountains). However, at nearby Hidden Tank, the Los Vallos Formation is only represented by a 75-m thick section of the Torres Member (Lucas and Krainer, 2012).

Six, dolomite-dominated intervals of the Torres Member can be readily correlated from the Yeso Group type section in Socorro County to the Fra Cristobal Mountains (see below). However, correlation farther south, into the Caballo Mountains is more difficult, due to facies and thickness changes within the Los Vallos Formation and stratigraphic relief of the unconformity at the base of the overlying San Andres Formation (Kottowski et al., 1956; Lucas and Krainer, 2012).

PALEONTOLOGY

In central New Mexico, the Yeso Group has long been perceived of as a relatively unfossiliferous interval. Marine fossils are rare and typically poorly preserved. Those that are good age indicators, such as conodonts and fusulinids, are not present because of the very shallow and often harsh (evaporitic, hypersaline) marine facies. Carbonate rocks (mostly dolomitized limestone) of the Yeso Group do locally contain fragments of bivalves, brachiopods, gastropods, echinoderms, smaller foraminifers, ostracods, rare bryozoans and locally abundant calcareous algae and cyanobacteria (locally forming oncoids). In the Fra Cristobal Mountains, at Massacre Gap, dolomitic limestones of the Torres Member contain species of the following foraminiferal genera: *Baryshnikova?*, *Calcitornella*, *Eotuberitina*, *Geinitzina*, *Globivalvulina* (Fig. 8G), *Hemigordiellina*, *Midiella*, *Orthovertellopsis* (Fig. 8I) and *Palaeonubecularia*. Calcareous algae are represented by species of the genera *Gymnocodium* (gymnocodiacean alga: Fig. 8H, J), *Paraboueina* (codiacean alga) and *Wagonellopsis* (dasycladacean alga). Currently, we (SGL and KK) are working with D. Vachard to present complete documentation of these Yeso microfossils.

In contrast to the rarity of marine fossils, terrestrial to shallow marine fossils are abundant in some of the siliciclastic strata of the Yeso Group, particularly in the Arroyo de Alamillo Formation. During the last few years, about 170 specimens of plant fossils, invertebrate traces and tetrapod footprints have been found by us at 20 localities in the Yeso Group in the Sandia Mountains, the Lucero uplift, the Joyita Hills east of Socorro as well as in the Oscura and Caballo Mountains. Most of the sites are stratigraphically low in the Yeso Group, i.e., near the base of the De Chelly and Arroyo de Alamillo formations.

The stratigraphically highest record of terrestrial fossils of the Yeso Group in central New Mexico is at NMMNH (New Mexico Museum of Natural History) localities 8715 and 8716 in the Sierra de la Cruz area east of Socorro, where fossils occur in pale reddish to greenish-gray beds ~45–50 m above the base of the Arroyo de Alamillo Formation. Invertebrate trace fossils from NMMNH locality 6708 in the Joyita Hills previously referred to floodplain deposits of the upper Abo Formation (Minter and Lucas, 2009) are reinterpreted as shallow marine ichnia in the upper part of the Arroyo de Alamillo Formation (~ 34 m below the base of the Los Vallos Formation).

Plant remains are rare and usually poorly preserved in siliciclastic strata of the Yeso Group. An exception to this are two sites in the Joyita Hills (NMMNH localities 8715 and 8741) discovered in 2012 that yielded some better preserved imprints, including an *in situ* callipterid pteridosperm and various conifer remains (Fig. 9). At both sites, root traces (Fig. 9E) are abundant, indicating that plants must have been common, at least locally. The dominance of non-hygrophilous forms in this flora might be an expression of discontinuous water supply.

Invertebrate traces are remarkably diverse in siliciclastic strata of the lower Yeso Group but not have yet been studied in detail. The invertebrate ichnofauna includes taxa that are well-documented from the underlying Abo Formation and equivalent red beds such as *Augerinoichnus*, *Diplichnites*, *Diplopodichnus*, *Sphaerapus*, and *Stiallia* (Fig. 10A-E; Minter et al., 2008; Minter and Braddy, 2009; Voigt et al., 2013; Lucas et al., 2013). Beyond this, there are specimens that can be compared to *Helminthoidichnites* (NMMNH P-66095/96), *Lithographus* (NMMNH P-66302), *Scoyenia* (NMMNH P-47869), *Stiaria* (NMMNH P-66358), *Striatichnium* (NMMNH P-66094), and *Treptichnus* (NMMNH P-66095/96). This assemblage is clearly dominated by arthropod traces and can be referred to the *Scoyenia* ichnofacies of lowland to coastal plain environments (Minter and Braddy, 2009; Buatois and Mángano, 2011; Voigt et al., 2013).

In contrast to the typical red beds of the Abo Formation, the Arroyo de Alamillo Formation in the Joyita Hills east of Socorro contains beds with invertebrate traces that are more common in marine environments (Fig. 10F-G). One layer with *Thalassinoides* burrows up to 18 cm in diameter occurs 8 m above the base of the Arroyo de Alamillo Formation in the Cerros de Amado area and can be laterally traced for several hundred meters but does not have an equivalent at other places of the Joyita Hills. The *Thalassinoides* bed clearly indicates a shallow marine incursion near the base of the Yeso Group. A nearshore environment is also evident with respect to the assemblage of *Cruziana problematica*, *Diplichnites gouldi*, *Diplichnites* isp., *Monomorphichnus* isp., *Palaeophycus tubularis*, *Rusophycus carbonarius*, and *Striatichnium* cf. *S. natalis* described by Minter and Lucas (2009) from NMMNH locality 6708.

The lower Yeso Group of central New Mexico yields tetrapod tracks we assign to *Batrachichnus salamandroides*, *Varanopus curvidactylus*, *Erpetopus willistoni*, and *Dromopus lacertoides* (Fig. 11), which most likely were produced by small temnospondyls, captorhinomorphs, and araeoscelid diapsids. There are at least two more kinds of captorhinomorph tracks from the Arroyo de Alamillo Formation (Fig. 11D-E), but specimens are too poorly preserved for an ichnotaxonomic assignment. The tetrapod ichnofauna of the basal Yeso Group is characterized by its predominance of captorhinomorph tracks (*Varanopus curvidactylus*, in particular) that are more common than the otherwise ubiquitous *Dromopus*. The scarcity or even absence of temnospondyl (*Batrachichnus*, *Limnopos*), reptiliomorph (*Amphisauropus*, *Ichniotherium*) and “pelycosaur” (*Dimetropus*, *Tambachichnium*) tracks may be partially due to the nearshore environment (Voigt and Lucas, 2012). A tetrapod ichnofaunal change, however, is unambiguously indicated by the rarity of *Batrachichnus* and *Dimetropus* that together with *Dromopus* constitute more than 95% of all tetrapod tracks of the slightly older but facially similar Robledo Mountains Formation in the Robledo Mountains of southern New Mexico (Lucas and Heckert, 1995; Lucas et al., 1998, 2011; Minter and Braddy, 2009; Voigt et al., 2013). According to preliminary analyses of the vertical distribution of tetrapod tracks in the Early Permian of New Mexico, this ichnofaunal change almost coincides with the Abo-Yeso boundary and was probably climatically driven. We (SV and SGL) will present a more detailed analysis of this ichnofaunal change elsewhere.

AGE

None of the fossils now known from the Yeso Group in central New Mexico are precise age indicators. Thus, stratigraphic position and

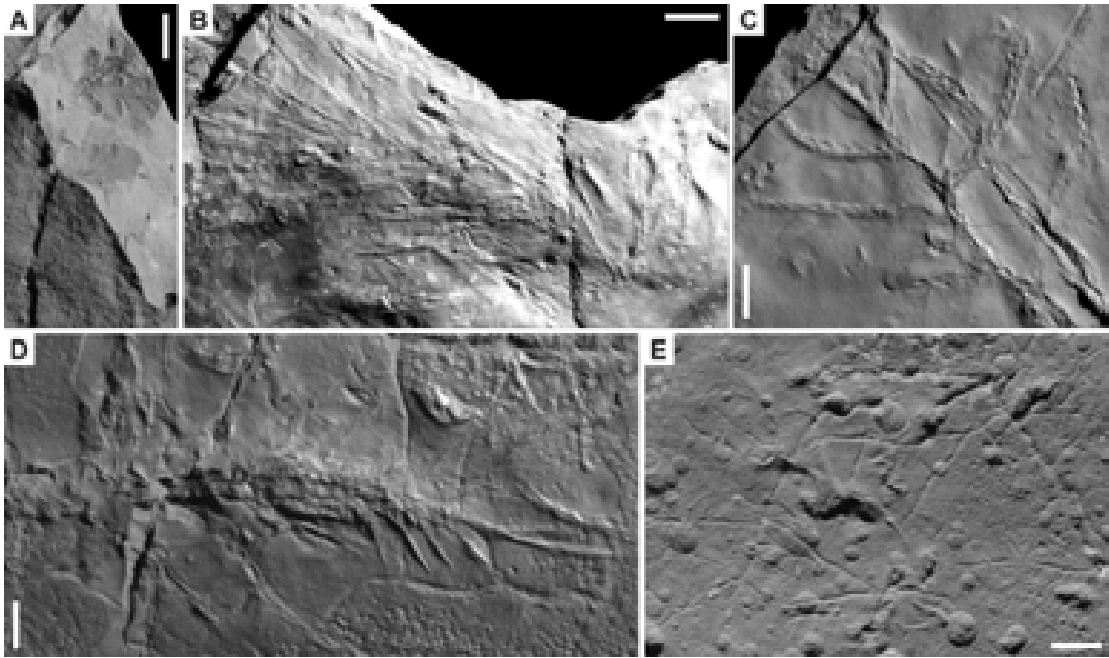


FIGURE 9. Plant fossils and root traces from the lower Yeso Group (Arroyo de Alamillo Formation) of central New Mexico. **A**, Callipterid pteridosperm, *in situ* preservation. **B**, Branch fragment of *Dicranophyllum* sp. **C**, Walchian conifer remain. **D**, Conifer cone impression. **E**, Horizontal root traces on surface with raindrop imprint marks. **Material:** **A**, NMMNH P-66114; **B**, NMMNH P-66320; **C**, NMMNH P-66323; **D**, NMMNH P-66324; **E**, NMMNH P-66099. **A**, **E**, NMMNH locality 8715, **B-D**, NMMNH locality 8741. Scale bars equal 1 cm.

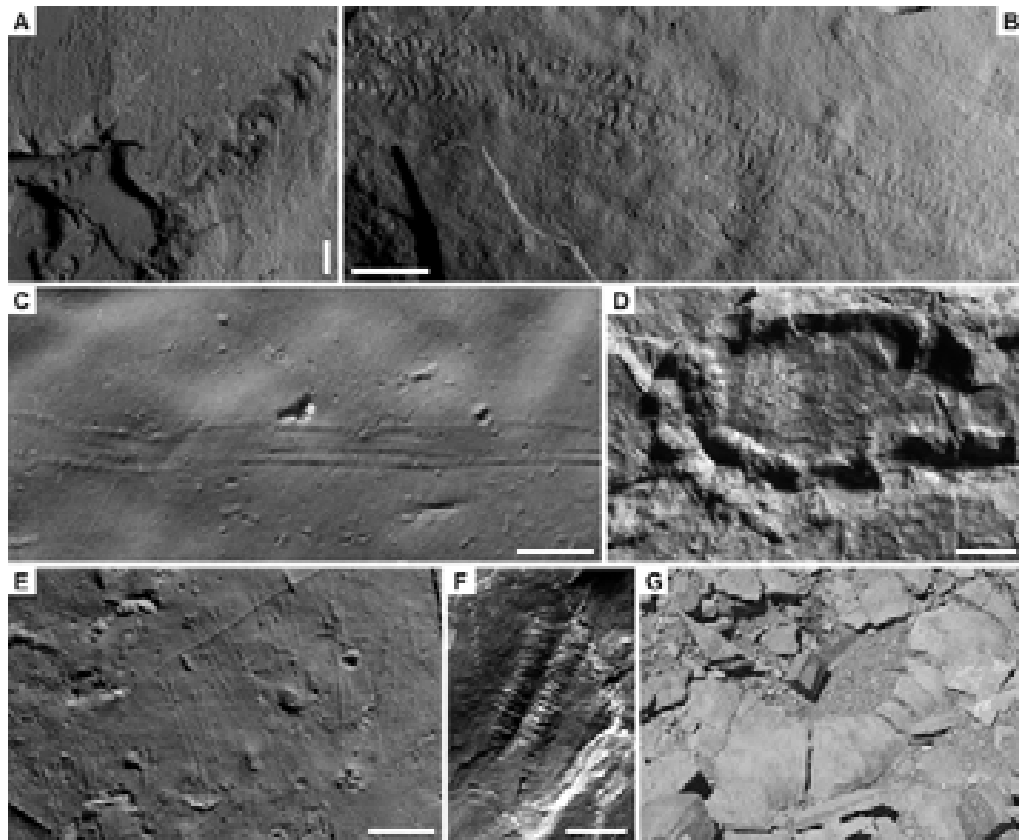


FIGURE 10. Invertebrate traces from the lower Yeso Group (Arroyo de Alamillo Formation) of central New Mexico. **A**, *Augerinoichnus helicoidales*. **B**, *Diplichnites* isp. **C**, *Diplopodichnus biformis* with transition into *Diplichnites* isp. **D**, *Sphaerapus larvalis*. **E**, *Stiallia pillosa*. **F**, *Cruziana problematica*. **G**, *Thalassinoides* isp. **Material:** **A**, NMMNH P-66224; **B**, NMMNH P-66269; **C**, NMMNH P-66108; **D**, NMMNH P-66359; **E**, NMMNH P-24673; **F**, NMMNH P-51591. **A**, NMMNH locality 8738, **B**, NMMNH locality 8742, **C**, **E**, NMMNH locality 8715, **D**, NMMNH locality 8741, **F**, NMMNH locality 6708, **G**, field photograph, Cerros de Amado, east of Socorro. **G** is preserved in full relief, all other specimens in convex hyporelief. Scale bars equal 1 cm. Burrow cast in **G** ~18 cm in diameter.

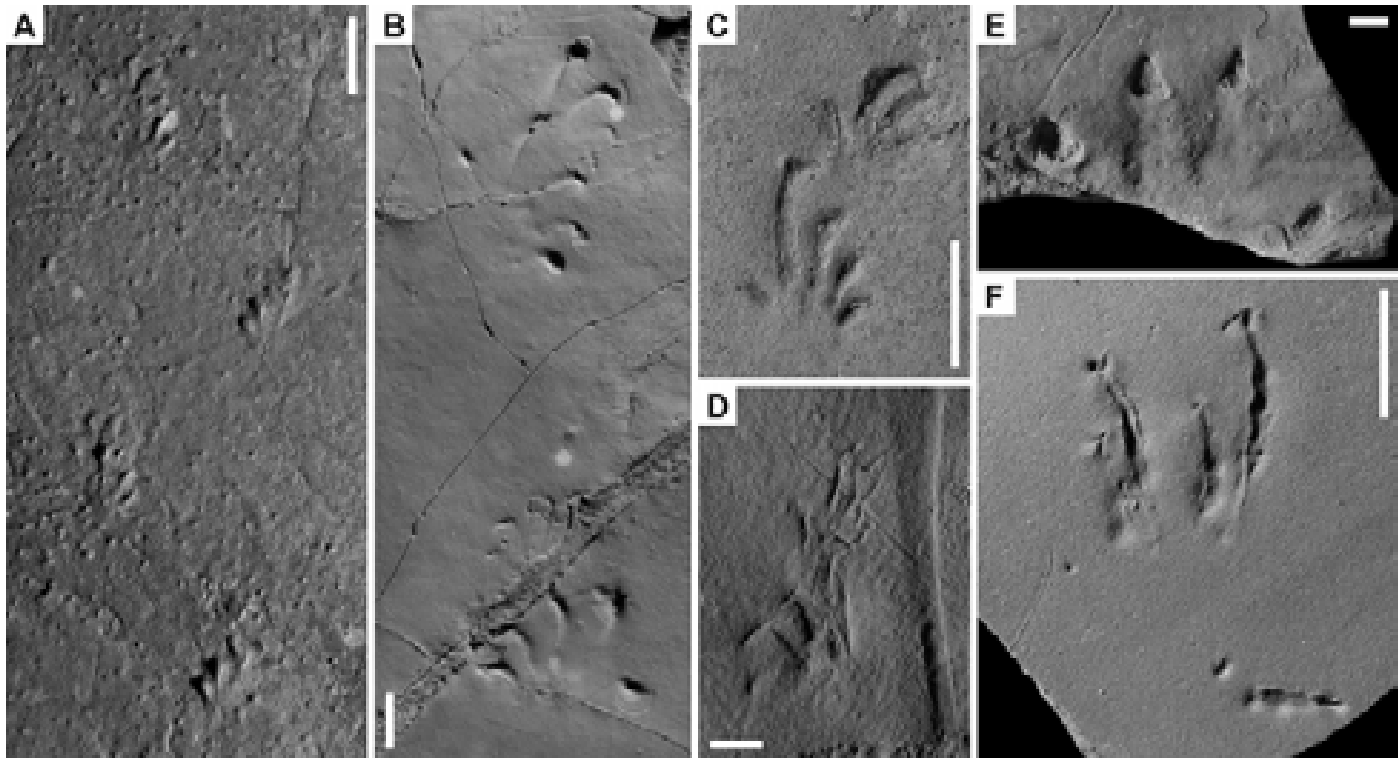


FIGURE 11. Tetrapod footprints from the lower Yeso Group (Arroyo de Alamillo Formation) of central New Mexico. **A**, *Batrachichnus salamandroides*, trackway of four manus-pes sets. **B**, *Varanopus curvidactylus*, incomplete trackway of two manus-pes sets. **C**, *Erpetopus willistoni*, left manus-pes set. **D**, Captorhinomorph tracks indet., left manus-pes set with tail/body trace. **E**, Captorhinomorph track indet., incomplete imprint of right pes or manus. **F**, *Dromopus lacertoides*, right manus-pes set. **Material:** **A**, NMMNH P-66338; **B**, NMMNH P-66265; **C**, NMMNH P-24592; **D**, NMMNH P-24591; **E**, NMMNH P-24673; **F**, NMMNH P-62516. **A-B**, NMMNH locality 8741, **C-F**, NMMNH locality 2771. Scale bars equal 1 cm.

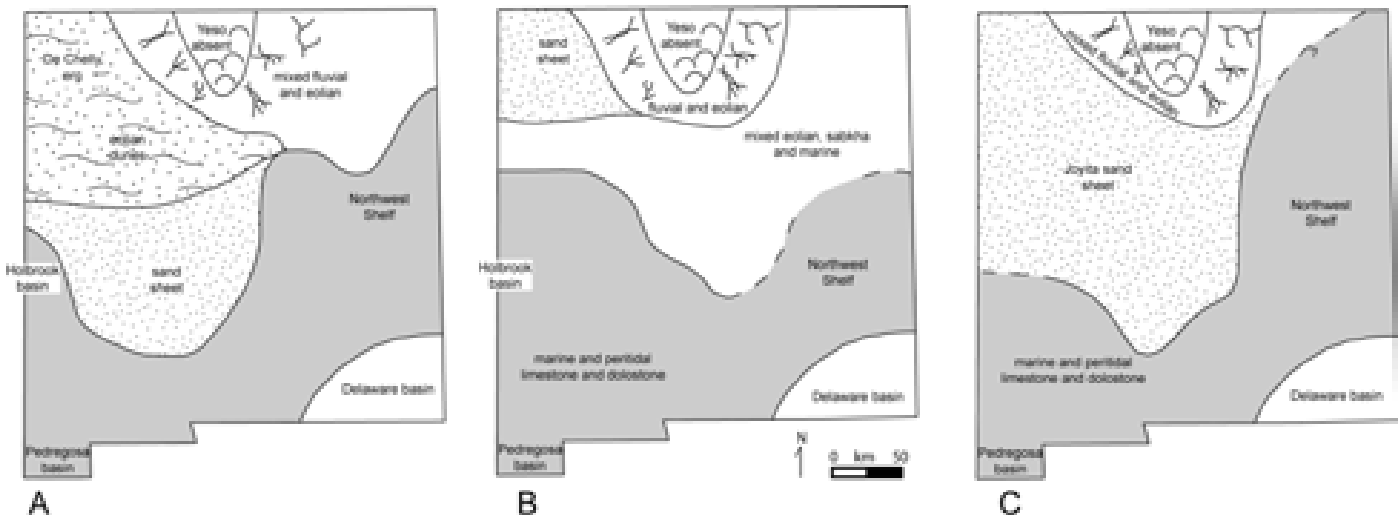


FIGURE 12. Generalized paleogeography of New Mexico during deposition of the Yeso Group. **A**, De Chelly-Arroyo de Alamillo formations. **B**, Torres and Cañas members of the Los Vallos Formation. **C**, Joyita Member of Los Vallos Formation. Modified from Mack and Dinterman (2002).

regional stratigraphic relationship are the basis for assigning the Yeso an early-middle Leonardian age (e.g., Lucas et al., 2005; Lucas and Krainer, 2012)

DEPOSITIONAL ENVIRONMENTS

General

Yeso Group deposition took place along the vast northwestern shelf of the Permian basin during part of Early Permian (Leonardian) time (Fig. 12). To the northwest, the De Chelly erg, which covered much of

the Four Corners and southern Colorado Plateau, represents the northwestern limit of the Yeso lithosome. To the southeast, in the Permian basin, Yeso strata grade into/interfinger with marine strata of the Bone Spring Formation. The vast area in between these extremes was of low relief but was nonetheless one variably covered by coastal dunes and sandflats, wadis, sabkhas and shallow marine shelves.

De Chelly and Arroyo de Alamillo formations

During deposition of the De Chelly and Arroyo de Alamillo formations, marine carbonates were deposited in the Delaware and Pedregosa

basins to the south, grading into shallow-marine siliciclastic, carbonate and evaporate sediments with intercalated eolian deposits towards the northern and northwestern shelf, and into eolian sand sheets in west-central New Mexico, eolian dune fields in the northwest and ephemeral streams and eolian deposits in north-central New Mexico (Fig. 12A). Deposition of the Arroyo de Alamillo Formation took place by eolian and fluvial processes on an arid coastal plain during a time of regionally low sea level (Baars, 1974; Stanesco, 1991; Mack and Suguio, 1991; Mack and Dinterman, 2002; Lucas and Krainer, 2012).

Los Vallos Formation

Deposition of the Torres and Cañas members of the Los Vallos Formation was in cyclically shallow marine, sabkha and eolian coastal plain environments (Fig. 12B). In Socorro and much of Sierra counties, the Torres and Cañas members of the Los Vallos Formation are composed of similar lithologies that are arranged to form six depositional cycles (Figs. 4-5, 13). Various types of dolomite formed in a shallow marine, mostly restricted environment, as indicated by a low-diversity biota, partly in evaporitic settings documented by vuggy dolomite containing gypsum and halite, and rarely under more normal, open marine conditions represented by dolomite containing a diverse fossil assemblage. Coquina layers are interpreted as storm deposits (tempestites).

Gypsiferous siltstone and gypsum formed in sabkha environments, and siltstone to fine-grained sandstone are interpreted as eolian deposits. Hunter and Ingersoll (1981) interpreted the evaporitic sediments of the Cañas Member as deposits of a hypersaline lagoon that covered an area of approximately 25,000 km², though this is a minimum as the Cañas Member has a much greater areal extent than recognized by Hunter and Ingersoll (see, for example, Colpitts, 1989).

Joyita Member deposition again records eolian and fluvial processes during a time of regionally low sea level (Fig. 12C). Thus, during late Yeso time, during deposition of the Joyita Member, an extensive eolian sand sheet prograded towards the south to overlie shallow marine, particularly evaporitic deposits (Mack and Dinterman, 2002).

YESO DEPOSITIONAL CYCLES

One of the most striking aspects of Yeso regional lithostratigraphy is the consistent architecture of the Los Vallos Formation from the Lucero uplift in Valencia County (Lucas and Zeigler, 2004), through the Yeso type section in Socorro County (Lucas et al., 2005) to Massacre Gap in the southern Fra Cristobal Mountains of Sierra County (Lucas and Krainer, 2012). Thus, the Torres Member of the Los Vallos Formation at these three sections contains six obvious transgressive-regressive cycles marked by six intervals of dolomite separated by thicker intervals of siltstone, gypsum and minor carbonate and sandstone beds.

The dolomite intervals in the Torres Member represent periods of relatively high sea-level when large areas were covered by a shallow sea with locally increased salinity. During periods of lowered sea-level, evaporitic and eolian sediments were deposited. At the Yeso type section and at Massacre Gap, strata above the Torres Member are a relatively thick gypsum-dominated interval (the Cañas Member) overlain by red-bed sandstones and siltstones, the Joyita Member. This architectural similarity allows a straightforward correlation of the Yeso Group from the Yeso type section to Massacre Gap (Fig. 13).

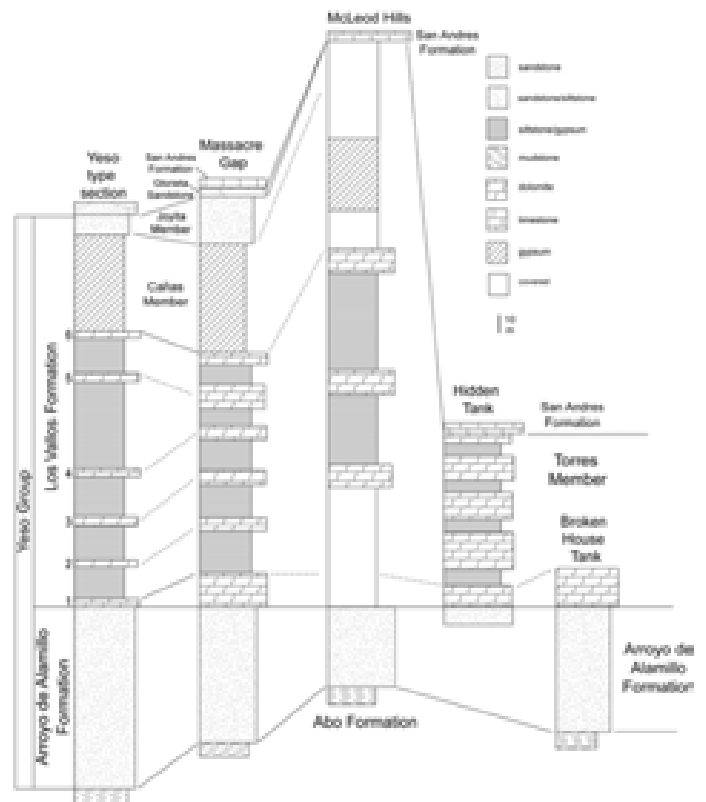


FIGURE 13. Correlation of selected Yeso Group sections, including the type Yeso section in Socorro County and the Yeso Group section in the Fra Cristobal and Caballo Mountains of Sierra County (from Lucas and Krainer, 2012). Note continuity of dolomite beds from Yeso Group type section in Socorro County to Massacre Gap in the Fra Cristobal Mountains of Sierra County, a distance of ~ 100 km.

The cycles of deposition of the Torres and Cañas members of the Los Vallos Formation are striking in their areal extent. They may well correlate to at least some of the third-order cycles evident in the platform deposits of the broadly correlative Leonardian Hess Formation in West Texas (Ross and Ross, 1995). These cycles may be the product of glacio-eustatic sea-level fluctuations. Although the maximum extent of the Gondwana glaciation was earlier, during the Late Pennsylvanian-Wolfcampian, smaller ice sheets existed in Australia and Siberia throughout the late Early Permian and Middle Permian (Isbell et al., 2003; Fielding et al., 2008), which may have caused minor sea-level fluctuations sufficient to drive the cyclic succession evident in the Los Vallos Formation of the Yeso Group.

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At the Lower Permian Yeso Group type section, northeast of Socorro, the middle part of the section is the Torres Member of the Los Vallos Formation. Seen here, these strata are interbedded gypsum, siltstone and thin, laterally persistent limestone beds. The evident cyclicity of these strata may be driven by eustatic sea-level cycles, but a match to published Early Permian sea-level curves is not obvious.

THE LOWER PERMIAN GLORIETA SANDSTONE IN CENTRAL NEW MEXICO

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Abstract—In central New Mexico, the Lower Permian Glorieta Sandstone is a relatively thin (< 100 m thick), prominent, sandstone-dominated stratigraphic unit near the top of the Permian section. The Glorieta Sandstone is generally a yellowish brown to light gray, very fine to fine-grained, well-sorted quartzose sandstone that is often crossbedded but also includes beds that are ripple laminated and/or tabular bedded. Beds of limestone and gypsum are often intercalated with sandstone strata and are thus mapped with (included in) the Glorieta Sandstone. The Glorieta Sandstone is either a persistent sandstone unit intercalated between early-middle Leonardian Yeso Group strata (below) and late Leonardian-Guadalupean San Andres Formation strata (above) or multiple sandstone bodies interbedded with the lower part of the San Andres Formation. We describe three characteristic sections of the Glorieta Sandstone in central New Mexico, at Cedar Crest in Bernalillo County, Chupadera Mesa in Torrance County and Mesa del Yeso in Socorro County. Published studies of sedimentology of the Glorieta Sandstone are in apparent conflict because they suggest it is either of dominantly shallow marine or of eolian origin. The Glorieta Sandstone is continuous to the northwest with the Coconino erg of northern Arizona. It was deposited mostly by eolian processes but includes eolian sands that were reworked and redeposited in shallow marine environments.

INTRODUCTION

In central New Mexico, one of the most distinctive Permian stratigraphic units is the Glorieta Sandstone. Much less than 100 m thick, the Glorieta is dominantly yellowish brown, very mature quartzarenite that forms a cliff or ledge near the top of the local Permian section. Here, we present a brief synopsis of Glorieta stratigraphy and sedimentation in central New Mexico (Fig. 1).

HISTORY OF STUDY

Keyes (1915 a, b) first used the term Glorieta Sandstone as the “main body of the Dakotan around the southern end of the Rocky Mountains.” He thus considered it to be of Cretaceous age, and evidently took the name from Glorieta Mesa in eastern Santa Fe-southwestern San Miguel Counties (Fig. 1). However, other workers rapidly recognized the Glorieta Sandstone (usually as the upper member of Lee’s [1909] Yeso Formation) as of Permian age (e. g., Rich, 1921). Note, though, that Baker (1920) regarded the Glorieta as of Triassic age (he correlated it to the Upper Triassic Santa Rosa Formation), an incorrect correlation rejected by Rich (1921), among others.

Darton (1928) was unable to consistently separate the Yeso and San Andres formations of Lee (1909) and the Glorieta Sandstone of Keyes (1915 a, b). He, instead, combined them into the Chupadera Formation, a unit abandoned long ago (Lucas, 2009). Keyes (1935) first recommended abandoning the term Chupadera Formation. But, what really undermined the Chupadera Formation were U.S. Geological Survey maps in central New Mexico that distinguished and separately mapped the Yeso, Glorieta and San Andres formations (Kelley and Wood, 1946; Wilpolt et al., 1946; Wood and Northrop, 1946; Wilpolt and Wanek, 1951).

Needham and Bates (1943) designated a type section of the Glorieta Sandstone on Glorieta Mesa near Rowe in San Miguel County (Figs. 1-2). Here, they described it as 41 m of white, gray and buff quartzose sandstone above Yeso strata and below limestone of the San Andres Formation. Needham and Bates (1943, p. 1664) concluded that “on account of its wide distribution, persistence of lithology, bold topographic expression, and stratigraphic importance, the Glorieta is considered to be a formation.” Nevertheless, many subsequent workers recognized the Glorieta as a member of the San Andres Formation (e.g., Read



FIGURE 1. Map of New Mexico showing key Glorieta Sandstone outcrops referred to in the text.

et al., 1944; Wilpolt and Wanek, 1951; Bachman, 1953; Kelley, 1971, 1972; Milner, 1976, 1978). This is largely because at many locations the Glorieta Sandstone is either a thin sandstone unit (<10 m thick) at the base of the San Andres Formation section or it is two or more sandstone intervals interbedded with limestones of the lower part of the San Andres Formation. That is to say, the Glorieta intertongues with the lower San Andres. Thus, by the 1950s, the Glorieta Sandstone was recognized across central New Mexico as a prominent interval of mostly quartz-rich

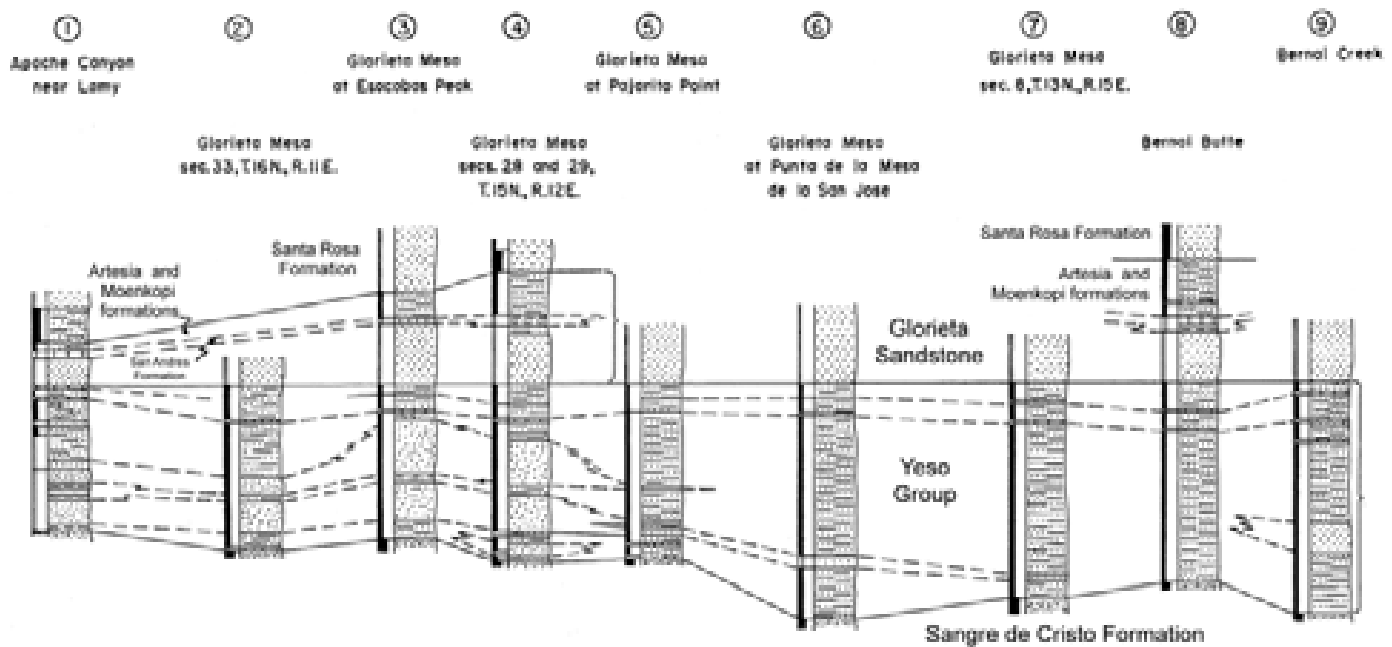


FIGURE 2. Stratigraphic sections from near Lamy (Santa Fe County) across Glorieta Mesa to Bernal Creek (San Miguel County), showing distribution and thickness of the Glorieta Sandstone in its type area (after Baltz et al., 1956).

sandstone (sometimes, because of secondary silicification, incorrectly referred to as a metaquartzite) of Early Permian (Leonardian) age.

However, earlier, Lang (1937, p. 850), working in southeastern New Mexico, had named the Hondo sandstone member (of the Chupadera Formation), noting that "it has never been definitely shown that this particular sandstone is wholly or in part the sandstone of Glorieta Mesa, in northern New Mexico." Lang (1937) described the Hondo Sandstone as "about 56 feet [17 m] thick" and "variably streaked yellowish to brownish red... composed of coarse white quartz grains... crossbedding is apparent." Needham and Bates (1943) considered the Hondo Sandstone to be the same as the Glorieta Sandstone, and we concur. Thus, Lang (1937) evidently renamed the Glorieta Sandstone lithosome in southeastern New Mexico the Hondo Sandstone. Despite this, Harbour (1970) claimed that the Glorieta and Hondo are separate lithostratigraphic units and regarded both as members of the San Andres Formation. Harbour (1970) thus applied the name Glorieta Sandstone to a sandstone interval at the base of the San Andres Formation and restricted the name Hondo to a stratigraphically higher sandstone interval in the lower San Andres Formation across parts of Otero and Lincoln Counties.

STRATIGRAPHIC NOMENCLATURE

Although many workers during the 1940s-1970s considered the Glorieta Sandstone to be a member of the San Andres Formation, since the 1980s the trend has been to regard it as a separate formation (e.g., Woodward, 1987; Colpitts, 1989; Lucas et al., 1999; Lucas and Zeigler, 2004; Mack and Dinterman, 2002; Kues and Giles, 2004). This has particularly been the case in quadrangle mapping by STATEMAP from the Jemez Mountains of Sandoval County through the Cerros de Amado of Socorro County. Connell (2008), who compiled all of the STATEMAP results from the Albuquerque area, is a good example of recent use of Glorieta Sandstone as a formation-rank unit.

We conclude that the Glorieta is a lithologically distinctive unit almost always mappable at a scale of 1:24,000. Therefore, the Glorieta Sandstone merits formation rank as either a sandstone sheet beneath the San Andres Formation or a basal sandstone sheet and stratigraphically higher sandstone bed(s) interbedded with carbonate/gypsum beds that are lithologies characteristic of the Yeso Group and/or the San Andres

Formation. Distinction of an upper sandstone of Glorieta lithology as the Hondo Sandstone Member of the San Andres Formation appears to have merit only locally (Harbour, 1970). However, the regional interrelationships of Glorieta sandstone bodies and lithologies characteristic of the San Andres Formation merit more study, along the lines of Milner (1978).

LITHOSTRATIGRAPHY

Lithology

The Glorieta Sandstone is generally a very fine to fine-grained, well-sorted quartzose sandstone that is often crossbedded but also includes beds that are ripple laminated and/or tabular bedded (Figs. 3-7). The sandstone is locally very strongly cemented and so well indurated that it has the texture of a metaquartzite. Typically, the sandstone is yellowish brown to light gray. Beds of limestone and gypsum are often intercalated with sandstone strata and are thus mapped with (and included in) the Glorieta Sandstone.

In thin section (Fig. 3), sandstone of the Glorieta Sandstone is commonly fine-grained with grain-size ranging mostly from 0.1 to 0.3 mm. Most of the sandstones are well sorted, and the detrital grains are subrounded to rounded; larger grains may be well rounded. Monocrystalline quartz is by far the most common grain type. Other grain types such as polycrystalline quartz, chert, feldspar, rock fragments and micas (muscovite) are rare. Polycrystalline quartz also includes grains composed of elongate crystals ("stretched metamorphic"), indicating a metamorphic source rock. Detrital feldspar grains are fresh or appear as altered grains, some of which may be completely altered to clay minerals, forming a "pseudomatrix." Most common is potassium feldspar (including microcline and untwinned grains), and plagioclase is rare. Accessory minerals are tourmaline and zircon.

Due to the mineralogical composition most of the sandstones plot into the field of quartzarenite, and some into the field of subarkose after the classification scheme of Pettijohn et al. (1987). At Cedar Crest (Sandia Mountains: Fig. 4), sandstone from the uppermost sandstone interval of Glorieta lithology contains a higher amount of detrital feldspar, particularly microcline, but this sandstone still plots into the field of sublitharenite.

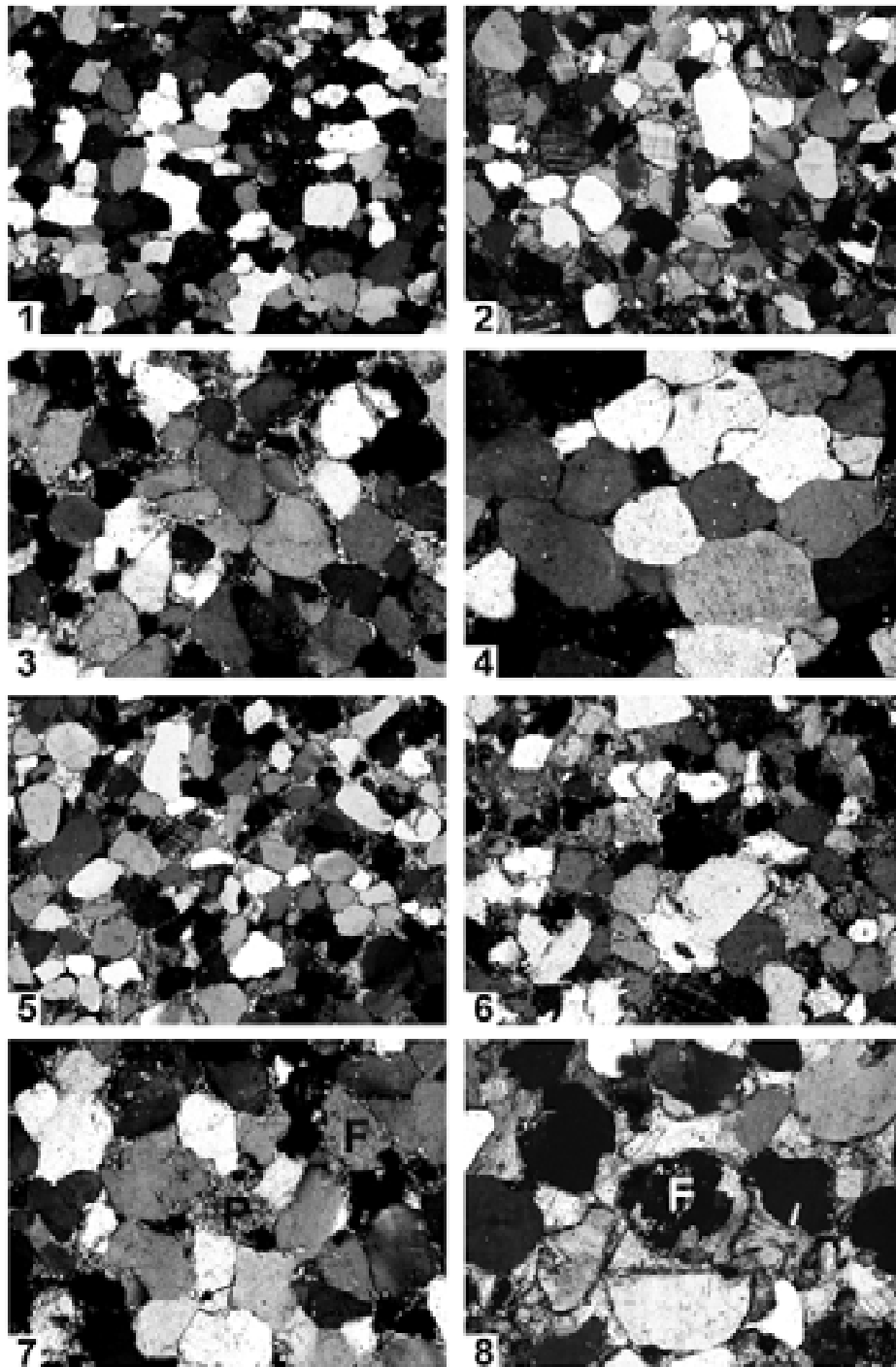


FIGURE 3. Thin section photographs showing texture and mineralogical composition of the Glorieta Sandstone. All photographs taken under polarized light. **1**, Well-sorted sandstone (quartzarenite) composed of quartz grains and very rare feldspar grains, cemented by authigenic quartz overgrowths. Sample SGL 12, Mesa del Yeso, width of photograph is 3.2 mm. **2**, Well-sorted sandstone (subarkose) composed of subrounded to rounded quartz grains and many potassium feldspar grains (mostly microcline) cemented by coarse, blocky calcite. Sample GSM 37, Cedar Crest, width of photograph is 3.2 mm. **3**, Well-sorted quartzarenite containing small amounts of matrix, partly pseudomatrix (center right). Some of the quartz grains display authigenic overgrowths. Sample SGL 12, Mesa del Yeso, width of photograph is 1.2 mm. **4**, Well-sorted quartzarenite cemented by quartz that occurs as very visible authigenic overgrowths on the detrital quartz grains. Sample SGL 12, Mesa del Yeso, width of photograph is 1.2 mm. **5**, Well-sorted and well-rounded quartzarenite containing a few small detrital feldspar grains. Sandstone is cemented by coarse calcite, and quartz overgrowths are absent. Sample GSM 22, Cedar Crest, width of photograph is 3.2 mm. **6**, Quartzarenite, cemented by quartz (thin overgrowths around some of the detrital quartz grains) and calcite cement. Sample GSM 14, Cedar Crest, width of the photograph is 1.2 mm. **7**, Quartzarenite composed of abundant quartz grains, rare feldspar (F), some matrix, and a grain in the center that is completely altered to clay minerals (pseudomatrix, P), probably a former detrital feldspar grain. The sandstone is cemented by authigenic overgrowths (arrow). Sample GSM 26, Cedar Crest, width of the photograph is 1.2 mm. **8**, Quartzarenite with rare potassium feldspar grains. The feldspar grain in the center (F) is partly replaced by calcite cement that randomly also replaces detrital quartz grains. Sample GSM 37, Cedar Crest, width of the photograph is 1.2 mm.

Cedar Crest (SMNHC)

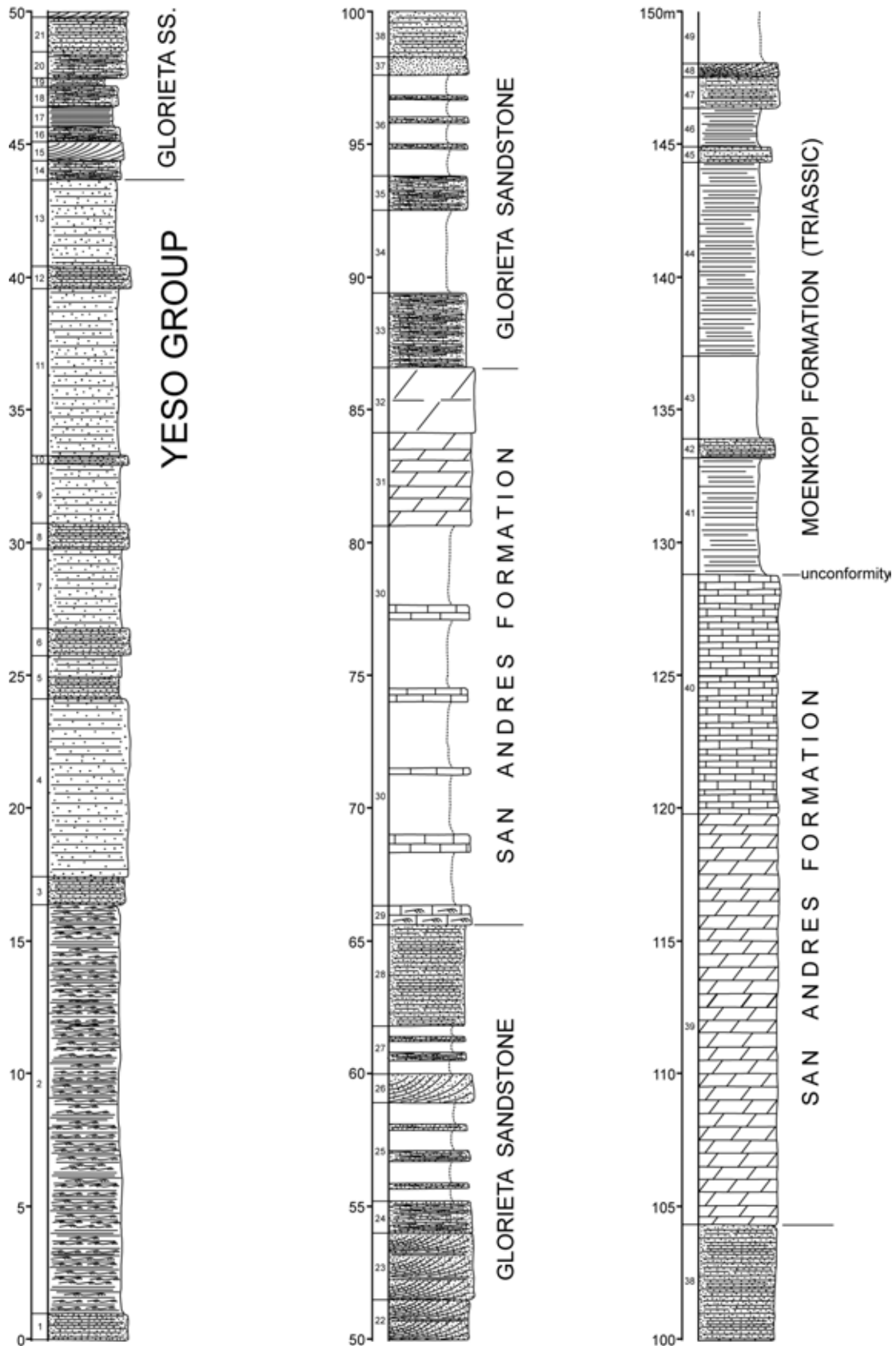


FIGURE 4. Measured stratigraphic section of Glorieta Sandstone at Cedar Crest (Sandia Mountain Natural History Center) in Bernalillo County. See Figure 5 for legend to lithologic symbols.

Chupadera Mesa (Deer Canyon Preserve)

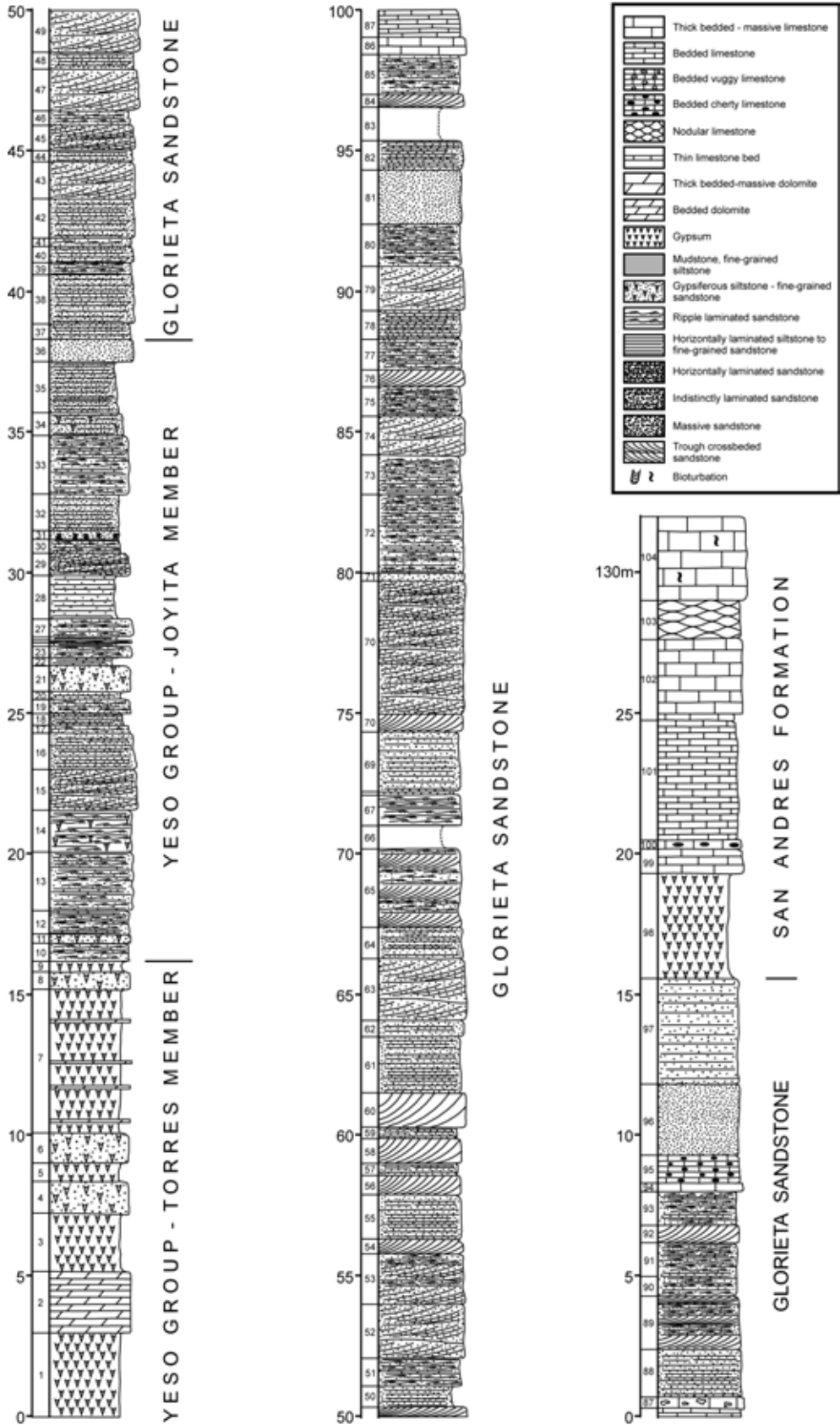


FIGURE 5. Measured stratigraphic section of Glorieta Sandstone at Chupadera Mesa (Deer Canyon Preserve) in Torrance County.

At Mesa del Yeso east of Socorro (Fig. 6), sandstone of the Glorieta Sandstone is commonly cemented by quartz, which occurs as authigenic overgrowths around detrital quartz grains. Some of the sandstone beds contain small amounts of clayey matrix, including pseudomatrix formed by the alteration of detrital feldspar grains (Fig. 3.3). Calcite cement is rare.

At Cedar Crest different types of sandstone can be distinguished with regard to their diagenesis:

1. Sandstone containing small amounts of matrix and cemented by quartz that occurs as quartz overgrowths. Calcite cement is very rare or absent.

2. Sandstone locally containing small amounts of matrix and cemented by coarse, blocky calcite cement that replaces feldspar and quartz grains. Quartz overgrowths on detrital quartz grains are absent.

3. Sandstone in which many quartz grains display thin authigenic overgrowths; the remaining pore space is filled with calcite cement that may randomly replace detrital feldspar and quartz grains.

The textural parameters (sorting, rounding) and mineralogical composition of the sandstone samples we studied of the Glorieta Sandstone (Fig. 3) are very similar to those of sandstones studied by Milner (1978) from different localities in New Mexico. This indicates that the composition and texture of the Glorieta Sandstone is very uniform over large areas.

Contacts

Across central New Mexico, the Glorieta Sandstone has a sharp contact on underlying Yeso Group strata (Figs. 4-7). This contact may locally appear to be gradational and interfingering regionally (Milner, 1978). Glorieta Sandstone strata also have a complex and interbedded relationship with overlying, carbonate-dominated strata of the San Andres Formation. This suggests that the lower and upper contacts of the Glorieta may be conformable. However, the base of the San Andres Formation is reported to be a regional, transgressive unconformity (e.g., Sarg and Lehman, 1986). Whether that unconformity is at the base or within the Glorieta Sandstone merits further study.

Thickness

In central New Mexico, the Glorieta Sandstone is 0 to 100 m thick. In general, thickness increases to about more than 150 m in a belt running from Guadalupe County northeast into southern Harding County toward the southeast (Milner, 1978, p. 10). In parts of Sierra County, the Glorieta Sandstone is either absent (Caballo Mountains: Kelley and Silver, 1952; Lucas and Krainer, 2012) or thin (< 10 m thick) and/or patchy (Fra Cristobal and San Andres Mountains: Kottlowski et al., 1956; Lucas and Krainer, 2012). To the north, in Socorro and Valencia Counties, the Glorieta is a relatively thick and persistent unit, with some interbedded carbonate and gypsum beds (Fagrelus, 1982; Lucas and Zeigler, 2004). Farther north, in Bernalillo and Sandoval Counties, the Glorieta is again relatively thin, but to the northeast, in San Miguel County, the formation reaches thicknesses in excess of 100 m (Milner, 1978). To the west, the Glorieta Sandstone appears to merge with the Coconino Sandstone of northern Arizona (Middleton et al., 1990).

Distribution

The Glorieta Sandstone is present across central New Mexico, from Sandoval County to Sierra County (Fig. 1). Here, we describe in detail three measured sections of the Glorieta Sandstone in central New Mexico: (1) at the Sandia Mountain Natural History Center in Cedar Crest, Bernalillo County; (2) on the flank of Chupadera Mesa near Mountainair in Torrance County; and (3) overlying the Yeso Group type section in Socorro County (Figs. 4-7).

Cedar Crest

Cedar Crest lies on the east flank of the Sandia Mountains on the

north side of Interstate 40. At Cedar Crest, the Glorieta Sandstone overlies the San Ysidro Formation of the Yeso Group as a sandstone unit that separates it from limestone and dolomite of the lower part of the San Andres Formation (Fig. 4). Approximately 22 m above the base of the San Andres Formation, another sandstone sheet of Glorieta lithology, 18 m thick, is present. STATEMAP thus mapped the Glorieta Sandstone here as two distinct units intercalated in the lower San Andres Formation. However, another stratigraphic interpretation would be to assign the upper sandstone interval to the Hondo Sandstone Member of the San Andres Formation.

The lower sandstone unit of the Glorieta Sandstone is composed of the following lithofacies: (1) horizontally laminated sandstone, up to 3.8 m thick; (2) ripple-laminated sandstone, up to 1.2 m thick; (3) wavy to horizontally laminated sandstone, up to 1 m thick; (4) sandstone and siltstone displaying trough crossbedding, up to 2.5 m thick; and (5) red siltstone, up to 0.7 m thick. The lower sandstone unit is overlain by 21 m of strata characterized by resistant beds of muddy limestone and thick-bedded to massive dolomitic limestone assigned to the San Andres Formation. The overlying sandstone unit is composed of horizontally laminated, ripple-laminated and massive sandstone.

Chupadera Mesa

At Deer Canyon Preserve on the western flank of Chupadera Mesa near Mountainair, we measured an interval ~ 77 m thick of Glorieta Sandstone overlying the Joyita Member of the Los Vallos Formation of the Yeso Group (Fig. 5). In the upper part of this Glorieta section are two thin (< 5 m thick) limestone intervals (tongues) of San Andres lithology. The base of the main body of the San Andres Formation is a gypsum bed.

The main lithologies of the Glorieta Sandstone at the Deer Canyon Preserve section are: (1) horizontally-laminated sandstone, thin to thick bedded (tabular) – individual units are up to 2 m thick; (2) ripple-laminated sandstone, up to 1.9 m thick; (3) massive sandstone, rare, up to 2.5 m thick; (4) crossbedded sandstone, individual units are up to 5.4 m thick (eolian crossbedding?); (5) sandstone with large-scale eolian crossbedding (rare), 0.7 m thick; and (6) two covered intervals, 0.8 and 1.2 m thick. In the upper part of the Glorieta Sandstone two limestone intervals are intercalated that are 2.3 and 1.3 m thick and composed of laminated, thick-bedded lime mudstone and thin-bedded wackestone.

Mesa del Yeso

Mesa del Yeso in Socorro County is the type section of the Yeso Group (Needham and Bates, 1943; Lucas et al., 2005). Here, about 87 m of Glorieta Sandstone overlie the Joyita Member of the Los Vallos Formation of the Yeso Group (Figs. 6-7). These Glorieta strata are sandstone except for one thin interval of limestone about 55 m above the base of the Glorieta Sandstone.

The Glorieta Sandstone is overlain by bedded gray limestone of the San Andres Formation. The color of the Glorieta Sandstone is dominantly pale yellow brown to dark yellow brown; subordinately, the sediments are colored red, dark brown and rare pale green. The grain size is dominantly fine-grained sandstone, and subordinately siltstone. The following main lithotypes are recognized: (1) massive, fine-grained sandstone (0.3-1.6 m); (2) horizontally laminated siltstone to fine-grained sandstone, partly indistinctly laminated (0.2-2.6 m); (3) indistinctly wavy laminated fine-grained sandstone (0.5-5.6 m); (4) fine-grained sandstone displaying small-scale cross-bedding (dm-scale) (0.5-2.1 m); and (5) fine-grained sandstone displaying large-scale cross-bedding (> 1m), individual units are 0.7-5.9 m thick. A few beds of red mudstone-siltstone (0.2-1.2 m) and red, fine-grained sandstone, partly crossbedded, intercalated in red mudstone-siltstone (0.6-1.1 m), are part of the Glorieta section. In the upper part of the Glorieta Sandstone, two thin gray limestone beds are intercalated (2 and 20 cm thick). Rarely, burrows are observed in fine-grained sandstone in the upper part of the Glorieta Sandstone.

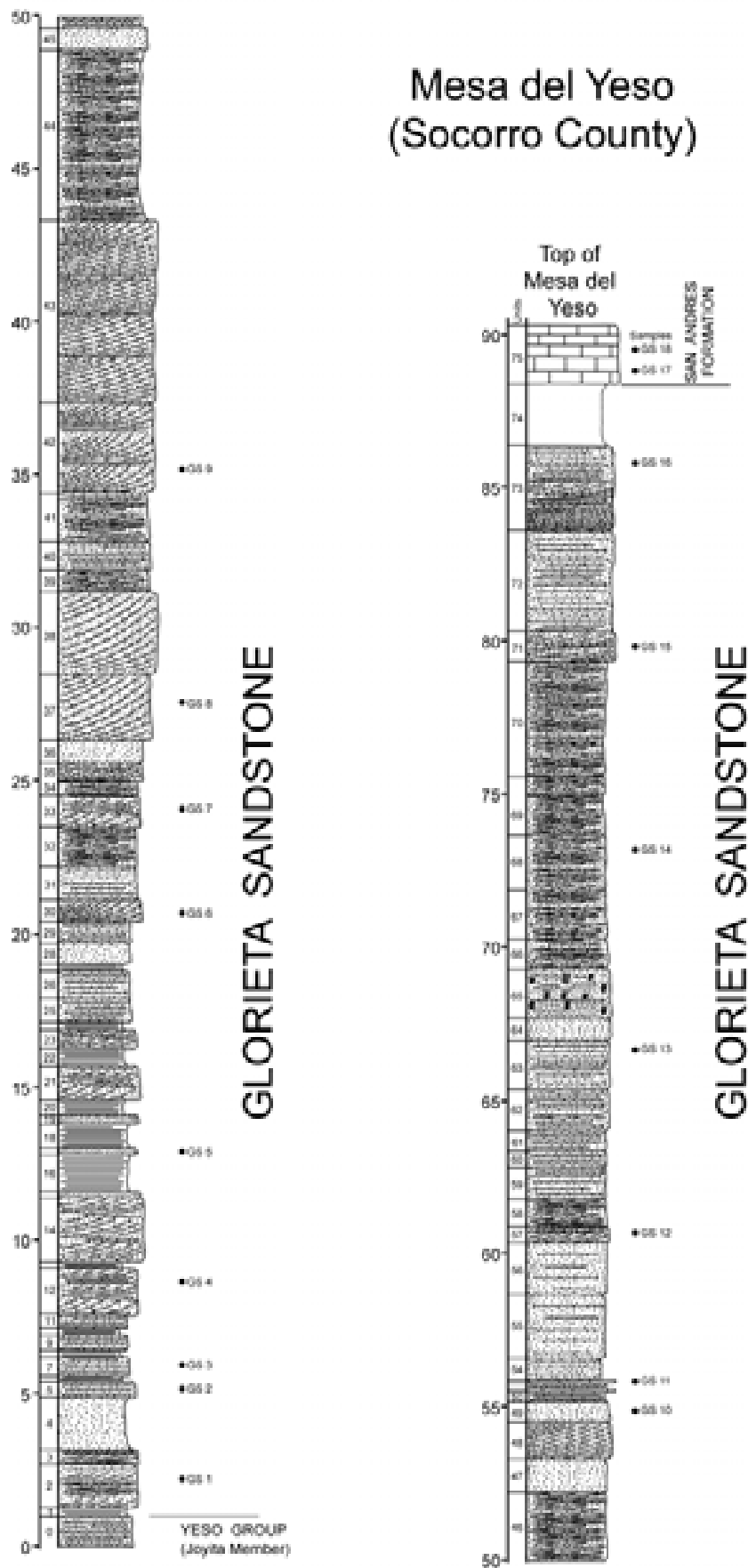


FIGURE 6. Measured stratigraphic section of Glorieta Sandstone at Mesa del Yeso in Socorro County. See Figure 5 for legend to lithologic symbols.

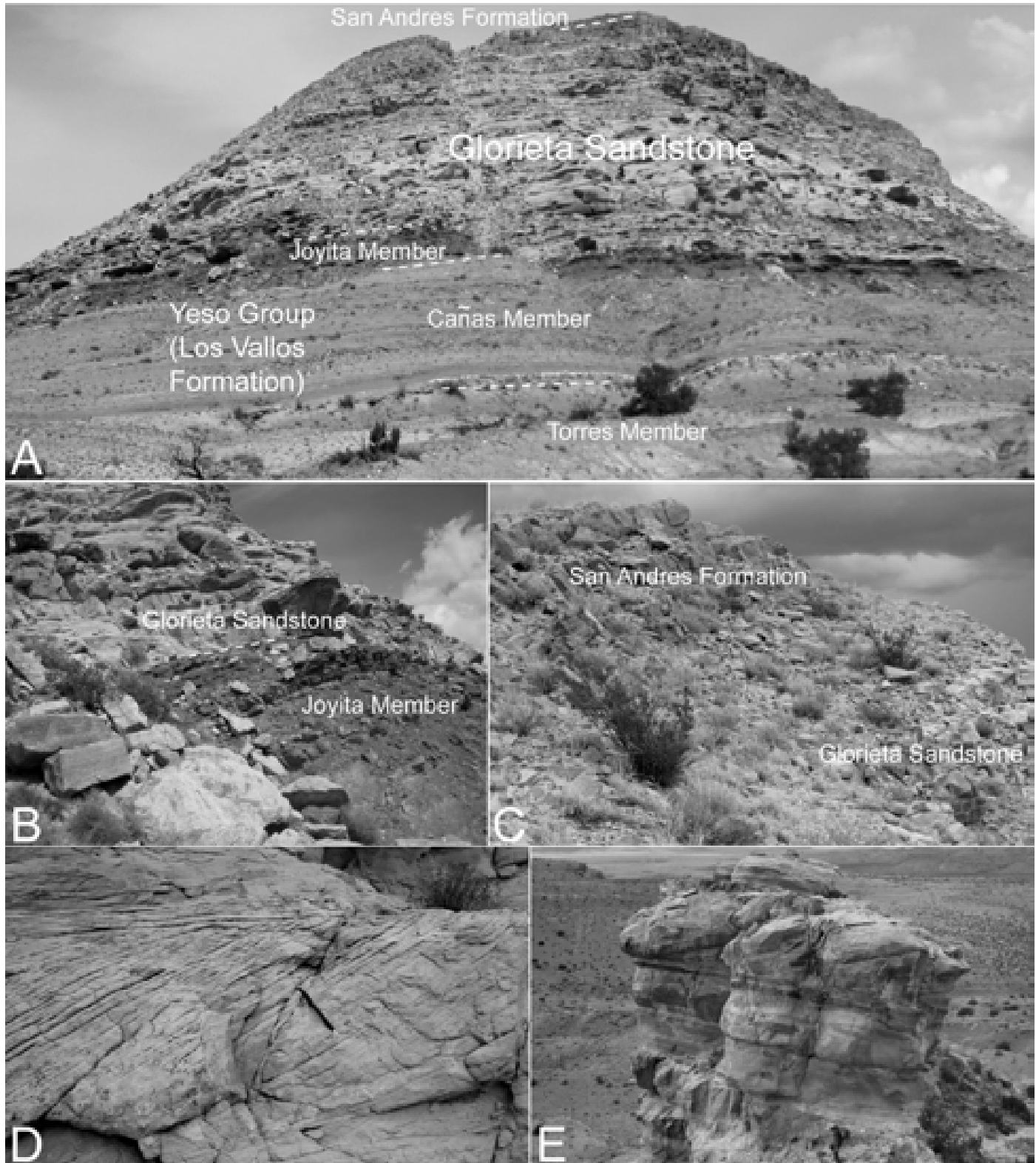


FIGURE 7. Photographs of Glorieta Sandstone outcrops at Mesa del Yeso in Socorro County. **A**, Overview of section, showing Glorieta Sandstone between Yeso Group and San Andres Formation. **B**, Closeup of sharp contact of light-colored quartzose sandstone at base of Glorieta Sandstone on red-bed siltstone and sandstone of the Joyita Member of the Los Vallos Formation (Yeso Group). **C**, Rubbly slope marks contact of light colored quartzose sandstone at top of Glorieta Sandstone and overlying lime mudstones at base of San Andres Formation. **D-E**, Two views of large scale, eolian crossbedding in the Glorieta Sandstone, closeup (pen for scale) (**D**) and outcrop scale (person for scale) (**E**).

PALEONTOLOGY AND AGE

No age-diagnostic fossils have been reported from the Glorieta Sandstone. Some bioturbation and marine foraminifers (seen in thin section) appear to be the only fossil record reported from the formation (e.g., Milner, 1978; Krainer et al., 2012). Therefore, its age relationships must be based on its stratigraphic position and regional stratigraphic relationships. Thus, the Glorieta Sandstone overlies and is overlain by strata of Leonardian age, so it can be confidently assigned a Leonardian age. Given its relationship to the overlying San Andres Formation, a late Leonardian age assignment is justified. As was outlined by Nelson and Hook (2005), the Glorieta has been correlated with the Duncan Sandstone in Oklahoma and the San Angelo Formation in Texas. Both of these units are upper Leonardian on the basis of fossils found within and in bracketing units. The Toroweap Formation, which directly overlies the Coconino Sandstone at the Grand Canyon of northern Arizona, also contains a late Leonardian fauna (Turner, 1990). Thus, regional correlations of the Glorieta Sandstone also support assignment of a late Leonardian age.

DEPOSITIONAL ENVIRONMENTS

The depositional environment of the Glorieta Sandstone has been interpreted as either dominantly shallow marine (Baars, 1961, 1962, 2000; Kelley, 1971; Milner 1978; Dinterman, 2001) or as dominantly eolian (Bauer, 2011; Bauer and Mack, 2011). Baars (1961, 1962, 2000) pointed out that the Glorieta Sandstone is very similar to the Coconino Sandstone of the Colorado Plateau, although sedimentary structures of the Glorieta Sandstone indicate that most of it was deposited in a subaqueous environment. He argued that very large-scale cross-stratification, which is typical of the eolian Coconino Sandstone, is almost absent in the Glorieta Sandstone, and that the primary dip of the cross-strata is considerably lower than in the Coconino, indicating deposition in a high-energy subaqueous environment. According to Baars, most of the Glorieta Sandstone was deposited in a beach to upper neritic marine (shoreface) setting; large-scale cross-stratification is interpreted to represent offshore sand bars. Locally, eolian conditions were present (Baars, 1962, 2000). Baars (2000) suggested that the sediment of the Glorieta Sandstone was derived from the Mazatzal uplift in north-central Arizona.

Milner (1978), in a detailed study of the Glorieta Sandstone in Lincoln County, eastern New Mexico, pointed out that cross-stratification, ripples and parallel stratification are the common sedimentary structures. He noted that the ripples lack internal crossbedding and that tabular and trough crossbedding are common and occur as low-angle cross-sets, which is indicative of beach to upper shoreface deposits. He noted that parallel stratification is rare and that massive sandstone beds are locally common. Rarely, bimodal cross-stratification (herringbone crossbedding) is present. Paleocurrent analysis demonstrated that the predominant paleocurrent direction was towards the south-southwest. Milner (1978) concluded that in eastern New Mexico most of the Glorieta Sandstone was deposited along N-NE to S-SW trending coastlines that were dominated by eastward and southward prograding barrier-island complexes during sea-level lowstands. Sediment was transported by longshore drift along the coastline, and the common cross-stratification formed in foreshore to upper shoreface, coastal dune and tidal channel environments. Sea-level rise caused deposition of carbonate sediments in tidal flat, restricted, open marine, and evaporitic environments.

Recently, Bauer (2011) and Bauer and Mack (2011) interpreted sandstone with ripple lamination, large-scale planar crossbedded sandstone and massive sandstone of the Glorieta Sandstone to be of eolian origin, and small-scale trough-crossbedded sandstone as deposits of shallow interdunal streams. Based on zircon ages, Bauer (2011) and Bauer and Mack (2011) conclude that the sediment of the Glorieta Sandstone was derived from the Appalachians and interior craton to the east and transported westward by winds.

We lack the data to resolve these evident disagreements about the overall depositional system represented by the Glorieta Sandstone. Instead, we can offer some observations based on our local, preliminary work. At Mesa del Yeso, the Glorieta Sandstone is composed of stacked sets of medium- to large-scale crossbedded sandstone, separated by bounding surfaces that are almost horizontally or inclined at low angle (Figs. 6-7). Large-scale, tabular-planar and wedge-planar crossbedding is separated by horizontal to low-angle bounding surfaces. In the lower part of the formation, tabular-planar crossbedded sandstone units bounded by planar-bounding surfaces are present. Crossbedded intervals are separated by horizontally and ripple-laminated sandstone intervals.

We interpret the crossbedded sandstone intervals as eolian dune deposits. Foresets dip mostly towards the southwest, indicating that the sand dunes formed in a relatively unimodal wind regime and that the crossbedded sandstones may represent transverse or barchanoid dunes (cf. Ahlbrandt and Fryberger, 1982; Fryberger, 1990; Brookfield and Silvestro, 2010).

We interpret the horizontally and ripple-laminated sandstones composed of ripple and grainfall strata as interdune deposits. Ripples in the Glorieta Sandstone are poorly preserved, and we did not observe ripple foresets (as did Milner, 1978). During the migration of wind ripples no avalanching occurs down the lee slope, as is the case in subaqueous ripples, so ripple foresets are absent in wind ripples (e.g., Schenk, 1990).

The red mudstone-siltstone with intercalated trough-crossbedded sandstone in the lower part of the formation represents fine-grained overbank deposits with intercalated small fluvial channels. The thin intercalated limestone in the upper part indicates a short marine transgression. Such short periods of marine transgression are observed in other sections such as at Deer Canyon Preserve along Chupadera Mesa (Fig. 5).

We did not recognize clear evidence of marine sandstone deposits within the Glorieta Sandstone at Mesa del Yeso, although in the San Andres Mountains near Rhodes Pass, at the type section of the San Andres Formation, the Glorieta Sandstone is of shallow marine origin as is indicated by the presence of foraminiferans (Krainer et al., 2012). Also, the herringbone-crossbedding observed by Milner (1978) indicates that locally the Glorieta Sandstone was deposited in a shallow marine, tidally influenced environment.

The Glorieta Sandstone is an eastward extension of the Coconino erg of northern Arizona (e.g., Kues and Giles, 2004). We believe that it is primarily of eolian origin, especially in its northwestern and western outcrops on the Colorado Plateau portion of New Mexico. However, to the southeast and east, deposition of parts of the Glorieta Sandstone took place in marginal marine settings. Part of the evident disagreement among the sedimentological studies may stem from the fact that some eolian sands of the Glorieta depositional system were reworked and redeposited in shallow marine settings in southern and eastern New Mexico. Thus, what workers like Kelley (1971) saw as a marginal marine Glorieta Sandstone shoreline facies interfingering with shallow marine carbonates of the transgressing San Andres sea, are actually eolian sands that were deposited in everything from coastal dunes to offshore bars and shoals early in the San Andres transgression.

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At the Lower Permian Yeso Group type section, northeast of Socorro, the upper part of the section consists of a thick, gray and orange interval of gypsum (the Cañas Member of the Los Vallos Formation) overlain by the red-bed clastics of the Joyita Member of the Los Vallos Formation. The Joyita Member is the uppermost unit of the Yeso Group, and is overlain here by the Lower Permian Glorieta Sandstone (yellow-brown sandstones that make up the shoulder of the hill). The mesa is capped by dark brown/gray limestones of the Lower Permian San Andres Formation.

THE PERMIAN SAN ANDRES FORMATION IN CENTRAL AND WESTERN NEW MEXICO

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Abstract—The San Andres Formation is a Permian lithostratigraphic unit primarily composed of limestone that is exposed discontinuously across approximately two-thirds of New Mexico, with sparse outcrops in the central and western parts of the state in Bernalillo, Valencia, Socorro, Cibola and McKinley Counties. The formation thickens southward from a minimum of 1-10 m in northern New Mexico to as much as 181 m in south-central New Mexico. Most carbonate beds of the San Andres Formation have muddy textures and consist of either lime mudstone, muddy wackestone or dolostone; some gypsum and sandstone beds are also present, particularly in the lower part of the formation. The San Andres Formation unconformably overlies the Glorieta and Coconino Sandstones at most outcrops but locally overlies the older Yeso Group. Thus, the base of the San Andres Formation is an unconformity of regional extent that can be traced across much of New Mexico and into Texas as far to the east as the Midland Basin, and perhaps into Oklahoma where the homotaxial equivalent of the San Andres Formation is the Blaine Formation. This unconformity marks the beginning of the great Guadalupian (Middle Permian) cycle of deposition, which actually began during the latest Leonardian (Early Permian) with the onset of San Andres deposition. The upper contact of the San Andres Formation is also an unconformity overlain by younger Permian (Kaibab and Artesia formations) or Triassic (Moenkopi Formation) strata. Locally, carbonate beds of the San Andres Formation yield a diverse marine microfauna (mostly algae and foraminifers) and macrofauna (brachiopod dominated, but including gastropods, bivalves and some cephalopods, among others). The age of the San Andres Formation has mostly been determined from fusulinid and conodont data obtained in southeastern New Mexico and West Texas that indicate that it is of late Leonardian and early Guadalupian (primarily Roadian) age. In central and west-central New Mexico, however, all data suggest that only the Leonardian part of the formation is preserved.

Deposition of the San Andres Formation in central and western New Mexico occurred on the Northwestern Shelf platform of the adjacent Permian Basin and can be characterized by several distinct depositional stages. The first is an initial transgression from east to west over an extensive and possibly uninterrupted eolian dune field. This was followed by recurring episodes of deposition of reworked sands and, near the shelf margins, submarine channelization with accompanying debris deposits from both submarine flows and adjacent terrestrial sources. Following this stage, the San Andres Formation was in a relatively quiescent tectonic setting with the development of a diverse marine shelf environment combined with local shallow areas near the shelf perimeter. Eventually, the San Andres sea regressed eastward, with the westerly transgression of the Kaibab Formation overlapping the regressive top of the San Andres Formation, possibly as far east as western Texas.

INTRODUCTION

The San Andres Formation is a late Early Permian (Leonardian) formation primarily composed of limestone that is exposed discontinuously across approximately two-thirds of New Mexico (Kottlowski, 1969) and into Arizona, with sparse outcrops in the central and western parts of the state, in Bernalillo, Valencia, Socorro, Cibola, and McKinley Counties (Fig. 1). The San Andres Formation in this part of New Mexico was deposited on both the central and distal margins of the Northwestern Shelf platform, an extension of the Delaware Basin of southeastern New Mexico and western Texas (Fig. 2). The shelf platform was a unique depositional environment influenced both by base level changes in worldwide sea levels and, perhaps more importantly, by terrestrial processes that occur along the perimeter of the platform. Although the primary focus of this article is central and western New Mexico and adjacent Arizona, the full extent of the San Andres Formation is considered.

Many authors have studied the San Andres Formation as a groundwater resource for the early expansion of inter-continental railroad construction and the associated development that rail brought to an other-

wise undeveloped countryside, but also because the San Andres Formation forms one of the more prolific hydrocarbon reservoirs of the Permian Basin in New Mexico (Dutton, 2004). The studies of the San Andres Formation have resulted in many questions and much speculation regarding the presence of apparently anomalous stratigraphic units, the configuration of the platform margin(s) and the presence of fossil invertebrate species not typically encountered in the adjacent Delaware Basin environment. Equally frustrating in the search for answers to the many questions is the lack of outcrops surrounding the southwestern, western and northern margins of the San Andres Formation platform environment and its boundary with the adjacent Kaibab Formation of the Colorado Plateau.

Additional research on the shelf margins reveals a close interaction of the San Andres Formation platform environment with the adjacent Kaibab Formation throughout the depositional history of both formations (Brose, 2011). By recognizing the interaction of these two distinctly different depositional environments, a better understanding of the Northwestern Platform depositional history is provided, and many of the various stratigraphic and paleontological questions can be solved.

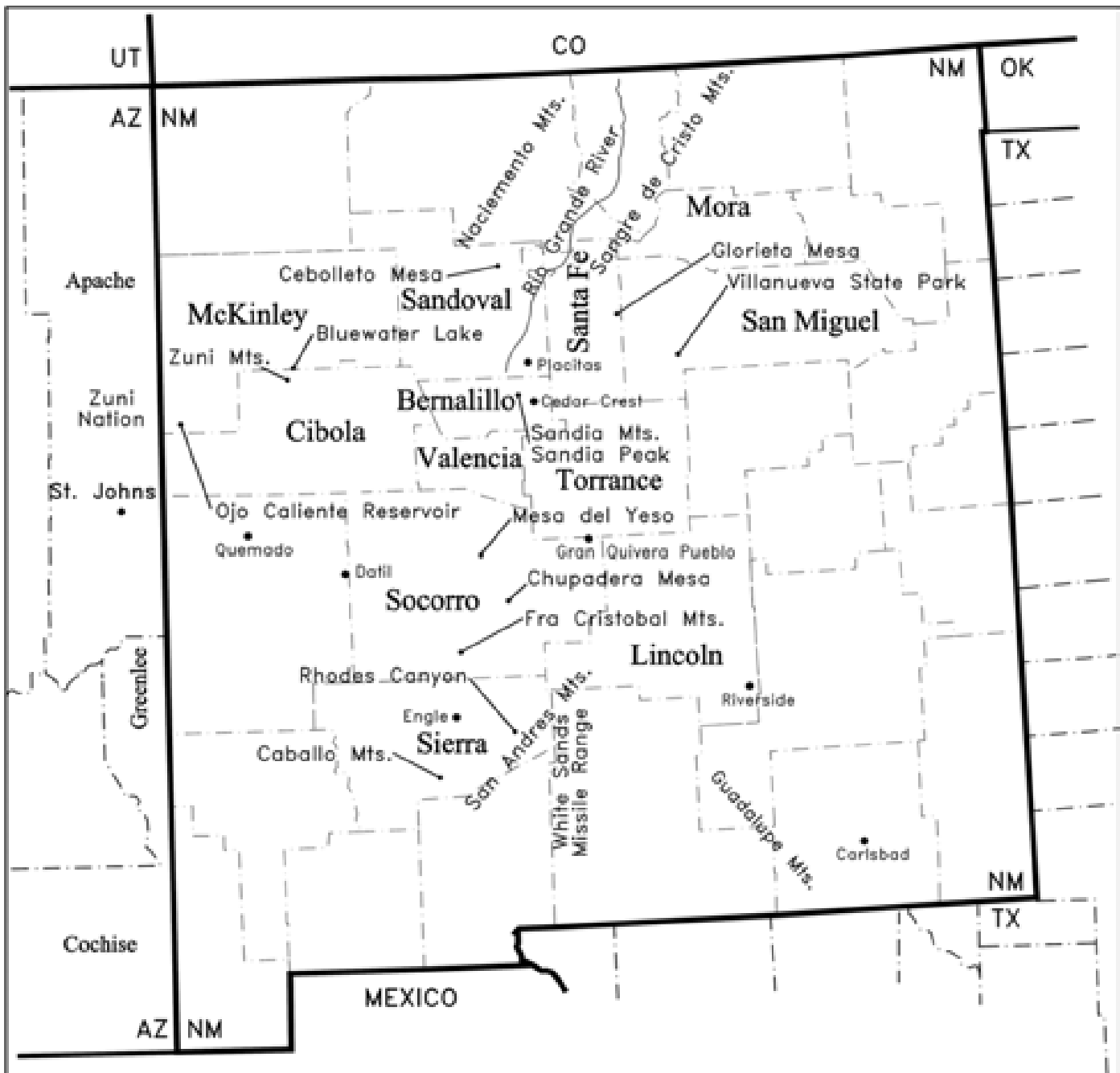


FIGURE 1. Map of New Mexico showing key locations of San Andres Formation outcrops mentioned in the text.

EARLY UTILIZATION OF THE SAN ANDRES FORMATION

The karstic limestones of the San Andres Formation provide a valuable source of groundwater throughout New Mexico. The earliest use of springs as a resource was likely associated with Paleo-Indian sites dating back as early as 10,000 B.C. (Williams, 1986). In McKinley County in western New Mexico, both Peach Springs and Ramah show evidence of Eden and Folsom (9000 to 8000 B.C.) occupation at springs emanating from the San Andres Formation. As climatological conditions dried during the Archaic period (approximately 6000 B.C. to the birth of Christ), the springs emanating from the San Andres Formation certainly became more important to the indigenous peoples, especially those groups using the higher (2500 m) mountains. These sites may have become less

important through the Classic Chaco period (about A.D. 950 through approximately A.D. 1100), when upstream/uphill migration terminated (Williams, 1986), but again would have gained importance during the upland period (post-A.D. 1100) when migration again occurred into the highlands to escape the drought conditions occurring in the basin lowlands.

The first documented use of the San Andres Formation for mined resources is likely associated with the Spanish Missionary arrival in the early A.D. 1600's. Although earlier visited by Juan de Oñate in the period of 1598-1604, Fray Santander organized the Pueblo Indians inhabiting the "Pueblo de Las Humanas," located in central New Mexico near the Rio Grande, to construct a large missionary complex (Hackett, 1937). It was here that evaporite deposits (primarily gypsum) of the San

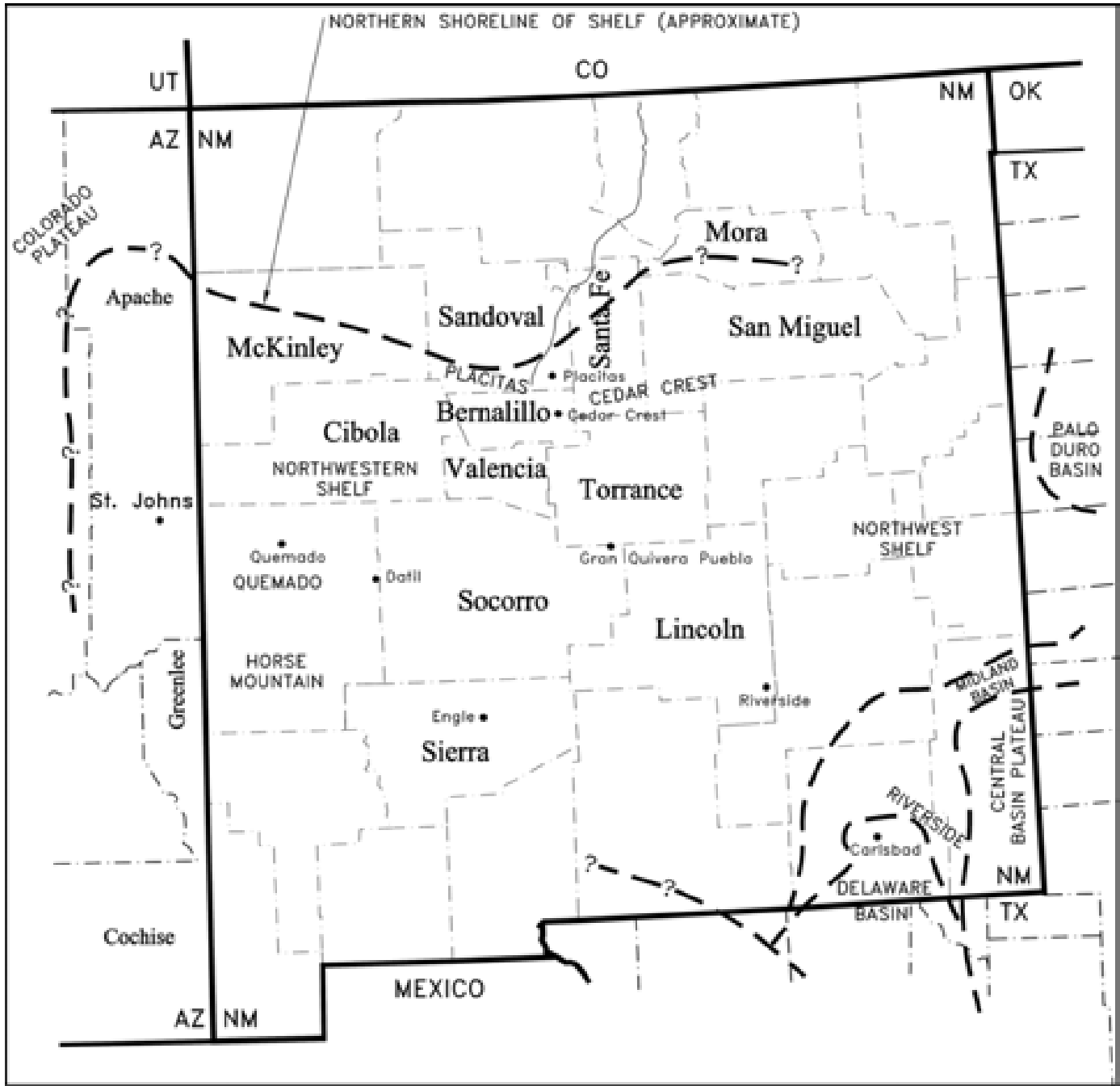


FIGURE 2. Location of Permian basin and Northwestern Shelf platform, in which the San Andres Formation was deposited.

Andres Formation, formed during the last regression of the sea across the shelf platform, were mined and used as windows for the structures. The Pueblo de Las Humans did not receive its current name, Gran Quivira Pueblo in Valencia County, until it was abandoned in the late 1600's.

PREVIOUS STUDIES

The presence of Permian-age rocks in the southeastern portion of New Mexico was first identified by George G. and Benjamin F. Shumard (1858) following a study of outcrops in the Guadalupe Mountains located southwest of Carlsbad, New Mexico, as part of Captain John Pope's railroad survey party. As with most railroad investigations of the mid-1800's, the purpose of the survey was to find adequate groundwater

supplies to support the development of rail lines. Although others (Jenney, 1874; Tarr, 1892) subsequently mapped the Guadalupe Mountains, it wasn't until 1901 when Girty (1902, 1908) collected the faunal assemblages that established the Permian age of the rocks of the southern Guadalupe Mountains.

The first use of the terminology "San Andreas" limestone was by Lee (1909), who took the name from the San Andres Mountains, misspelling it. Although the spelling was modernized by Darton (1928), he did not specifically isolate the San Andres limestone and instead combined the underlying Yeso Formation (including the Glorieta Sandstone) and the San Andres limestone into the "Chupadera Formation." The type section of the San Andres Formation was specified by Needham

and Bates (1943) in the San Andres Mountains of south-central New Mexico. The type section was further studied by Kottlowski et al. (1956) and Lindsay (1994), with a more recent revision to the stratigraphy using detailed analyses of thin sections completed by Krainer et al. (2012).

The identification and presence of the San Andres Limestone became an important discovery for subsequent oil field development in the 1920's. The San Andres Limestone was early recognized as a major reservoir for hydrocarbons within the Permian Basin located in western Texas and southeastern and central New Mexico and forms the most prolific play of the Northwestern Shelf environment. As currently understood the Permian Basin comprises an area of approximately 194,250 km², with the San Andres Formation constituting the largest play in the Paleozoic stratigraphy (Dutton, 2004; Krainer et al., 2012). The Permian Basin is an asymmetrical sink in which the deeper Delaware Basin to the west is separated from the shallower Midland Basin by the Central Basin Platform (Nassir, 2006). The Delaware Basin is in turn bounded to the west by the Northwestern Shelf. The westernmost boundary of the Northwestern Shelf has been poorly understood because of cover by younger sedimentary and volcanic rocks and scarcity of subsurface data (Peirce, 1989), particularly as it appears to extend into east-central Arizona (Brose, 2011).

HISTORY OF STRATIGRAPHIC NOMENCLATURE

As noted above, the San Andres Formation was first named by Lee (1909) for exposures located in the San Andres Mountains of southeastern New Mexico. Lee (1909, p. 12) named the "San Andreas limestone" as the uppermost unit of the Manzano Group of Herrick (1900). He took the name from "San Andreas Mountain, at the north end of which it is typically developed, as described in the section on page 29" (Lee's, 1909, fig. 7; see our Fig. 2). Lee (1909, p. 12) described the "San Andreas limestone" as "essentially marine limestone, which is often cherty and poorly fossiliferous, although several localities were found where fossils are abundant." He recognized its distribution across much of central New Mexico, from Mesa del Yeso in Socorro County through the Fra Cristobal, Caballo and San Andres Mountains of Sierra County.

The original misspelling of San Andres as "San Andreas" by Lee (1909) was corrected by Darton (1928). However, Darton (1928) was unable to separate the Yeso and San Andres formations regionally, so he combined them in his Chupadera Formation, a stratigraphic concept abandoned in the 1940s (Lucas, 2009).

Needham and Bates (1943), in one of the classic articles of New Mexico geology, properly regarded Lee's (1909, his fig. 7) section as the type section of the San Andres Formation. Lee (1909, p. 29) gave the location of this section in the northern San Andres Mountains as "in the canyon through which the road passes from Engle to Rhodes's ranch." Also known as Rhodes Pass or Rhodes Canyon, this site is presently within the White Sands Missile Range. Needham and Bates (1943, p. 1665) fixed the type section there in a small canyon south of the road, providing a specific location in sec. 29, T12S, R 2E (their Fig. 1). Needham and Bates (1943, p. 1665) described the type section of the San Andres Formation as 181 m of limestone with two thin (0.6 m thick) sandstone beds near its base, overlying the Glorieta Sandstone (their Fig. 2). Significantly, the top of the San Andres Formation is eroded at the type section, so it is not a complete section of the formation. Needham and Bates (1943) also noted the presence of numerous brachiopods (*Dictyoelostus*, etc.) in the type section. They were aware of a much broader geographic extent of the San Andres Formation than was Lee (1909), probably due mostly to the work of Darton (1928).

Kottlowski et al. (1956, p. 60-62, 89-90, fig. 11) described in detail a section of the San Andres Formation located ~ 1.6 km west of the type section. Here (center E1/2 sec. 30, T12S, R2E), the San Andres Formation was reported to be 174 m thick, also has an eroded top and is much better exposed than at the type section. Lindsay (1994, p. 133)

therefore referred to it as "a much better exposed reference section," and redescribed the reference section and interpreted it as a transgressive systems tract overlain by a highstand systems tract divisible into 47 shallowing upward parasequences and 13 parasequence sets. Again, this section is incomplete, in that the top is eroded.

Krainer et al. (2012) redescribed this reference section as ~ 130 m thick, dividing it into a lower, hydrocarbon-bearing bedded-limestone interval, a middle thicker-bedded and massive limestone interval and an upper interval of bedded limestone with some chert (Fig. 3). Petrographic study revealed that the most common microfacies of the San Andres Formation at the reference section is bioclastic wackestone, and that muddy textures dominate the section. Ammonoids indicate that the reference section of the San Andres Formation is of Leonardian age, although, again, the top of the formation has been eroded. The detailed lithostratigraphy and petrography of Krainer et al. (2012) did not support Lindsay's (1994) interpretation of the reference section as a transgressive systems tract overlain by a highstand systems tract, both composed of numerous parasequences. Instead, Krainer et al. (2012) suggested that San Andres Formation limestone was deposited in a normal marine environment of dominantly low to moderate water turbulence within the photic zone with open circulation. Well washed grainstones formed under high-energy conditions are rare, and only a few shallowing-upward parasequences are developed. The reference section of the San Andres Formation is an incomplete section of the lower, Bonney Canyon Member of the formation.

In southeastern New Mexico, Kelley (1971) used the occurrence of allochthonous detritus to separate the San Andres Formation into three members (in ascending order), the Rio Bonito Member, the Bonney Canyon Member and the Fourmile Draw Member (Fig. 4). The Rio Bonito Member consists of massive, thick-bedded limestone, and the entire reference section of the formation (Krainer et al., 2012) can be referred to that member. The overlying Bonney Canyon Member is the middle, typically thin-bedded dolomitic and limestone interval of the San Andres Formation, and the Fourmile Draw Member consists of the uppermost, evaporitic portion of the section.

Numerous early publications (Lloyd, 1929; Crandall, 1929; Blanchard and Davis, 1929; Cartwright, 1930; Fiedler and Nye, 1933; Lang, 1937) developed facies relationships and suggested new formal stratigraphic names for the San Andres Formation. However, it was not until the work of King (1944) that detailed mapping enabled a stratigraphic correlation of the rocks in the Delaware Basin to the Guadalupe Mountains reef zone of the Northwestern Shelf. Within the Northwestern Shelf environment, deposition of the San Andres Formation reportedly alternated between normal marine shelf carbonates and eolian environments. Although composed primarily of limestone, dolomite, and sandstone, thick sections of anhydrite deposits are common in this shelf environment.

Subsequent studies have been completed throughout the south-central, central, western and northern portions of the state to identify the San Andres Formation, determine the paleodepositional environment of the shelf margins, and resolve disparities between formation descriptions provided by different authors (e. g., Needham and Bates, 1943; Kottlowski, 1956; Smith, 1954; Read and Wanek, 1961; Brady, 1962; Harbour, 1970; Rascoe and Baars, 1972; Sarg, 1986; Ward, 1986; Colpitts, 1989; Peirce, 1989; White, 1989; Lindsay, 1994; Pranter, 2004). The San Andres Formation crops out extensively in the southeastern portion of New Mexico, but its outcrop belt becomes disrupted westward crossing the Rio Grande rift system. To the west of the Rio Grande rift, the San Andres Formation is encountered irregularly on outcrop and is primarily isolated along subsequently (post-Permian) uplifted sections of strata in regional mountain highlands throughout central and western New Mexico.

In general, the published information has documented the depositional environment(s) associated with the Northwestern Shelf San Andres Formation. Although the interpretations have provided a macroscopic

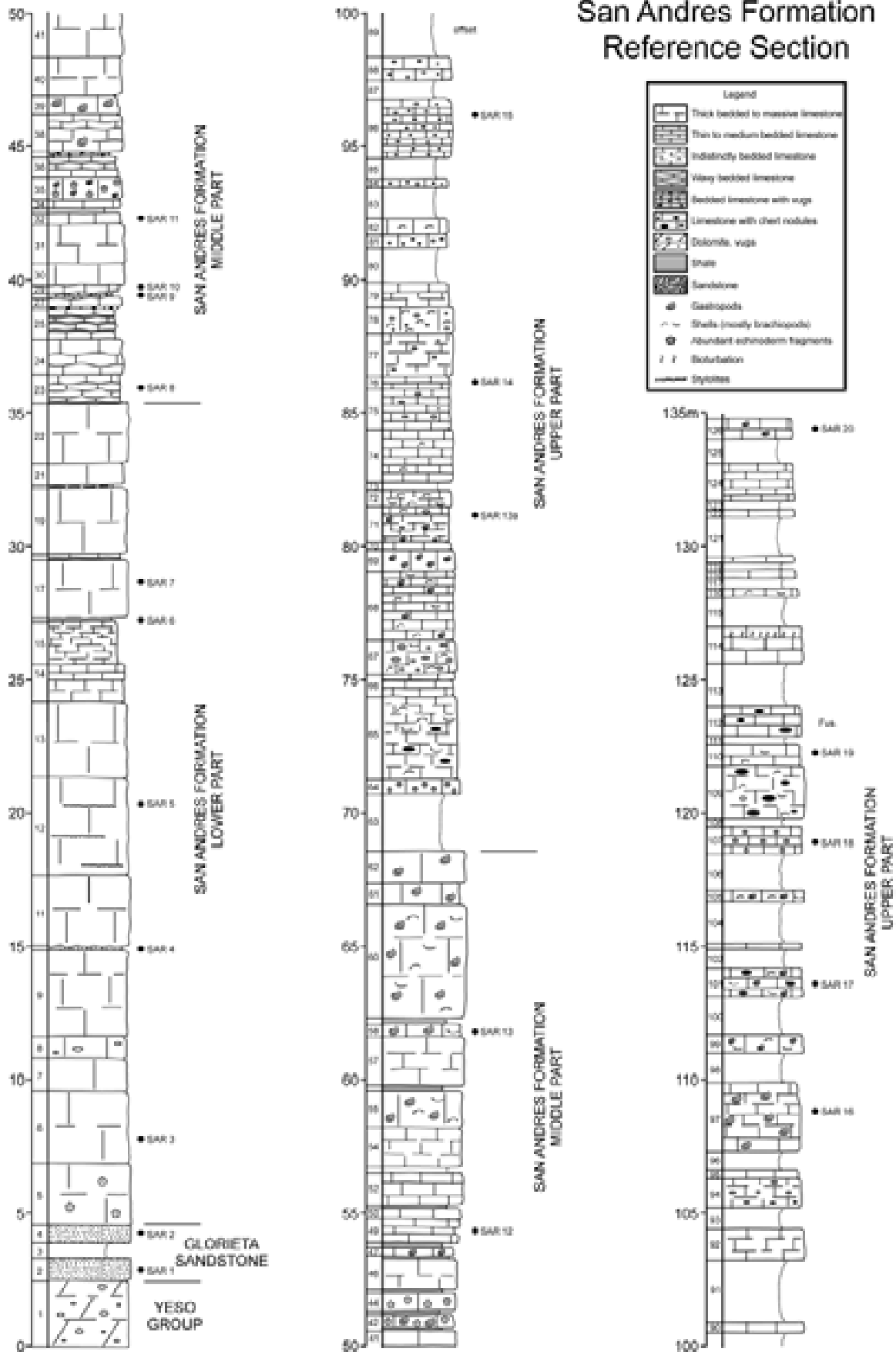


FIGURE 3. Reference section of the San Andres Formation (from Krainer et al., 2012).

understanding of the shelf environment, recent work (Kues and Lucas, 1989; Lucas, 1991; Brezinski, 1991; Brose, 2011) has focused more closely on the margins of the shelf environment. These evaluations have assisted in interpreting several key regionally anomalous deposits in both the lower San Andres Formation and in overlying younger Permian depositional environments.

In particular, most researchers confirm the presence of a sand interval in the lower portion of the San Andres Formation. This interval has been termed the "Hondo Sandstone Member" of the San Andres Formation (Lang, 1937; Harbour, 1970) and has been used by most authors as a marker bed to separate the San Andres into lower and upper intervals. Based on thin section analyses, discussed below, it is likely that the sandstone intervals represent reworked sandstone of the Glorieta Formation and, farther west, Coconino Sandstone of the Colorado Plateau, both of which were redeposited in the marine San Andres sea. This interpretation is consistent with regional observations that suggest numerous high-energy events likely transported sands from the shelf margins into the adjacent marine environment, and helps explain the presence of marine fauna in some of the sands (Brose, 2011; Krainer et al., 2012). It also suggests that a separate formational nomenclature for the Hondo sands may be appropriate in the vicinity of the type section, but it is more likely that the reworked deposits in the lower section(s) of the San Andres Formation are not of significant thickness to be interpreted as a separate member of the formation. Some other workers (e.g., Kelley, 1971; Milner, 1978) have included the Glorieta Sandstone in the San Andres Formation, although this trend has not been followed in the majority of New Mexico.

Similarly, the stratigraphy of the upper contact of the San Andres Formation has been interpreted differently by many authors, and a clear resolution of the depositional history is only recently emerging (Brose, 2011). The stratigraphic unit capping the San Andres Formation in much of north-central New Mexico is the Artesia Formation, formerly termed the Bernal Formation by Charles B. Read in 1939, using the acronym "Pb" in a geologic map of north-central New Mexico (Read and Andrews, 1944; Read et al., 1944) and later formally published as a lithostratigraphic name on a map legend by Bachman (1953). The Bernal was described as a fine-grained clastic unit of mostly orange siltstone and massive sandstone (Lucas and Hayden, 1991). This terminology was used by Wood and Northrop (1946) for exposures in the Nacimiento Mountains, by Bachman (1953) in the Ocate area of northwestern Mora County, and by Baltz et al. (1956) in the Sangre de Cristo Mountains.

According to Baars (1962), the Bernal Formation was an accepted name for the youngest Permian strata in northern New Mexico, and was being applied to similar strata in south-central New Mexico (e.g., Smith and Budding, 1959; Wilpolt and Wanek, 1961). However, Tait et al. (1962) recommended the abandonment of the term Bernal Formation in favor of inclusion as the Grayburg and Queen formations of the Artesia Group (also see Kelley, 1972). Lucas and Hayden (1991) described a type section of the Bernal Formation and recommended rejection of the term, replacing it with Artesia Formation (undifferentiated). Correlation of Bernal to Artesia indicates that it represents an early Guadalupian cycle of deposition younger than the Leonardian-Guadalupian cycle during which the San Andres Formation was deposited.

Recent publications (Kues and Lucas, 1989; Brezinski, 1991) suggest that the upper portion of the San Andres Formation contains a macroinvertebrate fauna characteristic of the adjacent Kaibab Formation of the Colorado Plateau. Indeed, most published accounts of Bernal Formation outcrops are strikingly similar to those of the Kaibab Formation, and it is postulated that the Kaibab Formation encompasses the youngest Permian strata in west-central New Mexico, and that appears to be the case in southeastern New Mexico throughout the Guadalupe Mountains eastward into western Texas (Brose, 2011). The Kaibab Formation of the Colorado Plateau is immediately overlain by the Triassic Moenkopi Formation, which also overlies the Bernal Formation.

LITHOSTRATIGRAPHY

Various researchers have divided the San Andres Formation into lithostratigraphic units. Local lithostratigraphic units defined by different authors may be useful for correlation in adjacent or nearby San Andres Formation outcrops, but regional variations in depositional environments do not enable correlation of each member throughout New Mexico.

Contacts

The majority of the Lower Permian section in central New Mexico is assigned to the Abo Formation and overlying Yeso Group (Lucas et al., 2005). Stratigraphically, the Glorieta Sandstone directly overlies the Yeso Group. The Yeso Group in central New Mexico is primarily a two-part unit. The lower part is sandstone-dominated and is assigned to the DeChelly Sandstone in parts of northern New Mexico, and the upper part is a mixture of siliclastics (mostly siltstone), evaporites (gypsum) and at many locations thin carbonate beds, most of which are dolostones. These strata are typically overlain at the very top of the Yeso by red-bed siliclastics that are part of the Joyita Member of the Los Vallos Formation. The Glorieta Sandstone, along with the adjacent Coconino Sandstone of the Colorado Plateau, formed an extensive eolian dune system resting atop the Yeso Group across what is now northern and central New Mexico. Both the Glorieta Sandstone and adjacent Coconino Sandstone are quartzarenites composed primarily of dominantly monocrystalline quartz, with cementation inclusive of both calcite and quartz (Bills, 2000; Krainer et al., 2012; Lucas et al., this volume: Glorieta Sandstone).

The San Andres Formation rests unconformably above the Glorieta and Coconino sandstones, where present. Both the Glorieta and Coconino sandstones may also be locally absent, with the San Andres Formation unconformably resting on the older Yeso Group, as is the case in much of the Caballo and San Andres Mountains. Both Kelley (1971) and Milner (1978) recognized the possibility that the sands encountered in the lower portion(s) of the San Andres intertongue with San Andres Formation deposits, suggesting a rapidly changing sea level that enabled eolian deposits to cover previously deposited marine strata (Fig. 4). As observed on outcrop throughout New Mexico, it appears more likely that the sands encountered in the lower portions of the San Andres Formation are reworked Glorieta and Coconino sands that have been redeposited in the marine environment. This is substantiated by the presence of marine fauna in the sands, wave sets in sands with little or no scouring of the underlying sea floor, and variable thicknesses of the sands predicated upon the proximity of the adjacent shoreline.

The base of the San Andres Formation is an unconformity of regional extent that can be traced across much of New Mexico and into Texas as far to the east as the Midland basin, and perhaps into central Texas east of the Midland basin and Oklahoma where the homotaxial equivalent of the San Andres Formation is the Blaine Formation. This unconformity marks the beginning of the great Guadalupian (Middle Permian) cycle of deposition, which actually began during the latest Leonardian (Early Permian) with the onset of San Andres deposition.

To date, a thorough study of the distribution of reworked eolian sands in the San Andres Formation has not been conducted although Milner (1978) addresses this subject. A significant volume of the dune deposits were deposited offshore within the depositional systems of both the San Andres Formation and the adjacent Kaibab Formation. A detailed examination of the thicknesses, depositional pattern(s) and location(s) of these sands would provide a clearer understanding of the relative highs and lows eventually covered by the transgressing San Andres sea.

Similarly, the upper contact of the San Andres Formation was subject to erosion as the Kaibab Formation depositional system transgressed easterly across the retreating San Andres sea (Brose, 2011). Evaporite deposits are common to both formations, suggesting alternating periods of inundation and desiccation of transitional shelf sequences.

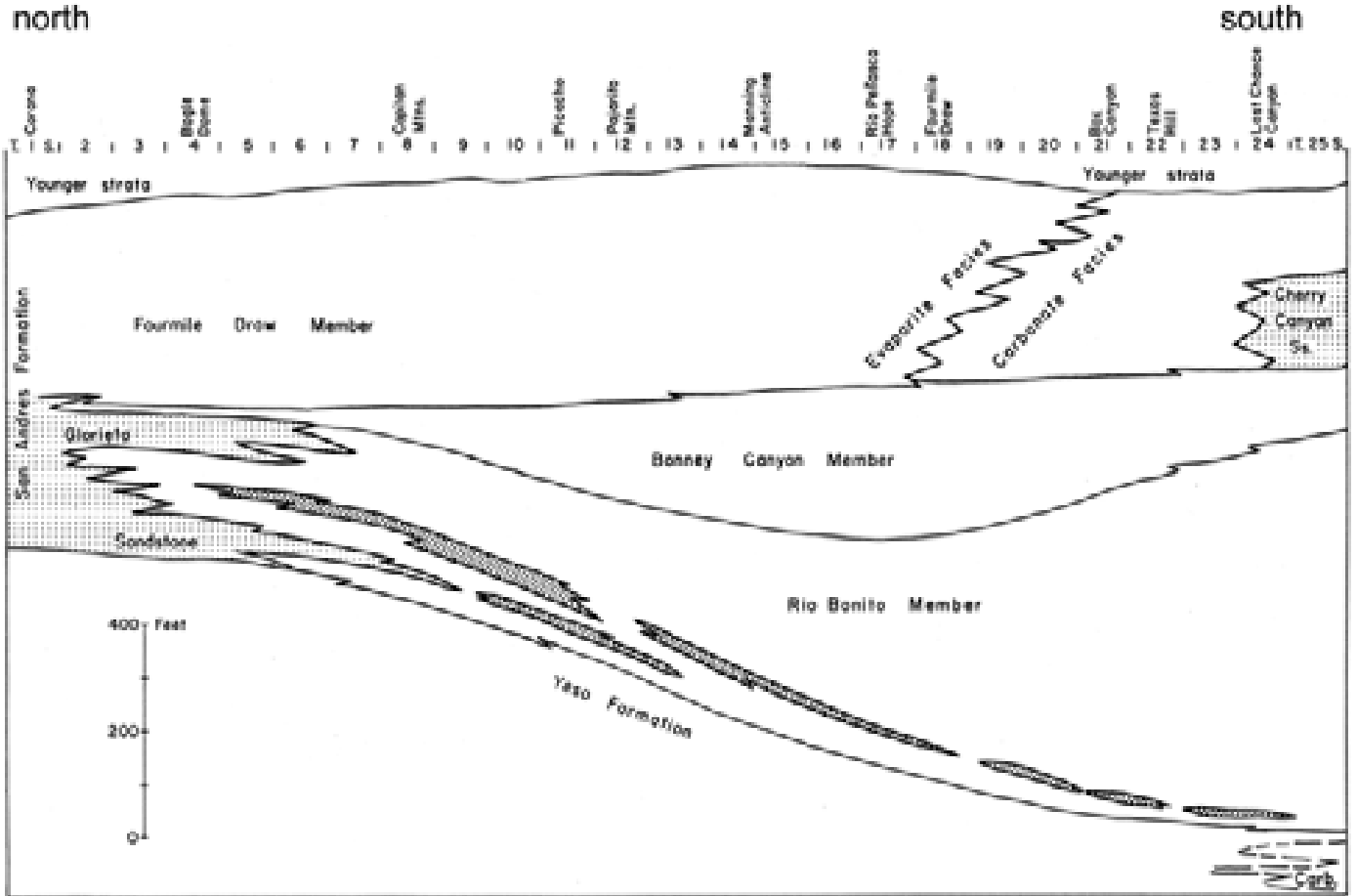


FIGURE 4. North-south stratigraphic diagram of San Andres Formation from Corona in Torrance County, New Mexico, to Last Chance Canyon in Eddy County, New Mexico, showing inferred stratigraphic relationships of members of San Andres Formation (from Kelley, 1971).

This interpretation is further complicated by the presence of overlying strata that include younger Permian deposits (the Kaibab and Artesia formations) and the Triassic Moenkopi Formation. Irrespective of these studies, the presence of Kaibab faunal assemblages as far to the east as the Sacramento Mountains (Brezinski, 1991; Brose, 2011) supports a Kaibab Formation transgression over much of central and southern New Mexico.

The Triassic Moenkopi Formation unconformably overlies the Permian sequence in central New Mexico and on the adjacent Colorado Plateau, and marks the boundary between the Paleozoic and Mesozoic eras. The Moenkopi Formation was deposited upon an eroded Permian landscape of significant topographic relief. Locally, the Moenkopi Formation can be seen infilling karstic sinkholes within the San Andres Formation, with both the San Andres Formation and Moenkopi Formation undergoing both plastic and brittle deformation (Smith, 1954; Lucas et al., 2003). The Moenkopi Formation deposits appear to be more obvious on outcrop along the western and northern portions of the shelf margins, although further field activities are warranted to confirm this observation. The presence of both plastically-deformed San Andres Formation limestones and clays and silts of the Moenkopi Formation suggests that dissolution of the lithified San Andres Formation limestone occurred shortly after deposition, with variable conditions for cementation of the underlying formation occurring both locally and regionally (Brose, 2011).

Thickness

The thickness of the San Andres Formation varies depending on

the depositional location within the shelf, the thickness of reworked sands deposited on the otherwise quiescent seafloor, and subsequent erosion of the upper surface prior to the deposition of the Permian Kaibab and/or Artesia formations and/or Triassic Moenkopi Formation. Because the San Andres Formation commonly caps mesas and plateaus, its upper surface is widely erosional. At the type section in the San Andres Mountains, as described by Needham and Bates (1943), the thickness of the San Andres Formation is 181 m, although the top of the section is eroded. At the reference section, also located in the San Andres Mountains, the San Andres Formation reaches a thickness of 135 meters (Krainer et al., 2012), although the top of this section is also missing. Elsewhere, the San Andres is documented as ranging from 100 to 350 meters in thickness, with typical sections likely averaging closer to 150 meters across the central, north-central and western portions of the state (e.g., Colpitts, 1989; Lucas and Zeigler, 2004; Shomaker, 1971).

A section of the San Andres Formation in the Horse Mountains, south of Datil, New Mexico, consists of 115 m of gray and pinkish-gray micrites and dolomites (Lucas and Kues, 1994). In the Lucero uplift of Valencia County, Lucas and Zeigler (2004) described a section of San Andres Formation that is approximately 109 m thick. This section is approximately half gypsum, and limestone and siltstone make up most of the rest of the section.

The San Andres Formation thins near the western and northern margins of its outcrop area. A measured section near St. Johns, Arizona is less than 5 m thick, and this thickness may represent the combined San Andres Formation and Kaibab Formation. On the northern flank of the Sandia Mountains, the San Andres is only 5 m thick (Lucas et al., 1999),

but it is thicker to the south (Fig. 5). The San Andres Formation is present in the Sangre de Cristo Mountains and on top of the Glorieta Mesa in Santa Fe and San Miguel counties, where the San Andres Formation thins to a maximum of 10 meters (Baltz, et al., 1956) and appears in irregular and isolated outcrops. Farther east, near Villanueva State Park, the San Andres Formation thins to 1-7 m (McLemore, 1996). Foster (1972) indicates that the San Andres Formation and Glorieta Formation continue easterly, where the combined section reaches a maximum thickness of over 385 m.

Lithology

Detailed descriptions of the San Andres Formation are available for the east-central and southeastern portions of New Mexico. The descriptions typically focus on the distal (easterly) portion of the North-western Shelf and into the deeper Delaware Basin. The San Andres Formation here is the primary reservoir for hydrocarbons and forms the major play (Dutton et al., 2004). These analyses generally point to an increasingly evaporitic trend westerly (Pitt et al., 1981).

The lithology of the reference section of the San Andres Formation (Lindsay, 1994) was recently studied by Krainer et al. (2012) (Fig. 6). Although incomplete, approximately 135 m of section are encountered at the reference locality. The lower portion of the section (approximately 30 m) is composed of massive limestone beds that are indistinctly bedded. The limestone is gray to dark gray, muddy, and bituminous. There are several thin (1-10 cm) siltstone intervals in the lower section(s). The middle portion of the section (approximately 33 m) is also gray to dark gray, thin to thickly bedded massive limestone with local thin shale layers. The limestone is muddy and commonly bituminous, and large fossils (such as brachiopods) are rare. The upper part of the reference section (approximately 67 m) is composed of medium- to thickly-bedded limestone and crinoidal wackestone. This part of the section contains abundant fossil assemblages, is devoid of hydrocarbons that appear in the lower two sections, and vugs infilled with calcite are common.

White and Kelly (1989) provide a regional description of the San Andres Formation, primarily referencing the groundwater resources associated with this regional aquifer system, and developed isopachs of the Glorieta Sandstone and thicknesses of the San Andres Formation (together with depositional environments) for portions of Catron, Cibola, Sandoval, Socorro and Torrance Counties. Their description of the lower unit of the San Andres Formation includes thin-to-thick, medium-bedded dolomites, variegated shales and siltstone with sandstone stringers similar to the underlying Glorieta Sandstone. The middle unit is described as a white-to-pinkish, medium to coarse, massive sandstone, with the upper unit consisting of massive, white-to-pink, cliff-forming limestone. Although thicknesses of individual units were not provided, the overall section is reported to be 123 m in the Lucero basin. Once again, the description of the San Andres Formation appears to reflect a combination of San Andres Formation and Kaibab Formation.

Colpitts (1989) supplemented White and Kelly's general description with a Permian reference section of the southeastern Zuni Mountains in Cibola County. Here, the San Andres Formation consists of approximately 45 m of a lower calcareous dolomite, a middle cross-bedded sandstone unit and an upper dolomitic limestone and calcareous dolomite unit. The calcareous dolomites are medium brownish gray to light gray, and are fossiliferous in the upper unit. Colpitts (1989) suggests that the middle crossbedded sandstone likely represents a "tongue" of the Glorieta Sandstone.

An outcrop along the northern portion of the Zuni Mountains, near Bluewater Lake, was documented by Brose (2011). Here, the stratigraphy closely matches that of the San Andres Mountains reference section described by Krainer et al. (2012), with a lower, hydrocarbon-bearing sequence that is disrupted by high energy deposits, including sands deposited in a marine environment, and an upper section that is not hydrocarbon enriched but contains abundant fossils.

Near Placitas in Sandoval County and just north of Sandia Peak, Lucas et al. (1999) briefly described the San Andres Formation as an interval of grayish-orange sandy limestone approximately 5 m thick. Here, the San Andres Formation outcrops are "patchy," being locally absent beneath the unconformity at the base of the Triassic Moenkopi Formation. It is possible that these outcrops include Kaibab Formation sands and silts above the gray San Andres Formation limestone.

To the south in the Sandia uplift, at Cedar Crest, we report here a section of San Andres Formation that has prominent sandstone beds of Glorieta lithology interbedded with very muddy limestones (Fig. 5). Thus, at Cedar Crest, the Glorieta Sandstone overlies the Los Vallos Formation of the Yeso Group as a sandstone unit that separates it from limestone and dolomite of the lower part of the San Andres Formation. Approximately 22 m above the base of the San Andres Formation another sandstone sheet of Glorieta lithology, 18 m thick, is present. Carbonate strata of the San Andres Formation here are resistant beds of muddy limestone and thick-bedded to massive dolomitic limestone.

Farther to the north, along Glorieta Mesa and into the Sangre de Cristo Mountains, the San Andres Formation consists of dark gray, thin- to medium-bedded limestone that has a bituminous odor on fresh breaks (Baltz et al., 1956). Farther to the east, near Villanueva State Park, the San Andres Formation consists of gray to brown, fine-grained limestone with local silty interbeds (McLemore, 1996).

Distribution

The San Andres Formation extends from the northern edge of the reef complex that fringes the Delaware Basin and onto the Central Basin platform in West Texas and southeastern New Mexico westward across what was the vast paleoenvironment of the Northwest Shelf and inclusive of the eastern portion of the adjacent Colorado Plateau (Fig. 7). San Andres Formation deposits are observed on outcrop as far north as the southern portion of the Sangre de Cristo Mountains in Santa Fe and San Miguel counties (Baltz, 1959). White and Kelly (1989) illustrated the northern limits of the San Andres Formation in their figure 2 (p. 332), with the northern shoreline extending in a general east-west direction just south of the northern McKinley and Sandoval county lines.

Although disrupted by the subsequent rifting associated with the Rio Grande, the San Andres Formation is observed on outcrop in the mountains that border the eastern side of Cebollete Mesa, in the Zuni Mountains and near Ojo Caliente Reservoir on the Zuni Reservation. The San Andres Formation has been observed in outcrop near St. Johns, Arizona, which appears to be its westernmost exposure (Wilson et al., 1960).

Stearns (1962) first noted the presence of San Andres Formation in the Horse Mountains south of Datil, New Mexico (Lucas and Kues, 1994). Limited outcrops to the south, from central New Mexico westerly through Arizona, make a precise delineation of the paleo-shoreline difficult to reconstruct. Foster (1964) identifies the presence of the San Andres Formation in exploratory boreholes in Catron County to the south, but little is published regarding the southerly shoreline south of this area.

PALEONTOLOGY AND AGE

At the reference section located in the San Andres Mountains, Krainer et al. (2012) encountered a similar faunal assemblage in the upper part of the San Andres Formation as described by Kottlowski et al. (1956, p. 61) that includes brachiopods and molluscs (including numerous nautiloids) of Leonardian age. Thin section analyses of samples obtained from the reference section include echinoderm fragments, bryozoans, brachiopod shells, bivalves and gastropods. Less common are tubular foraminifers, and a few samples contain rare trilobite fragments.

In Lincoln County, east of the location of the type and reference sections in the San Andres Mountains, Milner (1976) reported a normal marine assemblage of brachiopods, crinoids, bryozoans, echinoids, and cephalopods from the San Andres Formation. Milner also mentions the

Cedar Crest (SMNHC)

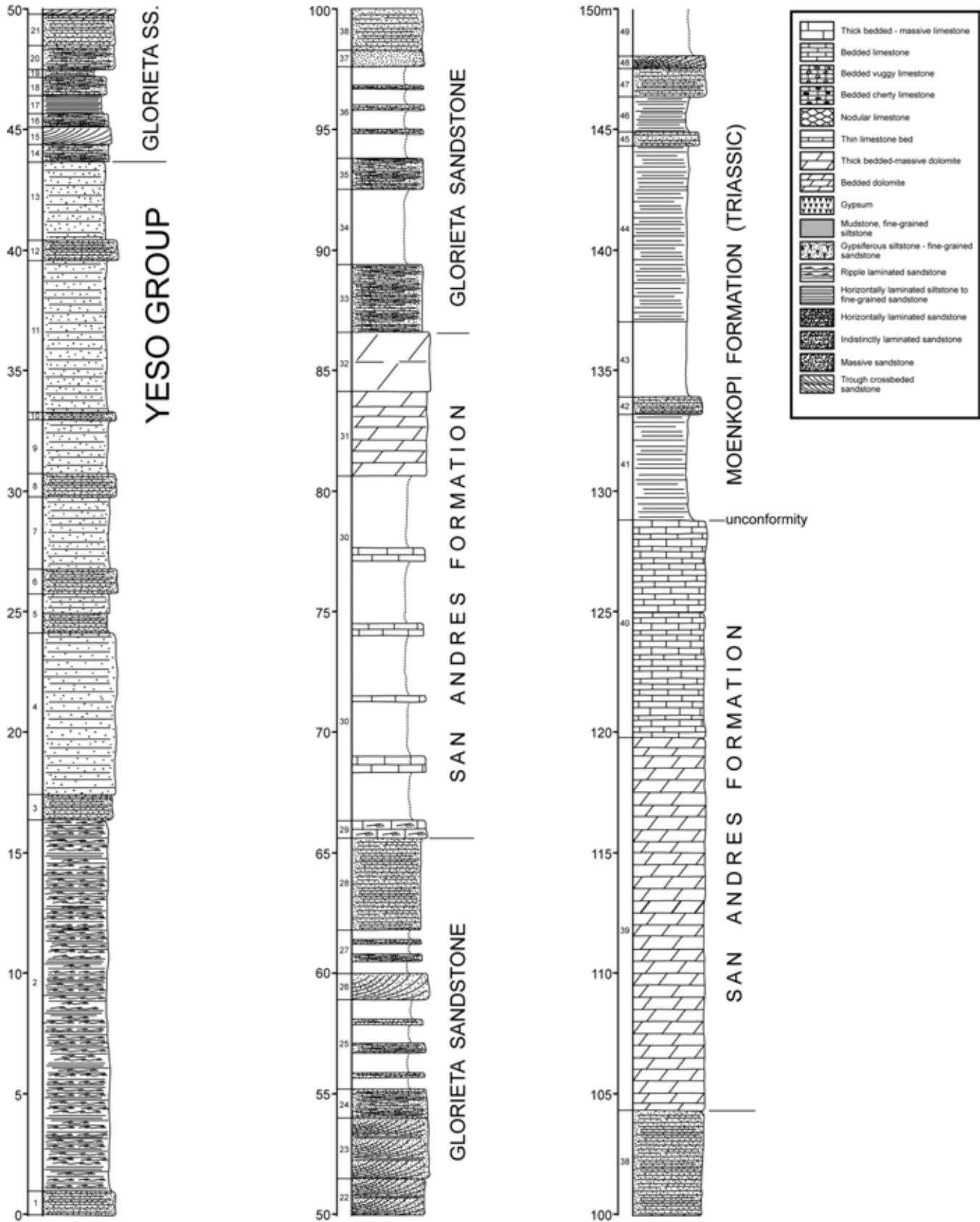


FIGURE 5. San Andres Formation near Cedar Crest, New Mexico, illustrating intertonguing of Glorieta and San Andres.

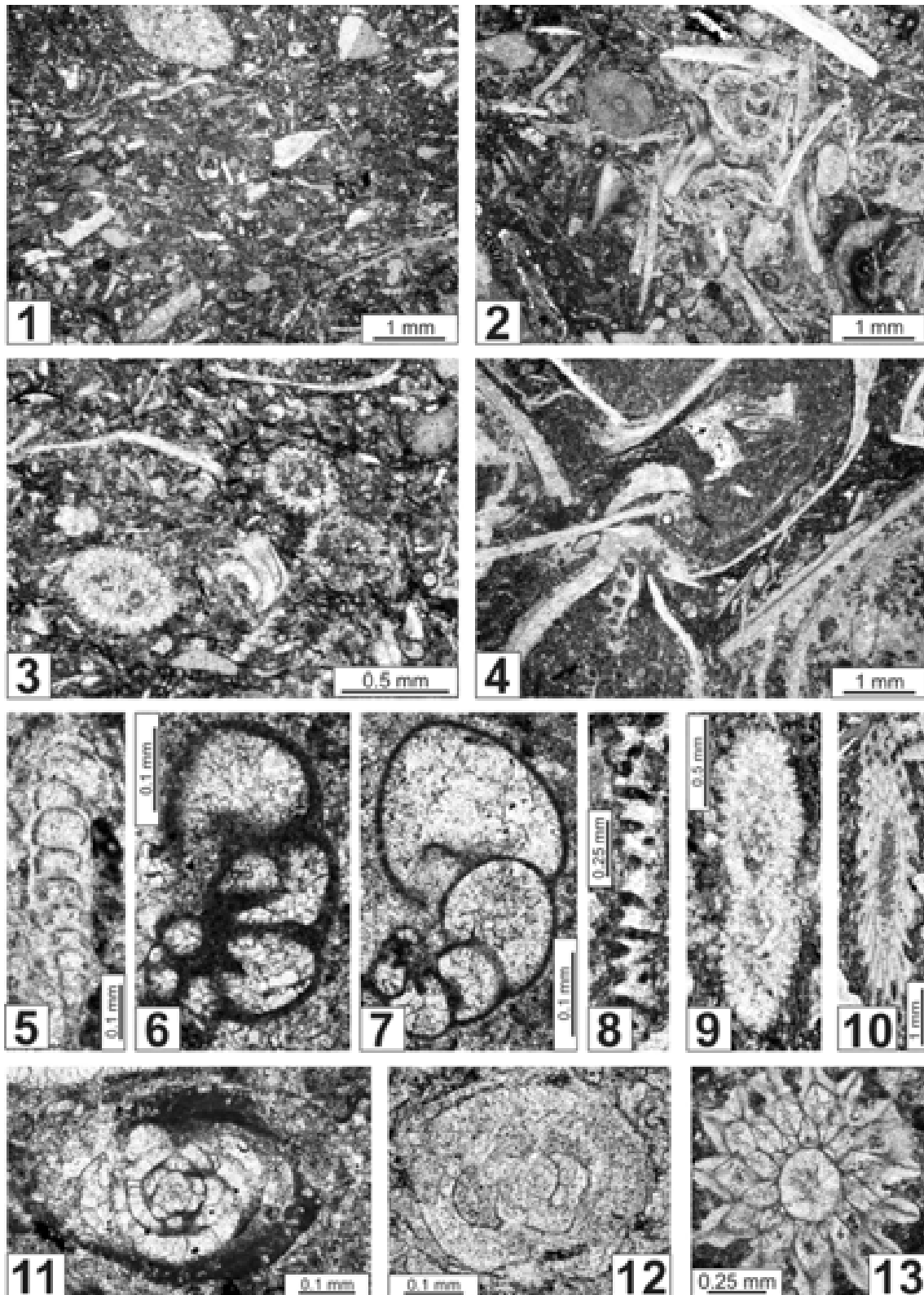


FIGURE 6. Common microfacies and microfossils in thin section of the San Andres Formation at the reference section (SAR) and McLeod Hills (MLY). **1**, Bioclastic wackestone containing a diverse fossil assemblage of echinoderms, brachiopods, bryozoans, smaller foraminifers (*Globivalvulina*), gymnocodiacean algae and other recrystallized skeletons. Sample SAR 7. **2**, Rudstone containing recrystallized skeletons of echinoderms, gastropods, brachiopods, smaller foraminifers and calcareous algae. Sample SAR 13. **3**, Wackestone containing many recrystallized gymnocodiacean algae, ostracods, smaller foraminifers and other skeletal grains. Sample SAR 8. **4**, Floatstone containing large recrystallized shell fragments, a few bryozoans and other skeletal grains embedded in micritic matrix. Sample SAR 9. **5**, *Geinitzina*, sample SAR 8. **6-7**, *Globivalvulina* sample MLY 3 and 6. **8**, *Paraboueina*, sample MLY 6. **9**, *Gymnocodium*, sample SAR 8. **10, 13**, Trepostome bryozoans, sample SAR 10. **11**, *Midiella*, sample MLY 5. **12**, *Orthovertellopsis*, sample SAR 11.

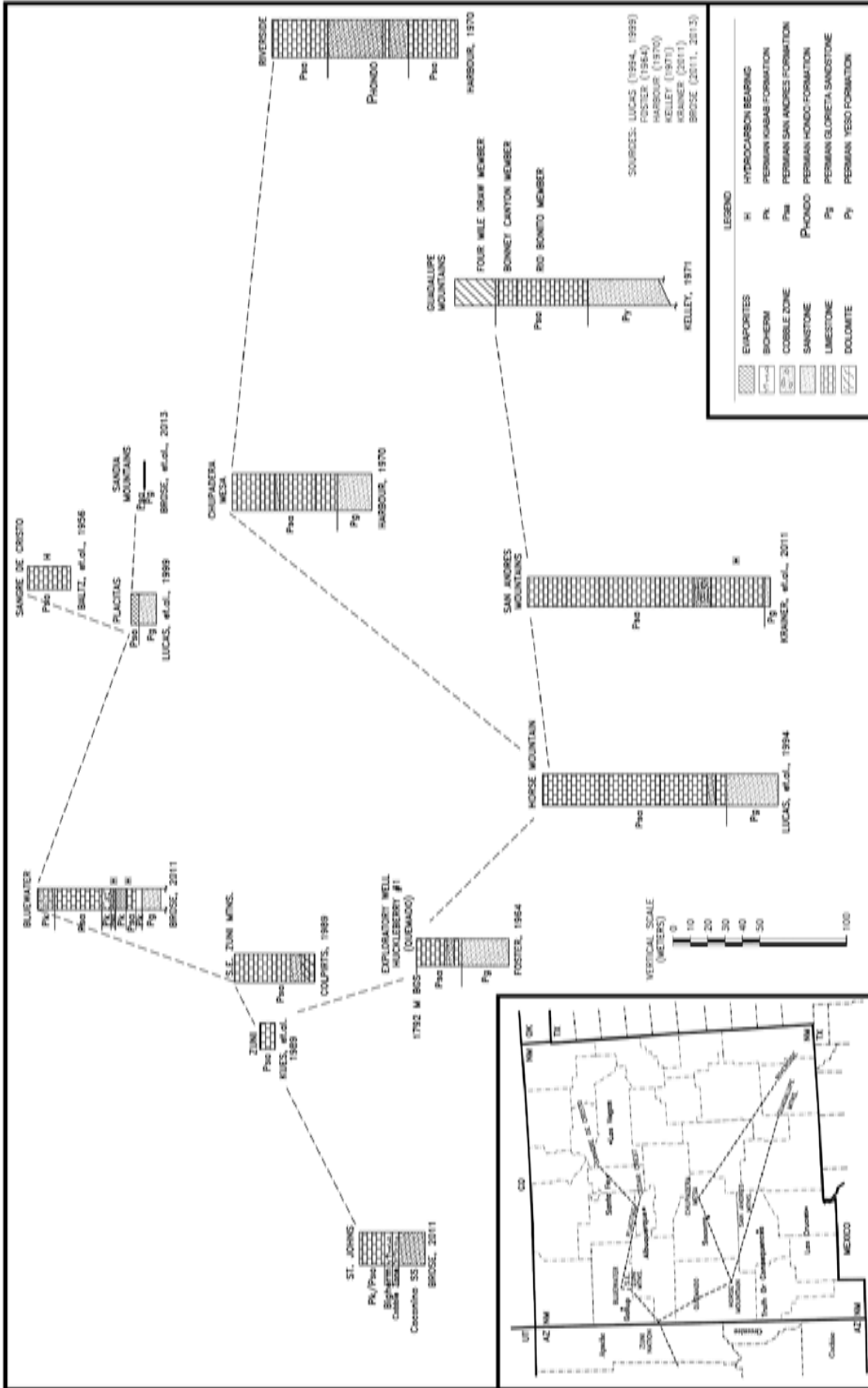


FIGURE 7. Fence diagram of San Andres Formation in New Mexico.

presence of a restricted marine assemblage that includes ostracods, gastropods, foraminifers, pelecypods, and spirorbid worm tubes.

Near Bluewater Lake in Cibola and McKinley counties, the lower portion of the San Andres Formation contains a few large fossils. The predominant faunal assemblage located in the upper portion of the San Andres Formation consists of brachiopods and bivalves. Fossils common to both the San Andres Formation and the Kaibab Formation were also encountered near an outcrop south of St. Johns, Arizona, although the stratigraphy is mapped as San Andres Formation (Kues and Lucas, 1989). Species include brachiopods, bivalves, gastropods, scaphopods and nautiloids.

Near Placitas in Sandoval County, Lucas et al. (1999) indicate that faunal assemblages are not observed in the isolated San Andres Formation outcrops. Similarly, at Cedar Crest (Fig. 5) there are no fossils of note in the San Andres Formation.

The age of the San Andres Formation has mostly been determined from fusulinid and conodont data obtained in southeastern New Mexico and West Texas. These data indicate that the San Andres Formation is of late Leonardian and early Guadalupian (primarily Roadian) age (Kerans et al., 1993 and references therein). In central and west-central New Mexico, however, all data suggest that Guadalupian strata, if ever deposited, have been eroded.

SEDIMENTATION/DEPOSITIONAL ENVIRONMENTS

The San Andres Formation can be characterized by several distinct depositional stages. The first was an initial transgression from what is today east to west over an extensive and possibly uninterrupted eolian dune field (the greater Coconino erg) extending from eastern New Mexico across the current southwestern Colorado Plateau. This was followed by a recurring period of deposition of reworked sands and, near shelf margins, submarine channelization with accompanying debris deposits from both submarine flows and adjacent terrestrial sources. Following this stage, the San Andres Formation was in a relatively quiescent phase with the development of a diverse shelf faunal environment combined with local shallow areas near the perimeter of the shelf environment. Eventually, the San Andres sea regressed eastward, with the transgression of the Kaibab Formation overlapping the regressive San Andres Formation possibly as far west as western Texas.

As just noted, the initial transgression of the San Andres sea from what is today east to west occurred over an extensive and possibly uninterrupted eolian dune field that extended from eastern New Mexico into northern Arizona, southern Utah and Colorado (?). The dune field was truncated to the north by a highland extending from north-central New Mexico through the northern portion of the Colorado Plateau and northerly into the Harrisburg Mesa area of southern Utah (Baars, 1962; White and Kelley, 1989; Brose, 2011). Clastic boulder and cobble conglomerates, conglomeratic sandstones and stream deposits drained southerly into the marginal shelf environment, resulting in channelization of stream deposits, deposition of gravels within eolian dune sets, and localized scouring of the loose and wind-blown sands, as illustrated in Figure 7.

Faunal assemblages and thicknesses suggest that the San Andres marine environment was an offshore, normal marine setting for most of the time of deposition of the formation (Kues and Lucas, 1989; Krainer et al., 2012). Given the transitional nature of the encroaching marine environment, nearshore hypersaline lagoons, nearshore shallow environments and shallow shelf conditions predominated along and near the

shoreline(s). This environment was periodically interrupted by the deposition of windblown sands sourced from the adjacent eolian dune fields and shallow marine near-shore environments. These were formed during the late Leonardian-Guadalupian transgression, which was the last major marine transgression on the North American craton (Krainer et al., 2012). Published descriptions of the San Andres Formation (Baltz, 1956) indicate that the northernmost extent of the San Andres marine environment likely terminated near the Sangre de Cristo Mountains. Baltz's accounts of a bituminous odor associated with freshly exposed surfaces are consistent with outcrops of the lowest section of the San Andres Formation at other locations.

Eventually, the transgressive marine environment appears to have breached near-shore bioherm deposits associated with the adjacent Kaibab marine environment along the western margin of New Mexico. The second period of San Andres development is associated with numerous, repetitive high-energy events along the shelf margin. These events resulted in the transport and deposition of sands from the adjacent dune fields, clastic flows from the adjacent highlands southward into embayments and open marine shelf environments, and events that resulted in Bouma-type deposits along shelf margins, scouring of the marine seafloor and transport of large clastics some distance easterly onto the adjacent San Andres seafloor. The resulting turbidity caused by these events can likely be observed as far away as the San Andres Mountains, where muddy textures dominate the lower sequences of the San Andres stratigraphy (Krainer et al., 2012) and appear to represent the distal, easternmost boundaries of the perimeter event(s).

Outcrops along the Arizona-New Mexico border suggest that a possible high (Pierce, 1989) separating the two (San Andres and Kaibab) marine environments may have been breached on several occasions (Brose, 2011), resulting in the comingling of the San Andres and Kaibab marine environments. The distal deposition resulting from these events appears to have reached the northern shore near the Sangre de Cristo Mountains.

Following this period of redundant, high-energy events, the San Andres Formation was once again in a relatively quiescent phase with the development of a diverse shelf faunal environment combined with local shallow areas (backwater lagoons, embayments, bioherms and tidal flats) near the perimeter of the shelf environment. Marine fauna proliferated as evidenced by the presence of a relatively uninterrupted depositional sequence that resulted in as much as 100 m of gray limestone containing an abundance of diverse faunal assemblages as far west as the New Mexico-Arizona border. It appears unlikely that the San Andres marine environment extended north to the Sangre de Cristo Mountains at this time because the upper part(s) of the section(s) are not present, likely the result of erosion.

The last San Andres Formation sequence is associated with the regression of the San Andres marine environment eastward. Evaporite deposits formed on the regressing marine environment as the shelf margins extended easterly, encroaching into the Midland basin. At some point, either coincidentally or immediately following the San Andres marine regression, the Kaibab marine sequence breached the topographic high separating these two environments and the Kaibab marine environment transgressed easterly, covering the San Andres Formation (Brose, 2011).

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THE PERMIAN ARTESIA FORMATION IN CENTRAL NEW MEXICO

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Abstract—In central New Mexico, the stratigraphically highest Permian strata belong to the Middle Permian (Guadalupian) Artesia Formation (= “Bernal Formation”). Artesia Formation strata are the thin (10-30-m-thick), unconformity-bounded edge of the Artesia Group of southeastern New Mexico, between the Lower Permian San Andres Formation and the Middle Triassic Moenkopi Formation. The Artesia Formation in central New Mexico is mostly reddish brown and reddish orange siltstone and fine-grained sandstone with a few beds of gypsum and dolomite/limestone. Many clastic beds are gypsiferous, bedforms are usually tabular or ripple laminated, and some beds are bioturbated. For decades, overlying Triassic red beds of the Moenkopi Formation were included in the Artesia Formation, but these strata are readily distinguished from Artesia Formation red beds. There are no Artesia Formation outcrops west of the Rio Grande in central New Mexico; its westernmost outcrops are near Lamy in Santa Fe County and east of Socorro in Socorro County. The Artesia Group represents deposition on the vast evaporitic to siliciclastic shelf landward of the great reefs that fringed part of the western edge of the Permian basin. Artesia Formation strata in central New Mexico are the remnant edge of this vast shelf.

INTRODUCTION

In central New Mexico, the stratigraphically highest (and therefore youngest) Permian strata belong to the Middle Permian (Guadalupian) Artesia Formation. In parts of the northern highlands and Rio Grande Valley of New Mexico, strata of the Artesia Formation were long referred to as Bernal Formation and often confused with Triassic strata. However, recognition of the presence of the Triassic Moenkopi Formation across central New Mexico (e. g., Lucas and Hunt, 1987; Lucas and Hayden, 1989a, b, 1991; Lucas 1991a, b, 2004; Lucas and Heckert, 1994, 1995, 1996; Spielmann and Lucas, 2009) clarified the distribution of the Artesia Formation. Here, I present a brief overview of current understanding of the stratigraphy of the Artesia Formation in central New Mexico (Fig. 1).

HISTORY OF STUDY

Across central New Mexico (principally Sandoval, Santa Fe, San Miguel, Bernalillo, Valencia and Socorro Counties), a relatively thin succession of siliciclastic red beds immediately overlies the Lower Permian San Andres Formation. Early workers (e. g., Lee, 1909; Darton, 1928) regarded these red beds as Triassic in age. However, U. S. Geological Survey geologist Charles. B. Read first identified at least some of these red beds as Permian strata during the 1930s. Read intended to coin the name Bernal Formation for these Permian strata.

Read first informally used the term Bernal Formation in 1939. He planned to use the name in a U.S. Geological Survey map of north-central New Mexico (Read and Andrews, 1944; Read et al., 1944), but at that time, the Survey did not want to adopt new stratigraphic names. Instead, Read et al. (1944) used the acronym “Pb” to identify the “upper member” of the San Andres Formation. In the map legend, they described the unit as “a fine-grained clastic member.” Measured sections with the map indicate this unit is mostly orange siltstone and massive sandstone with a bed of white gypsum. The maximum thickness of the upper member of the San Andres Formation reported by Read et al. (1944) was about 66 m (218 ft) at “Bernal Butte and Chapelle Butte, near Bernal, New Mexico” (their section 35). Here, the “upper member” rests on limestone of the San Andres Formation and is overlain by sandstone that Read et al. (1944) termed “Dockum Formation.”

Two years later, Wood and Northrop (1946) mapped the upper clastic member of the San Andres Formation, identified as “Pb” in their legend, in the Nacimiento Mountains of Sandoval County. Kelley (1949,

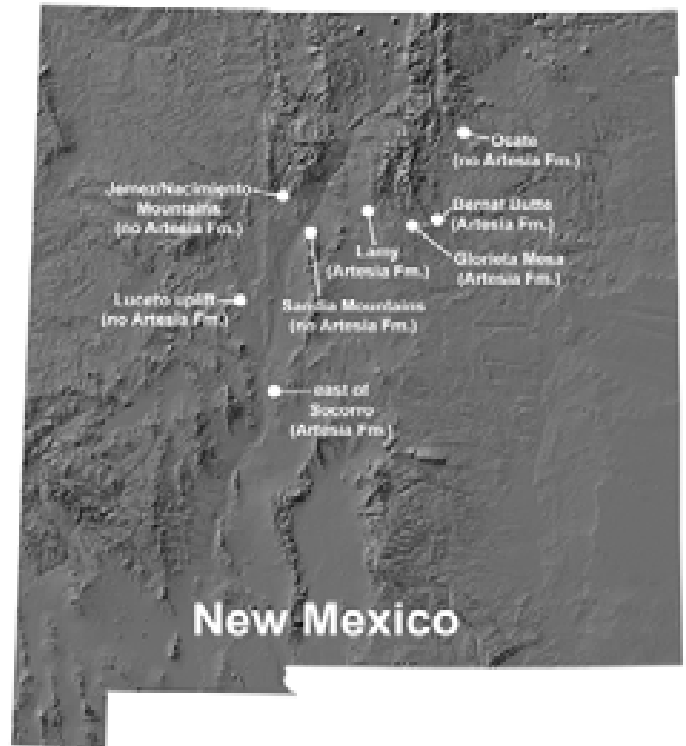


FIGURE 1. Map of New Mexico showing distribution of key outcrops of Artesia Formation and Moenkopi Formation strata referred to in the text.

fig. 2) first published the term Bernal in a table, apparently believing that Read had already defined it formally. However, it was Bachman (1953) who first formally introduced the term Bernal Formation on a map legend in an area where no “Bernal” strata are present. In so doing, Bachman (1953) described the Bernal as brownish-red siltstone and fine-grained sandstone and mapped its supposed distribution in the Ocate area of northwestern Mora County (Fig. 1). However, he failed to describe a type section, although Bachman (1953) noted “the type sequence is near the villages of Bernal and Chapelle, San Miguel County, and the formation has been recognized over much of central and northwestern New Mexico.”

The term Bernal Formation was used frequently in the guidebook to the seventh field conference of the New Mexico Geological Society held in 1956. Baltz et al. (1956, especially p. 42) identified “Bernal Butte” in a photograph labeled to show the extent of the Bernal and adjacent units. Baltz and Bachman (1956, p. 101-102) summarized the lithology and distribution of the Bernal Formation in the southeastern Sangre de Cristo Mountains and noted its correlation with the Guadalupian Whitehorse Group of southeastern New Mexico. In 1958, Read and Hayes discussed the nomenclatural problems in eastern New Mexico and correlation of the Bernal, Whitehorse, and Chalk Bluff formations.

By 1962, Bernal Formation was a well-accepted name for the youngest Permian strata in northern New Mexico (e. g., Baars, 1962). It also was being applied to the youngest Permian strata in south-central New Mexico (e. g., Wilpolt and Wanek, 1951; Anonymous, 1955; Smith and Budding, 1959; Smith, 1964; Weber, 1964). Nevertheless, Tait et al. (1962) presented a unified stratigraphic terminology for Guadalupian clastic/evaporitic units in southeastern New Mexico. They proposed the term Artesia Group to encompass five formations (Grayburg, Queen, Seven Rivers, Yates, and Tansill, in ascending order). Because the Bernal Formation is equivalent to the lower part of the Artesia Group (principally the Grayburg and Queen formations), Tait et al. (1962) recommended abandonment of the term Bernal. Few subsequent authors followed this recommendation (Dixon, 1967 is an exception), and Bernal Formation continued to be used for the youngest Permian strata in north and south-central New Mexico into the 1980s.

In the mid-1980s, working with A. P. Hunt and S. N. Hayden, I realized that many “Bernal Formation” outcrops in north-central New Mexico are Triassic Moenkopi Formation (Lucas and Morales, 1985; Lucas et al., 1985; Lucas and Hunt, 1987; Lucas and Hayden, 1989a, b, 1991). For this reason, Lucas and Hayden (1991) described the type section of the Bernal Formation and thus clarified the differences between Permian “Bernal” strata and Triassic Moenkopi strata (Fig. 2). This clarification allowed recognition of the actual distribution of “Bernal” and Triassic strata across central New Mexico, which is briefly summarized here.

STRATIGRAPHIC NOMENCLATURE

The term Artesia Formation should replace Bernal Formation across central New Mexico (Tait et al., 1962; Lucas and Hayden, 1991). Careful distinction should be made between Artesia strata and siliciclastic red beds of the Moenkopi Formation (see below and Figures 2-3). Even though the Artesia strata of central New Mexico are correlative to all or parts of three Artesia Group formations in southeastern New Mexico (Fig. 4), my data do not suggest that Artesia strata in central New Mexico can be consistently subdivided into member- or bed-level units.

LITHOSTRATIGRAPHY

Lithology

The Artesia Formation in central New Mexico is mostly reddish brown and reddish orange siltstone and fine-grained sandstone. Minor lithologies are gypsum and dolomite. Many clastic beds are gypsiferous. Bedforms are usually tabular or ripple laminar, and some beds are bioturbated. Lucas and Hayden (1991) provided some petrographic data on Artesia Formation strata that indicate siltstone of the formation is calcite cemented, more than 95% quartz and contains minor feldspar, mica and authigenic anhydrite. Carbonate beds of the Artesia Formation are mostly dolomitic lime mudstone.

For decades, overlying Triassic red beds of the Moenkopi Formation were included in the Artesia (“Bernal”) Formation. However, these strata are readily distinguished from Artesia Formation red beds. Thus, the Moenkopi strata display typical fluvial sedimentary structures, mostly trough crossbeds, are mostly grayish-red, and are dominantly immature sandstones (lithic graywackes and litharenites) with lesser

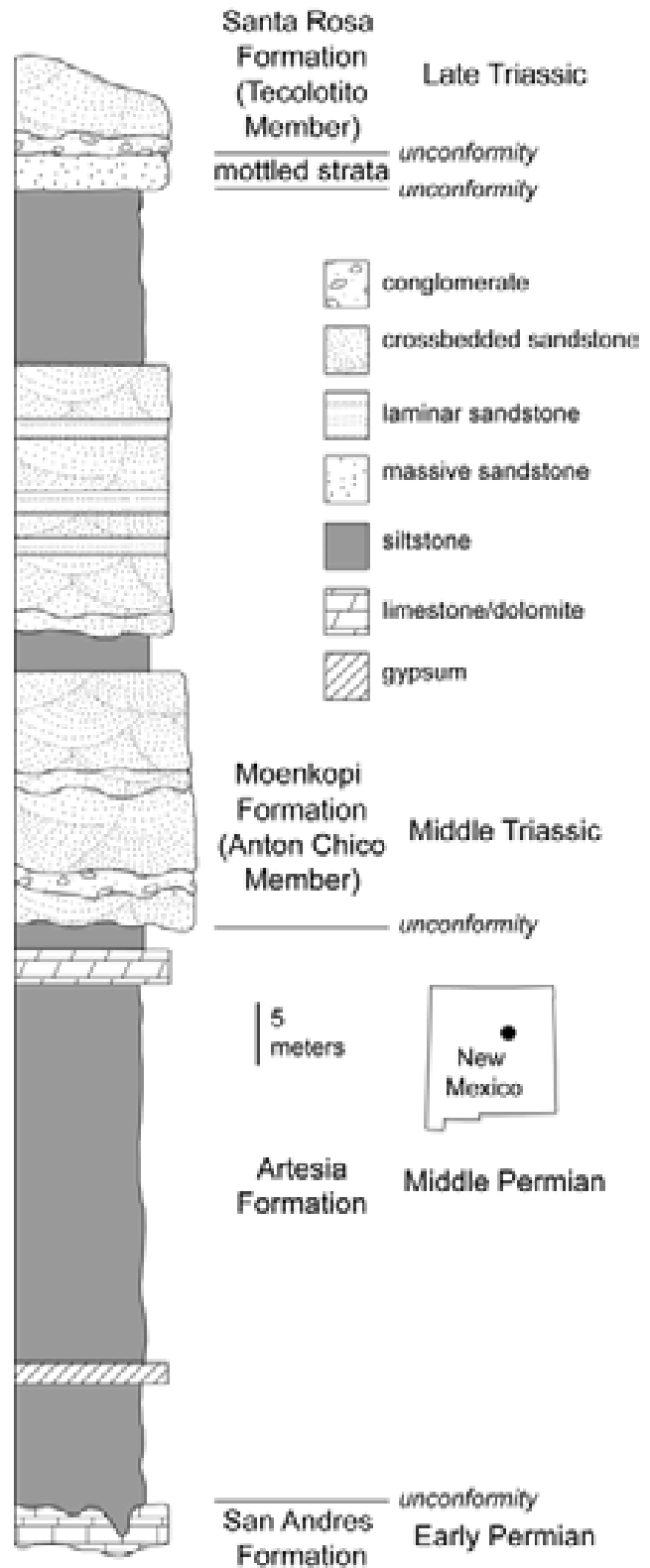


FIGURE 2. Type section of the “Bernal Formation” at Bernal Butte near Las Vegas, in San Miguel County (after Lucas and Hayden, 1991). This section of the Artesia (= Bernal) Formation is characteristic in consisting mostly of siltstone with thin beds of dolomite and gypsum. The base of the Artesia Formation is a karsted, unconformable contact on the underlying Lower Permian (Leonardian) San Andres Formation. Fluvial sandstone and conglomerate of the Middle Triassic Moenkopi Formation unconformably overlie the Artesia Formation strata.

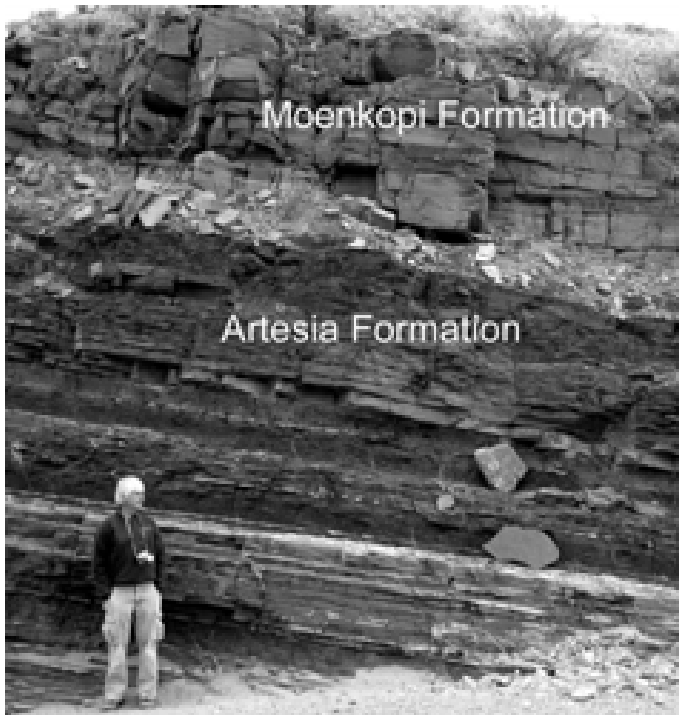


FIGURE 3. Characteristic outcrop of Artesia Formation and Moenkopi Formation strata, near Carthage in Socorro County. Note the easy distinction of flat-bedded Artesia Formation siltstones and sandstones from overlying crossbedded and pebbly Moenkopi Formation sandstones. The contact of the two formations is at the rubbly zone just below the Moenkopi cliff.

amounts of siltstone, mudstone and intraformational conglomerate composed of siltstone and nodular limestone-pebbles. In contrast, Artesia Formation red beds are typically laminar, bioturbated or massive, reddish brown and reddish orange and are dominantly mature sandstones and siltstones that are often gypsiferous and are intercalated with a few beds of gypsum and dolomite/limestone.

Contacts

Across its outcrop area, the Artesia Formation is an unconformity-bounded unit. Its basal contact with underlying San Andres Formation limestone is a sharp and irregular paleo-karst marked by dissolution and brecciation (Fig. 2). This regional unconformity is in part responsible for the relatively thin preserved sections of the San Andres Formation in parts of northern New Mexico.

The contact of the Moenkopi Formation above the Artesia Formation is a sharp surface at which channelform sandstone and conglomerate at the base of the Moenkopi is scoured into underlying Artesia strata. Paleotopographic relief on this unconformity can be up to a few meters locally.

Thickness

Thickness of the Artesia Formation across central New Mexico is typically 10-30 m (Lucas, 1991a, b; Lucas and Hayden, 1991; Spielmann and Lucas, 2009). In general, the Artesia lithosome thickens to the southeast to a maximum of about 520 m in the “type” well log in Eddy County, southeastern New Mexico (Tait et al., 1962) (Fig. 4).

Distribution

In north-central New Mexico, there are no outcrops of the Artesia Formation north and west of Santa Fe. Outcrops in the Jemez Mountains of Sandoval County and the Sandia Mountains of Bernalillo County (Fig. 1) long termed “Bernal Formation” (e.g., Baars, 1962; Kelley and

Northrop, 1975; Woodward, 1987) pertain to the Triassic Moenkopi Formation (Lucas and Hayden, 1989a, b; Lucas, 1991a; Lucas and Heckert, 1995, 1996; Lucas et al., 1999). Outcrops to the northeast of Santa Fe, in the Ocate area and vicinity in Mora County (Fig. 1), mapped as “Bernal Formation” (e.g., Bachman, 1953; Baltz and O’Neill, 1984, 1986), are also Moenkopi Formation strata (Lucas et al., 1990). However, in western San Miguel County, across Glorieta Mesa to Bernal Butte, the type section of the “Bernal Formation,” the Artesia Formation is exposed as a relatively thin unit between the San Andres or Glorieta formations and the Moenkopi Formation (Lucas and Hayden, 1991; Lucas, 1995), though these strata were earlier mapped together with the Moenkopi Formation as “Bernal Formation” (e.g., Read et al., 1944; Johnson, 1970, 1974; Budding, 1972). North of Albuquerque, the westernmost outcrops of the Artesia Formation are in the vicinity of Lamy in Santa Fe County (Lucas, 1991a; Lucas and Hayden, 1991) (Fig. 1).

South of Albuquerque, there are also no Artesia Formation outcrops west of the Rio Grande. For example, in the Lucero uplift of Valencia County (Fig. 1), Moenkopi Formation strata disconformably overlie the San Andres Formation (Lucas and Heckert, 1994; Lucas and Zeigler, 2004; Lucas et al., 2004). South of Albuquerque, the Artesia Formation is present between the San Andres and Moenkopi formations east of the Rio Grande in Socorro County (Fig. 1) and in Lincoln County (Lucas, 1991b; Spielmann and Lucas, 2009). Farther to the south, in Sierra County, Cretaceous rocks rest directly on the San Andres Formation.

Although there are no mapped Artesia Formation outcrops west of the Rio Grande, R. Brose (written communication, 2013) has suggested that the Artesia Formation may be (at least in part) equivalent to strata termed Kaibab Formation in western New Mexico and eastern Arizona. At these locations, Kaibab strata overlie and may intertongue with the underlying San Andres Formation. The Kaibab, in turn, is capped by the Triassic Moenkopi Formation and was deposited in paleoenvironments similar to those in which the Artesia Formation was deposited. Further work needs to be undertaken to decipher the relationship of the Kaibab and Artesia formations.

PALEONTOLOGY, AGE AND CORRELATION

The only fossils I have observed in Artesia Formation strata in central New Mexico are nondescript bioturbation. The age and correlation of the Artesia strata thus are based on their stratigraphic position and regional stratigraphic relationships. These indicate that the Artesia Formation is the equivalent of the lower part of the Artesia Group in southeastern New Mexico, primarily the Grayburg and Queen formations (e.g., Tait et al., 1962; Kelley, 1972a, b) (Fig. 4). These are strata of Middle Permian (Guadalupian, mostly Wordian) age, so the Artesia Formation can be assigned a Middle Permian age.

DEPOSITIONAL ENVIRONMENT

No detailed sedimentological studies of the Artesia Formation have been undertaken in central New Mexico. But, in a general sense, the Artesia Group represents deposition landward of the great reefs that fringed part of the western edge of the Permian basin (e.g., Tait et al., 1962; Silver and Todd, 1969; Fekete et al., 1986; also see Hill, 1996 and references cited therein). The Artesia Group strata were thus deposited on the vast evaporitic to siliciclastic shelf behind those reefs (Fig. 5). From southeast to northwest, Artesia Group strata thin, change from evaporitic to siliciclastic and are progressively truncated by the sub-Triassic unconformity (Fig. 4).

Artesia Formation strata in central New Mexico are thus the remnant western and northwestern edges of this vast shelf (Fig. 5). This marginal marine shelf origin of the Artesia Formation is supported by the presence of dolomites and evaporites (deposited in sabkha environments with periodic inundations of sea water) interbedded with siltstone and

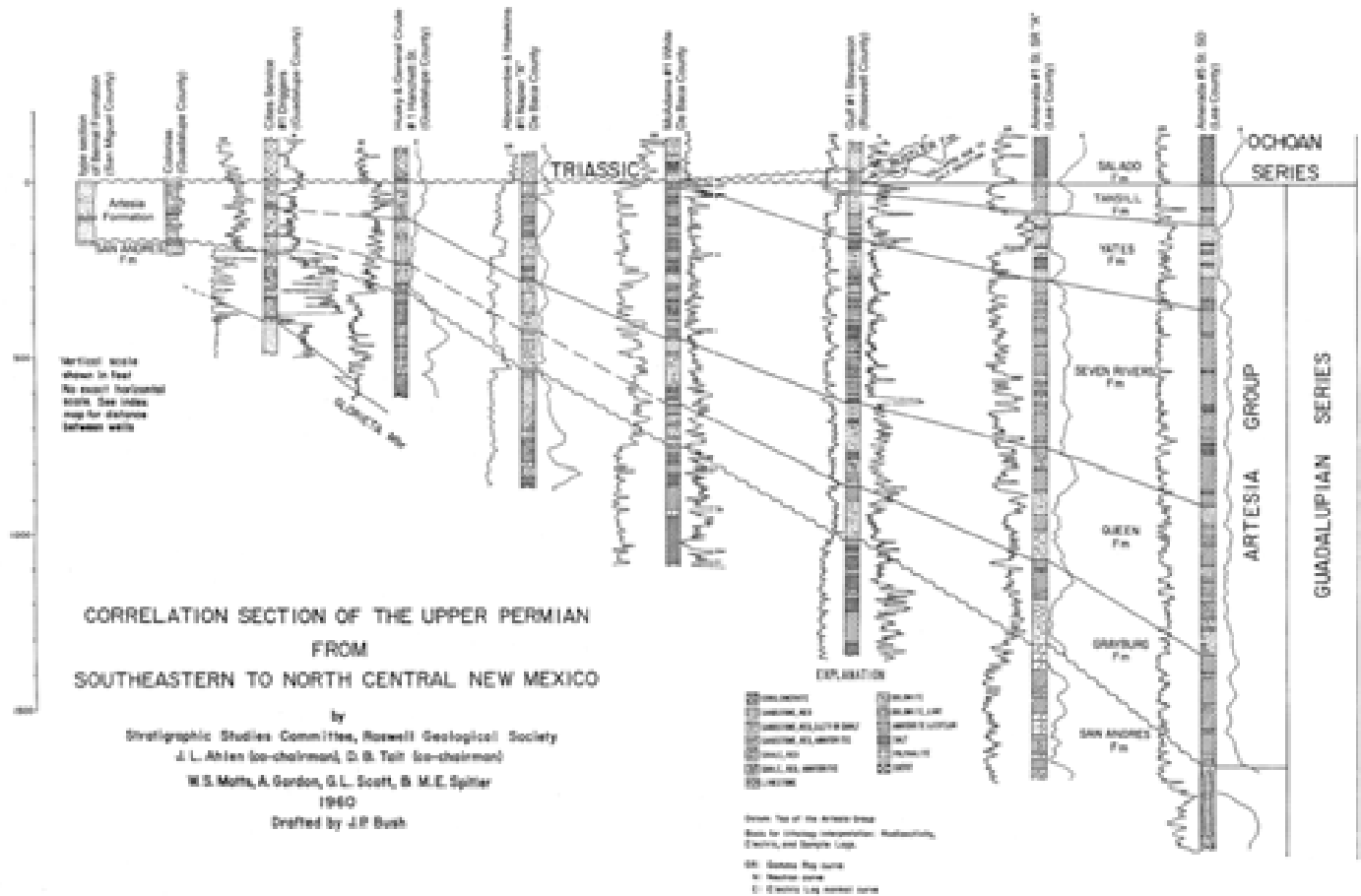


FIGURE 4. Cross section across part of eastern New Mexico, showing stratigraphic relationships of Artesia Group and Artesia Formation strata, modified from Tait et al. (1962, fig. 3). The stratigraphic cross section shows the northwest thinning of Artesia Group units that are progressively being truncated at the sub-Triassic unconformity. The correlation indicates that the Artesia Formation of central New Mexico (two sections to far left) is equivalent to parts of the Grayburg, Queen and Seven Rivers formations of the Artesia Group to the southeast.

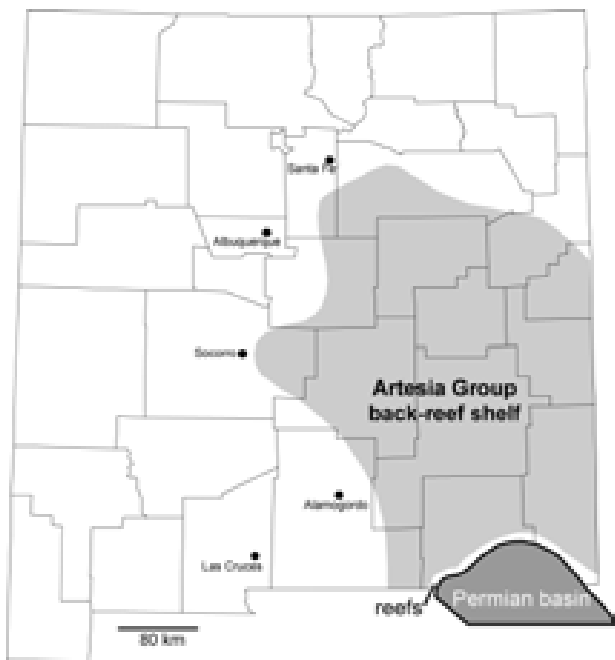


FIGURE 5. Paleogeographic map of New Mexico during Guadalupian time (modified from Kues and Giles, 2004, fig. 15) showing extent of Artesia Group/Formation back-reef shelf deposits.

sandstone (deposited during shoreline retreat).

So, this is a classic “carbonate rimmed shelf,” where the Capitan reef complex and associated platform define the northwestern margin of the Delaware Basin (Permian Basin). The reef complex separates the shallow-water deposits to the north-northwest (Artesia Group) from the basinal deposits to the southeast. Facies indicates that behind the rim (reef complex) a broad, shallow shelf existed that was protected from the open sea. Immediately behind the rim (reef complex) mostly carbonate sediments were deposited in a restricted shallow marine environment with poor circulation and increased salinity in a tidal flat-lagoonal environment. This facies towards the north-northwest grades into coastal sabkha environments with evaporitic sediments and fine-grained siliclastic sediments with minor carbonate deposits, and finally into coastal plain deposits dominated by fine-grained siliclastic sediments – Artesia Group/Formation strata.

ACKNOWLEDGMENTS

I dedicate this article to the memory of Steven N. Hayden (1946-2010). During the late 1980s and early 1990s, Steve’s work on Artesia and Moenkopi strata in central New Mexico extended from the Sandia Mountains to Gallup, and was integral to understanding the regional stratigraphy of these units. I also had much collaboration and assistance in the field from Andrew Heckert, Adrian Hunt, Larry Rinehart and Justin Spielmann, for which I am grateful. Richard Brose and Karl Krainer provided helpful reviews of an earlier draft of this paper.

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UPPERMOST PENNSYLVANIAN BURSUM FORMATION NEAR CIBOLA SPRING, SEVILLETA NATIONAL WILDLIFE REFUGE, SOCORRO COUNTY, NEW MEXICO

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Abstract—Mixed carbonate and siliciclastic deposits of the Bursum Formation near Cibola Spring record the transition from marine to terrestrial depositional environments that occurred in central New Mexico near the Pennsylvanian-Permian boundary. The deposits are about 40 m thick, are well exposed, and contain age-diagnostic fossils (fusulinids). A brief examination of these rocks was conducted during geological mapping efforts in this area, the results of which are summarized in this note. Good access to exposures near Cibola Spring makes this an attractive place for more detailed, future investigation of Bursum strata in the region.

INTRODUCTION

Pennsylvanian strata of the Bursum Formation are well exposed in the vicinity of Cibola Spring on the Sevilleta National Wildlife Refuge (SNWR), Socorro County, central New Mexico (Fig. 1). Surface exposures of the rocks are displayed within a geologically rich setting that includes overlying and underlying Paleozoic sedimentary rocks, faults and folds related to Laramide and superimposed Rio Grande rift tectonic activity, Paleogene igneous intrusions (mafic dikes and sills), a ~3 Ma basalt flow (Black Mesa), and are easily accessible within short walking distances of roads on the eastern side of the SNWR. Just to the west of Cibola Spring the Bursum is relatively undeformed and its complete thickness is exposed at closely spaced outcrops. Since our initial visit to the site in 2011, SNWR personnel have led field trips to the area for the general public in order to showcase some of the geological features that are present on the refuge. In this note we provide a brief description of these uppermost Carboniferous rocks as they are well exposed, fossiliferous, easily accessible on public land (a permit is required to conduct research on SNWR), and are in proximity to outcrops of the Bursum that have been described in detail by Krainer and Lucas (2009). We present generalized graphic sections of the Bursum Formation and a few photographs of common fossils. The intent is to provide information to both the layperson and to specialists who may want to conduct detailed studies of the Bursum Formation and the Carboniferous-Permian transition at this site.

SETTING AND PREVIOUS WORK

The study area (Fig. 1) is located near the eastern margin of the Rio Grande rift, an elongate N-S trending structural feature of regional extent resulting from Neogene crustal extension. Rift-bounding fault-block uplifts (mountains) alternate on either side of the rift, which extends across New Mexico from Las Cruces on the south to the border with Colorado north of Taos. The closest rift-bounding uplift (Los Pinos Mountains) lies a short distance to the northeast of the study area. As discussed in many reports (see Lueth et al., 2009 for discussions and literature citations), the area also contains well-exposed geological evidence for earlier (late Cretaceous-Paleogene) compressional tectonics (Laramide orogeny) as well as late Paleozoic Ancestral Rocky Mountain (ARM) tectonic activity. Indeed, local (Joyita uplift to the west) as well as regional (Pedernal landmass to the east) uplifts and intervening troughs associated with the ARM orogeny undoubtedly played a role in sedimentation patterns during deposition of the Bursum Formation in this area (Thompson, 1954; Kottlowski and Stewart, 1970; Beck and Johnson, 1992; Krainer and Lucas, 2009).

During the time of deposition of the Bursum Formation (roughly a few million years at the end of the Carboniferous Period), fluctuations of sea level due to relatively high-frequency glacial-interglacial episodes

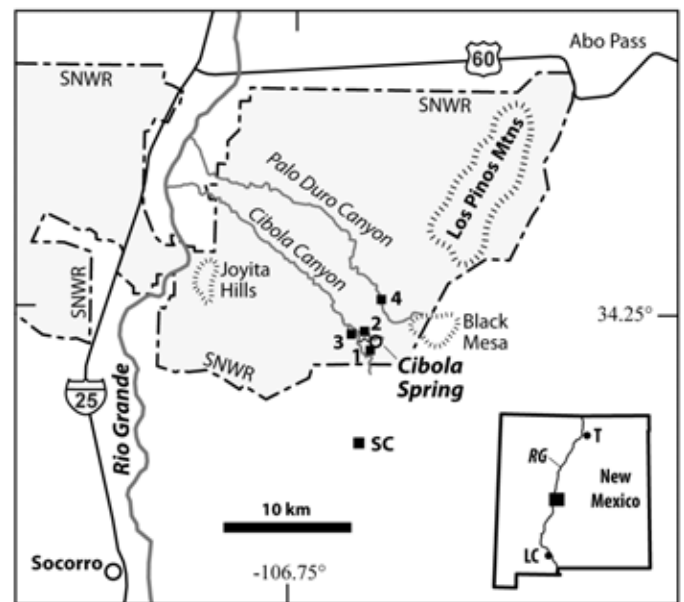


FIGURE 1. Map showing location of measured stratigraphic sections of the Upper Pennsylvanian Bursum Formation near Cibola Spring. Sections are depicted in Figure 2; SC is the Sierra de la Cruz stratigraphic section described by Krainer and Lucas (2009). Outline labeled SNWR is the Sevilleta National Wildlife Refuge boundary. In the inset map, T is Taos, LC is Las Cruces, and RG is the Rio Grande.

are interpreted to have taken place (Rygel et al., 2008). Such rises and falls, superimposed on ARM tectonic activity, probably influenced the timing, location, and extent of marine and non-marine deposition in the Bursum Formation.

Today, the study area lies within a semi-arid desert setting, characterized by grasses, shrubs and a scattering of small trees (mostly juniper), at an elevation of about 1700 m (5577ft). Downcutting of the Rio Grande to the west began after about 0.8 Ma, and the area has been deeply incised by large tributaries with headwaters in the mountains to the east. The bedrock geology is thus well exposed, although hillslopes are commonly covered with up to a few meters of colluvium, and fluvial terraces at various levels above modern valley floors are capped by gravels derived from the east. Nonetheless, good outcrops of bedrock are present, and it is not difficult to piece together details of the Paleozoic stratigraphy from adjoining exposures.

General geological observations for the region east of the Rio Grande in the vicinity of Socorro were presented by Darton (1928), and

reasonably detailed geologic maps of the area were prepared by Wilpolt et al. (1946) and Wilpolt and Wanek (1951) as part of the U.S. Geological Survey Oil and Gas Investigation series. The maps prepared by the USGS workers, despite their reconnaissance nature, remain useful when becoming acquainted with the geology of this region. It was in the Wilpolt et al. (1946) compilation that the name "Bursum Formation" was coined. Although other names for this rock unit have been proposed, this is the name that has ultimately been accepted for the rocks representing the marine-terrestrial transition that occurred in central New Mexico at the end of the Pennsylvanian (Lucas et al., 2002).

Also, during the mid-20th century the prolific American fusulinid specialist and stratigrapher, M.L. Thompson, conducted a considerable amount of work studying upper Paleozoic rocks in New Mexico. Especially pertinent to the strata discussed here is his 1954 treatise titled "American Wolfcampian fusulinids." In particular, many of the fusulinid species that have been recognized from the Bursum in this region were first named and described by Thompson (1954).

As discussed in greater detail in a number of publications (see, for example, Lucas et al., 2000; Wilde, 2006), the presence of schwagerinid fusulinids in the Bursum Formation in Socorro County (and elsewhere in New Mexico) led to the assignment of a Permian age (Wolfcampian Stage) to these rocks. Now, with refinements to the global stratigraphic time scale and improved biostratigraphic correlations between the North America and Eurasian continents, the consensus appears to be that the Bursum Formation is Upper Carboniferous, that is, of Pennsylvanian age. The problem appears to remain as to what stage name should be applied to schwagerinid-bearing Bursum beds. This is a topic that may be addressed by other papers in this volume.

Our first trip to the Bursum outcrops west of Cibola Spring were sparked by a short description of the area in Kottlowski and Stewart's (1970, p. 24) report regarding the ARM Joyita uplift (Kottlowski and Stewart referred to the Cibola Spring outcrops as their Palo Duro Canyon section, but it actually lies in the headwaters of Cibola Canyon). The first reference to the Bursum Formation in the area, however, was probably given by Thompson (1954), wherein he describes collecting fusulinids from highly calcareous shale near the east-west ranch road on the south side of the Los Pinos Mountains. Among other observations, he named a new species of fusulinid, *Schwagerina pinosensis*, from the specimens collected at that time.

During the past 15 years Lucas and Krainer and their colleagues have systematically examined, described, and discussed Bursum strata throughout New Mexico, and their published works are a "must read" for anyone interested in details about this rock unit. The papers authored by Lucas and Krainer cited herein (Lucas et al., 2000, 2002; Krainer and Lucas, 2009) provide a good introduction to the literature on the Bursum Formation in New Mexico.

LITHOLOGY

The Bursum Formation in central New Mexico overall represents a transition from dominantly marine (Pennsylvanian Atrasado Formation) to terrestrial, fluvial coastal plain (Permian Abo Formation) depositional environments (Lucas et al., 2000). The transition took place during several transgressive-regressive cycles of marine incursions and prograding terrestrial depositional environments. The formation thus exhibits a variety of lithologies ranging from fossiliferous normal marine limestones and shales, through marginal marine lagoonal, estuarine, and deltaic calcareous muds and coarser-grained siliciclastic deposits, to continental fluvial floodplain muds and channel sands. This lithologic variability is one of the notable characteristics of the Bursum Formation and makes it especially interesting to study from a paleoenvironmental standpoint. In the Cibola Spring area, the Bursum is about 40 meters thick and consists of about 76 percent siliciclastics (mostly mudstone or shale) and 24 percent limestone (Fig. 2). Conglomerates are rare, although a fairly persistent interval of intraformational (limestone-clast) conglomerate, locally up to 2 meters thick, is present in broad channels approximately 26 meters above the base.

Colors of siliciclastic mudstones and shales to sandy siltstones range from gray to greenish-gray and purplish-gray to reddish-brown. Sandstones are typically reddish-brown on weathered outcrops, although in some areas sands and minor pebbly sands are pale yellowish-brown. The percentage of sandstone varies laterally, ranging from about 3 to 10 percent of the total siliciclastic fraction depending on locality within the area. Limestones are typically light to dark gray, although some beds weather yellowish-brown, or shades of greenish-gray or purplish where they are intercalated with similarly colored shales.

A discontinuous, but laterally persistent limestone bed (wackestone to packstone), about a meter thick, is present a few meters above the base of the Bursum in the study area. The unit is bioturbated, giving it a nodular appearance on many outcrops (which is common for many of the thin to medium limestone beds in the Bursum Formation in this area). It is underlain and overlain by purplish-gray to reddish-brown shale and sandy siltstone. In some areas this limestone is not present where crossbedded fluvial sandstones and pebble conglomerates up to a few meters thick fill channels over hundreds of meters wide or more. Yet we have identified a similar limestone bed more than four kilometers to the northeast of section 1 in Figure 1. At the localities where it is present, it contains at the top a remarkable abundance of small, triticitid fusulinids.

A laterally persistent interval of medium-bedded wackestone-packstone and lesser gray shaley interbeds and partings several meters thick, is present just below the middle of the section. It forms an extensive cuesta west of Cibola Spring, which is probably near Thompson's (1954) fusulinid sampling locality. Locally it is quite fossiliferous, containing a variety of macroinvertebrates (especially crinoid stems and brachiopods), as well as assemblages of lower Wolfcampian fusulinids. We refer to this as the "medial limestone." The lateral persistence, resistance to weathering and fossil content of this carbonate unit make it recognizable throughout the area. It probably corresponds to Thompson's (1954) "main limestone" of the Bursum in his descriptions of the area. It also forms prominent cuestas just to the south of Highway 60 near Abo Pass, some 30 km to the northeast of the study area. Both Stark and Dapples (1946) and Kottlowski and Stewart (1970) comment on the fact that macroinvertebrate shells in the interval are commonly partially replaced with red silica.

Another notable limestone bed, about 0.4 m thick, is present in the upper part of the section. It is typically a bioturbated dark gray skeletal wackestone, and overlies gray shale. The presence of yellowish-brown patches and stains on weathered outcrop commonly impart a nodular appearance to the limestone. The base of this limestone bed immediately overlies and incorporates, in its lower several centimeters, a remarkable abundance of nodules, each up to several centimeters in diameter, consisting in part of encrusting foraminifera and clotted (microbial) micrite. Terms such as "oncoids" or "algal biscuits" (which are restricted to demonstrably algal constructs by modern definition) do not strictly qualify for these nodular structures (Peter Scholle, pers. commun.); we refer to them here as "foram-microbial nodules."

Siliciclastic facies are generally discontinuous, commonly changing significantly over relatively short distances. For example, the triticitid-bearing limestone bed noted above near the base of the Bursum is present within a several-meter-thick interval of purplish to red shales at localities separated by over 3.5 km in both Cibola and Palo Duro Canyons. Details of the lithostratigraphic succession of the basal Bursum are similar at these two localities (Fig. 2). About midway between these localities, as noted above, crossbedded fluvial sand and pebble conglomerate fill channels that may have eroded the limestone in that area.

Within the Cibola Canyon area there is a distinctive, ~4-meter-thick sequence of red, micaceous, sandy siltstone that appears to be persistent laterally. This siliciclastic interval begins about 7 meters above the base of the formation at section 1. Another persistent siliciclastic interval consisting of up to 2 meters of cross-bedded, arkosic fluvial sandstone beds and intercalated siltstone and shaley beds and partings is present near the top of the Bursum in the area, above the foram-microbial nodule limestone and below the highest marine limestone beds in the

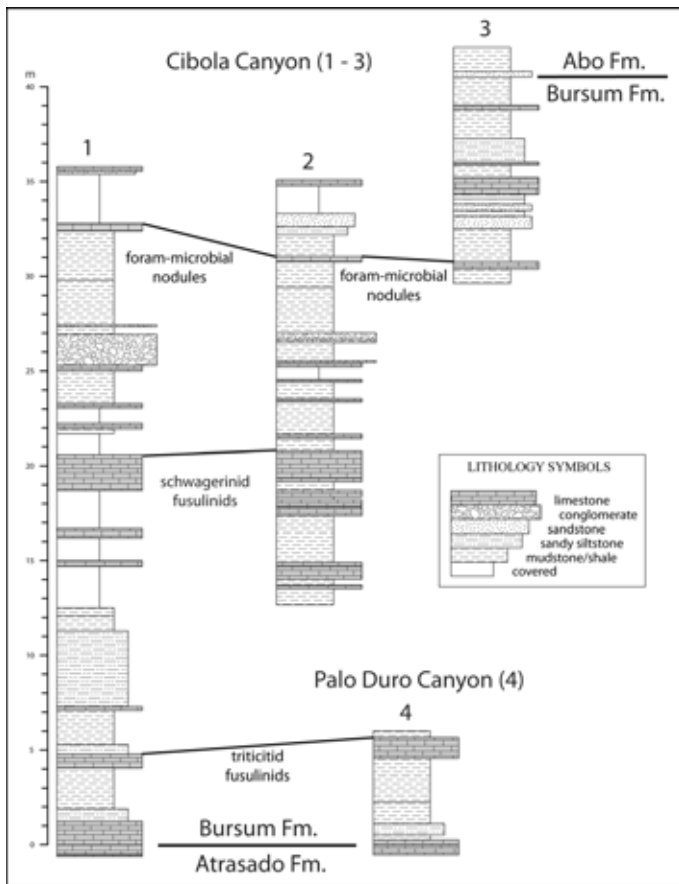


FIGURE 2. Generalized stratigraphic sections of the Bursum Formation near Cibola Spring. Tie lines indicate units that can be readily traced throughout the area. See the Appendix for coordinates of the localities.

Bursum. It forms a resistant cliff on steep hillslopes and can be traced on aerial photographs of the area.

DEPOSITIONAL ENVIRONMENTS

As noted above, Lucas and Krainer and their colleagues have made significant contributions to our understanding of the Bursum Formation in central New Mexico, including inferred depositional environments of the different lithofacies that are represented. Krainer and Lucas (2009, Fig. 3) indicate that, in general, purplish and reddish siliciclastic muds suggest deposition in fluvial-floodplain environments, and that gray calcareous shales were deposited in marine environments. cursory examination of a few of the gray siliciclastic muds in the Bursum revealed marine fossils including brachiopods, fusulinids and foram-microbial nodules; we did not observe marine invertebrates (or terrestrial plant fossils, for that matter) in the purplish or red-colored mudstones, although little time was expended in that effort. Fine-grained (micritic) carbonate nodules and stringers are present in many of the red-bed muds, which are generally regarded as pedogenic by Krainer and Lucas (2009). One of us (DWL) found scraps of bone, presumably from a terrestrial quadruped, in one of the fluvial sandstones interbedded with red-bed siliciclastic muds near the base of the section. In short, without taking the time to look carefully for fossil evidence, an initial outcrop-based assessment of depositional environment (marine vs. terrestrial) of the siliciclastic muds based on color seems reasonable. Nonetheless, difficulties in interpretation arise when one encounters thin to medium beds of limestone, some of which are clearly normal marine carbonates, intimately intercalated with purplish or reddish shales. Perhaps the large magnitude (100+ m), high frequency (Milankovitch-scale) shifts in sea level that have been documented for the Late Pennsylvanian-Early Permian (e.g., Rygel et al.,

2008) may explain some of these apparently abrupt lithologic shifts.

Similarly, it may be difficult to pinpoint the depositional environment (marine vs. terrestrial) for some of the limestone-clast conglomerates one encounters in the Bursum Formation. Krainer and Lucas (2009) discuss depositional models for marine vs. terrestrial limestone-clast conglomerates.

The *Triticites*-bearing limestone bed just above the base of the Bursum is of interest in part because it is commonly brecciated and is overlain and underlain by purplish-to-red siliciclastic muds that presumably represent deposition in a floodplain environment. The limestone itself is clearly marine. The breccia consists of the fossiliferous limestone, which is broken into small (mm- to cm-scale) interlocking angular fragments that are separated by films of siliciclastic mud similar in color to the surrounding muds. The degree of brecciation varies between localities. These observations suggest that the limestone was deposited in normal marine water, and was then exposed subaerially and subjected locally to desiccation shrinkage and brecciation.

FOSSILS

Marine strata in the Bursum Formation in the vicinity of Cibola Spring contain a variety of fossil invertebrates. Limestones are commonly wackestones or packstones and, in addition to the small skeletal fragments, contain an abundance of large macroinvertebrate fossils including brachiopods, bivalves, and echinoderms, with rare gastropods and corals. We have already mentioned the presence near the top of the sequence of a laterally extensive bed of well-developed foram-microbial nodules, the nuclei of which consist of brachiopod or mollusk shells. Among the microinvertebrate fauna, ostracodes and foraminifera (including fusulinids) are represented.

A few of the more common macroinvertebrate fossils from the medial limestone are pictured in Figure 3, along with foram-microbial nodules collected from the upper part of the Bursum Formation. A small collection of macroinvertebrates from the medial limestone has been obtained by the SNWR for public display in their visitor center, and have been identified by Barry Kues (University of New Mexico). More information about macroinvertebrates in the Bursum Formation at its type locality in Socorro County is presented in the paper by Kues (2002). Peter Scholle (NM Bureau of Geology and Mineral Resources) has collected, slabbed, and had thin sections made of the foram-microbial nodules, and is presently in the process of studying them. Toomey et al. (1988) present the results of their investigation of similar, so-called "algal biscuits" from Lower Permian limestones in southern Kansas-northern Oklahoma.

A preliminary examination of the fusulinids we collected from the Bursum supports previous reports (Thompson, 1954; Kottlowski and Stewart, 1970) regarding the assemblages and age assignments based on these assemblages. In particular, the presence of schwagerinid fusulinids in the medial limestone indicated to previous investigators an early Wolfcampian age assignment. As far as we can ascertain, these previous investigators did not examine the fusulinids in the limestone bed near the base of the Bursum. Detailed documentation of this assemblage will thus have to await examination by a specialist. Our preliminary observations, limited to examination of polished slabs of this limestone and a few thin sections, suggest that the assemblage contains species of small *Triticites*; schwagerinids were not observed. Photographs of fusulinids from the Bursum Formation in the vicinity of Cibola Spring are included in Figure 3.

DISCUSSION

The main focus of our work in the eastern part of the SNWR has been geological mapping. Here we have drawn attention to uppermost Pennsylvanian strata that are exposed over several square kilometers along the headwaters of Cibola and Palo Duro Canyons. Aside from the brief (but informative) references in the literature cited above, Bursum exposures at this site have received little documentation.

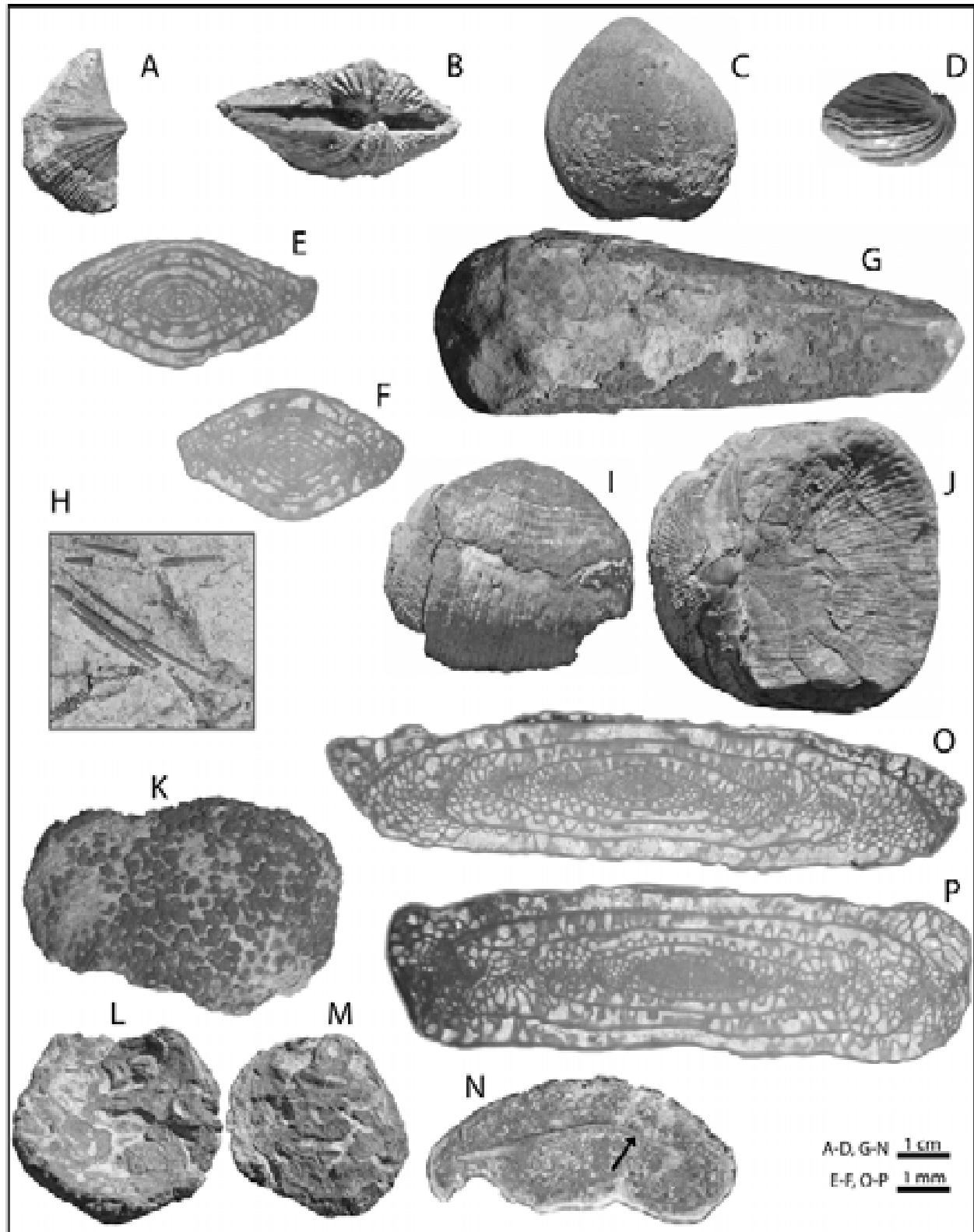


FIGURE 3. Selected marine fossils from the Bursum Formation near Cibola Spring. A-D and I-J are brachiopods. The *Triticitis* (E-F) are from near the base of the Bursum, the foram-microbial nodules (K-N) from near the top, and the other fossils are from limestone beds near the middle of the formation (see Fig. 2). A, *Neospirifer*, fragment of pedicle valve. B, *Neospirifer*, posterior view. C, *Composita*, pedicle valve. D, *Composita*, side view. E-F, *Triticitis*, axial sections of fusulinids from the base of the Bursum Formation. G, *Aviculopinna*, small fragment of the large conical bivalve. H, Echinoid spines. I, *Reticulatia*, fragment of pedicle valve. J, *Reticulatia*, oblique brachial view. K, Foram-microbial nodule, top view. L-M, Foram-microbial nodule, two interior halves that split along the surface of the fossil shell that forms the nucleus of the nodule, showing traces of what appear to be burrows or borings (the lighter colored areas). N, Foram-microbial nodule, vertical cut through the nodule shown in Fig. 3K, showing cross section of the nucleus (arrow), an invertebrate shell. O-P, *Schwagerina*, axial sections of fusulinids from the middle of the Bursum Formation.

The Bursum Formation in the vicinity of Cibola Spring was deposited along the north side of the "Joyita Hills basin," a narrow north-south oriented structural depocenter that existed during the ARM orogeny in Bursum time (Krainer and Lucas, 2009). The ~40 m thickness of the Bursum in this area is consistent with the Bursum isopach map for the region constructed by Krainer and Lucas (2009, Fig. 7). The site lies only 8 km to the north of the Sierra de la Cruz section measured by Krainer and Lucas (2009), where the Bursum is nearly twice as thick. It would be interesting to revisit the Sierra de la Cruz locality to conduct a comparison of the two stratigraphic successions. For example, are the two main fusulinid-bearing intervals near Cibola Spring also present in the Sierra de la Cruz sequence? Does the limestone associated with the foram-microbial nodules extend as far southward as Sierra de la Cruz? These three marker horizons at the Cibola Canyon site superficially appear to fall within the 3 depositional sequences (DS-1 to DS-3) proposed by Krainer and Lucas (2009) for the Sierra de la Cruz and neighboring localities. If reasonable correlations of marker beds between these two localities are apparent, examination of localities farther away in the depocenter of the Joyita Hills basin may be warranted. Although siliciclastic intervals are variable in terms of thickness and lithology over short lateral distances, it would be interesting to know how far distinctive marine carbonate beds can be traced.

We see little direct evidence for notable unconformities at the bottom or top of the Bursum in the sections examined during this study, such as obvious discordance of bedding or extensive basal conglomerates at the bases of the Bursum or the overlying Abo formations. This does not preclude the existence of unconformities between the formations. Indeed, Krainer and Lucas (2009), based on examination of outcrops over a widespread area, present strong arguments for Atrasado-Bursum and Bursum-Abo unconformities in central New Mexico.

A notable feature of the Atrasado-Bursum contact at a few localities in the Cibola Spring area is that the uppermost Atrasado limestone beds are silicified and overlying carbonate beds in the Bursum are not. For example, in the vicinity of section 4 (Fig. 2) highly silicified limestones in the upper Atrasado are directly overlain by up to 0.8 m of non-silicified, nodular limestone, which are in turn overlain by reddish, sandy siliciclastic mud clearly assignable to the Bursum Formation. It is not apparent to us what this local transition from silicified to non-silicified limestone implies, but it is tempting (as we have indicated in sections 1 and 4 of Figure 2) to include the overlying, nodular limestone within the Bursum, and to wonder how the silicification of the underlying Atrasado beds happened (and what amount of time was involved) if the abrupt lithologic change does, in fact, represent an hiatus.

The contact between the Bursum and Abo formations is exposed at section 3 (Fig. 2). Here a 0.25 m wackestone in the upper part of the Bursum is separated from overlying, brick-red sandstones and mudstones typical of the Abo Formation by 1.5 m of pale purplish gray mudstone. Although the color change from drab gray mudstone to typical Abo-red mudstone/sandstone is abrupt, there is no conglomerate at the base of the Abo at this locality.

Our investigation of the Bursum Formation near Cibola Spring identifies two zones with an abundance of fusulinids. One is near the base of the formation, and the other midway up in the section. These zones may correlate with other fusulinid-bearing limestones in the sections of the Bursum described by Krainer and Lucas (2009). Near Cibola Spring, *Triticites* are present in the lower zone, and the upper zone contains species of *Schwagerina* (in addition to *Triticites*). The fusulinids from the basal zone apparently have not been examined by specialists, but a preliminary supposition is that they may represent an upper Virgilian assemblage, whereas the schwagerinid-bearing beds above are lower Wolfcampian, as indicated in the previous studies cited above. Thus, the Bursum Formation in this area may represent deposition across the Virgilian-Wolfcampian boundary of the older literature.

CONCLUSIONS

About 40 m of mixed marine-terrestrial strata of the uppermost Pennsylvanian Bursum Formation are well exposed in the vicinity of Cibola Spring on the SNWR in central New Mexico. Marine deposits are fossiliferous, and some of the limestone intervals can be traced over the extent of the local outcrop belt, and perhaps over a much larger area. Two distinct fusulinid assemblages are present, one near the base and one near the middle of the section, which may be important for precise chronology of the sequence. Access to the area, which is administered by the U.S. Department of Fish and Wildlife, is facilitated by good roads, making this an attractive study site for detailed bio- and litho-stratigraphic investigation of the Bursum Formation.

ACKNOWLEDGMENTS

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APPENDIX

Map coordinates of measured sections depicted in Figure 2 (UTM meters, Zone 13S, horizontal datum NAD 83):

1. Base at 345068E, 3788551N. Top at 344950E, 3788324N.
2. Base at 344386E, 3789459N. Top at 344266E, 3789499N.
3. Base at 343578E, 3789635N. Top at 343610E, 3789568N.
4. This short section is exposed in the wall of the drainage at 345753E, 3792196N.

CONODONTS OF THE ATRASADO FORMATION (UPPERMOST MIDDLE TO UPPER PENNSYLVANIAN), CERROS DE AMADO REGION, CENTRAL NEW MEXICO, U.S.A.

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Abstract—A series of conodont faunas from the Atrasado Formation, Socorro County, New Mexico, permits biostratigraphic correlation of the eight members of the formation with the Midcontinent North America cyclothem successions. The Atrasado Formation is 290 m thick and composed of alternating siliciclastic sediments (shale, siltstone, arkosic sandstone) and different types of fossiliferous, partly cherty limestone. The Atrasado Formation is divided into the Bartolo, Amado, Tinajas, Council Springs, Burrego, Story, Del Cuerto and Moya members (in ascending order). The basal Bartolo Member is late Desmoinesian in age and the overlying Amado Member includes the Desmoinesian-Missourian boundary. The Tinajas Member ranges from the early Missourian to near the base of the Virgilian. The Council Spring, Burrego, Story, Del Cuerto and Moya members are early to middle Virgilian in age. Although some conodont faunas of the Atrasado Formation can be correlated approximately with individual Midcontinent cyclothem, direct matching of eustatic events cannot be demonstrated because of regional tectonic effects on the Atrasado lithologic succession.

INTRODUCTION

Conodont faunas are poorly known from the thick Middle and Upper Pennsylvanian sections in New Mexico, where sparse fusulinid collections have provided limited biostratigraphic and chronostratigraphic control. Lucas et al. (2009) first reported conodonts from the Pennsylvanian succession in the Cerros de Amado region of Socorro County, New Mexico, including sparse faunas from the upper Middle to Upper Pennsylvanian Atrasado Formation. Since that time, the Atrasado Formation at three of the original sections was resampled for conodonts (Cerros de Amado A, Cerros de Amado C, and Minas de Chupadera), and a fourth section of the Atrasado Formation was added (Ojo de Amado) (Fig. 1). Here we describe the distribution of conodonts in the Atrasado Formation and discuss the correlation of the lithostratigraphic members of the Atrasado to the conodont zonation of the cyclothem Middle and Upper Pennsylvanian succession in the Midcontinent region of North America (Barrick et al., 2004, 2013).

LITHOSTRATIGRAPHY AND SAMPLING

Lucas et al. (2009) provided a detailed review of the Pennsylvanian stratigraphy in the Cerros de Amado and vicinity (Fig. 1). They described the Atrasado Formation in this area as 290 m of interbedded siliciclastics (mostly shale and arkosic sandstone) and varied limestones divided into eight members (ascending order): Bartolo, Amado, Tinajas, Council Spring, Burrego, Story, Del Cuerto and Moya. The Atrasado Formation ranges from late Desmoinesian to Virgilian in age and overlies the Desmoinesian Gray Mesa Formation and is unconformably overlain by the lower Wolfcampian Bursum Formation.

Conodonts reported by Lucas et al. (2009) from the Atrasado Formation were obtained from sampling incomplete, but stratigraphically overlapping sections of the Atrasado Formation. In 2012, we resampled these sections and other strata of the Atrasado Formation in the Cerros de Amado. Our resampling focused on the four sections sampled by Lucas et al. (2009): Minas de Chupadera (Fig. 2), Cerros de Amado A (Fig. 3), Cerros de Amado B (Fig. 4) and Cerros de Amado C (Fig. 5). In resampling these we note the following:

1. Lucas et al. (2009, fig. 13) misinterpreted stratigraphic assignments in the lower part of their Minas de Chupadera section. Thus, the lower part of the section (Fig. 2, units 1-35) belongs to the Tinajas

Member, as shown here, not to the upper Bartolo, Amado and Tinajas members, as shown by Lucas et al. (2009).

2. The Cerros de Amado A section presented here (Fig. 3) is the same as presented by Lucas et al. (2009, fig. 11).

3. We added additional section to the Cerros de Amado B section of Lucas et al. (2009, fig. 12). Thus, units 47-90, strata of the Council Spring, Burrego and Story members of our section (Fig. 4), are newly measured and sampled.

4. Similarly, we added section to the Cerros de Amado C section of Lucas et al. (2009, fig. 8). Thus, units 178-184, strata of the Amado and Tinajas members of our section (Fig. 5), are newly measured and sampled.

We also sampled a section at Ojo de Amado (Fig. 6) not described by Lucas et al. (2009). The Ojo de Amado section is newly measured and sampled and will be described in detail elsewhere.

Approximately 80 different beds were sampled for conodonts from the five overlapping sections of the Atrasado Formation, with an average sample size of about 2-3 kg. We preferentially sampled carbonate packstone and grainstone beds, as well as thin packstone lenses in mudstone beds, because our experience has shown that these lithotypes are more likely to yield biostratigraphically useful conodont faunas from the New Mexico Pennsylvanian sections. However, where these lithotypes were not available or greater stratigraphic resolution was desired, carbonate mudstones were also sampled, but with limited success.

CONODONT FAUNAS

Bartolo Member

Few conodonts were recovered from the Bartolo Member at two sections, Cerros de Amado A and Cerros de Amado C (Figs. 3, 5). The age of the base of the Bartolo Member is constrained by the late Desmoinesian fauna obtained from the uppermost beds of the underlying Garcia Member of the Gray Mesa Formation at the Amado C section. A few meters below the top of the Garcia Member (unit 148), is an abundant and diverse fauna that includes *Swadelina neoshoensis* Lambert, Heckel and Barrick, 2003 and *Neognathodus roundyi* (Gunnell, 1931), as well as abundant *Idiognathodus* elements (Lucas et al., 2009). This fauna is characteristic of the late Desmoinesian *Sw. neoshoensis* Zone of the Midcontinent region, which represents the upper, but not uppermost portion of the Marmaton Group (Fig. 7; Barrick et al., 2004, 2013).

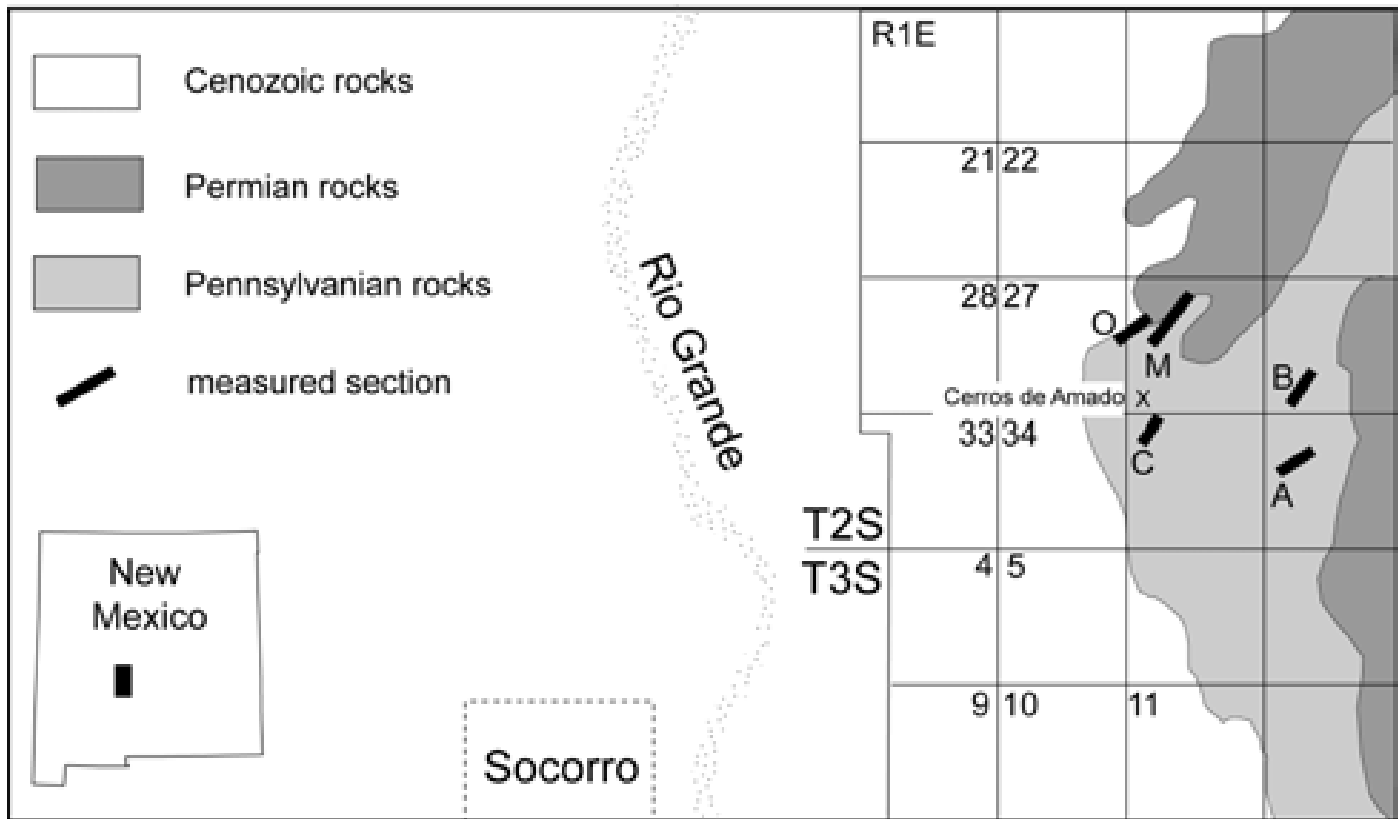


FIGURE 1. Simplified geologic map of the Cerros de Amado area, Socorro County, New Mexico (after Lucas et al., 2009), showing locations of measured sections of Pennsylvanian strata. Measured sections are: A, Cerros de Amado A (Fig. 3); B, Cerros de Amado B (Fig. 4); C, Cerros de Amado C (Fig. 5); M, Minas de Chupadera (Fig. 2); O, Ojo de Amado (Fig. 6). Grid shows legal land survey; numbered squares are sections, nominally one mile (1.6 km) on a side.

Thin limestone beds in the upper part of the Bartolo Member yielded mostly small faunas of *Adetognathus*, *Ellisonia*, and *Hindeodus* elements, all shallow-water forms. At the Amado C section, about 15 m below the top (unit 165), a few small *Idiognathodus* P₁ elements that resemble juvenile forms of *I. swadei* Rosscoe and Barrick, 2009a occur with rare *Neognathodus expansus* (Jones, 1941). At the Amado A section, a limestone about 33 m below the top (unit 13) produced *I. swadei*, *N. expansus*, and *N. roundyi*, and a *Swadelina* morphotype similar to *Sw. neoshoensis*. A higher limestone, 13 m below the top of the member (unit 22), yielded a slightly different association of *I. expansus* (Stauffer and Plummer, 1932) and *Sw. nodocarinata* (Jones, 1941).

All these species indicate a latest Desmoinesian age, but provide little additional time resolution. The fauna of the highest Midcontinent Desmoinesian cyclothem, the major Lost Branch cyclothem (Fig. 7), is the best known, and it is characterized by *Swadelina nodocarinata*, *Neognathodus expansus* and *N. roundyi*, as well as *Idiognathodus expansus* and *I. swadei* (Rosscoe, 2008). However, all of these species may range below the Lost Branch cyclothem into at least the underlying intermediate Norfleet cyclothem, if not lower (Barrick et al., 2004, 2013).

Amado Member

The Amado Member yielded a series of low abundance, low diversity shallow-water faunas that range in age from the latest Desmoinesian into the early Missourian based on *Idiognathodus* species. Elements of *Hindeodus*, *Adetognathus*, and less common *Ellisonia* occur in most samples. The base of the Amado at the Amado A section (unit 30) produced a small fauna of *I. expansus* and a few *Swadelina nodocarinata*, two species that became extinct near the end of the Desmoinesian. However, the basal beds of the Amado Member at the Amado C section (unit 169) produced mostly small elements of *I. swadei*, a species that ranges

from the late Desmoinesian into the early Missourian. At the Amado C section, *I. swadei* occurs with *I. harkeyi* Gunnell, 1933 in unit 171, an association known from as low as the latest Desmoinesian South Mound-Checkerboard minor cycle in the Midcontinent region (Fig. 7), although both species range higher and are more characteristic of the early Missourian (Rosscoe and Barrick, 2009a,b, 2013).

At both the Amado A and Amado C sections, samples from higher levels in the Amado Member yielded diagnostic early Missourian species. *Idiognathodus eccentricus* (Ellison, 1941), which first appears in the basal Missourian Exline intermediate cyclothem in the Midcontinent (Rosscoe et al., 2009a, b), occurs in the Amado A section, unit 44. Slightly higher, in unit 48, small morphotypes with advanced medial nodosity, *I. turbatus* Rosscoe and Barrick, 2009a, and reduced lobes, *I. corrugatus* Gunnell, 1933, appear. This association suggests a level equivalent to the major Hertha cyclothem. By the top of the Amado Member, units 54 and 56, a more abundant and diverse variety of *Idiognathodus* species are present (*I. turbatus*, *I. corrugatus*, and *I. gemmiformis* Gunnell, 1933). These species indicate an age equivalent to at least the Hertha cyclothem and perhaps even the overlying major Swope cyclothem (Fig. 7). Near the top of the Amado C section, *I. turbatus* is definitely present in unit 175, in addition to a number of smaller *Idiognathodus* elements difficult to assign to species.

The Amado Member appears to be a complex composite unit, comprising a series of thin limestone units and interbedded shale that correlates to parts of four or more Midcontinent cyclothem: the latest Desmoinesian Lost Branch and Checkerboard-South Mound cyclothem, and the earliest Missourian Exline, Hertha, and perhaps even the Swope cyclothem (Fig. 7). The Amado Member corresponds approximately to the Midcontinent Moscovian-Kasimovian Boundary Composite Sequence, as discussed in Rosscoe and Barrick (2013).

Minas de Chupadera

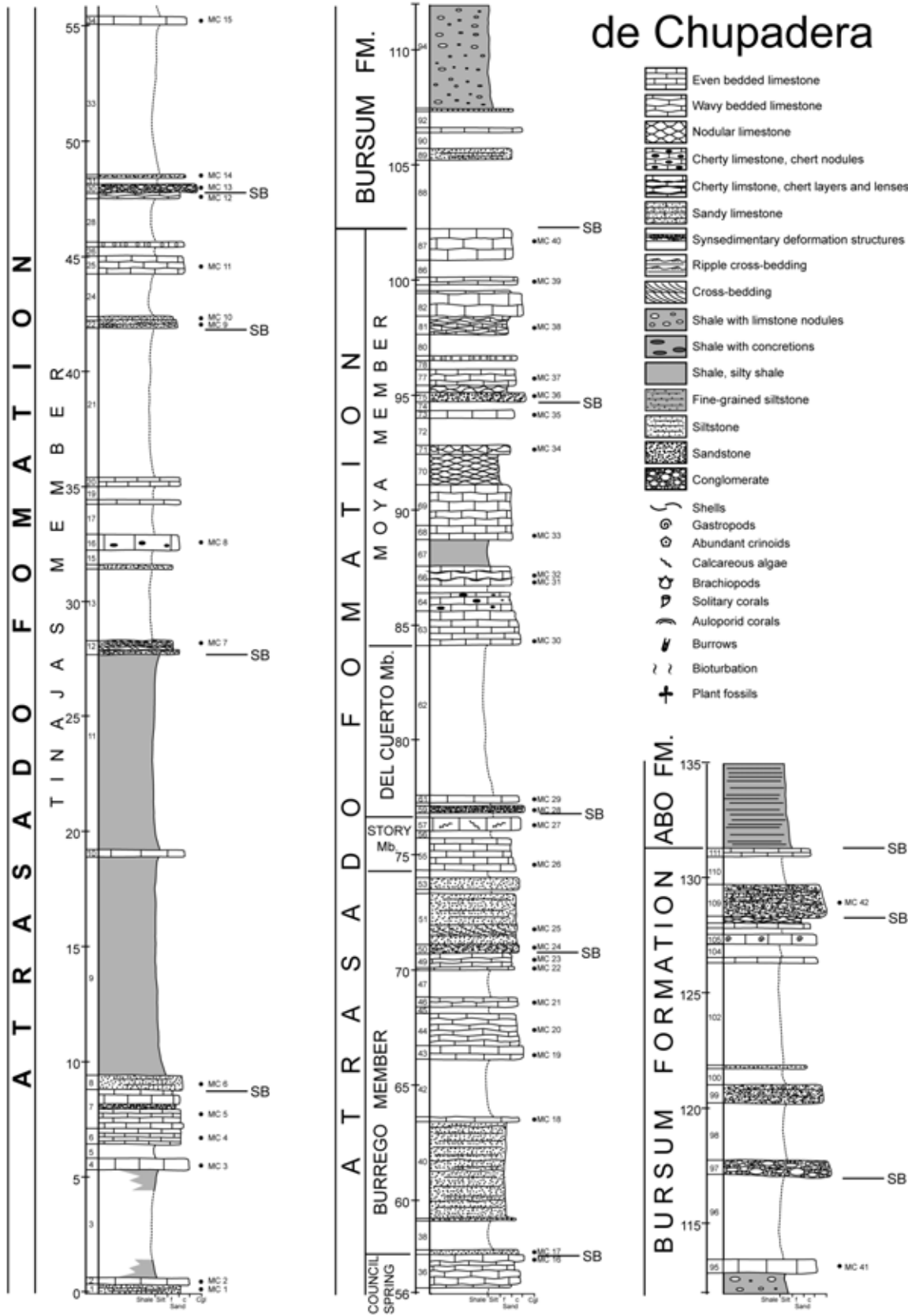


FIGURE 2. Measured section of part of the Atrasado Formation and the Bursum Formation at the Minas de Chupadera section. Section measured in the NW ¼ SW ¼ sec. 26, T2S, R1E. Conodont samples are referenced to the unit numbers shown on left of the lithologic column. Thin section levels are indicated by sample numbers to right of lithologic units (for example, “MC 18”).

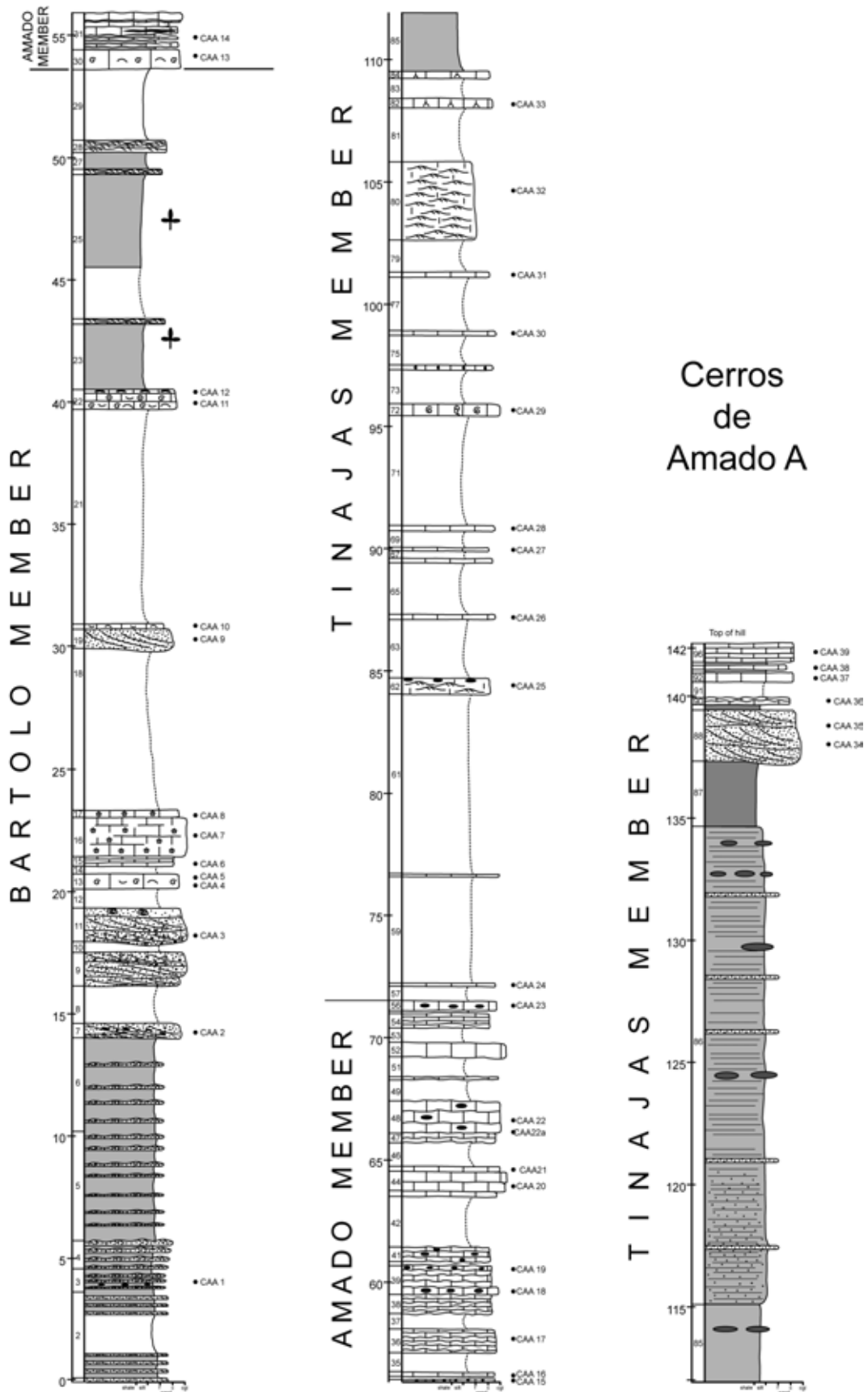


FIGURE 3. Measured section of part of the Atrasado Formation at the Cerros de Amado A section. Section measured in the S ½ NW ¼ sec. 36, T2S, R1E. For legend to lithologic symbols see Figure 2. Conodont samples are referenced to the unit numbers shown on left of the lithologic column. Thin section levels are indicated by sample numbers to right of lithologic units (for example, “CAA 24”).

Tinajas Member

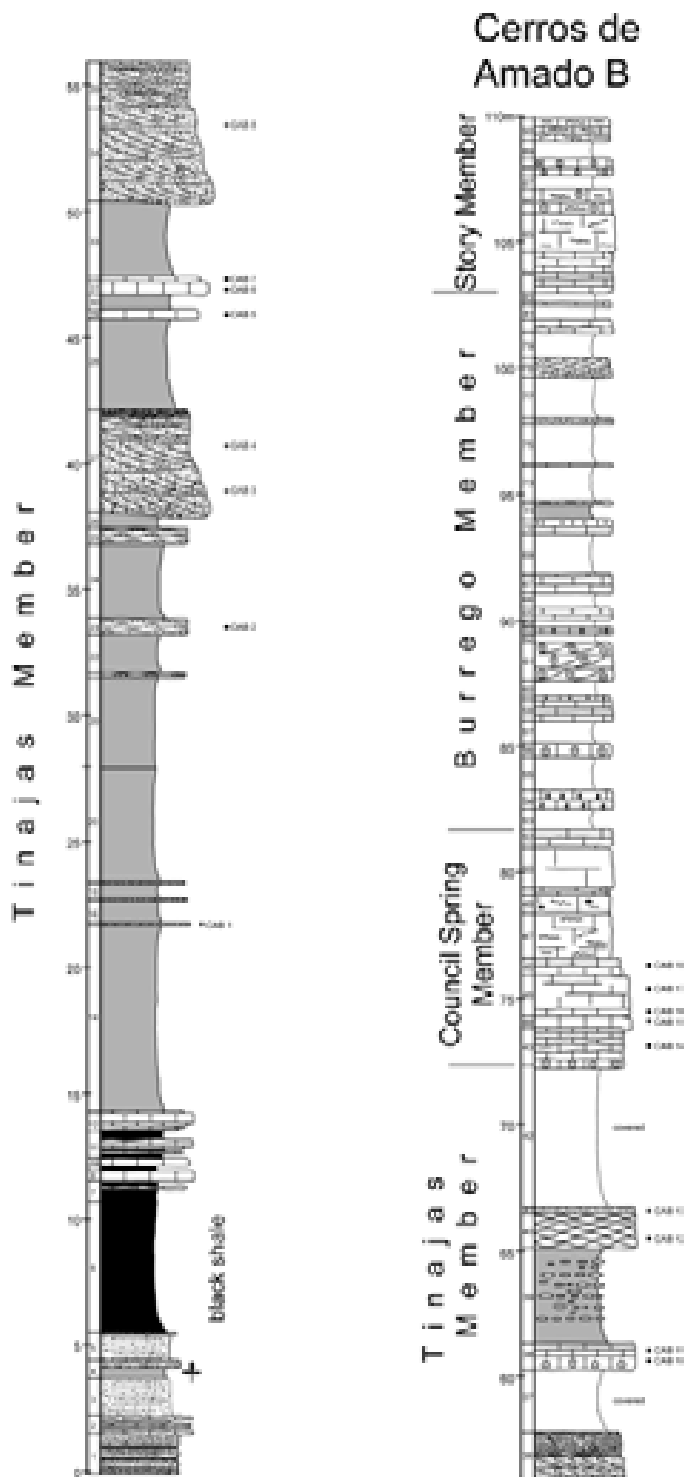


FIGURE 4. Measured section of part of the Atrasado Formation at the Cerros de Amado B section. Section measured in the SE $\frac{1}{4}$ SW $\frac{1}{4}$ sec. 25, T2S, R1E. For legend to lithologic symbols see Figure 2. Conodont samples are referenced to the unit numbers shown on left of the lithologic column. Thin section levels are indicated by sample numbers to right of lithologic units (for example, "CAB 12").

The few thin limestone beds in the lower to middle part of the Tinajas Member at the Amado A section yielded just a few undiagnostic conodonts. Near the base of the Tinajas Member at the Amado C section (unit 184) one sample produced a small fauna that contained large *Idiognathodus* P₁ elements that may be examples of early *I. magnificus* Stauffer and Plummer, 1932, and a small number of members of the *Streptognathodus gracilis* group. The *Streptognathodus gracilis* fauna is better developed at the Ojo de Amado section (Fig. 6), where it ranges through about 12 m of the Tinajas Member (units 7-27). Most of the P₁ elements can be assigned to *S. elegantulus* Stauffer and Plummer, 1932 (no nodes) and *S. gracilis* Stauffer and Plummer, 1932 (a single node on the caudal margin). A small number of P₁ elements of *S. corrugatus* Gunnell, 1933 (two or more caudal nodes) occur, and a few examples of *S. excelsus* Stauffer and Plummer, 1932 (caudal and rostral lobes) are present. The *S. gracilis* Zone in the Midcontinent region spans six intermediate to major cyclothem of the middle Missourian (Barrick et al., 2004, 2013). The abundant early forms of *I. magnificus* in the lower Tinajas Member at the Amado C section suggest a possible position low in the *gracilis* Zone for this sample.

The next highest diagnostic conodont fauna in the Tinajas Member lies over 15 m above the beds that bear the *S. gracilis* fauna in the Ojo de Amado section, and in the lower parts of the Cerros de Amado B and Minas de Chupadera sections. At the Ojo de Amado section, an abundant fauna that can be assigned to the middle Missourian *Idiognathodus eudoraensis* Zone was recovered from unit 64, and smaller faunas occur up into unit 68, through a thickness of about five meters. This fauna is restricted to a three-meter interval at the Amado B section, only a middle bed of which (unit 10) contains abundant conodonts. At the Minas de Chupadera section, an abundant basal bed (unit 2) is separated by a five-meter-thick clastic interval from an upper bed (unit 4) that yielded a much smaller fauna. Typical P₁ elements of *I. eudoraensis* Barrick, Heckel, and Boardman, 2008 occur at all three sections, but in different abundance relative to P₁ elements of the *Streptognathodus firmus* group.

In units 64 and 65 at the Ojo de Amado section, *I. eudoraensis* is subordinate to a variety of *Streptognathodus* morphotypes. The *Streptognathodus* morphotypes include P₁ elements that appear to represent the transition from the older Eurasian species *S. isakovae* Goreva and Alekseev, 2006 to *S. firmus* Kozitskaya, 1978 (in Kozitskaya et al., 1978). *Streptognathodus isakovae* possesses a slender P₁ element with an extremely long carina separated from the high margins (parapets) by deep adcarinal grooves. No lobes are present. The carina extends about two-thirds of the length of the platform, but two to four transverse ridges cross the dorsal end of the platform. In the younger species, *S. firmus*, the P₁ element is similar in shape, but the carina extends to near the dorsal end of the platform, either as a solid carina or as a carina and a few discreet nodes. A distinctive *Idiognathodus* P₁ element occurs in unit 64 that is absent or rare in other samples. This P₁ element (*Idiognathodus* sp. P, Fig. 9.19, 9.20) has a moderately long carina, about one-half of the platform length separated from the margins by shallow adcarinal grooves. The dorsal platform is flat, and 5 to 6 transverse ridges cross the platform, the most ventral of which show some medial nodosity. Lobes are absent, or a single node be present on the rostral or caudal sides, unlike the middle Missourian *S. oppletus* Ellison, 1941, which has small, but distinct caudal and rostral lobes. The Ojo de Amado specimens also resemble *S. pictus* Chernykh, 2002 from the late Kasimovian of the Urals, but *S. pictus* has a weakly developed caudal lobe.

In the higher beds at the Ojo de Amado section (unit 68) and at the Amado B and Minas de Chupadera sections, *Idiognathodus* elements dominate the fauna and in addition to *I. eudoraensis*, more characteristic *Idiognathodus* P₁ elements with complete transverse ridges, short carinas, and reduced lobes occur. The species assignment of these other *Idiognathodus* morphotypes is unclear, but they may include forms that have been assigned to *I. toretzianus* Kozitskaya, 1978 (in Kozitskaya et

Cerros de Amado C

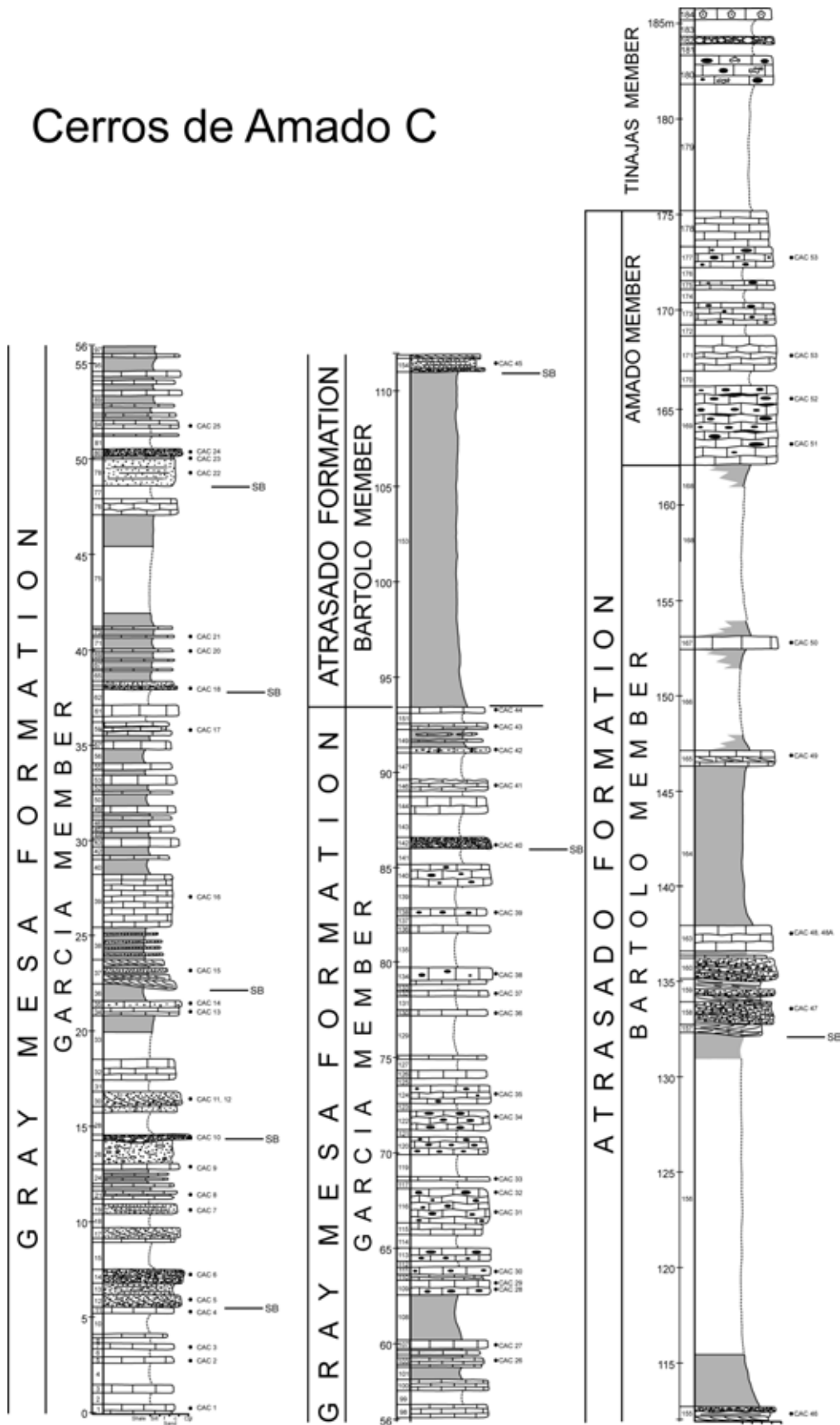


FIGURE 5. Measured section of parts of the Gray Mesa and the Atrasado formations at the Cerros de Amado C section. Section measured in the NW ¼ NW ¼ sec. 35, T2S, R1E. For legend to lithologic symbols see Figure 2. Conodont samples are referenced to the unit numbers shown on left of the lithologic column. Thin section levels are indicated by sample numbers to right of lithologic units (for example, “CAC 10”).

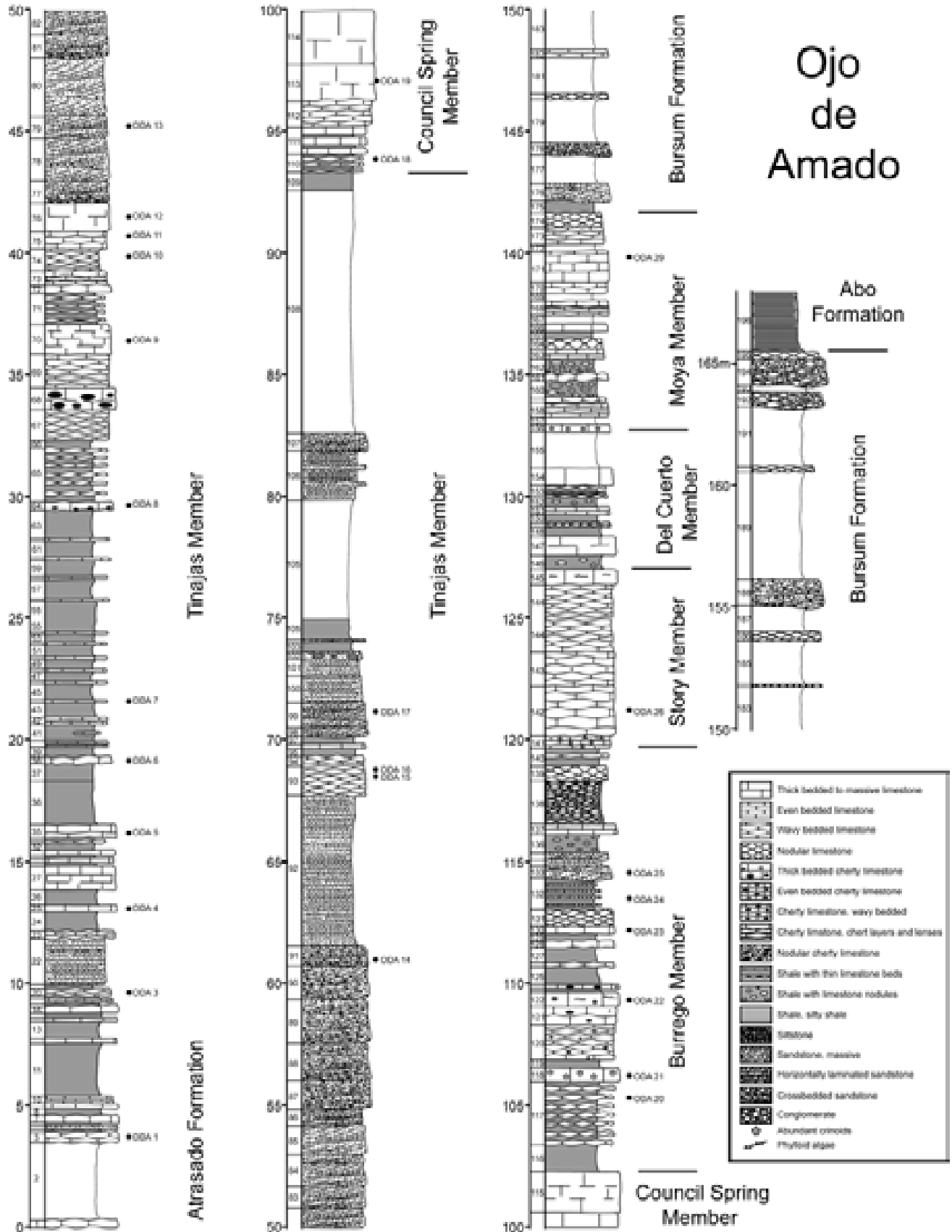


FIGURE 6. Measured section of part of the Atrasado Formation and the Bursum Formation at the Ojo de Amado section. Section measured in the NE ¼ sec. 27, and NW ¼ sec. 26, T2S, R1E. For legend to lithologic symbols see Figure 2. Conodont samples are referenced to the unit numbers shown on left of the lithologic column. Thin section levels are indicated by sample numbers to right of lithologic units (for example, “ODA 26”).

STAGE	CONODONT ZONE	CERROS DE AMADO	MIDCONTINENT CYCLOTHEM	KANSAS		
WOLFCAMP.	<i>S. binodosus</i>	BURSUM FORMATION	U. HUGHES CREEK	COUCIL GOVE GROUP (lower part)		
	<i>S. farmeri</i>		L. HUGHES CREEK AMERICUS			
	<i>S. flexuosus</i>		Five Point	ADMIRE GROUP		
	<i>S. bellus</i>		Falls City Brownville			
VIRGILIAN	GZHELIAN	ATRASADO FORMATION	MOYA MEMBER *	WABAUNSEE GROUP		
			<i>S. virgilicus (sensu stricto)</i>		DEL CUERTO MBR.	
			<i>S. vitali</i>		STORY MEMBER *	
			<i>I. simulator</i>	BURREGO MBR. *		
	KASIMOVIAN		<i>S. zethus</i>	COUNCIL SPRING MEMBER *	UPPER TOPEKA Curzon Hartford DEER CREEK Avoca	SHAWNEE GROUP
			<i>I. eudoraensis</i>	TINAJAS MEMBER *	LECOMPTON	
			<i>S. gracilis</i>	TINAJAS MEMBER *	Spring Creek Clay Creek OREAD	DOUGLAS GROUP
					Toronto CASS	
					latan	
				<i>I. confragus</i>	TINAJAS MEMBER *	South Bend STANTON
<i>I. cancellosus</i>	TINAJAS MEMBER *	Plattsburg Wyandotte IOLA DEWEY Cherryvale Hogshooter	KANSAS CITY GROUP			
<i>I. turbatus</i>		DENNIS				
<i>I. eccentricus</i>		AMADO LIMESTONE MEMBER *	SWOPE	BRONSON GROUP		
MOSCOWIAN	<i>Sw. nodocarınatus</i>	BARTOLO MEMBER *	HERTHA			
	<i>Sw. neoshoensis</i>	GARCIA FORMATION (upper part) *	Exline	PLEASANTON GROUP		
			Checkerboard LOST BRANCH Norfleet	MARMATON GROUP (upper part)		
DESMOINES.			ALTAMONT Farlington			

FIGURE 7. Correlation of members of the Atrasado Formation with the Midcontinent conodont zones (left) and the succession of Midcontinent cyclothem in Kansas (right). The asterisks indicate the levels of the most biostratigraphically significant conodont faunas in the Atrasado Formation. Numerous unconformities may occur within the members and between the members of the Atrasado Formation; see Figures 2-6. See text for discussion of the conodont-bearing levels.

al., 1978) in Eurasia or *I. "post-magnificus"* in the Midcontinent (Barrick et al. 2004, 2013). Most of the *Streptognathodus* P₁ elements present are close to typical *S. firmus*, and a small number of *S. gracilis* group P₁ elements were recovered.

The middle Tinajas *I. eudoraensis* fauna can be readily assigned to the lower part of the Midcontinent *eudoraensis* Zone (Fig. 7, Barrick et al., 2004, 2013), although the Tinajas fauna includes a greater variety of forms than have been reported from the Stanton cyclothem in the Midcontinent region (Ramirez, 2008). The association of species and morphotypes in unit 64 at Ojo de Amado suggests a position near the boundary of the upper part of the late Kasimovian *I. toretzianus* and *S. firmus* zones in Russia (Goreva and Alekseev, 2010). The older *S. isakovae* overlaps with *I. eudoraensis* in the upper part of the *toretzianus* Zone, just before the first occurrence of *S. firmus*. The overlying unit 68 at Ojo de Amado and faunas at the other sections where *S. firmus* is the typical morphotype, clearly represent the *firmus* Zone.

The upper part of the Tinajas Member is dominated by siliclastics, but interbedded are some limestone units from which small collections of *Streptognathodus pawhuskaensis* Harris and Hollingworth, 1933 were obtained. Lower samples tend to have morphotypes with longer carinas (Minas de Chupadera, unit 27; Amado B, unit 38) and higher samples yielded forms with the more typically short carinas, deep troughs and some accessory nodes (Minas de Chupadera, unit 34, Amado B units 40, 40A; Ojo de Amado, unit 93, middle). In the Midcontinent region, collections with *S. pawhuskaensis* morphotypes with longer carinas occur just above the *eudoraensis* fauna (South Bend cyclothem), whereas the more typical *S. pawhuskaensis*, including those with accessory nodes, are more characteristic of the latest Missourian Iatan and earliest Virgilian Cass cyclothem (Fig. 7).

Council Spring Member

Few conodont elements were recovered from the algal-dominated limestones of the Council Spring Member at two sections, Ojo de Amado and Amado B. The most common P₁ elements are those of *Idiognathodus* species with strongly reduced lobes. A few bear both small caudal and rostral lobes and are assigned to *I. toretzianus*, but most possess only a caudal lobe, which is characteristic of *I. lobulatus* Kozitskaya, 1978 (in Kozitskaya et al., 1978). The remainder of the fauna comprises a few *S. pawhuskaensis*, *Hindeodus*, and *Adetognathus* elements. The age of the Council Spring Member is thus weakly established by the poorly documented range of *I. lobulatus*, which appears in the early Virgilian/early Gzhelian *I. simulator* Zone in the Midcontinent (Oread cyclothem, Fig. 7; Barrick et al., 2004, 2013; Boardman et al., 2006) and in Russia (Goreva and Alekseev, 2010).

Burrego Member

In the Burrego Member at the Amado B (unit 56) and Minas de Chupadera (unit 48) sections, the sparse conodont faunas contain the first occurrences of *Streptognathodus ruzhencevi* (Kozur, 1976, in Kozur and Mostler, 1976) and *S. vitali* Chernykh, 2002. These two very similar species both possess a long carina and differ only in the relative length of the carina. In *S. ruzhencevi*, the carina extends nearly the full length of the platform, whereas in *S. vitali* the carina extends over half the length of the platform, but not to near the dorsal tip. The adcarinal grooves tend to be shallower, the dorsal medial groove narrower and shallower, and the transverse ridges on the margin broader in both species, compared with morphotypes of *S. pawhuskaensis*, with which they occur. A single specimen of *Idiognathodus simulator* (Ellison, 1941), diagnostic of the early Gzhelian, was recovered from unit 48 in the Minas de Chupadera section. *Idiognathodus tersus* Ellison, 1941 appears higher in the Burrego at the Amado B section (unit 63). Unlike most *Idiognathodus* species, *I. tersus* completely lacks lobes. A thin carbonate packstone in the lower part of the Burrego Member at the Ojo de Amado section (unit 118) provided a large *S. pawhuskaensis* fauna and just a few *S. ruzhencevi*/*S. vitalis* and *I. tersus* elements. *Streptognathodus ruzhencevi* and *S. vitali*

appear in the Queen Hill cyclothem in the Midcontinent region (Fig. 7; Boardman et al., 2006) and define the base of the *S. vitali* Zone in the Russian conodont succession (Chernykh, 2002; Goreva and Alekseev, 2010). *Idiognathodus tersus* and *I. simulator*, however, first occur in the older *simulator* Zone in Russia and the Midcontinent and range into the *vitali* Zone.

Story Member

Elements of *Hindeodus* and *Adetognathus* dominate faunas from the Story Member, but a smaller number of *Ellisonia* and *Streptognathodus* elements occur. *Streptognathodus* P₁ elements from the base of the Story Member at the Ojo de Amado section (unit 141) comprise the transition between *S. vitali* and *S. virgilicus* Ritter, 1995. Although both names are in common use, especially in Russia (e.g., Goreva and Alekseev, 2010), the exact line of distinction between the two species is not clear. Some specimens retain the longer carina characteristic of *S. vitali*, but in others the carina has shortened to less than one-half of the platform length and are here assigned to *S. virgilicus*, following Barrick et al. (2013). In addition to the shortened carina, *S. virgilicus* possesses broader margins with longer transverse ridges and a flatter dorsal platform with a narrow medial groove. A smaller number of P₁ elements of *S. pawhuskaensis* are also present.

The appearance of *Streptognathodus virgilicus* permits assignment of the Story to the Gzhelian *virgilicus* Zone, which follows the *vitali* Zone in the Russian conodont zonation. The level of the first occurrence of *S. virgilicus* in the Midcontinent succession is not well documented. Ritter (1995) reported the first occurrence of *S. virgilicus* in the Queen Hill cyclothem, but Barrick et al. (2013) indicate that the first occurrence of this species is in the intermediate Avoca cyclothem, just above the Queen Hill cyclothem (Fig. 7).

Del Cuerto Member

Small, undiagnostic faunas comprising *Hindeodus*, *Adetognathus*, *Ellisonia*, and a few tiny *Streptognathodus* elements were recovered from the Del Cuerto Member.

Moya Member

The few samples from the Moya Member at the Ojo de Amado and Minas de Chupadera sections yielded faunas that are characterized by abundant *Hindeodus*, common *Streptognathodus* morphotypes, and rarer *Adetognathus* and *Ellisonia* elements. The Moya fauna includes a mixture of *S. pawhuskaensis* and *S. virgilicus* group P₁ elements. The typical features of *S. virgilicus* are better developed in the Moya fauna than in the underlying Story fauna. The dorsal end of the platform is relatively low and the platform outline is broader. The groove is narrow, and the margins bearing transverse ridges are broader, especially the rostral margin, which makes the platform shape more curved and asymmetrical. P₁ elements with a more rounded dorsal end on which the outline part of the transverse ridges are reclined dorsally occur. These rounded specimens are likely examples of *S. holtensis* Ritter, 1994, but the Moya specimens tend to be more elongate than the material illustrated by Ritter (1994) from the Holt Shale (upper Topeka cyclothem) in Kansas. No age-diagnostic conodonts were obtained from the upper part of the Moya Member nor from the overlying Bursum Formation.

DISCUSSION

Correlation of the conodont faunas of the Atrasado Formation with the Midcontinent conodont zonation and to the succession of Midcontinent cyclothem (Fig. 7) must be approached with some caution. These correlations do not indicate that we have confidently identified in the Atrasado Formation the same eustatic events that produced any individual Midcontinent cyclothem. Instead, our correlations provide only the approximate time equivalence of the members of the Atrasado with the Midcontinent cyclothem, based on what is known about the ranges of the conodont species in the Midcontinent region.

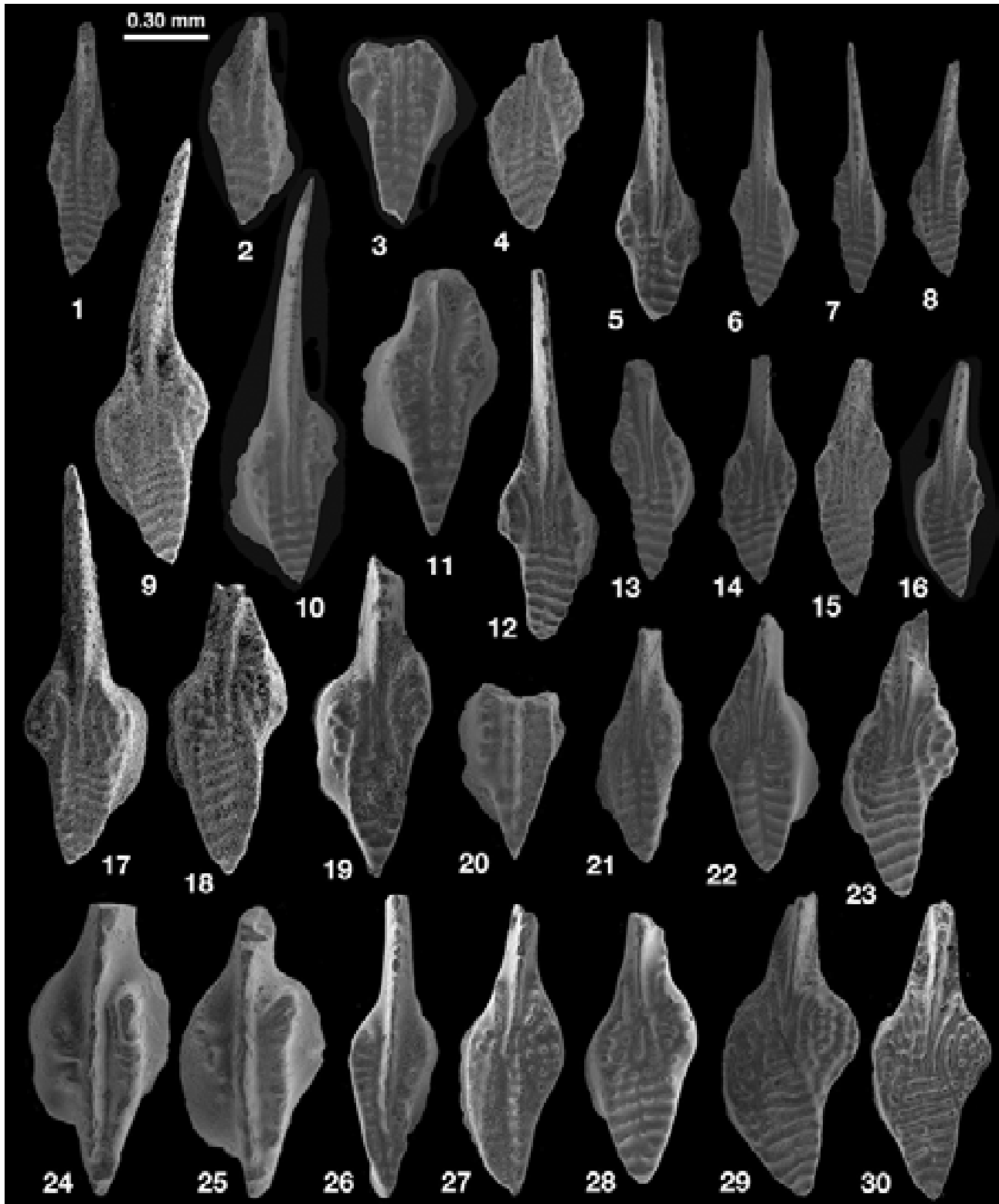


FIGURE 8. Conodonts from the Bartolo and Amadao Members of the Atrasado Formation. All illustrations are upper views of P_1 elements. Scale bar = 0.30 mm (X50). Specimens are repositied in the Department of Geosciences, Texas Tech University. **1**, *Idiognathodus gemmiformis* Gunnell, 1933?: Amado A 54. **2, 8**, *Idiognathodus corrugatus* Gunnell, 1933: **2**, Amado A 54; **8**, Amado A 48 middle. **3-5, 11**, *Idiognathodus turbatus* Rosscoe and Barrick, 2009: **3, 5**, Amado C 175; **4**, Amado A 54; **11**, Amado A 56 top. **6**, *Idiognathodus eccentricus* (Ellison, 1941): Amado A 44a. **7, 12**, *Idiognathodus sulciferus* Gunnell, 1933?: **7**, Amado A 44 base; **12**, Amado C 169. **9, 13-14, 23, 29-30**, *Idiognathodus swadei* Rosscoe and Barrick, 2009: **9**, Amado C 171; **13**, Amado A 44 base; **14**, Amado A 22; **23, 29-30**, Amado A 13. **15-16**, *Idiognathodus expansus* Stauffer and Plummer, 1932: Amado A 30. **17-19**, *Idiognathodus harkeyi* Gunnell, 1933: **17-18**, Amado C 171; **19**, Amado C 169. **20, 24-25**, *Neognathodus expansus* (Jones, 1941): **20**, Amado C 165; **24-25**, Amado A 13. **21-22**, *Swadelina nodocarinata* (Jones, 1941): **21**, Amado A 30; **22**, Amado A 22. **26**, *Neognathodus roundyi* (Gunnell, 1931): Amado A 13. **27-28**, *Swadelina neoshoensis* Lambert, Heckel, and Barrick, 2003: Amado A 13.

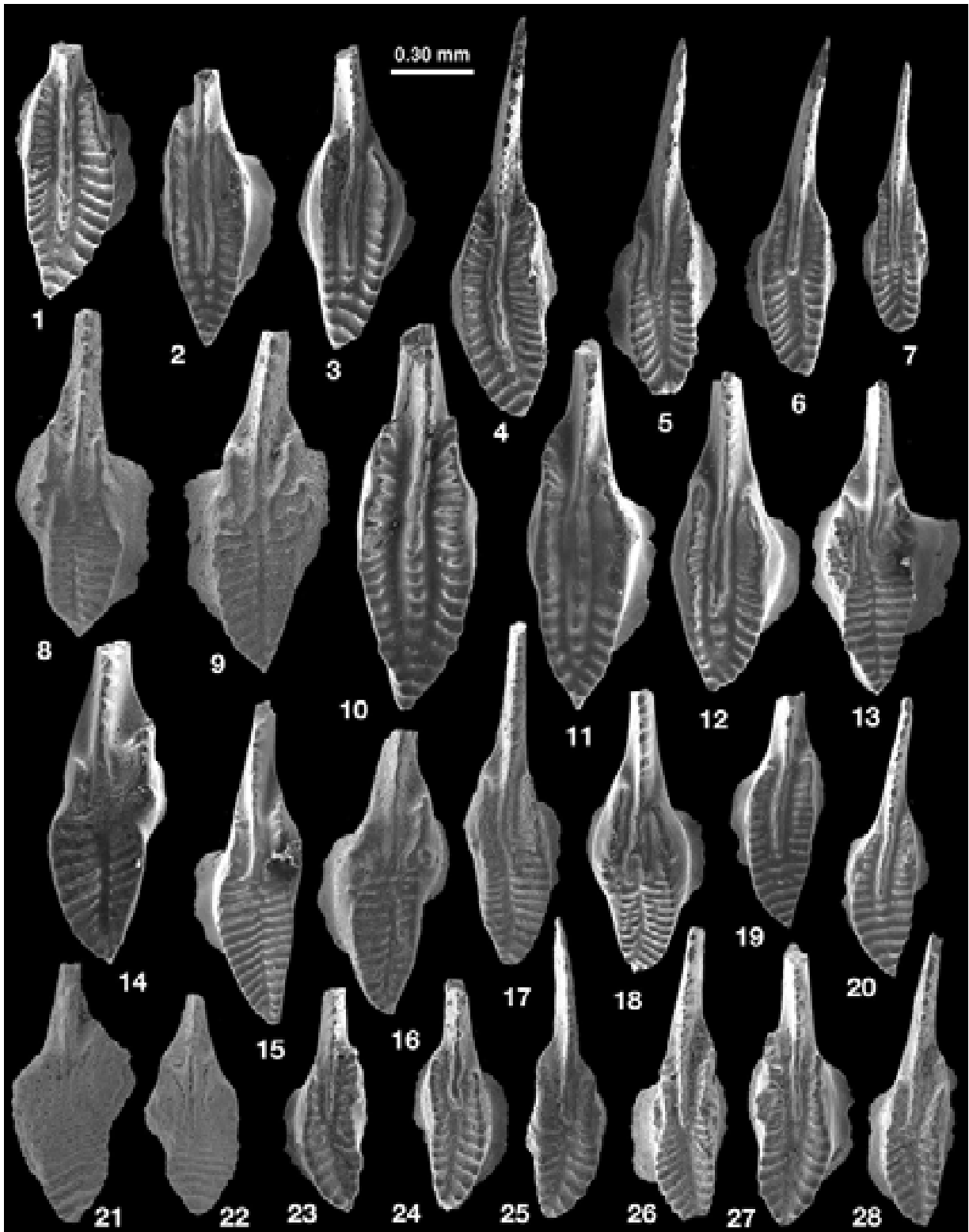


FIGURE 9. Conodonts from the Tinajas Member of the Atrasado Formation. All illustrations are upper views of P_1 elements. Scale bar = 0.30 mm (X50). Specimens are deposited in the Department of Geosciences, Texas Tech University. **1, 3-4**, Transitional forms between *Streptognathodus isakovae* Goreva and Alekseev, 2006 and *S. firmus* Kozitskaya, 1978: **1, 3**, Ojo de Amado 65; **4**, Ojo de Amado 64. **2, 10-12**, *Streptognathodus firmus* Kozitskaya, 1978: **2, 11, 12**, Ojo de Amado 64; **10**, Ojo de Amado 65. **8-9, 13-18**, *Idiognathodus eudoraensis* Barrick, Heckel and Boardman, 2008: **8-9**, Amado B 10; **13, 15, 18**, Ojo de Amado 65; **14, 17**, Ojo de Amado 64; **16**, Minas de Chupadera 2. **19-20**, *Idiognathodus* sp. P: Ojo de Amado 64. **21-22**, *Idiognathodus magnificus* Stauffer and Plummer, 1932?: Amado C 184. **23**, *Streptognathodus gracilis* Stauffer and Plummer, 1932: Ojo de Amado 7. **24, 28**, *Streptognathodus elegantulus* Stauffer and Plummer, 1932: **24**, Ojo de Amado 14; **28**, Ojo de Amado 14. **25-26**, *Streptognathodus excelsus* Stauffer and Plummer, 1932: Ojo de Amado 7. **27**, *Streptognathodus corrugatus* Gunnell, 1933: Ojo de Amado 14.

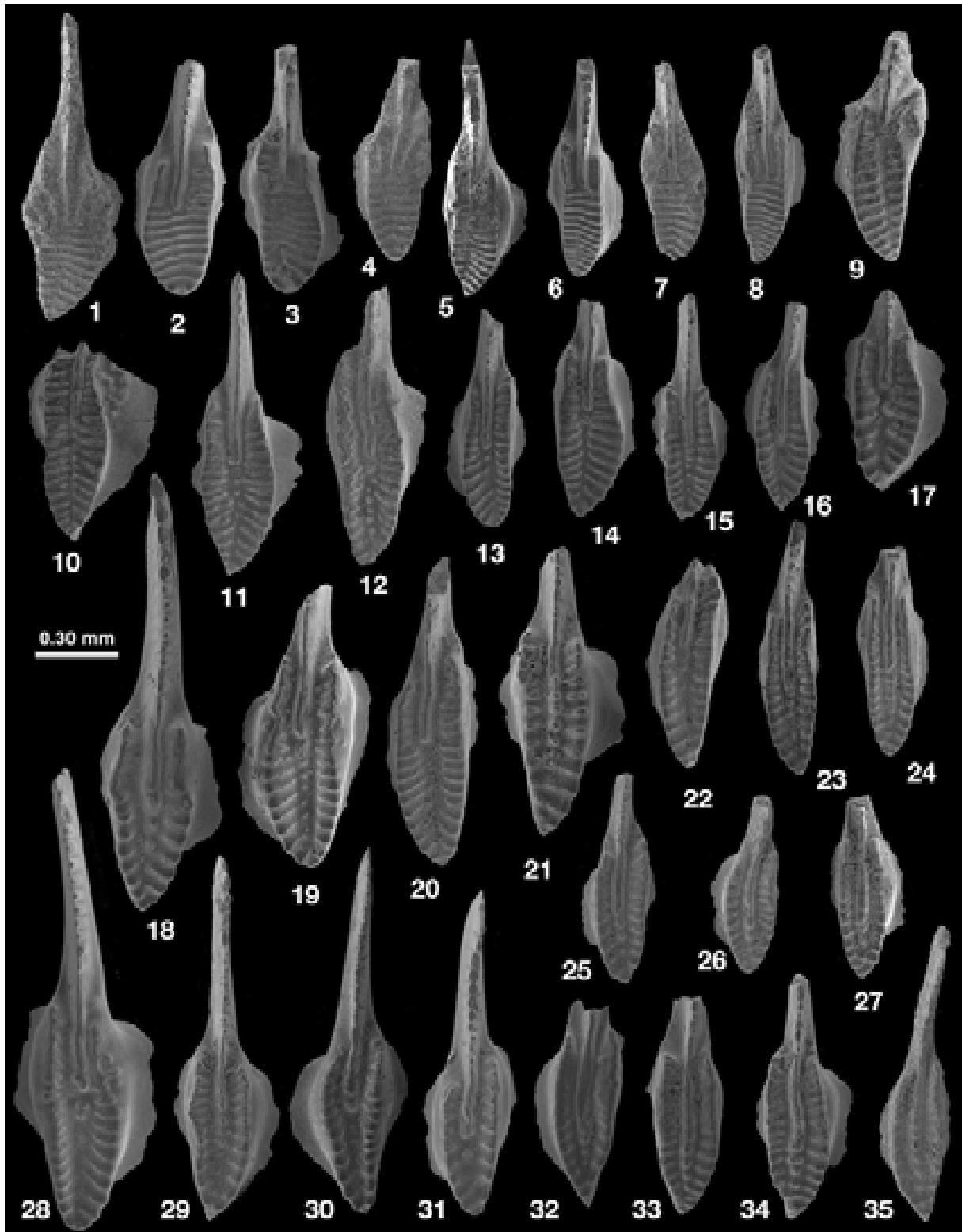


FIGURE 10. Conodonts from the Council Spring, Borrego, Story, and Moya members of the Atrasado Formation. All illustrations are upper views of P_1 elements. Scale bar = 0.30 mm (X50). Specimens are repositied in the Department of Geosciences, Texas Tech University. **1, 4**, *Idiognathodus toretzianus* Kozitskaya, 1978: **1**, Amado B 46 base; **4**, Amado B 44. **2-3, 6**, *Idiognathodus tersus* Ellison, 1941: **2, 6**, Minas de Chupadera 55; **3**, Ojo de Amado 118. **5, 7-8**, *Idiognathodus lobulatus* Kozitskaya, 1978: **5**, Ojo de Amado 110; **7**, Ojo de Amado 111; **8**, Amado B 63. **9**, *Idiognathodus simulator* (Ellison, 1941): Minas de Chupadera 48. **10-12, 15-16**, *Streptognathodus virgilicus* Ritter, 1995: **10-12**, Minas de Chupadera 79; **15-16**, Ojo de Amado 159. **13-14, 17**, *Streptognathodus holtensis* Ritter, 1994?: **13, 17**, Ojo de Amado 159; **14**, Minas de Chupadera 79. **18-20, 22, 28-31**, *Streptognathodus pawhuskaensis* Harris and Hollingsworth, 1933: **18, 28**, Minas de Chupadera 34; **20, 22**, Ojo de Amado 140; **29, 31**, Ojo de Amado 117; **30**, Minas de Chupadera 27. **21, 23-26, 32**, *Streptognathodus vitali* Chernykh, 2002: **21**, Ojo de Amado 118; **23-24**, Ojo de Amado 141; **25**, Ojo de Amado 111; **26**, Amado B 54; **32**, Amado B 63. **27, 33-35**, *Streptognathodus ruzhencevi* Kozur, 1976: **27**, Minas de Chupadera 48; **33-34**, Amado B 63; **35**, Amado B 56.

The Midcontinent conodont zonation (Barrick et al., 2004, 2013) is based on the succession of conodont species and morphotypes in the core shales (transgressive and lower highstand system tracts) of successive cyclothems. The conodont fauna of each core shale accumulated during a relatively small interval of the cycle and formed in response to the environmental conditions specific to the core shale in the Midcontinent region, most often the black shale facies. The intermediate and major cyclothems likely represent short time intervals, approximately 100,000-year cycles of eustatic sea-level rise and fall, but the conodont species do not range through the entire cyclothem. The conodont zones, however, tend to coincide with the 400,000-year groupings (Heckel, 2006). The first and last occurrences of many index species correspond primarily with the major cyclothems, whereas associated minor and many intermediate cyclothems contain less diagnostic conodont faunas. The discontinuous distribution of abundant conodont-bearing beds in the Midcontinent makes it difficult to determine the full ranges of species.

The success of the Midcontinent conodont zonation in correlating strata from the Midcontinent region into the Illinois Basin and the Appalachian Basins, as well as north-central Texas (Barrick et al., 2004, 2013) depends largely on the dominance of the eustatic control on depositional patterns in these regions. Core shales in each region formed in response to the same eustatic event and were populated by similar conodont faunas, permitting ready correlation of the eustatic event across all basins. The emphasis on identification of the few index taxa used to correlate the core shales has obscured significant differences in faunas within individual cyclothems across the outcrop belt within one basin (e.g.,

Roscoe and Bader, 2010), and across basins (Heckel et al., 2012). The model of “digital correlation” for global correlation (Heckel et al., 2007) relies heavily on the recognition of index taxa, often rare outside of the Midcontinent region, to identify major eustatic events and then matching of adjacent cycles to the Midcontinent cyclothem succession.

The Pennsylvanian succession in New Mexico comprises a more complex pattern of cyclical deposition where tectonic events associated with the Ancestral Rocky Mountains superimposed an independent series of local sea-level events on the eustatic sea-level pattern. As a consequence, one cannot easily ascertain that any observed stratigraphic cycle is eustatic or tectonic without extensive mapping of facies patterns, physical correlation of cycles, and local biostratigraphic control. Even so, the demonstration that a major stratigraphic cycle can be correlated across one mountain range and even into adjacent ranges, does not entail an eustatic origin for the cycle. Regional tectonic events may generate areally extensive, short-lived deepening of basins that produce regional “cyclothems,” as well as uplifts that prevent the deposition of major eustatic cyclothems. At this point in time, we do not possess sufficient information to distinguish eustatic sea-level events from tectonic sea-level events in the Cerro de Amado region.

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THE GALLINA WELL LOCALITY, AN EARLY PERMIAN (MIDDLE WOLFCAMPIAN) VERTEBRATE FOSSIL SITE IN SOCORRO COUNTY, NEW MEXICO

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Abstract—The Gallina Well locality is a Lower Permian vertebrate body and ichnofossil site located approximately 20 km northeast of Socorro, New Mexico in the Joyita uplift. The locality is situated stratigraphically low in the Scholle Member of the Abo Formation and is Coyotean in age. The fossiliferous beds are reddish-brown, fluviially-deposited, calcrete-pebble conglomerate and mudrock. Vertebrate body fossils from the site include paleoniscoid fish; the temnospondyl amphibians *Eryops* sp., *Trimerorhachis* sp., *Platyhystrix* sp., and *Zatrachys* sp.; a skull fragment of the lepospondyl *Diplocaulus* sp.; postcranial of the diadectid *Diadectes* sp.; a captorhinid skull and postcranial skeleton; and specimens of the eupelycosaur *Ophiacodon* sp., *Sphenacodon* sp. and *Dimetrodon* sp. The coprolite ichno-assemblage includes *Dakryonocopros arroyoensis*, *Alococopros triassicus*, *Heteropolacopros texaniensis* and amorphous coprolites. The Gallina Well locality yields the most diverse and extensive vertebrate body fossil and coprolite assemblage of Early Permian age known from southern New Mexico. Its basic composition differs little from the pelycosaur-dominated assemblages found to the north, indicating some uniformity of the Coyotean vertebrate fauna across New Mexico.

INTRODUCTION

New Mexico has an extensive terrestrial vertebrate body and ichnofossil record from the Lower Permian, accessible by means of the numerous outcrops of sandstone- and mudstone-dominated red beds throughout most of the state. Tracks and trackways are nearly ubiquitous, though vertebrate body fossils are seemingly more common to the north, with pelycosaur-dominated assemblages from the Jemez Mountains, Arroyo del Agua area and El Cobre Canyon (e.g., Berman, 1993; Lucas et al., 2005a, b, 2010, 2011, 2012a).

Approximately 20 km northeast of Socorro, in the Joyita uplift, the Gallina Well locality yields the most diverse and extensive vertebrate body fossil and coprolite assemblage of Early Permian age known in southern New Mexico. First discovered by one of us (DSB) in 1980, the site has produced an assemblage of vertebrate animals nearly as diverse as any locality of Early Permian age from northern New Mexico (Fig. 1). However, few of the vertebrate fossils from the Gallina Well Locality have been formally documented in print, and no analysis or correlation of the total vertebrate assemblage has previously been made. Here we examine and document the totality of vertebrate remains from the Gallina Well locality in the collections of the New Mexico Museum of Natural History and Science and the Carnegie Museum of Natural History.

Institutional abbreviations: AMNH, American Museum of Natural History, NY; CM, Carnegie Museum of Natural History, Pittsburgh, PA; NMMNH, New Mexico Museum of Natural History and Science, Albuquerque, NM; UCMP, University of California Museum of Paleontology, Berkeley, CA.

GEOLOGY AND STRATIGRAPHIC CONTEXT

In the Joyita uplift area of Socorro County, New Mexico, the Abo Formation is well exposed, up to 300 m thick and consists of two members. The lower, Scholle Member is characterized by slope-forming mudrocks with interbedded pebble conglomerates and sandstones. The upper, Cañon de Espinosa Member is sandstone dominated with lesser mudrocks. The Gallina Well locality is situated stratigraphically low in the Lower Permian Scholle Member of the Abo Formation (Fig. 1). It sits 46 m above the underlying Bursum Formation, which yields fusulinids that are early Wolfcampian (Newwellian) in age (Spielmann et al., 2009;

Kraimer and Lucas, 2009; Lucas et al., 2009). The fossiliferous beds are reddish-brown, fluviially-deposited, calcrete-pebble conglomerates (Spielmann et al., 2009) and mudrock. The coprolite ichnoassemblage from Gallina Well is comparable to that known from Early Permian fossil sites in the Chama Basin of north-central New Mexico and from Texas (Cantrell et al., 2012). The body fossil assemblage is very similar to that from the Early Permian vertebrate fossil sites in central and northern New Mexico.

VERTEBRATE PALEONTOLOGY

Taxa reviewed here from the Gallina Well site include a paleoniscoid fish, the temnospondyl amphibians *Eryops* sp., *Trimerorhachis* sp., *Platyhystrix* sp., and *Zatrachys* sp., the lepospondyl *Diplocaulus* sp., the diadectomorph *Diadectes* sp., a captorhinid, and the eupelycosaur *Ophiacodon* sp., *Sphenacodon* sp. and *Dimetrodon* sp. In describing the state-wide distributions of vertebrates from the Lower Permian Abo Formation in New Mexico, the locality designations given to the outcrop areas of their occurrences follow those given by Lucas et al. in this volume: (1) Jemez Mountains; (2) Abo Pass, which includes the Los Pinos Mountains locality of Berman (1993); (3) the Fra Cristobal/Caballo mountains, limited to just the Caballo Mountains by Berman (1993); and (4) Lucero uplift, but limited here to only the Upper Pennsylvanian Red Tanks Member of the Bursum Formation. This list is expanded to include four additional localities: (1) El Cobre Canyon and (2) the Arroyo del Agua area of the Chama Basin, Rio Arriba County, north-central New Mexico. In El Cobre Canyon the fossil-bearing levels include the lower El Cobre Canyon Formation and the overlying Arroyo del Agua Formation of the Cutler Group (Lucas et al., 2005b), whereas in the Arroyo del Agua area locality all the fossil-bearing levels except one are in the El Cobre Canyon Formation (Lucas et al., 2005a, b). It is important to point out, however, that in El Cobre Canyon the Pennsylvanian-Permian boundary is stratigraphically high in the El Cobre Canyon Formation, essentially restricting the Upper Pennsylvanian to the underlying El Cobre Canyon Formation and the Lower Permian to the overlying Arroyo del Agua Formation, whereas in the area of Arroyo del Agua the exposed part of the El Cobre Canyon Formation is assessed as Lower Permian (Lucas et al., 2005a, b); (3) Lower Permian Sangre de Cristo Formation of north-central New Mexico at scattered sites along the

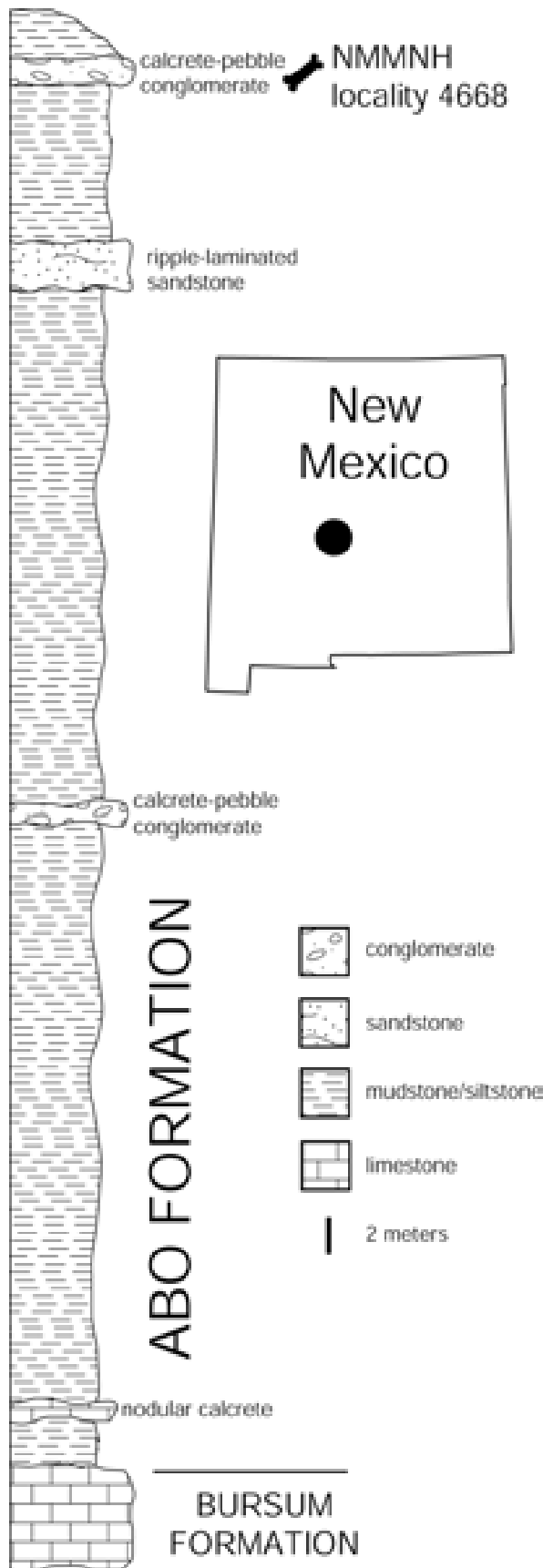


FIGURE 1. Stratigraphic section showing the location within New Mexico and the position of the Gallina Well locality (NMMNH locality 4668) in the lower portion of the Abo Formation (from Spielmann et al., 2009).

Pecos River valley drainage from Glorieta Pass at Glorieta on the eastern border of Santa Fe County southeastward to about 8 km south of Ribera, San Miguel County. These exposures are being designated as the Pecos River valley locality, following Berman (1993); and (4) scattered sites in the Lower Permian Abo Formation near Tularosa, Otero County, along the western escarpment of the Sacramento Mountains and referred to here as the Tularosa locality (Vaughn, 1969). The Gallina Well site lies in the Abo Formation outcrop in an area designated here as the Cerros de Amado-Joyita Hills locality by Lucas et al. (this volume), which was referred to by Berman (1993) as the Socorro locality.

Class Osteichthyes
Subclass Actinopterygii
Infraclass Actinopteri
Order Paleonisciformes
Family indet.

Referred specimens: CM 87689, portion of fin ray.

Description: A small patch of fin ray, about 1 cm², consisting of in place lepidotrichia (Berman, 1993).

Discussion: CM 87689 was eroded free from an unknown level, but obviously originated from the mudrock-sediment exposure below the calcrete-pebble conglomerate that nearly caps the Gallina Well section (Spielmann et al., 2009). Paleoniscoids are rare in the Lower Permian of New Mexico and recorded from only two other localities: a skeleton of a small paleoniscoid, seemingly related to *Progyrolepis*, from the Upper Pennsylvanian/Lower Permian Arroyo del Agua locality (Langston, 1953), and numerous skeletal fragments (teeth, scales, denticulated elements of the jaw and palate) from the Upper Pennsylvanian Lucero uplift locality (Harris et al., 2004).

Class Amphibia
Amphibia indet.
Fig. 2A-B

Referred specimen: NMMNH P-66814, right premaxilla fragment (Fig. 2A-B).

Description: NMMNH P-66814 (Fig. 2A-B) is a small anterior skull fragment that consists of the right premaxilla with ~12 marginal teeth. The fragment also contains the anterior margin of a circular nasal opening. The premaxilla is ornamented with polygonal pitting.

Discussion: The polygonal pitting is consistent with other contemporaneous amphibians, however, the relatively small and incomplete fragment precludes a more definitive assignment.

Order Temnospondyli
Family Eryopidae
Genus *Eryops*
***Eryops* sp.**
Fig. 2C-H

Referred specimens: NMMNH P-35703 (Fig. 2E-H), NMMNH P-65401 (Fig. 2C-D), two intercentra.

Description: These intercentra (Fig. 2C-D) are severely weathered and of two distinct sizes, suggesting they are either from two different individuals or from different areas of the axial skeleton. The larger of the two, NMMNH P-35703 (Fig. 2E-H), has a square ventral margin.

Discussion: *Eryops* is a common element of vertebrate assemblages of Lower Permian deposits, widely distributed across North America, that often include superbly preserved specimens. Occurrences of *Eryops* in New Mexico, however, are typically limited to single elements, fragmentary remains or partial, articulated postcranial skeletons and only rarely skulls, which have been recorded from the Upper Pennsylvanian, as well as the Lower Permian: the Lower Permian localities of the Jemez Mountains, Arroyo del Agua, and Pecos River Valley; and the Upper Pennsylvanian Lucero uplift locality. Most notable specimens

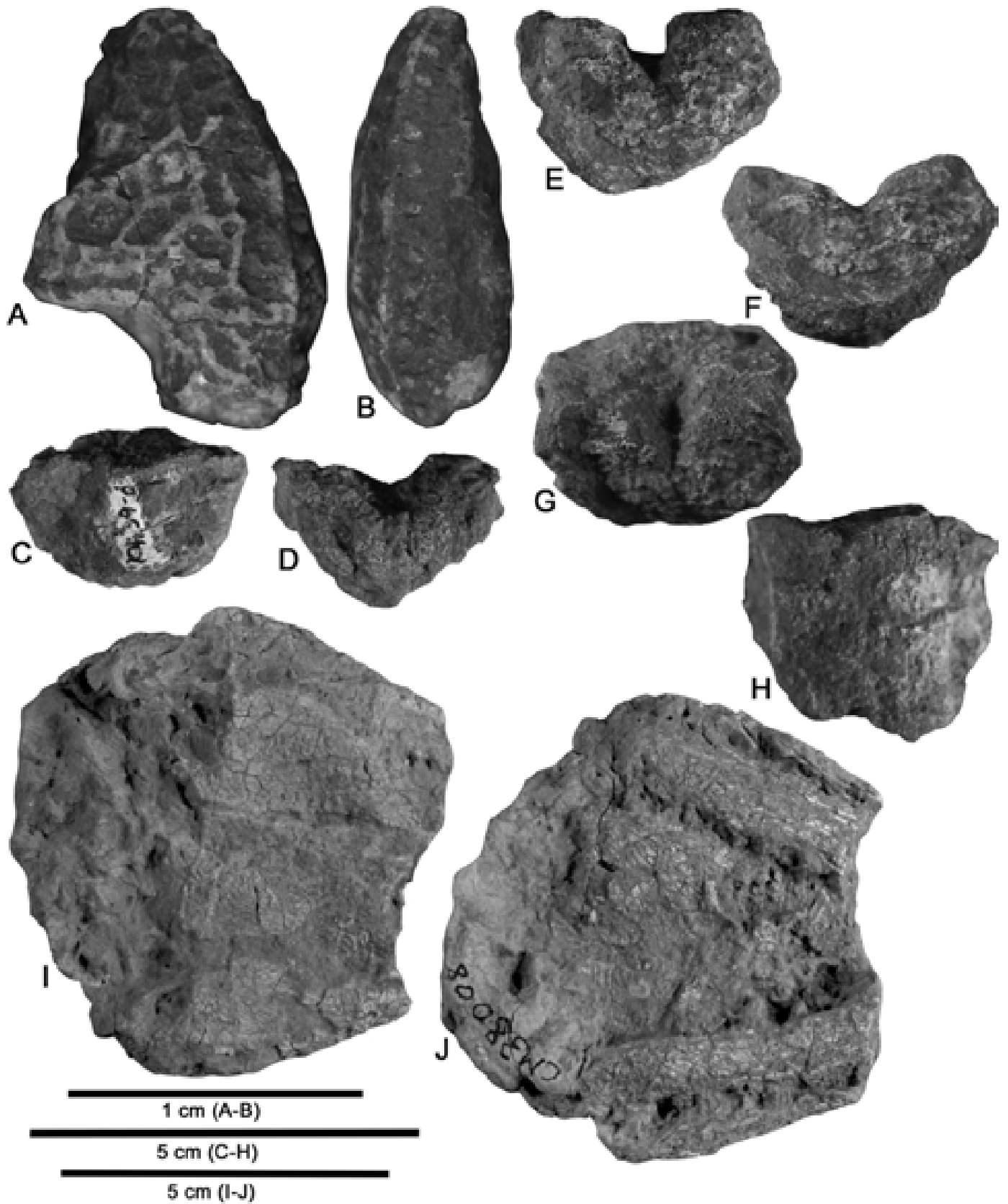


FIGURE 2. **A-B**, Amphibia indet., NMMNH P-66814, right premaxilla in **A**, dorsal and **B**, lateral views. **C-D**, *Eryops* sp., NMMNH P-65401, intercentrum in **C**, ventral and **D**, posterior views. **E-H**, *Eryops* sp., NMMNH P-35703, intercentrum in **E**, anterior, **F**, posterior, **G**, dorsal and **H**, ventral views. **I-J**, *Trimerorhachis* cf. *T. sandovalensis*, CM 38008, incomplete skull in **I**, dorsal and **J**, palatal views.

are two, nearly complete, superbly preserved skulls, one including the right mandible, and the disarticulated appendicular elements (CM 34906) from the Arroyo del Agua locality, and the other (NMMNH P-46379) from the Upper Pennsylvanian El Cobre Canyon locality, which documents the only Pennsylvanian occurrence of *Eryops* (Werneburg et al., 2010).

Family Trimerorhachidae
Genus Trimerorhachis
Trimerorhachis cf. *T. sandovalensis*
Fig. 2I-J

Referred specimens: CM 38008 (Fig. 2I-J), CM38010, partial skulls.

Description: One of the partial skulls of *Trimerorhachis* (CM 38008, Fig. 2I-J) consists of the posterior portions of the skull and mandibles, whereas the other (CM 38010) consists of the anterior portions of the skull and mandibles.

Discussion: As is the case with most of the other Gallina Wells specimens, these partial skulls are too poorly preserved to be positively identified at the species level, and are tentatively referred to *Trimerorhachis sandovalensis*, because this is the only species of *Trimerorhachis* recorded from the state. The holotype of *T. sandovalensis* (CM 38025) consists of a nearly complete skull with which is articulated a large portion of the anterior postcranium, from the Abo Formation of the Jemez Mountains (Berman and Reisz, 1980).

Trimerorhachis is frequently encountered, often on the basis of excellently preserved specimens, from Early Permian deposits distributed widely across North America. However, its occurrences in the Lower Permian of New Mexico, other than the holotype of *T. sandovalensis* and the Gallina Well specimens, are limited to poorly preserved skulls from the Lower Permian Jemez Mountains, Abo Pass, and Fra Cristobal/Caballo mountains localities and a single specimen (CM 38030) from the Upper Pennsylvanian Lucero uplift locality (Harris et al., 2004).

Superfamily Dissorophidea
Family Dissorophidae
Genus Platyhystrix
Platyhystrix cf. *P. rugosus*
Figs. 3A-D, 4A-C

Referred specimens: NMMNH P-35702 (Fig. 3A), dermal bone fragment; NMMNH P-63350 (Fig. 4A-C), incomplete skull; NMMNH P-65006 (Fig. 3B), dermal bone fragment; NMMNH P-65599 (Fig. 3C), indet. bone fragment; NMMNH P-66816 (Fig. 3D), dermal bone fragment.

Description: All of the Gallina Well specimens are referred to *Platyhystrix rugosus*, as it is the only known species of the genus. The various dermal bone fragments (NMMNH P-35702, P-65006, P-65599, P-66816) have a consistent ornamentation of shallow, irregular pits bordered by low ridges (Fig. 3A-D), whereas on the incomplete skull (NMMNH P-63350; Fig. 4A-C) a nodular or papillose sculpturing unique to the species occurs along the lateral borders of the skull table and adjoining portions of the cheek between the orbit and otic notch, as well as on the vertebral neural spines. The partial skull is heavily concreted and partially encased in matrix. The left side of the skull is visible, including the orbit and the narial margin, however, few sutures can be discerned. In these areas the dermal bones exhibit the nodular or papillose sculpturing unique to the species. The right mandible is crushed against the right side of the skull and is visible in ventral view. A single mandibular tooth is preserved in cross section.

Discussion: *Platyhystrix* has been recorded from all of the vertebrate-bearing localities recognized here except Fra Cristobal/Caballo mountains and Lucero uplift. Until the greater part of a skull with associated vertebrae of *Platyhystrix rugosus* (AMNH 11544), collected by David Baldwin from the Cutler Group of the Arroyo del Agua area in

1881, was described by Berman et al. (1981), its cranial anatomy remained unknown, and most importantly, its dissorophid assignment was speculative. Prior to that description, occurrences of *Platyhystrix* were based almost entirely on its unique presacral vertebrae, which are commonly encountered in the Lower Permian red beds of the Southwest, as well as in Texas (Berman et al., 1981). Langston (1953) also listed five small *Platyhystrix* skull and jaw fragments from the (presumably) El Cobre Canyon Formation at the Arroyo del Agua locality that were collected by field parties from the University of California, Berkeley, in 1934 and 1935. His assignment was based on their extremely papillose sculpturing. The easily recognized, unique neural spines of *Platyhystrix* are greatly elongated, laterally compressed, and distally expanded in the sagittal plane with a distinctive nodular or papillose sculpturing or ornamentation and form a high dorsal sail. The Gallina Well *Platyhystrix* skull is only the second to be reported.

Family Zatrachyidae
Genus Zatrachys
Zatrachys serratus
Fig. 3E-F

Referred specimens: CM 38007 (Fig. 3E-F), right posterior portion of skull and mandible.

Description: Right cheek portion of the skull roof (Fig. 3E-F), which extends posteriorly from the posterolateral margin of the orbit and also includes the underlying portion of the palate and posterior end of the mandible. Although the bone surface is severely weathered, making the tracing of sutures impossible, its general anatomy eliminates any doubt as to its identification. Furthermore, the complex sculpturing pattern of well-developed ridges, troughs, and tubercles of *Zatrachys* is unmatched in any other Permian amphibian.

Discussion: *Zatrachys serratus*, the only species of the genus, has been identified from two other Lower Permian localities in New Mexico: fragmentary cranial remains from the Jemez Mountains locality and a large series of well-preserved adult skulls from the Arroyo del Agua locality (Langston, 1953). *Z. serratus* is well known from numerous, excellently preserved skulls from the Lower Permian of Texas and Oklahoma (Olson, 1965; Schoch, 1997), and recently has been described on the basis of the left half of the skull roof table posterior to the orbit and skull-roof fragments from the Upper Pennsylvanian-Lower Permian of West Virginia (Urban and Berman, 2007). *Zatrachys* is posited as obligatory aquatic, especially in view of Langston's (1953, p. 396) interpretation of the depositional conditions in which the New Mexico specimens were preserved in the Arroyo del Agua area that suggest "quiet pools provided the normal habitat of adults of this species." The rarity of such habitats in the Lower Permian of New Mexico may explain the few occurrences of *Zatrachys*.

Order Nectridea
Family Diplocaulidae
Genus Diplocaulus
Diplocaulus sp.
Fig. 3G-H

Referred specimens: CM 38011 (Fig. 3G-H), partial skull.

Description: Harris et al. (2005) described CM 38011 (Fig. 3G-H), which consists of the right posterior horn-like portion of the skull of *Diplocaulus*, including elements of the parietal, squamosal, quadratojugal, postparietal, tabular, opisthotic, exoccipital and pterygoid.

Discussion: Other than the Gallina Well specimen, which was described as too incomplete to assign to a known species (Harris et al., 2005), occurrences of *Diplocaulus*, although locally abundant in the Clear Fork Group of Texas, are quite rare in the Lower Permian of New Mexico: a single vertebra from the Jemez Mountains locality; numerous, articulated strings of vertebrae, some associated with cranial elements, from the Abo Pass locality; and isolated vertebrae from the Pecos River valley

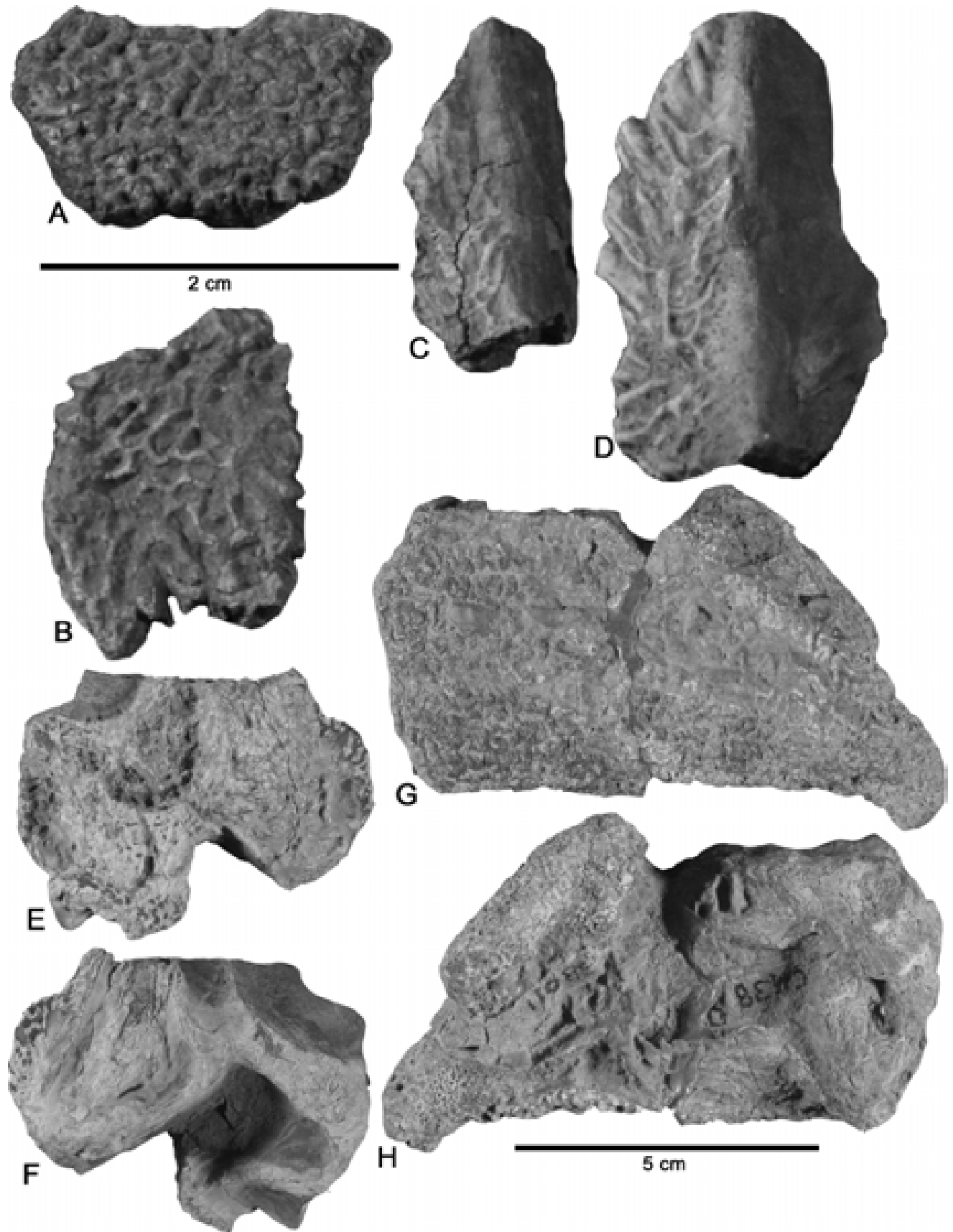


FIGURE 3. **A-D**, *Platyhystrix* cf. *P. rugosus*, dermal bone fragments in external view, **A**, NMMNH P-65006, **B**, NMMNH P-35702, **C-D**, NMMNH P-65599. **E-F**, *Zatrachys serratus*, CM 38007, right posterior portion of skull and mandible. **G-H**, *Diplocaulus* sp., CM 38011, right side of skull in **G**, dorsal and **H**, ventral views.

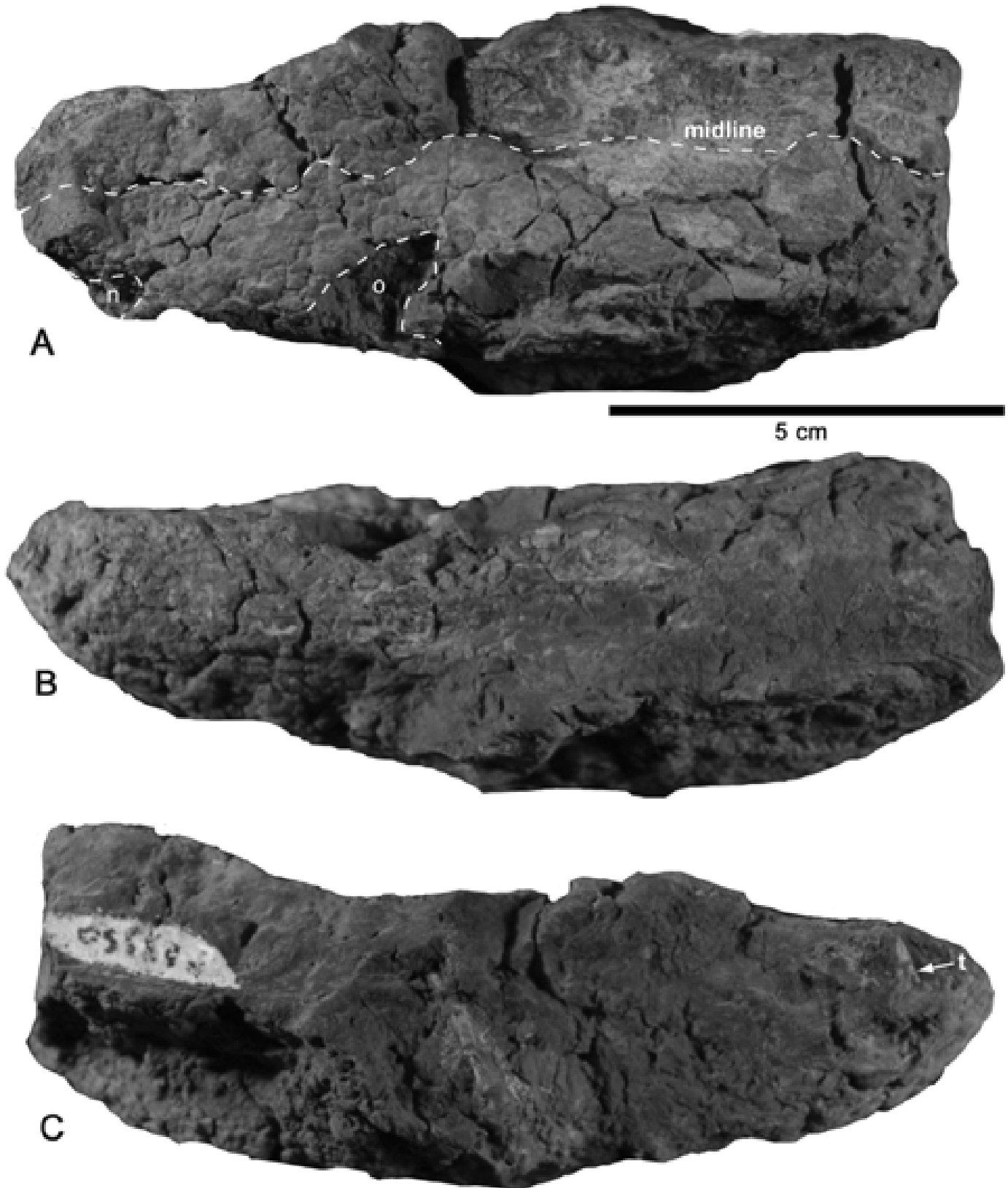


FIGURE 4. *Platyhystrix* cf. *P. rugosus*, NMMNH P-63350, skull in A, dorsal, B, left lateral and C, cross-sectional views. The midline suture is highlighted as well as key morphologic features. **Abbreviations:** n, narial opening; o, orbit; t, tooth.

locality. The Abo Pass specimens all occur in a freshwater limestone believed to be a small pond or lake deposit, which also contained specimens of the lungfish *Gnathorhiza* and the amphibian *Trimerorhachis* (Berman, 1976, 1979; Berman et al., this volume). *Diplocaulus* has been interpreted as obligatory aquatic, inhabiting persistent and temporary ponds/lakes and rivers. The rarity of such habitats in the Lower Permian of New Mexico may explain why only a single, articulated specimen of *Diplocaulus* has been found in New Mexico.

Order Diadectomorpha

Family Diadectidae

Diadectes sp.

Fig. 5A-E

Referred specimens: CM 73361 (Fig. 5A-C), left dentary, NMMNH P-62989 (Fig. 5D-E), nearly complete vertebra.

Description: The dentary, CM 73361 (Fig. 5A-C), appears to be complete except for the loss of teeth. The single vertebra, NMMNH P-62989 (Fig. 5D-E), is missing the tip of its neural spine and the tips of its left pre- and postzygapophyses, but is otherwise complete. The neural arch is partially concreted, obscuring some details.

Discussion: *Diadectes*, a large, highly terrestrial herbivore, is widely distributed across the United States in Lower Permian deposits, which in New Mexico include the Jemez Mountains, El Cobre Canyon, Arroyo del Agua, and Pecos River valley localities. These records are based mainly on isolated or associated elements, but also include five undescribed, partial, articulated skeletons. These are four adult postcranial skeletons, three (CM 38036, 38045, 41700) of which are from the Arroyo del Agua locality and one (CM 38041) from the Jemez Mountains locality; the fifth specimen is a nearly entire skeleton of a small juvenile (CM 38036) that also includes large parts of the skull from the Arroyo del Agua locality.

Also noteworthy, a partial skull with braincase of a diadectid (NMMNH P-35751) has also been recovered from the Upper Pennsylvanian at the Lucero uplift locality (Harris et al., 2004). Unfortunately, poor preservation prevents identification below the family level.

Class Amniota

Order Synapsida

Suborder Eupelycosauria

Family indet.

Fig. 5F-U

Referred specimens: NMMNH P-65565 (Fig. 5L-P), vertebra; NMMNH P-66811 (Fig. 5F-H), centrum; NMMNH P-66813 (Fig. 5I-K), vertebra; NMMNH P-66819 (Fig. 5Q-U), ungual phalanx.

Description: NMMNH P-65565 is a heavily concreted vertebral centrum and neural arch with a prominent ventral keel. NMMNH P-66811 is a mediolaterally crushed vertebral centrum missing its neural arch. NMMNH P-66813 is a well preserved vertebral centrum with a heavily concreted neural arch and a ventral keel. NMMNH P-66819 is a single ungual phalanx with a spherical flexor muscle insertion and a prominent ventral keel.

Discussion: It is strongly suspected that these elements pertain to either *Ophiacodon* or *Sphenacodon*, but poor preservation prevents positive identification.

Family Ophiacodontidae

Genus *Ophiacodon*

Ophiacodon sp.

Fig. 5V-X

Referred specimens: CM 87691, a complete femur; NMMNH P-65018 (Fig. 5V-X), proximal portion of the left femur.

Description: In size and morphology, both elements are consistent with the femur of the *Ophiacodon navajovicus* specimen NMMNH

P-43121, the most complete specimen known, consisting of mainly disarticulated elements of a single postcranial skeleton from the Upper Pennsylvanian of the El Cobre Canyon locality (Harris et al., 2010). The *Ophiacodon* identification is most convincingly argued by the presence of a secondary or posterior adductor ridge (or crest) of the shaft of CM 87691.

Discussion: Two species of *Ophiacodon* from New Mexico are considered valid (Romer and Price, 1940; Reisz, 1986): *O. navajovicus*, restricted to the Upper Pennsylvanian of the El Cobre Canyon locality; and *O. mirus*, restricted almost entirely to the Lower Permian of the Arroyo del Agua locality. Notable specimens include a pelvis and hind limb elements (CM 76896) and the greater portion of a partly articulated skeleton (CM 38034) from the Lower Permian Arroyo del Agua locality.

In this volume, Berman et al. describe a partial, articulated skeleton (CM 47771) of a large *Ophiacodon* from the Lower Permian Pecos River valley locality that undoubtedly represents a new form for the state. It includes portions of two sacral vertebrae with fused ribs, the centrum of the last presacral vertebra, partial pelvis, the left hind limb and most of the pes, and a few probable anterior caudal ribs. It is the largest *Ophiacodon* specimen to be reported in New Mexico and is most comparable to the Texas Lower Permian *O. retroversus* on the basis of size and a few anatomical features, but also shares a few features with the much smaller *O. mirus*. Although possibly representing a new species, a specific assignment of CM 47771 is deferred until more complete material becomes available. Finally, dorsal vertebrae of *Ophiacodon* have been described from the Lower Permian Jemez Mountains and Tularosa localities (Lucas et al., 2012; Vaughn, 1969, respectively).

Family Sphenacodontidae

Genus *Sphenacodon*

Sphenacodon ferox?

Fig. 6A-H

Referred specimens: NMMNH P-65020 (Fig. 6D-E), proximal end of left femur; NMMNH P-65029 (Fig. 6A-C), distal end of left femur; NMMNH P-66810 (Fig. 6F-H), complete vertebra; CM 87692, distal end of femur.

Description: In size and morphology these elements (Fig. 6A-H) are consistent with *Sphenacodon ferox*. The proximal end of the femur, NMMNH P-65020 (Fig. 6D-E), lacks an intertrochanteric fossa on its proximal articular surface, and its internal trochanter is not greatly extended anteriorly so as not to be visible, and is hidden in dorsal view of the femur; both of these features contrast with the femora of *Ophiacodon navajovicus* illustrated by Harris et al. (2010).

The distal head of the femur, NMMNH P-65029 (Fig. 6A-C), is much narrower compared to the broadly expanded head in *Ophiacodon*. Furthermore, the larger posterior condyle is much more greatly developed and extends much farther distally beyond the anterior condyle than in *Ophiacodon*. The complete vertebra, NMMNH P-66810 (Fig. 6F-H), has a moderately sized, rectangular neural spine that compares well with those in *Sphenacodon*, and is easily distinguished from the elongated neural spines of *Dimetrodon* and the broader, shorter neural spines in *Ophiacodon*.

Discussion: *Sphenacodon* is a very common element of the Upper Pennsylvanian-Lower Permian vertebrate assemblages of New Mexico: Upper Pennsylvanian and Lower Permian of the El Cobre Canyon locality; Lower Permian of the Jemez Mountains, Tularosa, and Pecos River valley localities; and the Upper Pennsylvanian of the Lucero uplift locality (Spielmann et al., 2010). Only two species of *Sphenacodon* are recognized, *Sphenacodon ferox* and *S. ferocior*, both of which occur, including their holotypes, in New Mexico, and until recently could only be distinguished on the basis of the much larger size, more stout build, and greater length of the neural spines of the latter species, which was undoubtedly the apex carnivore during the Early Permian in New Mexico (Romer and Price, 1940; Reisz, 1986). However, with the description of a recently discovered, superbly preserved skull from the Lower Permian

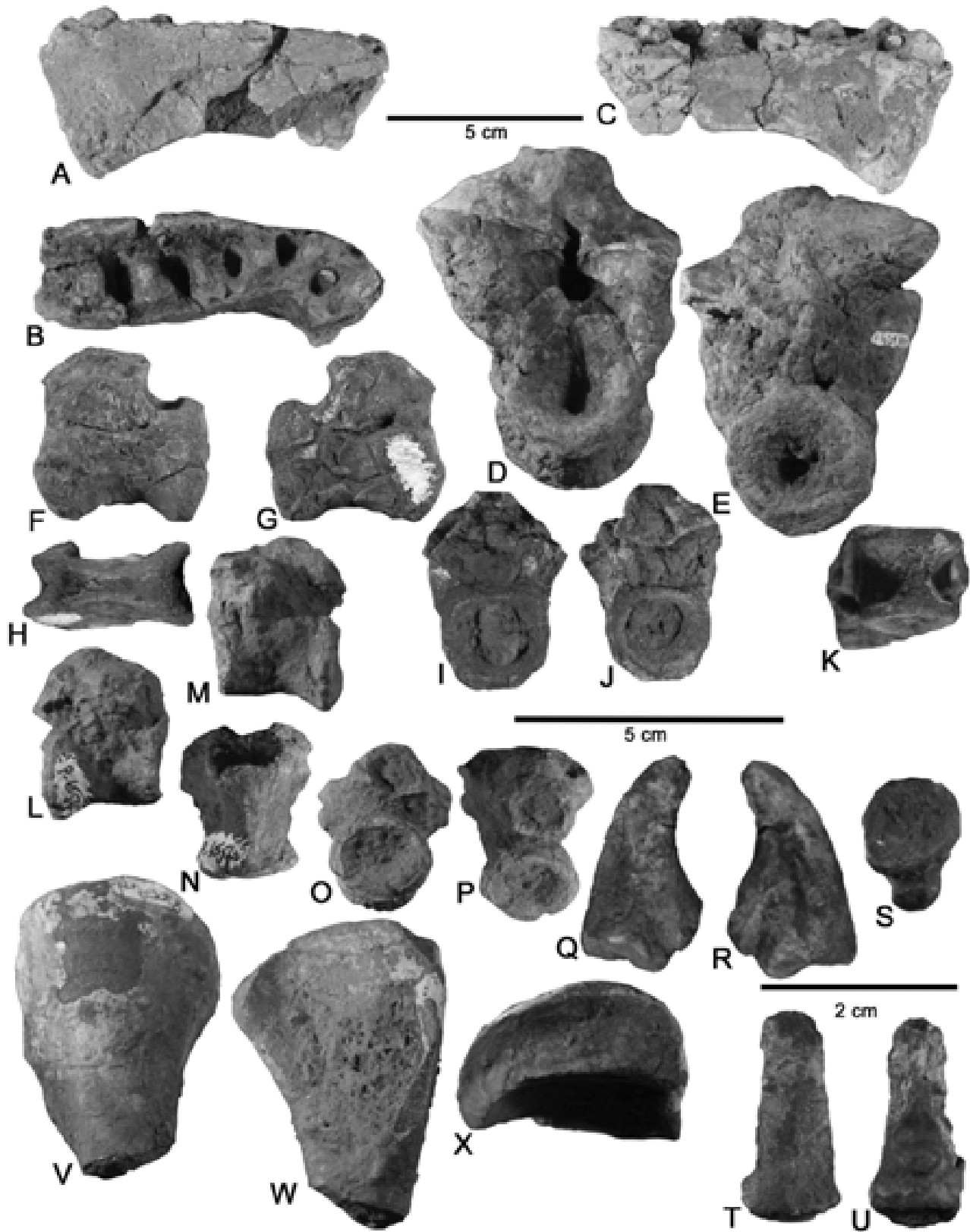


FIGURE 5. A-C, *Diadectes*, CM 73361, left dentary fragment in A, lateral, B, occlusal and C, medial views. D-E, *Diadectes*, NMMNH P-62989, dorsal vertebra in D, anterior and E, posterior. F-H, Pelycosauria indet., NMMNH P-66811, vertebral centrum in F, left lateral, G, right lateral and H, ventral views. I-K, Pelycosauria indet., NMMNH P-66813, incomplete dorsal vertebra in I, anterior, J, posterior and K, ventral views. L-P, Pelycosauria indet., NMMNH P-65565, vertebral centrum in L, right lateral, M, left lateral, N, ventral, O, anterior and P, posterior views. Q-U, Pelycosauria indet., NMMNH P-66819, ungual phalanx in Q, medial, R, lateral, S, articular, T, dorsal and U, ventral views. V-X, *Ophiacodon*, NMMNH P-65018, proximal left femur in V, dorsal, W, ventral and X, proximal views.

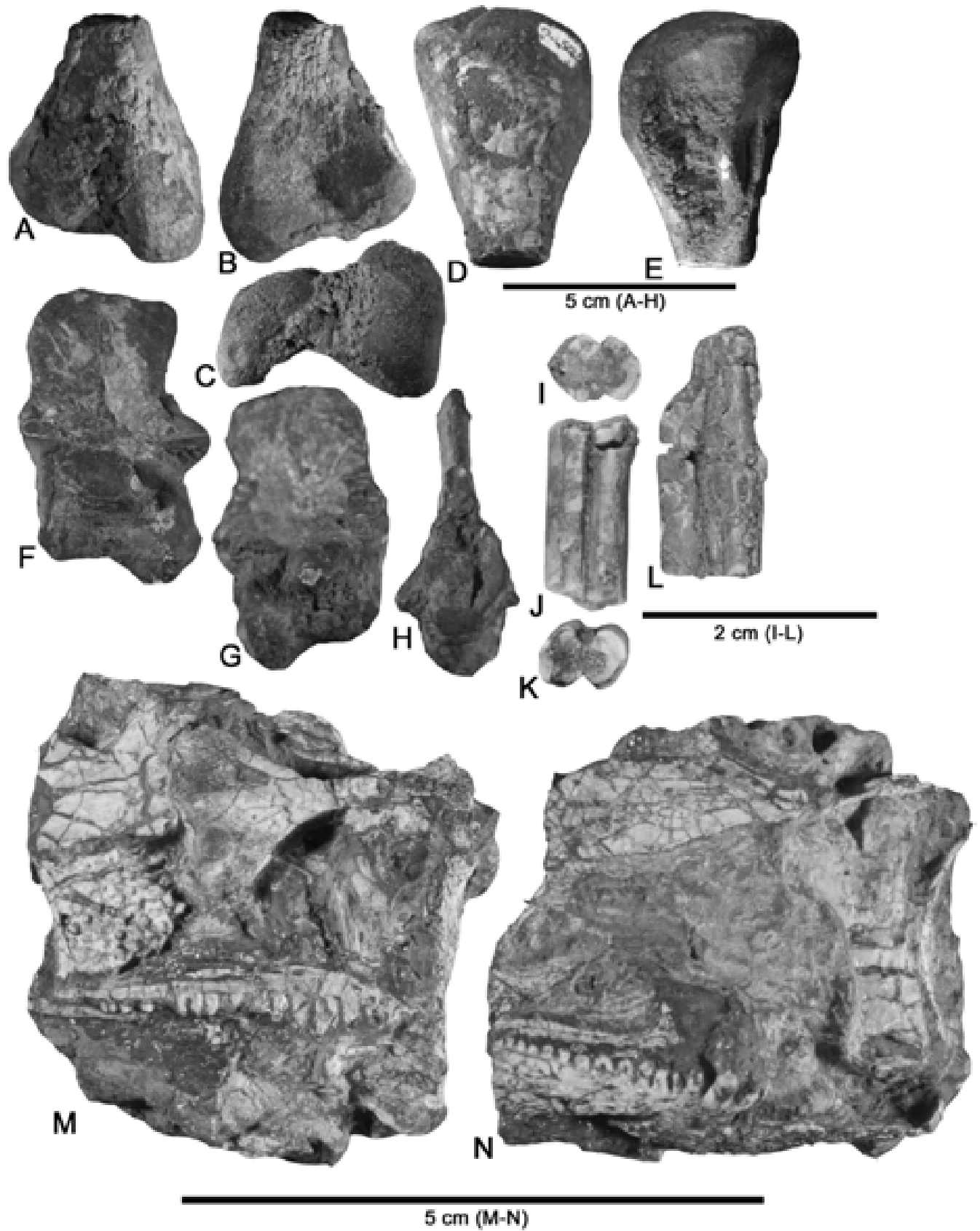


FIGURE 6. A-C, *Sphenacodon*, NMMNH P-65029, distal end of left femur in A, dorsal, B, ventral and C, distal views. D-E, *Sphenacodon*, NMMNH P-65020, proximal end of left femur in D, dorsal and E, ventral views. F-H, *Sphenacodon*, NMMNH P-66810, dorsal vertebra in F, left lateral, G, right lateral and H, anterior views. I-L, *Dimetrodon* sp., NMMNH P-35713, isolated neural spine fragments in J, L, lateral and I, K, axial views. M-N, Captorhinid, CM 41707, associated cranial and postcranial elements.

Arroyo del Agua Formation of the El Cobre Canyon locality, Spielmann et al. (2010) recognized 11 cranial features that distinguish the two species. Confident species assignments of *Sphenacodon* specimens, however, have been limited to: *S. ferox* from the Upper Pennsylvanian and Lower Permian of the El Cobre Canyon locality and Lower Permian El Cobre Canyon Formation of the Arroyo del Agua locality, and *S. ferocior* from the Lower Permian Jemez Mountains and Arroyo del Agua localities (Romer and Price, 1940; Reisz, 1986). The carpus and tarsus of *S. ferox* were described in detail for the first time on the basis of an incomplete, partially articulated skeleton from the Lower Permian El Cobre Canyon Formation of the Arroyo del Agua locality (Henrici et al., 2005).

Although occurrences of the two species overlap temporally (Lucas et al., 2005b), with both co-occurring in the latest Pennsylvanian Virgilian and earliest Permian Wolfcampian Coyotean land-vertebrate faunachron of Lucas (2005, 2006), only *S. ferox* extends into the younger, earliest Permian, Seymouran faunachron. Outside of New Mexico elements identified as sphenacodontid or *Sphenacodon* cf. *S. ferocior* have been reported from the Lower Permian Halgaito and Organ Rock formations of southeastern Utah and neighboring areas (Vaughn, 1962, 1964).

Genus *Dimetrodon*

Dimetrodon sp.

Fig. 6I-L

Referred specimens: NMMNH P-35713 and NMMNH P-62985 (Fig. 6I-L), two isolated neural spines.

Description: Cantrell et al. (2011) described these specimens. Cross-sections of the neural spines (Fig. 6I-L) exhibit the unmistakable, unique figure-8 outline characteristic of the genus.

Discussion: Previous records of *Dimetrodon* from the Lower Permian Abo Formation in New Mexico include: (1) vertebral fragments identified as *Dimetrodon* cf. *D. limbatus* from the Caballo Mountains locality (Vaughn, 1969; Lucas et al., 2012b); (2) vertebral fragments identified as *Dimetrodon* cf. *D. limbatus* from the Cerros de Amado near the Joyita Hills, central Socorro County (Lucas et al., 2009); (3) vertebrae assigned to *Dimetrodon* sp. from the Jemez Mountains locality (Madalena et al., 2007); and (4) three incomplete, loosely articulated vertebrae and the neural spine of a fourth (NMMNH P-36219) and a neural spine (NMMNH P-32519) have been described and assigned to *Dimetrodon* cf. *D. milleri* from the Upper Pennsylvanian of the Lucero uplift locality (Harris et al., 2004). Most importantly, a large portion of the partially articulated dorsal sail and elements of the left mandible (CM 26565) from the Jemez Mountains locality formed the basis of a small, new species, *Dimetrodon occidentalis* (Berman, 1977). It differs from other species of comparable size in the proportions of the centra, in exhibiting a marked decrease in the cross-sectional area of the distal portions of the neural spines, but still reaching an estimated greatest height of 360 mm. *D. occidentalis* is most comparable to *D. milleri*, with a greatest spine height of 390 mm and a estimated total skeletal length of 1990 mm. *D. milleri*, however, can be distinguished from *D. occidentalis*, as well as all other members of the genus, in having a round, cross-sectional outline of the neural spines, rather than the typical figure-8 outline.

Reptilia

Order Captorhinomorpha

Family Captorhinidae

Fig. 6M-N

Referred Specimens: CM 41707 (Fig. 6M-N), a left maxilla, left jugal, anterior half of the right mandible, a presacral vertebra, ribs, greater part of the left humerus, distal half of the right femur, and several unidentified fragments.

Description: The individual elements are randomly associated and densely concentrated in a small, strongly indurated, red, concretion-like mass (Fig. 6M-N).

Discussion: Records of Early Permian captorhinid reptiles in

New Mexico are sparse. The first to be described, *Puercosaurus obtusidens* Williston, 1916, was based on three poorly preserved, incomplete specimens, a left dentary and two skulls, collected from the Arroyo del Agua locality. It was declared a *nomen dubium*, however, by Berman and Reisz (1986), because the holotypic dentary is indeterminate, and there is no basis for accepting that it and the two skulls found at a different locality and referred to the species by Williston (1916) are conspecific. Four additional, indeterminate captorhinid specimens were described by Berman and Reisz (1986) from the same locality: (1) two crushed and very incomplete skulls referred to "*Puercosaurus obtusidens*" by Williston (1916); (2) a partial skull; (3) 14 dorsal vertebrae, most of which are articulated in strings of two or three, associated with ribs and appendicular elements; and (4) a partial left maxilla, small portions of both dentaries, presacral vertebra, and a left humerus. Numerous indeterminate fragments of maxillae and dentaries were also reported by Berman and Reisz (1986) from the Pecos River valley locality.

The only recognized species from the Arroyo del Agua locality, *Rhiodenticulatus heatoni*, was described as a new genus and species by Berman and Reisz (1986), which was based on three specimens: the holotype (UCMP 35757), consisting of a partial, articulated skeleton that includes the skull with mandibles and large portions of the axial and appendicular skeletons, and two paratypes, one (UCMP 40209) consisting of the greater portions of the skull and joined mandibles, and the other (UCMP 40210) consisting of three small portions of a partial, articulated postcranial skeleton. It is not unlikely that one or more of the five indeterminate specimens from the Arroyo del Agua locality could pertain to *R. heatoni* or a new form.

VERTEBRATE COPROLITES

Cantrell et al. (2012) described the vertebrate coprolites from the Gallina Well locality. The presence of *Dakryonokopros arroyoensis* at the Gallina Well locality extends the geographic range of the ichnogenus from the Lower Permian (middle Wolfcampian to middle Leonardian) strata of Texas into the Lower Permian (middle Wolfcampian) strata of New Mexico (Cantrell et al., 2012). *Dakryonokopros arroyoensis* occurs in conjunction with *Heteropolacopros texaniensis* in the Lower Permian strata of Texas as it does at the Gallina Well locality. *Alococopros triassicus* coprolites mark the earliest occurrence of the ichnogenus (known previously from the Early Triassic) and refute previous assertions that longitudinally-striated coprolite forms were produced by stem archosauromorphs (Cantrell et al., 2012). A comparable coprolite ichnoassemblage from north-central New Mexico was documented by Hunt et al. (2005) from three Lower Permian localities in the Arroyo del Agua area of the Chama Basin: the VanderHoof quarry, Welles quarry and Quarry Butte locality.

DISCUSSION

The middle Wolfcampian age of the Abo Formation in the Joyita Hills area was determined based on the early Wolfcampian age of the underlying Bursum Formation (cf. Krainer and Lucas, 2009), vertebrate biostratigraphy, and regional lithostratigraphic correlation.

Lucas (2005, 2006) established land-vertebrate faunachrons (LVFs) to divide Permian time based on the first appearances of key vertebrate taxa. Unfortunately, none of the taxa found at Gallina Well are temporally restricted enough to provide a precise age assignment. Thus, *Zatrachys* occurs in the Coyotean, Seymouran and Redtankian LVFs, *Trimerorhachis* occurs in the Coyotean through the Redtankian LVFs, and *Diplocaulus* occurs from the Coyotean through Littlecrotonian LVFs. Thus, these taxa range throughout much of the Early Permian. The presence of *Sphenacodon* material in the quarry suggests a Coyotean or Seymouran age for the assemblage, and a Coyotean age is most likely given how stratigraphically low the assemblage is in the Abo Formation (Fig. 1). Indeed, the mixed terrestrial-aquatic vertebrate fauna and ichnoassemblage from the Gallina Well locality is very similar to the Coyotean fossil assemblages from El Cobre Canyon, Arroyo del Agua and Jemez Springs, in north-central New Mexico. This suggests unifor-

mity of the Coyotean vertebrate fauna and ichnoassemblage across much of New Mexico.

The classic Early Permian collecting localities of New Mexico are primarily located in the northern half of the state (see review by Berman, 1993). Examples include the Jemez area (Lucas et al., 2012a), Arroyo del Agua (Lucas et al., 2005a) and El Cobre Canyon (Lucas et al., 2010). Gallina Well is the exception to this rule, and represents a major Early Permian fauna in the southern part of New Mexico. The faunas between the northern localities and Gallina Well are consistent and there are no distinctly northern or southern taxa. The Gallina Well locality yields the

most diverse and extensive vertebrate body fossil and coprolite assemblage of Early Permian age known from southern New Mexico. Its basic composition differs little from the pelycosaur-dominated assemblages found to the north, indicating some uniformity of the Coyotean vertebrate fauna across New Mexico.

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FLORA OF THE LOWER PERMIAN ABO FORMATION REDBEDS, WESTERN EQUATORIAL PANGEA, NEW MEXICO

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Abstract—The Lower Permian, Cisuralian age Abo Formation and its equivalents of New Mexico contain a low diversity flora dominated by several species of walcchian conifers and the peltasperm *Supaia thinnefeldioides*. A variety of much less common taxa includes the coniferophytes *Cordaites* and *Dicranophyllum*, the peltasperms *Supaia anomala*, *Brongniartites/Glenopteris*, *Autunia*, *Rhachiphyllum*, and *Gigantopteridium*, the possible cycads *Taeniopteris*, cf. *Russelites/Yuania*, and cf. *Plagiozamites*. A single site is dominated by *Glossopteris/Lesleya*-shaped foliage. A few specimens of rare, more typically wetland taxa, are present, particularly the pteridosperm *Neuropteris*, the sphenopsids *Sphenophyllum* and calamitalean stems, and the marattialean tree fern *Pecopteris*. Total species diversity is about 27, which is remarkably low given that plants were identified at 172 red beds sample sites covering a minimum area of about 83,000 km², from the Robledo Mountains and Fort Bliss in southern New Mexico, to the Zuni Mountains and Jemez Mountains in the north. This raises the question of taphonomic bias; conifers and supaioids had unusually stiff, resistant construction, which may have differentially favored their survival in depositional systems. A number of factors weigh against an interpretation of special biases, unique to these deposits. Abo red beds are sedimentologically heterogeneous, not consisting of one particular kind of environment for plant preservation. Yet, conifers and supaioids remain the most commonly represented plants throughout, both in still-water dense-debris accumulations and in active flow deposits where remains are generally isolated. Additionally, conifers and supaioids, although occurring together in many deposits, do not co-dominate sites, and mostly occur in isolation from one another, suggesting very different ecological distributions. Rare taxa, in contrast, mostly occur as singletons, or in assemblages that they dominate, less commonly mixed with conifers and supaioids, suggesting patchy distributions on the landscape, not preservational biases as the cause of their rarity and the generally low diversity.

INTRODUCTION

The change in floral composition from the Pennsylvanian to the Permian is well documented in the Euramerican parts of central and west-central Pangea, present day western North America through Europe. Here, assemblages of xeromorphic plants replaced wetland plants as the most commonly encountered vegetation of equatorial lowland basins. This change was driven principally by long-term, directional, if oscillatory, changes in the seasonality of rainfall and perhaps mean annual rainfall, across most of central and western Pangea, beginning in the Middle Pennsylvanian (Moscovian) and continuing through the Late Permian.

In the westernmost parts of Pangea, represented today by western North America, the rise in the frequency of occurrence of seasonally dry vegetation in basinal lowlands took place much sooner than in the peat-forming coal basins of the central portions of the supercontinent, beginning in the west at least by the Middle Pennsylvanian and perhaps even earlier in some regions (Tidwell and Ash, 2003). By the Middle Pennsylvanian, such plants as conifers, the pteridosperm *Sphenopteridium*, the cycadophytes *Charliea* and *Taeniopteris*, cordaitaleans and other plants that typify seasonally dry substrates had become widespread and abundant in the western reaches of Pangea and were penetrating into the central regions, especially during the drier portions of glacial-interglacial cycles (Plotnick et al., 2009; Falcon-Lang and DiMichele, 2010; Falcon-Lang et al., 2011). By the Pennsylvanian-Permian transition, this changeover was essentially complete in most of the western reaches of the equatorial zone, resulting in widespread development of a complex, seed-plant dominated vegetation that was highly

habitat differentiated, given its regional and sometimes local variability (Tidwell and Ash, 2004; DiMichele and Chaney, 2005). And, by the Early Permian, a widespread, conifer and peltaspermous pteridosperm vegetation dominated large portions of seasonally dry areas of the central and western Pangean continent (Kerp and Fichter, 1985; Kerp, 1996; Galtier and Broutin, 2008; Tabor et al., 2013). Existing side-by-side with this seasonally dry substrate flora was a now-depauperate wetland vegetation, probably confined to riverine corridors, stream floodplains and lakesides, where persistent soil moisture was present (DiMichele et al., 2006).

The Early Permian Abo Formation red beds, and their equivalents, in central New Mexico formed as a widespread inland to coastal fluvial complex. The formation is composed of fine-grained sandstone- and siltstone-filled channels alternating with well developed vertic, and often calcic paleosols (Mack et al., 1991, 2010; Mack, 2003, 2007), indicating a seasonal climate, with long or intensely dry periods. It has long been known as the source of a low diversity, conifer-dominated paleoflora (Hunt, 1983; Lucas et al., 1999), one that occurs in close association with vertebrate and invertebrate trackways (Lucas and Heckert, 1995) and occasional vertebrate skeletal remains (Lucas et al., 1999, 2009). The flora is preserved mainly as impressions, which has made it of limited interest for studies of plant morphology, there being no structural preservation, little or no cuticle, and with finer details of surface morphology often obscured. Nonetheless, there are reports of well preserved plants at some locations (e.g., Hunt, 1983; DiMichele et al., 2007), and these floras can, in many instances, be linked to depositional environments, permitting understanding of much of their taphonomy. Furthermore, the very low diversity of the Abo Formation red beds flora, especially the

conspicuous and overwhelming dominance of conifers and supaioid plants, is of interest. This low diversity is spotlighted further by the presence of higher diversity gray, generally siltstone-to-sandstone facies in the lower Abo Formation throughout much of its outcrop area. These gray deposits, inter-tongued with or laterally equivalent to the Abo red beds, not only contain a higher diversity flora than that of the red beds, but that flora differs in significant ways from that of the red beds. Consequently, the high dominance and low diversity of the Abo red beds over a spatially large area, is noteworthy and perplexing.

GEOLOGY OF THE ABO FORMATION AND EQUIVALENTS

The Abo Formation, and its correlatives in New Mexico, crop out in a north-south band through the center of New Mexico, primarily along the margins of the Rio Grand rift system, from near the southern borders of the state on the Ft. Bliss Military Reservation, near El Paso, Texas and the Robledo Mountains, near Las Cruces, to northern New Mexico, in the Zuni Mountains near Gallup and the southern Jemez Mountains west of Santa Fe (Fig. 1). The type area (Fig. 1 – labeled “Abo Pass”) is within Abo Pass at the southern end of the Manzano Mountains in western Torrance County (Lucas et al., 2005, 2013). In the vast majority of the Abo outcrop area the formation is entirely of terrestrial (fluvial) origin (Lucas et al., 2005, 2012). In southern New Mexico the terrestrial Robledo Mountains Formation, an Abo lithological equivalent, interfingers with marine units of the Hueco Group in its lower parts (Fig. 2) (Krainer and Lucas, 1995; Krainer et al., 2009; Mack, 2007; Voigt et al., 2013). The intertonguing of the siliciclastic Abo red-bed lithosome and the carbonate-dominated Hueco Group lithosome extends across southern New Mexico, in Doña Ana, southern Sierra and Otero counties, and is seen on outcrop in the Robledo, Doña Ana, San Andres, southern Caballo, Jarilla and Sacramento Mountains (e.g., Kottlowski et al., 1956; Bachman and Hayes, 1958; Pray, 1961; Lucas et al., 1995, 2002, 2012).

In a general sense, the Abo Formation (including its terrestrial equivalents, Fig. 2) can be divided into a lower third dominated by mudstone and channel-form sandstone, the Scholle Member, and a siltstone-sandstone-dominated upper two thirds, the Cañon de Espinosa Member, marking a generally fining upward profile. In the Scholle Member, sandstone channels, many with conglomeratic basal lags and erosional basal contacts, are incised into thick, clay-rich paleosols of vertic character, often with carbonate nodules. Channel sands may be flanked by thin, tabular, flat-bottomed sheet fine sands and siltstones. The Cañon de Espinosa Member consists dominantly of fine-grained sheet sandstones and siltstones separated by mudstones, generally with pedogenic overprinting of vertic and sometimes calcic character.

In many of the sheet sandstones and siltstones from throughout the formation, the upper parts of beds, sometimes just a few centimeters and occasionally significantly more, consist of an upward-fining sequence of siltstone/sandstone and siltstone/claystone interbeds. The finer grained sediments frequently show evidence of flow under partially exposed conditions (rills and small runoff features), or complete exposure, such as mud cracks, often of multiple generations on a single surface, and raindrop imprints. On some of the finer grained surfaces, footprints and trackways of vertebrates and invertebrates are common (Lucas and Heckert, 1995). These finer-grained layers also often include plant fossils, from isolated specimens to abundant masses of foliage and branches (DiMichele et al., 2007). Plant fossils also occur, however, in the sandstones themselves, often crossing bedding planes, suggesting transport and rapid deposition. As with the finer grained layers, plant fossils in sandstones and siltstones may occur in isolation or as part of more concentrated accumulations.

In many parts of the Abo outcrop area, exposures do not permit exact assessment of position within the formation. This results from the influence of cover, faulting, and erosion, which may obscure the base or top, and, in the case of faulting and erosion, often remove parts of the formation from the local outcrop.

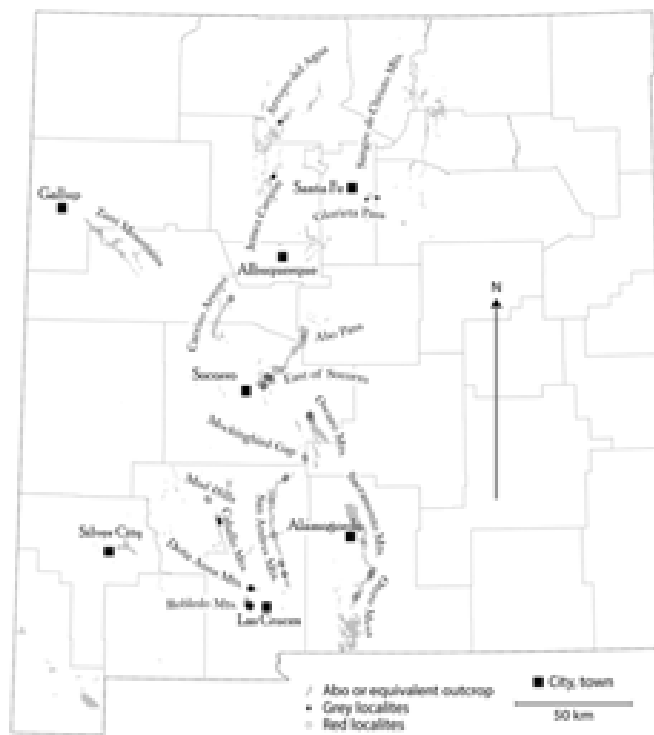


FIGURE 1. Outcrop area of the Abo Formation and its equivalent units in New Mexico. Sites from which fossil plants were collected are marked by open circles (red beds) or black crosses (gray beds).

MATERIALS AND METHODS

Abo Formation exposures were examined throughout the outcrop area (Fig. 1). The objective was to find and identify plant fossils, their environment of deposition, and, as far as possible, their position within the Abo Formation (lower, middle, upper). Due to the often incomplete nature of Abo Formation exposure, correlation from one exposure to another was confined to a broad classification of upper and lower, sometimes middle. Basal exposures in the immediately overlying Yeso Group also were examined in many places.

When plant fossils were found, their geographic position was recorded. A geologic description was made, and the outcrop was photographed. As often as possible, plant-fossil occurrences were put into the larger context of position within the local Abo section.

If it proved possible to collect specimens, a collection was made from a plant-bearing site. Because of the nature of the rocks, it was not always possible to remove specimens, particularly where the matrix was massive and the fossils were not on a bedding surface that could be removed with hand tools. In the many instances where the flora was monotypic and the number of specimens that could be collected large, only a voucher or two was removed to serve as a record of that locality. As often as possible, specimens were photographed in the field, particularly if they could not be removed. In some cases, plant fossils were found in debris fans derived from weathering of several siltstone/sandstone beds, and thus could not be related with absolute certainty to a particular layer; in these instances a “float” collection was made, and so labeled.

Specimens referred to in this study are housed either in the paleobotanical collections of the National Museum of Natural History, Smithsonian Institution, Washington, DC (USNM), or the paleontological collections of the New Mexico Museum of Natural History and Science, Albuquerque, NM (NMMNH).

Chama Basin		Jemez Springs		Cerro de Amado		Caballo Mountains		Robledo Mountains		AGE	
Yeso Group (De Chelly Sandstone)		Yeso Group (De Chelly Sandstone)		Yeso Group (Arroyo de Alamillo Formation)		Yeso Group (Arroyo de Alamillo Formation)		Apache Dam Formation		Leonardian	
Cutler Group	Arroyo del Agua Formation	Abo Formation	Cañon de Espinoso Member	Abo Formation	Cañon de Espinoso Member	Abo Formation	Cañon de Espinoso Member	Robledo Mountains Formation	Hueco Group	late (Lenoxian)	Wolfcampian
	El Cobre Canyon Formation		Scholle Member		Scholle Member		Scholle Member	Shalem Colony Formation		middle (Nealian)	
		Bursum Formation	Bursum Formation	Bursum Formation	Horquilla Formation	early (Newwellian)					
	Guadalupe Box Formation	Atrasado Formation				Virgilian					

FIGURE 2. Stratigraphic nomenclature and correlation of the Abo Formation and equivalent units from north-central New Mexico (left) to south-central New Mexico (right). The Pennsylvanian-Permian boundary, at the time of writing of this paper, is at the Newwellian-Nealian contact.

RESULTS

Fossils were collected from 172 discrete red bed sites and cover the full range of the Abo Formation, both geographically and stratigraphically. Many areas, and some deposits, were collected more heavily than others, there being more individual collections made; in some instances these collections were replicates, collected from a discrete and identifiable sedimentary unit. However, the great majority of collections are isolated occurrences, most often from beds that either could not be traced over any distance, or, more often, that had plant fossils only at one or a few widely spaced points. Thus, we report the data for the full set of collections instead of agglomerating them by site, where possible.

Composition of the Flora

A comparison of the red beds and gray beds floras is summarized in Table 1.

Only three species occur at more than 10% of the red beds sampling sites: *Walchia piniformis* (49%), *Otoviclia hypnoides* (difficult to separate from *Brachyphyllum?*, 13%), and *Supaia thimfeldioides* (25%) (Table 1).

When looked at from the viewpoint of major evolutionary lineages, only three of the lineages – conifers (82%), supaioid peltasperms (28%), and callipterid peltasperms (12%, the most abundant of which individually is the small form of *Autunia conferta*, 9%) – occur at more than 10% of the collecting sites.

Diversity at any one collecting site averages about 2 species, with a maximum of 7 species from any field site. Nine species were found at one site that was collected heavily over a large area from multiple sites by Adrian Hunt (1983) in his study of the Abo Formation flora in parts of Socorro County.

TABLE 1. Percentage occurrences of taxa at sites examined in the Abo Formation red facies and gray facies. **** = Taxa that occur commonly in both red and gray facies. ^^ = Taxa that occur commonly in only one facies.

Taxon	Red Beds (% of n=172)	Gray Beds (% of n=30)
Conifers****	82****	73****
<i>Walchia piniformis</i> ****	49****	50****
<i>Otoviclia hypnoides</i> ****	13****	10****
<i>Calamitochia americana</i>	5	10^^
<i>Walchia</i> sp.****	26****	20****
<i>Calamitochia</i> sp.	0	23^^
<i>Cordaites</i> sp.	2	40^^
Supaioids	29^^	10
<i>Supaia thimfeldioides</i>	25^^	7
<i>Supaia anomala</i>	0^^	0
Callipterids	12	37^^
<i>Autunia conferta</i> ****	9****	10****
<i>Brachyphyllum lyrastipula</i>	1	10^^
<i>Brachyphyllum schenckii</i>	1	20^^
<i>Lodveia</i> sp.	0	13^^
<i>Sphenopteridium</i> sp.	0	10^^
<i>Pennsiperis</i> sp.	3	13^^
<i>Nerospirites</i> sp.	1	10^^
<i>Odontopteris</i> sp.	0	20^^
<i>Nerodontopteris</i> sp.	0	10^^
<i>Alentopteris</i> sp.	0	10^^
<i>Sphenophyllum</i> sp.	2	13^^
<i>Pecopteris</i> sp.	2	20^^
Calamitalean stems	2	33^^
<i>Sigillaria brandii</i>	0	13^^
Average Collection Richness	1.9 taxa	4.7 taxa
Total Richness (whole plant proxy)	28	37

Total diversity consists of 28 discrete species, more or less, depending on how the small filicalean ferns are subdivided, whether the conifer morphotypes are reliable indicators of discrete species, and how many species are tied up in the calamitalean stems, and so on. The total red beds flora consists of the following: *Walchia piniformis*, *Otoviccia hypnoides* (+ *Brachyphyllum*?), *Hermitia schneideri*, *Culmitzschia americana*, *Culmitzschia speciosa*, *Dicranophyllum* sp., *Cordaites* spp., *Supaia thinnfeldioides*, *Supaia anomala*, *Brongniartites* sp., *Autunia conferta* (small form), *Autunia conferta* (large form), *Rhachiphyllum lyratifolia*, *Rhachiphyllum schenkii*, *Gigantopteridium americanum*, *Taeniopteris* (3 distinct morphotypes), *Charliea/Russelites/Tingia* sp., *Plagiozamites* sp., cf. *Glossopteris/Lesleya*, *Neuropteris* sp., indeterminate calamitalean stems, *Sphenophyllum*, *Pecopteris* sp., and small filicalean ferns.

Only a limited number (30) of Abo Formation gray-bed sites have been analyzed to date (Table 1), but these yield plants that differ greatly from those of the red beds in their patterns of dominance and diversity, although many species co-occur, suggesting a shared regional species pool. Gray beds are mainly confined to the lower part of the Abo Formation or its equivalents, are largely represented by channel fills of relatively coarse siltstones and sandstones, and may have been deposited primarily in braided channel belts, with wet braid plains. Nineteen taxa occur at $\geq 10\%$ of the sample sites. None of these, individually, are supaioids. Conifers as a group occur at 73% of sample sites; *Walchia piniformis* (50%), *Culmitzschia speciosa* (23%), *Otoviccia hypnoides* (10%), *Culmitzschia americana* (identified in 1938 by Rudolf Florin as *Lebachia americana*, according to collection notes; 10%) and *Ernestiodendron* (3%). Cordaitalean foliage occurs at 40% of sampling sites. Supaioids, collectively, occur at 10% of the sampling sites: *Supaia thinnfeldioides* (7%) and cf. *Brongniartites* (7%). Seven types of callipterids occur collectively at 37% of sampling sites, including *Autunia conferta*, small (10%) and large (3%) forms, *A. naumannii* (3%), *Rhachiphyllum lyratifolia* (10%), *R. schenkii* (20%), *Lodevia* sp. (13%), and *Dichophyllum flabellifera* (3%). The lyginopterid pteridosperm *Sphenopteridium* cf. *manzanitanum* (10%), not present in the red beds, is present at 10% of gray beds sites. Possible cycadophytes include *Taeniopteris*, which occurs at about 14% (combining two forms) of sites, whereas it is present at no more than 2% of red beds localities, and *Plagiozamites*, a single occurrence (3%). At gray beds sites, medullosan pteridosperms also are common elements, including *Odontopteris* sp. (20%), *Mixoneura* sp. (3%), *Neurodopteris auriculata* (10%), *Neuropteris* sp. (10%), *Callipteridium* (7%), *Alethopteris* sp. (10%), and *Pseudomariopteris* (3%). In addition, the tree fern foliage *Pecopteris* is present at 20% of sites examined, whereas it is present at only 2% of red beds sites. Together with the pecopteroids, the most reliable indicator of local wet landscape areas are the calamitaleans. This group is represented by foliage forms *Annularia* sp. and *Asterophyllites equisetiformis*, each present at 6% of collecting sites, and by calamitalean stems, which occur at 33% of sites examined. The lycopsid *Sigillaria brardii*, almost certainly an indicator of swampy landscape conditions, is entirely absent from the red bed samples, but occurs at 13% of the gray facies sample sites.

There also are several sites in the lower Abo Formation that can be described as dolomitic limestones containing plant fossils. Only one of these sites was examined in detail, "Site Flood" from the Abo equivalent, Community Pit Formation (see Fig. 2), of the Robledo Mountains (Lucas et al., 2013). This flora is entirely different from any other found in the Abo Formation or its equivalents. The most common elements in the flora are a conifer that is likely a previously undescribed voltzialean (Cindy Looy, personal communication, 2012) and the callipterid *Dichophyllum flabellifera*. Also present, but uncommon, are cordaitalean leaves, and *Annularia spicata* calamitalean foliage.

Illustration of the Flora

Representatives of the flora of the Abo red beds are illustrated in

Figures 3-18, approximately in the order of importance of the taxa in terms of their relative frequency of occurrence.

Figures 3 through 5 are conifers. The taxonomy of conifers is among the most difficult among late Paleozoic fossil plants, rivaling that of the marattialean ferns. Thus, the names below are effectively place holders. Conflicts in taxonomic practice (Visscher et al., 1986; Mapes and Rothwell, 1991) have, in some instances, created communication difficulties with regard to these plants. Because we are dealing nearly entirely with vegetative remains, lacking cuticle and, in many cases, detailed surface features of any kind, we have adopted a modified system based largely on that of Visscher et al. (1986) and Broutin and Kerp (1994). *Hermitia schneideri* is not illustrated; originally described as *Walchia schneideri*, it was transferred to *Hermitia* by Visscher et al. (1986).

Figure 3. *Walchia piniformis*

Conifer specimens identified as *W. piniformis* are the most commonly encountered plant fossils in the Abo Formation. These are for the most part fragmentary, but entire branches up to 80 cm in length, and large branch fragments, such as those illustrated, are common. These specimens have small, narrow, elongate leaves that curve inwardly acropetally.

Figure 4. *Otoviccia hypnoides*/cf. *Brachyphyllum*

A large number of conifer specimens have small leaves of triangular shape that are adpressed to various degrees to the axis. In some specimens, these leaves are so closely adpressed that they may be difficult to distinguish in mold-cast type preservation. The specimens illustrated in Figure 4.1-4.3 illustrate this condition. Such specimens are similar to those that Mamay (1967) designated *Brachyphyllum? densum*, from the Lower Permian of north-central Texas, beds that may correlate to some part of the uppermost Abo Formation. In other instances, such as the specimen illustrated in Figure 4.4, the leaves are less closely pressed against the supporting branch, though they are still generally triangular in shape. This is the more common morphology of the two, which we have designated *Otoviccia hypnoides*. Figure 4.5 is a specimen that probably belongs to *O. hypnoides*, but illustrates some of the difficulties encountered when identifying these fragmentary specimens.

Originally described as *Walchia hypnoides*, this species was re-named and transferred to *Otoviccia* on the basis of reproductive morphology. It differs from *Walchia* in having two fertile scales per dwarf-shoot; sterile and fertile scales are morphologically very similar, unlike the pattern in *Walchia*, where they differ in form (Kerp et al., 1990).

Figure 5. *Culmitzschia americana* and *Culmitzschia* cf. *speciosa*

Housed in the USNM collections are specimens from the Pennsylvanian-Permian boundary in central New Mexico identified in 1938 by Rudolf Florin as *Lebachia americana*; this species was later transferred to *Culmitzschia* (Clement-Westerhof, 1984). It is not certain if Florin examined these specimens in Washington, DC, or if they had been sent to him for assessment. Specimens of the type illustrated in Figure 5.1 and 5.2 were designated by Florin (1939) as *Lebachia americana*. We use the name *C. americana* here as a designation for this morphology, characterized by conifer stems bearing relatively small, straight, needle-like leaves. These leaves rarely show any acropetal inward curvature and, in fact, may arc slightly away from the stem instead of toward it, as illustrated in Figure 5.1. The specimen illustrated in Figure 5.2 also demonstrates preservation in a thin mud layer, or "drape" that also partially covers the branch fragment.

Figure 5.3 illustrates the most robust type of conifer foliage found in the Abo. Leaves are relatively large and curve inward at their tips acropetally. We have designated such specimens *Culmitzschia* cf. *speciosa*.

Neither of these forms are common in the Abo Formation red beds.

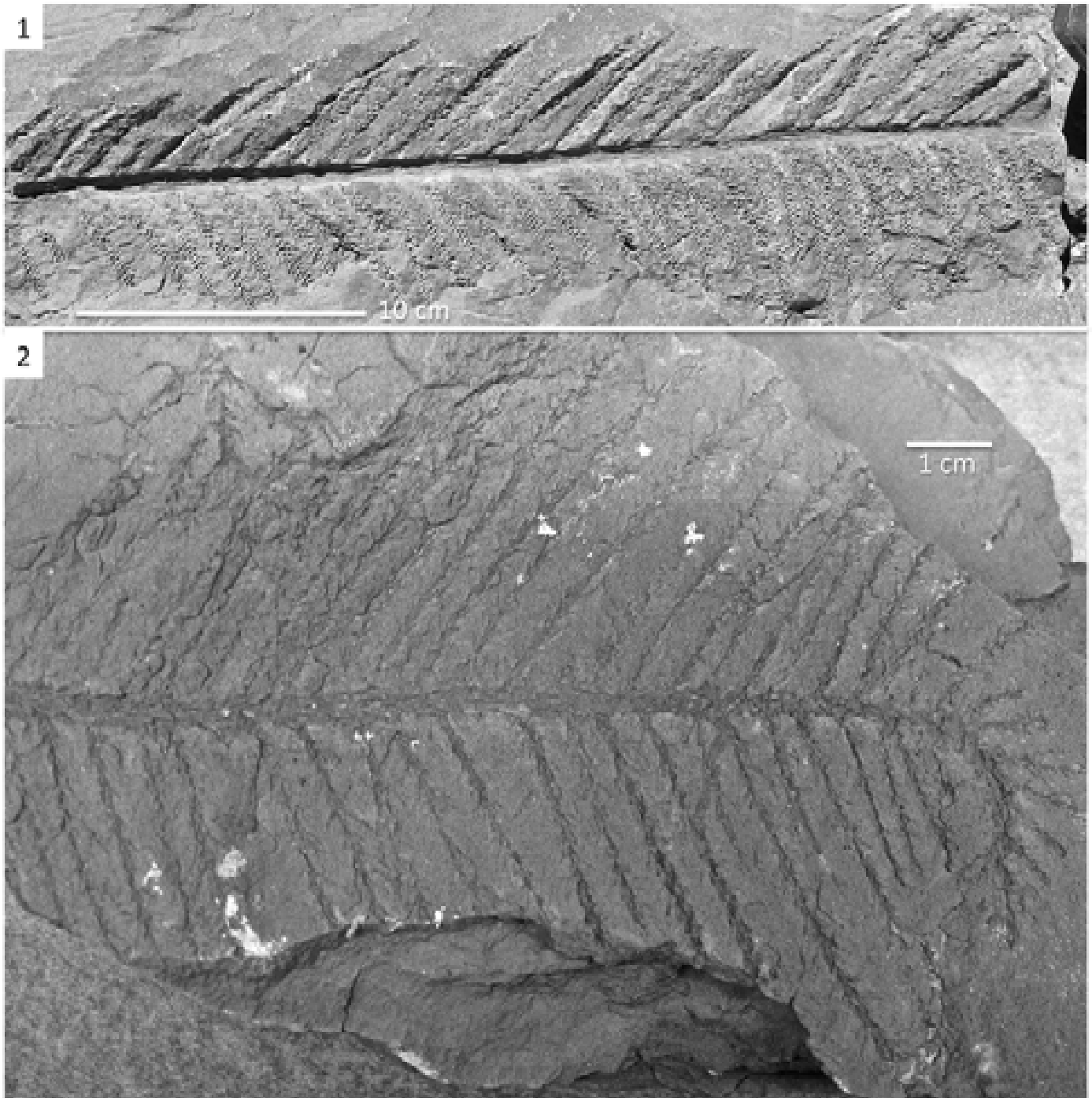


FIGURE 3. *Walchia piniformis*. 3.1, USNM specimen 558251; USNM locality 43557. 3.2, USNM specimen 539454; USGS locality 8979.

Figure 6. *Supaia thinnfeldioides*

Next to the conifers, the supaioid peltasperms, particularly *S. thinnfeldioides*, are among the most commonly encountered plants in the Abo Formation red beds. *Supaia thinnfeldioides* is characterized by small, forked fronds (Fig. 6.1), in which the venation is often obscure, perhaps due to the thickness or leathery nature of the leaf in life (e.g., Figs. 6.1 and 6.3). In some instances, however, the venation is visible and is characteristically arcing at a high angle (Fig. 6.2). Many species of *Supaia* were described by White (1929) from deposits of the Hermit Shale, in the Grand Canyon of Arizona. Many of these may be varieties of a single species; we have been very cautious in making determinations based on

White's original work and have treated the wide range of morphological variation found in Abo specimens as *S. thinnfeldioides*, which is the most common and abundant of White's species in the Hermit Shale collections.

Figures 7 and 8. *Supaia anomala*

In his 1929 Hermit Shale monograph, White described specimens as *Supaia anomala* that were quite different in overall morphology from the many other species of *Supaia* he described. As discussed in DiMichele et al. (2007), *S. anomala* appears to be a chimera of at least two forms, one of which, the most diagnostic for this species, likely falls outside the

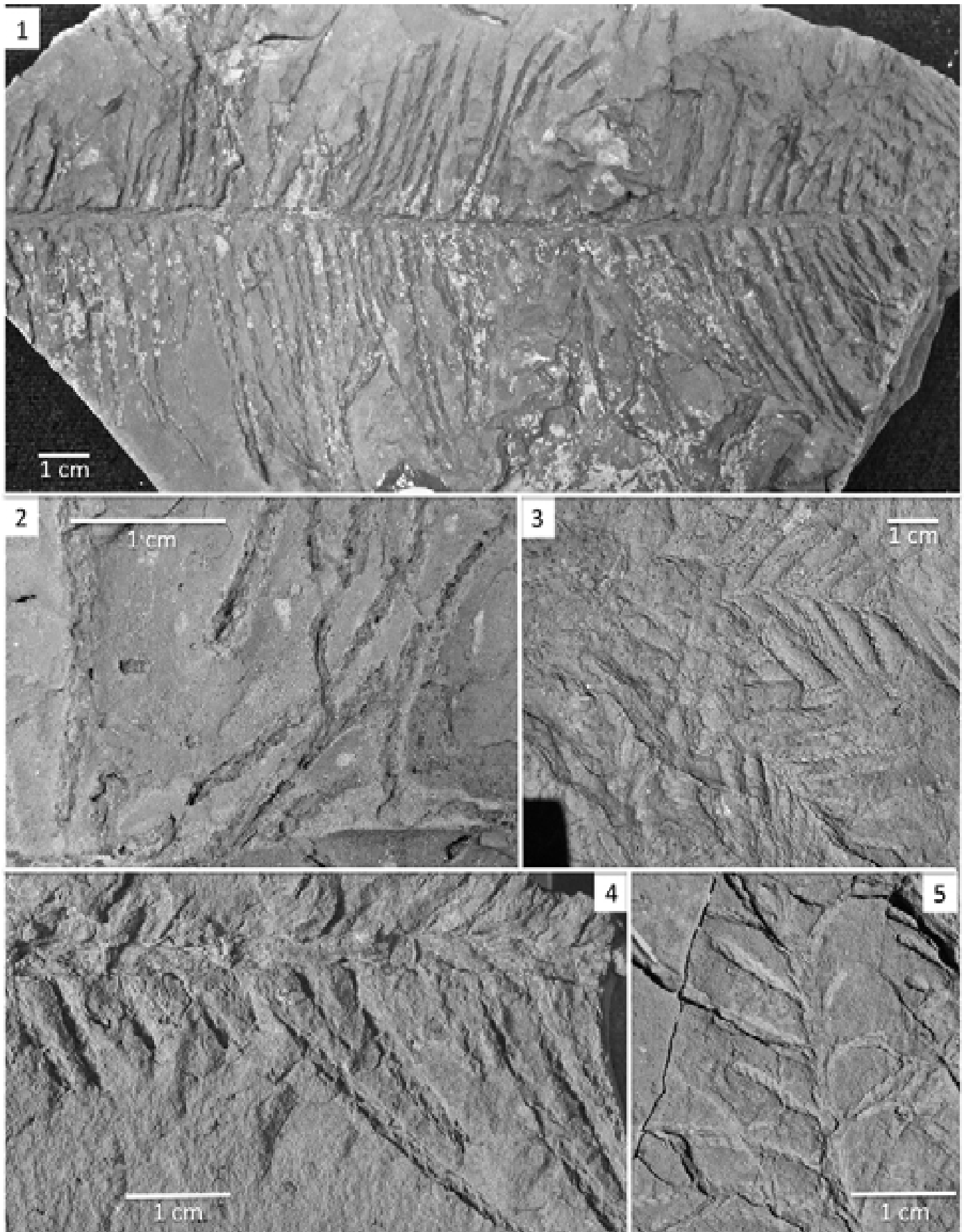


FIGURE 4. cf. *Brachyphyllum* sp./*Otovicia hypnoides*. **4.1**, USNM specimen 558269; USGS locality 8981. **4.2**, USNM specimen 558270; USNM locality 43441. **4.3**, Field Photograph; USNM locality 43559. **4.4**, NMMNHS P-42802; NMMNHS Cañoncito de la Uva locality. **4.5**, Field Photograph; USNM locality 43560.

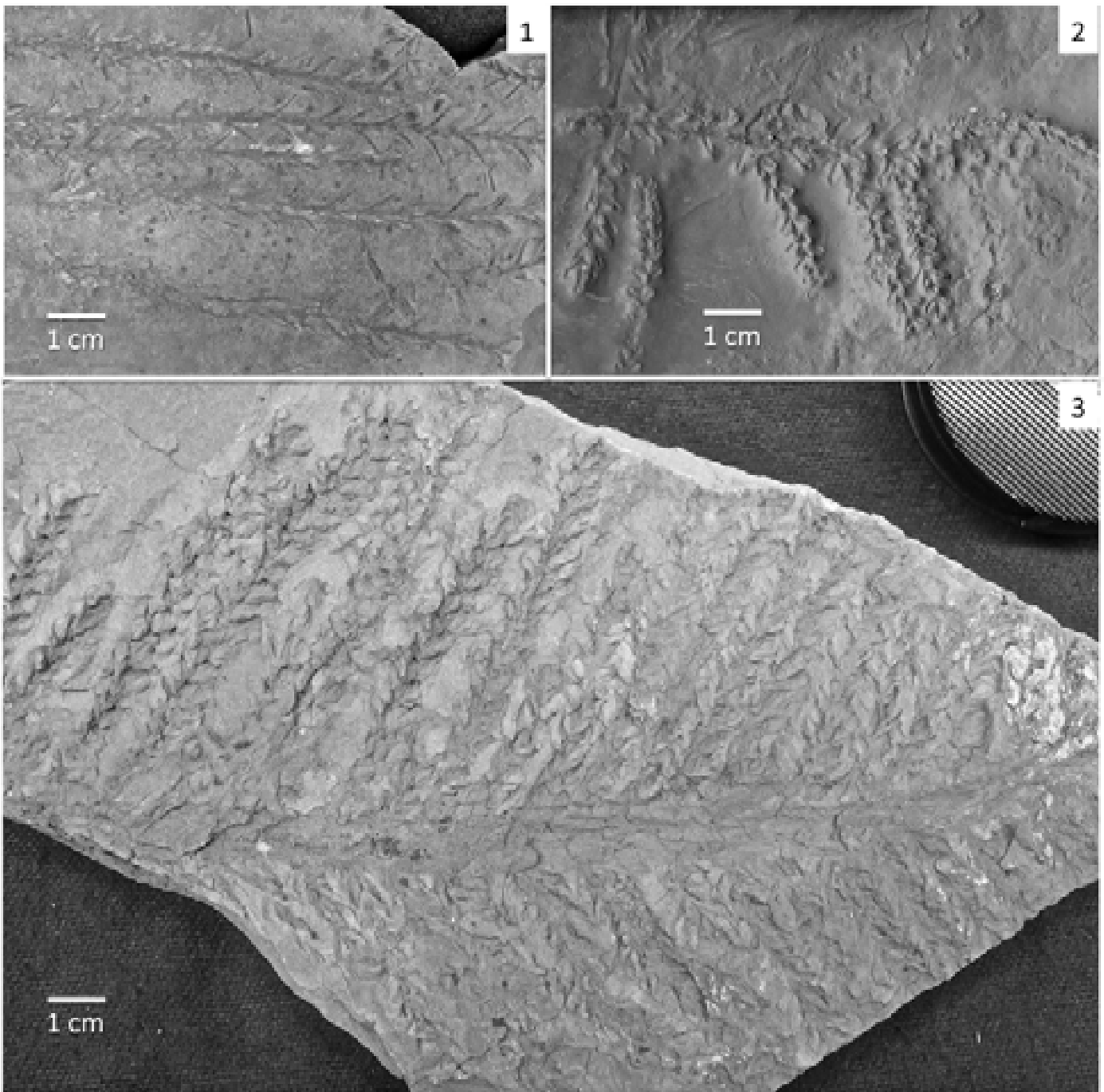


FIGURE 5. *Culmitzschia americana*. **5.1**, USNM specimen 558272; USGS locality 8993. **5.2**, USNM specimen 558299; USNM locality 42340; *Culmitzschia* cf. *speciosa*. **5.3**, USNM specimen 558271; USNM locality 43429.

generic concept of *Supaia*. This leaf-type consists of a ribbon-like lamina that flanks the petiole and all orders of branching of the rachis. The leaf is forked at the top of the petiole (the length of which we have not been able to determine) into two mirror-image segments (Figs. 7.1, 8). Divisions of each of the resulting primary rachises produce a pseudomonopodial arrangement with lateral, ribbon-like secondary rachises/pinnae departing at high angles. In some cases, these rachises/pinnae may again fork (Fig. 7.2). As seen in Figures 7.2 and 8, venation of the laminae is arching and slightly S-shaped.

Figure 9. *Autunia conferta*

Callipterids are the third most common major group of plants in

the Abo red beds, and the majority of these occurrences are of *Autunia conferta*. This iconic callipterid, formerly treated as an index species for the base of the Permian, is widespread across Europe and North America in rocks of latest Pennsylvanian and early Permian age. Most of the specimens in the Abo have been found in isolation and in the uppermost portions of the unit. Two morphologies occur that can be attributed to this species. Kerp (1988) described and illustrated a large range of variation in this species; large population sizes suggest overlap between the larger and smaller forms but, nonetheless, these are clear size centroids that are more or less common in different regions of Pangea. For example, in Europe, the Dunkard Group of the Appalachian Basin, and the Abo Formation of New Mexico, the small form of this species is by far the

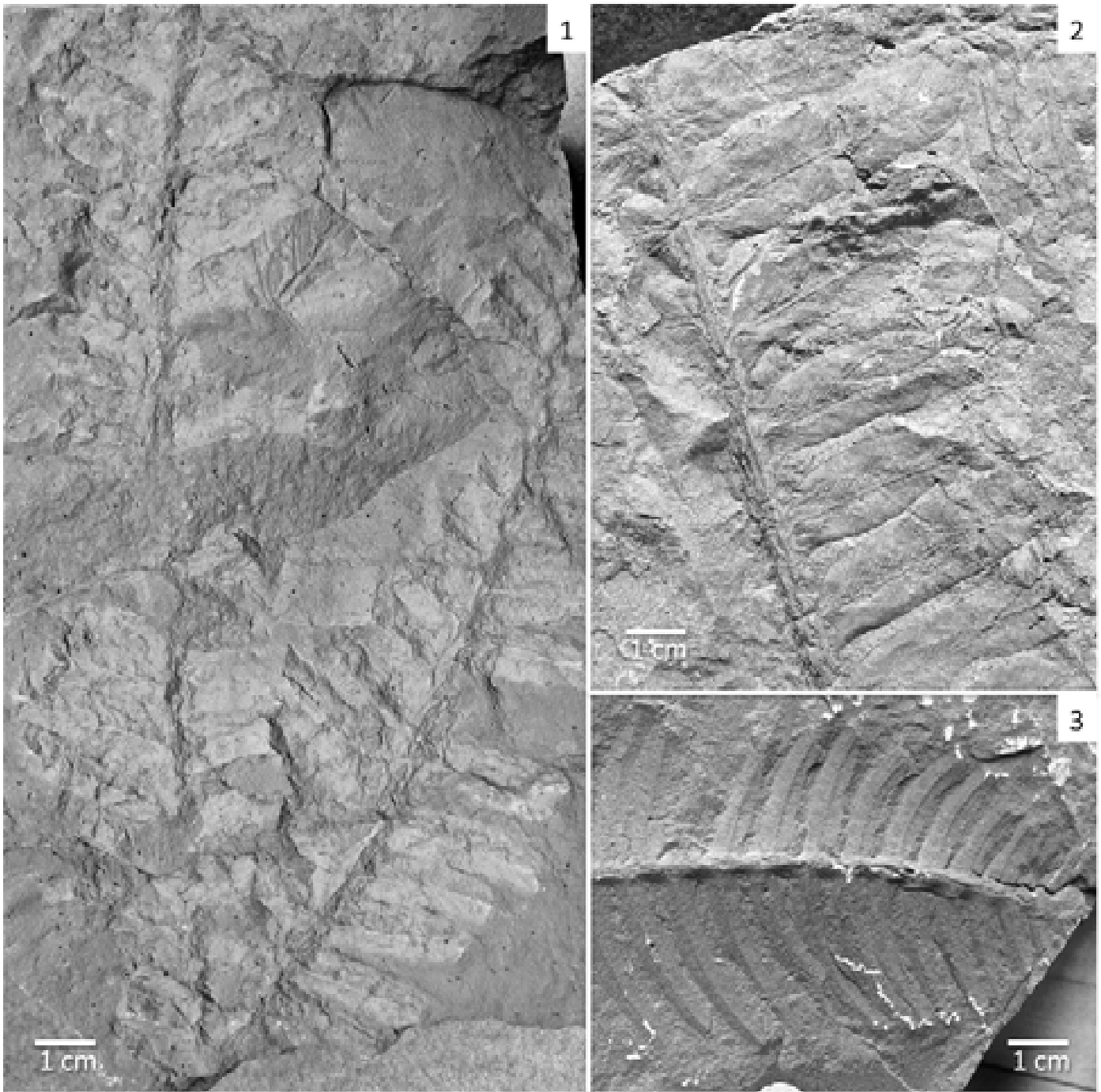


FIGURE 6. *Supaia thinnfeldioides*. 6.1, USNM specimen 558273; USGS locality 8979. 6.2, USNM specimen 558274; USNM locality 42110. 6.3, USNM specimen 558252; USGS locality 8977.

most commonly encountered. However, in the north-central Texas Pennsylvanian-Permian sequence, the large form is almost exclusively found. The specimen illustrated in Figures 9.1 and 9.2 is very small but could be part of a frond of a young plant. Differences in size in *A. conferta* specimens are mostly related to their habitat; those growing in relatively humid conditions appear to have had larger, sometimes pinnatifid pinnules, whereas those growing in more water-stressed environments appear to have had smaller pinnules (Kerp, 1988).

Figures 9.1-9.3 illustrate fragmentary specimens of the small form of *A. conferta*. The white arrow in Figure 9.2 points to rachial pinnules characteristic of *A. conferta*. Note also the prominent midveins, indica-

tive of vaulting of the lateral pinnule laminae, regularly triangular pinnule shape, bluntly pointed pinnule apices, and elongate terminal pinnules.

Figure 9.4 illustrates the larger form of what is probably *A. conferta*, though this is still a small specimen, from the terminal portion of a lateral pinna. It differs somewhat from the smaller forms in the slightly acropetal constriction of the pinnule lamina, the more elongate nature of the pinnules, and the angular midvein that appears to cross the pinnule lamina from its point of insertion on the basipetal side to its apical position somewhat to the acropetal side of the lamina. The midveins are prominent, the lamina vaulted, the insertion angular, and the laminar veins steep and relatively straight.

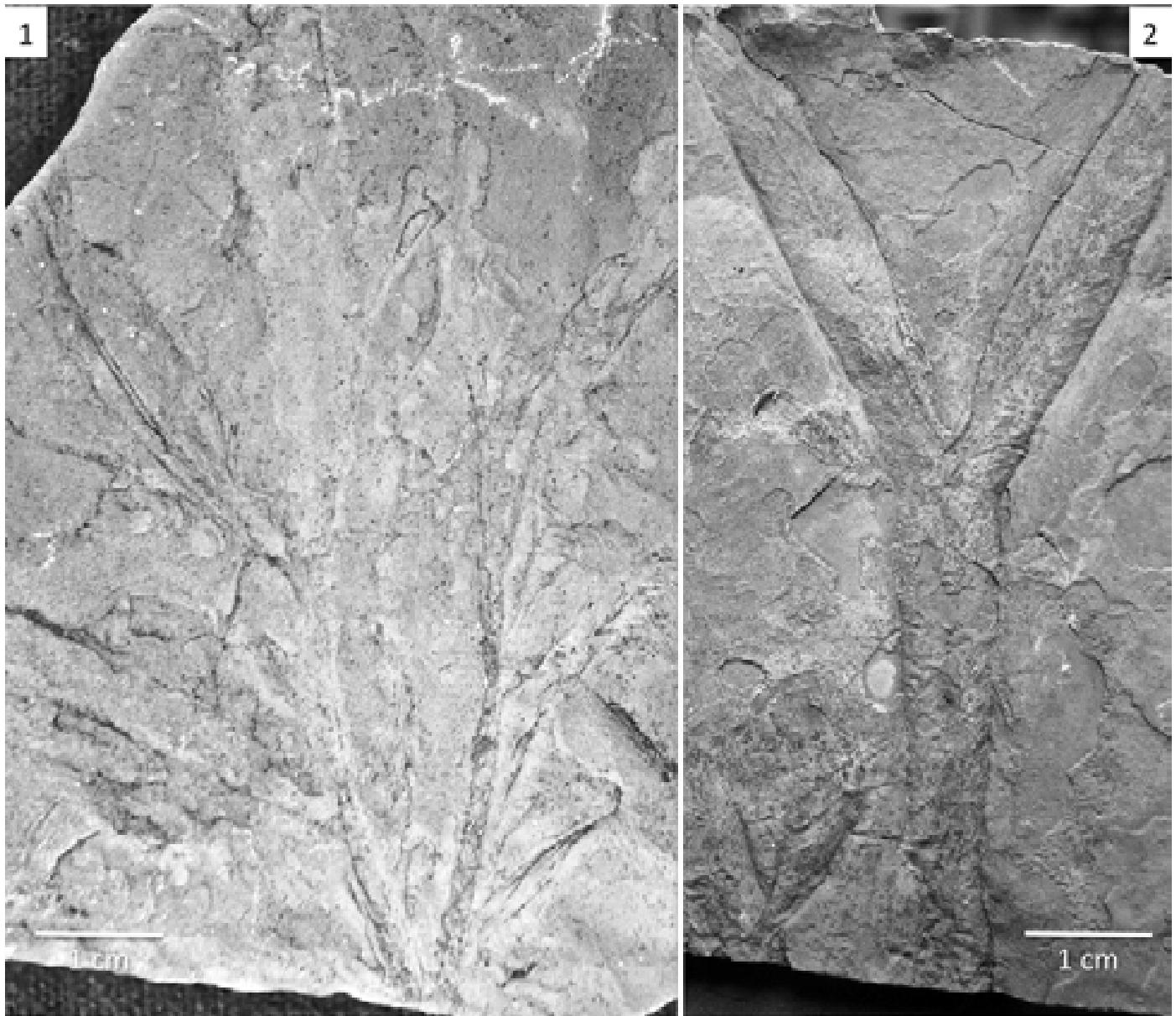


FIGURE 7. *Supaia anomala*. 7.1, USNM specimen 558275; USGS locality 8980. 7.2, USNM specimen 558276; USNM locality 42265.

Figure 10. *Brongniartites* sp. and *Gigantopteridium americanum*

Both of these plants are possibly peltasperms, based on their foliar morphology.

Brongniartites (probably an invalid name – see Naugolnykh, 1999) is part of a complex of likely related “supaioid” plants that can present identification challenges when preserved as fragmentary material (see DiMichele et al., 2005). The specimen illustrated in Figure 10.1, although incomplete, shows the acropetally expanding pinnule size and what appears to be an unforked primary frond axis, typical of what we have classified as *Brongniartites* sp. Another possible identification of leaves of this type is *Glenopteris*, another probable peltasperm with an unforked frond, found in the southwestern United States (Krings et al., 2005).

The *Gigantopteridium americanum* specimen (Fig. 10.2) shows the typical gross architecture of an American gigantopterid. The strong grooves that run slightly angularly through the lamina to the margins mark the position of the main lateral, or secondary veins. The lateral veins can be seen in some parts of the lamina, despite the generally poor

preservation of the specimen (Fig. 10.2, white arrows) and are of a ragged, fasciculate form, typical of *Gigantopteridium*. Suture veins are very difficult to recognize because of the preservation in sandstone.

Both of these taxa are very rarely encountered in the Abo red beds. The gigantopterid was the dominant element at the site where it was collected.

Figure 11. *Comia craddockii*

Comia is another suspected peltasperm that appears as part of a major radiation of this lineage during the later Pennsylvanian and Early Permian (DiMichele et al., 2005). The plant was not identified positively in the red Abo Formation facies. At a single location in the Doña Ana Mountains of southern New Mexico, however, a few specimens of this plant occurred in buff siltstones of the Robledo Mountains Formation (see Fig. 2). The only other reports of its occurrence are from north-central Texas, in the late Wolfcampian or early Leonardian (Sakmarian-Artinskian) (Mamay et al., 2009). The leaf of this plant shows a pinnate architecture, visible in Fig. 11.1. Higher magnification images of that



FIGURE 8. *Supaia anomala*. USNM specimen 558277; USNM locality 42265.

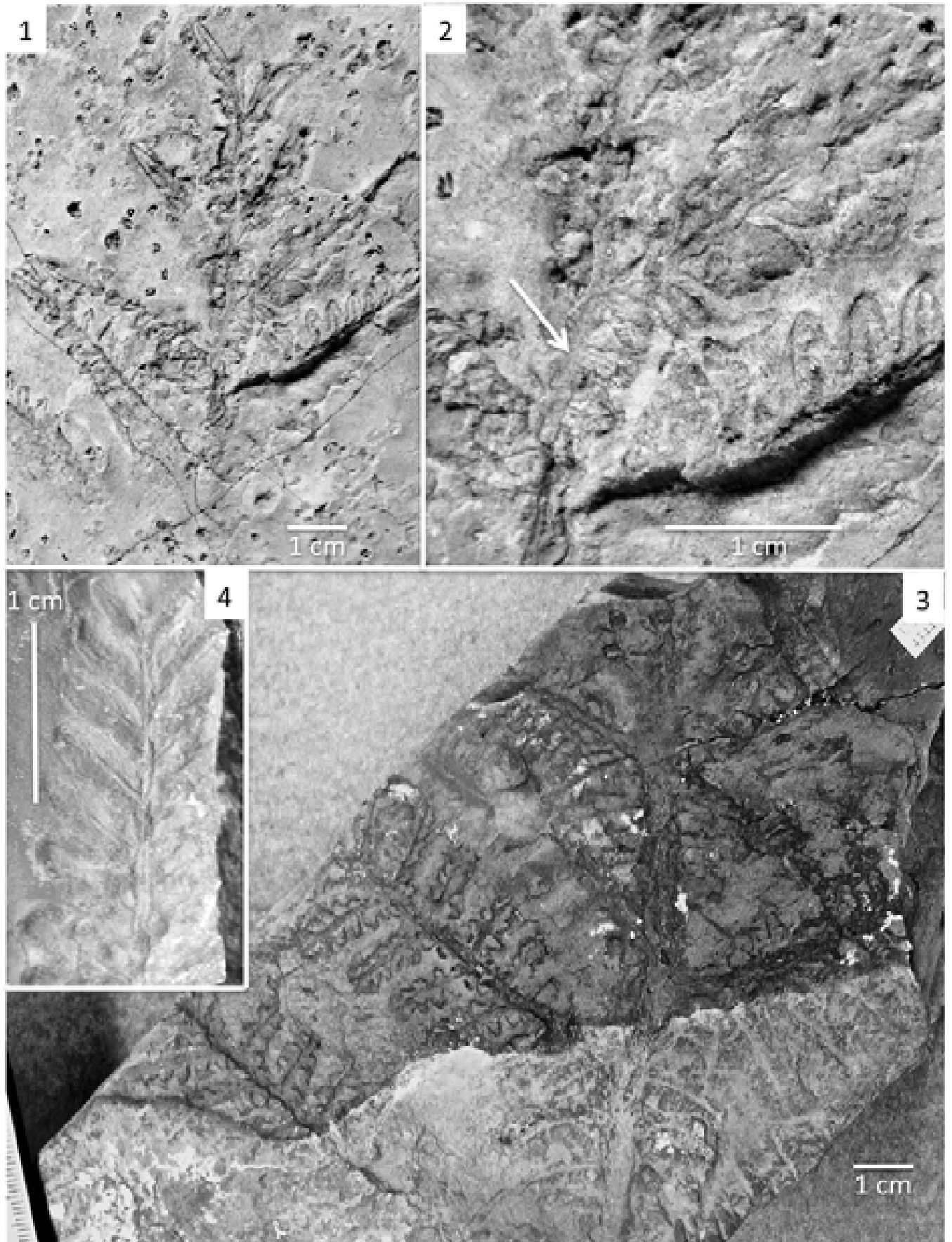


FIGURE 9. *Autunia conferta*. **8.1**, USNM specimen 543955; USNM locality 42251. **8.2**, USNM specimen 543955; USNM locality 42251. **8.3**, USNM specimen 558268; USNM locality 43447. **8.4**, USNM specimen 558278; USNM locality 42264.

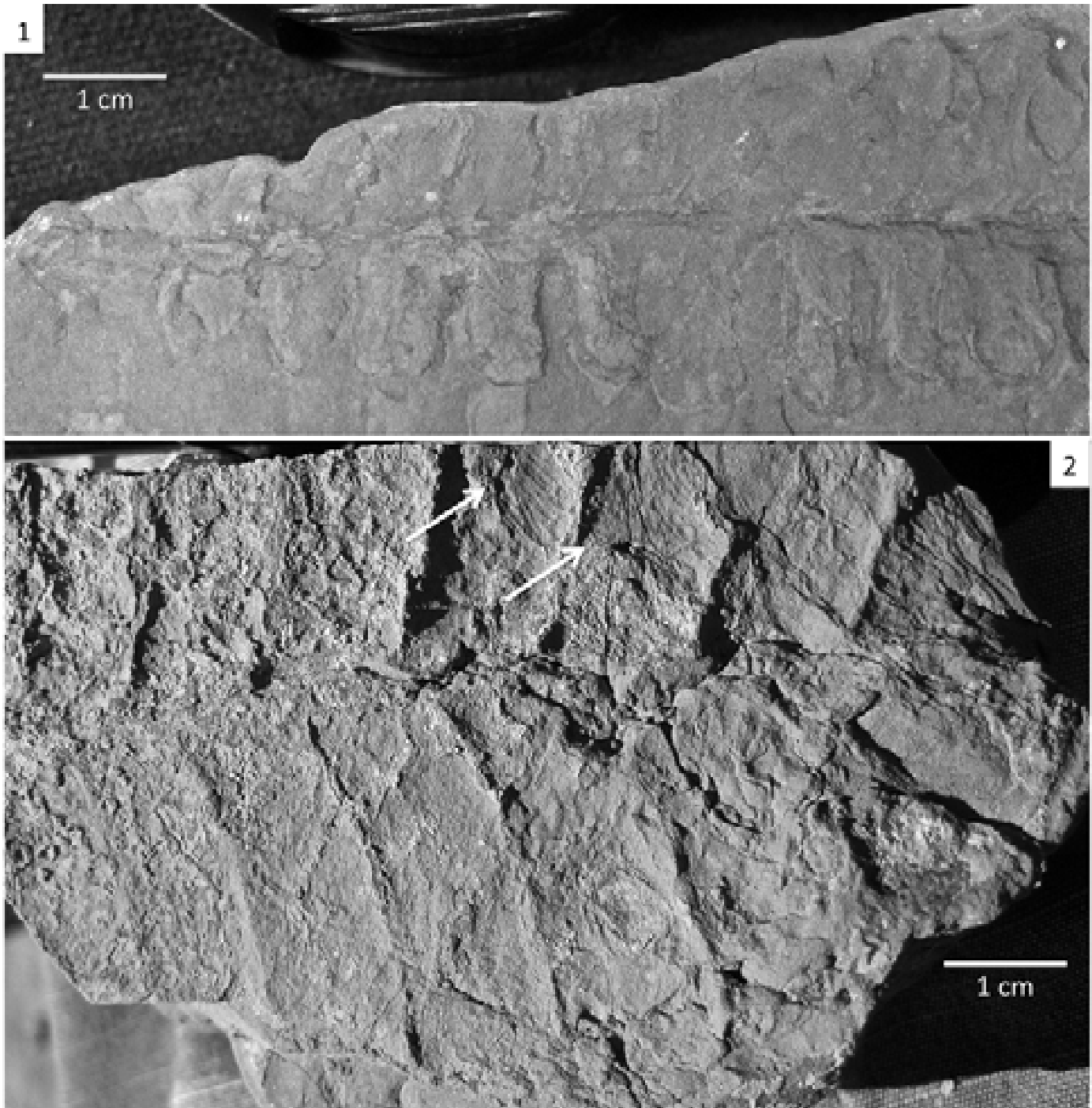


FIGURE 10. cf. *Brongniartites* sp./*Glenopteris* sp. **10.1**, USNM specimen 558280; USGS locality 8980. *Gigantopteridium* sp. **10.2**, USNM specimen 558281; USNM locality 43562.

same specimen (Fig. 11.2) and another (Fig. 11.3) demonstrate the characteristic fasciculate organization of the 4th order veins.

Figure 12. *Taeniopteris* spp.

The taxonomy of *Taeniopteris* below the generic level is suspect; our own attempts to sort out species morphometrically, based on large collections, and using features of venation and leaf shape, have failed to define groups that can be repeatedly recognized in the field or in collections. Overlap among various measurable characters is high and inconsistent from specimen to specimen. There are a number of described species in the genus (see discussion in Wagner and Martinez Garcia,

1982), and attempts have been made to bring order to these (Remy and Remy, 1975). The genus includes both ferns and seed plants, although the larger forms all appear to be of seed-plant affinities.

There are several kinds of *Taeniopteris* in the Abo Formation red beds, judged by the gross morphology of the leaves. None are common. Figure 12.1 is noteworthy because several *Taeniopteris* leaves appear to be attached to a common axis in a regular helix and to be apetiolate; in our experience, this is the only documented occurrence of such attachment in American specimens. Figure 12.2 shows a typical leaf apex with strongly developed, straight lateral veins. The significance of this trait, however, is uncertain, given that angle of veins can vary strongly among specimens

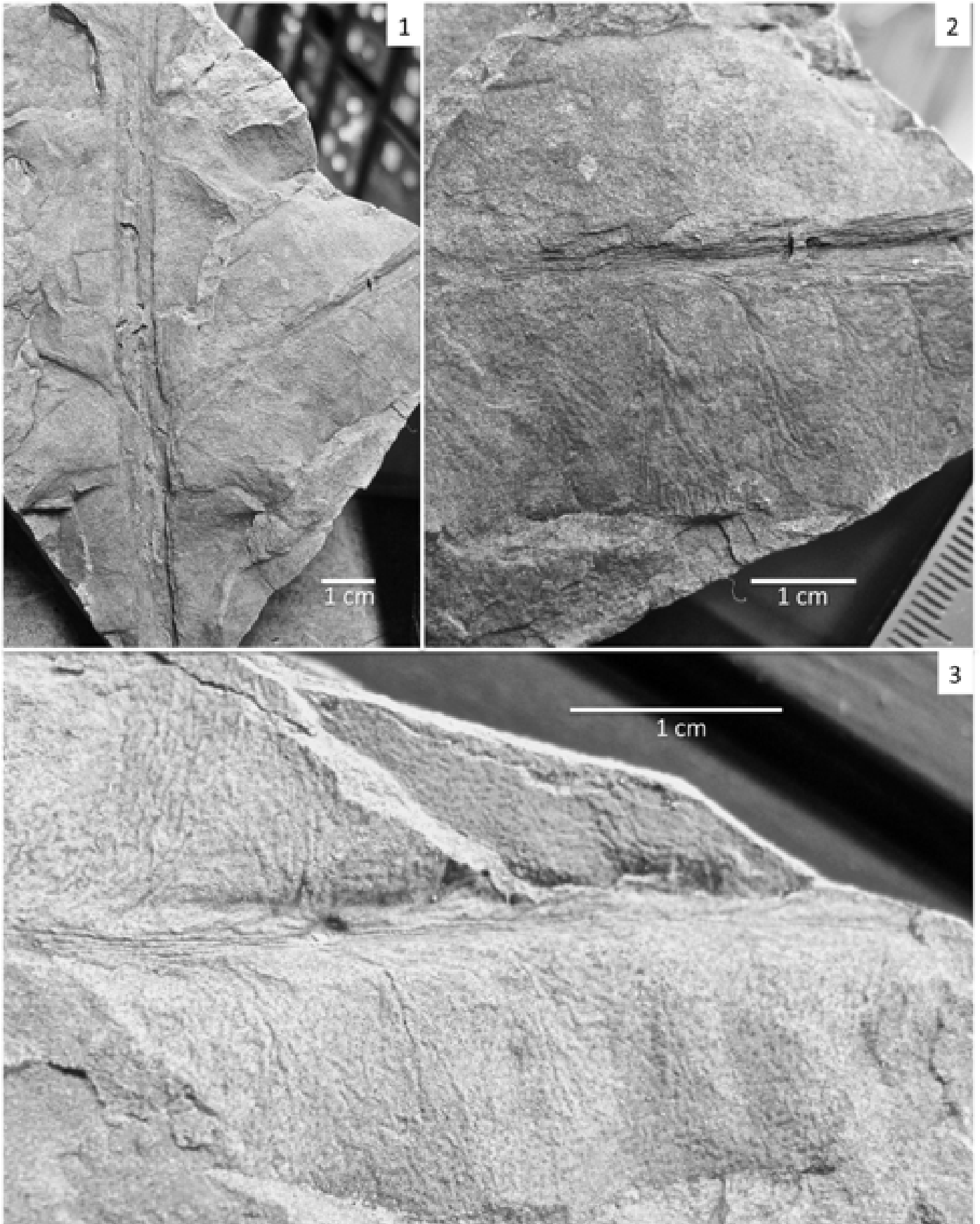


FIGURE 11. *Comia craddockii*. **11.1**, USNM specimen 558282; USNM locality 42271. **11.2**, USNM specimen 558282; USNM locality 42271. **11.3**, USNM specimen 558283; USNM locality 42271.

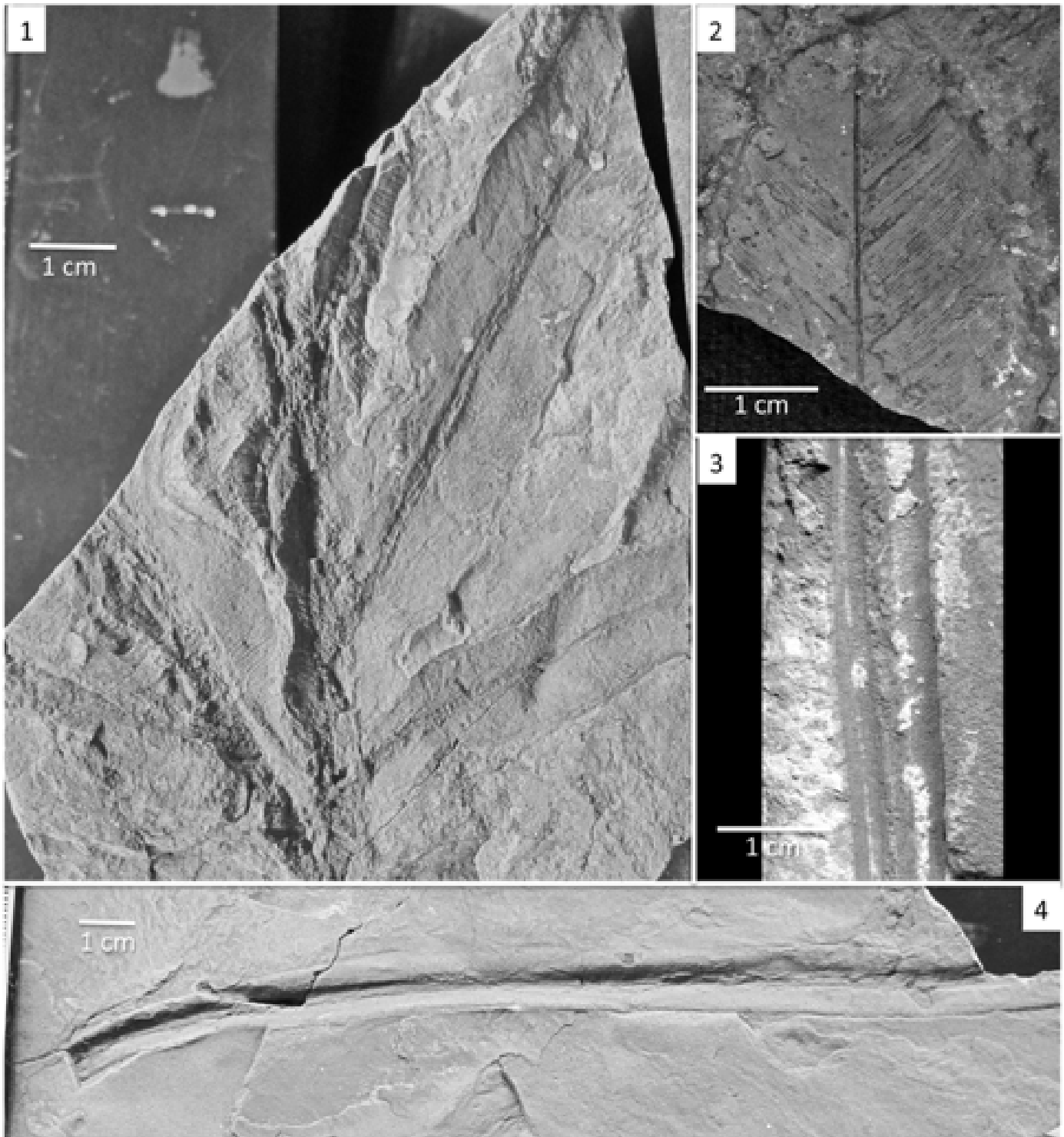


FIGURE 12. *Taeniopteris* spp. **12.1**, USNM specimen 543956; USNM locality 42255. **12.2**, USNM specimen 558284; USNM locality 43435. **12.3**, USNM specimen 558285; USNM locality 42255. **12.4**, USNM specimen 558286; USNM locality 42255.

of *Taeniopteris* from the same fossil population (one, sedimentologically constrained deposit at one collecting site), as can the number of vein bifurcations. The specimens illustrated in Figures 12.3 and 12.4 are part of the same assemblage as is the specimen illustrated in Figure 12.1. The veins in these leaves are faint, but can be seen in Figure 12.3, where they are slightly angular in their insertion and very straight. As with the leaves in the specimen illustrated in Figure 12.1, they appear to be apetiolate and to have a marked basal abscission zone (Fig. 12.4).

Figure 13. cf. *Glossopteris/Lesleya*

Undoubtedly, the most perplexing specimens found in the Abo red beds are from a site or proximate sites in the Zuni Mountains in northern New Mexico. The collections in the USNM were made in the first half of the 20th century by the University of New Mexico, and later by the U.S. Geological Survey from places we have not been able to relocate exactly. They are preserved in red, fine-grained sandstone that appears to have been deposited rapidly; some of the specimens have

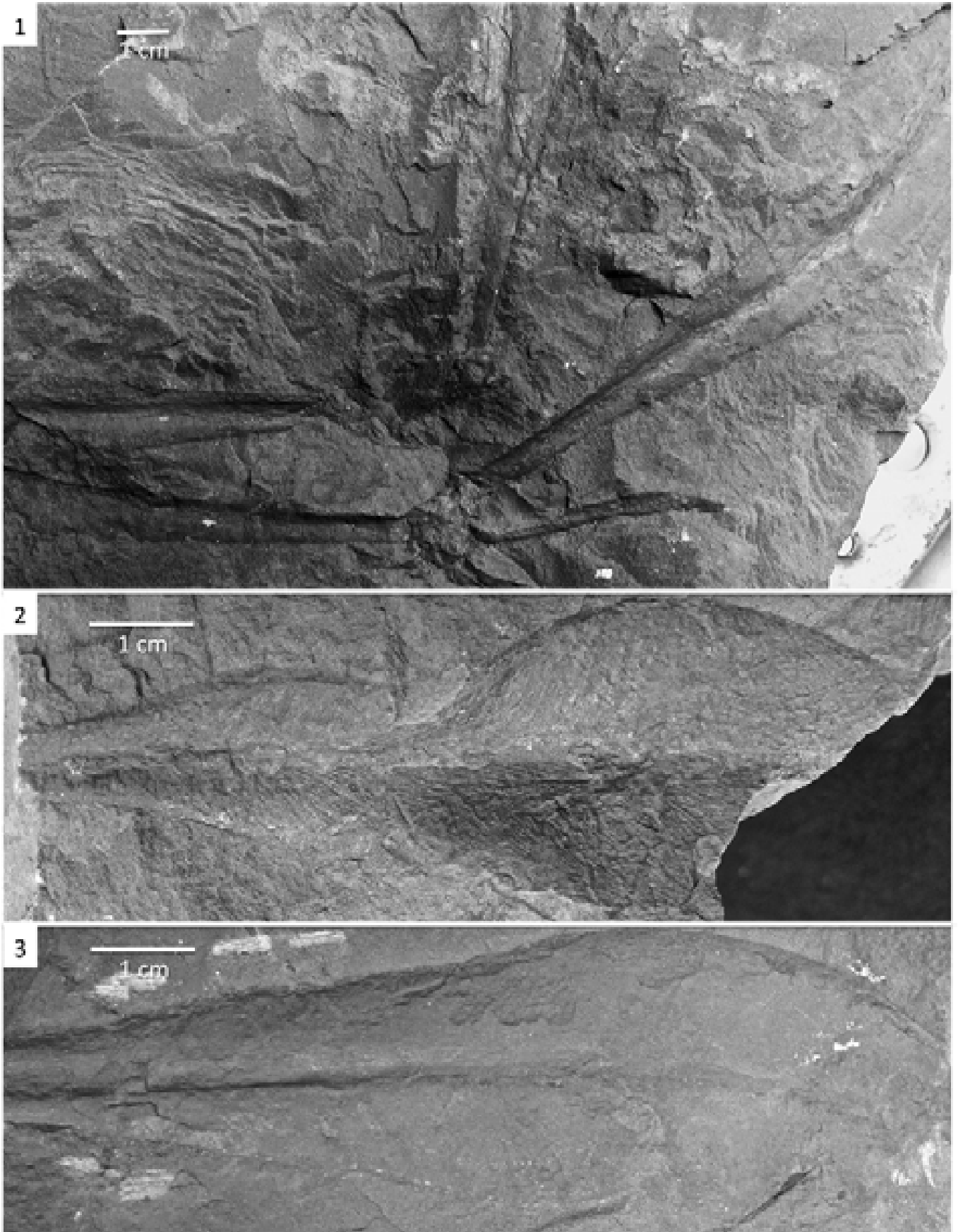


FIGURE 13. cf. *Glossopteris/Lesleya* sp. **13.1**, USNM specimen 559287; USNM locality 43680. **13.2**, USNM specimen 558288; USNM locality 43680. **13.3**, USNM specimen 588289; USNM locality 43680.

leaves still in attachment to stems, or branches, reflecting either burial of branches broken off in storms, or burial of small plants in situ (Fig. 13.1). In either instance, this further suggests that the plants were living in environmental settings close to actively flowing streams. As can be seen from Figures 13.1-13.3, the leaves are large, apetiolate, spatulate in shape with entire margins, rounded apices, and a strong midvein that runs from the base to the tip of the leaf. Lateral veins, though not well preserved, can be seen in Figure 13.2; these appear to be arching, open-dichotomous, and multi-forked. The specimens most closely resemble the early and early Middle Pennsylvanian plant *Lesleya* (Leary, 1990) or some forms of the Southern Hemisphere Permian plant *Glossopteris*. The closest comparison is with *Lesleya*, an enigmatic plant that apparently favored moisture-stressed habitats, possibly developed on limestone soils. Like the Abo specimens, *Lesleya* leaves are obovate, apetiolate, and have arching, open dichotomous venation. Although similar in shape to *Glossopteris*, the Abo Formation leaves do not have the reticulate venation that characterizes this genus. Species of *Glossopteris* were dominant in many temperate-climate environments of Gondwana, where the plant has been shown to be seasonally deciduous (e.g., Gould and Delevoryas, 1977). More northerly occurrences also are known from the Permian of Oman (Broutin et al., 1995; Berthelin et al., 2003, 2006) and southeastern Turkey (Wagner, 1962; Archangelsky and Wagner, 1983); both of these peri-Tethyan areas were positioned at higher latitudes during the Permian, and in tropical climates. Both the Turkish and Oman floras are a mixture, mainly of Cathaysian and Gondwanan taxa; in Oman some Euramerican taxa also have been recorded. There also are reports of leaves of *Glossopteris*-type morphology from the Jurassic (Delevoryas, 1969), so, convergence must be considered a real possibility, in the absence of reproductive organs.

Figure 14. *Dicranophyllum* sp. and *Plagiozamites* sp.

These two genera occur very rarely in the Abo red beds, and are elements typical of seasonally dry assemblages.

Dicranophyllum is a coniferophyte (Rothwell and Mapes, 2001). Found mainly as vegetative remains, with foliage often attached to stems, the plants are characterized by narrow leaves that undergo one or more bifurcations (Barthel, 1977; Wagner, 2005). There is only a single vein per leaf, which appears to be the case in the specimens found in Abo sandstones (Fig. 14.1). A relatively large number of species have been described, including *D. readii* from the Missourian-age Kinney Brick Quarry flora in the Manzanita Mountains of Bernalillo County, New Mexico (Mamay, 1981). Wagner (2005) notes, however, that due the rarity of this genus, and its long stratigraphic range (known from the Late Mississippian through the Early Permian), there probably were far fewer species than actually described. Barthel and Noll (1999) have reconstructed the plant as a small to medium sized shrub with a “bottle brush” habit – a mainly straight stem surrounded by a dense covering of leaves. The Abo plant is broadly similar to *Dicranophyllum gallicum*, which is the most commonly found form in Upper Pennsylvanian and Lower Permian deposits.

Plagiozamites is another long ranging, widespread genus, occurring through much of the later Pennsylvanian and into the Permian, across the Pangean equatorial region. Its affinities are most likely with the *Noeggerathiales*, an enigmatic group of heterosporous lower vascular plants (Wang et al., 2009). Only a single specimen is present in the Abo Formation exposures and collections we examined (Fig. 14.2). *Plagiozamites* leaves were compound. The individual pinnules are flat, elongate, and have a broad attachment to the pinna rachis. As visible in the specimen illustrated in Figure 14.2, the veins run parallel to the long axis of the pinnule and terminate at the bluntly rounded, slightly asymmetrical pinnule apex. In well preserved specimens, the curved, apical region of the pinnule is slightly crenulate, with veins ending in the crenulations. This specimen is similar to *Plagiozamites planchardii*, which, though rare, is widely reported in North America, including early Late

Pennsylvanian (Missourian) occurrences in the Appalachian Basin (Bassler, 1916) and New Mexico (Mamay, 1990).

Figures 15, 16, 17. Wetland elements in the Abo red-beds facies.

A small number of taxa typically found in wetland floral associations occur in the Abo red beds. All are rare, but occurrences are widespread.

The most common of these wetland elements are the pectopteroid ferns, two specimens of which are illustrated in Figure 15. The pectopteroids are a very widespread plant group that began significant increases in dominance and diversity during the Middle Pennsylvanian and became very diverse and abundant in wetland floras during the later Pennsylvanian (Cleal et al., 2012). In the Permian, particularly in drier landscapes, the pectopteroids persisted, but became both much rarer and less diverse. They probably grew in streambanks or local swampy areas where substrate moisture remained high, even if overall climate was strongly seasonal (DiMichele et al., 2006).

Also found widely, but generally not identifiable to species, were stem remains (Fig. 16.1) and possible foliar remains (Fig. 16.2) of the bushy scrambling/climbing sphenopsid *Sphenophyllum*. In all instances where such material has been found, it appears to have been buried rapidly in fine-grained sandstone, probably reflecting growth in stream side settings where occasional floods entombed local vegetation. The foliage illustrated in Figure 16.2 is perhaps assignable to the calamitaleans, as a species of *Annularia*; these specimens bear a vague resemblance to *A. spicata* in the fusiform shape of their leaves, though they are oversized for that particular species. In addition, it is possible that the leaves have multiple veins, which would rule out an assignment to *Annularia* and suggest, rather, affinities with *Sphenophyllum*, though no species assignment is possible.

Figure 16.3 illustrates one of only two specimens of *Sphenopteris*, the foliage of filicalean ferns, found in this study. The poor preservation of the specimen is due, in part, to a cover of fine clay, which has obscured the surfaces of the pinnules. Small, ground cover, climbing or epiphytic ferns are essentially absent from the Abo red-beds flora. Rare occurrences suggest that they were present locally on the landscape, but the near absence of these plants over such a large landscape area does not appear to be reflective of a persistent taphonomic filtering effect.

Also vanishingly rare are medullosan pteridosperm remains, a group generally rare in the Lower Permian. Figure 17 illustrates one of the very rare occurrences of foliage possibly attributable to *Neuropteris*, though no definitive species assignment is possible. Attribution to *Neuropteris* is supported by the narrow insertion on the supporting rachis, a distinct midvein that extends about 2/3 of the length of the pinnule lamina, arched lateral veins and the slightly auriculate basipetal margin of the pinnules.

Figure 18. Miscellaneous elements.

Figure 18 is a collection of miscellaneous elements identified in the Abo red beds. Figure 18.1 is one of several specimens, found at a single locality, that appear to be strobili borne at the end of an axis. The axis is obscure, but appears to be small, closely adpressed, and of triangular shape, similar to that we have attributed to *Otoviccia hypnoides* or cf. *Brachyphyllum*.

Small, bivalved or possibly trivalved (one valve may still be buried in the rock) sporangia are illustrated in Figure 18.2. Although uncommon, these sporangia occur widely, in large numbers on bedding planes, and often in association with callipterid foliage. No attachment has been identified, however, so, at present, their affinities remain unknown.

Calamitalean remains are nearly absent in the red beds facies. Figure 18.3 is a suspect calamitalean stem, given its strong ribs and suggestion of a node at the far left hand end. Given its relatively small size, however, it also may be a *Sphenophyllum* stem (or not a sphenopsid at all).

The flabellate, evenly dichotomized specimen illustrated in Figure

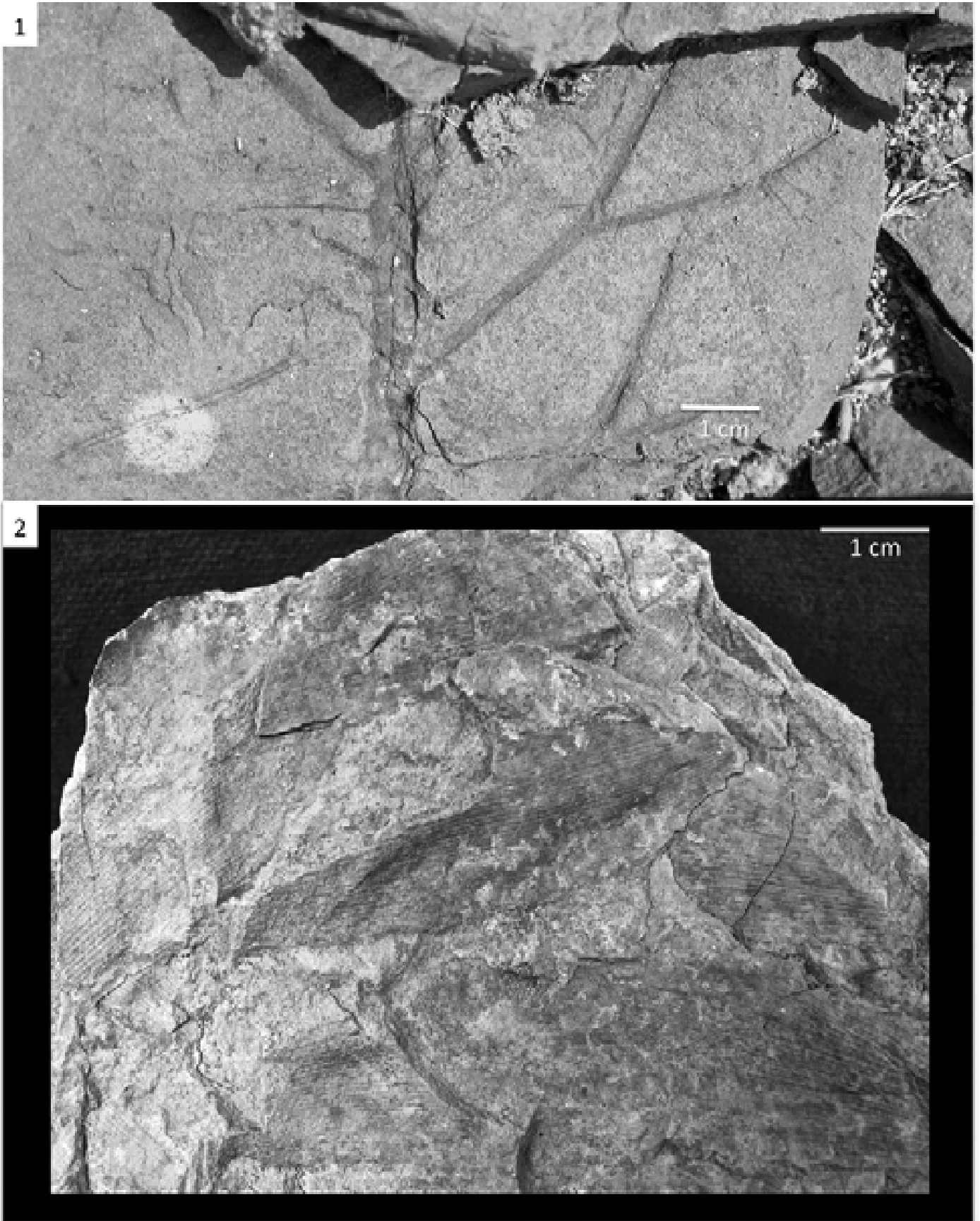


FIGURE 14. *Dicranophyllum* cf. *gallicum*. 14.1, Field photograph; USNM field locality NM2012-03. *Plagiozamites* cf. *planchardii*. 14.2, USNM specimen 558290; USNM locality 43452.

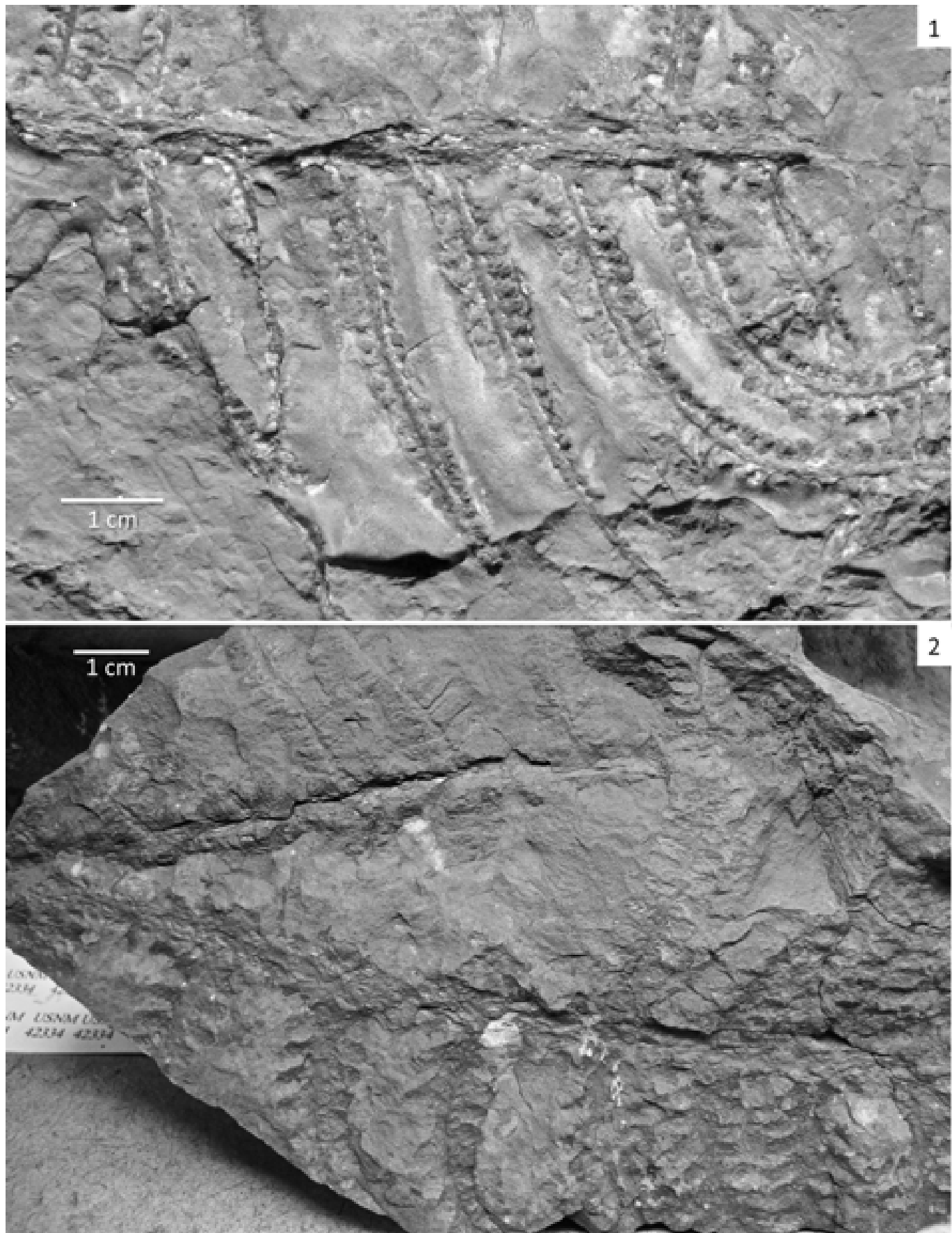


FIGURE 15. *Pecopteris* sp. **15.1**, USNM specimen 558291; USNM locality 42263. **15.2**, USNM specimen 558292; USNM locality 42334.

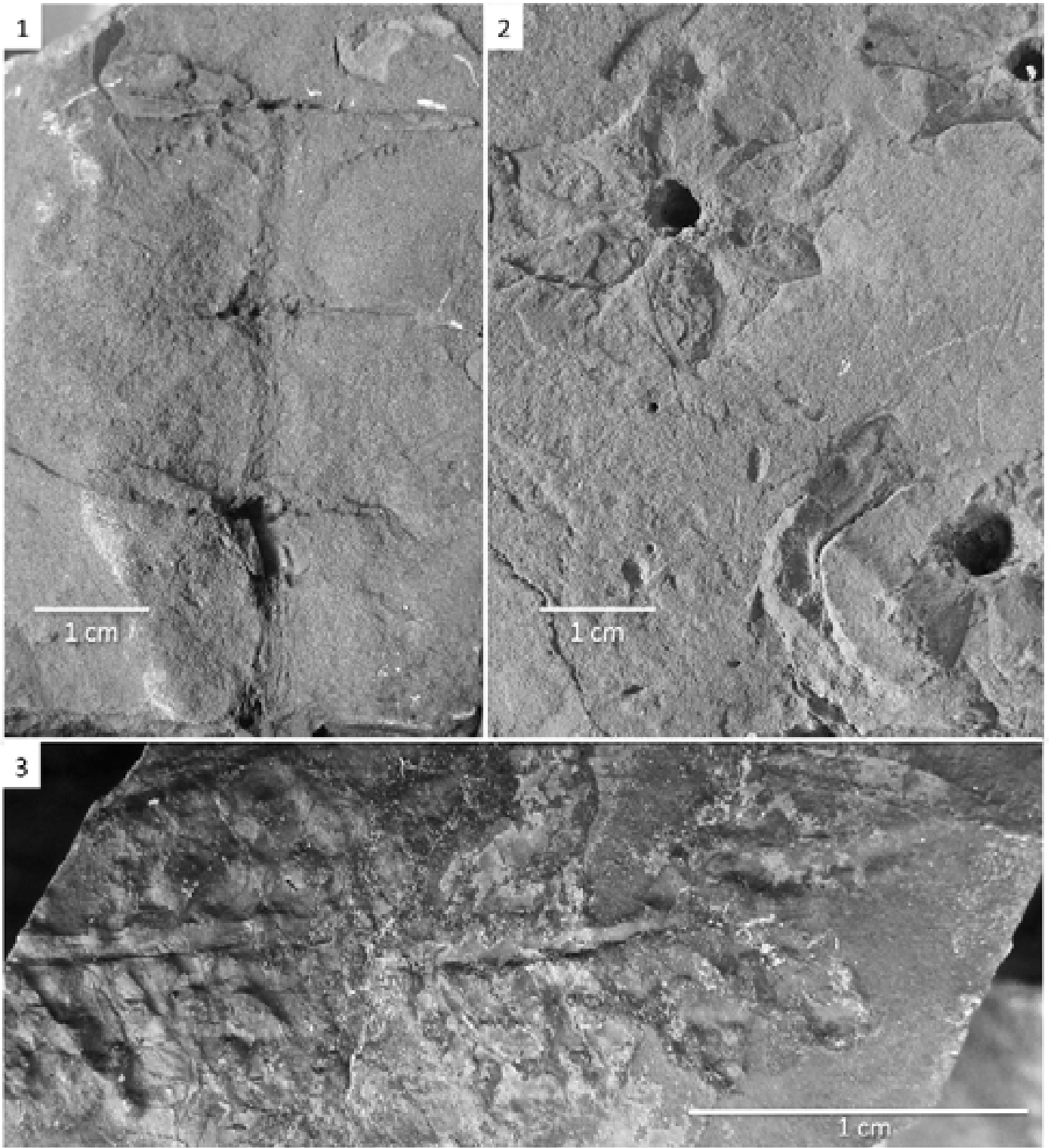


FIGURE 16. *Sphenophyllum* sp. **16.1**, USNM specimen 558293; USGS locality 8952. **16.2**, USNM specimen 558294; USGS locality 8952. *Sphenopteris* sp. **16.3**, USNM specimen 558295; USNM locality 42264.

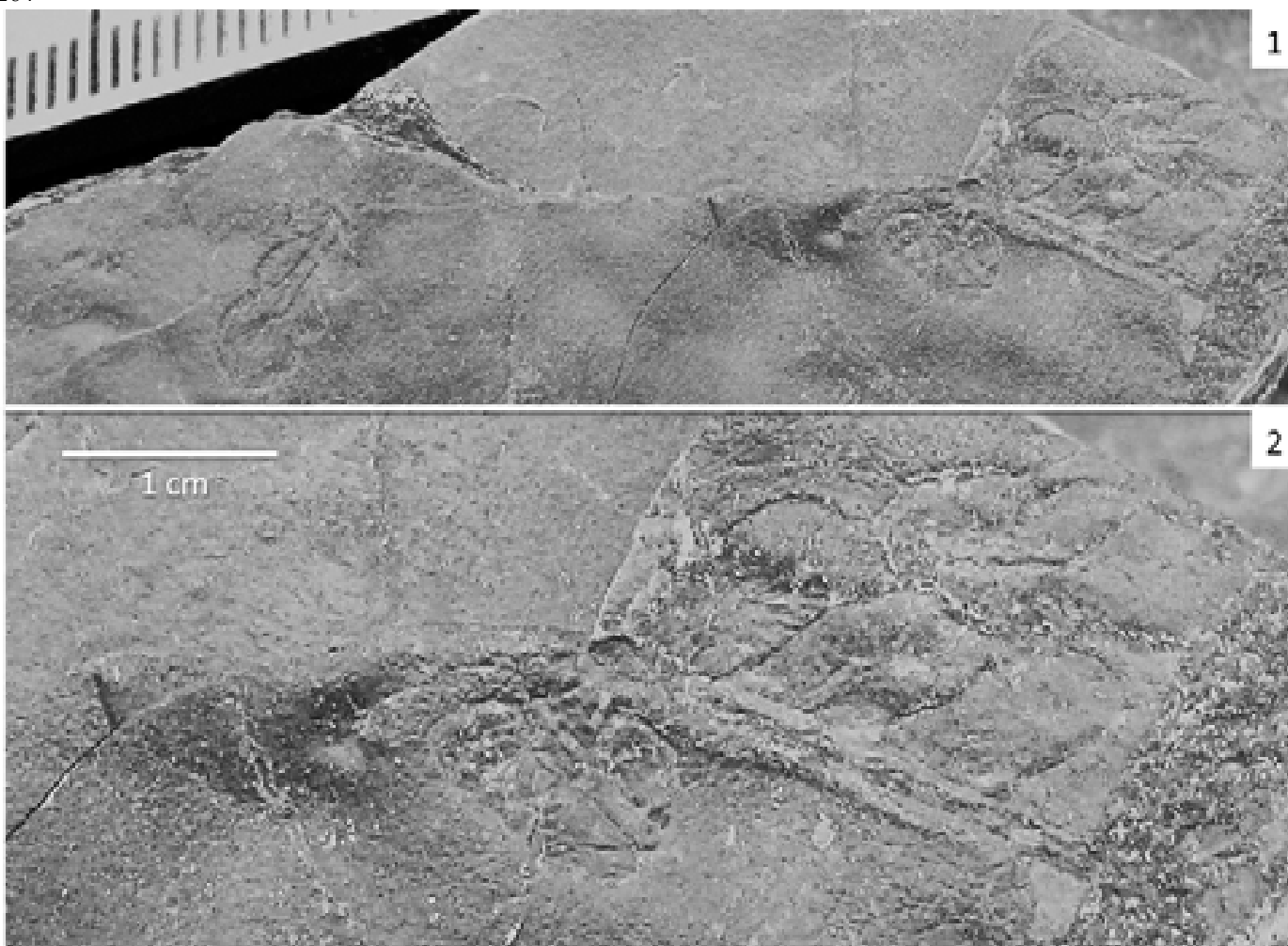


FIGURE 17. *Neuropteris* sp. 17.1-17.2, USNM specimen 558279; USNM locality 42264.

18.4 may be a liverwort or an alga. It is one of several such specimens from the same location as the cf. *Glossopteris/Lesleya* occurrences, in northern New Mexico.

DISCUSSION

The Abo red beds and temporally equivalent, lithologically similar units in central New Mexico are remarkable for the low diversity flora they yield over such a large area. Based on paleosols and sedimentological architecture (Mack, 2003, 2007; Mack et al., 2010), the Abo Formation red-bed landscape formed under some degree of rhythmicity, likely reflective of Early Permian, glacial-interglacial cyclicality, which would have affected regional base-level and climate. The paleogeographic position of the Abo beds, in western equatorial Pangea, may have contributed strongly to this signature, as a result of paleoatmospheric circulation patterns (Tabor and Poulsen, 2008) that created a strongly monsoonal climate. Thus, the region likely was marked by moisture seasonality, varying in its intensity through the duration of any given cycle (represented by the alternation of paleosols and sandstone-siltstone beds, and, where present, limestones – Mack, 2007; Krainer and Lucas, 1995; Lucas et al., 2013).

The monotonous nature of the Abo Formation red-bed flora is belied by several deposits that are strikingly different from the norm; these are often dominated by, or enriched in, plants that are not found widely or, in several cases, at any other sampling sites. The most unusual of these is a site in the Zuni Mountains of northern New Mexico (USGS Locality 8931 and USNM locality 43680 - This site was collected by

geologists from the University of New Mexico, prior to 1930, and again in the 1930s by paleobotanists from the U.S. Geological Survey; unfortunately, we have been unable to relocate the outcrops based on the limited records remaining). The collection is dominated by a plant that is very similar to *Lesleya*, and somewhat less to *Glossopteris* (Fig. 13) in its obovate leaf shape. Leaves still in helical attachment to branches suggest either that the plant grew in stream side environments where it was buried rapidly in place during floods (in which case the plants themselves would have been quite small), or that branches were broken from plants living close to the site of burial and rapidly entombed, possibly during storms. In the southern part of the state, a single deposit was found (USNM Locality 42255) that contained a quite peculiar form of narrow, taeniopterid leaf (Fig. 12.3, 12.4), also possibly in attachment to stems. The specific identity of this leaf is not known to us. Also in the south, in the Robledo Mountains, a single deposit was found (USNM Locality 43662) dominated by the gigantopterid *Gigantopteridium americanum* (Fig. 10.2); gigantopterids have been mentioned in various studies of New Mexico geology (Bachman and Hayes, 1958; Hunt, 1983), but we identified them from only two collecting sites. These kinds of occurrences are in keeping with studies of modern landscapes that indicate that most species, even those that are rare in number of occurrences or average abundance, are somewhere abundant to even locally dominant (e.g., Murray and Lepschi, 2004).

The question arises about whether the monotony of conifer and supaioid dominance in the Abo Formation flora is a taphonomic happenstance. Both kinds of plants produced very leathery leaves and/or

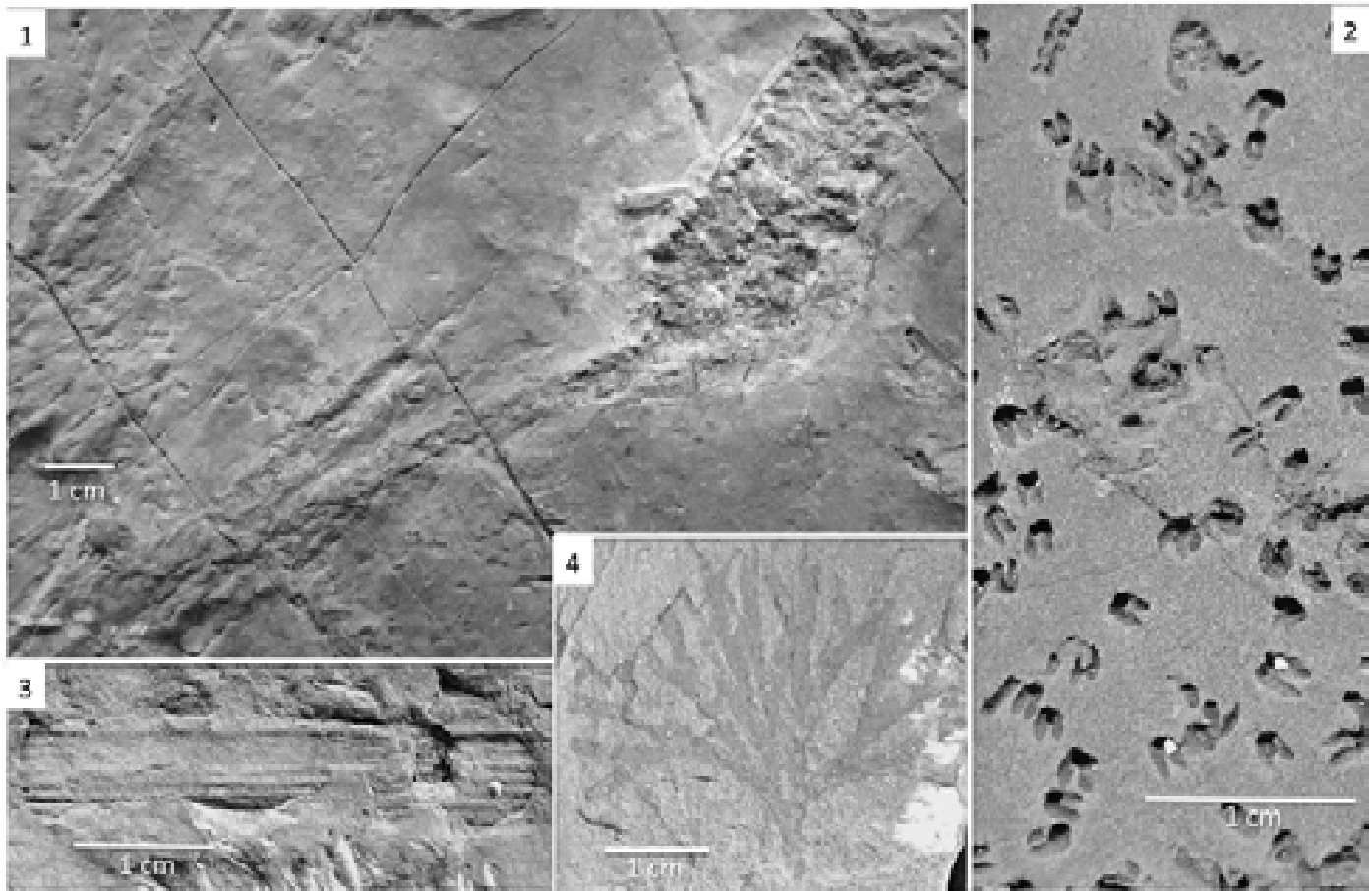


FIGURE 18. Conifer cone? **18.1**, USNM specimen 558296; USNM locality 42251. Sporangia of indeterminate affinity. **18.2**, USNM specimen 558297; USNM locality 42251. Calamite stem? **18.3**, USNM specimen 558298; USNM locality 42107. Thalloid thallus. **18.4**, USNM specimen 536578; USGS locality 8931.

branches, which presumably were able to withstand some degree of water transport prior to burial. This is indicated by not-infrequent specimens disposed across bedding planes, especially in fine sandstones with trough cross bedding, suggesting that the plant litter was rapidly deposited and buried. However, such occurrences are equaled or exceeded by mats of leaves of both groups, often in clay partings in the upper parts of fining upward fills of tabular sheet-sand channel belts. Such deposits indicate limited transport, deposition of large amounts of plant material under relatively quiet water conditions, probably settling from suspension. Furthermore, the rare elements in these floras, such as pectopteroids, neuropterids, or sphenopsids, occur in many cases with the more abundant plants, conifers, in particular, indicating that there does not seem to have been a persistent taphonomic filter removing them and leaving only conifers and supaioids. In addition, the rarity of plants that would be expected to occupy wet floodplains and streambanks, especially the robustly constructed calamitaleans, common as stem casts in many depositional environments (DiMichele and Falcon-Lang, 2012), suggests that they were absent from much of the landscape and confined to quite geographically and spatially limited wetter areas. In fact, the discovery at an Abo outcrop in Sierra County New Mexico, of autochthonous *Supaia thinnfeldioides* plants in periodically flooded streamside settings, indicates that these plants could tolerate strongly seasonally dry environments while surviving periods of flooding and partial burial (DiMichele et al., 2012).

As it relates to the possibility that the red-beds flora is simply a taphonomically filtered version of the same species pool found in the more diverse gray beds flora, consider the following. Cordaitaleans are very rare in the red-beds deposits. These plants had very tough, resistant

foliage capable of surviving transport. They also are known as elements of both coastal peat forming swamps (Raymond et al., 2010) and extrabasinal areas, under seasonal climates (Falcon-Lang and Bashforth, 2004). Thus, they had very wide environmental tolerances at the clade/evolutionary lineage level. Were the low diversity and composition of the red beds floras simply a taphonomic happenstance of transport, one might expect cordaitaleans to be a prominent part of the flora, especially since they occur at many gray Abo plant sites. Additionally, the leaves of supaioids and the branches of conifers are not similar in shape and size, and probably would not have had similar hydraulic characteristics under transport. Furthermore, these dominant elements tend to be separate in most deposits, suggesting that although they lived in proximity on floodplains, they probably were habitat differentiated to a fairly strong degree (DiMichele et al., 2007). And, there are elements of the red-beds floras that have not been found in the gray beds, including the peculiar “*Supaia*” *anomala* (Figs. 7 and 8), a plant with ribbon-like leaves, the *Glossopteris/Lesleya*-like plant from northern New Mexico (Fig. 13), and the *Taeniopteris*-like plant from southern New Mexico (Fig. 12). These patterns suggest that there were similarities between the gray and red beds species pools, but these were different in ways that would not be expected were taphonomic filtering the sole, or predominant explanation.

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Near Gallina Well, northeast of Socorro, a typical section of the lower part of the Lower Permian Abo Formation consists of red-bed mudstones and crossbedded sandstones. These are strata of fluvial origin. Note the lenticularity of the large sandstone channel complex in the foreground.

AN UPDATE ON THE FLORA OF THE KINNEY QUARRY OF CENTRAL NEW MEXICO (UPPER PENNSYLVANIAN), ITS PRESERVATIONAL AND ENVIRONMENTAL SIGNIFICANCE

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Abstract—The flora of the Kinney Brick Pit, Missourian age, central New Mexico, is revised. Total species diversity, estimated conservatively, is about 30. The flora is dominated by conifers and pteridosperms, and shows a preservational bias in favor of robust remains. Its varied composition reflects the various different habitats in the vicinity of the lagoonal (?) depositional environment. The various taxa identified are almost entirely the same as those known from Lower Rotliegend (Autunian) floras in Europe (*Annularia spicata* megafloral Zone).

INTRODUCTION

The Kinney Quarry site in central New Mexico is one of the few fully described Late Pennsylvanian Missourian-age floras in North America. The macroflora has been described in detail by Mamay and Mapes (1992), and particularly unusual taxonomic elements were singled out for description by Mamay (1981, 1990, 1992). Additional comments on the macroflora were included in the synthesis of Lucas et al. (2011), who determined the age of the deposit to be middle Missourian rather than Virgilian, as had long been asserted. Willard (1992) has described the palynoflora.

The flora appears to have been drawn from a coastal setting and deposited in a quiet, shallow embayment. These are certainly drifted remains preserving a mixture of plants from different habitats. It is emphasized that the collections available were made without record of the degree to which they are quantitatively similar to unbiased field collections, or of the degree to which unidentifiable fragments (comminuted plant debris) were excluded from the collections. Although plant fossils are known to come from at least two levels in the quarry, from an overall plant-bearing interval of less than 20 m thickness, collections were not systematically separated by collecting site or level. Consequently, it has not been possible to determine, from the collections themselves, if there were changes in the plant assemblages through time. Two color-variant lithologies are most common, and their floras do differ to some degree, suggesting variations of possible taphonomic or ecologic origin.

In this review, we re-illustrate the macroflora and provide some notes on the original identifications, updating some of these and adding a few additional taxa. The review is based on the collections made by Charles B. Read, Sergius H. Mamay, and Arthur D. Watt, of the U.S. Geological Survey, presently part of the collections of the National Museum of Natural History (NMNH). We stress that this is not a comprehensive systematic revision of the flora, but is intended as critical notes and an illustrated guide.

FLORA

The flora of the Kinney site is described below, and the various elements of this flora are listed in Table 1.

Lycopsids

Sigillaria brardii Brongniart (Fig. 1.1)

Lycopsid remains are exceptionally rare in the collections. Nearly all can be attributed to *Sigillaria brardii* (Figure 1.1) and range from small scraps of leaf cushion to larger pieces of bark with faint leaf-bases visible. This arborescent lycopsid is common throughout the Late Pennsylvanian

(Stephanian) wherever originally swampy, particularly organic-accumulating environments are represented. The large amount of variation in this species (and associated nomenclature) is reflected in the exhaustive monograph by Weiss (1893) (see also the discussion and enormous list of synonymy in Crookall, 1966). One of the North American synonyms is *Sigillaria approximata* Fontaine and White (1880).

A fragment with small leaf scars (Fig. 1.2) cannot be attributed confidently to any particular kind of lycopsid, but may represent part of a small ground cover or shrubby plant. Fragments of what may have been small lycopsids appear in many floras; no doubt, given the diversity of this group today, and its apparent long evolutionary history, small lycopsids were part of many floras.

Sphenopsids

The calamitalean sphenopsids are common elements of most late Paleozoic floras, generally in a context of wet environments. Only *Calamites gigas* Brongniart appears to have been adapted to drier conditions (Barthel and Kerp, 1992). Three different kinds of calamitalean remains are encountered in the Kinney floral assemblage, including foliage, fertile remains (strobili) and stems/rhizomes (Fig. 2.1), the latter generally as casts. These are not uncommon in the Kinney locality, but are taxonomically insignificant. Strobili are often quite spectacular, but difficult to interpret as adpressions with respect to sporangial attachment. Adpressed leaf whorls are generally easier to interpret, particularly where flattened whorls allow the shape and proportions of the individual leaves to be observed.

Reproductive organs

Calamitalean strobili are rarely preserved in their entirety. Three different kinds are recorded from Kinney, identified by Mamay and Mapes (1992) as *Calamostachys?* cf. *ludwigii*, *Calamostachys?* *pedunculata* and *Asterophyllites* sp. However, considerable doubt was expressed with regard to these identifications, which were considered provisional.

Palaestachya? sp. (Fig. 2.2)

A fairly well preserved succession of small strobili attached to a slender axis was figured by Mamay and Mapes (1992, fig. 2I) as *Calamostachys?* *ludwigii* Weiss. There is, indeed, certain resemblance with this Westphalian taxon, but the bracts in the strobili in Kinney are more curved and not reflexed as in *C. ludwigii*. The pointed bracts largely overlap, extending upward to about two thirds of the next leaf whorl. Sporangia are poorly preserved and relatively small, thus producing a lax aspect. The position of the sporangia appears axillary.

Lycopsids

- Sigillaria brardii* (stem)
- Unidentified small lycopsid (stem)

Sphenopsids

- Calamites* spp. (stem)
- Palaeostachya thuringiaca* (reproductive organ)
- Palaeostachya* sp. (reproductive organ)
- Calamostachys spicata* (reproductive organ)
- Annularia spicata* (foliage)
- Annularia spinulosa* (foliage)
- Annularia carinata* (foliage)
- Annularia sphenophylloides* (foliage)

Ferns

- Nemejcopteris feminaeformis* (foliage)
- Danaeites emersonii* (foliage)
- Pecopteris potonieii* (foliage)
- Pecopteris* cf. *oreopteridia* (foliage)
- Pecopteris* cf. *monyii* (foliage)
- Pecopteris* spp. (foliage)
- Remia pinnatifida* (foliage)
- cf. *Sphenopteris* (*Discopteris*?) *hadrophylla* (foliage)

Pteridosperms

- Macroneuropteris scheuchzeri* (foliage)
- Neuropteris ovata* (foliage)
- Neuropteris cordata* (foliage)
- Alethopteris schneideri* (foliage)
- cf. *Blanziopteris praedenta* (foliage)
- Mixoneura subcrenulata* (foliage)
- Mixoneura* cf. *gimmii* (foliage)
- Neurodontopteris auriculata* (foliage)
- Neurocallipteris planchardii* (foliage)
- Pseudomariopteris cordato-ovata* (foliage)
- Sphenopteridium manzanitanum* (foliage)

Coniferophytes

- Walchia piniformis* (stems, foliage)
- ?*Hermitia schneideri* (stems, foliage)
- Culmitzschia* cf. *C. speciosa* or *C. laxifolia* (stems, foliage)
- Ernestiodendron filiciforme* (stems, foliage)
- Dicranophyllum readii* (foliage)
- Gomphostrobus* sp. (reproductive organ)
- Cordaites* spp. (foliage)
- Cordaianthus* sp. (reproductive organ)
- cf. *Podozamites* sp. (stem, foliage)

Noeggerathiales

- Plagiozamites rochei* (foliage)
- Charliea manzanitana* (foliage)

Cycadopsids?

- Taeniopteris* sp. (foliage)

***Palaeostachya thuringiaca* (Weiss) Barthel (Fig. 2.3)**

A single, pedunculate strobilus (figured as *Calamostachys?* *pedunculata* Weiss by Mamay and Mapes, 1992, fig. 2J) shows relatively short, elongate sporangia. Bracts of successive whorls apparently do not overlap and barely clasp the sporangia. The compact aspect of this strobilus is produced by the proximity of successive whorls of sporangia. The insertion of sporangia can only be guessed at, but appears to be axillary.

This very complete specimen compares well with *Palaeostachya thuringiaca* as figured most recently by Barthel (2004) from the Lower Rotliegend (Autunian) of Manebach, Thuringia (Germany), in particular his Abb. 36. No attribution to any particular foliage has been made beyond conjecture.

***Calamostachys spicata* (Gutbier) Remy and Remy (Figs. 3.1, 3.2)**

A single specimen figured as ?*Asterophyllites* sp. by Mamay and Mapes (1992, fig. 2H) shows a *Calamites* stem with longitudinal ribbing, to which are attached several slender “branches” possessing well spaced whorls of tiny upswept leaves or bracts (although sporangia are absent). These thin “branches” are placed in such a manner as to suggest insertion on the stem nodes, and some of them are certainly attached.

The most obvious comparison is with *Calamostachys spicata* as figured and described by Remy and Remy (1975) as a special variety, which has not been accepted as such by later authors. The most complete specimen figured by Remy and Remy (1975, taf. 12, fig. 1) was partially refigured by Barthel (2012, abb. 2) with a very instructive enlargement (abb. 2a) showing curved bracts similar to the leaves of *Annularia spicata* (Gutbier) Schimper.

It may well be that *Asterophyllites dumasii* Zeiller, with *Calamostachys strobili*, as figured and described by Zeiller (1892) from Autunian strata at Brive, Massif Central, France, should be regarded as identical to *Annularia spicata* (even though Zeiller did not seem to consider this possibility). Kerp (1984) pointed out that the remains figured as *A. dumasii* probably represented strobili from which the sporangia had been removed. In this case, the upswept, sideways flattened foliar elements identified as *Asterophyllites* would be bracts rather than leaves. The bracts would be less “fleshy” than the leaves (see below under *A. spicata*), although the broadly curved shape is identical. Although Kerp’s interpretation of the “*Asterophyllites*” as strobili without preserved sporangia is accepted by the present writers, there is some doubt about his identification of “*dumasii*” for the material he described from Sobernheim, Saar-Nahe Basin, Germany. This material from Sobernheim shows stiffer, straighter bracts, and there is also a suggestion that the positioning of its sporangia might fit *Palaeostachya* rather than *Calamostachys*. However, sporangial insertion is not always straightforward, as interpreted from adpressions. It is noted that Gastaldo (1981) referred to Good (1975) for regarding *Calamostachys* and *Palaeostachya* as artificial genera; Gastaldo cited *Calamostachys spicata* as an intermediate case.

***Annularia spicata* (Gutbier) Schimper (Figs. 3.3, 3.4)**

The small, rather “fleshy,” pointed and curved leaves of this species are illustrated quite well by Mamay and Mapes (1992, fig. 2E) under the name of *Annularia* cf. *asteris* Bell. The latter is another species characterized by small, pointed leaves. However, these are narrower and not quite as “fleshy” as those of *Annularia spicata*, which are also more distinctly curved.

Annularia spicata has been profusely illustrated from various Lower Rotliegend (Autunian) localities in eastern Germany by Barthel (2012), who also discussed its synonymy and the various records of this species in the general type area (Gutbier’s types are from the Erzgebirge in Saxony). This species is widespread in Europe and characteristic for the highest Stephanian, i.e. the Autunian Substage (see Wagner and Álvarez-Vázquez, 2010a – an earlier Stephanian record of *A. spicata*, op. cit., table 8, is due to an error of transcription). In the Appalachian region of North America, it is recorded from the Conemaugh (post-Mahoning), Monongahela and Dunkard by Blake et al. (2002, fig. 2 – range chart).

The specimen with slightly larger leaves, figured as *Annularia* cf. *radiata* Brongniart by Mamay and Mapes (1992, fig. 2D, refigured here as Fig. 4.1) also may be *Annularia spicata*, although it is larger than usual for this species and the leaf shape is atypical. The specimen bears some resemblance to specimens described and illustrated by Mamay and Read (1956) as *Phyllothea paulinensis* from the Spotted Ridge flora of central Oregon, which is of indeterminate late Paleozoic age (although the present authors disagree on the nearness of this comparison). Attribution of the Spotted Ridge specimens to *Phyllothea* is questionable, given that the name originated for material from Gondwana, with which the Oregon specimens are not wholly congruent. As in *P. paulinensis*, whatever its true affinities may be, the leaves are consistently wider at their bases, taper to a narrow, elongate, acuminate apex, have a reflexed posture, and



FIGURE 1. Lycopside. **1.1**, *Sigillaria brardii*, USNM specimen 450764 (Mamay and Mapes, 1992, Fig. 2A). **1.2**, Unidentified lycopside axis, possibly a fertile section, USNM specimen 558126. Scale bar in mm, applies to both specimens.

may be borne at nearly right-angles to the stem. Some give the impression of being fused in their basalmost parts, though this is uncertain. In addition, Mamay and Read (1956) described whorls of up to 20 leaves. No whorls with that high a leaf count are present in this single specimen from Kinney.

Annularia spinulosa Sternberg (Fig. 4.2)

Barthel (2004) has argued that *Annularia spinulosa* should be preferred to the more commonly used name *Annularia stellata* (Schlotheim) Brongniart, not only because the International Code has declared Schlotheim's names illegitimate, but also because Schlotheim's type specimen could not be located (and is possibly lost). It also appears that *A. spinulosa* has pointed leaves, whereas specimens identified as *A. stellata* in the literature often show blunt leaf endings. The remains with several leaf whorls as found in the Kinney locality conform to the characters of *A. spinulosa* as figured from its type area in Saxony (Lower Rotliegend of Döhlen Basin).

This specimen was identified as *Annularia* cf. *pseudostellata* Potonié by Mamay and Mapes (1992, Fig. 2F). The latter, however, has more parallel-sided leaves in laxer whorls than occur in *Annularia spinulosa*.

Annularia carinata Gutbier (Fig. 4.3)

Two anisophyllous leaf whorls are shown with closely spaced mucronate leaves showing their maximum width in the upper third of the leaf length.

This is a highly characteristic species figured and described from the Lower Rotliegend (Autunian) of Saxony (e.g. Barthel, 2004, abb. 19, 20). It has been synonymized with *Annularia mucronata* Schenk (op. cit.). This species is regarded as widespread but uncommon with a stratigraphic range from Siberian to Autunian.

Annularia sphenophylloides (Zenker) Gutbier (Fig. 4.4)

A single, anisophyllous leaf whorl has been illustrated under this name by Mamay and Mapes (1992, fig. 2G). This long-ranging species (from late Bolsovian upwards throughout the entire Stephanian) is both common and easily recognized by its rather short, spatulate, mucronate leaves, which are normally very closely spaced. The earliest occurrence of *Annularia sphenophylloides* in the Appalachian region is recorded as upper Kanawha by Blake et al. (2002). This would correspond to upper Bolsovian in Europe.

Ferns

Identifiable fern remains are rare in the plant assemblage. However, due to various factors, the degree of difference between the abundance of ferns in the original vegetation and that in the fossil assemblage is difficult to quantify. Fern foliage in general is usually thin-limbed and species characterized by more sturdy laminae are relatively uncommon. Combined with prolonged maceration in water, non-robust construction will lead to decay of tissues and disintegration of fern foliage, particularly if there is no thick cuticle to hold pinnules together. Ground cover and low stature ferns, in particular, tend to be undersampled in most fossil floras, particularly when there was a well developed canopy that shielded low-growing plants from the effects of wind (Scheibling, 1980).

Pennsylvanian ferns are preferably classified on sporangial structures. This is particularly relevant for the incised pinnules with fan-shaped venation, which are regarded as sphenopteroid (with reference to *Sphenopteris* Brongniart, a morphogenus). Certain genera with sphenopteroid foliage (e.g. *Discopteris*) have more sturdy laminae and are thus more capable of resisting disintegration. The other large group of fern-like foliage, the pectopteroids (after *Pectopteris* Brongniart, another morphogenus), may also be classified on sporangiate structures. However, the most common synangia, *Scolecopteris* (*Asterotheca*), occur in

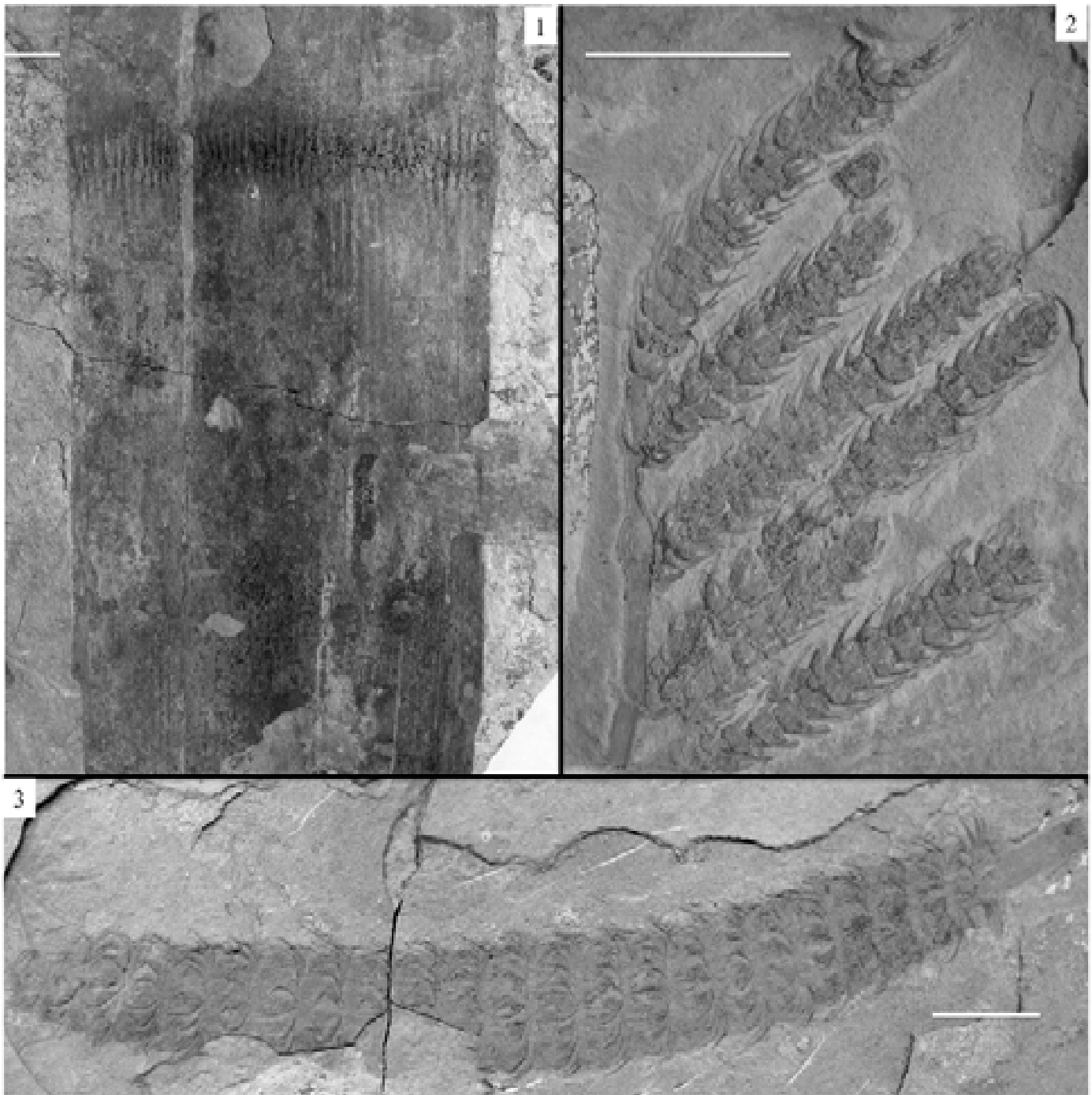


FIGURE 2. Calamitalean stems and cones. **2.1**, *Calamites* stem, USNM specimen 450766 (Mamay and Mapes, 1992, Fig. 2C). **2.2**, *Palaeostachya thuringiaca*, USNM specimen 450772 (Mamay and Mapes, 1992, Fig. 2I). **2.3**, *Calamostachys tuberculata*, USNM specimen 450773 (Mamay and Mapes, 1992, Fig. 2J). Scale bars = 1 cm.

many different forms of pectopteroids, which may be subdivided in more detail on pinnule and venation development.

Pecopteroid fern remains are uncommon in the Kinney plant assemblage. On the whole, these are small fragments with poorly preserved venation, making identification difficult. A notable exception is a quite characteristic species (here identified as *Pecopteris potonieii* Nemejc) that Mamay and Mapes (1992) recorded as *?Pecopteris* sp. (their Fig. 3B, G, non A). These specimens are fertile (*Asterotheca*) and it seems that the fairly large specimens may owe their cohesion to the sporangial cover, which lent strength to the pinnules.

It must be emphasized that preservational conditions may have

thinned out the total representation of floral elements to the detriment of the ferns.

***Nemejcopteris feminaeformis* (Schlotheim) Barthel (Fig. 5)**

Zygoterids, a large group of ferns with habits ranging from small trees to groundcover (Barthel, 1968; Phillips and Galtier, 2005), are represented by small fragments of foliage belonging to *Nemejcopteris feminaeformis*. This foliage is very recognizable, with elongate pinnules, denticulate pinnule margins, and highly angled straight simple veins that terminate in the marginal teeth; thus, even small debris can be identified. *Nemejcopteris feminaeformis* occurs throughout the Stephanian, from Barruelian upwards.

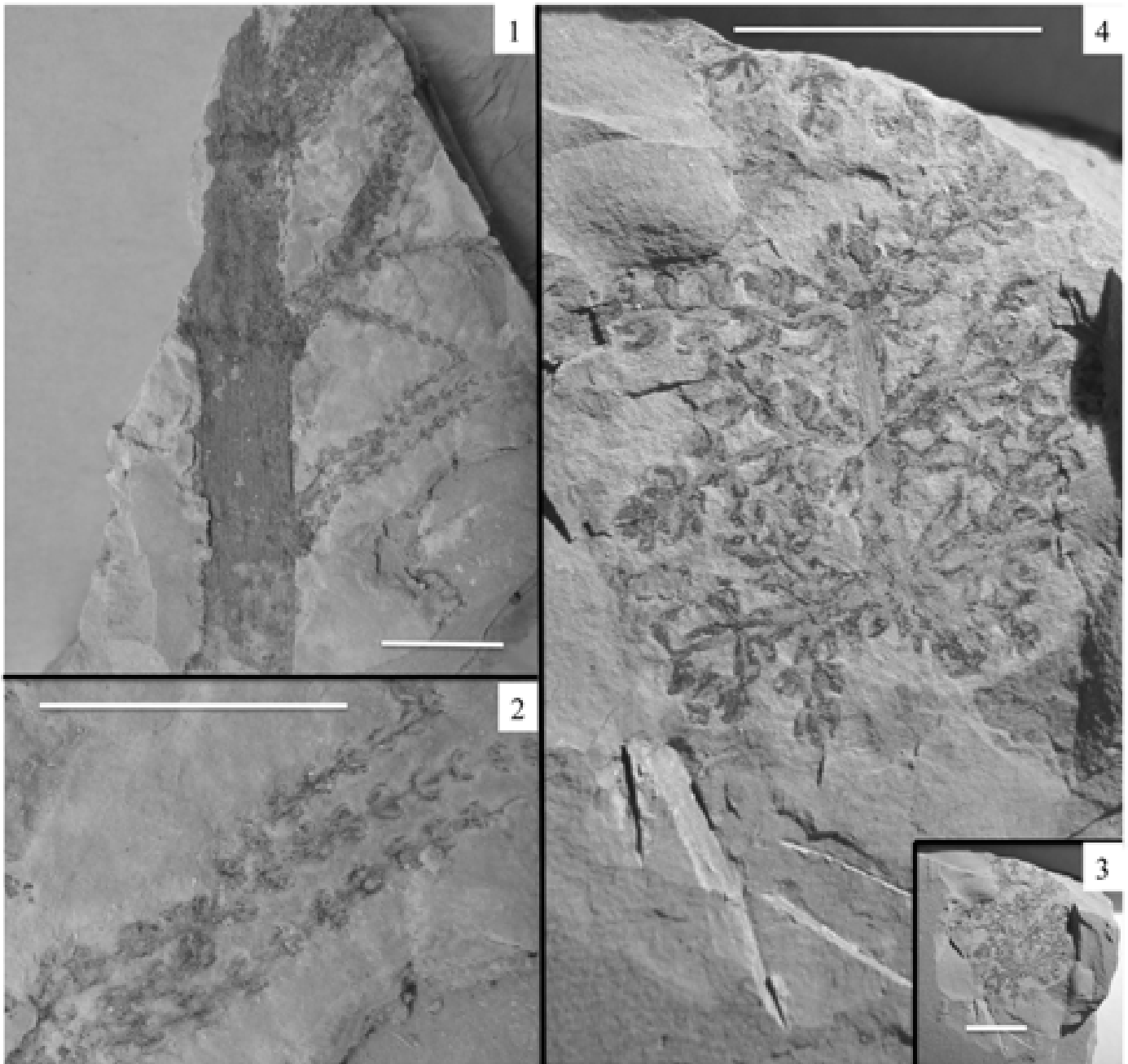


FIGURE 3. Calamitalean cones and foliage. **3.1**, *Calamites* stem bearing cones of *Calamostachys dumasii* (Zeiller) Jongmans, USNM specimen 450771, the laterally borne appendages on the lower part of the main stem are cone axes bearing only bracts; the sporangiophores have been shed, (Mamay and Mapes, 1992, Fig. 2H). **3.2**, *Calamostachys dumasii* cones bearing only bracts, USNM specimen 450771, higher magnification view of specimen in **3.1**. **3.3-3.4**, *Annularia spicata*, USNM specimen 450768 (Mamay and Mapes, 1992, Fig. 2E). Scale bars = 1 cm.

***Danaeites emersonii* Lesquereux (Fig. 6.1).**

Clearly a marattialean fern, fertile examples of *Danaeites* can be identified by the elongate synangia (fused groups of sporangia) on the lower surfaces of pinnules. *Danaeites emersonii* is most easily recognized on fertile remains, but can also be distinguished on the relatively broad, closely spaced pinnules with a rounded apex and at least twice bifurcate vein groups. *Danaeites* is characteristic of mid-Stephanian (Saberian) and younger rocks (Wagner and Álvarez-Vázquez, 2010a).

***Pecopteris (Asterotheca) potoniei* Nemejc (Figs. 6.2, 7)**

Among the remains of pecopterid foliage are fairly large specimens, such as that figured by Mamay and Mapes (1992, fig. 3B, re-

illustrated here, Fig. 7), which shows the subtriangular shape of widely spaced pinnules and gradually tapering pinnae with a small, poorly individualized apical pinnule. The pinnule lamina is covered by sporangia in a stellate pattern (*Asterotheca*, *Scolecopteris*). The veins are almost invariably obscured. In the rare cases where they are not (Fig. 6.2), the widely spaced lateral veins regularly fork once, with a second bifurcation occurring in the basal basiscopic part of larger pinnules.

The nearest comparison is with *Pecopteris (Asterotheca) potoniei* Nemejc, from the Rotliegend of Thuringia, Germany. This form, misidentified as *Pecopteris pennaeformis* Brongniart by Potonié (1893, taf. IX, figs 1-2), was renamed by Nemejc (1940, p. 14-15). It has the subtriangular, well-spaced pinnules that are partially fused at the base with adjacent pinnules. Lateral veins are generally once forked and rarely

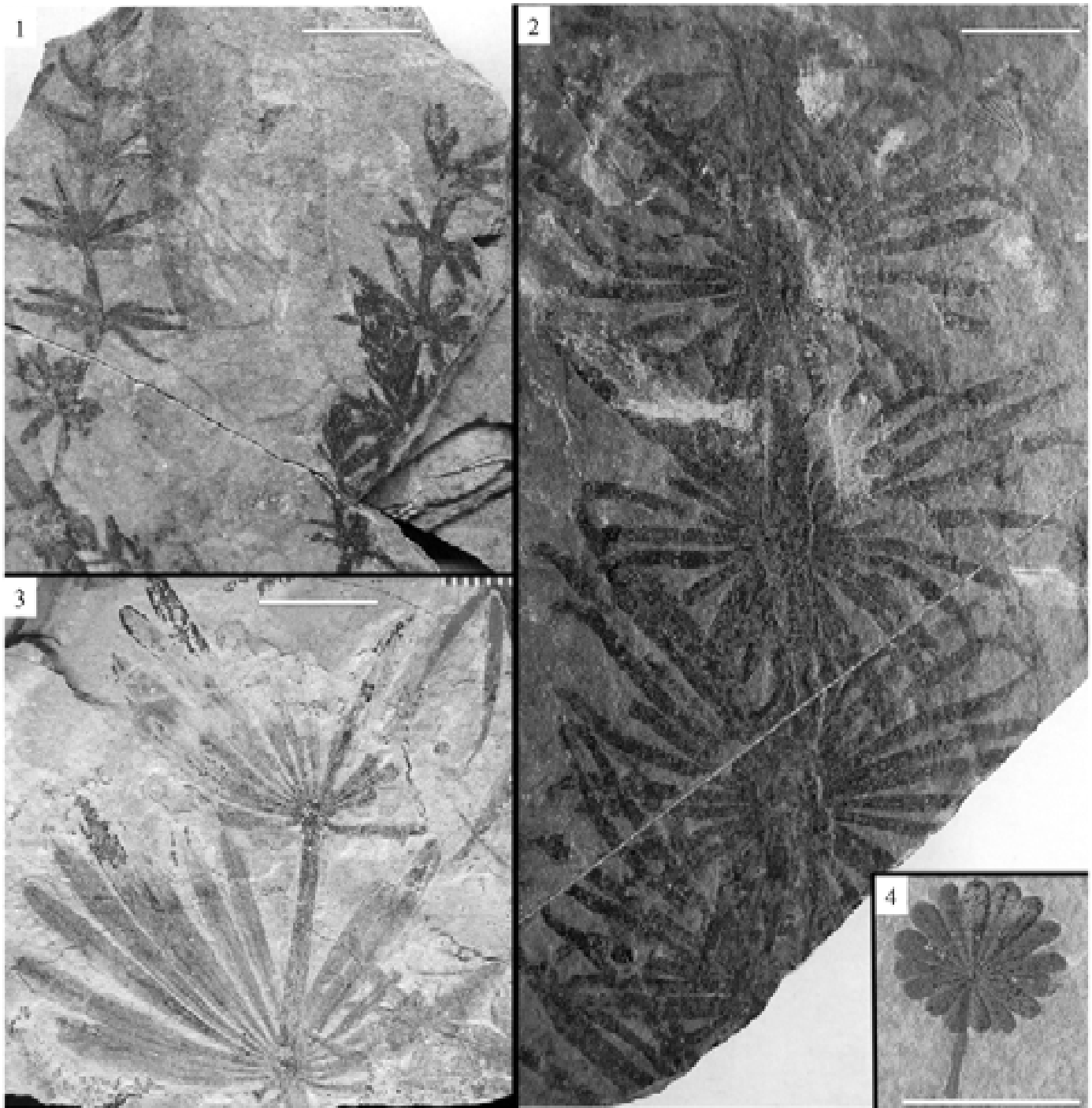


FIGURE 4. Calamitalean foliage, *Annularia*. **4.1**, *Annularia* cf. *spicata*, USNM specimen 450767 (Mamay and Mapes, 1992, Fig. 2D). **4.2**, *Annularia* *spinulosa*, USNM specimen 450769 (Mamay and Mapes, 1992, Fig. 2F). **4.3**, *Annularia* *carinata*, USNM specimen 558132. **4.4**, *Annularia* *sphenophylloides*, USNM specimen 450770 (Mamay and Mapes, 1992, Fig. 2G). Scale bars = 1 cm.

fork once again in the basal part of the longer pinnules. The size and shape of both pinnules and pinnae are the same for the Thuringian species and the material at hand. It is therefore proposed to accept a direct identification.

The relative frequency of these remains at Kinney (about 5% of the specimens in the collection) and their fairly large size probably can be explained as a result of increased cohesion due to the presence of sporangia, which would have made these (invariably fertile) remains more resistant to disintegration on (prolonged) maceration in water.

Pecopteris cf. *oreopteridia* (Schlotheim) Brongniart (Fig. 8.1)

A penultimate pinna fragment shows gradually fused, rapidly tapering pinna terminals with a poorly individualized, subtriangular apical pinnule. Poorly preserved lateral pinnules are up to twice as long as broad, with rounded apices, and tending toward a subtriangular shape. Veins are barely discernible, but a well developed midrib seems to be present, and there is a suggestion of rather widely spaced laterals.

The size and shape of last order pinnae characterized by rather short, gradually fused terminals allow comparison with *Pecopteris*

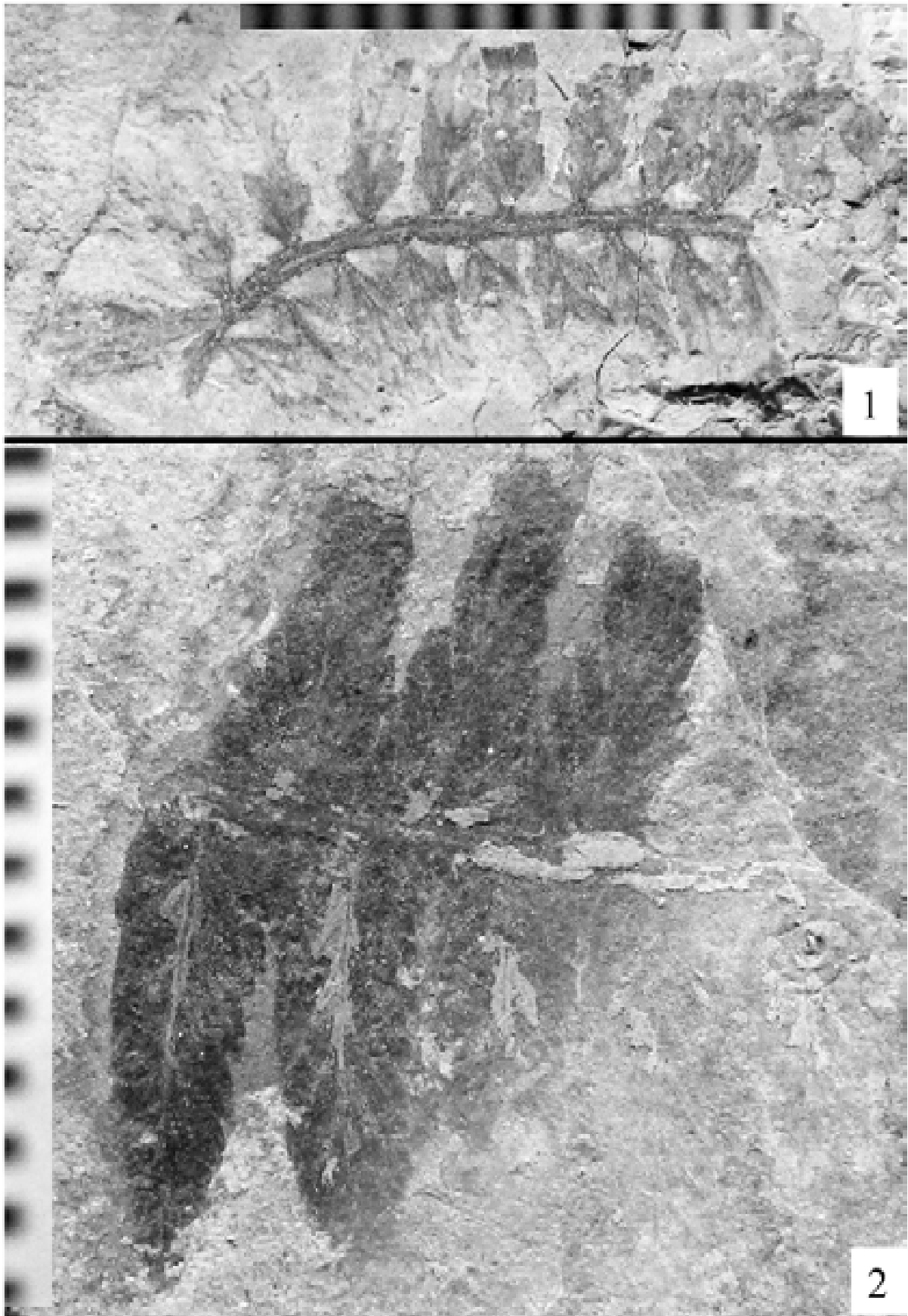


FIGURE 5. Fern foliage. *Nemejcopteris feminaeformis* foliage. 6.1, USNM specimen 558134 (Mamay and Mapes, 1992, Fig. 3F). 6.2, USNM specimen 450779. Scale bars in mm.

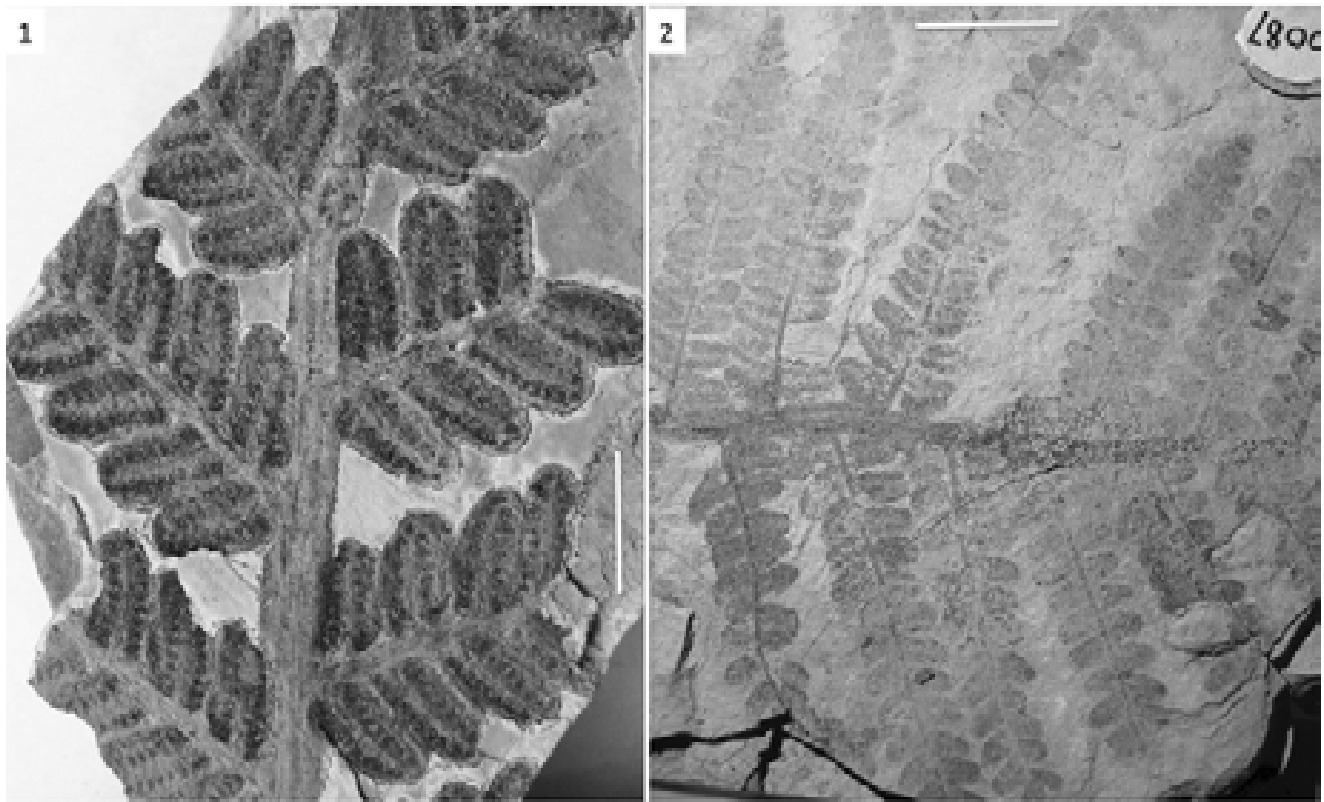


FIGURE 6. Fern foliage. **6.1**, *Danaeites emersonii*, fertile, USNM specimen 450776 (Mamay and Mapes, 1992, Fig. 3C). **6.2**, *Pecopteris* cf. *potonie*, USNM specimen 558263. Scale bars = 1 cm.



FIGURE 7. Fern foliage, *Pecopteris (Asterotheca) potonie* Nemejc, fertile. **7.1**, Terminal portion of a pinna, USNM specimen 450780 (Mamay and Mapes, 1992, Fig. 3B), scale bar in mm. **7.2**, Higher magnification view of specimen in **5.1**, scale bar = 1 cm.

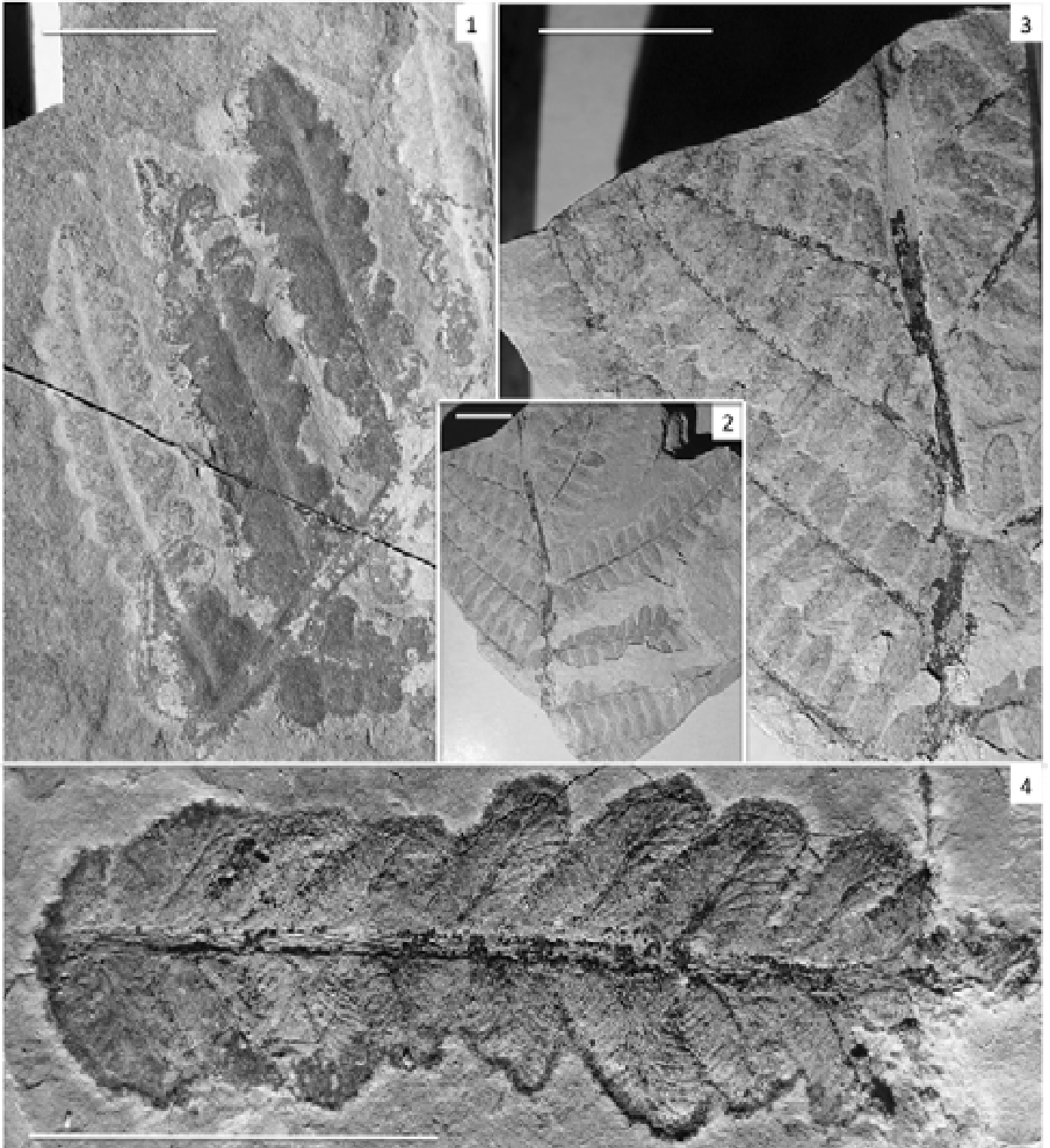


FIGURE 8. Fern foliage. **8.1**, *Pecopteris* cf. *oreopteridia*, USNM specimen 558266. **8.2**, *Pecopteris* cf. *monyi*, USNM specimen 558262. **8.3**, Enlargement of **8.2**, *Pecopteris* cf. *monyi*, USNM specimen 558262. **8.4**, *Pecopteris* sp., USNM specimen 558255. All scale bars = 1 cm.

(*Scolecoperis oreopteridia* as figured by Barthel (1980, taf. I) from its type area. The pinnule shape is also similar. A definite identification is precluded by the lack of detail with regard to venation.

***Pecopteris* cf. *monyi* Zeiller (Figs. 8.2, 8.3)**

A fairly large fragment of a penultimate pinna shows a broad rachis with up to six pinnae inserted on either side. No pinna terminals are preserved. Pinnules are subtriangular, apparently thin-limbed, with a thin, decurrent midrib. Lateral veins are indistinct, but possibly rather widely spaced. The subtriangular shape of pinnules with an apparently thin lamina, and showing a decurrent midrib, suggest *Pecopteris monyi*, as described from the Massif Central, France. This is a reasonably common species in the Stephanian of Europe. In the absence of details about the lateral veins, the specimen in hand can only be identified tentatively. It is noted that *Pecopteris monyi* needs a critical revision in view of the fact that several pecopterids with a sloping midrib have been figured under this name in the literature.

***Pecopteris* sp. indet. (Fig. 8.4)**

This is a small fragment, only 2.5 cm long, which shows parallel-sided pinnules with a thin, slightly decurrent midrib and once bifurcate veins. There is a certain resemblance to *Pecopteris oreopteridia* (Schlotheim) Brongniart, but this pinna fragment is too poorly preserved to insist on the determination. There is a suggestion of a small apical pinnule, such as occurs in *Pecopteris densifolia* Göppert, but this seems to be a fortuitous superposition of pinna fragments rather than an organic connection.

***Remia pinnatifida* (Gutbier) Knight emend. Kerp et al. (Fig. 9.1)**

A single pinna fragment of the penultimate order shows a sturdy rachis on which short, bluntly terminated last order pinnae are inserted at a high angle. Some pinnules are preserved as small, though well individualized, elements with a vaulted lamina showing deeply immersed veins consisting of a short midrib and oblique, simple and once bifurcate laterals. A portion of this specimen may be fertile, although this is open to interpretation, given the state of preservation.

Remia pinnatifida occurs from Siberian (Knight, 1985) to Autunian (e.g. Barthel, 2005; Kerp et al., 1991) in Europe (compare Wagner and Álvarez-Vázquez, 2010a). It does not seem to have been recorded previously from North America. The specimen recorded here has been transported and it not well preserved, but is sufficient to show the salient characters of this species.

cf. *Sphenopteris* (*Discopteris*?) *hadrophylla* Knight (Fig. 9.2)

A poorly preserved specimen consisting of two parallel pinnae showing fairly robust pinnules with subrounded, perhaps angular lobes is compared with this species, which was described from Siberian and Stephanian B strata in NW Spain (Knight and Wagner, 2012). Pinnule laminae are too faintly preserved (as a result of maceration) to show the lobate pinnule margins in sharp outline. Only remnants of the venation can be seen, thus allowing limited recognition. As far as can be seen, the veins appear to fork once and are strongly apically directed. Following bifurcation, the resultant veins curve in the same direction, thus forming a rather closed space between them. In the smaller pinnules, the veins are almost concave upward and a midvein is not well developed.

Sphenopteris hadrophylla is characterized by its sturdy pinnules with a vaulted lamina. This would have allowed fragments of this species to be preserved where other sphenopterids with a thinner lamina would have fallen apart.

Pteridosperms

The pteridosperms, also known as seed ferns, are a common element overall in the Kinney flora. They are represented by several taxa and present an array of variation that is challenging to sort out when dealing with fragmentary specimens.

***Macroneuropteris scheuchzeri* (Hoffmann) Cleal, Shute and Zodrow (Figs 10.1, 10.2)**

This species is represented by large, elongate, linguoid pinnules with characteristic short, apically directed surface hairs (see Laveine and Behlis, 2007). This plant is typical of wetland habitats (Stull et al., 2012), and in some Virgilian floras, such as that from north-central Texas, the taxon is found in dense, nearly monospecific accumulations in environments transitional between clastic to peat-forming swamps (DiMichele et al., 2005; Tabor et al., 2013). The presence of *M. scheuchzeri* in the Kinney flora, as with *Sigillaria*, suggests that standing water or periodically flooded swamps were present in places on the landscape. *Macroneuropteris scheuchzeri* has an appreciably longer range in North America than it has in Europe, where it occurs in strata up to and including the upper Cantabrian (Wagner and Álvarez-Vázquez, 2010a).

***Neuropteris ovata* Hoffmann (Figs 11, 12, 13)**

This taxon encompasses a considerable range of variation, some of which has been segregated at the species or subspecies level (see illustrations and discussion in Castro, 2005). In its most characteristic form (Figs. 11 and 12.2), pinnules of *Neuropteris ovata* are short and tongue-shaped, with a distinct basal “auricle” – a small extension of the lamina directed basiscopically. The lateral veins are fine, dense, and arch toward the margin from a thin, often weakly developed midvein. In the terminal portions of pinnae, lateral pinnules may become broadly attached to the rachis and fuse to various degrees (see particularly Fig. 12.2). There is considerable variation in this basic ground plan, expressed in the size and shape of the terminal pinnules, lateral pinnule size and shape, and degree to which pinnules near the termini of pinnae are fused to the rachis or to one another (see, for example, the large size of the *N. ovata*-like pinnules illustrated in Fig. 13, or the specimens illustrated in Fig. 20, the affinities of which could be with *N. ovata* or *Neurodopteris auriculata* – see below). Given the range of variation encompassed by specimens assigned to this species at different times and places, it might be concluded that *Neuropteris ovata* is in fact a species complex comprising several subtly different species.

The specimens collected from Kinney Quarry show the relatively broad pinnules with a dense venation that is characteristic of *Neuropteris ovata* var. *grandeuryi* of Wagner (1963). This refers to *Neuropteris flexuosa* Grand'Eury (non Sternberg) as figured by Bertrand (1930) from the Cévennes in the French Massif Central. It is a rare element in the classic Stephanian floras of the Massif Central (figured as *Neuropteris ovata* by Doubinger et al., 1995, from Saint Étienne, and misidentified as *Neuropteris heterophylla* Brongniart by Zeiller, 1890, pl. XII, fig. 1 from Épinac). *Neuropteris ovata* is a common species in the coastal Stephanian basins of NW Spain, up to and including the Siberian substage level (*sensu* Wagner and Álvarez-Vázquez, 2010a), and occurs less frequently higher up (e.g., Stephanian C of Puertollano, Wagner, 1985). It is shown as ranging throughout the Allegheny, Conemaugh, Monongahela and Dunkard of the Appalachian Basin in North America (Blake et al., 2002). This species appears to have been closely tied to swampy, frequently peat-forming habitats (Stull et al. 2012).

***Neuropteris cordata* Brongniart (Fig. 14)**

A poorly preserved fragment of the terminal part of a penultimate pinna shows elongate pinnules, 2.5 to 3 times longer than broad, with subparallel margins and a bluntly pointed apex. Pinnule attachment is neuropteroid. The terminal pinnule is relatively short and more or less individualized, with a rounded apex. In the lower part of this specimen, small, subcircular pinnules are separated off from the basal parts of pinnules in the process of being converted into last order pinnae.

The pinnule lamina seems fairly sturdy, but not vaulted, with a thin midrib marked clearly for about two thirds of the pinnule length. Lateral veins arise at a narrow angle and curve away from the midrib to reach the pinnule margin slightly obliquely.

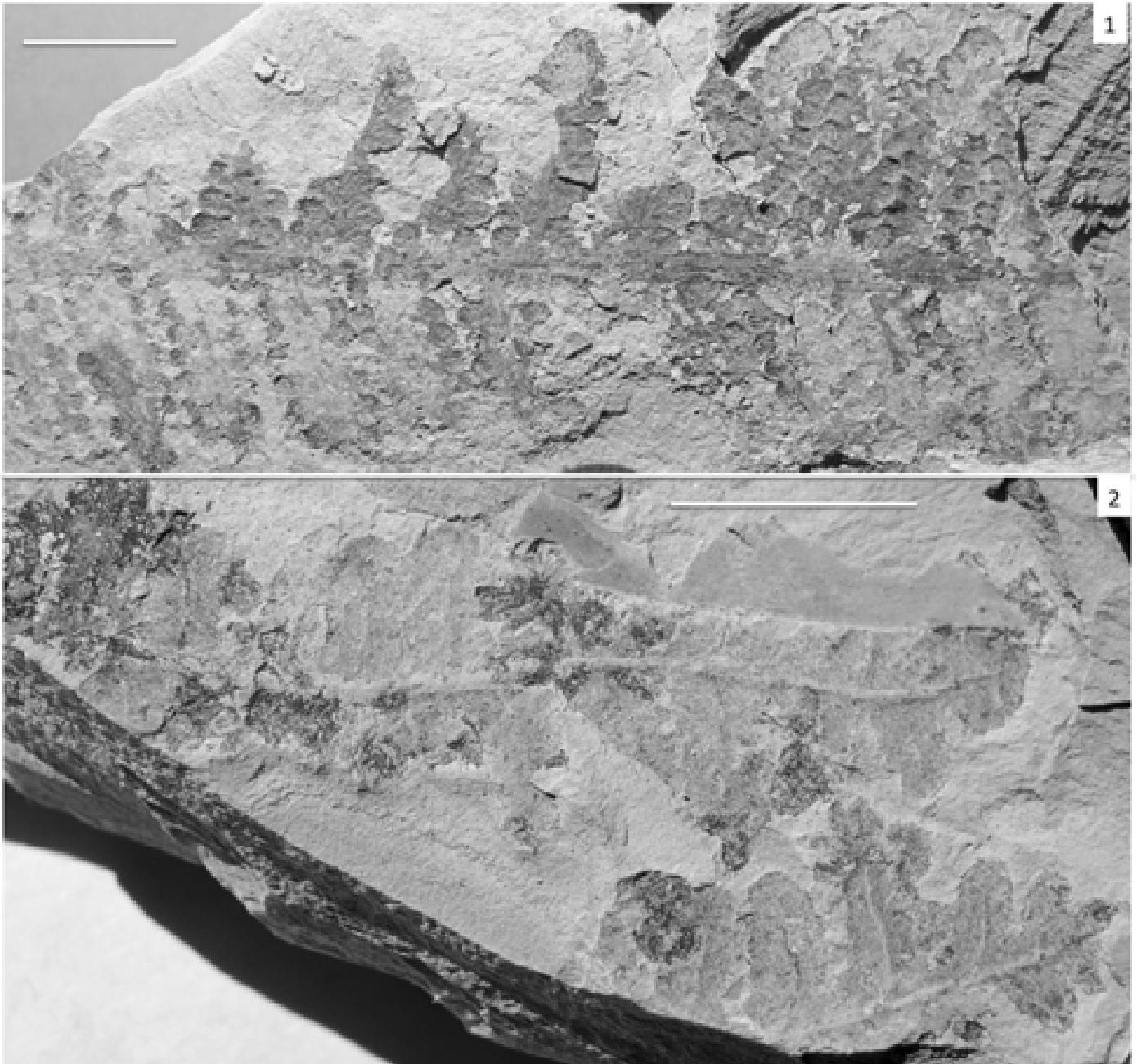


FIGURE 9. Fern foliage, *Remia* and *Sphenopteris*. **9.1**, *Remia pinnatifida*, USNM specimen 558256. **9.2**, cf. *Sphenopteris* (*Discopteris*?) *hadrophylla*, USNM specimen 558265. Scale bars = 1 cm.

The nearest comparison is with *Neuropteris cordata* as figured by Barthel (1976, tafn 33-34) from the Lower Rotliegend (Autunian) of Saxony. Barthel's photographs and drawings illustrate a range of morphological variation into which the specimen from Kinney fits without difficulty.

Cleal and Shute (1995) regarded *Neuropteris cordata* as a poorly known species (Brongniart's holotype is a single pinnule), which they considered as possibly the same as *Neuropteris pseudoblissii* Potonié. This is not a foregone conclusion however. At least two different species have been recorded in the literature as *Neuropteris cordata*, and Cleal and Shute (1995) are correct that this species is in need of revision. With regard to the specimen from Kinney Brickpit, it suffices that this specimen fits the well illustrated material recorded as *Neuropteris cordata* from Germany.

***Alethopteris schneideri* Sterzel (Fig. 15)**

A single specimen in the NMNH collections can be referred to the genus *Alethopteris*. It is only a small fragment (Fig. 15) but it seems fairly characteristic of this later Stephanian or early Rotliegend plant (compare Barthel, 1976), which is regarded as latest Stephanian by Wagner and Álvarez-Vázquez (2010a). As is typical of this species, the pinnules are inserted obliquely, are elongate with obtusely rounded ends, have a fairly deep acroscopic incision, and a relatively dense venation. The specimen also compares to some extent with *Alethopteris zeilleri*, although the latter's venation is coarser and the pinnule insertion is more nearly perpendicular.

cf. *Blanzopteris praedentata* (Gothan) Krings and Kerp (Fig. 16)

Several neuropteroid pinnules from the Kinney collections re-



FIGURE 10. Pteridosperm foliage. **10.1-10.2**, *Macroneuropteris scheuchzeri*, USNM specimen 450786 (Mamay and Mapes, 1992, Fig. 5B). **10.3**, *Neurodopteris auriculata*, USNM specimen 450789 (Mamay and Mapes, 1992, Fig. 5D). Scale bars = 1 cm.

semble *Blanziopteris praedentata* (Krings and Kerp, 1999). All are fragmentary, but show the pointed shape and crenulate margin of this rare species. This species was formerly treated as *Neuropteris* by Gothan (1909). Cleal and Shute (1995) treated it as *Sphenoneuropteris*; however, *Sphenoneuropteris* Shchegolev, with two species, is a synonym of *Ulvopteris crenulata* (Brongniart) (unpublished data, RHW). *Blanziopteris* has variable pinnule shapes, strongly dependent on position within the frond. Importantly, pinnules may have denticulate margins, particularly in the upper one-third. Veins are sparse, widely spaced, strongly marked and steeply arching. They terminate in marginal crenulae or teeth in those pinnules in which crenulate margins occur. Terminal pinnules are elongate and tongue shaped. Krings and Kerp (1999) noted that the species is only known conclusively from the Massif Central of France, and questioned other reports (from Spain, Portugal and Russia). The taxon was erroneously identified in the Kinney flora as *Neuropteris*

pseudoblissii (listed in Lucas et al., 2011, p. 18). *Neuropteris pseudoblissii* also is Stephanian in its stratigraphic distribution, the species being typified on material from Commentry in France (Crookall [1959] identified a single Westphalian/Duckmantian specimen from Britain, but that seems to have been in error). Krings and Kerp (1999) make a strong case for a climbing growth habit in *B. praedentata*.

***Mixoneura subcrenulata* (Rost) Zeiller (Figs. 17.1, 17.2, 18)**

A large number of specimens in the Kinney collection is attributed to the group of species within the genus *Mixoneura*, originally defined as a subgenus of *Odontopteris* Brongniart by Weiss (1870). It is quite different from *Odontopteris sensu stricto* (see discussion in Wagner and Castro, 1998, and Laveine and Dufour, in press). Although a larger number of species are currently attributed to this genus than Weiss originally

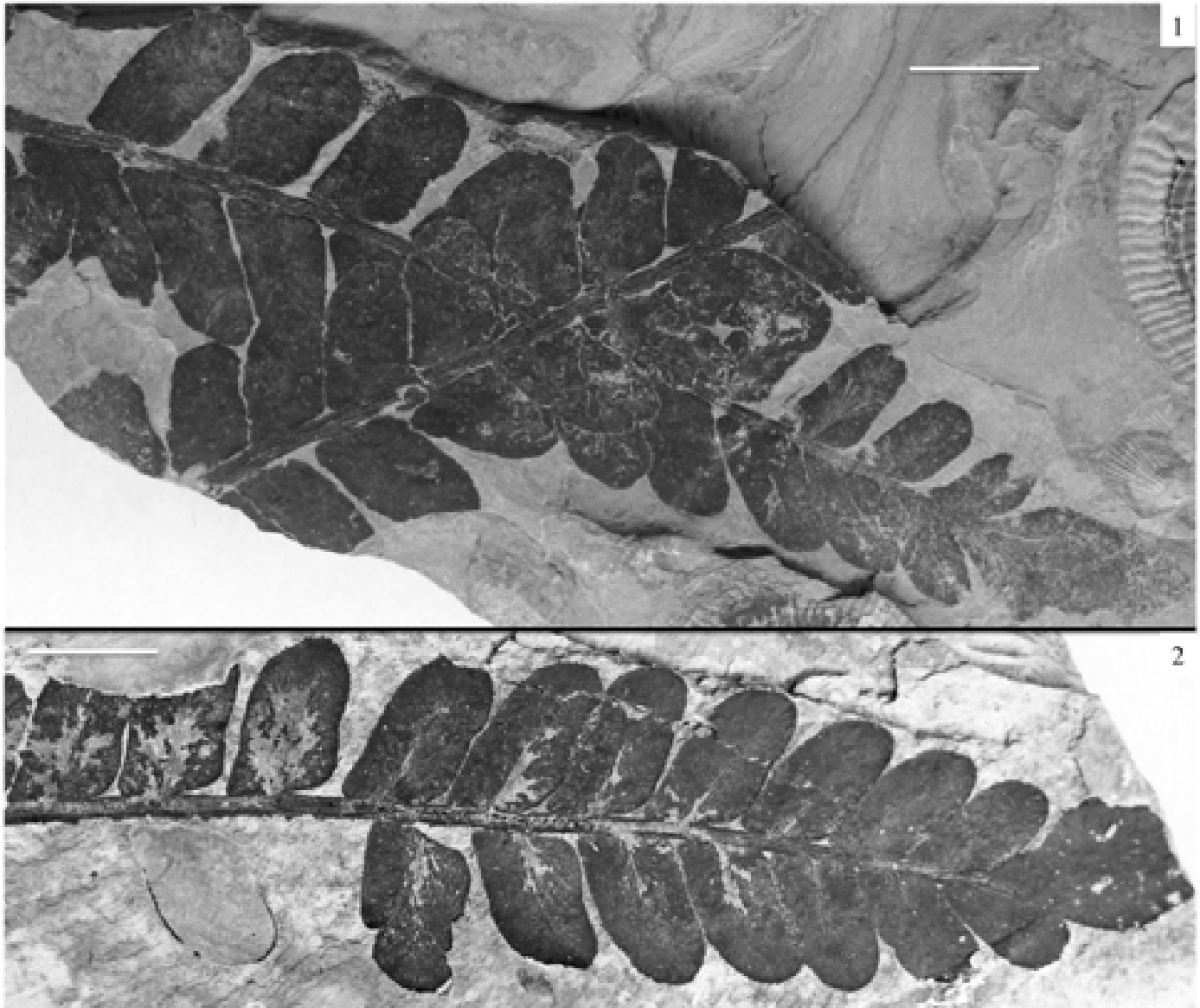


FIGURE 11. PteridospERM foliage, *Neuropteris ovata*. **11.1**, USNM specimen 450784 (Mamay and Mapes, 1992, Fig. 4E). **11.2**, USNM specimen 588131. Scale bars = 1 cm.

envisaged, the most common elements of *Mixoneura* belong to the *subcrenulata-lingulata* group and that around *Mixoneura schlotheimii* (Brongniart) (= *osmundaeformis* Schlotheim), with *Mixoneura pseudoschlotheimii* de Maistre as a possible synonym. Essentially, *M. subcrenulata* and *M. lingulata* are characterized by largely fused, rather massive terminals, whereas *M. schlotheimii* shows more gradually tapering, less fused terminals.

Doubinger and Remy (1958) compared *Mixoneura subcrenulata* with *Mixoneura lingulata* from upper Stephanian strata (including Lower Rotliegend) in Germany and France. They distinguished *M. lingulata* primarily on the more largely fused terminals and subdivided *M. subcrenulata* into three different varieties, which they suggested might be regional varieties. The subtleties of varietal differentiation may be difficult to grasp, but it is clear that the more commonly found *M. subcrenulata* has a longer recorded range. This range is extended downward into lower Cantabrian by a single find reported by Iwaniw (1985). For a recent figuration and discussion of *Mixoneura subcrenulata* from France, see Krings et al. (2000). These authors (op. cit., p. 387) regarded *Mixoneura lingulata* as a successor to *Mixoneura subcrenulata*, thus

suggesting a link between these two taxa.

The taxon represented at Kinney shows an array of morphologies that seem to have been derived from different parts of fronds. One population of specimens consists of pinnae bearing clearly individualized pinnules, each with a rounded shape, a broad basal attachment, and multiple vein entries into the lamina (Fig. 17.1, 17.2, 18.7). Such pinnae typically have large, irregularly triangular, blunt apical pinnules (Fig. 17.1, 17.2). The other population consists of frond segments in which pinnae comprise small pinnules at the base, these fusing into elongate tongue-shaped laminar elements with strong midribs, which terminate the pinnae (Figs. 18.1 – 18.4). In some specimens, this transition is very subdued, and the free pinnules are little more than lobes at the pinna base (Figs. 18.5, 18.6). The more fused pinna segments probably are from more distal, apical portions of the frond, whereas pinnae with more abundant and better-developed free pinnules at their base are from more proximal parts. The mixoneuroids, as a group, had highly variable frond morphologies (e.g., *Odontopteris schlotheimii* as figured by Barthel and Amelang, 2011, fig. 3), with an array of pinnule shapes that, if viewed in isolation, could easily be placed into several species, or even different

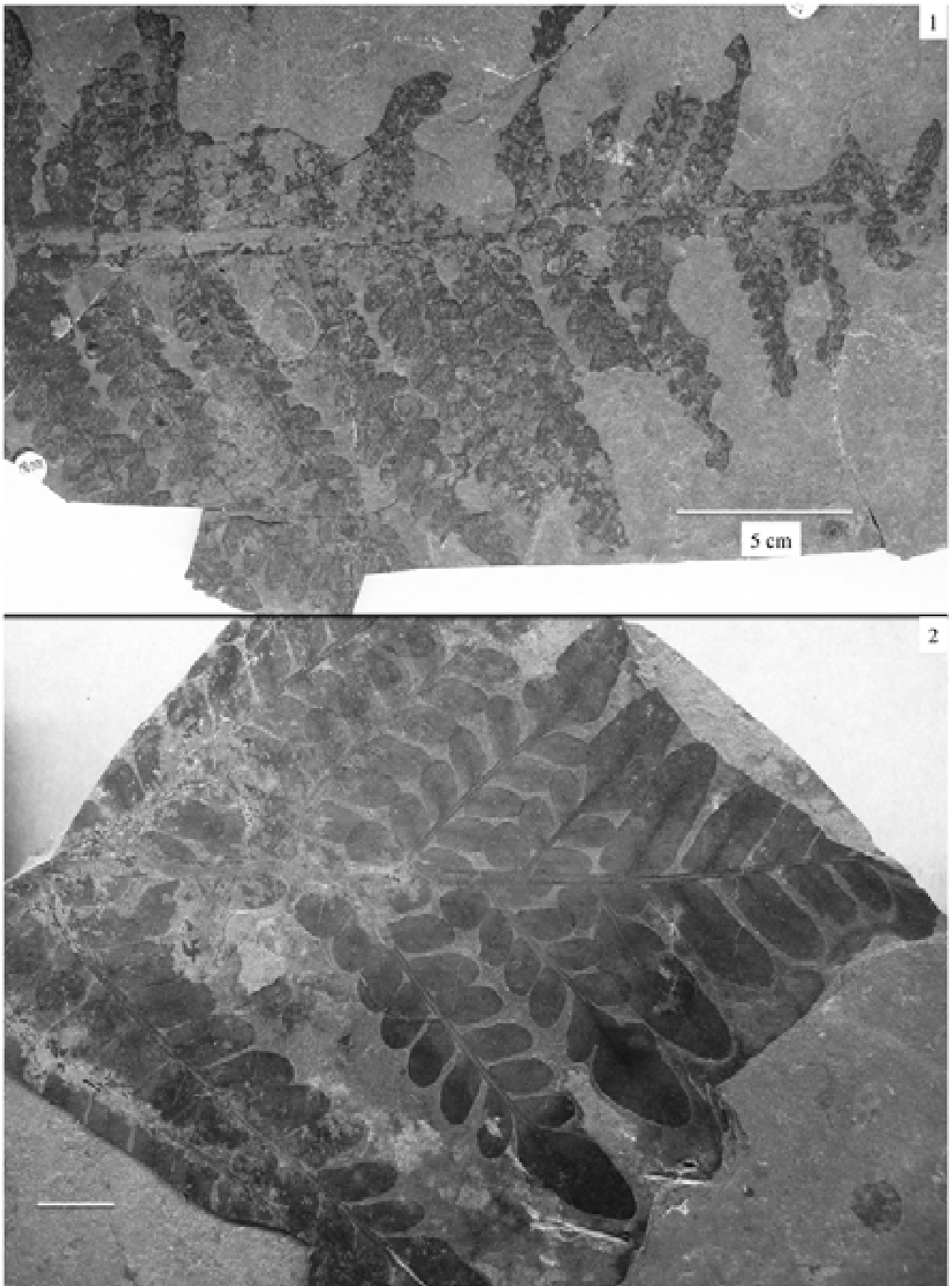


FIGURE 12. Pteridosperm foliage, *Neuropteris ovata*. **12.1**, Large specimen, USNM specimen 422631 (part of left, upper portion of this specimen was illustrated by Mamay and Mapes, 1992, Fig. 4A; specimen edge is a break that has since been repaired). Scale bar = 5 cm. **12.2**, USNM specimen 450781 (Mamay and Mapes, fig. 4C). Scale bar = 1 cm.



FIGURE 13. Pteridosperm foliage, *Neuropteris* cf. *ovata*. 13.1-13.2, USNM specimen 588130. Scale bars = 1 cm.

genera. A large number of specimens is needed in an assemblage to ascertain the range of variability of a single species.

Some of the specimens from Kinney Quarry assigned to *Neuropteris ovata* by Mamay and Mapes (1992; e.g., their fig. 4D) might belong to *Mixoneura subcrenulata*. In comparison with typical *Neuropteris ovata*, these examples show more ovoid pinnules in overall aspect that are not auriculate and are broadly attached to the rachis. A midrib is not developed in all the lateral pinnules of these specimens. They also display rather massive, spoon-shaped terminals.

***Mixoneura* cf. *gimmii* (Remy and Remy) Wagner (Fig. 17.3)**

Within the group of *Mixoneura subcrenulata-lingulata*, there is a third species, *Mixoneura gimmii*, which is quite similar to *Mixoneura subcrenulata*, were it not for a preponderance of neuropteroid pinnules. First described by Remy and Remy (1958), from the Lower Rotliegend of Manebach, Thuringia, it has later been recognized, with abundant

material, from the Autunian of Valdeviar, SW Spain (Wagner and Mayoral, 2007, fig. 25). One specimen from Kinney (Fig. 17.3), showing an apparently quite massive pinna terminal (only partially preserved) and partly mixoneuroid and partly neuropteroid ovoid pinnules, is most comparable. This specimen was figured previously by Mamay and Mapes (1992, fig. 4B) as *Neuropteris* and compared with *Neuropteris heterophylla* Brongniart. Although the latter does show mixoneuroid pinnules in part of the frond, the shape and proportion of the pinnules are quite different, and the comparison with *Neuropteris heterophylla* has been rejected (Lucas et al., 2011). A second specimen with ovoid pinnules and a fairly massive terminal (Fig. 30.3) may possibly be ascribed to this species as well.

The problem with the mixoneuroid fragments in the Kinney flora is the great difficulty in determining intraspecific variation. Because this is a transported and mixed assemblage of plant fragments of varied provenance, it is unlikely that similar plant remains might have originated from the same parent plant.

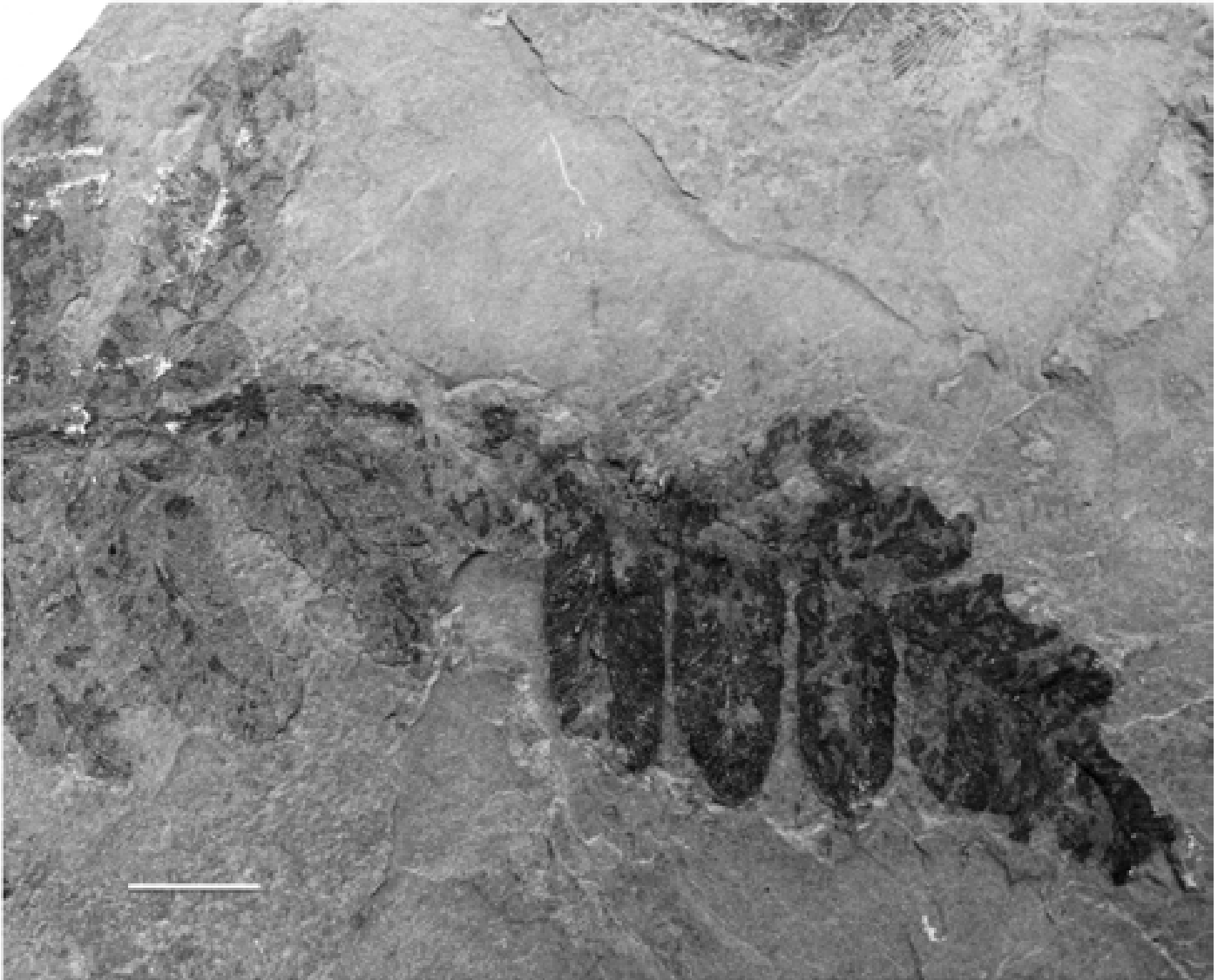


FIGURE 14. Pteridosperm foliage, *Neuropteris cordata*, USNM specimen 558148. Scale bar = 1 cm.

***Neurodontopteris auriculata* (Brongniart) Potonié (Figs. 10.3, 19.1, 19.2, and possibly Fig. 20)**

Among the most common fragmentary pteridosperm remains in the Kinney flora are those of a neuropteroid very similar to that described by Knaus and Lucas (2004) from the Carrizo Arroyo locality in central New Mexico (latest Pennsylvanian to earliest Permian). The plant has highly variable foliage morphology, with pinnule attachment ranging from neuropteroid to mixoneuroid. Knaus and Lucas (2004) recognized the different fragments as belonging to a single species due to the preservation of relatively large frond segments. However, the frond reconstruction of Knaus and Lucas (2004, fig. 1) is likely not justified by the fragmentary remains figured. With regard to the specimens encountered in Kinney Quarry, the pinnules are generally elongate with rounded apices and bases (Fig. 19.1, 19.2). They have dense lateral veins and a variably distinct midrib. Some pinnules are markedly elongate with both basisopic and acrosopic auriculae (Fig. 10.3, Fig. 20). These auriculate forms enter into the range of variation of *Neurodontopteris auriculata* (Laveine and Dufour, in press), particularly in their variable pinnule shapes and insertion angles, a dense but often somewhat variable lateral venation, and a midrib that is strong at the base and dissipates in the upper third of the pinnule.

Neurodontopteris auriculata was distinguished as a separate generic entity by Potonié (1893) on the basis of possessing both neuropteroid and odontopteroid pinnule attachment. He compared with *Mixoneura* Weiss, which was based on “*Odontopteris*” *subcrenulata* (Rost) Zeiller, and admitted that small fragments of the latter might be confused with partial remains of *Neurodontopteris auriculata*. He quite correctly saw Brongniart’s species as more neuropteroid than odontopteroid, while *Mixoneura* was taken out of *Odontopteris* (*sensu lato*). Cleal and Shute (1995) briefly discussed *Neurodontopteris auriculata*, which they apparently accepted as a monotypic genus, but did not compare it with *Mixoneura*, presumably because they regarded the latter as an odontopterid and, thus, not immediately comparable. Laveine and Dufour (in press) recently discussed the frond morphology and variation in pinnule shape and attachment of *Neurodontopteris auriculata*.

Doubinger et al. (1995, p. 199, 202) figured *Neurodontopteris auriculata* from its type area, the Saint Étienne Basin in central France (Brongniart, 1831, p. 236), and referred it to *Neuropteris osmundae* (Artis) Kidston, following a synonymy published by Crookall (1959). However, this rare Westphalian species is known too incompletely for a proper identification with *Neurodontopteris auriculata*.

Neurodontopteris auriculata is also a rare element, although ap-

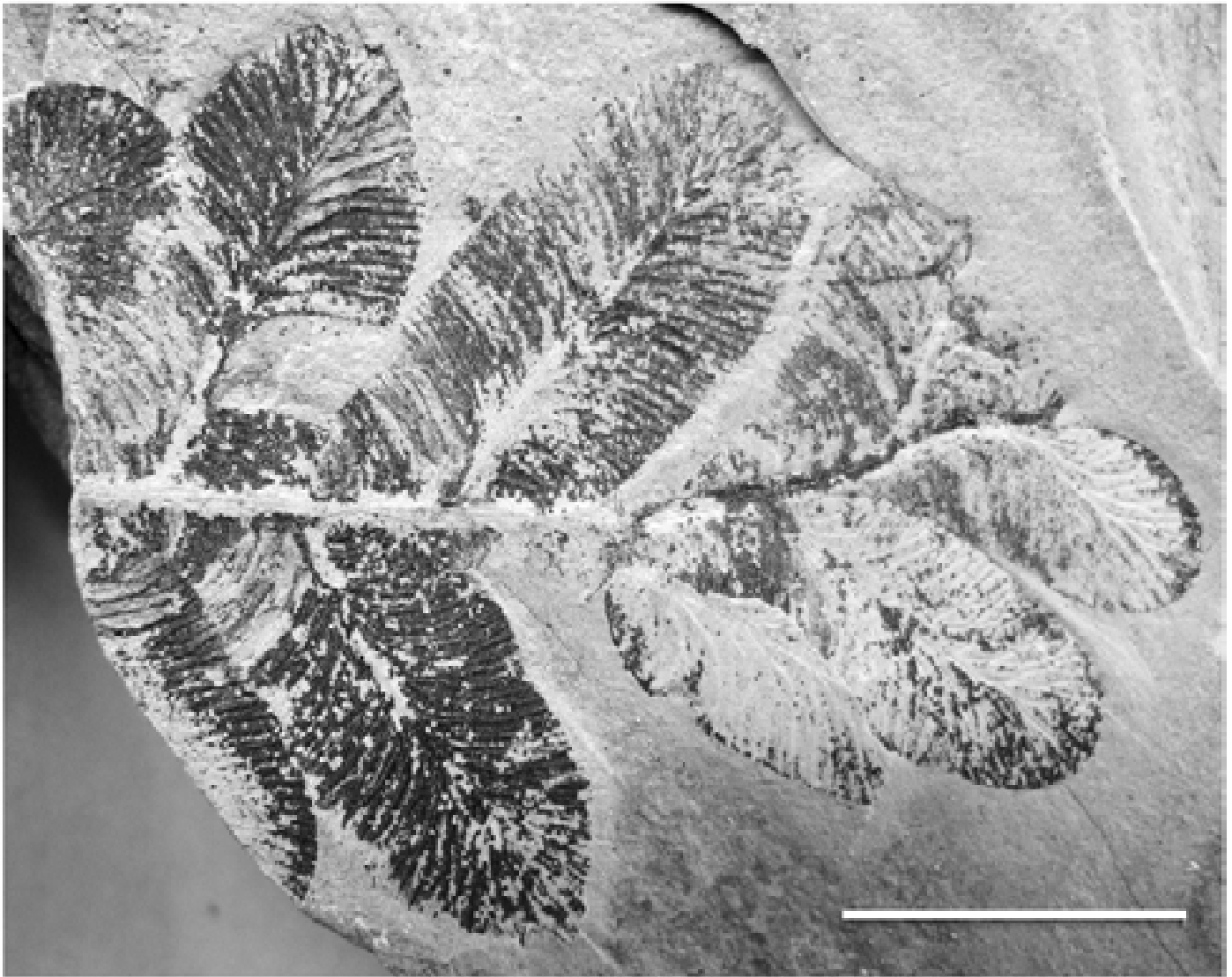


FIGURE 15. Pteridosperm foliage, *Alethopteris schneideri*, USNM specimen 558145. Scale bar = 1 cm.

parently quite widespread. It is commonly reported from high Stephanian strata, but there is a mid-Stephanian (Saberian) record from NW Spain (Wagner, 1963), albeit on very fragmentary remains. It may be an element of drier habitats, although in a general “coal-measure” context.

***Neurocallipteris planchardii* (Zeiller) Cleal, Shute and Zодrow (Fig. 19.3)**

A small number of specimens in the collections conform fairly well to the taxon *Neurocallipteris planchardii* (Figs 19.3) (see Zeiller, 1906; Doubinger et al., 1995; Laveine and Dufour, in press). The specimens have neuropteroid pinnules with a narrow attachment to the rachis, a cordate base, and almost parallel lateral margins that taper toward a rounded apex. The lateral venation is dense and arcs strongly to meet the margin at almost right angles, a strongly marked midrib is present, reaching to near the pinnule apex. *Neurocallipteris planchardii* is a rare element, occurring throughout the Stephanian, but often recorded from single pinnules, which have not always been identified beyond doubt.

***Pseudomariopteris cordato-ovata* (Weiss) Gillespie, Clendening and Pfefferkorn (Figs. 21, possibly 30.1)**

Pseudomariopteris Danz -Corsin belongs to the callistophytalean pteridosperms, a group of small, scrambling to climbing forms (Krings

and Kerp, 2000). In the Kinney flora, only a few fragmentary specimens could be attributed to this genus, specifically to *P. cordato-ovata*. The specimens conform to this species, as re-diagnosed by Krings and Kerp (2000), in having small, more or less ovoid pinnules with a constriction above the wide basal attachment and a thick-lamina, which often obscures the venation, leaving only the short midrib well marked (Fig. 21). No pinnae with the characteristic tendril-like terminals have been collected. Elsewhere in the paleoequatorial belt, *P. cordato-ovata* is a common element (Wagner and  lvarez-V zquez, 2010a; Tabor et al., 2013). The species has a long stratigraphic range, extending from the upper Middle Pennsylvanian (upper Asturian) into the highest Upper Pennsylvanian (Autunian) (Wagner and  lvarez-V zquez, 2010a).

***Sphenopteridium manzanitanum* Mamay (Figs. 22.1, 22.2)**

Fairly abundant material of a small bifurcate frond with sphenopteroid foliage has been described by Mamay (1992) as a new species of *Sphenopteridium*, a generic attribution normally applied to plants of Mississippian age. Mamay’s determination of *Sphenopteridium manzanitanum* is based on the bifurcate frond morphology, the deeply incised rather thick-limbed pinnules with a fan-shaped venation and, above all, the faint imprint of a fertile structure comparable to that of *Diplopteridium teilianum* Walton and identified as *Telangiospis*. Mamay’s

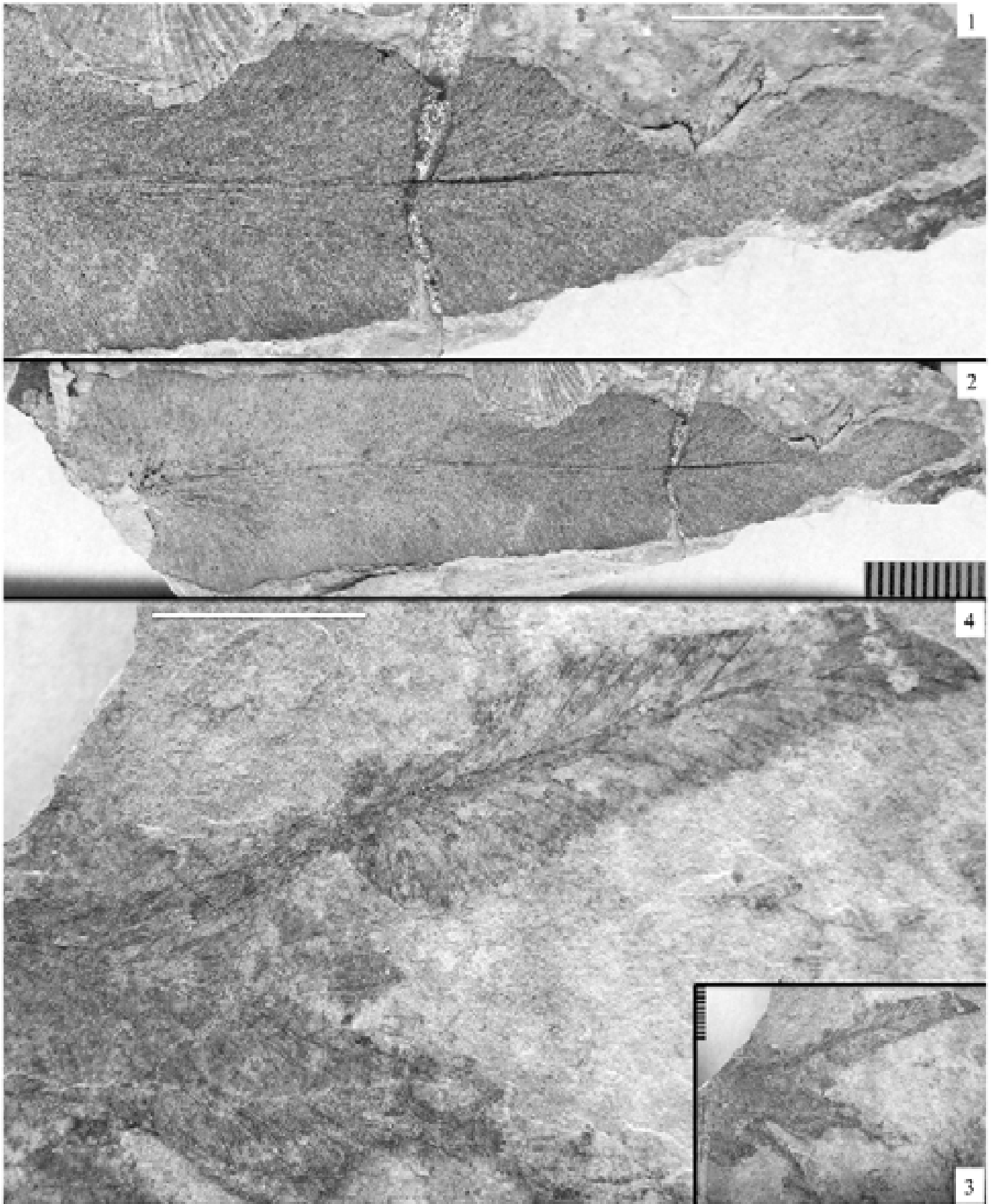


FIGURE 16. Pteridosperm foliage, cf. *Blanziopteris praedentata*. **16.1-16.2**, USNM specimen 558135. **16.3-16.4**, USNM specimen 558136. Scale bars in **13.1** and **13.4** = 1 cm. Scale bars in **13.2** and **13.3** in mm.

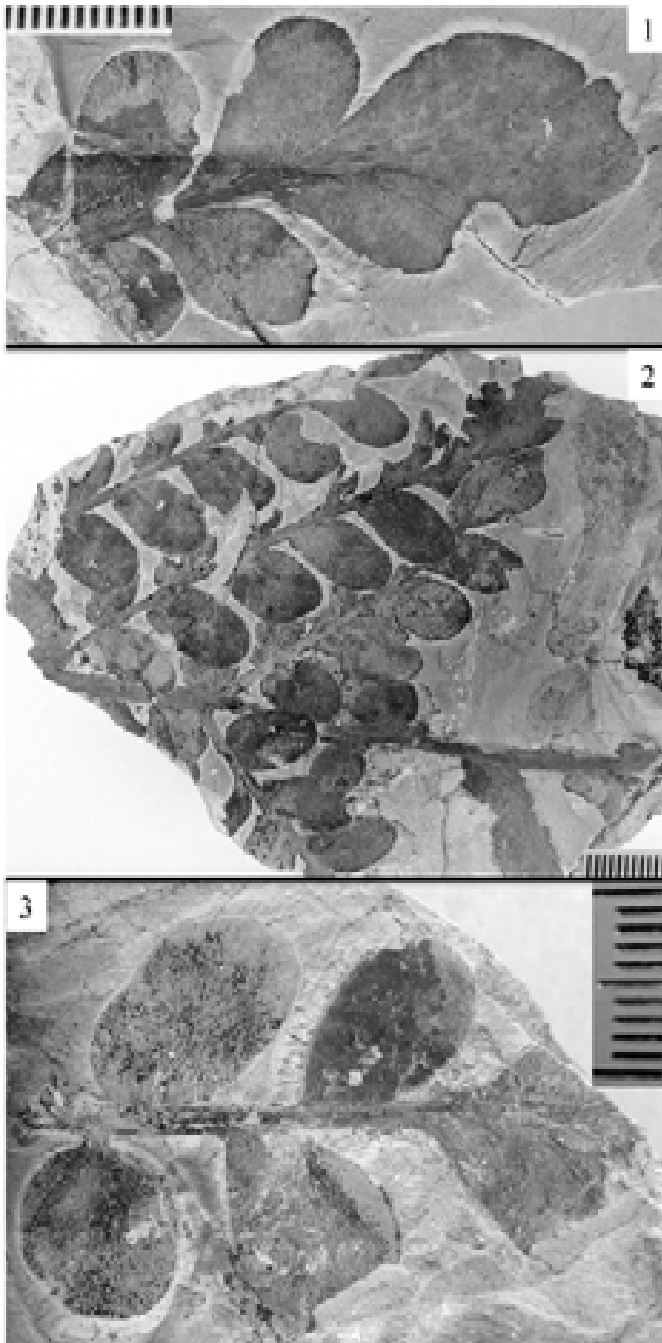


FIGURE 17. Pteridosperm foliage. **17.1**, *Mixoneura subcrenulata*, USNM specimen 450788, note marginal arthropod feeding damage, especially on large terminal pinnule (Mamay and Mapes, 1992, Fig. 3E). **17.2**, *Mixoneura subcrenulata*, USNM specimen 450782, note marginal arthropod feeding damage, especially on large terminal pinnule (Mamay and Mapes, 1992, Fig. 4D). **17.3**, *Mixoneura* cf. *gimmii*, USNM specimen 450783 (Mamay and Mapes, 1992, Fig. 4B). Scale bars in mm.

description is impeccable, but it is admitted that the fertile material is poorly preserved and that its attachment to the vegetative frond has not been demonstrated (even though its relative position is suggestive).

Mamay's attribution of this material to *Sphenopteridium* was surprising because this genus was regarded as exclusively Mississippian. However, the present writers regard this attribution as worthy of serious consideration, and his suggestion that the Kinney material would represent a "lazarus" taxon is quite reasonable. However, Mamay (1992) did

not compare his specimens with *Sphenopteris germanica* Weiss (1879). This Lower Rotliegend plant shows the same bifurcate frond structure (Remy and Remy, 1977, bild 87c; Rössler and Barthel, 1998, pl. X, fig. 2) and pinnule shape and venation as *Sphenopteridium manzanitanum*, with a degree of morphological identity that allows considering both species as probably one and the same. Fertile axes with synangia composed of partly coalesced sporangia have been described as *Schuetzia anomala* by Geinitz (1863), and assigned to *Sphenopteris germanica* by Gothan (1937). The latter pointed out that *Schuetzia anomala* possesses bifurcate axes with short transverse bars (op. cit., taf. 27, figs 2, 3), which are also seen in the associated remains of *Sphenopteris germanica*. Gothan (1937) also mentioned the consistent association of *Schuetzia* with *Sphenopteris germanica* in several Lower Rotliegend localities in Germany. Remy and Rettschlag (1954) macerated spores from *Schuetzia anomala*, which they regarded as a male sporangiate structure. Although the attribution of this presumed male synangiate structure to "*Sphenopteris*" *germanica* cannot be regarded as fully secure, the circumstantial evidence points in this direction. It would also suggest that "*Sphenopteris*" *germanica* is a pteridosperm.

The fertile specimen associated with *Sphenopteridium manzanitanum* is too poorly preserved to allow a straightforward identification with *Schuetzia anomala*, but there is a vague resemblance. It is clear that the taxa discussed above need more complete material with well-preserved fertile structures in connection with the vegetative frond. The existing information can only be regarded as tentative.

However, there is no reason not to follow Mamay (1992) in using *Sphenopteridium* as a generic identification for the specimens, which he described as a new species, but which could be identical with "*Sphenopteris*" *germanica*. The present writers thus conclude that this plant should be referred to as *Sphenopteridium germanicum*, a formal synonymy of which will need to be presented elsewhere. It also is reasonable to regard this taxon as a probable pteridosperm.

Coniferophytes

Walchian Conifers (Fig. 23)

At least two, and possibly three, types of "walchian" conifers occur as a conspicuous part of the Kinney flora (Fig. 23). The reader is referred to Looy (2007) for a discussion of various conifer groups, the complexities of conifer taxonomy, and their phylogenetic relationships.

Mamay and Mapes (1992) assigned the Kinney conifer remains to two species, *Walchia piniformis* (Schlotheim) Sternberg and *Hermitia* (*Walchia*) cf. *schneideri* (Zeiller) Kerp and Clement-Westerhof. We concur with the *W. piniformis* identification and Figure 23.3 is a re-illustration of their specimen (Mamay and Mapes, 1992, Fig. 8A). *Walchia piniformis* is a very widely occurring species in the Pennsylvanian and Early Permian throughout the Pangaeian tropics. The specimen Mamay and Mapes (1992, Fig. 8D) identified as *Hermitia* (*Walchia*) cf. *schneideri* based on its long, lax leaves is re-illustrated here as Figure 23.1. The long leaves may reflect its position on a lateral branch, perhaps as the principal axis, or as the stem of a small individual of *W. piniformis*, rather than being representative of a distinct species. Comparable variation can be observed in juvenile specimens of some modern araucarian conifers with plagiotropic branching patterns, such as *Araucaria heterophylla*, the Norfolk Island Pine. Somewhat more distinctive is a specimen (Fig. 23.2) that we attribute to the genus *Culmitzschia* (Mamay and Mapes, 1992, Fig. 8B), identified the specimen as *W. piniformis*). This type of conifer foliage has been reported widely in rocks of Stephanian through Early Permian age in Europe (e.g., Kerp et al., 2007; Bercovici and Broutin, 2008; Šimunek and Martínek, 2009; Wagner and Álvarez-Vázquez, 2010a) and in the western USA (e.g., Chaney and DiMichele, 2007; Lucas et al., 2012; Voigt et al., 2013). As with *Walchia*, *Culmitzschia* is almost always preserved in sedimentary settings and in floristic associations that suggest seasonally dry climates, though not necessarily without significant moisture availability. The described species most

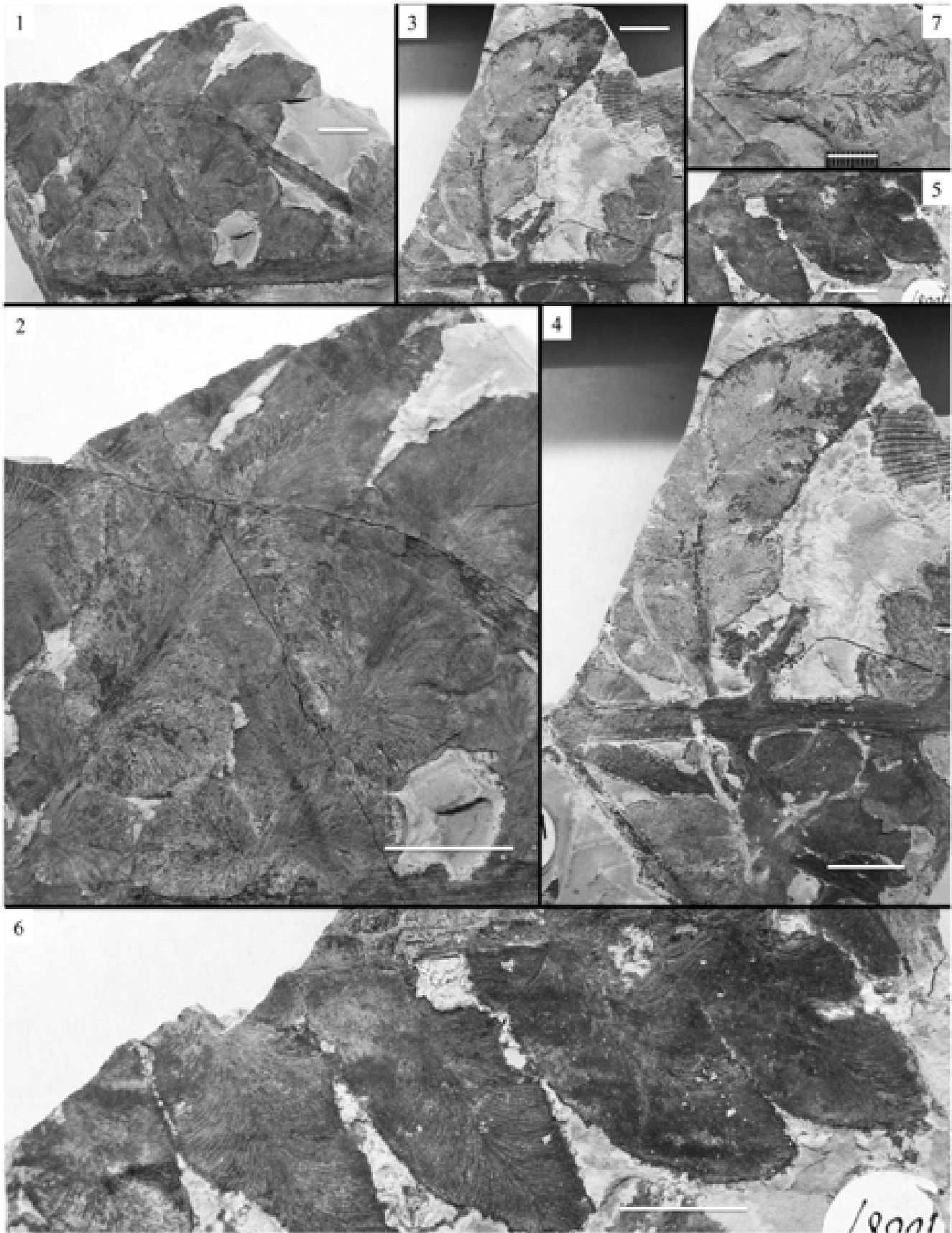


FIGURE 18. Pteridosperm foliage, *Mixoneura subcrenulata*. **18.1-18.2**, USNM specimen 558139, Note evidence of marginal damage due to possible arthropod feeding. **18.3-18.4**, USNM specimen 558147. **18.5-18.6**, USNM specimen 558124, Note evidence of marginal damage due to possible arthropod feeding. **18.7**, USNM specimen 558149. Scale bars = 1 cm.

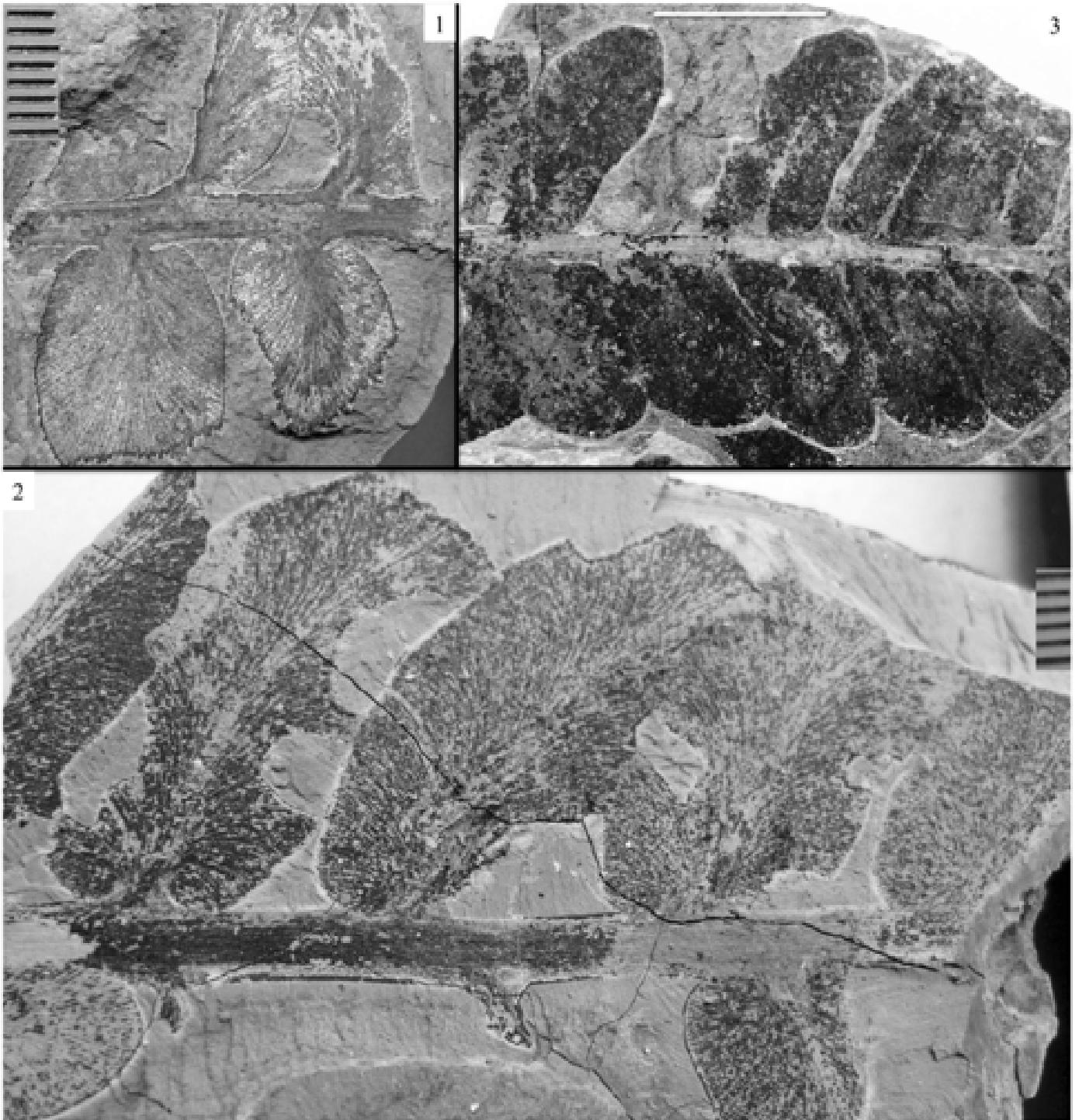


FIGURE 19. Pteridosperm foliage. **19.1**, *Neurodontopteris auriculata*, USNM specimen 558143. Scale bar in mm. **19.2**, *Neurodontopteris auriculata*, USNM specimen 558142, note marginal arthropod feeding damage. Scale bar in mm. **19.3**, *Neurocallipteris planchardii*, USNM specimen 558141. Scale bar = 1 cm.

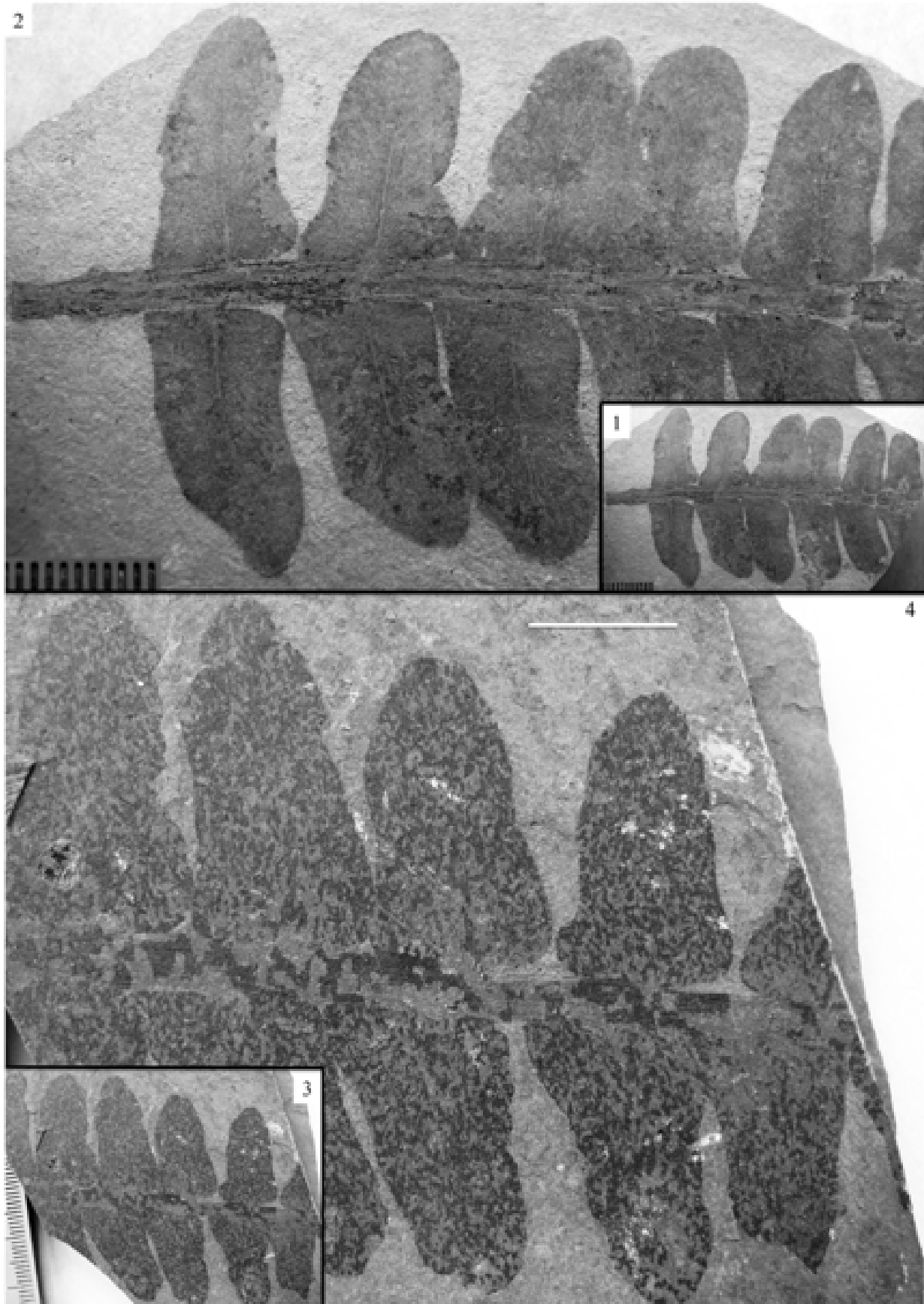


FIGURE 20. Pteridosperm foliage, *Neurodontopteris* cf. *auriculata*. **20.1-20.2**, USNM specimen 450789 (Mamay and Mapes, 1992, Fig. 5E). **20.3-20.4**, USNM specimen 558127. Scales in **20.1**, **20.2** and **20.3** in mm. Scale bar in **20.4** = 1 cm.

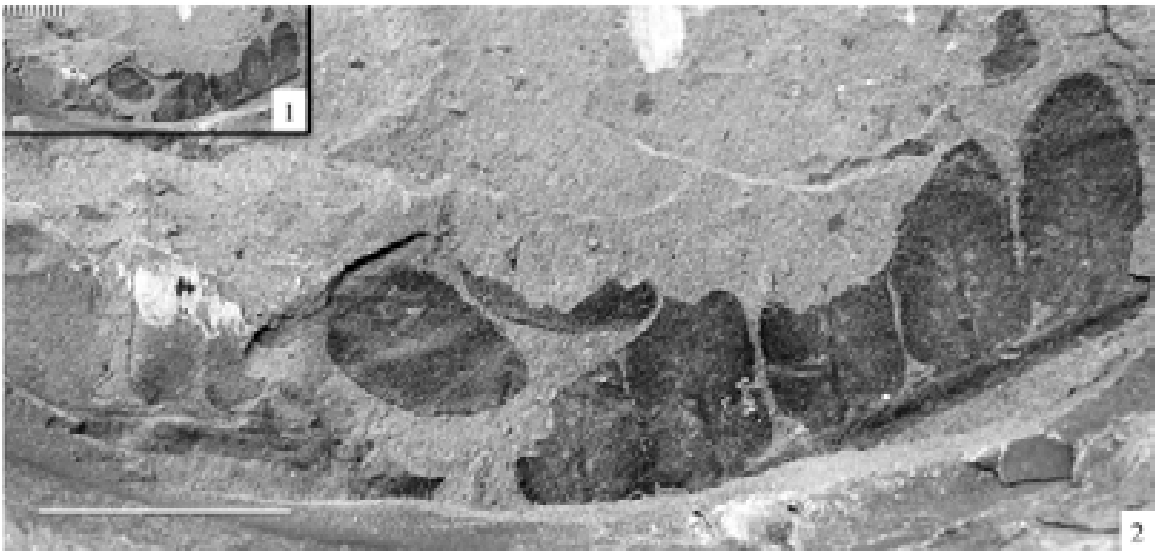


FIGURE 21. Pteridosperm foliage, *Pseudomariopteris cordato-ovata*. 21.1-21.2, USNM specimen 558125. 21.1 scale bar in mm. 21.2 scale bar = 1 cm.

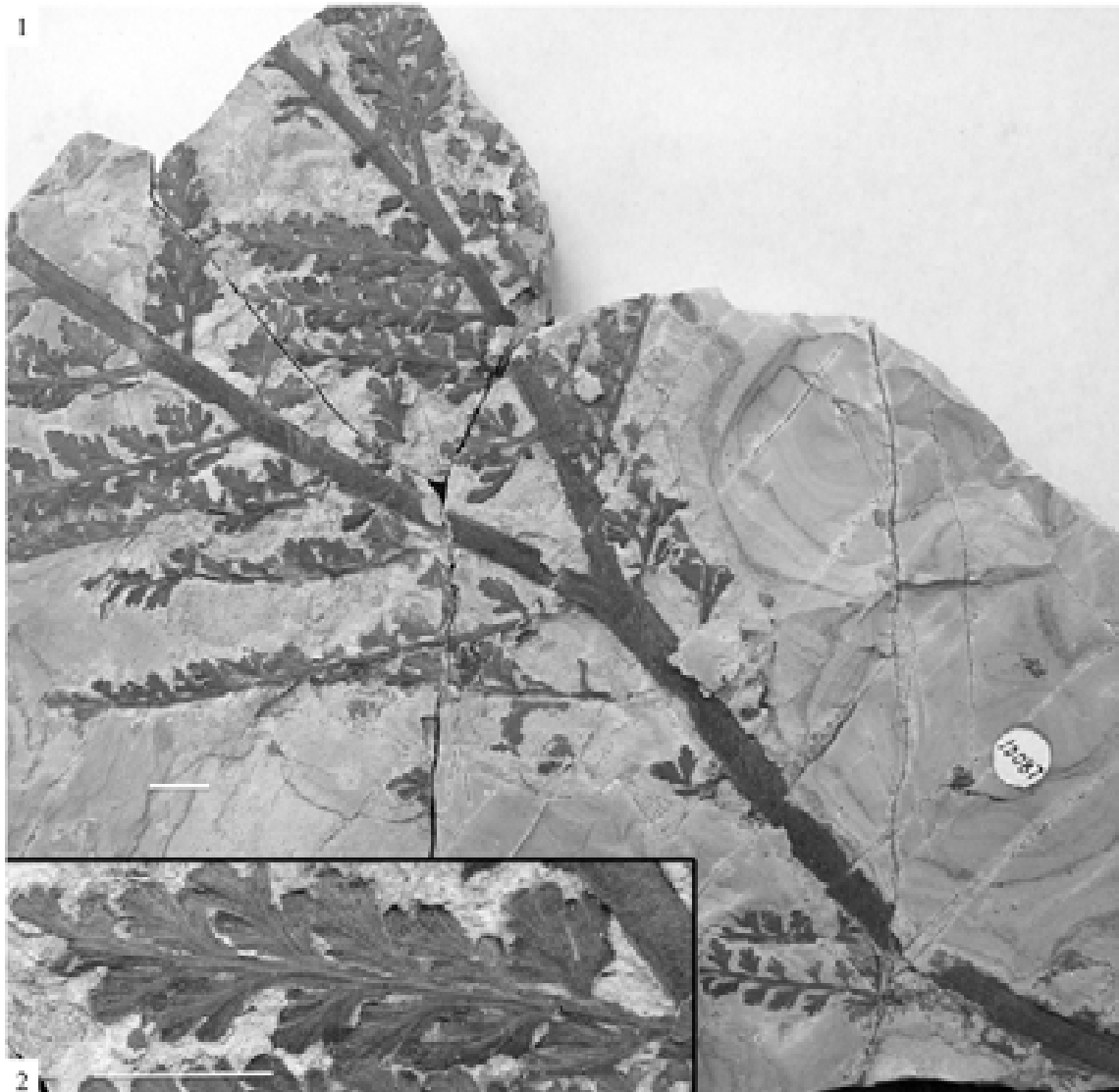


FIGURE 22. Pteridosperm foliage, *Sphenopteridium manzanitanum*. 22.1-22.2, Holotype, USNM specimen 450796 (Mamay and Mapes, 1992, Fig. 6E). 22.2 is an enlargement of a pinnae from 22.1. Scale bars = 1 cm.

similar to those in the Kinney collections are *Culmitzschia laxifolia* (Florin) Clement–Westerhof and *Culmitzschia speciosa* (Florin) Clement–Westerhof.

***Ernestiodendron filiciforme* (Schlotheim) Florin (Fig. 24)**

Also in the Kinney collections are rare specimens that are morphologically similar to *Ernestiodendron filiciforme* (Fig. 24). The fragments bear leaves that are inserted nearly normal to the axis and curve strongly upward over the last third of their length. The dark color of the specimens is typical of *Ernestiodendron* and reflects the thickness of the leaf in life. *Ernestiodendron*, like *Walchia* and *Culmitzschia*, occurs widely in both European/North African (in addition to the above citations, see: Kerp et al., 1990; Hmich et al., 2006; Galtier and Broutin, 2008) and American (e.g., Hotton et al., 2002) floral assemblages, generally with other conifers and in association with physical conditions and plant types indicative of seasonally dry climatic conditions. Hernández-Castillo et al. (2003) suggested that *Ernestiodendron* may have been a tree of small stature. The presence of *Ernestiodendron* in the Kinney flora is, to our knowledge, the oldest reported occurrence of this genus in North America.

***Dicranophyllum readii* Mamay (Fig. 25)**

The long, repeatedly bifurcate leaves figured and described as *Dicranophyllum readii* by Mamay (1981) (see also Mamay and Mapes, 1992, Figs. 7E, F) have been assigned correctly to *Dicranophyllum*. Whether or not a new species should have been recognized on the basis of isolated leaves, is more debatable. *Dicranophyllum* has been described more completely by Barthel et al. (1998), recording axial remains as well as the characteristic long, repeatedly bifurcate, slender leaves. These authors also described the male and female cones inserted on the axes of this small gymnospermous tree, for which they provided a reconstruction. A long list of *Dicranophyllum* leaf taxa described from North America is quoted by Mamay (1981). Now that this plant is known more comprehensively, it may be regarded less useful to distinguish species of *Dicranophyllum* exclusively on size and bifurcation of the leaves, distinctive though they may be. A large number of species has been described in the past on the basis of leaf remains only (e.g. *Dicranophyllum glabrum* (Dawson) Stopes, *Dicranophyllum lusitanicum* (Heer) Lima). *Dicranophyllum readii* differs from the relatively common species *Dicranophyllum gallicum* Grand'Eury by its very long (up to 50 cm length), at least twice bifurcate leaves. These are also described as very slender and not, as it appears, quite as sturdy as those of other species, e.g. *Dicranophyllum gallicum*. In terms of size, *Dicranophyllum readii* is most similar to *Dicranophyllum glabrum* (see Wagner, 2005), which has leaves in excess of 20 cm length. The preservation of almost entire *Dicranophyllum* leaves in Kinney Quarry may be ascribed to the relatively sturdy nature of these slender leaves, making them resistant to maceration.

Dicranophyllum has a long stratigraphic range, encompassing the entire Pennsylvanian. It also has a wide geographic distribution throughout the paleoequatorial realm (from New Mexico in the west to China and Sumatra in the east) (e.g., Stockmans and Mathieu, 1939; Barthel and Noll, 1999; DiMichele et al., 2001; Ziegler et al., 2002; van Waveren et al., 2007; Gastaldo et al., 2009; Wagner and Castro, 2011).

Coniferophyte Reproductive Organs

Conifer reproductive organs are represented in the Kinney flora almost entirely by platyspermic (flat), winged seeds. However, once dispersed, the seeds cannot be solely attributed to conifers with confidence. A single specimen of the cone bract *Gomphostrobus* is illustrated in Figure 26.1.

Cordaitaleans

The cordaitaleans are an extremely widespread and diverse group

of plants in Pennsylvanian and Permian floras. They are reported from swampy habitats (Costanza, 1985; Raymond et al., 2010) and from environments with significant moisture limitation (Falcon-Lang, 2003, 2004). These plants are generally a common to abundant component of floras in moisture-stressed habitats, which appears to be the case for the landscape from which the Kinney flora was drawn. However, as noted by Mamay and Mapes, (1992), cordaitaleans are a rare component of the Kinney macroflora, represented only by a few leaf fragments (see their figure 7D) and a reproductive axis, *Cordaitanthus* (their figure 7C, reillustrated here as Figure 26.2). Interestingly, this stands in sharp contrast with the palynological analysis of Willard (1992), in which cordaitalean pollen is a dominant to abundant element, as is conifer pollen in some samples. The rarity of cordaitalean foliage in the macrofloral suite is an anomaly. Transport selection would not have normally eliminated this element.

Noeggerathiales

The noeggerathialeans are heterosporous, lower vascular plants; current understanding is reviewed by Wang et al. (2009), who also describe whole plants preserved in ash beds. A number of rare plants from Stephanian (Autunian) strata having vague resemblance to the Mesozoic Cycadales and Bennettitales have been assigned to this group, including *Sphenozamites rochei* Renault 1882, *Plagiozamites planchardii* (Renault) Zeiller 1894, and *Charliea manzanitana* Mamay 1990. In all cases the leaves are ovate to elongate with parallel veins and are narrow-based, not quite stalked and are inserted on straight, rather sturdy axes. In these taxa the leaves are positioned in two-dimensional, very flat arrays, such that the helical leaf arrangement is not immediately obvious. Differences between these leaf taxa are found in length-breadth ratios, entire as against incised distal margins and the ending of all veins on the distal margin, as against some veins abutting onto the lateral margins. These variables may allow segregation.

***Plagiozamites rochei* (Renault) comb. nov. (Figs. 27.1, 27.2)**

A second noeggerathialean element from the Kinney flora was described by Mamay and Mapes (1992) as *Plagiozamites planchardii* (Renault) Zeiller, a taxon described originally from the upper Stephanian of Commeny, south-central France (Renault and Zeiller, 1888-1890), and recorded subsequently from the Rotliegend of Trienbach (Weilerthal) in the Vosges (Alsace, France) by Zeiller (1894), who introduced the genus. Zeiller compared it with both *Noeggerathia* and *Zamites*, and considered it to be a cycadopsid, on the basis of leaf morphology, but noted the absence of fertile material. In North America, the species has been recorded from the Missourian Conemaugh Formation of the Appalachian Basin, USA (Bassler, 1916), which is approximately same age as the Kinney Quarry deposit, from an unidentified Pennsylvanian location in New Mexico in collections made by Charles B. Read (noted by Ash and Tidwell, 1982), and from the Lower Permian Abo Formation of southern New Mexico (DiMichele et al., 2013).

Plagiozamites planchardii is here identified with *Sphenozamites rochei* Renault, a European species recorded from high Stephanian strata in France and Spain (Renault, 1882; Wagner and Mayoral, 2007). *Sphenozamites* is another Mesozoic genus belonging to the cycadopsids. Both Renault (in Renault and Zeiller, 1888) and Zeiller (1894) overlooked *Sphenozamites rochei* Renault 1882, which seems the same species and has priority. The present writers thus suggest that this element should be recorded as *Plagiozamites rochei* (Renault) comb. nov. [Basionym: *Sphenozamites rochei* Renault, 1882, Archives botaniques du Nord de la France, p. 180-184, 1 fig.].

***Charliea manzanitana* Mamay (Figs. 27.3, 27.4)**

Mamay (1990) described some of the specimens from Kinney Quarry as *Charliea manzanitana*, which shows oval leaves with more-or-less parallel lateral margins and a distal margin with, characteristically, two incisions, dividing the apex into three lobes. The parallel veins are

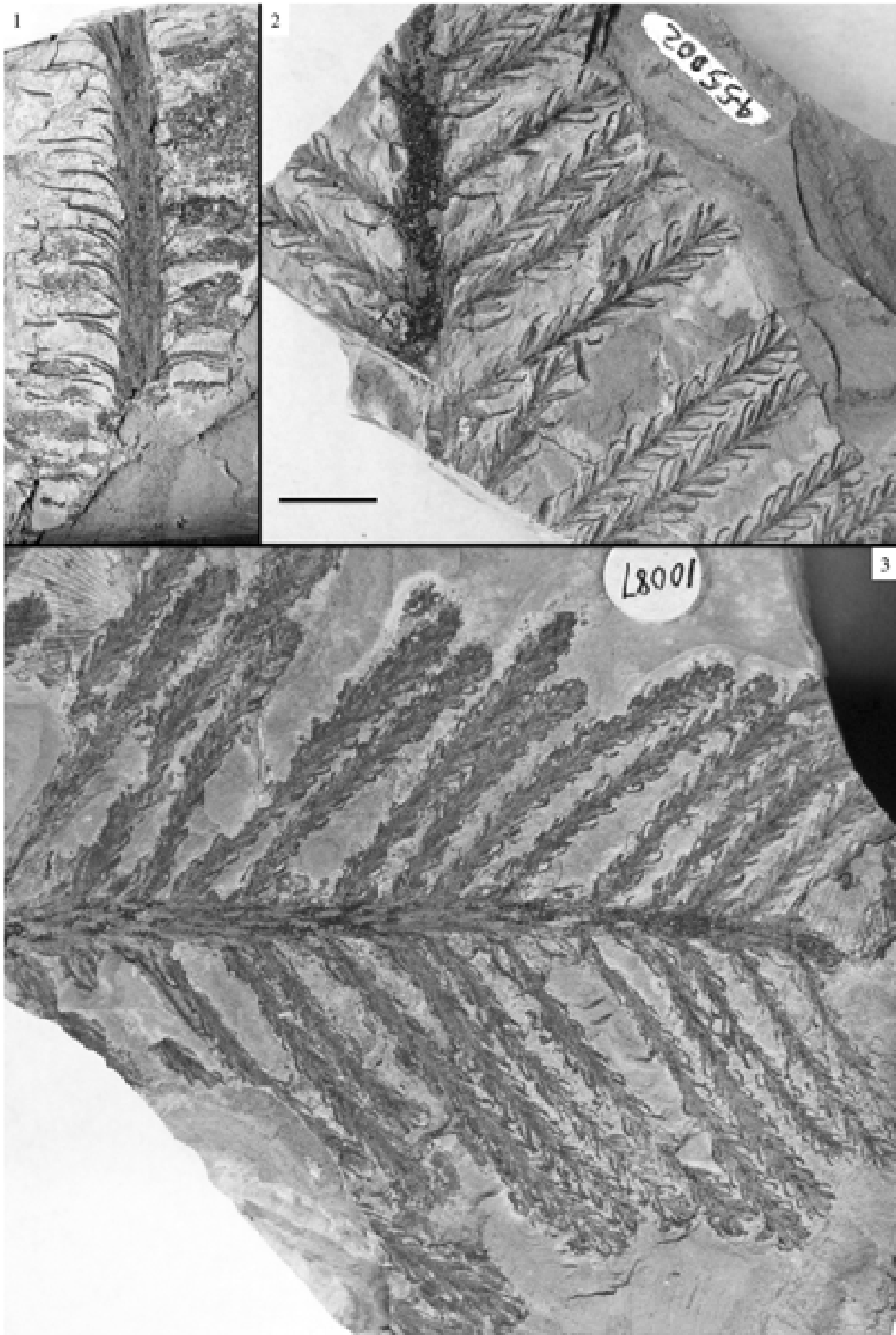


FIGURE 23. Conifers. **23.1**, *Walchia* sp., USNM specimen 455001 (Mamay and Mapes, 1992, Fig. 8D). **23.2**, *Culmitzschia* sp., USNM specimen 455002 (Mamay and Mapes, 1992, Fig. 8B). **23.3**, *Walchia piniformis*, USNM specimen 450803 (Mamay and Mapes, 1992, Fig. 8A). All same magnification, scale bar in **23.2** = 1 cm.

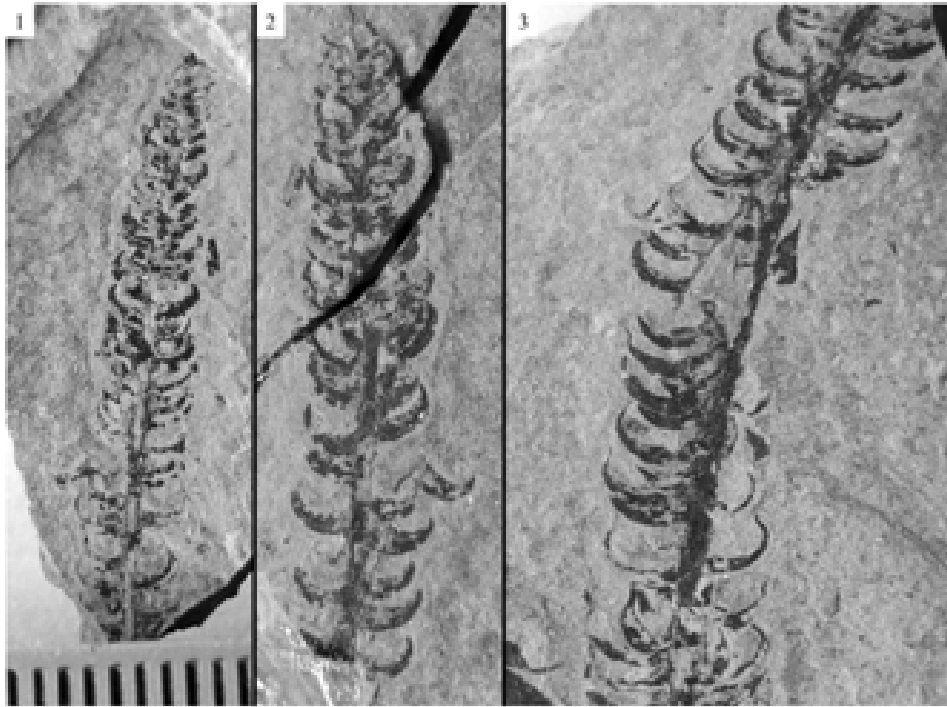


FIGURE 24. Conifers, *Ernestiodendron filiciforme*. **24.1**, USNM specimen 558128. **24.2**, USNM specimen 558129. **24.3**, USNM specimen 558133. All same magnification. Scale bar in **24.1** in mm.

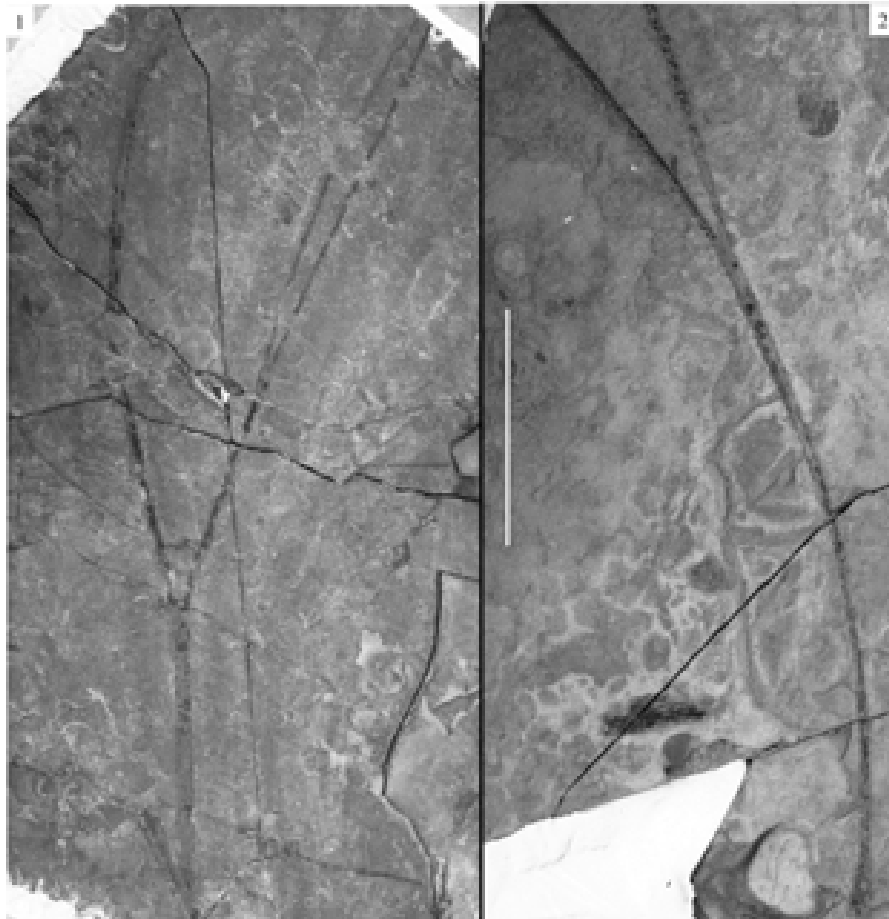


FIGURE 25. Conifers, *Dicranophyllum readii*. **25.1**, Holotype, USNM specimen 267277. **25.2**, USNM specimen 267281 (Mamay and Mapes, 1992, Fig. 7E). Both same magnification. Scale bar = 10 cm.

straight, display occasional dichotomies, and abut exclusively onto the distal margin. The most complete specimen of *C. manzanitana* is figured by Taylor et al. (2009, p. 661, fig. 16.22). Mamay compared *Charliea* with Chinese taxa such as *Tingia* and also with *Russellites* Mamay (an American genus described by Mamay, 1968, for specimens originally attributed to *Tingia* by Darrah, 1938, and later put into synonymy with *Yuania* by Wang and Chaney, 2010). Comparison also was made with *Noeggerathia*, a rare genus recorded from Westphalian strata in Bohemia, Czech Republic, and Saarland, Germany (Nemejc, 1928; Hirmer, 1940). *Noeggerathia foliosa* Sternberg shows terminal strobili (*Noeggerathiostrabus*), which allow recognition of the Noeggerathiales as a special group of heterosporous pteridophytes. A possible link to the progymnosperms is often assumed (Hirmer, 1940, compared with *Archaeopteris*), but this is not wholly accepted (see comments in Taylor et al., 2009, p. 497). Although Mamay (1968, 1990) suggested that *Russellites* and *Charliea* might be early cycadalean (compound) leaves, Tidwell and Ash (2003) demonstrated that what Mamay (1990) interpreted as pinnules actually are leaves. The Noeggerathiales seem to occur throughout the Pennsylvanian, but are rare fossils typically linked to better drained soils such as occur on hill slopes, but under wet climatic conditions.

Cycadopsids?

Two fragmentary specimens of *Taeniopteris* were discovered in the NMNH collections from Kinney Quarry (Fig. 28). This is a morphogenus characterized by leaves with a strong central vein and perpendicular to subperpendicular lateral veins, which may bifurcate near the central vein. Apart from *Taeniopteris jejuna* Grand'Eury, which Remy (1953) excluded as a fern with synangia (*Ilfeldia*), the majority of *Taeniopteris* might be cycadopsids (see Cridland and Morris, 1960; Gillespie and Pfefferkorn, 1986; Axsmith et al., 2003). *Taeniopteris* leaves are extremely variable in size. The veins may branch, and often do so relatively close to the midvein. The various species are distinguished on relative leaf width and vein density; they also are distinguished to a lesser extent on shape, since *Taeniopteris* leaves are normally found in fragmentary condition. This is certainly the case for the Kinney material, which is clearly affected by maceration and is highly fragmentary.

Remy and Remy (1975) attempted to tabulate the various characters such as leaf width, constitution of central vein, vein density and presence or absence of vein bifurcation. This useful exercise served to show that quite a number of species were introduced on inadequate material, and that the undoubtedly large amount of intraspecific variation was not taken into account sufficiently.

The two small fragments collected from the Kinney Quarry belong to a relatively broad-leaved form with dense venation (c. 35-40 veins/cm), oriented perpendicular to both the central vein and leaf margin. These characters fall within the range of *Taeniopteris fallax* (Göppert, 1864-65) and *Taeniopteris multinervia* (Weiss, 1869), as tabulated by Remy and Remy (1975). Leaf shape is unknown. *Taeniopteris multinervia* is the most common identification in Europe. However, Barthel (1976) placed *Taeniopteris multinervia* in synonymy with *Taeniopteris abnormis* (Gutbier, 1835). This synonymy was accepted by Wagner and Álvarez-Vázquez (2010a), and should probably be respected for the leaf fragments found in the Kinney Quarry. It is further noted that Halle (1927) placed *Taeniopteris smithii* (Lesquereux, 1879-80), from Alabama, USA, in synonymy with *Taeniopteris fallax*. In view of the fragmentary nature of the two specimens from Kinney, it seems most reasonable to record these as *Taeniopteris* cf. *abnormis*, this being the earlier described species. Whether or not the various species mentioned warrant distinction as separate taxonomic entities (see discussion in Wagner and Martínez-García, 1982), must await the availability of more complete specimens. *Taeniopteris lescuriana* (Fontaine and White, 1880) described from an equally fragmentary specimen in the Dunkard Group of West Virginia, may well be another synonym. Such remains, recorded as *Taeniopteris*

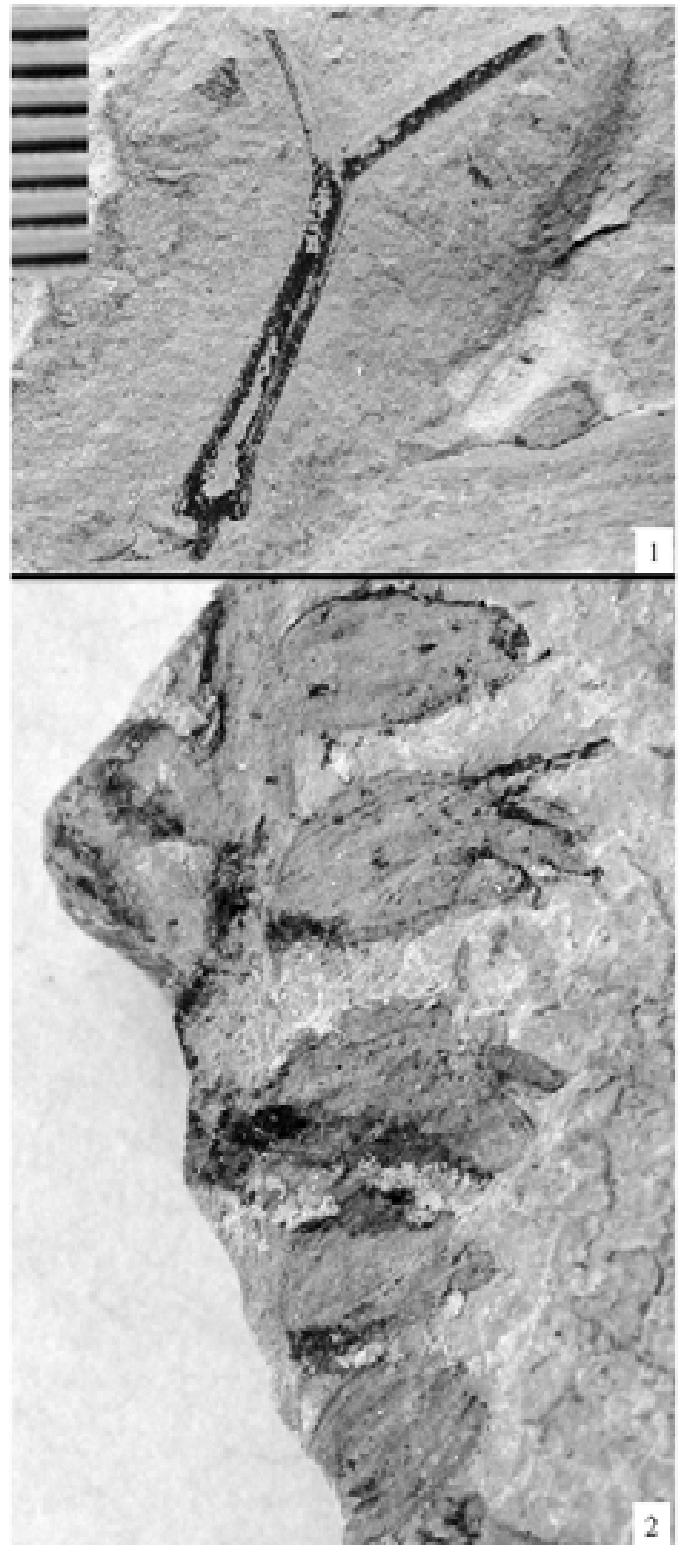


FIGURE 26. Coniferophyte reproductive structures. **26.1**, *Gomphostrobus* sp., USNM specimen 455003 (Mamay and Mapes, 1992, Fig. 8E). **26.2**, *Cordaianthus* sp., USNM specimen 450799 (Mamay and Mapes, 1992, Fig. 7C). Both same magnification. Scale bar in **23.1** in mm.

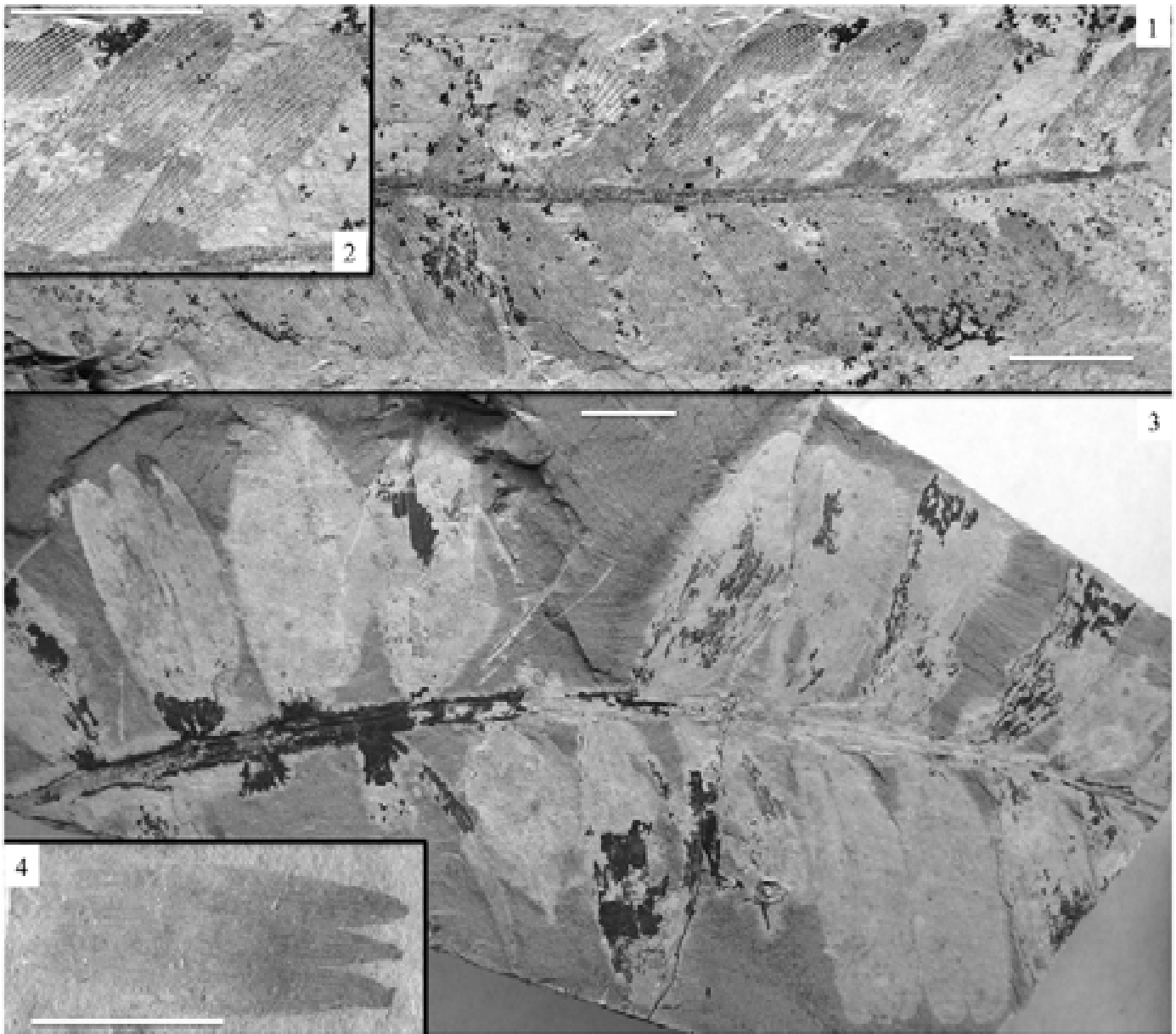


FIGURE 27. Noeggerathialean foliage. **27.1-27.2**, *Plagiozamites rochei*, USNM specimen 422621 (Mamay and Mapes, 1992, Fig. 9A). **27.3**, *Charliea manzanitana*, Holotype, USNM specimen 422628 (Mamay and Mapes, 1992, Fig. 9B). **27.4**, *Charliea manzanitana*, USNM specimen 422624. Scale bars = 1 cm.

abnormis Gutbier by Wagner and Álvarez-Vázquez (2010a), range from Stephanian B upward into the Autunian.

Incertae Sedis

There are several specimens of interest from Kinney Quarry that cannot be assigned to a species or genus with confidence.

The specimen illustrated in Figure 29 was described by Mamay (1990, Fig. 11), and Mamay and Mapes (1992, Fig. 9F) as *Podozamites*-like. The single specimen shows parallel-veined, elongate, entire leaves that clearly are helically arranged on the supporting stem. These leaves show distinct flexure, suggesting that they were borne in an orthotropic posture in life, rather than in a bilateral, plagiotropic orientation. In the features of its leaves, the specimen is similar to *Yuania* (*Russellites*) as described by Mamay (1968); however the leaves in *Yuania* (*Russellites*), as Mamay (1992, p. 15) notes, are “arranged in flat planes”, unlike this specimen, which is why he differentiated it. Although the leaf shape and

venation are similar to plant remains ascribed to the Noeggerthiales, *Podozamites* is a characteristically Mesozoic genus of conifers. The fragmentary specimen figured as *Podozamites* sp. by DiMichele et al. (2001, Fig. 3.3) from Texas also could well be a conifer, as claimed (op. cit.), but is more poorly preserved than the material from Kinney Quarry and might be a different taxon.

Figure 30.1 was illustrated by Mamay and Mapes (1992, Fig. 5H) as cf. *Neuropteris obliqua*. This identification was proposed tentatively and is certainly not correct. We suggest that it may be a large specimen of *Pseudomariopteris cordato-ovata*.

Figure 30.2 may be a small representative of *Neurodopteris auriculata*. The veins appear too arched to be a small example of *Odopteris*, and larger pinnules show a degree of midrib development. We note that the leftmost pinnule appears to have a rounded base.

Figure 30.3 is possibly attributable to *Odopteris gimmii*, as discussed above.

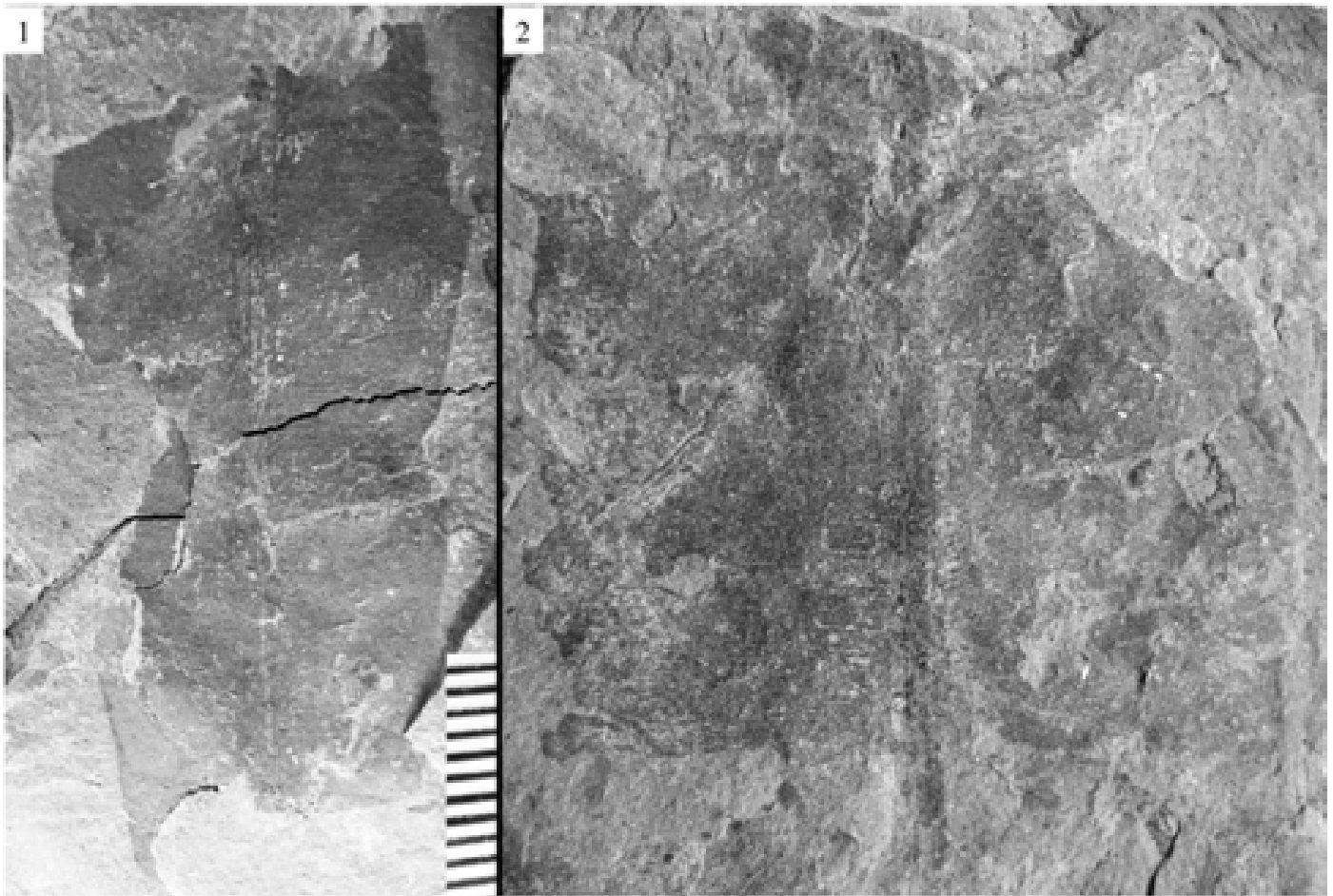


FIGURE 28. Cycadopsid? foliage, *Taeniopteris* cf. *abnormis*. **28.1**, USNM specimen 558146. **28.2**, USNM specimen 558144. Both same magnification. Scale bar in **27.1** in mm.

Figure 30.4 appears to be a small plantlet of indeterminate affinity, and the strands emanating from its base are interpreted as roots.

KINNEY FLORAL ASSEMBLAGE: PRESERVATIONAL ASPECTS

The floral remains from the Kinney Quarry have been collected from shales of brackish to marine facies. Several slabs with plant fragments also show the presence of *Dunbarella* shells, which suggest quiet waters of brackish salinity. The depositional environment has been interpreted as corresponding to an embayment in a tidally influenced estuary (Archer and Clark, 1992). This would imply that the plant fragments came down the estuary and drifted into the embayment, becoming waterlogged and settling on the mud floor in this sheltered environment. However, if the embayment could be seen as a more extensive lagoon, it would allow shallow streams draining the adjacent land area to bring in plant fragments of a certain size. Archer and Clark (1992) also mention bands with finely comminuted plant debris, such as one would expect of plant material subject to tidal action in an estuarine setting once entrained in the sediment. Residence time in water is apparent for many specimens from the evidence of frazzling and other damage to foliar remains due to tissue decay and by the apparent attachment of *Dunbarella* shells to many specimens.

The Kinney floral assemblage was collected from an unknown number of bands in a c. 14 m thick shale interval in the lower part of the Kinney Quarry (Willard, 1992, p. 49-50; Mamay and Mapes, 1992, p. 65; Lucas et al., 2011, fig. 6). The macroflora is wholly allochthonous, presenting a mixture of plants that, we presume from many other pub-

lished analyses of their ecological preferences, are from varied microhabitats on the original landscape. Willard (1992) examined two independently assembled collections, from two portions of the Kinney Quarry and found them to be palynologically distinct. One palynoflora, from the lower portion of the shale (the NMNH collection described in this paper) was a typical seasonally-dry, Late Pennsylvanian assemblage dominated by cordaitalean, conifer and pteridosperm pollen. The second collection was from the upper portions of the plant-bearing shale interval; it was dominated by small fern spores and a problematic palynomorph, the affinities of which are with Gondwana, and had a diversity of palynomorphs from sphenopsids and other lower vascular plants. Such a distinction was not detected in the NMNH Kinney macrofloral collections, and Mamay and Mapes (1992) explicitly state that they recognized no consistent spatial or temporal floristic compositional changes during their collecting activities. Unfortunately, there is no record of the different assemblages that may have been obtained from more than one fossiliferous bedding plane. This means that one can only judge from the total collection, which was brought together by different collectors (Lucas et al., 2011, p. 18). A museum collection may reflect a bias for the better preserved, larger specimens, though this depends strongly on the research objectives of the collector. Even so, the variety of specimens of different kinds shows that different habitats were represented in the area from which the plant remains originated.

The quiet “lagoonal” environment of deposition implies low-lying coastal conditions. The (occasional) presence of relatively large plant fragments capable of being imprinted on the mudrock (in some cases in surprising detail), suggests that their place of growth cannot have been very far away. Even though different habitats are inferred, these were



FIGURE 29. Problematica, possible coniferophyte, *Podozamites*-like shoot, USNM specimen 422630 (Mamay and Mapes, 1992, Fig. 9F). Scale bar = 1 cm.

probably all situated in one and the same general area. Elements possibly brought in from farther away include the conifers, and then only because of their robust construction, not because of some necessary confinement to “uplands”.

The local landscape may have included swamps as well as relatively drier ground, with better drained soils. Since transported plant remains are involved, implying a mixture, there is no direct evidence of different floral associations linked to specific habitats.

FLORAL COMPOSITION

A census of the NMNH collection reveals that there are subtle compositional differences between two facies, separable by color, one gray and one buff, otherwise similar in terms of plant preservation and host-rock characteristics. The facies color difference most likely indicates collection from different beds in the Kinney succession. Willard (1992) noted color differences between the two collections she analyzed palynologically. The most common plant in the gray facies is *Dicranophyllum*, which is much less common in the buff facies. Second most abundant in the gray facies are conifers, followed by neuropterid pteridosperms, and cordaitaleans. The buff facies is dominated by neuropterid pteridosperms and *Sphenopteridium*, with a large element of conifers. Pecopterid ferns, *Mixoneura* and *Sphenopteridium* are found nearly exclusively in the buff facies, and neuropterid foliage is about twice as abundant in the buff facies as in the gray.

Mamay and Mapes (1992) noted the relative frequency of conifer foliage. This is easily understood in terms of durability. Conifers are characterized by leaves with thick cuticles capable of withstanding prolonged immersion in water. Likewise, pteridosperm foliage is more likely to survive prolonged maceration than that of ferns, with generally thinner cuticles. In this respect, it is worthy of note that the relatively common (and comparatively large) pinna fragments of *Pecopteris potoniei* are invariably fertile, the inference being that the *Asterothea synangia* strengthened the pinnules, thus allowing for increased resistance to disintegration. It is also noted that a few of the other fern remains encountered as relatively large fragments occur as “ghosts”, with only the outline of pinnules preserved. This creates the impression of tissue decay in relatively large pinna fragments, which became waterlogged in very quiet, almost still waters. The absence of currents capable of breaking up weakened plant remains will have been an important factor for preservation.

In conjunction with a varied suite of sphenopsids, these fern remains may well have corresponded to wet environmental conditions in the source area, i.e. to a wet substrate floral habitat. This habitat is probably underrepresented as a result of the preservational bias. A genuine hygrophile element is *Sigillaria brardii*, a lycopsid tree that is known to have contributed significantly to Stephanian coal swamps (e.g., Willard et al., 2007; Tabor et al., 2013, and as confirmed by personal observation, RHW/CAV).

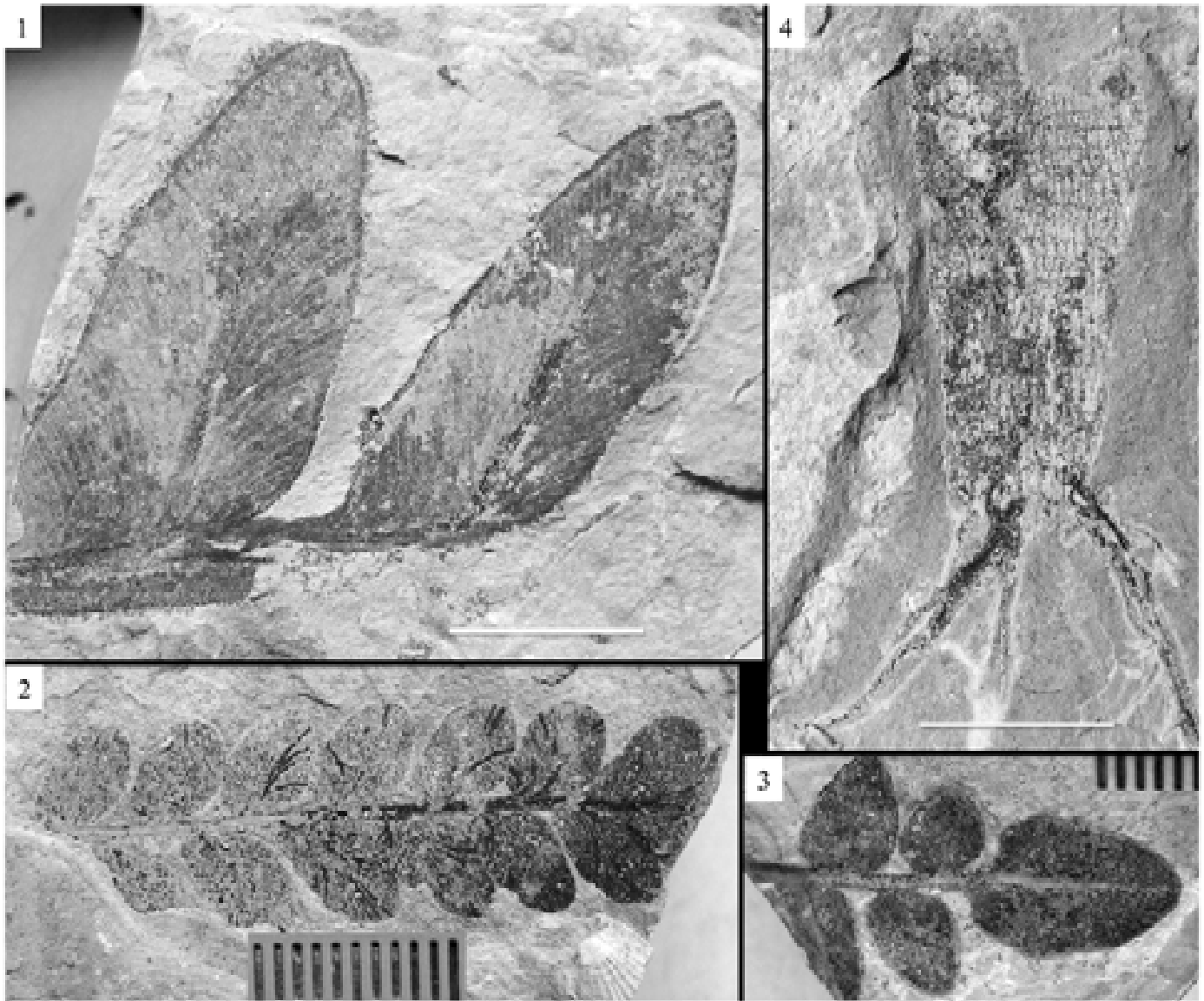


FIGURE 30. Problematica. **30.1**, cf. *Pseudomariopteris cordato-ovata*, USNM specimen 450792 (Mamay and Mapes, 1992, Fig. 9H). **30.2**, cf. *Neurodontopteris auriculata*, USNM specimen 558138. **30.3**, USNM specimen 558140, indeterminate small pinna terminal segment, possibly *Odontopteris gimmi*. **30.4**, USNM specimen 588137, possible small plantlet with basal roots. Scale bars in **30.1** and **30.2** = 1 cm. Scales in **30.3** and **30.4** in mm.

Kerp (1984) associated *Calamostachys dumasii* “*Asterophyllites*” *dumasii* with foliage of the *Annularia carinata*-type and stems referable to *Calamites gigas*. This is consistent with the stem and foliage assemblage found at Kinney Quarry. This association has important paleoecological implications, as *Calamites gigas* has been interpreted as a plant typical of Mediterranean-type seasonal climates (Naugolnykh, 2004), perhaps even with specializations for water storage under xeric conditions (Barthel and Rössler, 1994; Rössler and Noll, 2002).

The quite varied pteridosperm foliage (including *Sphenopteridium*?) could well belong to habitats of various different kinds. These plants are generally classed as mesophiles, not growing directly in swamps but living in close association, as well as including a category that possibly lived on hill slopes.

The species recorded as noeggerathialean (a controversial group of plants regarding their systematic position) belong to a habitat not normally represented in Pennsylvanian floras of the paleoequatorial belt. The noeggerathialeans occur throughout the Pennsylvanian, but are exceedingly rare, since they are apparently linked to higher ground. These are so-called “extrabasinal” elements, a term coined by Havlena (1970) for plants that lived outside the swamp area and directly associated

environments. One of the most significant papers with regard to the presence of noeggerathialean remains is that published by Leary and Pfefferkorn (1977), who recorded floral remains from the sides of a paleovalley of Early Pennsylvanian (mid-Namurian) age in Illinois, USA. More recently Wang et al. (2009) reported on plants of Early Permian age from China found buried in volcanic ash beds above a coal seam, indicating that they were growing, at the time of burial, in a swampy, peat-forming habitat, and thus indicating an expansion of the group, ecologically, in the Permian of western Pangea. There are many other published reports on this group, and those mentioned here may serve as examples and lead into the more extensive literature.

Conifers are regarded as favoring drier habitats, meaning better drained soils. The pattern of conifer occurrence during the Pennsylvanian is complex, however. In the western equatorial regions, these plants are prominent members of plant communities well back into the Middle Pennsylvanian. And conifer pollen is reported in rocks as old as latest Mississippian. Thus, this group was present at tropical latitudes, but rarely preserved in basinal lowlands, long before it rose to prominence in the later Pennsylvanian. Conifers become increasingly common, even in coal basins, during the Missourian across the west-central and western

paleoequatorial region (e.g., Cridland and Morris, 1963; Feldmann et al., 2005; Falcon-Lang et al., 2011). This reflects in part what has been described as a “taphonomic megabias”. The Pennsylvanian was a time of strong glacial-interglacial cyclicity with attendant changes in sea-level and climate. The widespread swamp environments that formed during the wetter phases of these cycles supported floral associations from a habitat inimical to the conifers, and also were the most favorable times for the preservation of plant material. In contrast, during the drier parts of cycles water tables were generally much lower, limiting the environments favorable for the preservation of plants (Gastaldo and Demko, 2011). Only when or where the wetter parts of ice/sea-level/climate cycles became drier, did conifers come to the fore. Whereas it is admitted that the presence of conifer remains does not constitute evidence of a late Pennsylvanian age (prime examples being reports of conifers from Moscovian strata in the Illinois coal basin – Falcon-Lang et al., 2009; Plotnick et al., 2009), the widespread occurrence of a varied conifer assemblage does provide a climatic signal. Taken in its proper context, this may have time-stratigraphic significance.

In summary, the composition of the Kinney floral assemblage shows the effect of preferential preservation of elements capable of withstanding prolonged maceration in water. In addition, it reflects a mixture of habitats, which shows that these plants originated from a varied source area, likely from an environmentally heterogenous landscape.

STRATIGRAPHIC INFERENCES

The taxonomic diversity of the Kinney floral assemblage, as with virtually any paleobotanical assemblage, must be reported with a caveat making note of the fragmentary nature of the plant fossil record and the difficulty of assembling dispersed organs into whole plants. Thus – what to count is always a major concern. At Kinney, there may be as many as 39 identifiable taxa, that is organs that can be placed into a Linnean taxonomic group of some rank. Using a more conservative approach, evaluating potential whole-plant species, this estimate may be closer to 30. In either case, this is an appreciable number for a late Paleozoic flora, but likely far from representing the total diversity of the original flora. A selection of these taxa is listed in Figure 31, together with their known stratigraphic ranges. In view of the probable synonymy between *Sphenopteridium manzanitanum* and *Sphenopteris germanica*, the well-known range of the latter is shown. Whereas these ranges are, on the whole, well established for hygrophile elements and associated mesophiles, this is certainly not the case for a number of species that Havlena (1970) classed as “extrabasinal”. Reference is made to the notable extension of the stratigraphic range of several species of classic Stephanian plants, recorded originally from high Stephanian strata in the non-marine coal-bearing successions of the Massif Central, France, which appear in substantially older sediments in the marine and marine-influenced Stephanian of NW Spain (Bouroz and Wagner, 1972; Iwaniew, 1985).

With reference to the hygrophile elements, it is *Annularia spicata*, above all, that is known to have a restricted range, corresponding to the Autunian Substage (as defined by Wagner and Álvarez-Vázquez, 2010a). *Annularia spinulosa* has a less restricted range, from Saberian upwards, but it is noted that this range may have to be revised when the published record of Stephanian examples of *Annularia stellata* has been scrutinized. *Annularia carinata* has been recorded from Stephanian B to Autunian. *Annularia sphenophylloides*, which is a much more common element, has a longer range, from upper Bolsovian to Stephanian C or even Autunian.

Macroneuropteris scheuchzeri does not range beyond upper Cantabrian in Europe, but has long been recognized as ranging throughout the Upper Pennsylvanian in North America. It may be the same form that was described as *Neuropteris hirsuta* by Fontaine and White (1880) from the lower Dunkard Group of the Appalachian Basin (in Blake et al., 2002; Wagner and Álvarez-Vázquez, 2010b). Perhaps, it should not

range beyond Stephanian C, but last occurrences are traditionally suspect.

Neuropteris ovata is another element of lowland flora, apparently linked to an alluvial plain environment. Originally regarded as characteristic of Westphalian D (Asturian Substage), it has long been recognized as occurring in Stephanian (Upper Pennsylvanian) strata of North America. In the more complete Stephanian record of NW Spain its highest occurrence is in the upper Saberian. However, a more extended range is apparent from its occurrence in Stephanian C coal-bearing deposits at Puertollano in south-central Spain. The form encountered in the Kinney Quarry seems to coincide with that recorded as var. *grandeuryi* Wagner from NW Spain. It may also be compared with *Neuropteris pseudovata* Gothan and Sze as recorded from the Stephanian C of North Portugal.

Neuropteris cordata, as identified in the present paper, refers specifically to the plant figured and described by Barthel (1976) from the Lower Rotliegend (Autunian) of Saxony, Germany. The use of this name in the literature is quite varied, making it pointless to attempt establishing a stratigraphic range.

Mixoneura subcrenulata and *Mixoneura gimmii* are two mesophile elements that are not normally part of the flora associated with swamp environments. Their usual occurrence is in the Autunian (and high Stephanian in general). However, the presence of *Mixoneura subcrenulata* has been discovered in lower Cantabrian (i.e. basal Stephanian) strata in a valley fill (Iwaniew, 1985), these being steep-sided valleys corresponding to an important topographic relief. It is more than likely that these species of *Mixoneura* are floral elements that lived on higher ground, not normally recorded in the context of alluvial plain facies. This implies that their total range is unknown.

More than likely, the same considerations apply to *Neurodopteris auriculata* and *Blanziopteris praedentata*. These are all Stephanian elements with rather infrequent occurrences depending on facies conditions. It is noted that *Neurocallipteris planchardii*, generally found in the higher Stephanian of Central France, has been reported from the upper Cantabrian (lower Stephanian) in the context of an allochthonous assemblage in marine strata (Wagner and Winkler Prins, 1970).

Alethopteris schneideri has only been recorded from the Autunian. Its habitat is unknown.

Another Autunian element is *Sphenopteris germanica*, the presumed equivalent of *Sphenopteridium manzanitanum*, but the habitat of this species is unknown. It might be an “extrabasinal” element. This is made particularly likely by its “Lazarus” status, its earlier occurrences being from late Mississippian-aged assemblages in habitats of seasonal drought.

Sphenopteris hadrophylla has been described from the Saberian and Stephanian B in NW Spain, and it has also been recorded (albeit tentatively) from the Narragansett Basin in eastern North America. Its possible presence in the Kinney Quarry needs to be confirmed by the discovery of more complete, better preserved specimens.

An important element among the peccopteroid ferns is *Danaeites emersonii*. Originally described from the Upper Pennsylvanian of the Appalachian region in North America (see Blake et al., 2002), it also has been recorded also from NW Spain where Wagner and Álvarez-Vázquez (2010a) state its presence from upper Saberian to Autunian.

Nemejcopteris feminaeformis is one of the most characteristic (and most easily recognized) zygopterid ferns in Stephanian strata, with a range from Barruelian to Autunian (earlier occurrences, as recorded in the literature, e.g. Zodrow, 1986, need scrutiny).

Remia pinnatifida is a well known Lower Rotliegend (Autunian) plant from Germany, where it seems to occur fairly commonly. Earlier occurrences are recorded from NW Spain (Knight, 1985; Castro, 2005) where it appears already in the Saberian.

Pecopteris potonie is only rarely recorded in the literature. Its type area is in the Lower Rotliegend (Autunian) of Thuringia, Germany, where it was originally misidentified. Nemejc (1940) also noted its pres-

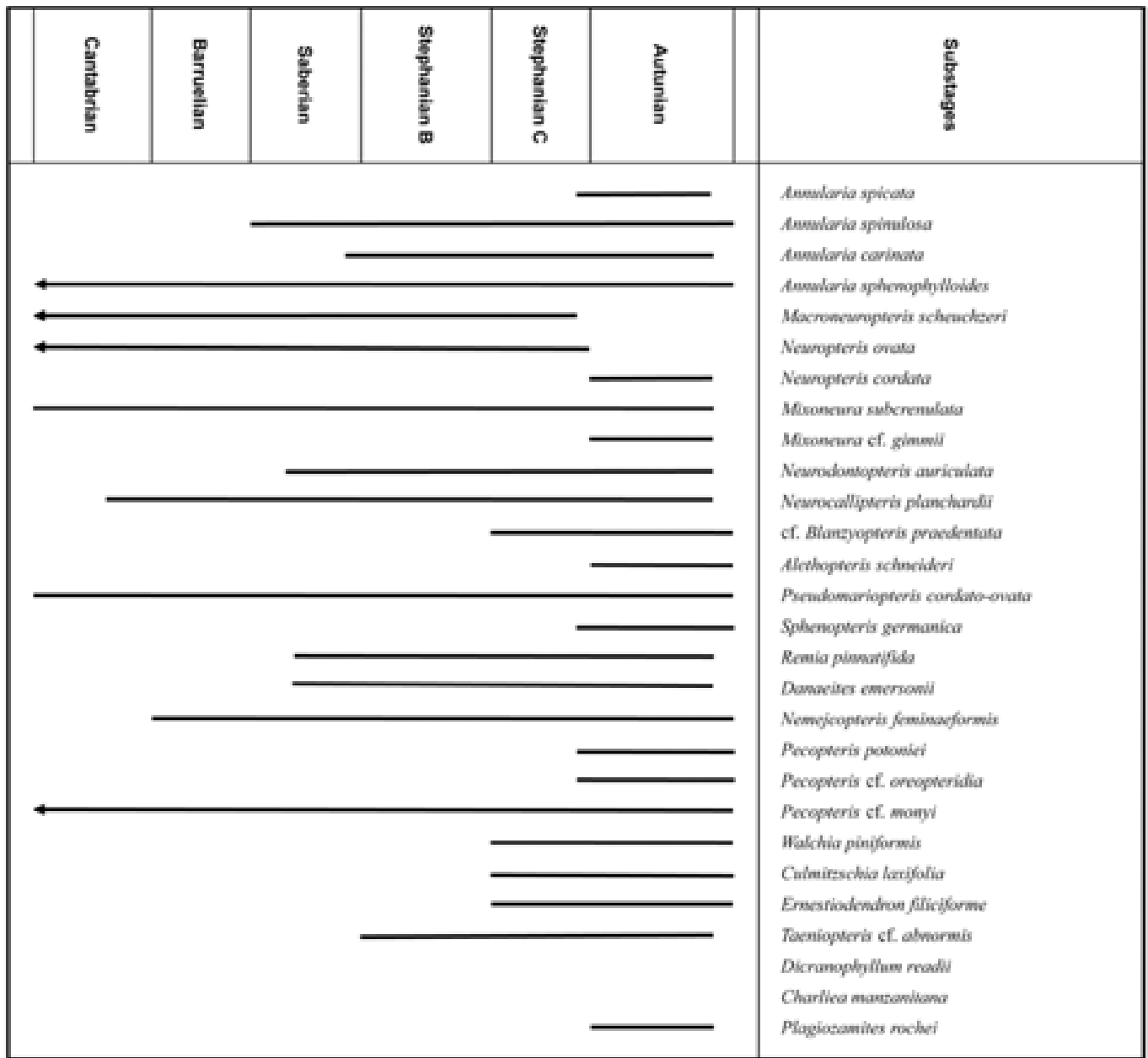


FIGURE 31. Known stratigraphic ranges of specifically identified taxa from the Kinney Quarry flora, set against the substages of the Stephanian Stage (practically equivalent to Upper Pennsylvanian Series in international usage: Kasimovian and Gzhelian stages). *Sphenopteris germanica* is shown as the probable equivalent of *Sphenopteridium manzanitanum*.

ence in the Permian of the Czech Republic (N.B. “Permian” should be interpreted in this case as Auturian, which is currently regarded as highest Carboniferous).

Pecopteris oreopteridia is characteristic of Lower Rotliegend (Auturian) strata. However, there is no guarantee that this species does not appear earlier. Records under this name should be subjected to a critical revision.

Pecopteris monyi is an oft-quoted species (but not always identified correctly), from the Stephanian of Europe. It is recorded from Asturian upwards into Stephanian C, with a total range that probably reflects the wide sense in which this species has been recorded in the literature. A critical revision of these records will be necessary.

Taeniopteris is another case in point. Excluding *Taeniopteris jejuna* Grand'Eury (not part of the Kinney assemblage), which is unique, the densely veined *Taeniopteris abnormis/fallax/multinervia*, of uncertain affinity, is clearly associated with non-wetland (especially peat-

forming) environments. This large-leaved *Taeniopteris* occurs most commonly in Auturian strata (and in the Permian), but it has been recorded as far down as Stephanian B (Wagner and Álvarez-Vázquez, 2010a). *Taeniopteris* fragments have been found in Bolsovian marine strata in NW Spain (van Loon, 1971), providing a glimpse of “extrabasinal” plants living beyond the alluvial plain.

The enigmatic gymnosperm, *Dicranophyllum*, may be another “extrabasinal” element. It ranges throughout the Pennsylvanian and into the Permian, including in the New Mexico Abo Formation.

Disputed noeggerathalean remains, such as *Plagiozamites rochei* and *Charliea manzanitana*, are “extrabasinal” elements, so scarce as to be of little stratigraphic significance (even though the published records are from high in the Stephanian).

The conifers are a special case, referred to earlier. Conifers are recorded palynologically as far down as Moscovian in the Paradox Basin of Colorado, USA (Rueger, 1996), and in coal basins of the American

midcontinent, and may well have occurred even earlier. Hygrophyte and mesophyte plant assemblages from the alluvial plain are unlikely to contain the remains of plants living on relatively high ground. In the context of Pennsylvanian floras, which, on the whole, have been recorded from extensive alluvial plains with widespread swamp environments, the conifers are likely to be so infrequent as to be virtually unrepresented during these wettest parts of glacial-interglacial cycles. It is no surprise that the conifer record becomes more abundant with the appearance of drier, occasionally semi-arid climate in parts of the Euramerican Realm of Permian times (but including the Lower Rotliegend-Autunian of the latest Carboniferous). The contrast between the drier facies of the Euramerican Realm and the wet facies of the Cathaysian Realm (including the Middle East) has been pointed out repeatedly.

It is still a matter of debate whether the diversification of conifer taxa, as recorded from the Autunian, has time-stratigraphic significance. Perhaps, it may be significant that the earliest known occurrence of *Ernestiodendron fliciforme* is from the Stephanian C of Valongo in the Douro Basin, North Portugal (Florin, 1940). *Walchia piniformis* is known from the Stephanian C upwards, while *Culmitzschia laxifolia* is also of Stephanian C and Autunian age.

The sum total of specifically identified taxa and their known ranges are set out in Fig. 31. Taken altogether, there is no doubt that this is a Lower Rotliegend type of assemblage as known from the Autunian of Europe. However, there is a strong environmental connotation. The classic paper by Gothan and Gimm (1930) has distinguished two different floral associations on the basis of habitat and climate. The implications for stratigraphic dating are now generally recognized, as is the need to stick to the same overall environmental conditions for a floral zonation. This automatically removes “extrabasinal” elements from biostratigraphic considerations. In the case of the mixed assemblage at Kinney, this affects a fairly considerable portion of the taxa identified. Even so, this flora has a late Stephanian aspect, either Stephanian C or Autunian. This is borne out by the presence of two pectopteroids first described from the Lower Rotliegend, and also by the presence of *Annularia spicata*. The latter is a distinctive species, unknown from strata below the Autunian.

One must therefore conclude, based on plant distribution patterns, that the flora is Autunian, with the possibility of Stephanian C. Reasons have been given by Wagner and Álvarez-Vázquez (2010a, p.

314) to translate this into the higher Gzhelian. It is noted that a fusulinacean assemblage found immediately below the Kinney plant-bearing strata has been dated as early to middle Missourian (Lucas et al., 2011), which is equated to part of the Kasimovian. This apparent contradiction between the marine faunal data and the terrestrial flora in immediately overlying deposits needs further investigation.

CONCLUSIONS

The Kinney floral assemblage, obtained from shallow brackish to marine (lagoonal?) strata, shows transport selection due to the different degrees of resistance to disintegration following upon decay with residence time in water. This has favored the preservation of conifers, pteridosperms, and (putative) noeggerathiales. Fern foliage may have suffered most of all and is likely underrepresented.

The plant remains record the presence of different habitats in the source area, and this is clearly apparent from the fairly large variety of plant remains collected at Kinney. Mesophyte and drier substrate, so-called “extrabasinal” plants are much in evidence. The transported nature of the occurrence limits the extent of any climatic inferences that can be drawn from this assemblage.

The floral assemblage is most similar to Autunian, possibly Stephanian C floras, known principally from Europe, but also widespread in North America. Indeed, the Kinney floral assemblage is almost entirely identical to those of the Lower Rotliegend (Autunian) in Europe.

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The man, the myth, the legend---of "puff baby," aka William DiMichele, seen here pushing the scientific envelope in 2009 in the Cerros de Amado of Socorro County, New Mexico.

A LARGE HETEROPOLAR COPROLITE FROM THE EARLY PERMIAN OF NEW MEXICO

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Coprofaunas with an abundance of spiral forms characterize non-marine redbeds of Early Permian age in northern (Hunt et al., 2005b) and central New Mexico (Cantrell et al., 2012), West Texas (Neumayer, 1904; Hunt and Lucas, 2005a-b, Hunt et al., 2005a, 2012b) and Oklahoma (Williams, 1972; McAllister, 1985; Hunt et al., 2012a). However, these coprolites are rarely longer than 80 mm, and the majority are shorter than 50 mm. The notable exception is the occurrence of large coprolites (> 100 mm in length) in the Vale Formation of Texas (Olson and Mead, 1982; Hunt and Lucas, 2005b; Hunt et al., 2005a). Here we describe a similarly large spiral coprolite from the Early Permian of northern New Mexico and discuss its relationships. NMMNH refers to New Mexico Museum of Natural History in Albuquerque, New Mexico. TMM refers to the Texas Memorial Museum collection at the University of Texas at Austin.

The Chama Basin in northern New Mexico yields significant Late Paleozoic vertebrate fossils, including coprolites, from several areas including in the vicinity of the village of Arroyo del Agua, Rio Arriba County. These are among the longest known and most extensively documented such fossils from North America (e.g., Romer, 1960; Berman, 1993; Lucas et al., 2005). One of the most significant Arroyo del Agua localities is the VanderHoof quarry (NMMNH localities 4699, 4700, 4706), which has yielded fossils of plants, fish, amphibians and reptiles (Langston, 1953; Romer, 1960; Berman, 1993; Eberth and Berman, 1993; Lucas et al., 2005). The VanderHoof quarry has also produced numerous coprolites, including *Heteropolacopros texaniensis*, *Hyronocopros amhipola* and amorphous coprolites (Langston, 1953; Eberth and Berman, 1993; Hunt et al., 2005b; Lucas et al., 2005).

The canyon of the Rio Puerco and its tributaries around Arroyo del Agua expose approximately 90 m of the upper part of the El Cobre Canyon Formation of the Pennsylvanian-Permian Cutler Group (Lucas and Krainer, 2005). They consist of brown arkosic sandstone in multistoried beds with relatively thin brown siltstone interbeds. Extraformational conglomerates characterize the El Cobre Canyon Formation, and they are composed of clasts of Proterozoic granite, quartzite and other metamorphic rock fragments (Lucas and Krainer, 2005).

Lucas (2002, 2005) proposed a formal tetrapod biochronology of the Permian, and the Lower Permian assemblage of the Arroyo del Agua region is part of the basis of this biochronological framework. The oldest interval of Permian time based on tetrapods is the Coyotean LVF (land-vertebrate faunachron), named for Coyote, New Mexico, near the VanderHoof quarry.

Many complete coprolites have been recovered from the VanderHoof quarry (Hunt et al., 2005b). In addition, the sample from this location includes numerous fragments of coprolites. Recently, several pieces have been re-assembled to reveal a large heteropolar coprolite. NMMNH P-34779 from NMMNH locality 4706 is missing the anterior end and is laterally flattened (Fig. 1). The anterior tip is missing. The coprolite has a preserved length of 106.5 mm and a maximum width of 46.5 mm and a maximum depth of 13.9 mm. The surface is abraded,

apparently as the result of Recent weathering. The coprolite is microspiral in morphology, with two coils in the short posterior spire (*sensu* Hunt and Lucas, 2012). In lateral view the coprolite resembles an acute-angle triangle.

Several large coprolites occur near the base of the Middle Leonardian (Redtankian LVF of Lucas, 2005) Vale Formation of Texas, the majority from the Sid McAdams locality, Taylor County (Olson and Mead, 1982; Hunt and Lucas, 2005b; Hunt et al., 2005a). One specimen (TMM 40339-12) is from the West Table Mountain locality (Hunt and Lucas, 2005b). Some of these complete and fragmentary coprolites are over 12 cm long. A minority of the large coprolites are not spiral, and two of these have been assigned to *Strophocopros valensis* (Hunt et al., 2005b). The majority of the large coprolites are microspiral heteropolar in form. They were initially identified as *Heteropolacopros texaniensis* (Hunt and Lucas, 2005b), but subsequently described as the new ichnotaxon *Megaheteropolacopros sidmcadamsi* (Hunt et al., 2005a). The incomplete holotype coprolite is 129 mm long and 54 mm in maximum width and has a subrounded cross section. The dimensions of this ichnotaxon are notable (more than double the length of *Heteropolacopros texaniensis*), but size alone is a problematic criterion for distinguishing an ichnotaxon. The other proposed diagnostic features of *Megaheteropolacopros sidmcadamsi* are also present in some specimens of *Heteropolacopros texaniensis* (maximum diameter at proximal end of posterior spire, spoon-shaped posterior spire). Thus, we consider these specimens to represent "*Megaheteropolacopros sidmcadamsi*" pending a needed revision of *Heteropolacopros texaniensis* and related ichnotaxa.

NMMNH P-34779 is flattened, abraded and incomplete but it is clearly similar to the holotype of "*Megaheteropolacopros sidmcadamsi*" in being heteropolar and microspiral and in having a length twice that of Early Permian specimens of *Heteropolacopros texaniensis*. Thus, we tentatively assign the specimen from the VanderHoof quarry to "*Megaheteropolacopros sidmcadamsi*." This is only the second occurrence of this ichnotaxon, with the other being the Sid McAdams locality, Taylor County, Texas. The range of "*Megaheteropolacopros sidmcadamsi*" now extends from the Wolfcampian (Coyotean LVF) of New Mexico to the middle Leonardian (Redtankian LVF) of Texas.

Olson and Mead (1982, p. 26) initially suggested that the largest (> 10 cm long) Texas coprolites "almost certainly pertain to *Dimetrodon*, the only known member of the fauna large enough to have produced them." However, Hunt and Lucas (2005b) noted that spiral coprolites can only be produced by an underived fish with a spiral valve, such as *Xenacanthus* sp., that is present at the VanderHoof Quarry, and that shark specimens from the Early Permian of Texas in the TMM collection include specimens (chondrocrania, Meckelian cartilages) of individuals with skulls that would have been more than 20 cm long and thus large enough to have produced "*Megaheteropolacopros sidmcadamsi*." Thus, the presence of "*Megaheteropolacopros sidmcadamsi*" at Arroyo del Agua suggests that large sharks were present in the Coyotean of New Mexico.

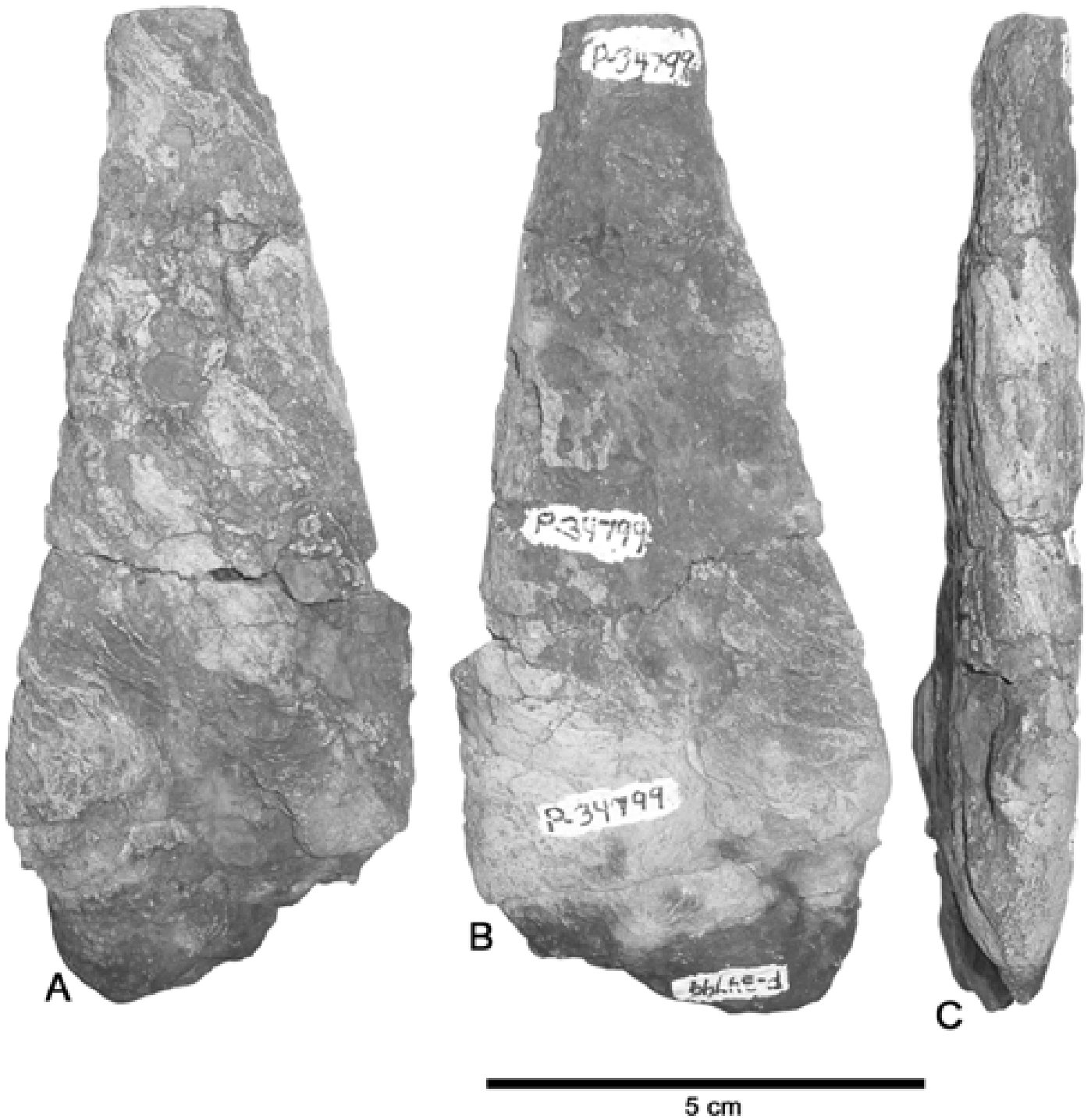


FIGURE 1. “*Megaheteropolacopros sidmcadamsi*,” NMMNH P-34779 from the VanderHoof quarry (NMMNH locality 4706), El Cobre Canyon Formation (Wolfcampian), Arroyo del Agua, Rio Arriba County, New Mexico. A-C, Coprolite in lateral views. Note spiral coils at posterior pole.

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In the headwaters of the Arroyo de la Presilla east of Socorro, the lower part of the Middle Pennsylvanian Gray Mesa Formation is the Elephant Butte Member. This unit is mostly ledge-forming limestone beds intercalated with slope-forming shale. Noteworthy, in the middle of the photograph, is a bed of trough-crossbedded quartzose sandstone, which sits at a sequence boundary (unconformity) within the section.

LATE PENNSYLVANIAN SEDIMENTATION IN THE WESTERN OROGRANDE BASIN (NEW MEXICO): “ICEHOUSE” SEDIMENTATION AND RAPID BASIN SUBSIDENCE IN THE SOUTHERN ANCESTRAL ROCKY MOUNTAINS

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Abstract—The Orogrande basin of southern New Mexico formed as one of the southernmost basins associated with the intracratonic deformation of the Permo-Pennsylvanian Ancestral Rocky Mountains. The mixed carbonate-siliciclastic, Upper Pennsylvanian (Virgilian) strata of the western Orogrande basin (New Mexico) record the time of most rapid subsidence in this basin, a time of high-amplitude glacioeustasy associated with Late Paleozoic icehouse conditions. Here, we focus on the siliciclastic facies and sequence stratigraphic attributes of strata along an updip (northern) to downdip (southern) transect in the poorly studied western Orogrande basin. These strata preserve information bearing on our understanding of (1) basin subsidence and intracratonic deformation related to the enigmatic Ancestral Rocky Mountain system and (2) eustatically driven sedimentation in a mixed carbonate-siliciclastic system, but remain only cursorily studied owing to their location on the restricted White Sands Mississippian Range (WSMR) in southern New Mexico.

Three distinct siliciclastic facies associations record (1) relatively isolated occurrences of fluvial processes throughout the study region, (2) transitional deltaic facies predominating in the central part of the transect, and (3) shallow-marine facies in southern regions. These facies occur within overall mixed carbonate-siliciclastic cyclic successions consisting of normal-marine carbonate strata truncated by surfaces of subaerial exposure in updip regions, and peritidal carbonate strata intercalated primarily with marine siliciclastic facies in downdip regions. These repeated successions of strata deposited during pervasive high-frequency cyclicity reflect extreme shifts in relative sea level associated with Late Paleozoic continental glaciation. Lowstands generally led to peritidal carbonate deposition (early) and fluvial siliciclastic deposition (late) in downdip regions with exposure of subtidal carbonate strata in updip regions, and highstands led to relatively starved deposition in downdip regions and progradational siliciclastic strata (early highstand) to normal subtidal carbonate deposition (late highstand) in updip regions.

The significant thickness (>800 m) and siliciclastic facies of the Virgilian section contrast with underlying, carbonate-dominated Missourian strata and record uplift of basement-cored source regions and very rapid subsidence of the Orogrande basin. Evidence that both the western and eastern margins of the basin were bounded by fault-controlled hingelines in an overall wrench system includes (1) rapid thickening of (primarily) strata in both the western and eastern regions of the basin, (2) paleocurrent shifts through time that record shifts in the regional tilt of the basin, and (3) evidence of contemporaneous seismicity. The sedimentation and subsidence patterns, including evidence for reversal of basinal tilt, and development of marked asymmetry, are most consistent with a basin origin linked to transpressional/transensional wrenching.

INTRODUCTION

The Permo-Pennsylvanian Orogrande basin of southern New Mexico is one of the southernmost basins associated with the tectonically enigmatic Ancestral Rocky Mountains (ARM; Fig. 1). The ARM orogeny remains enigmatic owing to its intraplate setting, but has been related to far-field effects of subduction and collisional suturing along the southeastern plate margin (Kluth and Coney, 1981; Kluth, 1986; Dickinson and Lawton, 2003; Soreghan et al., 2012), or possible subduction along southwestern Laurentia (Ye et al., 1996). The tectonic character and mechanics of subsidence within the Orogrande basin also remain enigmatic. The Ye et al. (1996) model of subduction along the southwestern plate margin depicted the Orogrande basin as a foreland basin. Wilson and Jordan (1988) interpreted the Orogrande basin as a north-trending pull-apart basin with some transcurrent motion along basin-bounding faults. Others have suggested that the Orogrande basin developed in a transpressional tectonic setting resulting from dextral transpressional shear along the southern margin of North America (Singleton, 1990; Beck and Chapin, 1995; Woodward et al., 1999).

Thick Upper Pennsylvanian strata of the Orogrande basin record the timing of peak subsidence and basin development (Soreghan 1994a), as well as global icehouse conditions. These strata are carbonate-rich along the eastern basin margin, and record a well-defined and relatively narrow shelf margin that includes well-known phylloid-algal carbonate buildups intercalated with fluvial and shallow marine siliciclastic strata (e.g., Soreghan and Giles, 1999a; Fig. 2). The Upper Pennsylvanian (Virgilian) strata of the Holder Formation in particular have been studied in detail along the eastern margin of the basin for >60 years (e.g., Plumley and Graves, 1953; Otte, 1959; Pray, 1961; Wray, 1962; Otte and Parks, 1963; Cys and Mazzullo, 1977; Toomey et al., 1977; Wilson, 1977; Mazzullo and Cys, 1979; Goldstein, 1988; Gordon, 1997; Rankey et al., 1999). A detailed study of these strata along the eastern basin margin led Wilson (1967) to develop the now classic concepts of “Cyclic and Reciprocal” sedimentation, a landmark precursor to our current understanding of the high-frequency sequence stratigraphy that characterizes strata of this age in many regions globally. These strata are quintessential examples of “icehouse” carbonates marked by the development of subaerial exposure surfaces on subtidal carbonates (e.g., Goldstein, 1988;

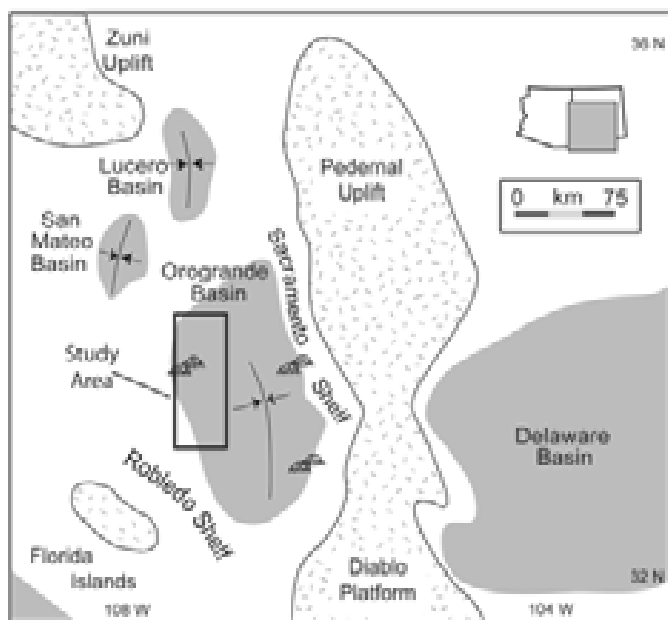


FIGURE 1. Uplifts and basins of the Late Pennsylvanian-Early Permian southwestern Ancestral Rocky Mountains (modified from Algeo et al., 1991; Soreghan and Giles, 1999).

Rankey et al., 1999), and an absence of peritidal carbonate strata, a pattern thought to prevail for icehouse systems in general (Wright, 1992; Read, 1995).

In contrast to the well-studied Permo-Pennsylvanian strata of the eastern Orogrande basin, correlatives of the central/western basin remain little studied owing to their location on the U.S. Army White Sands Missile Range (WSMR; Fig. 3), a region that has remained restricted and largely inaccessible since 1945. These strata of the Panther Seep Formation are remarkably siliciclastic-rich, with carbonate members that commonly exhibit peritidal facies (Kottlowski et al., 1956; Schoderbek, 1994; Soreghan 1994a; Soreghan and Giles, 1999a; Soreghan et al., 2000; Doherty et al., 2002; Seals et al., 2002). The siliciclastic facies here record tectonic uplift of source areas, as well as sediment dispersal patterns and thus basin gradients and changes in gradients. The carbonate strata contrast with those of the eastern basin by the locally common development of peritidal conditions, an unexpected divergence from the usual models of icehouse carbonate sedimentation (Pollitt et al., 2009). This paper focuses on depositional facies, paleocurrents and thickness patterns of siliciclastic facies in the uppermost Pennsylvanian section of the central/western Orogrande basin in order to better understand basin shape and subsidence, and address sequence stratigraphic patterns in a mixed carbonate-siliciclastic section of a shallow distal-ramp setting.

GEOLOGIC SETTING

The Orogrande basin of southern New Mexico developed in Pennsylvanian time in western equatorial Pangaea, as one of the southernmost basins of the late Paleozoic Ancestral Rocky Mountains system (Kluth and Coney, 1981; Kluth, 1986). In Mississippian time, prior to development of the Orogrande basin, this region recorded deposition on a southwardly dipping ramp generally controlled by the trend of the Transcontinental arch that represented the landward area to the north. In contrast, by Pennsylvanian time, uplift of the Pedernal landmass to the east and concomitant subsidence of the Orogrande basin created a well-defined shelf margin now well exposed in the Sacramento Mountains (e.g., Jordan, 1975). During Middle Pennsylvanian-Early Permian time, the basement-cored Pedernal uplift rimmed the basin to the east/northeast and separated the Orogrande basin from the Delaware basin and its Northwest shelf of eastern New Mexico and west Texas (Fig. 1). The Pedernal

		Northern Orogrande basin (Oscura Mountains)	Western Orogrande basin (southern San Andres Mountains)	Eastern Orogrande basin (Sacramento Mountains)
PERMIAN	Wolfcampian	Abo Fm	Abo Fm Hueco Ls	Abo Fm Hueco Fm
	"Bursumian"	Bursum Fm		Laborcita Fm
PENNSYLVANIAN	Virgilian		Panther Seep Fm	Holder Fm
	Missourian	Upper Madera Fm		Beeman Fm
	Desmoinesian	Lower Madera Fm	Lead Camp Limestone	Gobbler Fm

FIGURE 2. Pennsylvanian-Permian stratigraphy of the western (San Andres Mountains) and northern (Oscura Mountains) Orogrande basin; Fm = Formation (modified from Kottlowski et al., 1956; Otté, 1959; Bachman, 1968; Soreghan and Giles, 1999).

uplift continued southward as the Diablo platform. The eastern margin of the Orogrande basin is well exposed along the west face of the Sacramento Mountains, where mixed carbonate-siliciclastic strata of the Upper Pennsylvanian have been well studied (references above). The Pedernal uplift is thought to have provided the siliciclastic sediment source for the Pennsylvanian strata immediately to the west (e.g., Pray, 1959, 1961; Kottlowski, 1968; Dunning, 1978), although less is known regarding the source area for the siliciclastic strata in the western part of the basin.

More than 2,500 m of cyclic strata accumulated in the Orogrande basin during Pennsylvanian-Permian subsidence (Kottlowski, 1963), with >800 m of intercalated carbonate and siliciclastic strata preserved in the Upper Pennsylvanian Panther Seep Formation of the San Andres Mountains (western Orogrande basin; Fig. 2). Within the southwestern Orogrande basin, the Panther Seep Formation contains >50% siliciclastic facies in Ash Canyon (Fig. 3), whereas the proportion drops to ~25% in the northern part of the basin (Hembrillo Canyon, Rhodes Canyon, and the southern Oscura Mountains, Fig. 3). These strata include both continental and marine components, with the marine incursions presumably entering from the south (e.g., Kottlowski et al., 1975).

Within the Orogrande basin, stratigraphic names of the Upper Pennsylvanian section vary, and include the Panther Seep Formation (San Andres Mountains) and Madera and Bursum formations (Oscura Mountains), as depicted in Figure 2. The ages depicted here are based primarily on fusulinid biostratigraphy documented in the works of Kottlowski et al. (1956) and Soreghan (1992), together with new data collected in this study. However, fusulinid-bearing strata are increasingly rare in the upper part of the Hembrillo Canyon (Fig. 3), and in the Ash Canyon section; hence, ages in the southern study area are tentative, and the sections may range into the earliest Permian.

METHODS

The focus of this study is on siliciclastic strata within the mixed carbonate-siliciclastic Panther Seep and Bursum formations and correlative sections exposed at four localities within the Orogrande basin: Ash Canyon, Hembrillo Canyon, Rhodes Canyon, and the southern Oscura Mountains (Fig. 3). Previously published measured sections of the Middle Pennsylvanian to Lower Permian interval in Ash Canyon (Kottlowski et al., 1956; Schoderbek, 1994), Hembrillo Canyon (Soreghan, 1992), and Rhodes Canyon (Soreghan, 1992) were utilized for data on total section thicknesses and carbonate facies interpretations, but siliciclastic inter-



FIGURE 3. Present-day location map of study area in southern New Mexico (modified from Algeo et al., 1991; Soreghan and Giles, 1999).

vals of those sections were remeasured and described in greater detail for this study. The Oscura Mountains section was measured in detail using a Brunton compass and Jacob staff and described at decimeter (bed) scale. Siliciclastic intervals in each section were logged in the field and traced out laterally for at least one km to assess vertical and lateral changes in grain size and sorting, sedimentary structures, and bedding contacts and sandbody geometry. Paleocurrent measurements were taken from cross-stratification, parting lineations, scour marks, ripple cross-laminations, and trough axes; in the case of three-dimensionally exposed trough cross-beds, paleocurrent directions were determined following established procedures (DeCelles et al., 1983). Modal compositions of sandstones from representative siliciclastic facies were determined by visual estimates from thin sections.

Fusulinid biostratigraphy was used for stage-level age assignments for the sections, needed to document thickness variations through time for the Late Pennsylvanian. Fusulinids in thin sections from the southern Oscura Mountains section were identified by Dr. Greg Wahlman. These data combined with previous fusulinid-based age determinations from the study sections (Kottowski et al., 1956; Shell Oil Co., unpublished ca. 1960; Soreghan, 1992) were used for correlations. Missourian and Virgilian isopach maps were constructed by integrating these data with other published measured sections, subsurface and structural data in the Orogrande basin. A tectonic subsidence curve was made with the program DecemPowMac developed by M. Kominz using the measured section from Ash Canyon, which preserves the thickest section in the basin.

SILICICLASTIC FACIES AND PALEODISPERSAL

Siliciclastic strata of the Upper Pennsylvanian-Lower Permian Panther Seep Formation and its correlatives exhibit diverse facies throughout the basin. Figure 4 shows generalized stratigraphic sections of the Virgilian interval throughout the study area. Table 1 lists details of individual siliciclastic lithofacies. Analogous to correlative sections elsewhere in low-latitude Pangea, these strata exhibit a marked cyclicity, expressed

by intercalation of carbonate and siliciclastic strata. The cyclicity is perhaps best expressed in the Rhodes Canyon section (Fig. 4), which records approximately 15 cycles; a similar number occur in the other study sections, although the facies expression of such cycles varies, and cycle identification is subject to interpretation. Examples of vertical facies successions for the three siliciclastic facies associations are illustrated in Figure 5, but note that there are variations in facies not captured in these simplifications. The siliciclastic lithofacies are grouped into facies associations designated A, B, and C. Key carbonate lithofacies documented in previous studies from localities in the San Andres Mountains are summarized in Table 2; Figure 6 includes examples of the peritidal carbonate facies. Paleocurrent measurements are summarized in Table 3. As used below, the southern Oscura Mountains (OM) form the northern part of the study area, whereas the Rhodes and Hembriillo Canyon sections (RC and HM, respectively) form the north-central parts, and Ash Canyon (AC) forms the southern section (Fig. 3).

Facies association A: Upwardly Fining, Channel-form Sandstone

Facies association A (Table 1) consists of 5-10 m thick intervals containing basal conglomerate overlain by an upwardly fining interval of sandstone to siltstone, rarely overlain by red mudstone (Fig. 5). Sandstone units of this facies association are channel-form and pinch out laterally into siltstone and/or mudstone when <100 m. This facies association is most common in the lower part of the northern section (OM), and occurs more rarely in the north-central (RC/HC), and southern parts (AC) of the study area. The basal conglomerate contains sub-angular to well-rounded clasts of chert, lime/dolostone, granite, petrified wood, and carbonaceous plant debris; very coarse-grained quartz, and granitic cobbles occur only in the northern section (OM; Fig. 6C-D).

The basal conglomerate grades upward to upwardly fining packages of trough cross-bedded, dominantly arkosic, coarse- to medium-grained sandstone commonly overlain by climbing ripple cross-laminated and planar laminated fine-grained sandstone and siltstone with parting lineations commonly preserved on upper bedding surfaces. Lateral accretion surfaces occur locally. In the lower study interval, paleocurrents from trough cross-bedded sandstones record flow to the south-southwest, whereas paleocurrents from ripple laminations in these same upwardly fining packages exhibit flow to the northwest (Fig. 4). In contrast, sandstones of this facies association in the upper section exhibit flow to the south-southeast (Fig. 4). Rarely, red, nodular, very thinly bedded, calcareous mudstone intervals locally exhibiting slickensides and downwardly bifurcating (inferred root) traces overlie the siltstone. Basal contacts of this facies association are sharp, with either carbonate (subtidal in north, peritidal in south) or inferred marine mudstone, with the latter commonly overlying this facies association. In thin section, the sandstone facies are arkosic, with a framework composition averaging 40% quartz, 35% feldspar, 15% carbonate lithoclasts (mostly inferred pedogenic with rare marine lithoclasts or bioclasts), 5% metamorphic lithic fragments, and 5% mica.

This facies association (channel-form siliciclastic facies) is interpreted to record deposition in a fluvial environment. The lenticular geometry of the conglomerate and sandstone, and upwardly fining character, are consistent with deposition in fluvial systems (Miall, 1985, 1996). Downwardly bifurcating root traces and randomly oriented slickensides in the red mudstone reflect pedogenesis in floodplains. The apparent 90° shift in paleocurrent direction from southwesterly directed flow in trough cross-bedded sandstones to northwesterly flow in ripple laminations within the same upwardly fining interval is interpreted to represent overbank flow. The up-section shift in flow directions of channel sandstones from dominant southwesterly to southeasterly appears to record a shift in the direction of basin gradients, and possible source regions, explained later in more detail (Fig. 4).

Variations in clast type and size within basal conglomerates likely reflect proximity to source, with the northern regions (e.g., OM) recording more proximal deposition, consistent with the generally south-di-

TABLE 1. Summary table of siliciclastic facies and facies associations.

Lithofacies	Composition	Thickness, Bedding, and Sedimentary Structures	Contacts	Interpretation
<i>Facies Association A</i>				
Lenticular, upwardly fining coarse to medium sandstone	Arkosic sandstone (Q 40%, F 35%, carbonate lithoclasts (15%), and metamorphic lithics (5%), mica (5%). Common basal conglomerate with clasts (up to 15 cm) of chert, quartz, lime/dolostone, granite, petrified wood, and plant debris.	2-5 m thick; medium to thick bedded, poorly to moderately sorted, lenticular geometry; fines upward from basal conglomerate to medium sandstone; trough cross beds	Basal contacts sharp and erosive, common basal conglomerate; upper contacts gradational to overlying fine sandstone to siltstone	Fluvial channel-fill
Upwardly fining sandstone to siltstone	Same as above	1-4 m thick; thinly bedded, poorly to well sorted, sheet-like geometry, fines upward from fine sandstone to siltstone; climbing ripple cross-laminations, planar-laminations with parting lineations on upper bedding surfaces	Basal contacts gradational with underlying medium sandstone; upper contacts gradational where overlain by red mudstone, sharp where overlain by other lithofacies	Fluvial floodplain sandstone and siltstone
Red mudstone	Quartz silt, clay	Averages ~1 m thick; fissile to nodular; slickensides, vertical root traces, calcitic veining; observed only rarely in the Oscura Mountains section	Basal contacts are gradational with underlying fine sandstone to siltstone; upper contacts are sharp	Pedogenically modified fluvial floodplain
Lithofacies	Grain Types	Bedding and Sedimentary Structures	Contacts	Interpretation
<i>Facies Association B</i>				
Upwardly coarsening claystone to fine sandstone	Clay to very fine sandstone, fine to medium grained mica, calcite cemented, plant debris and impressions, rare marine fossils (i.e., pectinids)	20-40 m thick, successions coarsen up, thinly bedded to fissile with thin (<10 cm) interbeds of fine sandstone common at tops; climbing ripple cross-laminations, and planar laminations common	Sharp lower contact, commonly an exposure surface; sharp upper contact with upwardly fining medium to coarse sandstone or open marine limestone, or gradational with upwardly coarsening medium to coarse sandstone lithofacies	Pro-deltaic to mid-deltaic claystone and sandstone
Lenticular upwardly fining sandstone	Arkosic to subarkosic (Q 45%, F 35%, carbonate lithoclasts 15%, metamorphic lithics 5%), fine- to medium-grained, plant debris, pebble-size limestone and mudstone intraclasts	5-10 m thick; medium to thick bedded, poorly to well sorted; common basal conglomerate (limestone and mudstone intraclasts, matrix supported) upwardly fining medium-fine sandstone; channel-form geometry; trough cross bedding grading upward to ripple cross-laminae and planar laminae	Sharp to erosive lower contact with underlying coarsening upward mudstone to fine sandstone lithofacies; sharp to gradational upper contact commonly with biosiliciclastic limestone	Deltaic distributary channel sandstone
Upwardly coarsening sandstone	Same as above	5-10 m thick; medium to thick bedded, poorly sorted to moderately well sorted, laterally persistent; planar laminated grading upward to low-angle trough cross bedding; commonly bioturbated and/or soft-sediment deformed at top	Basal contacts commonly gradational, but rarely sharp, with underlying lithofacies; sharp upper contacts with overlying open-marine limestone	Deltaic bar sandstone

TABLE 1. Continued. Summary table of siliciclastic facies and facies associations.

Lithofacies	Grain Types	Bedding and Sedimentary Structures	Contacts	Interpretation
<i>Facies Association C</i>				
Hummocky cross-stratified sandstone	Subarkosic, calcite cement, finely micaceous, shell fragments, locally conglomeratic with small (< 5 cm diameter) angular limestone intraclasts	1.5-2.5 m thick; medium bedded, locally fines upward from medium sandstone to fine sandstone, moderately to well sorted, hummocky cross-stratified in fine sandstone, rare faint low-angle trough cross-beds	Basal contacts sharp with underlying strata; upper contacts gradational to planar laminated fine sandstone and/or siltstone	Storm dominated lower shoreface sandstone
Soft-sediment deformed medium to fine sandstone	Subarkosic, Q (35%, F 25%, carbonate lithoclasts 35%, metamorphic/volcanic lithics 5%), fine- to medium-grained mica, petrified wood and wood impressions common	4-10 m thick; medium to thickly bedded, poorly to well sorted, fines upward from medium to fine sandstone, and locally from a basal conglomerate to siltstone, trough cross beds and pervasive convolute laminae, flame structures	Basal contacts sharp and erosive; upper contacts gradational when overlain by fine sandstone to siltstone, but sharp when overlain by another lithofacies	Shallow-marine sandstone

rected paleocurrents. Abundant coarse-grained arkosic and micaceous sandstone, and cobble-sized granitic clasts in the north (OM) record drainage from the nearby basement-cored Pedernal uplift, whereas limestone intraclasts within medium- to fine-grained arkosic sandstones to the south record erosion of exposed shelfal and peritidal carbonates during sea level lowstands in positions more removed from basement-cored source areas.

Facies association B: Thick, Upwardly Coarsening Sandstone-Mudstone

Facies association B (Table 1) consists of 20-50 m thick upwardly coarsening successions of (basal) mudstone to fine-grained sandstone (Fig. 5, 6E-F). Facies association B is common in the northern part of the study area (OM, RC, HC), but absent in the south (AC). This association is micaceous with abundant carbonaceous plant debris and impressions (Fig. 6D), and local marine bivalve impressions (e.g., pectinids). This facies is overlain rarely by sharp-based and lenticular upwardly fining sandstone (more common in the lower part of the section), or grades abruptly to laterally persistent, upwardly coarsening sandstone (throughout the section). The lenticular, upwardly fining sandstone ranges up to 5-8 m thick and includes mudstone clasts at the base of trough cross-bedded medium-grained sandstone overlain by climbing ripple cross-laminated and planar-laminated fine-grained sandstone and siltstone. In contrast, laterally persistent upwardly coarsening sandstone units range up to 10 m thick and consist of medium bedded and planar-laminated medium-grained sandstone (base) to thickly bedded trough cross-stratified coarse-grained sandstone (top). Sandstones of facies association B are arkosic, with a framework mineralogy consisting on average of 45% quartz, 35% feldspar, 15% carbonate lithoclasts and allochems (mollusk fragments and rare ooids), and 5% metamorphic lithic fragments.

Basal contacts of facies association B typically are sharp above either a thin, dark, oncoidal wackestone (Soreghan, 1992) or above sub-aerial exposure surfaces (e.g., calcrete) developed on subtidal carbonate of the underlying cycle. Upper contacts grade over <20 cm into biosiliciclastic packstone to grainstone. Paleocurrents from this facies association low in the section exhibit southwesterly flow directions, while the upper section exhibits a southeasterly flow direction (Fig. 4).

This facies association is interpreted to represent deposition in a range of small (e.g., bay-fill) deltaic environments. The presence of both plant debris and marine biota in upwardly coarsening mudstone to silt-

stone is consistent with deposition in pro-deltaic to upper distal deltaic subenvironments, e.g. deltaic distributary channels, and bar sands. Analogous to the fluvial facies association, a shift from dominantly southwesterly flow directions in the lower section to southeasterly flow directions in the upper section indicates an apparent shift in the direction of basin gradients, and possible source regions from the northeast during early Virgilian time to northwest in late Virgilian time (Fig. 4).

Facies association C: Laterally Continuous, Fine- to Medium-Grained Sandstone

Occurrences of facies association C (Table 1) range in thickness from 1.5-15 m, and consist of well-sorted, medium- and fine-grained sandstone. Sandstones of this facies association are laterally persistent and traceable laterally for at least one kilometer. This facies association is more common in the southern part of the study area (AC), but also occurs in central regions (HC). Locally, these sandstones grade upward from basal medium-grained sandstone and are overlain by planar-laminated fine-grained sandstone to siltstone. Parting lineations occur locally on upper bedding surfaces and, more commonly, low-angle trough cross-bedding occurs. Rarely, in the upper parts of the sections in the southern (AC) and south-central (HC) regions, hummocky to swaley cross-stratified, fine-grained sandstone occurs. Both upper and lower bedding contacts are sharp, and sandstones are commonly overlain by peritidal carbonate, and locally marine silty carbonate mudstone, and underlain by marine silty carbonate mudstone or siltstone (our data, plus Shell Oil Co., unpublished, ca. 1960; Schoderbek, 1991; Fig. 6G-H). In the lower part of the study interval, paleocurrents from trough cross-bedded sandstone units indicate a southwesterly flow direction, whereas, in the upper part of the section, paleocurrents from trough cross-bedded sandstones and parting lineations on the upper bedding surfaces exhibit flow to the southeast. Sandstone framework composition averages 35% quartz, 25% feldspar, 35% carbonate lithoclasts and allochems (mud/wackestone lithoclasts, bioclasts including algae, brachiopods, bivalves, and ooids and peloids), and 5% metamorphic and volcanic lithic fragments.

The presence of hummocky to swaley cross-stratification, lateral persistence of sandstone, and abundance of marine bioclasts indicate deposition in a shallow marine environment. Hummocky to swaley cross-stratification records sedimentation within the shoreface. Medium- to fine-grained sandstone units within Ash Canyon were interpreted by Schoderbek (1991) to represent deposition in a fluvial system. Although

TABLE 2. Summary table of carbonate facies from the San Andres Mountains.

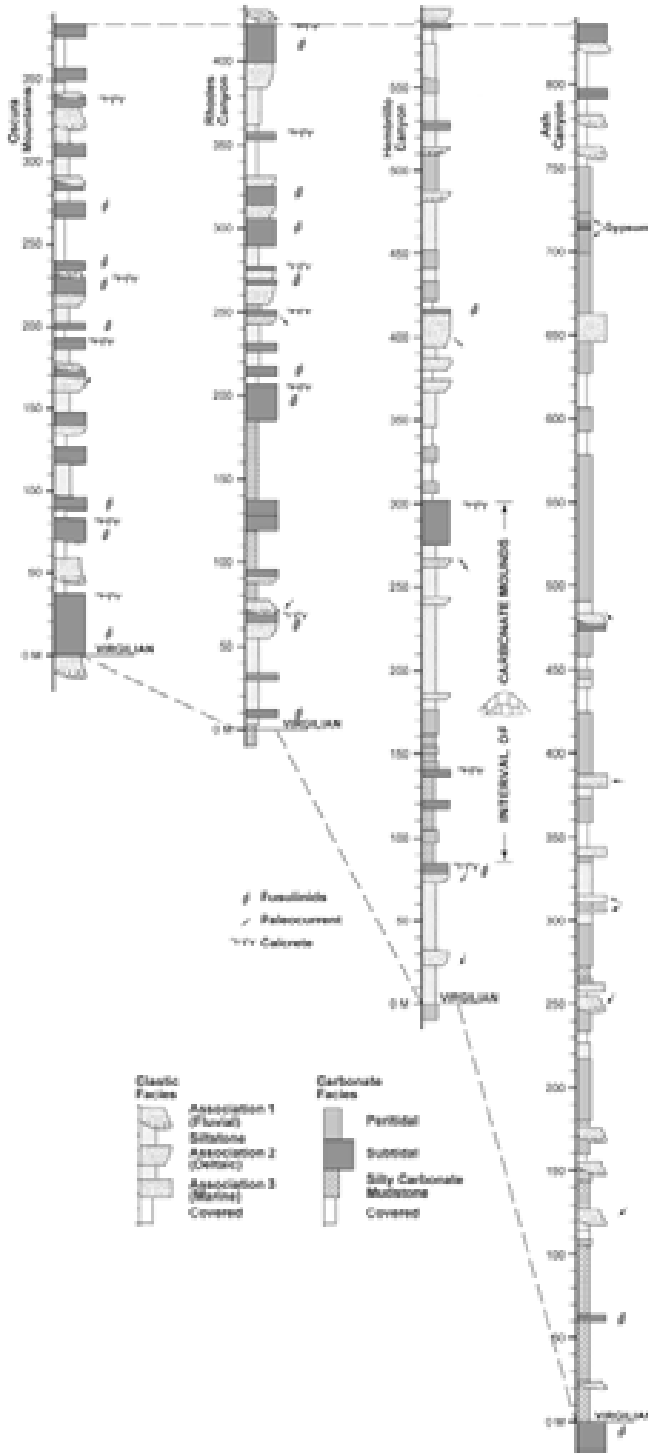


FIGURE 4. Generalized stratigraphic sections of Virgilian-aged strata in the western Orogrande basin (San Andres Mountains and Oscura Mountains); arrows indicate mean paleocurrent directions. Based on data from this study (details in Seals, 2002), supplemented with data from Kottlowski et al. (1956), Shell Oil Co. unpublished data (ca. 1960), Schoderbek (1991), and Soreghan (1992). The top datum is the approximate Abo interval. There are approximately 15 cycles within the Virgilian in each section, although there is not one pattern that characterizes a “cycle” everywhere; hence such determinations are subject to interpretation.

Facies	Composition	Interpretation	Geographic occurrence	Reference
Laminated Peloidal Carbonate	Biohercycliclastic wackestone, peloidal carbonate, laminated restricted subtidal to micrite and calcisiltite, with crystallog and algal laminae and stromatolites, desiccation sites and supratidal cracks	Very shallow, periodically or permanently exposed	Rhodes, Hembello, and Ash canyons	Soreghan, 1992; Schoderbek, 1994
Biohercycliclastic-Neobiohercycliclastic Grainstone	Skeletal and nonskeletal grains, grains of crinoids, forams, peloids, and mixed biohercycliclastic	Slightly restricted to open marine conditions	Rhodes and Hembello canyons, southern Oscura Mountains	Soreghan, 1992
Biohercycliclastic Packstone	40-60% biohercycliclastic debris	Shallow-marine environment with significant current agitation	Rhodes and Hembello canyons, Oscura Mountains	Soreghan, 1992
Fossiliferous Biohercycliclastic Wackestone	Micritic limestone containing a variety of microfossils	Marine environments at depth below significant current action	Rhodes and Hembello canyons, Oscura Mountains	Soreghan, 1992
Dark Oxidized Wackestone	Dark gray to black, weathers orange to rusty, thin (2-20 cm), 20-40% biohercycliclastic debris; distinctive diagenetic features	Carbon influenced, oxygen-poor marine environment	Hembello Canyon, early in Hembello Canyon	Soreghan, 1992
Algal Limestone	Fine, subparallel algal laminae, desiccation cracks	Deposition in arid peritidal environments, supratidal to intertidal	Ash Canyon	Schoderbek, 1994
Limestone Breccia	Massive, porous to dense, irregularly filled disconformity “paper”	Subaerial exposure of previous limestone within supratidal settings	Rarely in Ash Canyon	Schoderbek, 1994
Mud-dominated Limestone	Massive to finely laminated, micritic, rare allochems	Relatively “deeper” water	Ash Canyon	Schoderbek, 1994

we infer the presence of isolated fluvial units in this section, we favor a shallow-marine interpretation for the majority of the sandstone units in Ash Canyon, owing to their lateral continuity, abundant (locally up to 50%) marine allochems and bioclasts, and lack of continental attributes such as overbank deposits, and paleosols. Because this facies association records deposition in a shallow-marine setting, paleocurrents from sandstone units may reflect the influence of dominantly southwesterly directed marine currents and not necessarily indicate paleoslope direction.

Soft-Sediment Deformation in Siliciclastic Facies

Pervasive soft-sediment deformation (convolute bedding, deformed cross-bedding, and flame structures) occurs commonly in various facies within the southern-most part of the field area (AC), locally in the central-northern regions (HC and RC), and rarely in the northern-most part of the study area (OM). The soft-sediment deformation features common within the AC section occur on scales ranging from a few centimeters within a single bed to larger, meter-size deformation structures that locally affect the entire bed (Fig. 7). Convolute bedding structures are bound above and below by undisturbed bedding.

Owing to their presence in various facies, and random occurrence unrelated to any particular type of primary sedimentary structure, we propose these features can be interpreted as seismites (cf. Santos et al., 2012), as discussed in more detail in a later section.

SEQUENCE STRATIGRAPHY

Upper Pennsylvanian strata of the western Orogrande basin are highly cyclic, analogous to late Paleozoic “icehouse” sequences and cyclothems described from the mid-continent and elsewhere (e.g. Heckel, 1986; Veevers and Powell, 1987). Previous researchers have studied the sequence stratigraphic framework of the Upper Pennsylvanian for both the eastern Orogrande basin (Sacramento Shelf; Wilson, 1967; Van Wagoner, 1977; Raatz, 1996; Raatz and Simo, 1998; Rankey et al., 1999) and selected parts of the western Orogrande basin (Schoderbek, 1991, 1994; Soreghan, 1992, 1994a; Soreghan and Giles, 1999a).

In contrast to these previous studies, we focus here on the sequence stratigraphic architecture of an oblique transect spanning “high” (inner) shelfal settings in the northern region (OM) to distal basinal regions southward (AC). A series of well-developed phylloid-algal car-

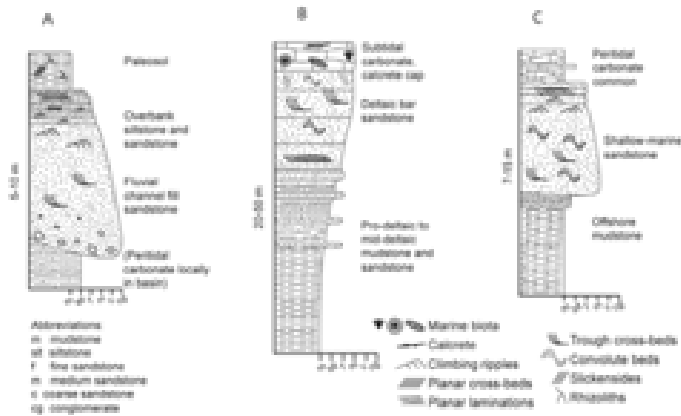


FIGURE 5. Example vertical logs of siliciclastic facies associations. **A**, Facies association A (fluvial strata). **B**, Facies association B (deltaic strata). **C**, Facies association C (shallow marine strata).

bonate mound buildups occur within the middle Virgilian interval in the central part of this transect (HC, Fig. 4; Soreghan and Giles, 1999a) that appear to record an inflection point on this depositional profile. Phylloid algal buildups also occur in the Virgilian section in the Sacramento Mountains (e.g., Toomey et al., 1977), inferred to reflect a narrow shelf margin here. Vertical facies successions and depositional environments within individual sequences (cycles) vary markedly across the study transect, but most sequences consist of a mixed carbonate-siliciclastic lithofacies. In the northern regions (OM, RC), these cycles commonly appear similar to cycles described for the Virgilian section of the eastern Sacramento shelf, in that they contain a subtidal carbonate member that is normal marine and bounded by surfaces of subaerial exposure. In southern regions (HC, AC), however, the facies successions mostly exclude normal-marine subtidal carbonate, and instead contain restricted carbonate facies in irregular successions, with thick intervals of silty carbonate mudstone in addition to the sandstone facies. In view of the shifts exhibited across the study transect, we address the sequence stratigraphy by region from north (landward) to south (seaward). Although we infer a general north (landward)-to-south (seaward) gradient for the study transect, out-of-plane sedimentation, especially of siliciclastic sediments, likely complicates the sequence stratigraphy.

Northern Region

Carbonate-dominated, upwardly shallowing sequences (10-20 m thick) are common in the lower part of the section in the northernmost locality. These sequences consist predominantly of normal-marine limestone (bioclastic grainstone, packstone, and wackestone; Table 2, Fig. 5) and locally include a thin (1-2 m) dark gray to black, fissile marine shale at the base. The basal shale typically is sharply overlain by a 2-10 m thick packstone or grainstone, capped by a calcrete. In one cycle, fluvial sandstone (facies association A) overlies the calcrete.

In contrast to the carbonate-dominated sequences in the lower part of the section, deltaic strata (facies association B) are common in sequences within the upper part of the section, with rare occurrences of fluvial facies. These sequences (10-20 m thick) include upwardly coarsening deltaic siliciclastic strata (facies association B, Fig. 5), abruptly overlain by 1-5 m of fossiliferous carbonate packstone or grainstone capped by a subaerial exposure surface (calcrete).

Central Region

Sequences in Rhodes Canyon and the lower part of the section in Hembrillo Canyon are ~50-70 m thick with ~20-55 m of deltaic siliciclastic strata (facies association B) overlain abruptly by 2-15 m of carbonate wacke-, pack- or grainstone (Figs. 4, 5), capped by calcrete. Commonly, a thin (0.2-2 m) dark oncoidal wackestone marks the base of the sequence

TABLE 3. Summary table of paleocurrent data.

Location	Meters from base of section	Type of sedimentary structure	Mean direction (azimuth)	Number of measurements
Oscura Mountains	260	Trough cross-limbs	203°	19
	303	Trough cross-limbs	238°	19
Rhodes Canyon	205	Trough cross-limbs	247°	20
	377	Trough cross-limbs	132°	18
Hembrillo Canyon	145	Trough cross-limbs	188°	15
	195	Tabular cross-beds	191°	19
	195	Trough cross-limbs	230°	33
	520	Trough axes	130°	19
Ash Canyon	160	Trough cross-limbs	246°	11
	292	Trough axes	222°	5
	292	Trough cross-limbs	231°	22
	345	Trough cross-limbs	250°	16
	348	Climbing ripples	277°	10
	424	Trough cross-limbs	280°	26
	520	Trough cross-limbs	144°	13

* low angle trough cross-limbs

(Soreghan, 1992, 1994a). In the middle to upper parts of the section in Hembrillo Canyon, sequences comprise 5-15 m of marine silty carbonate mudstone overlain by thin (< 2 m) restricted (peritidal) carbonate, similar to sequences in the southernmost area (described below).

Of all the regions studied within the San Andres Mountains (western Orogrande basin), the sequences of the northern and lower central regions appear most similar to those of the well-studied eastern Orogrande shelf. Key similarities include the occurrence of normal-marine carbonate strata abruptly capped by surfaces of subaerial exposure (cf. Goldstein, 1988; Rankey et al., 1999), with a distinct absence of peritidal facies.

Southern Region

Two fundamentally different types of sequences occur in this area and are distinguished by the presence of either a fluvial (rarely) or shallow-marine siliciclastic (more commonly) component (Fig. 7). Sequences with a basal fluvial component occur only twice in the lower part of the section and once higher in the section. These sequences range from 20-25 m thick, and consist of a basal fluvial unit up to 8 m thick (facies association A). The fluvial unit is abruptly overlain by 5-20 m of marine shale, which is sharply overlain by 1-3 m of peritidal carbonates with common algal laminations, gastropods, and desiccation cracks (also noted by Schoderbek, 1991). In at least one instance, the basal contact of the fluvial unit truncates the underlying cycle, evident by rip-up clasts (up to 5 cm diameter) of the subjacent peritidal carbonate within the basal conglomerate of the fluvial unit.

Marine siliciclastic-dominated sequences typically consist of a basal marine shale (7-20 m thick), sharply overlain by 7-15 m of shallow-marine sandstone (facies association C), sharply overlain by peritidal carbonate (1-3 m thick). These sequences typically range from 20-30 m thick.

Sequence Stratigraphic Interpretation

As established by others for icehouse sequences of the late Paleozoic in general, and the Orogrande basin in particular (previously referenced), the high-frequency sequences of the western Orogrande basin were controlled predominantly by high-amplitude, high-frequency glacioeustasy. Soreghan and Giles (1999b) documented glacioeustatic amplitudes in excess of 80-100 m for Virgilian carbonate-mound-dominated strata in this region (HC; Fig. 4). Such large amplitudes of sea level change in the eperic Orogrande basin resulted in the juxtaposition of normal-marine subtidal facies abruptly capped by subaerial exposure

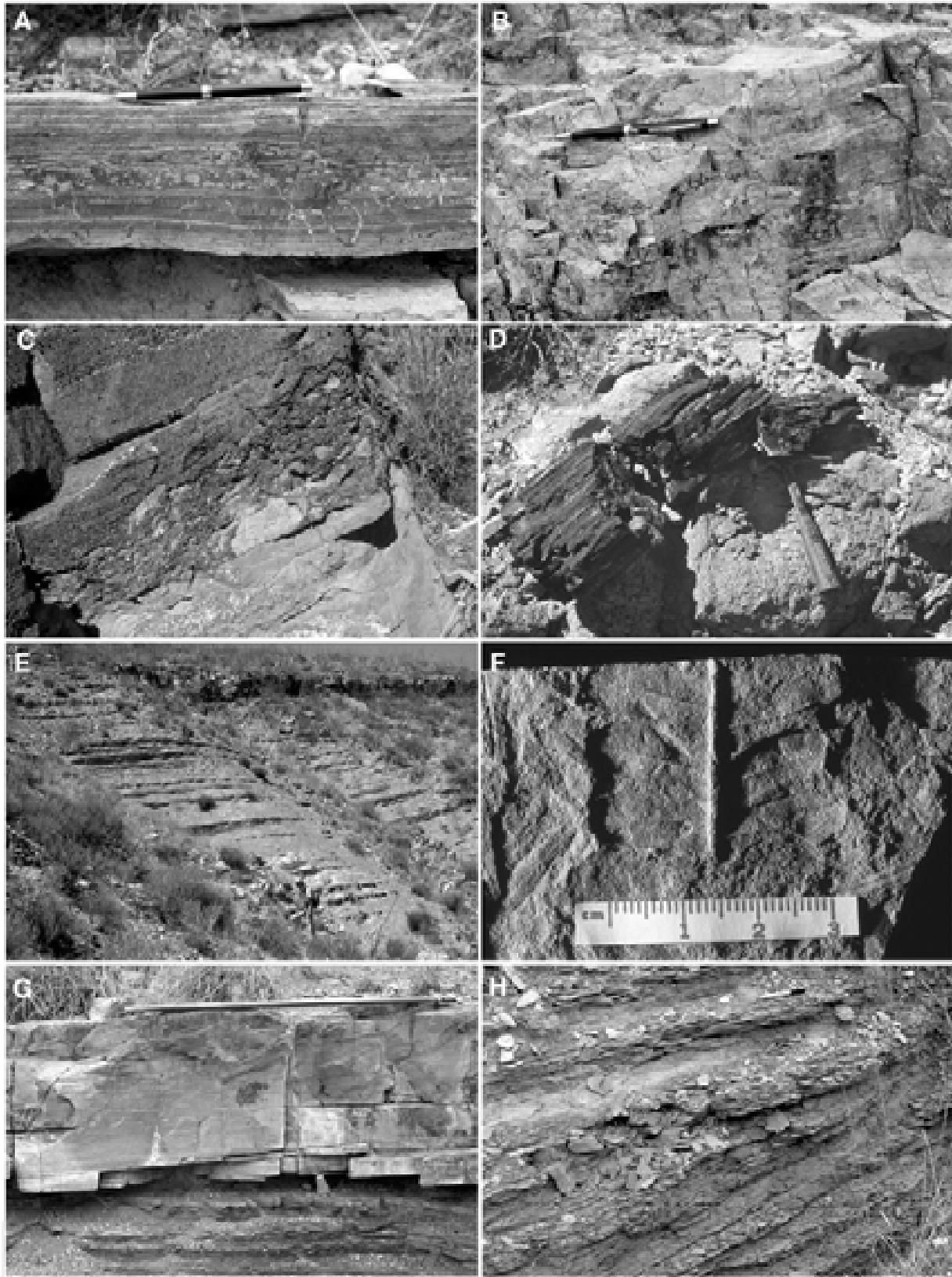


FIGURE 6. Photos of selected facies. **A-B**, Examples of peritidal carbonate facies, **A**, intraclastic laminite and **B**, irregularly (cryptalgal) laminated dolomicrite (Hembrillo Canyon). **C-D**, Facies association A (fluvial): **C**, Conglomerate at base of succession, with intraclasts of peritidal carbonate, erosively atop a peritidal carbonate (Ash Canyon); **D**, log of petrified wood within inferred fluvial sandstone (Oscura Mountains). **E-F**, Facies association B (deltaic): **E**, Thick interval of interbedded fine sandstone and siltstone (Hembrillo Canyon); **F**, plant impressions from this facies. **G-H**, Facies association C (marine): **G**, Bar sandstone abruptly atop the dark, silty carbonate mudstone facies (Hembrillo Canyon); **H**, detail of dark, silty carbonate mudstone facies common in the southern locations (note pen for scale; Hembrillo Canyon).

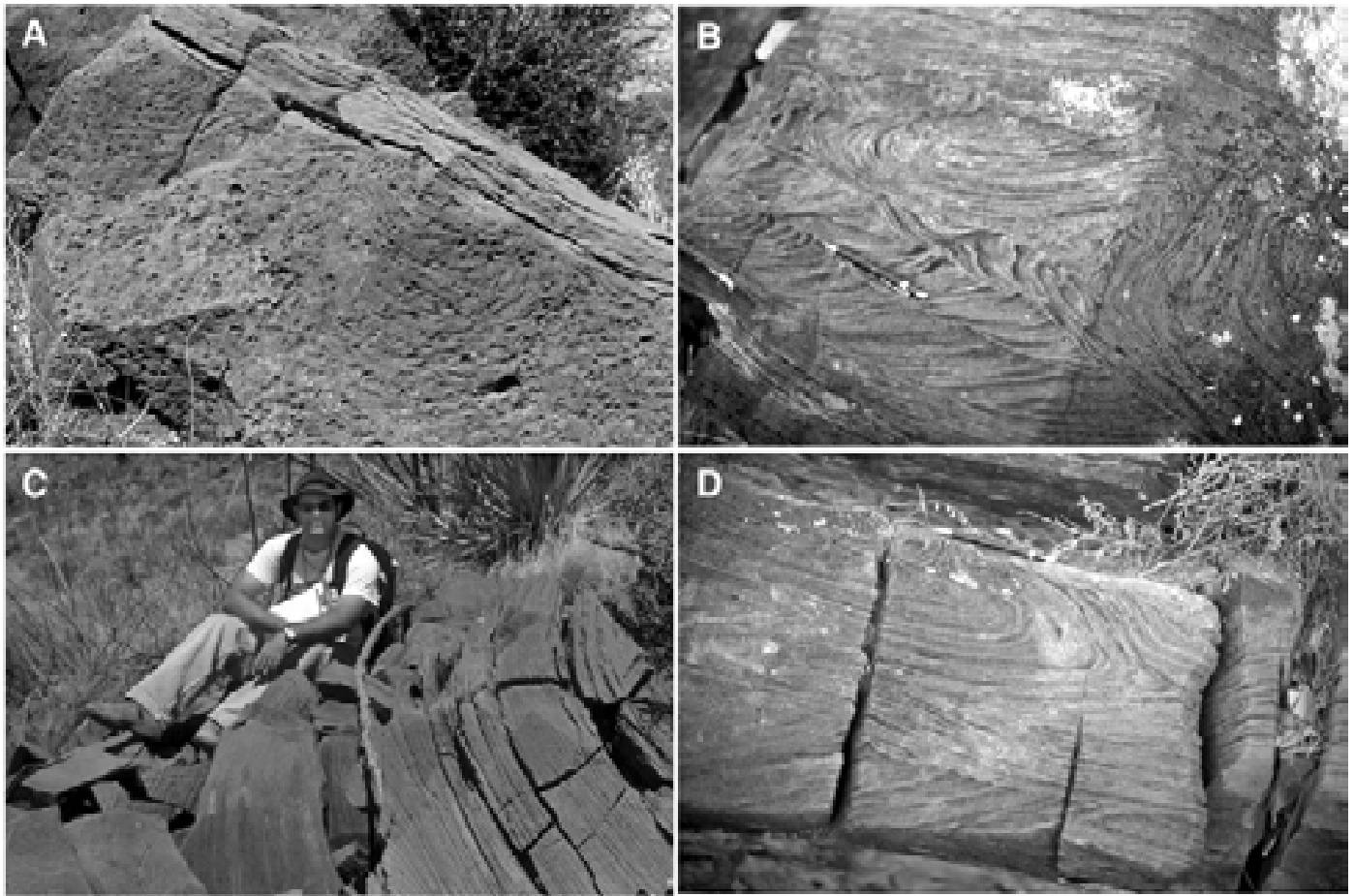


FIGURE 7. Photos of soft-sediment deformation in sandstone facies; all examples pictured are from the Ash Canyon region, within the basinal region of the Orogrande basin. These features are interpreted to reflect probable seismicity proximal to this region.

surfaces in shelfal regions, and restricted carbonates with marine and local continental siliciclastic strata in basinal settings. Wilson (1967) first described the model of “cyclic and reciprocal sedimentation” in which siliciclastic sediments bypassed the exposed eastern Orogrande shelf and entered the basin during sea level lowstands, whereas normal-marine carbonate strata accumulated across the shelf during sea-level highstands. This model has since been elaborated upon by others (e.g., Raatz and Simo, 1998; Rankey et al., 1999) for the narrow, relatively low-gradient shelf-slope transect exposed in the Sacramento Mountains. Figure 8 illustrates sequence stratigraphic interpretations, as detailed below. We note possible alternative interpretations for selected phases of these cycles.

Lowstand

During sea-level lowstand, shelfal areas, which were the site of normal-marine carbonate deposition during the previous highstand, were exposed, and calcretes developed atop subtidal carbonate facies, owing to an absence of accommodation on the shelf (Fig. 8A). Southward, in more basinal areas (AC region) restricted peritidal carbonates accumulated. During periods of extreme lowstand (3rd-order sea-level fluctuations), these basinward locales were also exposed. Hence, during extreme lowstands, siliciclastic sediments bypassed the exposed shelf as fluvial systems extended into basinal regions and, by late lowstand, locally incised peritidal carbonate strata. This is evinced locally by the occurrence of sandstone units with erosive bases containing intraclasts of peritidal facies (Figs. 6A, 8) lying directly above restricted peritidal carbonates. However, we acknowledge that fluvial deposition could also occur during other stages of the sea-level cycle in various regions along

the depositional transect, such as early transgression (e.g., Fig. 8B) and falling-stage times in mid-shelf locations, and landward locations in highstand phases.

Transgression

At late lowstand, siliciclastic sediment bypassed the exposed shelf as fluvial systems extended into basinal regions and locally incised underlying peritidal carbonates (Figs. 6A, 8). With transgression, the basin began flooding, marked locally by a transgressive lag atop fluvial facies, or oncoidal carbonate lags atop peritidal facies (basinward). It is possible that fluvial sediments were progressively trapped and stepped landward during transgression (Fig. 8B). Minimal sedimentation occurred in the basin, excepting some dark, silty carbonate mudstone facies in southern/basinal (AC and, to a lesser degree, HC) locations. Initial flooding across the shelf is recorded locally by transgressive lags of dark oncoidal wackestone (Soreghan, 1992) developed immediately above subaerially exposed carbonates.

Highstand

During early highstand, marginal marine and marine siliciclastic sediments accumulated (Fig. 8C), and by late highstand, submergence of the entire shelf pushed the shoreline northward, sequestering siliciclastic sediment landward and enabling accumulation of normal-marine carbonate in shelfal settings in the central and northern regions (Fig. 8C). We infer that the phylloid-algal mounds of the Hembrillo Canyon region (Fig. 4) accumulated during highstand phases. Basinward, relatively dark (organic-rich) offshore shale and silty carbonate mudstone accumulated as landward sequestering of detrital material significantly reduced siliciclastic influx into the basin (Wilson, 1967; Raatz and Simo, 1998).

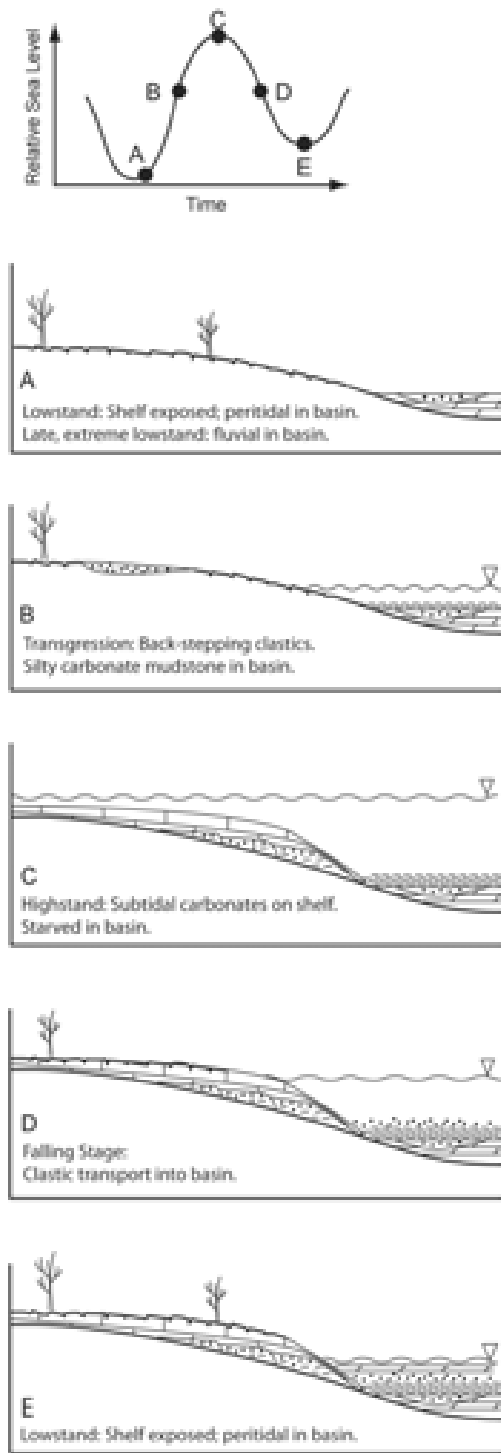


FIGURE 8. Depositional model for Late Pennsylvanian high-frequency sequences, with inferred sea level curve on each panel. **A**, During lowstand, exposure on the shelf correlates basinward to peritidal carbonate accumulation. During extreme lowstands, late-lowstand fluvial systems extend into the basin, locally incising peritidal carbonate strata. **B**, During transgression, fluvial systems may be preserved shelfward, and minimal (mudstone) deposition occurs basinward. **C**, By highstand, deltaic and marine siliciclastic sediments prograde, but ultimate complete submergence of the shelf causes sequestration of siliciclastic material landward, enabling normal-marine subtidal carbonate accumulation. **D**, As sea level falls, forced regression results, in some cases, in delivery of siliciclastic sediments to the basin. **E**, At late lowstand, the shelf is entirely exposed, creating paleosol development on highstand carbonate strata, and peritidal facies accumulate in the basin.

Falling sea level

As sea-level fell, upper shelfal regions were exposed and calcretes developed atop subtidal carbonate facies. The development of a sub-aerial exposure surface atop subtidal carbonate implies that exposure was not a result of vertical accretion to sea-level, but rather a direct response to a eustatic sea-level fall (Soreghan and Dickinson, 1994; Rankey et al., 1999; Fig. 8D). Accordingly, this scenario is consistent within a forced regressive system (Plint and Nummedal, 2000; Posamentier and Morris, 2000). Fluvial siliciclastics bypassed the exposed shelf and accumulated in shallow-marine environments within the basin (Fig. 8C). The occurrence of shallow-marine sandstone above highstand offshore shale and below lowstand peritidal carbonates is consistent with deposition during the falling stage system tract (FSST) of Plint and Nummedal (2000). Alternatively, such strata could reflect highstand progradation under conditions of fixed sea level.

UPPER PENNSYLVANIAN THICKNESS TRENDS

Data sources and approach

Missourian and Virgilian isopach maps (Figs. 9 and 10) were constructed using data from this study combined with previously published thicknesses from outcrop and subsurface sources (Table 4). For this study, Virgilian thicknesses were determined by combining the thicknesses of the Virgilian Panther Seep Formation and its correlatives throughout the region with the Bursum or Laborcita formations, as Wahlman and King (2002) and Kues (2002) advocated a latest Pennsylvanian (Virgilian) age for most of the Bursum and Laborcita formations. In the study area, however, biostratigraphic control is locally sparse. For example, the Panther Seep Formation includes upper Missourian strata (Soreghan, 1992, 1994a), and the Bursum Formation might locally include lowest Wolfcampian strata (B. Kues, personal comm., 2002). Regionally, however, insufficient biostratigraphic data exist to precisely locate the Missourian-Virgilian and Pennsylvanian-Permian boundaries. Accordingly we used lithostratigraphy as an approximation of chronostratigraphy and recognize that this may result in errors of up to 100 m (B. Kues, personal comm., 2002). Therefore stage thicknesses used in construction of the isopach maps are coded as high confidence (good biostratigraphic control, minimal post-depositional disruption) or low confidence (estimated thickness). In addition to thickness data, the ultimate isopach shape for the Virgilian reflects consideration of auxiliary data such as facies and paleocurrent data from this and previous studies (Fig. 10).

Missourian isopach patterns

The Missourian isopach map (Fig. 9) indicates that the Orogrande basin was thickest in the northern and central regions and thinned gradually toward the southwest. The lack of significant thickening across the region indicates that the Orogrande basin had not yet experienced significant tectonic development.

Virgilian thickness trends

The Virgilian isopach map (Fig. 10) differs from previously published maps (Meyer, 1966; Wilson and Jordan, 1988; Schoderbek, 1994) in that it (1) includes data points not available previously, and (2) depicts a shape based fundamentally on thickness data but further constrained by facies, paleocurrent, and structural data (details below).

In contrast to the subtle thickness trends of the Missourian isopach map, a generally rhombic shape emerges for the Virgilian Orogrande basin (Fig. 10), with margins that parallel pre-existing Precambrian structures identified in earlier studies (e.g., Karlstrom and Daniel, 1993; Daniel et al., 1995; Woodward et al., 1999; Heizler et al., 2000), and development of hingelines along the eastern, southeastern, and western margins. Abrupt thinning of Virgilian-aged strata updip and parallel to these Pre-



FIGURE 9. Missouriian isopach map, showing minimal development of the Orogrande basin; see Table 4 for data references.

cambrian structures may reflect Pennsylvanian reactivation of these structures, as demonstrated for analogous structures in ARM systems in northern New Mexico (e.g., Cather et al., 2005, 2006). The northern margin of the basin displays a generally north-south oriented ramp configuration with the basin depocenter located to the south, but appears to have shifted from the southwest (west of AC) to the southeast through Virgilian time, evident in part by a paleocurrent shift from southwest-directed flow during the early part of the Virgilian to southeast-directed flow during late Virgilian time. A thick siliciclastic-rich section in the Mud Springs Mountains to the northwest (Soreghan, 1992) exhibits easterly directed paleocurrent directions (Soreghan, unpublished data, 1992); together with thick subsurface sections immediately west of RC (data points 26, 27, 28, Fig. 13, Table 4), this pattern suggests the possibility of a siliciclastic fairway to the Orogrande basin entering from the northwest (Fig. 10). Abrupt thinning of Virgilian strata to the southeast may indicate the presence of a northeast-southwest oriented structural hingeline acting as a structural “sill” isolating relatively thin Virgilian strata in the Franklin and Hueco mountains in a southeastern depocenter. Indeed, thinning of uppermost Pennsylvanian (Virgilian)-lowermost Permian (Wolfcampian) strata, and facies changes in the vicinity of the Texas-New Mexico border (Meyer, 1966; Kottlowski, 1969) have led some to suggest the Orogrande basin may have been subdivided into two sub-basins (Simo et al., 2000).

DISCUSSION

Late Pennsylvanian Paleogeography of the Orogrande Basin

Data from facies associations, paleocurrent data, sequence stratigraphy, and thickness trends were integrated to construct paleogeographic maps of the Orogrande basin for Late Pennsylvanian time (Figs. 11 and 12A-D).

Missourian: During Missourian time, the Pedernal uplift to the east began to take the shape of a tilted fault block (Kottlowski, 1969), but subsidence remained minimal within the Orogrande basin. Fluvio-

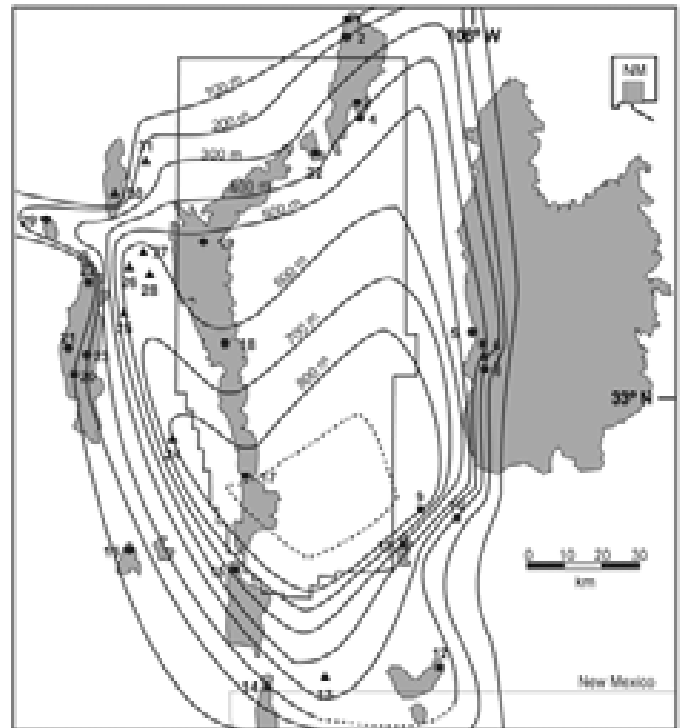


FIGURE 10. Virgilian isopach map, showing strong development of the Orogrande basin, with relatively well defined margins, and a thick sedimentary section; see Table 4 for data references.

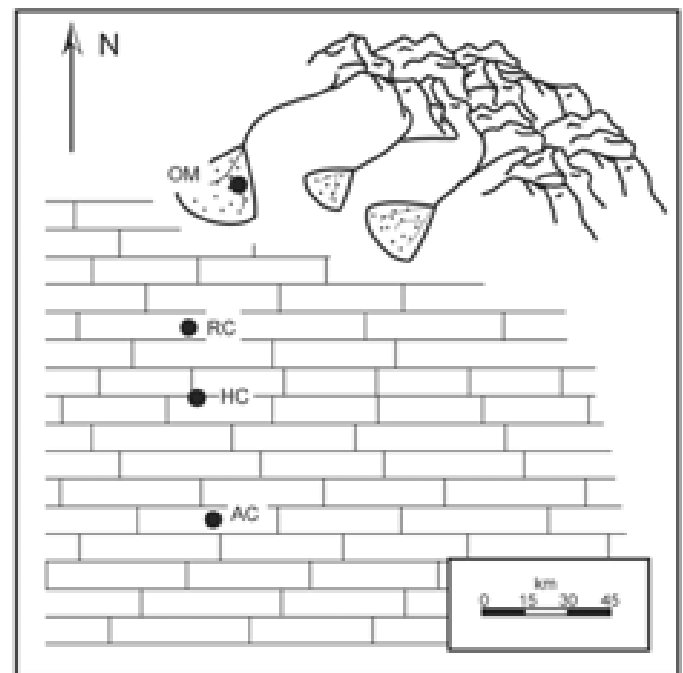


FIGURE 11. Paleogeography of the Orogrande basin during Missourian time, showing northeastern Pedernal uplift source region and predominantly carbonate facies across the basin (OM = Oscura Mountains, RC = Rhodes Canyon, HC = Hembrillo Canyon, AC = Ash Canyon).

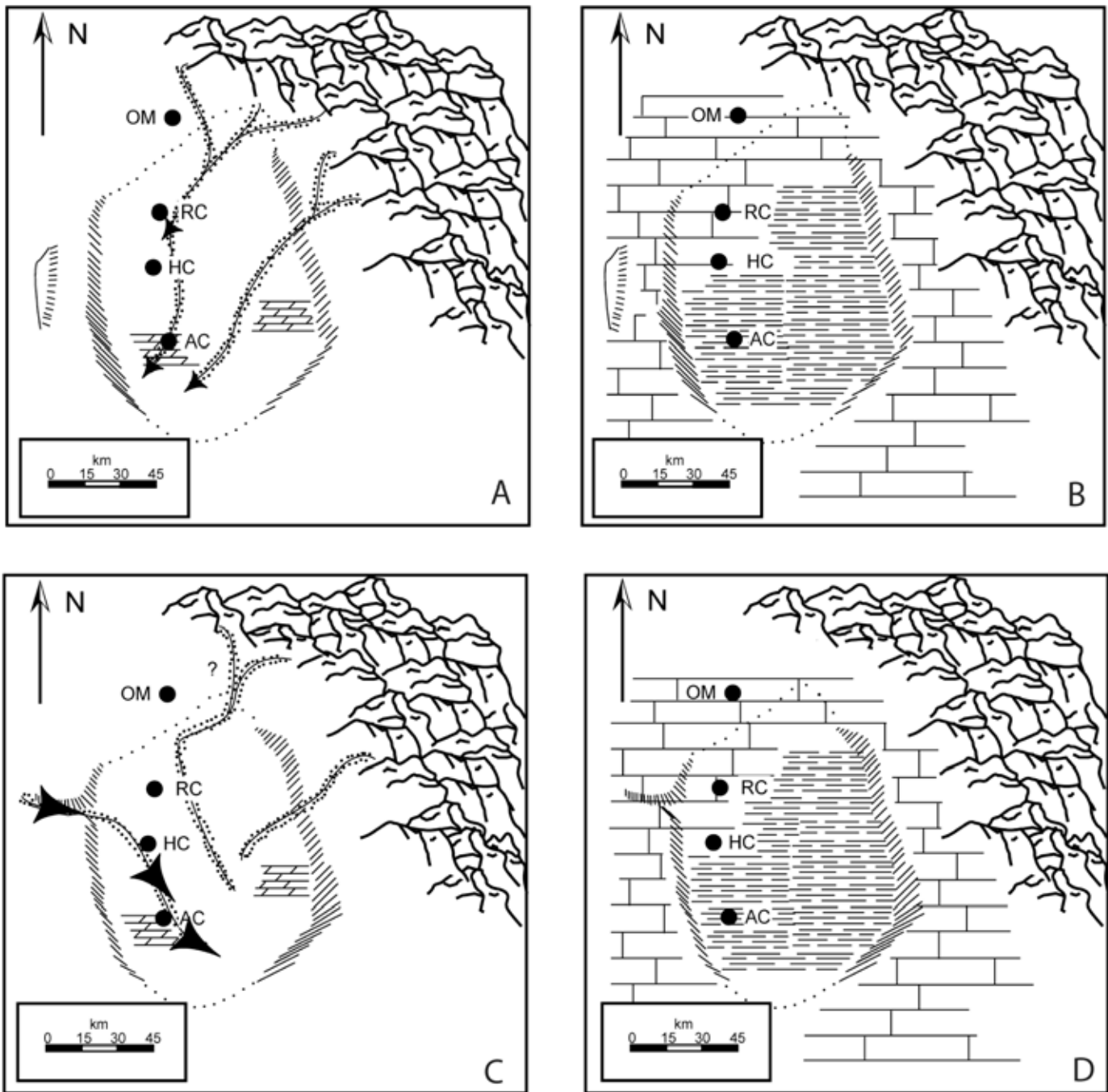


FIGURE 12. Paleogeography of the Orogrande basin during **A**, early Virgilian lowstand, **B**, early Virgilian highstand, **C**, late Virgilian lowstand; paleocurrent directions in Mud Springs Mountains locality to the west from Soreghan (unpublished data, 1992) and in Sacramento Mountains from Dunning (1978), and **D**, late Virgilian highstand (OM = Oscura Mountains, RC = Rhodes Canyon, HC = Hembriillo Canyon, AC = Ash Canyon; MS= Mud Springs Mountains).

deltaic and shallow-marine siliciclastics accumulated in the northern and eastern basin margins, and by late Missourian time siliciclastic influx occurred only in the northeastern part of the basin (OM). This is consistent with early Virgilian isopach patterns and paleocurrents that suggest a siliciclastic source to the northeast. These siliciclastic strata include arkosic sandstone and chert-pebble conglomerate derived from both sedimentary (carbonate) and Precambrian crystalline rocks progressively exposed in the denuded Pedernal uplift (Kottlowski et al., 1956; Kottlowski, 1960, 1968; Pray, 1961; Kottlowski and Benjar, 1971). Westward (Beeman Formation) and southward, the Missourian section comprises both normal-marine limestone, as well as some calcareous

shale (Raatz and Simo, 1998; Fig. 11). However, new fusulinid and conodont biostratigraphy on the Beeman (by G. Wahlman and J. Barrick, respectively), demonstrated that the top of the Beeman is of early-middle Virgilian age.

Virgilian: By Virgilian time, the Pedernal uplift formed a highland bordering the Orogrande basin to the east/northeast (Thompson, 1942; Kottlowski, 1960, 1968). Non-depositional to erosional domes developed locally along the eastern margin (e.g., Pray, 1961; Algeo, 1992) and western margin (e.g., Maxwell and Oakman, 1986; Soreghan, 1992; Lawton et al., 2002) of the basin resulting in intraformational unconformities and relatively thin and incomplete Virgilian sections in

selected parts of the present-day Caballo, Mud Springs, and Sacramento mountains (Figs. 3, 12A-B). Facies interpretations and paleocurrents indicate that, during early Virgilian time, fluvial and fluvio-deltaic systems drained the Pedernal highland to the east and northeast and flowed southwesterly toward the basin depocenter (Fig. 12A). Fluvial units low in the AC section (lower ~300 m) entered the basin only during lower (3rd) order sea level lowstands and exhibit southwesterly directed flow, indicating that the basin depocenter during early Virgilian time lay west of Ash Canyon. Even farther west, in the Robledo Mountains, the Virgilian section is thin and carbonate-rich (Wahlman and King, 2002). On the basis of this observation, we infer the presence of a fault-controlled margin west of Ash Canyon, that effectively ponded siliciclastics within the basin; southwesterly directed paleocurrents in Ash Canyon, proximal to this margin, record a marked basin asymmetry for this time. Correlative Virgilian facies trends on the eastern basin margin record growth of the north-northwesterly trending La Luz anticline (Pray, 1959; Wilson, 1967), reflecting the active tectonism of this phase. Marine siliciclastic sediment accumulated in the central part of the study transect (HC, RC) during transgression, and by highstand, shelfal regions around the margins of the basin were submerged, siliciclastics were sequestered landward, and normal-marine carbonate strata accumulated in shelfal regions (Fig. 12B).

In the Mud Springs Mountains to the northwest (Soreghan, 1992), paleocurrents from a fluvial sandstone unit in the upper Virgilian section indicate an easterly flow direction, and the section is significantly thicker than nearby carbonate-rich Virgilian sections in the Caballo and Fra Cristobal mountains immediately to the south and north, respectively (Fig. 10; Soreghan, 1992). Fluvial and fluvio-deltaic systems within the upper Virgilian sections in Ash, Hembrillo, and Rhodes canyons (Fig. 3) exhibit southeasterly directed flow, in contrast to the southwesterly directed flow of the lower Virgilian. These data suggest a shift in the basin gradient, and possible addition of a source region to the west-northwest by the late Virgilian; however, fluvial and fluvio-deltaic systems within the upper Virgilian sections in the northern region (OM) record a continued contribution of siliciclastics from source areas to the northeast. The relatively thick Mud Springs Mountains section, and southeasterly directed flow from fluvial and fluvio-deltaic systems in the central, southern, and northwestern parts of the basin record a period of major incision during the late Virgilian (Fig. 12c). The change to southeasterly directed paleocurrents records a shift in the basin depocenter from southwest of Ash Canyon in early Virgilian time to the south-southeast in the late Virgilian. This shift of the basin depocenter toward the steep, tectonically active eastern and southeastern margins of the basin records marked asymmetry similar to that of early Virgilian time, and a reversal of the early Virgilian regional tilt of the basin. Our data suggest that deltaic systems developed in the north and became increasingly common in the southern Oscura Mountains during the late Virgilian.

Late Pennsylvanian Subsidence in the Orogrande Basin

Previous workers have proposed various models for the tectonic origin of the Orogrande basin. Handschy (1985) and Handschy et al. (1985) favored a foreland basin model for both the Orogrande and Pedrogosa basins, suggesting they represent the northernmost extension of deformation associated with the foreland of the Ouachita-Marathon orogenic front. In contrast, Ye et al. (1996) proposed that asymmetric thickening of strata toward basement uplifts in various basins of the Ancestral Rocky Mountains (including the Orogrande basin) implies foreland basin deformation in the foreland of a subduction zone along the southwestern margin of North America during the Late Paleozoic. Goetz and Dickerson (1985) and Wilson and Jordan (1988) inferred an origin by crustal extension with some transcurrent movement related to a northwest-southeast trending transform margin along the southern edge of North America. Algeo and Wilson (1991) inferred that the Orogrande basin was a north-trending pull-apart basin resulting from cratonward overstep a broad dextral transform margin to the south. Algeo (1992)

later inferred that the faults and folds along the eastern margin of the Orogrande basin were the result of dextral transpressional shear during the Ouachita-Marathon orogeny. This is consistent with other interpretations of transpressional or transtensional basin tectonics proposed by Singleton (1990), Beck and Chapin (1995), Giles and Lawton (1996), and Lawton et al. (2002).

The presence of a thick (~800 m) section that includes southwesterly flowing fluvial systems in the southernmost region (AC) correlative with a thin carbonate-dominated section immediately to the west in the present-day Robledo Mountains is consistent with the possible presence of a discrete seismically active, fault-controlled margin between these localities, and records an asymmetry of the Orogrande basin that shifted during Late Pennsylvanian time from a southwesterly tilt to a southeasterly tilt. The prevalence of inferred seismites (soft-sediment deformation features) occurring in Ash Canyon coincides with the thickest Virgilian-aged section, and reinforces the interpretation of a western structural hingeline in this region, with reduced occurrence of seismites with distance from this region (cf. Greb and Dever, 2002).

Back-stripping analysis using data from the Ash Canyon section indicates that the Orogrande basin underwent very rapid subsidence in Virgilian time (Fig. 13), yet little evidence exists for very deep-water deposition here. The presence of peritidal facies episodically throughout the Virgilian section in the central and southern (HC and AC) regions records near-exposure conditions in basinal regions at lowstands. This, together with glacioeustatic fluctuations of 80-100 m (Soreghan and Giles, 1999b) imply that basinal water depths even at highstands were not excessive. Thick successions of the silty carbonate mudstone, and in some cases of peritidal facies, suggest that sedimentation largely paced creation of accommodation.

The possibility of seismic activity along both western and eastern boundaries, together with the occurrence of features such as 1) Pennsylvanian-aged transcurrent faults (eastern margin; Pray, 1961; Algeo, 1992), 2) intraformational unconformities reflecting apparent doming in both western and eastern margins (Pray, 1961; Maxwell and Oakman, 1986; Algeo, 1992), 3) a single, rapid pulse of subsidence during the late Virgilian (Fig. 13) resulting in a relatively small basin, are all consistent with basins formed in transcurrent tectonic settings. Accordingly, this southwestern terminus of the Ancestral Rocky Mountains appears to record complex and short-wavelength deformation within an overall wrench system.

CONCLUSIONS

Virgilian siliciclastic facies preserved in cyclic strata of the western Orogrande basin include (1) upwardly fining packages of lenticular sandstone interpreted to record fluvial deposition, (2) thick mudstone to sandstone inferred to record deltaic deposition, and (3) laterally continuous fine- to medium-grained sandstone interpreted to record shallow-marine deposition. These facies accumulated during rapid sea level fluctuations generally associated with high-frequency (4th-order) glacioeustasy that was prevalent during the Late Pennsylvanian icehouse interval. Sea level lowstands were marked by widespread exposure in shelfal regions, resulting in calcrete development atop subtidal carbonate strata, widely regarded as typical of icehouse cyclicity. In more basinal regions, however, peritidal carbonates accumulated during lowstands. During transgression to early highstand, marine siliciclastics were preserved in the central ramp region, whereas offshore mudstone dominated deposition in the basin. Shelfal areas were completely flooded during highstands, resulting in normal-marine carbonate accumulation. Preservation of peritidal facies is considered atypical of icehouse systems (e.g., Read, 1995), and are certainly absent in shelfal positions, where subtidal carbonates occur capped by lowstand exposure surfaces. However, basinal regions appear to have provided the accommodation needed at lowstands to preserve these facies. Raatz and Simo (1998) documented a similar relationship between shelfal and basinal cycles for the Beeman Formation (Missourian-early Virgilian) of the western Orogrande basin,

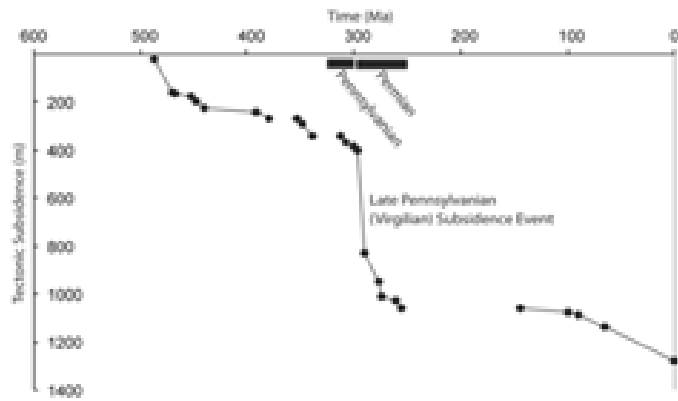


FIGURE 13. Tectonic subsidence through time for the Orogrande basin. See text for methods and discussion.

which they termed “inverse sedimentation.”

The Virgilian represents the most rapid pulse of subsidence of the Orogrande basin (Fig. 13). A distinct fault-controlled margin has long been inferred for the eastern basin margin (Sacramento shelf), but isopach and other data suggest possible fault-controlled hingelines on the western and possibly southern basin margins as well. Thickness patterns and paleocurrents indicate a southeastward shift of the basin depocenter during the Virgilian, recording an inversion of regional tilt in

the basin. Basin asymmetry and depocenter shifts, and active tectonism on both the western and eastern margins of the basin are consistent with probably transcurrent (transtensional/transpressional) motion in this part of the southern Ancestral Rocky Mountains.

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**THE ARTINSKIAN-KUNGURIAN (UPPER LOWER PERMIAN)
CALCAREOUS ALGAE AND SMALLER FORAMINIFERS OF
THE YESO GROUP AND SAN ANDRES FORMATION (NEW MEXICO, USA)**

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Although the Late Pennsylvanian and early Early Permian foraminifers are well-known in New Mexico, the late Early Permian (or Leonardian regional stage) remains poorly investigated. In general, the Leonardian regional stage is poorly known in the USA and all of the Americas. The Uralian equivalents of the Leonardian are Artinskian plus Kungurian; the Tethyan ones are Yakhtashian plus Bolorian.

The Leonardian regional stage was subdivided into three fusulinid biozones, from bottom to top: (a) PL1: zone of *Schwagerina* (now *Praeskinnerella*) *crassitectoria*; (b) PL2: zone of transitional *Parafusulina*; (c) PL3: middle zone of *Parafusulina*. The biozone PL1 is well-represented by the lower Bone Spring Formation in the Delaware Basin of southeastern New Mexico-West Texas, whereas to the west its upper part corresponds to the lower part of the Yeso Group. The biozone PL2 is represented by the middle Bone Spring Formation and the middle Yeso shales and carbonates. The biozone PL3 corresponds to the upper Bone Spring Formation and the upper Yeso Group. Moreover, PL3 contains the first *Boultonia* (now *Ogbinella*) *guadalupensis*, which were encountered in the subsurface in the San Andres Formation. These three biozones are considered to be either Artinskian and Kungurian or only Kungurian in age; the two first correspond to the Hessian regional substage and the last one to the Cathedralian; a zone of *Chalartoschwagerina hawkinsi* was eventually added at the base of the Leonardian and correlated with the latest Artinskian.

No biozonations with smaller foraminifers have been established in the Leonardian beds, and, in general, the knowledge of these groups is very poor in America. References are more numerous in the Urals and other regions of the former SSSR, as well as in the Arctic islands (Canada, Norway, etc.). This summary intends (1) to list a rich microflora of carbonate algae with interesting transitional forms recently recovered from Yeso and San Andres strata in central New Mexico, (2) to document an interesting microfauna of smaller foraminifers – Globivalvulinoidea, Miliolata and Nodosariata, and (3) to try to provide biostratigraphic and paleobiogeographic data.

We studied the Yeso Group in the southern Fra Cristobal Mountains where a complete section is exposed at Massacre Gap, and in the McLeod Hills of the southern Caballo Mountains, Sierra County, New Mexico. The San Andres Formation was studied in its reference section in the northern San Andres Mountains, approximately 1.6 km west of the type section near Rhodes Canyon, Sierra County, New Mexico. The Lower Permian Yeso Group is divided into the Arroyo de Alamillo Formation and Los Vallos Formation, whereas the Los Vallos Formation is further divided into Torres, Cañas and Joyita members. The Yeso Group is underlain by nonmarine redbeds of the Abo Formation and overlain by eolian and shallow marine strata of the Glorieta Sandstone.

At Massacre Gap in the Fra Cristobal Mountains, the Torres Member is approximately 134 m thick and composed of dolomite, gypsiferous siltstone, gypsum and siltstone to fine-grained sandstone. The succession is composed of six transgressive-regressive cycles marked by six intervals of dolomite separated by thicker intervals composed of siltstone, gypsiferous siltstone, gypsum and minor carbonate and sandstone beds. Dolomite occurs as bedded to massive, and is locally indistinctly laminated.

Calcareous algae and foraminifers were recognized in dolomite horizons 2 and 4. In dolomite horizon 2, algae and foraminifers occur in an oncoidal floatstone to rudstone, composed of oncoids up to 6 mm in diameter, coated grains and fossil fragments, including skeletons of bivalves and brachiopods, small gastropods, foraminifers, ostracods, echi-noderms and calcareous algae.

In dolomite horizon 4, a dolomitized wackestone is intercalated that is composed of peloidal matrix and skeletons of bivalves, gastropods, ostracods, crinoids, foraminifers and calcareous algae. Intercalated in dolomite horizon 4, there are also thin rudstone (coquina) layers composed of abundant bivalve and brachiopod shell fragments, minor amounts of gastropods and ostracods, and rare fragments of crinoids and calcareous algae.

The dolomite horizon in the middle of the Torres Member in the McLeod Hills of the southern Caballo Mountains contains mudstone to wackestone that is composed of peloidal micrite and a low diversity fossil assemblage of locally abundant ostracods, echinoderm fragments and a few smaller foraminifers. Dolomite of the Torres Member was deposited in a shallow marine, mostly restricted environment indicated by the low-diversified fossil assemblage. In this probable coastal sabkha, partly evaporitic conditions prevailed. Rarely, normal marine incursions into the sabkha were present, indicated by diversified fossil assemblages. Coquina layers can be interpreted as storm layers (tempestites).

The reference section of the San Andres Formation, exposed in the northern San Andres Mountains is approximately 130 m thick. The succession rests on very thin Glorieta Sandstone, has an eroded top and is divided into lower, middle and upper parts. The lower part is 30 m thick, composed of thick bedded to massive, and rare thin bedded, gray to dark gray, bituminous muddy limestone. The middle part measures 33 m and consists of thin to thick bedded limestone and massive limestone beds up to 4.3 m thick. Limestone is gray to dark gray and of muddy texture. The upper part is 67 m thick and composed of medium to thick bedded, partly indistinctly bedded, light to dark gray limestone; which rarely contains chert nodules. Richly bioclastic wackestone is the dominant microfacies, whereas subordinate ones are rudstone, floatstone and in the upper part grainstone. Smaller foraminifers and calcareous algae are present throughout the section and are particularly abundant in some samples (see list below).

The succession of bedded dolomite and fossiliferous limestone overlying the Yeso Group in the McLeod Hills, southern Caballo Mountains, is assigned to the San Andres Formation. The succession is approximately 22 m thick and composed of massive to bedded dolomite, and bedded (mostly 20-40 cm) dolomitic limestone and limestone. Individual limestone intervals are up to 1.6 m thick. Limestone is fossiliferous, containing echinoderm fragments (mainly crinoids), bryozoans, brachiopods, gastropods and rare corals. A few limestone beds contain chert nodules and silicified fossils. Limestones are composed of wackestone with a very diverse fossil assemblage (including algal wackestone) and bioclastic mudstone with a low diversity fossil assemblage. The microfacies and fossil assemblage indicate deposition in an open marine, shallow shelf environment of dominantly low to moderate water turbulence; grainstones indicate deposition under shallow water, high-energy conditions. Mudstone and wackestone containing vugs filled with calcite, which originally was gypsum or other evaporite minerals, and a less diverse fossil assemblage, indicate deposition in a shallow, restricted environment with increased salinity. Nautiloids and ammonoids indicate late Leonardian age. Up to now, we have not found smaller foraminifers at this locality, although they were important for the correlation with the late Kungurian (= Bolorian = *Misellina* fusulinid biozone).

The complete Artinskian-Kungurian assemblage of the Yeso Group and San Andres Formation is composed of rare cyanobacteria *Ellesmerella* sp. 1; common codiacean and gymnocodiacean algae *Boueina*? sp. 1; *B.*? sp. 2; *Gymnocodium* sp. 1; *G.* sp. 2; undetermined mastoporean? dasyclads; some microproblematica *Tubiphytes* sp. 1 and *Eotuberitina reitlingerae*; relatively common to common smaller foraminifers *Globivalvulina* cf. *apiciformis*; *G.* aff. *kamensis*; *G.* cf. *donbassica*; *G.* ex gr. *graeca*; *Calcivertella*? sp.; *Calcitornella* cf. *elongata*; *C.*? sp. 1; *Ammovertella* sp.; *Palaeonubecularia* sp.; *Pseudospira* aff. *ishimbaica*; *Hemigordiellina* cf. *elegans*; *H.* aff. *elegans*; *H.* cf. *simplex*; *H.* aff. *simplex*; *H.*? cf. *pseudopusilla*; *H.*? aff. *pseudopusilla*; *Hemigordius* cf. *saranensis*; *Orthovertella*? sp. 1; *O.*? sp. 2; *Neodiscus*? ex gr. *ovatus*; *N.*? sp. 2; *Protonodosaria*? sp.; *Nodosinelloides longa*; *N. netjachewi*; Gen. indet.; *Nestellorella*? sp.; *Geinitzina* ex gr. *postcarbonica*; and *Frondicularia* aff. *turae*.

The assemblages of the Yeso and San Andres remain constant during both stages, Artinskian and Kungurian, but we are investigating possible additional biomarkers. The algae appear endemic but are very interesting because they are transitional between several orders. Despite their general endemism, some foraminifers might be used for accurate correlations with the Urals stratotypes: *Globivalvulina* cf. *apiciformis*; *G.* cf. *donbassica*; *Hemigordius* cf. *saranensis*; *Nodosinelloides longa*; and *Geinitzina* ex gr. *postcarbonica*. Nevertheless, it is evident that the strong endemism of the foraminifers and calcareous algae in North America during the Middle and Late Permian begins as soon as the Artinskian-Kungurian.

THE DISSOROPHOID *MILNERERPETON HUBERI* (TEMNOSPONDYLI) FROM THE LATE PENNSYLVANIAN KINNEY BRICK QUARRY IN NEW MEXICO RESTUDIED – PALEONTOLOGY, PALEOENVIRONMENT, AND AGE

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Abstract—The amphibian *Milnererpeton huberi* (Hunt, Lucas and Berman, 1996) was found in the Late Pennsylvanian (Missourian) strata of the Tinajas Member of the Atrasado Formation at the Kinney Brick Quarry, Manzanita Mountains, central New Mexico. This probable branchiosaurid has two known specimens that are characterized by 17 diagnostic features. The unique character set for *Milnererpeton huberi* includes the outline shape of the supratemporal, low dorsal maxillary shelf, smooth ventral surface of the parasphenoid, branchial ossicles with two needle-like points, heavily ossified thoracic neural arches, and presacral count of 22 vertebrae. Larval *Milnererpeton huberi* combine many features of the families Branchiosauridae and Amphibamidae. The same combination of characters is seen in the branchiosaurid *Branchiosaurus salamandroides* and the amphibamid *Platyrhinops fritschi* from the Westphalian D (Moscovian, Middle Pennsylvanian) of Nýrany in Bohemia (Czech Republic). The simplest explanation for this observation is to assume that both larval amphibamids and larval branchiosaurids passed through nearly the same early ontogenetic stages of development. The lateral line system is traceable on the ventral skull roof of *Milnererpeton huberi*. The presence or absence of a well-mineralized (or ossified) notochordal canal may be useful to demonstrate sexual dimorphism. *Milnererpeton huberi* had an interesting combined feeding strategy of plankton feeding, using specialized branchial ossicles, and ate ostracodes. A short overview to the North American “branchiosaurs” is given here. Kinney Brick Quarry is a classic estuarine-lagoonal Konservat-Lagerstätte. *Milnererpeton* was not a marine amphibian, but rather an inhabitant of a freshwater body that was occasionally brackish. The presence of the freshwater/brackish water ostracod *Carbonita* in its intestine/stomach contents supports this contention. The age of the Kinney Brick Quarry is Missourian (early to middle Kasimovian) based on fusulinids and conodonts, which corresponds to Stephanian B of the West European Regional Scale based on the occurrence of the insect zone species *Syscioblatta allegheniensis*.

INTRODUCTION

The Kinney Brick Quarry in the Manzanita Mountains of central New Mexico is a world-famous locality for Late Pennsylvanian fossil plants, invertebrates, and vertebrates developed in Missourian strata of the Tinajas Member of the Atrasado Formation. Four small temnospondyl skeletons are known from this outcrop – in contrast to thousands of other fossil finds.

Berman (1973) described the trimerorhachid *Lafonius lehmanni* as the first tetrapod record from Kinney. He collected a second trimerorhachoid skeleton, which was briefly described as an indeterminate saurerpetontid by Hunt et al. (1992). In the same paper, a well preserved small amphibamid skeleton and an unprepared skeleton of unknown affinity were presented from the Kinney Brick locality. The amphibamid specimen was later described as *Milneria huberi* by Hunt et al. (1996), but the generic name was preoccupied by the recent bivalve *Milneria* (Carditidae). Therefore, Hunt et al. (2002) proposed the replacement name *Milnererpeton* for *Milneria* Hunt, Lucas and Berman, 1996. Werneburg and Lucas (2007) noted that this taxon is the first truly European branchiosaurid from North America. In Schoch and Milner (2008) the species *Milnererpeton huberi* was briefly discussed as a problematic form of possible branchiosaurid affinity, but which lacked critical diagnostic features.

We have subsequently acquired a better understanding of the difficulties in distinguishing between the closely related larval branchiosaurids and amphibamids (Werneburg, 2012), which presents an advantage for understanding the relationship of both clades in their early ontogenetic stages. Amphibamids and branchiosaurids underwent intensive study

and discussion during the last decade (Anderson et al., 2008a, b; Bourget and Anderson, 2011; Carroll, 2004; Clack and Milner, 2010; Fröbisch and Reisz, 2008; Fröbisch and Schoch, 2009a, b; Huttenlocker et al., 2007; Ruta et al., 2003; Schoch and Carroll, 2003; Schoch, 2004; Schoch and Rubidge, 2005; Werneburg et al., 2007; Ruta and Coates, 2007; Schoch and Milner, 2008; Werneburg, 2009, 2012). This great interest is based on the growing consensus that both clades have a close relationship to modern amphibians.

The first skeleton of *Milnererpeton huberi* was restudied in detail during the last few years. In 2005, the “unprepared skeleton of unknown affinity” (Hunt et al., 1992) was prepared by one of us (RW). Both specimens clearly belong to the same species and show many important features, which were not previously described. Thus, the lateral construction of the palate, shape of the otic notch, and structure of the branchial ossicles are now well known. For the first time a reconstruction of the ventral skull roof and palate in ventral view is given. It is also now possible to discuss the questions of sexual dimorphism and feeding strategy of these small, newt-like dissorophoids.

Of great interest is the controversial topic of whether *Milnererpeton* was a marine amphibian. Both skeletons were found with ammonites in a fossiliferous stratum representing an estuarine environment, but this may not have been their true habitat. The intercalation of marine and nonmarine beds in the Kinney Brick Quarry has also allowed a precise age dating using different methods based on marine and terrestrial fossils. A recent overview of the biostratigraphy of the Kinney Quarry will be given here.

Institutional Abbreviations: CM, Carnegie Museum of Natural History, Pittsburgh; FMNH, Field Museum of Natural History, Chi-

cago; MCZ, Museum of Comparative Zoology, Harvard University, Cambridge; NMMNH, New Mexico Museum of Natural History, Albuquerque; SMNS, Staatliches Museum für Naturkunde, Stuttgart; TU BAF, Technische Universität Bergakademie Freiberg; USNM, National Museum of Natural History, Washington; YPM, Peabody Museum, Yale University, New Haven.

OCCURRENCE AND PALEOENVIRONMENT

Strata exposed in the Kinney Brick Quarry provide a unique glimpse of a diverse and well-preserved Late Pennsylvanian lagoonal biota, and the quarry stratigraphy and paleontology was the subject of a symposium volume about 20 years ago (Zidek, 1992). A classic Konservat Lagerstätte, Kinney preserves soft tissues and other delicate structures of plants and animals not well known from correlative deposits (Lucas and Huber, 1991; Kues and Lucas, 1992). Fossils documented from the Kinney Brick Quarry include palynomorphs (Willard, 1992), a diverse, conifer-rich megafloora (Mamay, 1981, 1990; Ash and Tidwell, 1982; Mamay and Mapes, 1992), a shelly marine invertebrate assemblage that includes a few ammonoids but is dominated by brachiopods and the pectinacean bivalve *Dunbarella* (Clark, 1978; Archer and Clark, 1992; Kues, 1992a, b; Mapes and Boardman, 1992), syncarid and hoplocarid crustaceans (Schram and Schram, 1979), eurypterids (Kues, 1985), conchostracans (Kozur et al., 1992), ostracods (Kietzke and Kaesler, 1992), terrestrial arthropods, mostly diplopods and insects (Carpenter, 1970; Shear et al., 1992), conodonts (Krukowski, 1992; Lucas et al., 2011), a diverse assemblage of fishes, mostly acanthodians and palaeoniscoids (Zidek, 1975, 1992b; Gottfried, 1987a, b, 1992; Bardack, 1992; Huber, 1992; Schultze, 1992) and amphibians (Berman, 1973; Hunt et al., 1992, 1996, 2002; Werneburg and Lucas, 2007), as well as coprolites (Hunt, 1992; Hunt et al., 2012), and “fish eggs” (Mamay, 1994). The depositional setting of Kinney is interpreted to be that of an estuary fed by a river delta (Feldman et al., 1992; Lorenz et al., 1992).

Students at the University of New Mexico discovered the rich fossil biota of the Kinney quarry in 1961. During the 1960s, large collections were made of fishes, plants, and other fossils from the quarry, especially by D. Dunkle and S. Mamay of the National Museum of Natural History. Extensive fossil collections from the Kinney Brick Quarry are now housed at the Carnegie Museum of Natural History, National Museum of Natural History, New Mexico Museum of Natural History, University of Kansas Museum of Natural History, and University of New Mexico.

The quarrying operation has exposed about 30 m of the upper part of the Tinajas Member of the Atrasado Formation (Figs. 1-2). The active quarry floor typically is developed on black, platy, and laminar micrite with a high content of black clay containing an invertebrate fauna indicative of abnormal marine conditions. The upper 10 cm of this limestone are finely laminated and grade into the overlying shale. The basal shale exposed at the quarry is 0.4 m thick, olive gray to olive black, well indurated, calcareous, and highly fossiliferous. Part of this interval is highly fissile and not calcareous. Above it is 2 m of olive-gray, calcareous shale that contains thin bands of plastic clay. Individual laminae in these shales are continuous on strike for at least 10 m. Above, and in gradational contact with the underlying shale, are ~ 11 m of olive-gray and greenish-gray, laminar calcareous silty shale that are moderately bioturbated. These strata coarsen upward slightly and are marked by a rapid decrease in floral and faunal diversity near their base. Above them are ~ 13 m of olive-gray, silty shale, and claystone interbedded with lenticular, yellowish-orange and grayish-brown, subarkosic and sublitharenitic, laminated and ripple-laminated sandstone ledges. The section is completed by 1.0+ m of limestone-cobble conglomerate with clasts as much as 8 cm in diameter, and grayish-yellow, trough-crossbedded, subarkosic-sublithic arenite.

Combining physical stratigraphy with the fossil assemblages, we can interpret local sea level trends and identify several distinct depositional environments at the Kinney Brick Quarry (Fig. 1). These make up

a regressive sequence in which limestone grades up through prodelta and deltaic clastics with a capping delta-plain facies.

Thus, the basal micrite (Fig. 1, units 1-2) represents deposition in a nearshore marine environment that received some input of freshwater and clastic sediments. Note its lithology (especially the high black-clay content of this micrite) and unusual fauna (some stenohaline brachiopods and other groups, but dominated by euryhaline taxa, such as the inarticulate brachiopod *Lingula* and bivalves *Myalina* and *Solemya*). The black-clay content, terrestrial plant debris, and euryhaline elements of the fauna (especially abundant *Lingula*) are consistent with deposition near the shoreline with a significant freshwater input. The overlying highly fossiliferous shales (Fig. 1, units 3-4) were deposited in a calm lagoonal or estuarine environment with a significant fresh-water input. Uniform, fine grain size, fine lamination and lack of bioturbation, dark colors, and preservation of soft-bodied forms suggest deposition in quiet, oxygen-poor waters with restricted circulation. A lowland, pteridosperm-dominated flora and freshwater faunal elements (especially) suggest low salinity.

Overlying shales (Fig. 1, units 5-7) represent a similar facies, but probably with a greater fresh-water influence. Dominant elements are *Dunbarella*, an euryhaline bivalve, and terrestrial plants.

Overlying silty shales (Fig. 1, units 8-9) are interpreted by us to represent increased sedimentation rates in a prodelta estuarine environment, brought about by the onset of a significant fluvial discharge. A conifer-dominated floral assemblage abounds here, with sparse, small *Dunbarella*. Most of the relatively xerophytic floral assemblage may have been floated a short distance into an environment characterized by frequent shallow ponding and deposition on a surface better drained than the underlying shales.

Overlying laminated and ripple-laminated sandstone ledges and intercalated shales and claystones (Fig. 1, units 10-14) are interpreted by us as delta front, distributary mouth bars, and associated deposits. Unit 15 is a shale that shows marine influence indicated by the presence of *Lingula* and *Myalina*. This unit, and the overlying fluvial sandstone/conglomerate, may be the base of another transgressive sequence.

We thus interpret the stratigraphic sequence at the Kinney Brick Quarry as mostly reflecting a marine regression corresponding to the progradation of a clastic delta (Fig. 1). Shifting sediments from the delta probably isolated an embayment from normal marine conditions as a clastic wedge developed and extended seaward. Lagoonal conditions were established, and clastic input was initially restricted to clay-size particles. Eventually, the embayment was filled by silty shales from an advancing delta plain on which sand was later deposited. The onset of a subsequent transgression is documented by the highest strata in the quarry section.

DESCRIPTION AND COMPARISONS

The two amphibian skeletons from the Kinney Quarry in the NMMNH collection clearly belong to one species. Both specimens are of the same size and nearly the same proportions of the skull and skeleton (Table 1). They have the same scale morphology, short, relatively thick ribs, well-ossified neural arches, and, most importantly the same short and wide postorbital skull roof. On the other side a multitude of features complement one another.

Werneburg (2012) carefully compared the morphologies of the larval branchiosaurid *Branchiosaurus salamandroides* and amphibamid *Platyrrhinops frittschi* from the Westphalian D (Moscovian, Middle Pennsylvanian) of Nýrany in Bohemia (Czech Republic), but was unable to recognize definitive characters at this early stage of their development that could be used to distinguish between their families. Werneburg (2012) supported the theory that the branchiosaurids could be a clade of paedomorphic amphibamids (Schoch and Milner, 2008), but it is still too early for taxonomic conclusions at this stage of discussion. With the redescription of *Milnererpeton* it is necessary to restudy the small dissorhoids from the Stephanian B of Commeny and the latest Stephanian or earli-

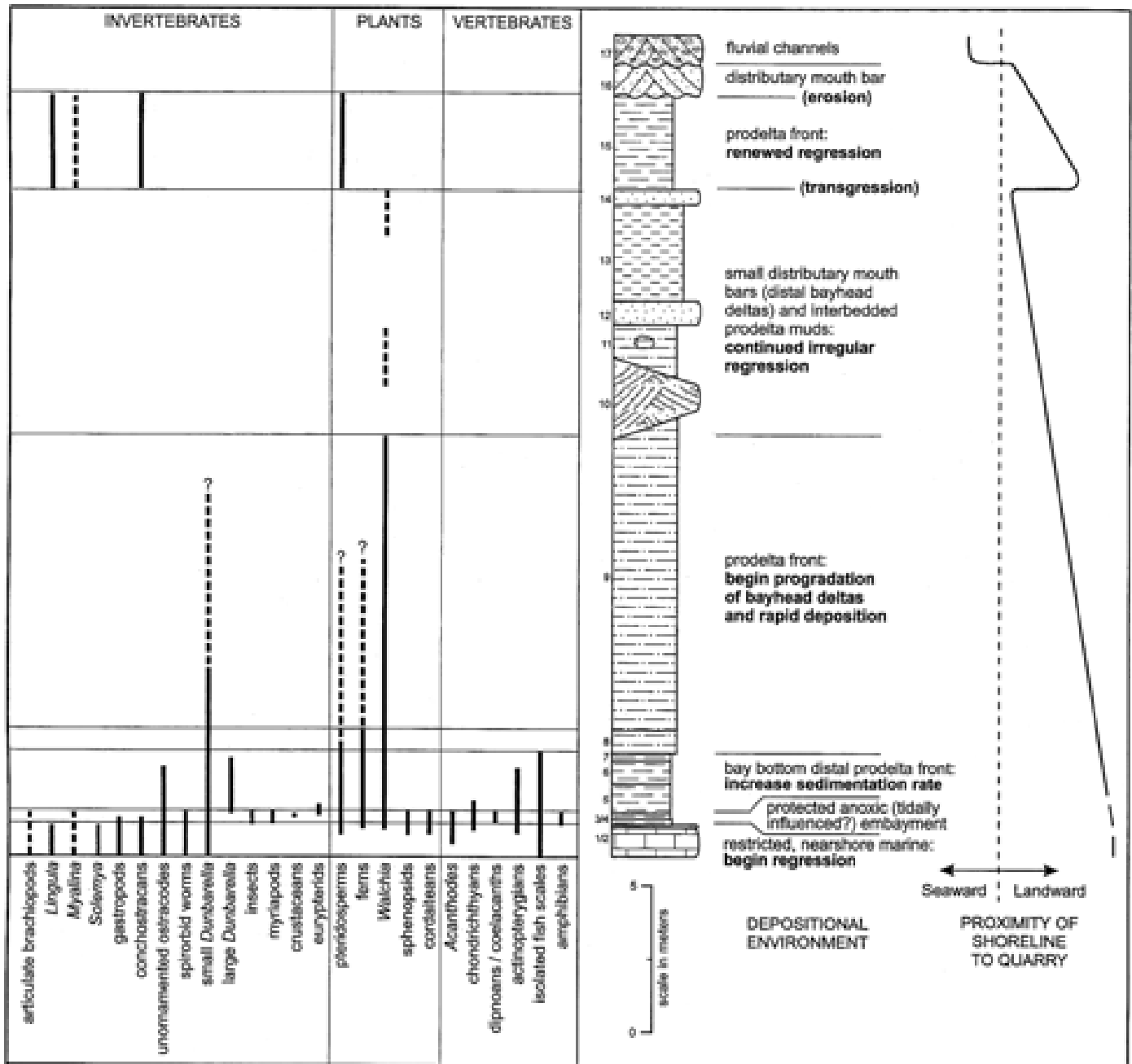


FIGURE 1. The section of the Late Pennsylvanian Kinney Brick Quarry.

est Permian of Montceau-les-Mines, both in France, in order to compare the large and partly metamorphosed branchiosaurids with the amphibamids.

Until that is done, we have decided to assign the small dissorophoid *Milnererpeton* to the family Branchiosauridae on the basis of some new features (especially the shape of the branchial denticle with two needles).

Order TEMNOSPONDYLI Zittel, 1888

Superfamily DISSOROPHOIDEA Bolt, 1969

Family ?BRANCHIOSAURIDAE Fritsch, 1879

Genus MILNERERPETON Hunt, Lucas and Berman, 2002

***MILNERERPETON HUBERI* (Hunt, Lucas and Berman, 1996)**

Figs. 3-12

1989 Trimerorhachid: Hunt et al., p. 26A.

1992 Amphibamid: Hunt et al., p. 211, figs. 1, 2, 3B, D.

1996 *Milneria huberi*: Hunt et al., p. 555-565, figs. 1-3.

2002 *Milnererpeton huberi*, Hunt et al., p. 125.

2007 "*Milnererpeton*" *huberi*, Werneburg and Lucas, p. 164A.

2008 *Milnererpeton huberi*, Schoch and Milner, p. 14.

Holotype: Specimen NMMNH P-3696a, b; compare figs. 3B, C, 4-8, 12B, D-F, I.

Type Locality: Kinney Brick Quarry in the Manzanita Mountains of central New Mexico, USA.

Type Horizon: Units 3-4, Missourian strata of the Tinajas Member of the Atrasado Formation, Kasimovian, Late Pennsylvanian (Fig. 1).

Referred Specimen: Specimen NMMNH P-15019; compare Figs. 3A, 9-11, 12A, C, G-H.

Diagnosis: For an early ontogenetic stage with a skull length of about 6.5 mm.



FIGURE 2. **Left**, overview of the Kinney Quarry section (2005). **Right**, fossil bearing units 1-5; amphibians come from unit 3/4 (2005).

Characters differing from most or next related branchiosaurid genera and species:

- (01) **a:** Supratemporal more triangular with pointed anterior end;
b: Supratemporal anterolaterally constricted due to the median expanded squamosal;
c: Supratemporal posterolaterally constricted due to the enlarged otic notch;
 (all three features shared only with *Schoenfelderpeton prescheri*, but differ from all other branchiosaurids).
- (02) Relatively short distance of otic notch to orbit (shared only with a few branchiosaurids such as *Branchiosaurus salamandroides*, *S. prescheri* and *Melanerpeton* (?*Apateon gracile*, but differs from all others)
- (03) Wide contact of postorbital and supratemporal (differs only from *S. prescheri* and “*Leptorophus*” *tener*)
- (04) Contact (very probable) of post- and prefrontal (shared only with *B. salamandroides*, *Apateon dracyiensis* and ?*S. prescheri*)
- (05) Dorsally directed maxillary shelf relatively shallow (differs from *B. salamandroides*, *A. dracyiensis*, *S. prescheri*, and *A. pedestris*, but shared with many other branchiosaurids)
- (06) Premaxilla bears 9 teeth, maxilla bears 25-30 teeth (differ from *B. salamandroides*, *S. prescheri*, *A. pedestris*, *M. humbergense* and others)
- (07) Ventral basal plate and cultriform process of parasphenoid are smooth (differs only from *S. prescheri* and *M. humbergense*)

TABLE 1. Biometric characters of *Milnererpeton huberi* (Hunt, Lucas and Berman, 1996). **Abbreviations:** H_p , postorbital midline length of skull from level of posterior margins of orbits; H_w , postorbital width of skull between lateral margins of supratemporals; IO_w , minimum interorbital width; S_p , midline skull length; $feml$, maximum femur length; hum_p , maximum humerus length; icl_p , maximum midline interclavicle length; il_p , maximum dorsoventral length of ilium; ra_p , maximum radius length; ti_p , maximum tibia length.

Specimen (Inv.-no.)	S_p in mm	IO_w/S_p	H_w/S_p	H_p/S_p	hum_p/S_p	ra_p/hum_p	$ti_p/feml_p$	icl_p/S_p	$il_p/feml_p$
NMMNH P-15019	6.5	0.28	0.91	0.30	0.49	0.56	0.56	-	0.62
NMMNH P-3696	ca 6.5	ca 0.30	ca 0.91	ca 0.30	ca 0.52	ca 0.50	ca 0.56	ca 0.40	ca 0.41

- (08) Hyobranchial skeleton belongs to eisfeldi-morphotype (shared only with *M. eisfeldi*, *M. arnhardti*, *M. (?A.) gracile* and ?*M. humbergense*)
- (09) Branchial ossicles with round basal plate and up to two needle-like processes (shared only with *B. salamandroides*; most Lower Permian branchiosaurids have brush-like multi-ended branchial denticles)
- (10) Very large and heavily ossified thoracic neural arches (shared only with *B. salamandroides*)
- (11) Presacral count of vertebrae is 22 (shared only with *B. salamandroides* and *M. (?A.) gracile*; most other Lower Permian branchiosaurids have 19-21)

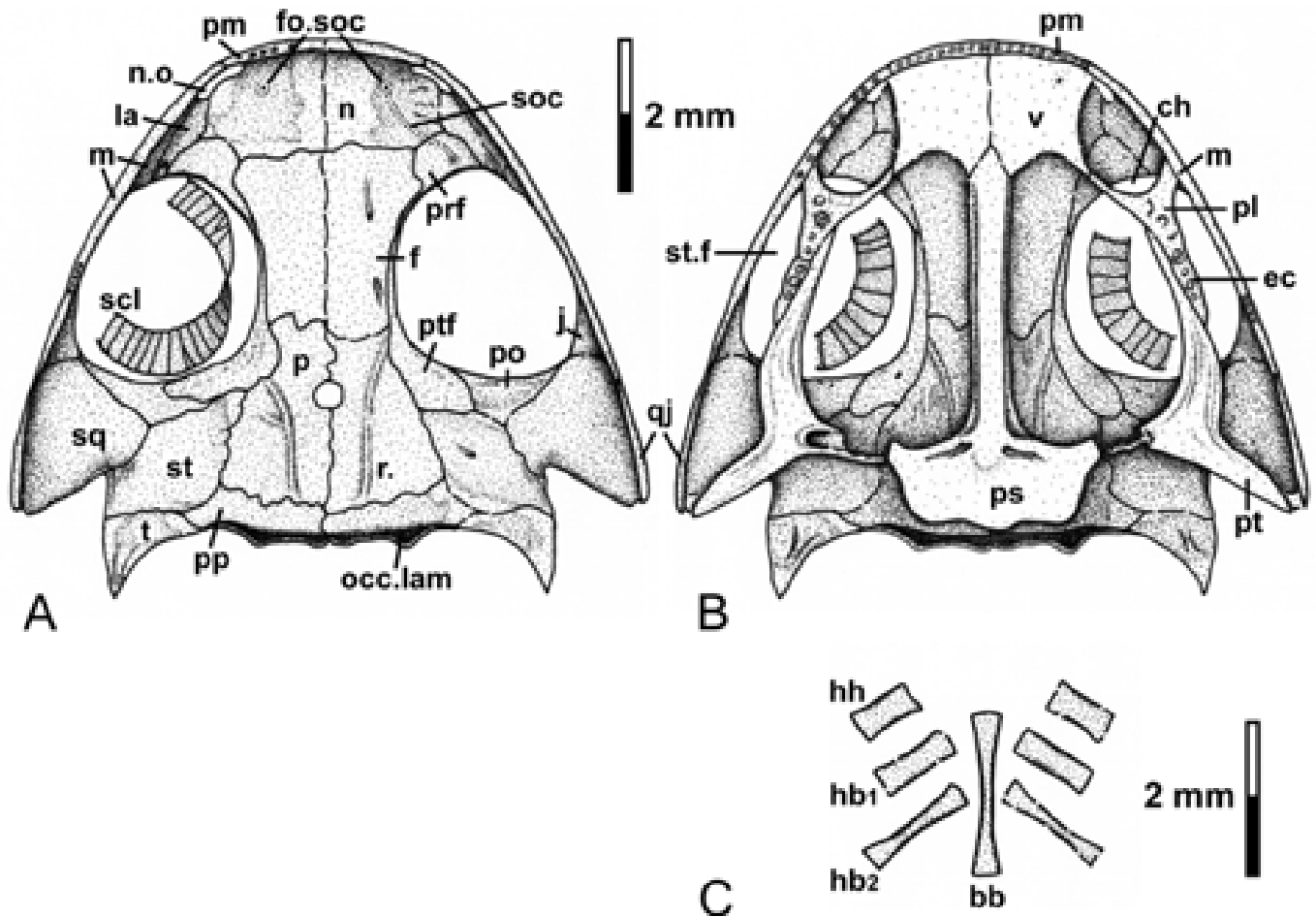


FIGURE 3. *Milnererpeton huberi* (Hunt, Lucas and Berman, 1996), reconstructions. **A**, Skull roof in ventral view, NMMNH P-15019. **B**, Skull roof with palate in ventral view, holotype NMMNH P-3696. **C**, Hyobranchial skeleton, NMMNH P-3696. **Abbreviations:** **bb**, basibranchial; **ch**, choane; **ec**, ectopterygoid; **f**, frontal; **fo.soc**, supraoccipital foramen; **hb1, 2**, hypobranchials 1 and 2; **hh**, hypohyal; **j**, jugal; **la**, lacrimal; **m**, maxilla; **n**, nasal; **n.o**, narial opening (naris); **occ.lam**, occipital rim (flanche); **p**, parietal; **pl**, palatine; **pm**, premaxilla; **po**, postorbital; **pp**, postparietal; **ptf**, postfrontal; **prf**, prefrontal; **ps**, parasphenoid; **pt**, pterygoid; **qj**, quadratojugal; **r**, “parietal ridge;” **scl**, sclerotical ring; **soc**, supraoccipital canal; **sq**, squamosal; **st**, supratemporal; **st.f**, subtemporal fenestra; **t**, tabular; **v**, vomer.

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| <p>(12) Metacarpal 2 and next proximal phalanx of digit 2 are more elongated than metacarpal 3 and the following proximal phalanx of digit 3
(shared only with <i>A. dracyiensis</i>, <i>A. kontheri</i> and <i>Tungussogyrinus bergi</i>)</p> <p>(13) Interclavícula elongated
(shared with <i>B. salamandroides</i>, <i>S. prescheri</i>, “<i>L.</i>” <i>tener</i> and some others, but differs from most others)</p> <p>(14) Ilium with relatively slender, posteriorly directed dorsal process
(shared only with <i>B. salamandroides</i>, but differs from Lower Permian branchiosaurids with a more massive and straight dorsal process)</p> <p>(15) Ventral scales heavy ossified
(shared only with <i>B. salamandroides</i> and <i>B. fayoli</i>, but differs from Lower Permian branchiosaurids with only slightly ossified scales).</p> <p>Characters differing from most or next related amphibamid genera and species:</p> <p>(01) a: Supratemporal more triangular with pointed end;
b: Supratemporal anterolaterally constricted due to median expanded squamosal (both features differ from most</p> | <p>amphibamids)</p> <p>c: Supratemporal posterolaterally constricted due to the enlarged otic notch
(shared with <i>Amphibamus</i> and <i>Platyrhinops</i> in adult stage, but differs in their larval stage, and differs from other genera)</p> <p>(04) Contact (very probable) of post- and prefrontal
(shared with the Late Pennsylvanian genera <i>Amphibamus</i> and <i>Platyrhinops</i>, but differs from the Permian amphibamids)</p> <p>(05) Maxillary shelf relatively slender
(differs from <i>Platyrhinops</i>, but shared with <i>Amphibamus</i> and others)</p> <p>(06) Premaxilla bears 9 teeth; maxilla bears 25-30 teeth
(differs from <i>Platyrhinops</i>, <i>Amphibamus</i> and others)</p> <p>(07) Basal plate and cultriform process of ventral parasphenoid are smooth
(differs from all larval and adult amphibamids)</p> <p>(08) Hyobranchial skeleton belongs to eisfeldi-morphotype
(differs from <i>Platyrhinops fritschi</i>, but not recorded in other amphibamids)</p> <p>(09) Branchial ossicles with round basal plate and up to two needle-</p> |
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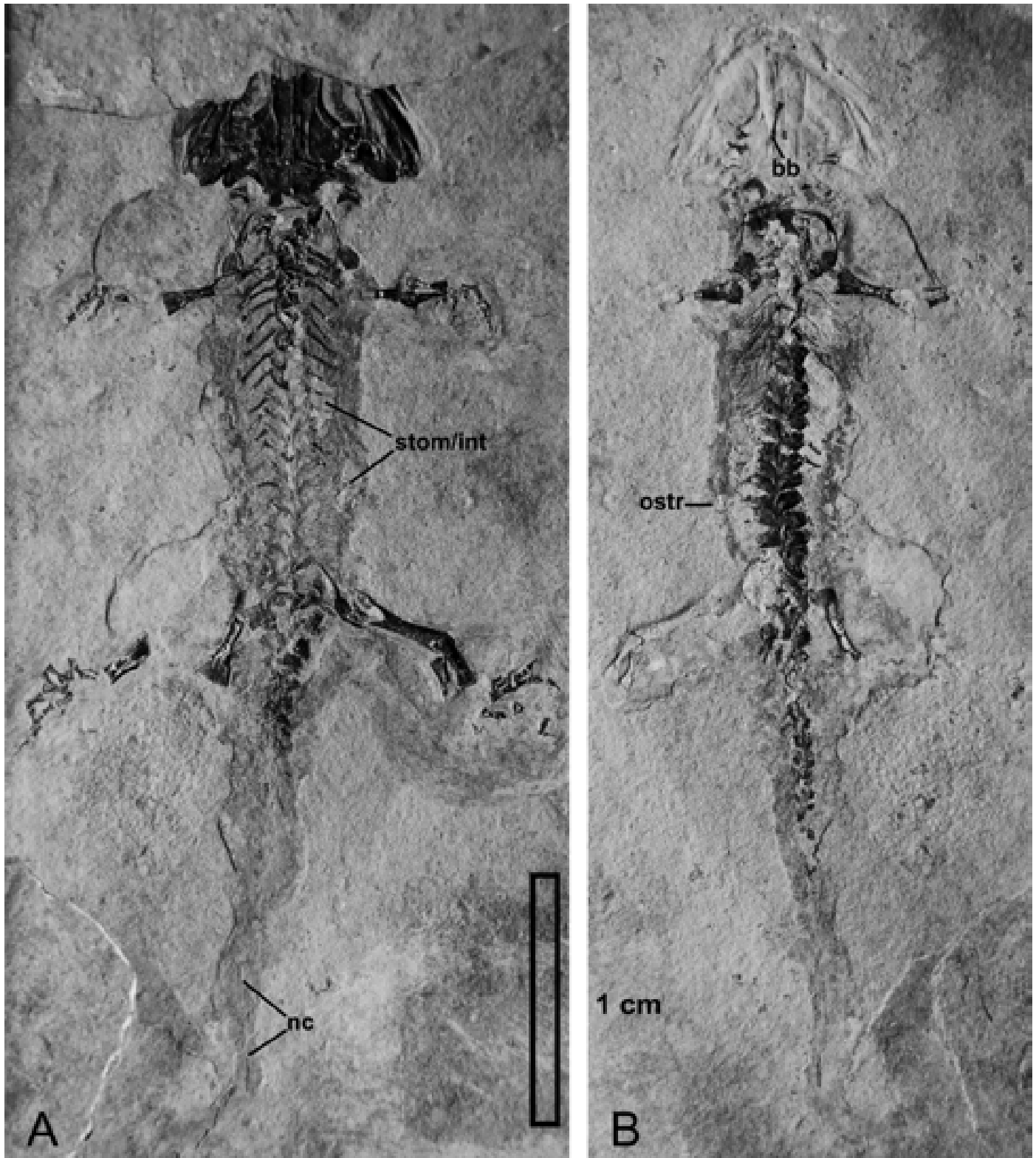


FIGURE 4. *Milnererpeton huberi*, skeleton. **A**, Holotype NMMNH P-3696a. **B**, Counterpart of the holotype NMMNH P-3696b, with hyobranchial skeleton and neural arches. **Abbreviations:** **bb**, basibranchial; **nc**, notochord; **ostr**, ostracod shells; **stom/int**, stomach or intestine.

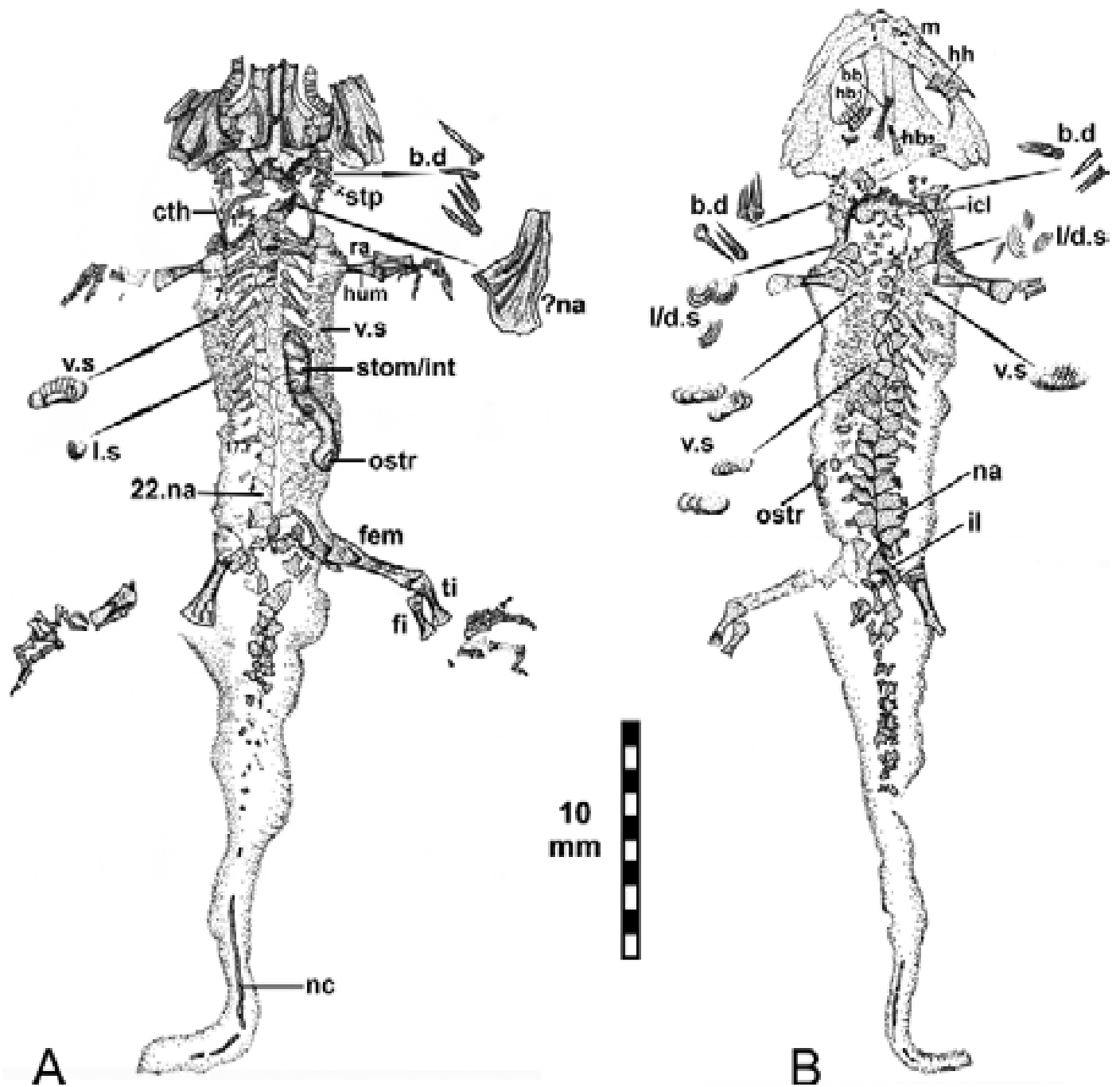


FIGURE 5. *Milnererpeton huberi*, skeleton. **A**, Holotype NMMNH P-3696a. **B**, Counterpart of the holotype NMMNH P-3696b, with hyobranchial skeleton and neural arches. **Abbreviations:** **bb**, basibranchial; **b.d**, branchial denticles; **cth**, cleithrum; **d.s**, dorsal scales; **fem**, femur; **fi**, fibula; **hb1, 2**, hypobranchials 1 and 2; **hh**, hypohyal; **hum**, humerus; **icl**, interclavicle; **il**, ilium; **l.s**, lateral scales; **m**, maxilla; **na**, neural arch; **nc**, notochord; **ostr**, ostracodes; **ra**, radius; **stp**, stapes; **stom/int**, stomach or intestine; **ti**, tibia; **v.s**, ventral scales.

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| | like processes
(differs from all amphibamids) | (shared only with <i>P. fritschi</i> ; differs from all other amphibamids) |
| (10) | Very large and heavily ossified thoracic neural arches
(differs from <i>Amphibamus grandiceps</i> ; similar in <i>P. fritschi</i>) | (15) Ventral scales heavy ossified
(only shared with <i>A. grandiceps</i> and <i>P. fritschi</i> ; differs from <i>?P. lyelli</i> and all Permian amphibamids) |
| (11) | Presacral count of vertebrae is 22
(differs from all amphibamids) | (16) Monocuspid marginal teeth
(differs from <i>Amphibamus</i> , <i>Platyrhinops</i> , <i>Doleserpeton</i> and <i>Tersomius</i> ; shared with other amphibamids) |
| (12) | Metacarpal 2 and next proximal phalanx of digit 2 are more elongated than metacarpal 3 and the following proximal phalanx of digit 3 | (17) Ventral basal plate of parasphenoid with foramina and transverse grooves for internal carotid arteries |

(shared only with *P. fritschi*; differs from other known amphibamids in this early stage, if known).

The combination of the features 1, 5, 7, and 9-11 may be a unique character set for *Milnererpeton huberi*.

Description

The two skeletons have the same skull length of about 6.5 mm. Even in comparison with the other branchiosaurids, this size is not large. The exoccipital, the neural arches, and the hyobranchial skeleton are well ossified. Therefore, both specimens of *Milnererpeton huberi* were most probably neotenic.

The following description of individual bones, mostly in ventral aspect, is given in relation to the most comparable branchiosaurids and amphibamids. Small differences between ventral and dorsal shapes of the skull roof bones exist.

Skull roof (Figs. 3A-B, 6, 11)

Both skulls are exposed in ventral view, and therefore, the dorsal sculpture pattern is unknown. The ventral surface of the skull roof is smooth, except for ridges, furrows and tiny foramina. The curved parietal ridges are most prominent and marked the lateral margin of the orbitotemporal region of the endocranium. This crest is named “Parietalleiste” (Pffannenstiel, 1932) or crista orbitotemporalis (Shishkin, 1973). Large foramina with a short furrow are traceable on the supratemporal, frontal, and parietal, but their occurrences are variable among specimens. Some small foramina are present on the postorbitals and postfrontals.

Especially interesting is the curved furrow or open canal on the inner surface of the dorsoventrally directed occipital rim (Fig. 11A, E-F), which may belong to the lateral line system. This clear record of the supraoccipital line on the ventral surface of the skull roof is surprising (Figs. 3A, 11A, C, E-F). It is a curved internal canal in the transparent bone of the left nasal and the left prefrontal exiting through one anterior (nasal) and one posterior (prefrontal) foramen. Such an internal canal of the lateral line system is well observable in fossil fish skulls, e. g., of the Triassic *Saurichthys*. On the ventral surface of the right nasal only the anterior foramen is preserved. It may be that the furrows on the frontals are also remnants of the supraoccipital sulcus.

The skull in general has a short, wide preorbital as well as a foreshortened postorbital region. The orbitae are large in this early stage. The sclerotic ring is well ossified. In one case this ring is well recorded (Fig. 11A) to give the possibility for restoration in 3D-view (Fig. 3A) with a wider ventral and a more closely dorsal caliber to support the eye ball (Fig. 3A). The ventral part of the sclerotic ring elements is as wide as the dorsal part, and ventrally few triangles are intercalated (Fig. 6C).

The nasals appear to be narrow, but the anteriorly widened frontals clearly indicate a wide contact with the nasals (Figs. 3A, 11). The nasal is much wider than the frontal, as is known from branchiosaurids and the Pennsylvanian amphibamid *Platyrhinops*. A short, blunt snout region is normal for such small specimens of *Milnererpeton*. The presence of a foramen and internal canal of the supraoccipital line on the left nasal, prefrontal, and frontal is discussed above. The length of the frontal is equal to that of the parietal. The frontals are obviously widened anteriorly, as in the Late Pennsylvanian amphibamids *Platyrhinops lyelli* of Linton and *P. fritschi* of Nýrány. The same feature is sometimes recorded in the trematopid *Mordex laticeps* and the micromelerpetonid *Limnogrirus elegans*, both of Nýrány (Werneburg, 2012, p. 16). The interorbital region is relatively wide ($IO_w/S_1 = 0.28-30$), as in *Branchiosaurus salamandroides* and some other branchiosaurids and amphibamids. Because of the anteriorly widened frontals the dorsal, orbital contribution of the prefrontal is reduced to a splint-like process, which contacts a complementary process of the postfrontal. The narrow posterior process of the prefrontal is characteristic of amphibamids (Schoch and Milner, 2008), but the same configuration is known in the

Lower Permian branchiosaurid *Apateon dracyiensis* with anteriorly widened frontals, and a splint-like posterior process of the prefrontal that clearly contacts that of the postfrontal (Werneburg, 2001, fig. 6). The differences in this feature complex are a result of the intraspecific variability of the listed disorophoids and thus are not just a feature of amphibamids. The prefrontal-postfrontal contact is not definite but is considered very probable in view of the extremely elongated anterior process of the postfrontal (Figs. 3A, 11A). A large contact between the postorbital and supratemporal is present. The parietal foramen is large, but in general relatively variable in size and position. The middle part of the parietal narrows anteriorly, allowing a wedge-shaped extension of the supratemporal to occupy the area between it and the postfrontal. Therefore, the supratemporal has more of a triangular instead of the more common trapezoidal outline. The lateral margin of the supratemporal is constricted two times. The anterior constriction of the supratemporal below the postorbital allowed the median expansion of the squamosal (Figs. 3A, 11A). The shorter posterior constriction of the lateral supratemporal is the natural margin of the enlarged otic notch for the expanded middle ear (Figs. 3A-B, 11A, D-F). This “supratympanic shelf” is characteristic of many amphibamids (but not all).

The shape of the supratemporal is identical to that of the Lower Permian branchiosaurid *Schoenfelderpeton prescheri* (Boy, 1986). Therefore, these features are not unique for one of the families, branchiosaurids or amphibamids. Unfortunately, this feature of the constricted supratemporal of *Schoenfelderpeton* is not coded in the character matrix for the phylogenetic analysis of the Branchiosauridae by Schoch and Milner (2008, table 1, fig. 5). In all other branchiosaurids, we do not find this feature.

The tabular has a large, posteriorly-directed tabular horn, as in *Branchiosaurus* and some *Apateon* species, and most recently described in the amphibamid *Platyrhinops fritschi* and the trematopid *Mordex laticeps* (Werneburg, 2012). Thus, a large tabular horn is not diagnostic only for branchiosaurids. On the ventral surface is a pair of ridges that entered posterior to the tip of the horn. The postparietals are relatively narrow anteroposteriorly, as are their occipital flanges or laminae, which are elongated in a dorsoventral direction. This is not a unique feature of *Milnererpeton huberi* as assumed by Hunt et al. (1992, 1996). A wide occipital rim is present also in small specimens of *Apateon dracyiensis* (Werneburg 2001, figs. 3a, 9a, 10a, b; skull length about 7-9 mm), *Apateon caducus* (Schoch, 1992, fig. 11) and *Platyrhinops fritschi* (Werneburg, 2012, figs. 9-10 and the adult shape in fig. 13c).

The squamosal and the concave rim of the lateral margin of the supratemporal surround the enlarged otic notch. The medial portion of the squamosal is widely expanded, but narrows to a short anteroposterior margin. Therefore, the distance between the otic notch and the orbit is relatively short, as is characteristic of most amphibamids and a few branchiosaurids. The quadratojugal is relatively narrow and elongated, possibly entering anteriorly to contact the maxilla (Figs. 3A-B, 6A-B). The quadratojugal is narrow, as is known in *Branchiosaurus salamandroides* (Werneburg, 2012, fig. 3C), but may become wider during ontogenetic development. In the Lower Permian branchiosaurids the quadratojugal is unossified in its earliest stages of development, but then becomes relatively wide in later stages of development. The quadrate condyle position is well anterior to the occipital articulation in this stage. The jugal has a narrow, short, anterior part, whereas the posterior part is incompletely ossified (Fig. 11A, E-F). The narrow lacrimal extends between the orbit and the naris (Figs. 3A, 11A, E-F). The naris is narrow and kidney-shaped because of the anterolateral extension of the nasal (Fig. 3A). The premaxilla outline is not well preserved. It bears 9 teeth. A midline fenestra of the premaxillae is uncertain. The maxilla is very elongated (Figs. 6A-B, 11A), as in *B. salamandroides*, and relatively slender with no broadened maxillary shelf, in contrast to *B. salamandroides* or *Platyrhinops fritschi* (Werneburg, 2012, figs. 3c, 4a, 6d, 11b, e). The maxilla bears about 25-30 teeth, which are monocuspid and not pedicellate (Fig. 11B).

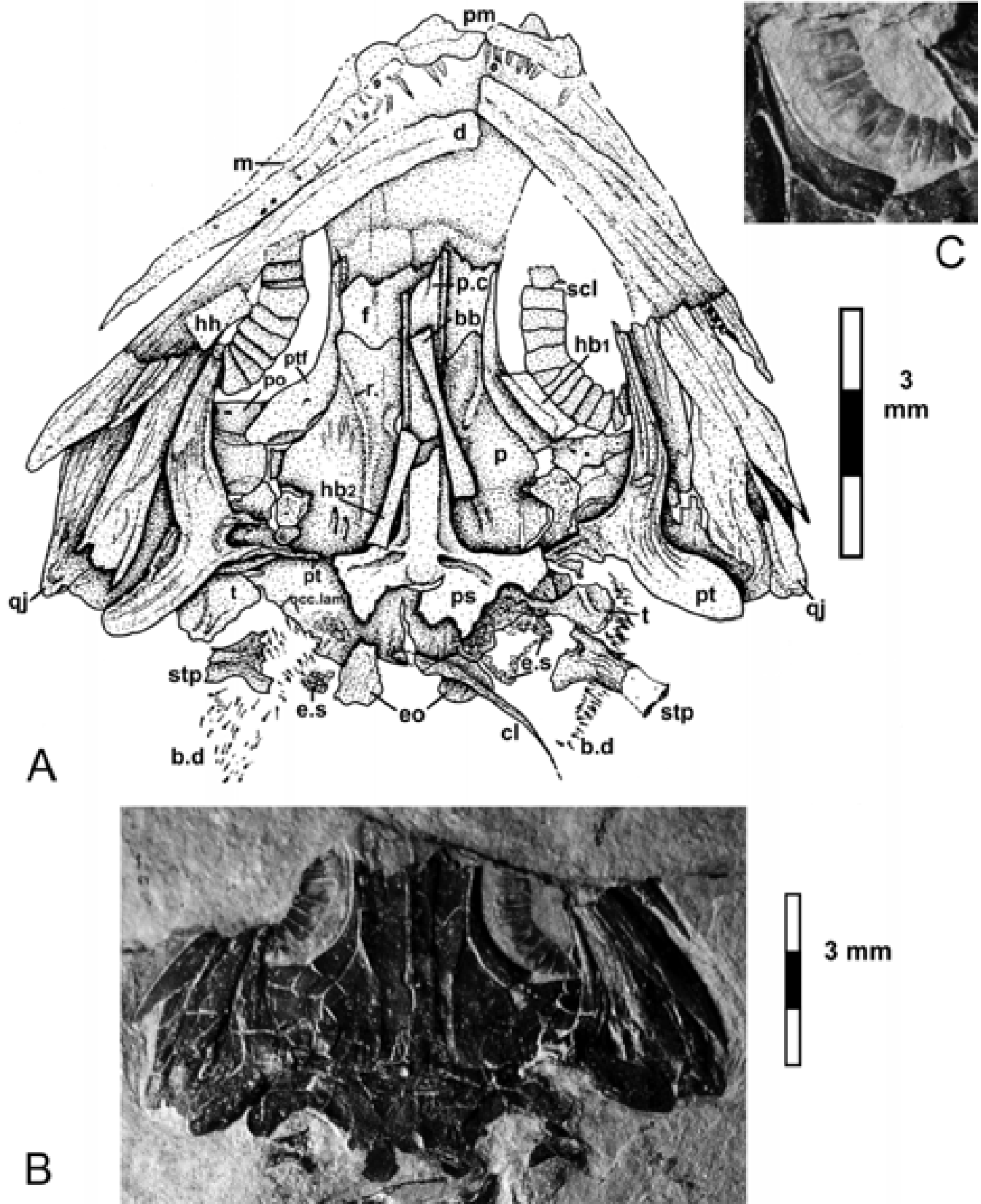


FIGURE 6. *Milnererpeton huberi*, skull of holotype NMMNH P-3696a; A, composite skull drawing from NMMNH P-3696a, B, skull of NMMNH P-3696a, C, orbit with sclerotic ring of NMMNH P-3696a. **Abbreviations:** bb, basibranchial; b.d, branchial denticles; cl, clavicle; d, dental; eo, exoccipital; e.s, endolymphatic sac; f, frontal; hb1, 2, hypobranchials 1 and 2; hh, hypohyal; m, maxilla; occ.lam, occipital rim (flanche); p, parietal; p.c, cultriform process of parasphenoid; pm, premaxilla; po, postorbital; ptf, postfrontal; ps, parasphenoid; pt, pterygoid; qj, quadratojugal; r, "parietal ridge;" scl, sclerotic ring; t, tabular.

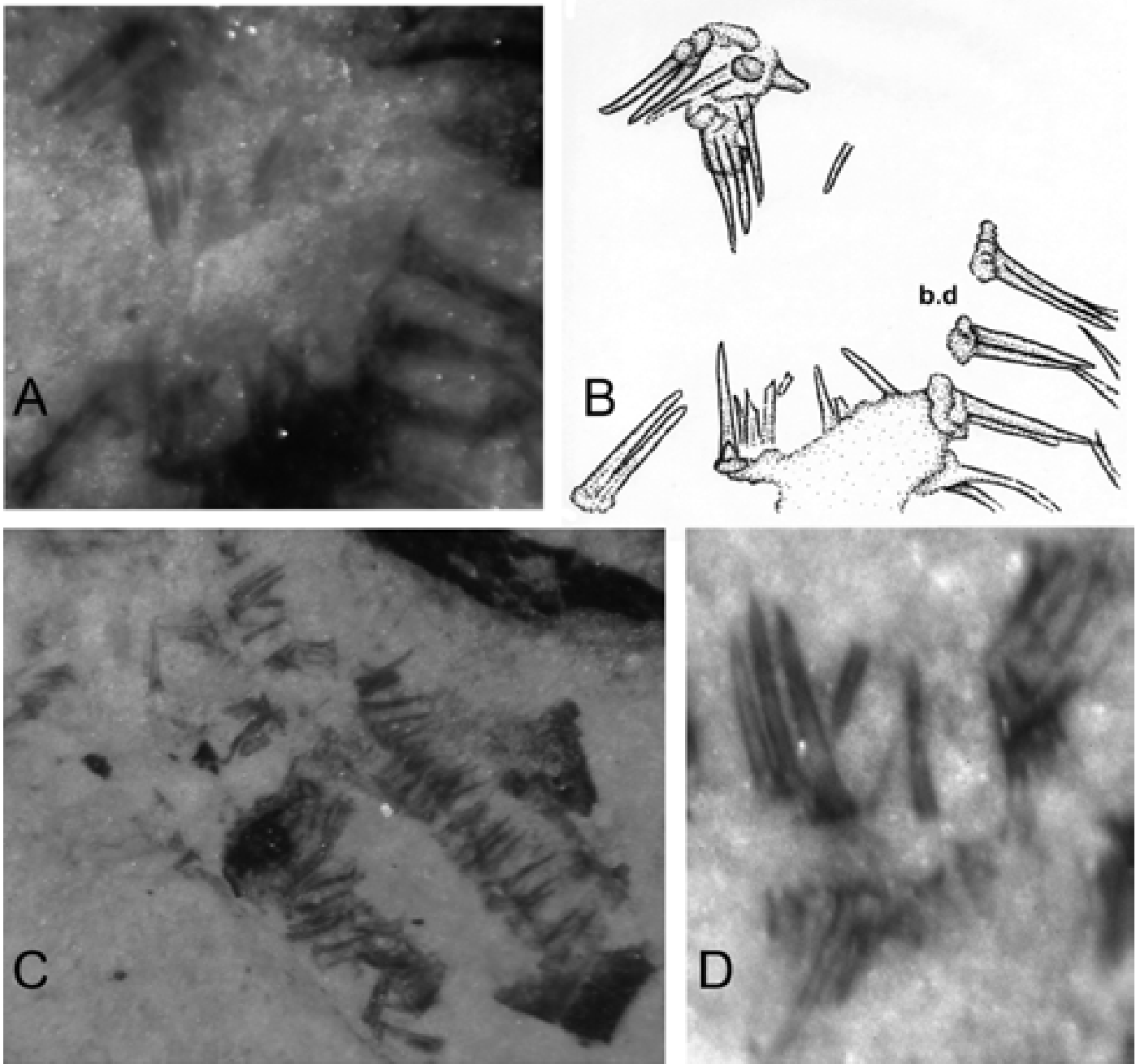


FIGURE 7. *Milnererpeton huberi*. A-D, branchial denticles with one and two spike-like process(es) are about 0.15 mm in length; holotype NMMNH P-3696a, b. **Abbreviation:** b.d, branchial denticles.

Palate (Figs. 3B, 6A-B, 11A, E-F)

The narrow cultriform process of the parasphenoid has a stepped, narrow lateral margin. It has a right angle union with the basal plate, whose width greatly exceeds the length. The plate bears well-developed foramina and transverse grooves for the internal carotid arteries, as well as pronounced basiptyergoid processes. No other foramina exist on the ventral basal plate, in contrast to *B. salamandroides* and *P. fritschi* (Werneburg, 2012, figs. 25d-f, 26d-f). The entire parasphenoid is smooth and bears no indication of sculpturing, which is an important difference from the presence of tubercles or tooth-like denticles exhibited by amphibamids. This is also the case in the smallest known specimen of *Platyrhinops lyelli*, skull length of about 13 mm (Clack and Milner, 2010, fig. 3), *P. fritschi* (Werneburg, 2012, figs. 26d-g; $S_1 = 7-13.9$ mm), and the smallest specimens of *Amphibamus grandiceps* (Milner, 1982, figs. 5b,

d; $S_1 = 5-7.0$ mm). Most branchiosaurids lack sculpturing on the parasphenoid, with two exceptions: *Schoenfelderpeton prescheri* (Boy, 1986) and *Melanerpeton humbergense* (Boy, 1978). Furthermore, Werneburg (2012, figs. 25h-j) has described the new dissorophoid taxon *Nyranerpeton amilneri*, from the Late Pennsylvanian (Westphalian D) of Nýrány, that is related to the micromelerpetontids and in some aspects is comparable with urodeleans, but whose ventral surface of the parasphenoid is also smooth. Therefore, the presence or absence of sculpturing of the parasphenoid is not a good diagnostic feature of the family.

The pterygoid has a medially elongated, mostly cartilaginous basal process, a relatively short quadrate ramus, and a curved, tapering, palatal ramus that extends along the posterior half of the medial margin of the palatine, and, therefore, widely separated from contacting the vomer.

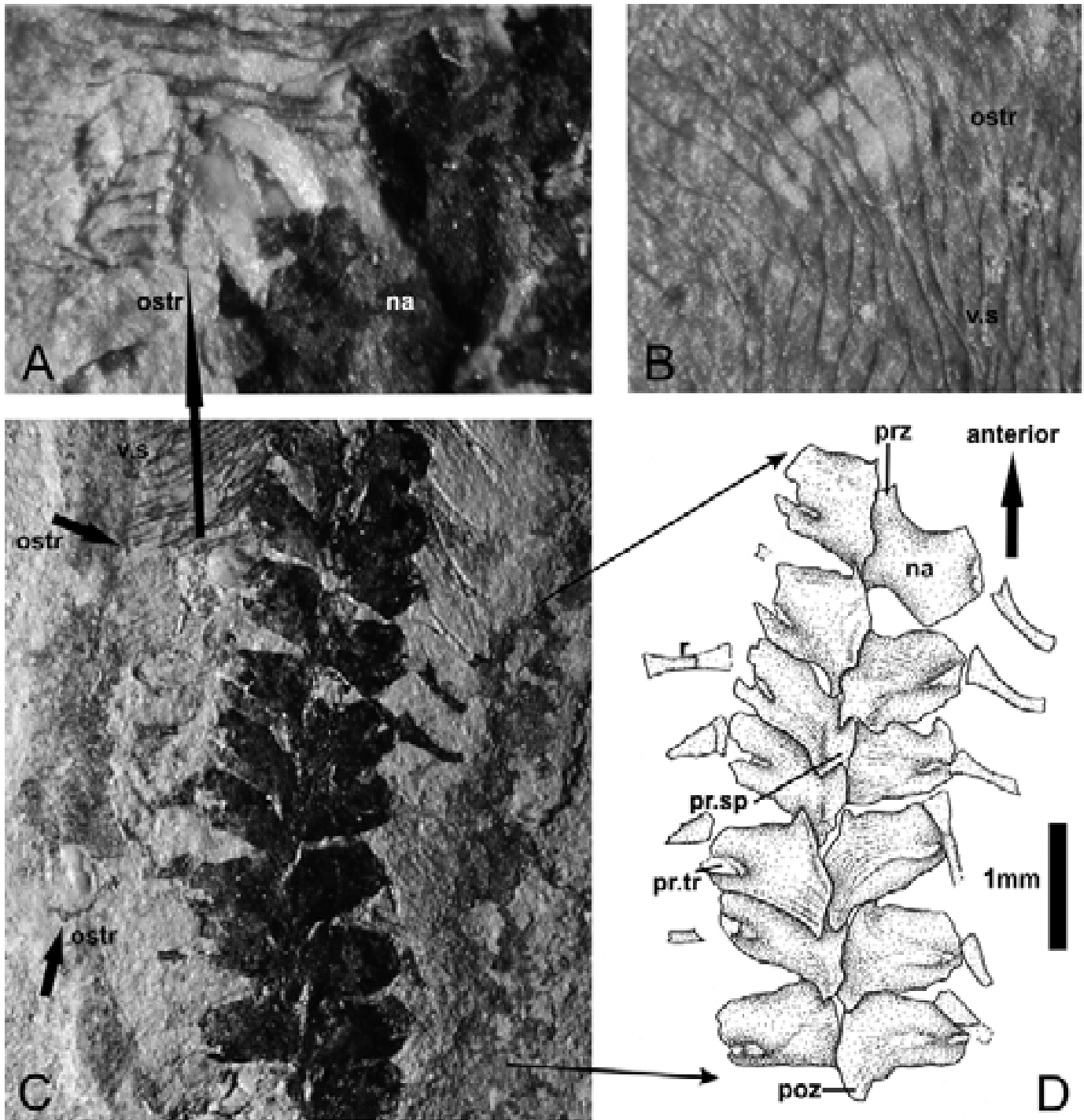


FIGURE 8. *Milnererpeton huberi*, thoracic region of the skeleton from the counterpart of holotype NMMNH P-3696b. **A**, Double-valved ostracode *?Carbonita* sp. interfingering with neural arch. **B**, Ostracode *?Carbonita* sp. covered by ventral scales. **C-D**, Heavily ossified neural arches and the stomach or intestine content of ostracodes *?Carbonita* sp.; ostracodes are about 0.7 mm elongated. **Abbreviations:** na, neural arch; ostr, ostracodes; pr.sp, spinose process; pr.tr, transverse process; r, rib; v.s, ventral scales.

The posterior margin of the vomers is incompletely preserved, but it is possible confidently reconstruct this area on the basis of neighbouring elements of the palate (Fig. 3B). A remnant of the vomerine fang is preserved in the counterpart of the holotype (Figs. 3B, 5B, 6A), which appears to be from near the anteromedial margin of the choane.

The palatine is Y-shaped with a narrow, elongated vomerine ramus. An anterolateral process contacts the maxilla. This feature, according to Schoch and Milner (2008), provides “robust support for the monophyly” of the family Branchiosauridae. As explained by Werneburg

(2012, p. 41-44), a Y-shaped palatine is known in branchiosaurids and also in amphibamids at an early ontogenetic stage. *Platyrhinops fritschi* (Werneburg, 2012, figs. 11b, 12c) and *Amphibamus grandiceps* (Milner, 1982, figs. 5a, d; Schoch, 2001, fig. 4) also exhibit this feature at growth stages in which skull length is from 5 to 14 mm. Consequently, the subtemporal fenestra is greatly elongated anteriorly. Larger amphibamid specimens have a narrow palatine, but in the absence of the lateral maxillary process and its slightly convex lateral margin, it has a greatly expanded contact with the maxilla. The same shape of palatine is known

for larger branchiosaurid specimens, such as the neotenic *Apateon dracyiensis*, with a skull length of 12-15mm (Werneburg, 2001, figs. 5-7), the neotenic *A. caducus* with a skull length of 38 mm (Fröbisch and Schoch, 2009, fig. 1), and *Melanerpeton* (?*Apateon*) *gracile* with a skull length of about 23 mm (Werneburg, 1991, fig. 5b). Therefore, the branchiosaurids and amphibamids have the same ontogenetic development of the palatine up to the stage in which the anterior portion of the subtemporal fenestra is lost. The palatine of *Milnererpeton huberi* bears four teeth; the shorter, narrower ectopterygoid has three.

Visceral ossifications and gills (Figs. 3C, 5, 6A, 7)

The holotypic stapes is well preserved (Figs. 6A-B). Its proximal base is widened and pierced by the large stapedial foramen. The stapedial shaft is stout and elongated. The shape is the same as in *A. dracyiensis* (Werneburg, 2001, fig. 16b) and is similar to that in *P. lyelli* (Clack and Milner, 2010, fig. 5).

The complete hyobranchial skeleton is preserved in the holotypic counterpart (Figs. 3C, 5B, and 6A). The median element is the very narrow basibranchial. The hypobranchial 2 is much more elongated than hypobranchial 1, but the latter element is more robust. The hypohyal is shorter and thicker than the hypobranchials. The proportions of the ossified hyobranchial elements correlate to the so-called “*eisfeldi*-type” in branchiosaurids (Werneburg, 2009, fig. 8). However, ossified hyobranchial skeletons are not only characteristic of neotenic branchiosaurids, as earlier presumed, but are also known in the amphibamid *P. fritschi* and the trematopid *Mordex laticeps* (Werneburg, 2012, figs. 8, 14b).

External gills are not preserved, but numerous branchial denticles are present. Unfortunately, not all denticles do lie arranged in distinctive rows allowing us to estimate their number, but the morphology of the branchial denticles is well preserved (Figs. 3, 6A- B, 7). The basal plate is circular or oval and bears one or two, needle- or spike-like process(es). The same type of branchial gill raker with one and two spikes on the denticles is known in the branchiosaurid *Branchiosaurus salamandroides*, but rarely reported (Werneburg, 2012, figs. 2a, c). Gill rakers with one preserved needle are known in the holotypic specimen of *B. salamandroides*, of *Platyrhinops fritschi*, *Mordex laticeps* (Werneburg, 2012, figs. 3c; 8c, 12b, 15), and in the Permo-Triassic branchiosaurid *Tungussogyrinus bergi* from the Tungus-Basin in Siberia (Werneburg, 2009, fig. 4B). It may be that the gill rakers with two needles is characteristic of branchiosaurids, as multi-denticle gill rakers have been described in many Lower Permian branchiosaurids. However, these have more elongated basal plates with brush-like ends (Boy, 1986, fig. 6; Werneburg, 2008, figs. 13b, 14b). It thus seems that the brush-like denticles represent the more derived type.

The lower jaw (Fig. 11A) is not well enough preserved to describe morphologic details.

Postcranial skeleton (Figs. 4-5, 8-10, 12)

The vertebral column has 22 presacral vertebrae based on the number of preserved ribs (Fig. 6A). This presacral number is similar to that in branchiosaurids (19-22), in contrast to *Platyrhinops fritschi* (23-25) and *Platyrhinops lyelli* (25-26, after Clack and Milner, 2010), but otherwise is nearly identical to that in *Amphibamus grandiceps* (20-21, after Milner, 1982). This feature is important in diagnosing genera and species, but not in differentiating the families Branchiosauridae or Amphibamidae. The paired neural arches are very large and heavily ossified (Figs. 4B, 6B, and 8C-D). The neural arches are preserved as lateral halves flattened into a horizontal plane, so the right and left halves of each arch pair are exposed in lateral view, with the spines facing one another along the midline. The transverse processes are well developed, and the spinose processes are low. These are similar in shape (and number) to the thoracic neural arches in *B. salamandroides* (Werneburg,

2012, figs. 3a, b, 4a). The Lower Permian branchiosaurids do not have such large and robust neural arches. Vertebral centra are not ossified at the ontogenetic stage represented by the *Milnererpeton* specimens.

The ribs are short, straight and have expanded ends. They are very similar to those in branchiosaurids and larval amphibamids. The ribs of the other two trimerorhachoid skeletons from the Kinney Quarry are significantly thinner.

The humerus is an elongated bone half the length of the skull (Table 1). The radius is also elongated, half or more than half of the length of the humerus. The length of the radius is very similar to those of *Branchiosaurus salamandroides*, *Platyrhinops fritschi* and *Mordex laticeps* (Werneburg, 2012, tables 1-3). Metacarpal 2 and the next proximal phalanx of digit 2 are more elongated than metacarpal 3 and the proximal phalanx of digit 3 (Fig. 12B). The elongated digit 2 of the manus is known from some branchiosaurids (*Apateon dracyiensis*, *A. kontheri*, *Tungussogyrinus bergi*; Werneburg, 2009, fig. 12) and amphibamids too (*P. fritschi*; Werneburg, 2012, figs. 33d, f). The phalangeal formula of the manus is 2-2-3-3, with the third digit being the longest.

The femur is only slightly longer than the humerus. The tibia is clearly longer than the half length of the femur (Table 1). The holotypic specimen is informative, demonstrating individual or intraspecific variability in the phalangeal formula of the pes, especially regarding the length of the third digit. The fourth digit of the left pes has three phalanges, with the fourth the longest digit, and the third digit of the right pes has four phalanges, with the third the longest digit (Figs. 12D, E). The phalangeal formula of the pes is 2-2-3/4-4-2.

Remains of the shoulder girdle are present in both skeletons. Only the anterior portion of the interclavicle is preserved (Figs. 6B, 12F), and the preserved left lateral margin of the anterior portion is concave. Therefore, the preserved portion of the interclavicle may represent only the anterior half; the posterior half has been reconstructed (Fig. 12F). This indicated that the interclavicle is elongated and comparable with that of *Branchiosaurus salamandroides*, *Platyrhinops fritschi* (Werneburg, 2012, figs. 28a, c, d, j-k) and micromelerpetontids. The anterior margin of the interclavicle is also feathered. The clavicle has a short, narrow, dorsal blade, and an oval, broadly ventral blade that projects medially (Figs. 10B, 11A). The cleithrum is an elongated, narrow rod. The scapulocoracoid is characteristically half-moon-like (Fig. 10B).

The pelvic girdle is represented by the ilium and ischium. The ventral part of the ilium is greatly expanded to accommodate the acetabulum in ventral view (Fig. 12G). The dorsal process is slightly curved posteriorly, and its dorsal part is greatly expanded. The ilium is comparable to those of the Late Pennsylvanian dissorophoids from Nýrány (Werneburg, 2012, figs. 30-31), whereas the ilium of most Lower Permian branchiosaurids is more massively ossified, and the ventral margin of the dorsal process is straight. The ischium is ossified as a small rectangular plate (Figs. 6, 10A).

The well ossified ventral scales cover completely the ventral thoracic region (Figs. 6, 8B, 10A). Each is much wider than long and has a sculpturing pattern of 3 or 4 concentric rings and straight, parallel marks (Fig. 6). The posterior margin is thickened, which affects the line-pattern of the ventral scalation. The dorsal and the slightly smaller lateral scales are nearly circular in outline, with 2 or 3 concentric rings (Figs. 6, 10A). The dorsal scalation pattern of the trunk is not so dense as on the ventral side.

The tail of *Milnererpeton huberi* is very elongated, with a length equal to the presacral body length. The notochord is well preserved in the distal half of the tail of the holotype specimen (Figs. 4-5). The mineralized or ossified notochord consists of narrow sticks arranged in a single row. This may be a sexually dimorphic feature (see below).

Surprisingly, the bleached stomach or intestine clearly containing few ostracod shells in both specimens (Figs. 4, 6, 9A, and 10A). The nature of these ostracods and the feeding strategy of *Milnererpeton* are discussed below.

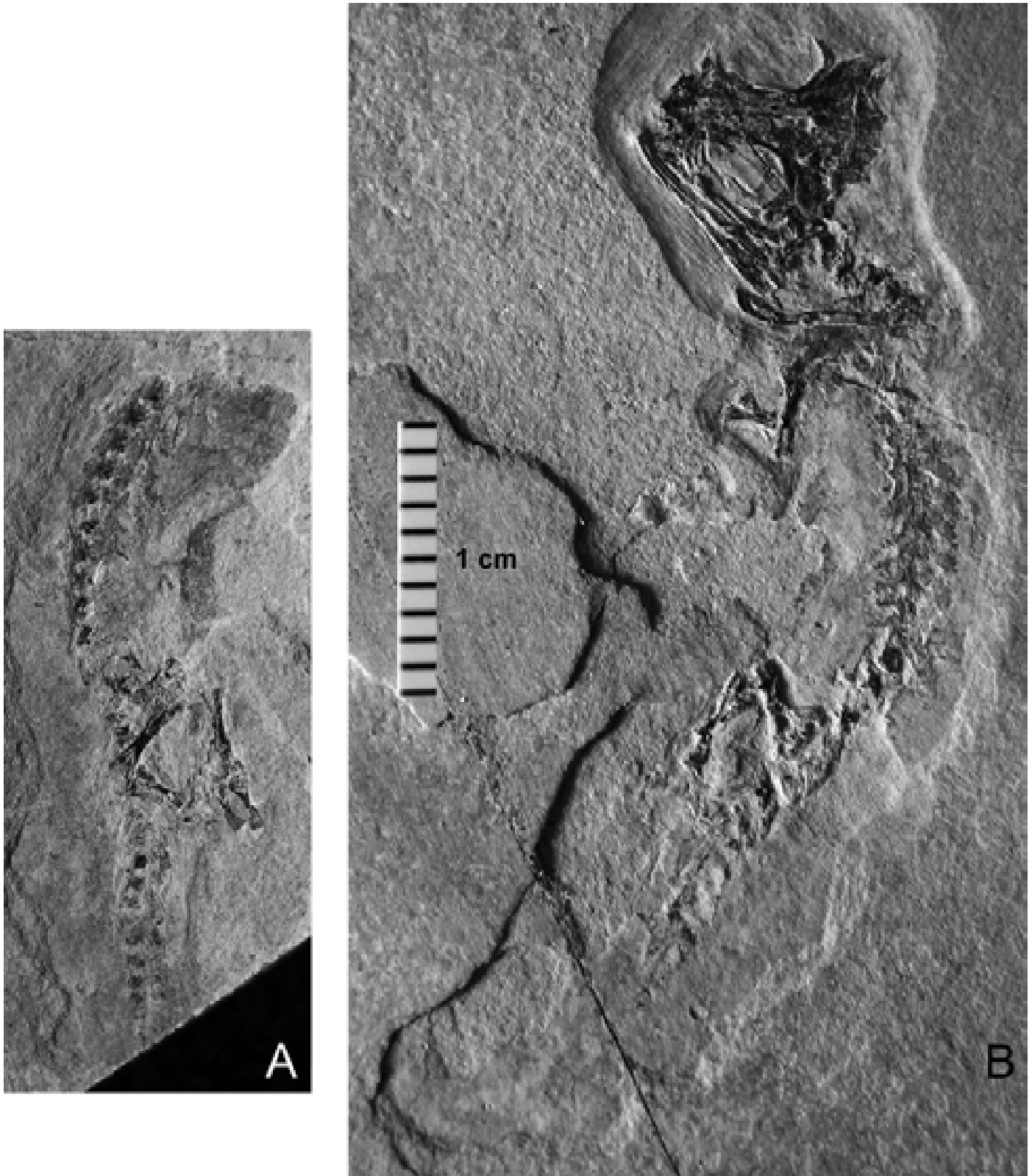


FIGURE 9. *Milnererpeton huberi*, second skeleton. **A**, NMMNH P-15019b. **B**, Counterpart of NMMNH P-15019a.

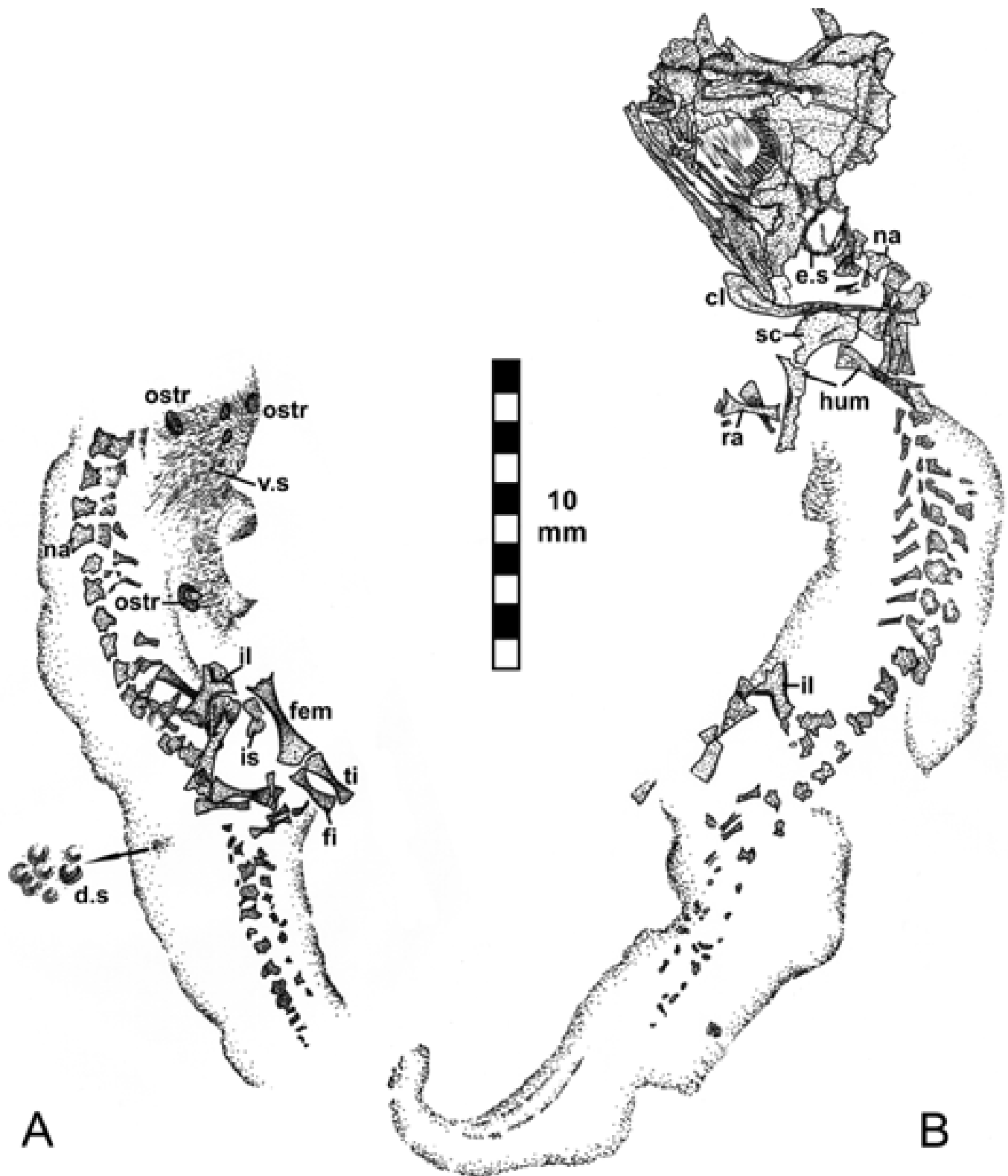


FIGURE 10. *Milnererpeton huberi*, second skeleton. **A**, NMMNH P-15019b. **B**, Counterpart NMMNH P-15019a. **Abbreviations:** **cl**, clavicle; **d.s**, dorsal scales; **e.s**, endolymphatic sac; **fem**, femur; **fi**, fibula; **hum**, humerus; **il**, ilium; **is**, ischium; **na**, neural arch; **ostr**, ostracodes; **ra**, radius; **sc**, scapulocoracoid; **ti**, tibia; **v.s**, ventral scales.

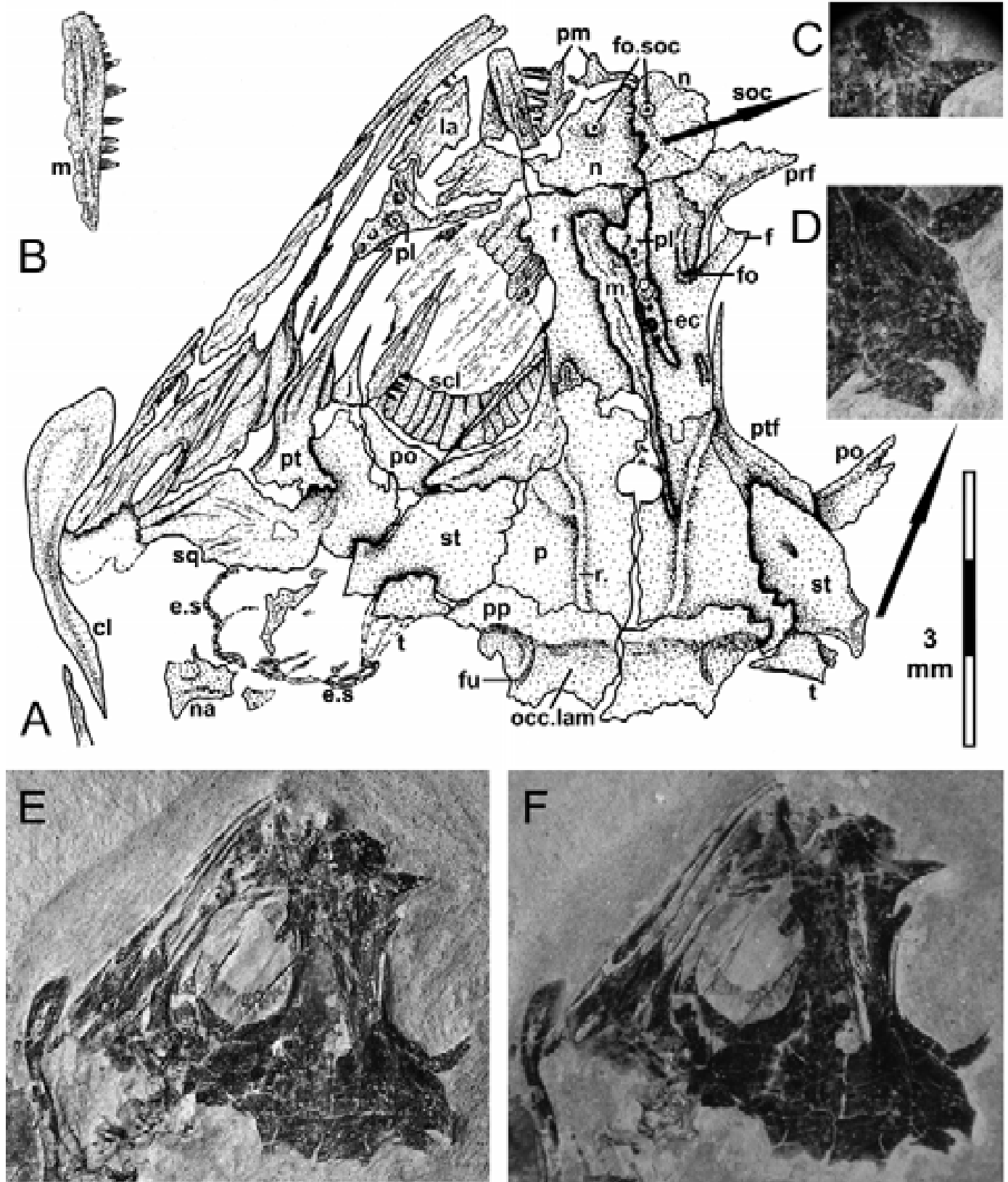


FIGURE 11. *Milnerpeton huberi*, skull of the second specimen NMMNH P-15019a. A, Ventral skull drawing. B, Maxilla with teeth. C, Foramina and internal canal of the supraoccipital lateral line in nasal and prefrontal. D, Supratemporal. E-F, Skull without and with covering of ethanol. **Abbreviations:** cl, clavicle; ec, ectopterygoid; e.s, endolymphatic sac; f, frontal; fo, foramen; fo.soc, supraoccipital foramen; fu, furrow; j, jugal; la, lacrimal; m, maxilla; n, nasal; na, neural arch; occ.lam, occipital rim (flanche); p, parietal; pl, palatine; pm, premaxilla; po, postorbital; pp, postparietal; ptf, postfrontal; prf, prefrontal; pt, pterygoid; r, "parietal ridge;" scl, sclerotic ring; soc, supraoccipital canal; sq, squamosal; st, supratemporal; st.f, subtemporal fenestra; t, tabular.

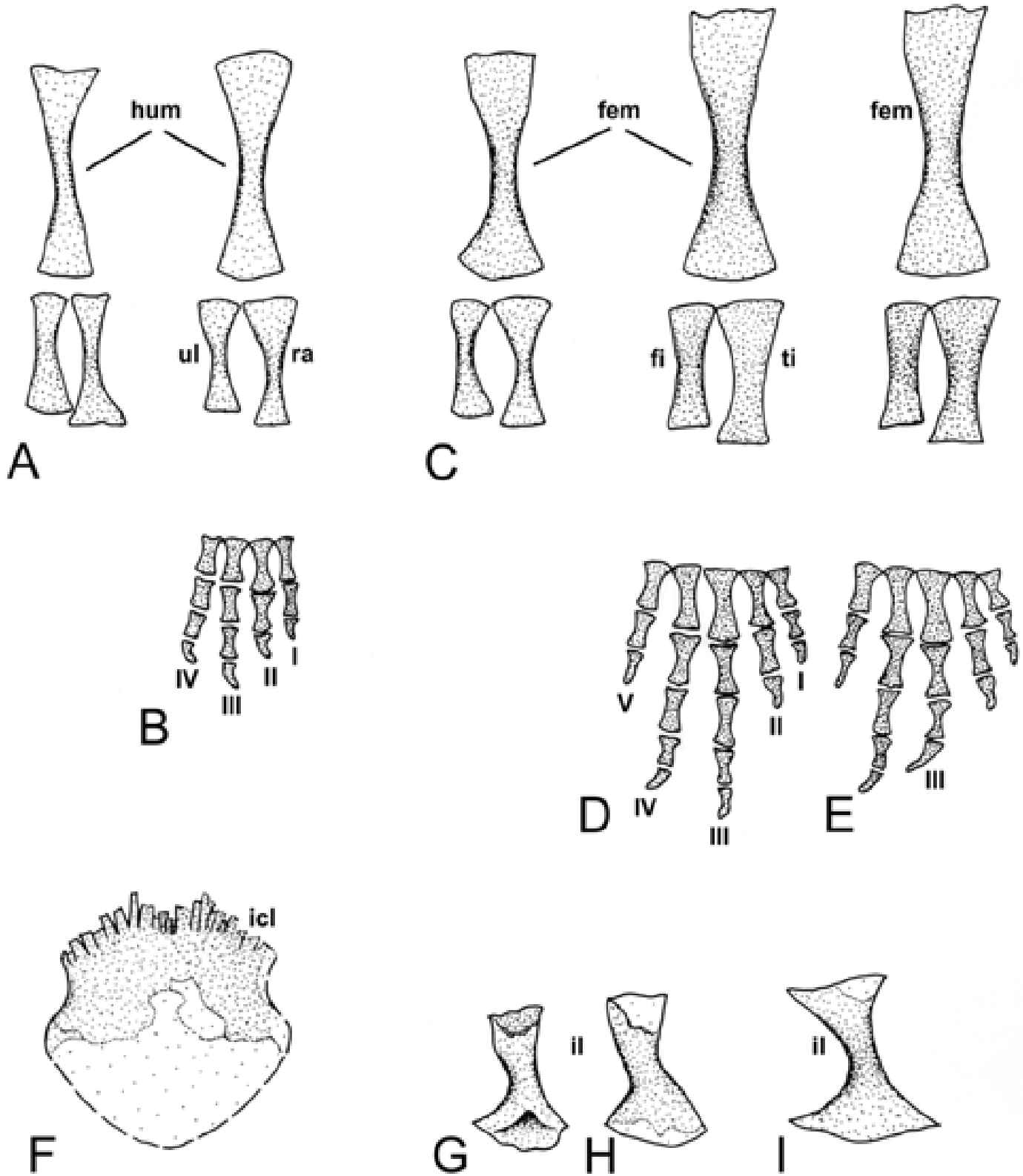


FIGURE 12. *Milnererpeton huberi*, fore- and hind limb, interclavicle and ilium in comparison. **A**, Forelimb, NMMNH P-15019. **B**, Forelimb, holotype NMMNH P-3696. **C**, Hindlimb, NMMNH P-15019. **D**, Pes, holotype NMMNH P-3696. **E**, Pes, holotype NMMNH P-3696. **F**, Reconstruction of the interclavicle, holotype NMMNH P-3696. **G**, Ilium in lateral view, NMMNH P-15019. **H**, Ilium in medial view, NMMNH P-15019. **I**, Ilium in medial view, holotype NMMNH P-3696. Drawings with different scale bars (compare figs. 4 and 9). **Abbreviations:** **fem**, femur; **fi**, fibula; **hum**, humerus; **icl**, interclavicle; **il**, ilium; **ra**, radius; **ti**, tibia; **ul**, ulna.

DISCUSSION

The revision of *Milnererpeton huberi* (Hunt et al., 1996) has provided several new results:

- Lateral construction of the palate with Y-shaped palatine.
- Shape of the otic notch.
- Reconstruction of the skull in ventral and palatal views;
- Remains of the lateral line system;
- Structure of branchial ossicles and hyobranchial skeleton;
- Reconstruction of the interclavicle;
- Exact determination of the presacral vertebrae count;
- Sexual dimorphism;
- Feeding strategy.

Milnererpeton huberi from the Kinney Brick Quarry is re-diagnosed based on the combination of 17 characters. Most important is the morphology of the: supratemporal (1), maxillary shelf (5), parasphenoid (7), branchial ossicles (9), and the thoracic neural arches (10), and the presacral count of 22 vertebrae (11).

Milnererpeton huberi is closely related to the oldest known branchiosaurid, *Branchiosaurus salamandroides* from the Pennsylvanian (Westphalian D) of Nýrany, Czech Republic (Werneburg, 2012). Both species share many characters (2-4, 7, 9-11, 13-17), but are distinguished by the characters 1, 5-6 and 12 in the diagnosis of *M. huberi*. In some aspects the Lower Permian *Schoenfelderpeton prescheri* (Boy, 1986) from Germany is comparable to *M. huberi* (1-2, ?4, 7, 13, 16-17), but many differences also exist (3, 5-6, 9-12, 14-15).

In a few characters *M. huberi* demonstrates a strong relationship with the Late Pennsylvanian amphibamids. Among these amphibamids larval specimens are closely comparable in sharing some primitive features (i.e., post- and prefrontal contact). *M. huberi* shares the features 1c, 4, 10, 12, 15 and 17 with *Platyrhinops fritschi* from the Czech Republic and the characters 1c, 4, 5 and 15 with *Amphibamus grandiceps* from Mazon Creek. However, there are many differences between *P. fritschi* (1a, b, 5-9, 11, 16) and *A. grandiceps* (1a, b, 6-11, 16).

It is possible to distinguish *Milnererpeton huberi* from related genera and species of dissorophoids. However, it is difficult to attribute the New Mexican species to the family Branchiosauridae or Amphibamidae. The knowledge of thousands of branchiosaurids (Boy, 1972 ff.; Schoch, 1992 ff.; Werneburg, 1983 ff.) and increasing number of larval amphibamids (Milner, 1982; Schoch, 2001; Werneburg, 2012) now allows an explanation for the degree of variability of widely recognized characters in the small-sized dissorophoids. Unfortunately, some of these characters ("clear cut features") can no longer be considered as valid.

No longer valid characters of the branchiosaurids are:

- Y-shaped outline of palatine (Schoch and Milner, 2008), shared with the amphibamid *Platyrhinops fritschi* and the trematopid *Mordex laticeps* (Werneburg, 2012; see discussion above);
- Presacral count of 21-22 vertebrae (Schoch and Milner, 2008), shared with *Amphibamus grandiceps* (20-21 after Bolt, 1979, p. 547, in Milner, 1982; or fewer?);
- Posteriorly elongated, large tabular horn (Schoch and Milner, 2008), shared with *P. fritschi* and *M. laticeps* (Werneburg, 2012);
- Hyobranchial skeleton ossified (e.g. Milner, 1982), shared with *P. fritschi*, *M. laticeps*, *Nyrannerpeton amilneri* (Werneburg, 2012) and others;
- Ribs with widened distal ends (Schoch and Milner, 2008), shared with *P. fritschi* (Werneburg, 2012);

No longer valid characters of *Amphibamus* + *Platyrhinops* and some other amphibamids are:

- Supratemporal laterally constricted by enlarged otic notch (Schoch and Milner, 2008),

shared by the branchiosaurid *Schoenfelderpeton prescheri* (Boy, 1986) and *Milnererpeton huberi*;

- Bicuspidity (Schoch and Milner, 2008), not known in all amphibamids;
- Frontal anteriorly widened (Clack and Milner, 2010), constricting prefrontal to a narrow posterior process (Schoch and Milner, 2008), shared by *M. laticeps*, the micromelerpetontid *Limnogyrinus elegans* (Werneburg, 2012), and *Milnererpeton huberi*;

The above observations support one conclusion: The larval amphibamids and branchiosaurids passed through the same early ontogenetic stages. A next obvious step in this discussion is to focus attention on large branchiosaurids to search for features that are characteristic of amphibamids. This will clarify if the Branchiosauridae are a clade of paedomorphic amphibamids (Schoch and Milner, 2008) or the recognition of the branchiosaurids as a subfamily of the Amphibamidae (Clack and Milner, 2010).

Milnererpeton huberi is presently designated as a probable branchiosaurid, because its branchial ossicles have two (multi-ended) denticles. The designation is supported by the unsculptured parasphenoid, which characterizes most of the branchiosaurids. The two specimens of *Milnererpeton huberi* were very probably neotenic forms, as are most of the branchiosaurids.

A lateral line system is easily recognized in many aquatic temnospondyls as sulci in the sculpture pattern of the skull roof. This character is normally not known in amphibamids and branchiosaurids, but may exist in some branchiosaurids (Werneburg, 1988, figs. 6b, 9d, 11c). However, the lateral lines are traceable in *Milnererpeton huberi*, but from the ventral skull roof. The supraoccipital line is preserved as an internal canal (meatus) with foramina at both ends of this meatus on the nasal and the prefrontal (and possibly also on the frontal). Another open canal or furrow is present on the inner surface of the occipital rim of the skull table.

The two described specimens of *Milnererpeton huberi* may exhibit a sexually dimorphic character in branchiosaur-like amphibians. The distal tail of the holotypic specimen presents a well mineralized or ossified notochord, and the second specimen has a well preserved tail region without notochord remains. Werneburg (2008) has reported notochords in more than 200 specimens of 11 species in branchiosaurids, micromelerpetontids, saurerpetontids, discosauriscids and others. He has reported a mineralized notochord in the morphotype of branchiosaurids, despite equally good soft part preservation of the specimens. In 110 thanatocenose specimens of the branchiosaurid *Apateon dracyiensis* with very good skin preservation, 42 specimens were recorded with (38%) and 68 specimens without (62%) a mineralized notochord.

Sexual dimorphism provides the simplest interpretation of this phenomenon. The mineralization of the notochord covering leads to a narrowing of the soft notochord and, therefore, to a general hardening of the tail region, which alleviates the undulation of the body and in general would increase mobility. A comparison of the branchiosaurids with the almost identical morphotype of present day salamander larvae shows that a strengthened tail region in males brings advantages in mating, swimming, and defence. Therefore, the holotypic specimen of *Milnererpeton huberi* may be a male (Figs. 4-5) and the second specimen a female (Figs. 9B, 10B).

Both specimens of *Milnererpeton huberi* had an interesting dual feeding strategy. They were equipped with a system of branchial ossicles, which bear one or two needle-like points. This derived branchial system allowed the animals to filter feed on planktonic prey. Additionally, they fed on ostracodes. This feeding strategy is demonstrated by the presence of ostracodes in the intestine or stomach content of both specimens. The feeding spectrum is well known from the Lower Permian branchiosaurids of the Thuringian Forest Basin in Germany and is extended by the New Mexican specimens with ostracods. *Apateon flagrifer*

preferred conchostracans (based on stomach contents) in an ontogenetic stage with a skull length of 10 mm (Werneburg, 1986). The branchiosaurids *Apateon kontheri*, *Melanerpeton eisfeldi* (Werneburg, 1988), and *Melanerpeton sembachense* (Werneburg, 1989) have small branchiosaurids and paleoniscids in their stomach or intestine content at an ontogenetic stage of a skull length of 15 up to 34 mm. In one case true cannibalism was demonstrated (Werneburg, 1989). An additional feeding strategy is known from the Permo-Triassic branchiosaurid *Tungussogyrinus bergi* from the Tungus-Basin in Siberia (Werneburg, 2009) in which tricuspid teeth were evidently used to scrape algae.

The probable branchiosaurid *Milnererpeton huberi* was found together with ammonites. This prompts the question: was *Milnererpeton* a marine amphibian? Clearly the answer is no, as the following topics of discussion would appear to indicate:

(1) Branchiosaurids like *Milnererpeton* were r-strategists that employed neoteny; that is small size and a high reproductive rate as life-history strategies (Boy and Sues, 2000). These newt-like amphibians can be regarded as an evolutionarily successful clade. Many thousands of specimens have been collected from the Lower Permian lake-lagerstätten in Germany, France, and the Czech Republic. Even widely distant occurrences of branchiosaurids, such as on the Isle of Sardinia (Italy), yielded hundreds of specimens (Werneburg et al., 2007), or at least the 30 specimens from the 8000 km distant Tungus-Basin in Siberia (Werneburg, 2009). Mass occurrences of these animals preserved in their life habits are clearly typical. On the other hand, only two specimens of the branchiosaurid-like *Milnererpeton huberi* are known after extensive excavations at the Kinney Brick Quarry Lagerstätte. Yet, they are accompanied by thousands of other fossils from units 3-4 (Fig. 1). Therefore, it is presumed that the bed with the thanatocenose was not the life environment of *Milnererpeton huberi*.

(2) Fortunately, we know the intestine or stomach contents of both specimens of *Milnererpeton huberi* included ostracods that must have lived in the same environment of these amphibians. All of the ingested ostracodes are tentatively identified as *?Carbonita* (K. Kietzke, written commun., 2013). According to Kietzke and Kaesler (1992, p. 131), the ostracode *Carbonita* inhabited nearly fresh to brackish water. This evidently was also the life habitat of *Milnererpeton huberi*: a fresh water body, which on occasions may have been temporarily brackish.

The probable branchiosaurid *Milnererpeton huberi* is the most definitive branchiosaurid known from North America. All other branchiosaurid records there are doubtful (compare Milner, 1982; Boy, 1987). A short overview of the North American “branchiosaurs” includes:

- “*Branchiosaurus darrahi*” Romer, 1939 (MCZ 1265), may be a true branchiosaurid. But very probably this specimen was not from the Dunkard Basin in Pennsylvania, as noted by Romer (1939), but the Autun-Basin of France (Werneburg and Spencer, 2007). The bituminized gray mudstone surrounding the thoracic region was retained as a very narrow envelope. Additionally, the characteristic old French preparation method to polish the skeleton with brushes is preserved in the holotypic specimen. This is strong evidence for the provenience of its origin as the Millery Formation of Autun. Such specimens of “*Protriton petrolei*” from Autun were incorporated into repositories of the world by the 1930s, as was the branchiosaurid “*Apateon petrolei*” (MCZ 1263) in the Museum of Comparative Zoology in Harvard University, Cambridge. It is supposed that two branchiosaurid specimens from Autun were originally in the MCZ collection.

- Larval dissorophoids from the Francis Creek Shale, Mazon Creek in Illinois, Late Pennsylvanian (Westphalian D):

Amphibamus grandiceps clearly is represented by two larval specimens with a skull length of 5-7.0 mm (Milner,

1982, figs. 4, 5a-d; Schoch, 2001, fig. 4; FMNH-PR664, FMNH-R38). They have a sculptured basal plate of the parasphenoid without internal carotid foramina, a low maxilla + premaxilla, a wide postorbital region (Boy, 1987, $H_w/S_1 = 0.90-98$), and a wide interorbital region ($IO_w/S_1 = 0.36-38$, postmetamorphic = 0.34).

Unknown dissorophoid – type A, skull length 4-5.0 mm, (“*?Branchiosaurus* sp.” described by Milner, 1982, figs. 5e, 6a; Boy, 1987, figs. 9b, c; YPM-803, YPM-802). They have a smooth basal plate of the parasphenoid with internal carotid foramina, maxilla + premaxilla with a high process, and a wide interorbital region.

Unknown dissorophoid-type B, skull length 7-8 mm, (Milner, 1982, figs. 5f-g, 6b; Boy, 1987, p. 100-101; USNM 4400). This unique specimen has a smooth basal plate of the parasphenoid without internal carotid foramina, probable low premaxilla, a narrow postorbital region (Boy, 1987: $H_w/S_1 = 0.65$) and a relatively narrow interorbital region ($IO_w/S_1 = 0.29$).

- Larval dissorophoids from Linton locality in Ohio, Late Pennsylvanian (Westphalian D), not known yet. The smallest published specimen of *Platyrhinops lyelli* has a skull length of about 13 mm and may belong to the juvenile stage of development (Clack and Milner, 2010, fig. 3).

AGE OF THE KINNEY BRICK QUARRY DEPOSIT

Some early workers (e.g., Stucky, 1967) assigned a Permian age to the Kinney fossils, based primarily on an unsubstantiated report of the “Permian index fossil” *Callipteris* (cf. Read and Mamay, 1964) from the quarry (Kelley and Northrop, 1975). In the 1990s (see articles in Zidek, 1992) most workers considered the quarry to be Virgilian in age, based on the incorrect stratigraphic placement of the quarry in the geologic mapping of Myers and McKay (1976), but not based on any direct fossil evidence from the quarry. About 3 m below the stratigraphic level of the Quarry a 0.3-m-thick fusulinid wackestone occurs (Fig. 1A), which is dated as early/middle Missourian, belonging to the upper part of the early Kasimovian. The conodont fauna from unit 1, the marine limestone (Fig. 1), is early to middle Kasimovian in age, and provisional assignment is made to the *Idiognathodus confragus* Zone of the Midcontinent conodont zonation (*Streptognathodus confragus* Zone of Barrick et al., 2004) by Barrick in Lucas et al. (2011) and Schneider et al. (this volume). Based on the occurrence of the insect zone species *Syscioblatta allegheniensis* at Kinney, the horizon corresponds in the European scale to the *Sysciophlebia grata-Syscioblatta allegheniensis* zone, which correlates to the *Branchiosaurus fayoli* amphibian zone situated in the Stephanian B of the west European regional scale (Schneider and Werneburg, 2006; Werneburg and Schneider, 2006; Schneider et al., this volume).

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In the Cerros de Amado, northeast of Socorro, algal bioherms are characteristic of the Upper Pennsylvanian (late Missourian) Council Spring Member of the Atrasado Formation. Here, near the Ojo de Amado, the Council Spring Member is a massive (up to 40-ft thick) bioherm that extends on strike for about 500 ft. It overlies slope-forming strata of the Tinajas Member, and is overlain by Burrego Member strata.

TAPHONOMY AND PALEOECOLOGY OF PENNSYLVANIAN FISHES FROM THE KINNEY BRICK QUARRY, NEW MEXICO, USA

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Abstract—The Kinney Brick Quarry in central New Mexico, USA, has long been considered a Late Pennsylvanian Lagerstätte for both its fossil flora and fauna. However, this is not true for all fish species found in the Kinney Quarry. We analyze the three largest collections of Kinney fish specimens, totaling 597 specimens, which are housed at the New Mexico Museum of Natural History, University of Kansas Museum of Natural History and the Carnegie Museum of Natural History. These three collections represent 95% of the fossil fish collected from the Kinney Quarry. Four classes and 24 species are represented. The highly abundant Acanthodii are represented by *Acanthodes kinneyi*. The highly abundant and diverse Actinopterygii are represented by *Platysomus schultzei*, *Platysomus* sp., *Amphicentrum jurgenai*, *Schizolepis manzanitaensis*, *Tanyrhynchichthys mcallisteri*, *Pyritocephalus lowneyi*, cf. *Bourbonnella* and seven undescribed palaeoniscoid morphotypes. The diverse but rare Chondrichthyes are represented by *Cobelodus aculeatus*, *Symmorium reniforme*, *Orthacanthus huberi*, *Peripristis* aff. *P. semicircularis* and ?*Listracanthus*. The diverse and rare Sarcopterygii are represented by Coelacanthidae gen. et. sp. indet., Rhabdodermidae gen. et. sp. indet., a new species of Rhabdodermidae, and *Sagenodus (Proceratodus) hlavini*. These fish can be divided into three ecological groups: (1) fish that lived in the estuary: *Acanthodes kinneyi*, *Platysomus*, *Amphicentrum jurgenai*, *Schizolepis manzanitaensis*, palaeoniscoids types 5, 6, "small scale" palaeoniscoid, *Cobelodus aculeatus*, Rhabdodermidae type 1; (2) the open marine visitors to the Kinney estuary: *Symmorium reniforme*, *Peripristis* aff. *P. semicircularis*, ?*Listracanthus*, Rhabdodermidae type 2 and Coelacanthidae gen. et sp. indet; and (3) the few freshwater species of fishes: *Orthacanthus huberi*, coelacanthid indet. and *Sagenodus hlavini*. There are a few fish of uncertain paleoecology: *Pyritocephalus lowneyae*, *Tanyrhynchichthys mcallisteri*, and most of the undescribed morphotypes of palaeoniscoids. The Kinney fish assemblage can be compared to modern fish communities that live in lagoons in northern Queensland, Australia. This comparison indicates that the fish fauna at Kinney quarry closely fits that of a deep-water estuarine fish community. This suggests that the water in the Kinney estuary was not less than 1 m and no more than 10 m in depth, and most likely between 5-7 m in depth. Small fish are not common at Kinney, and fish diversity is relatively low; thus, Kinney is not an ideal Lagerstätte for fish.

INTRODUCTION

The Kinney Brick Quarry is an approximately 11.8 acre quarry in the Manzanita Mountains near Albuquerque, New Mexico (Fig 1). The sediments in the quarry represent a Lagerstätte developed in the Late Pennsylvanian Tinajas Member of the Atrasado Formation. The exposed section at the quarry is 28 m thick and consists of limestone, shale and sandstone (Fig. 2) (Kues and Lucas, 1992; Lucas et al., 2011). The Kinney fossil fish assemblage is concentrated in the lower 0.5 m of strata exposed at the quarry (Fig. 2), in calcareous shales (unit 3-4) directly above the basal lime mudstone (units 1 and 2) of the quarry floor (Kues and Lucas, 1992).

The sediments exposed at the Kinney Quarry represent a delta prograding into an estuary (Kues and Lucas, 1992; Lorenz et al., 1992), so that water was becoming fresher towards the top of the stratigraphic column (Fig. 2). The lower, thinly bedded shales exposed at the Kinney Quarry have been interpreted as a protected estuary because of the lack of bioturbation, thinly laminated sediments and large amount of organic matter in the lower part of the section, which suggest that the bottom waters were anoxic (Feldman et al., 1992; Archer and Clark, 1992). The estuary was stratified, with low energy water. The river that delivered sediments to the prograding delta front was a freshwater source that lowered the salinity of the water. Kinney is unusual because of the mixture of fossils of organisms that are only freshwater in origin (e.g., land plants, insects, amphibians) with organisms that are only marine in origin (e.g., brachiopods, marine bivalves, cephalopods). These organisms either coexisted in the estuary and/or were transported into the estuary (Huber, 1992a).

The remarkable fossil assemblage from the Kinney Quarry has been fairly well studied in the 1992 New Mexico Bureau of Mines and Mineral Resources Bulletin 138 (Zidek, 1992a), but there are still some questions that remain unanswered, such as water depth of the Kinney estuary and the ecological structure of the fish community. In this paper we examine each of the species of fish found in the quarry to assess whether it was autochthonous or allochthonous to the Kinney estuary. Based on this, we compare the fish that lived in the Kinney estuary to modern lagoonal/estuarine fish faunas to estimate the depth of water. Also, from this analysis we can reconstruct the paleoecology of the Kinney estuary and gauge the quality of the Kinney Lagerstätte for fish preservation.

MATERIAL AND METHODS

We examined the Kinney Brick Quarry collections at the New Mexico Museum of Natural History and Science (NMMNH), University of Kansas Museum of Natural History (KUVP) and the Carnegie Museum of Natural History (CM). We surveyed a total of 597 fish specimens from the Kinney Quarry from these three museums. NMMNH has the largest collection, with 302 specimens; KUVP has 238 specimens in their collection; and CM has 52 in their collection (Fig 3). There are 9 described and 13 undescribed species of fish from the quarry (Table 1). The fish fauna consists of 5 species and 8 undescribed morphotypes of Actinopterygii, 1 species of Acanthodii, 5 species of Chondrichthyes, and 4 species of Sarcopterygii (Fig. 3B-C). Actinopterygian fish dominate both in abundance and diversity. Acanthodians are highly abundant but not diverse. Chondrichthyan fish are diverse but very rare. Sarcopterygii are diverse and members of this clade are fairly rare.

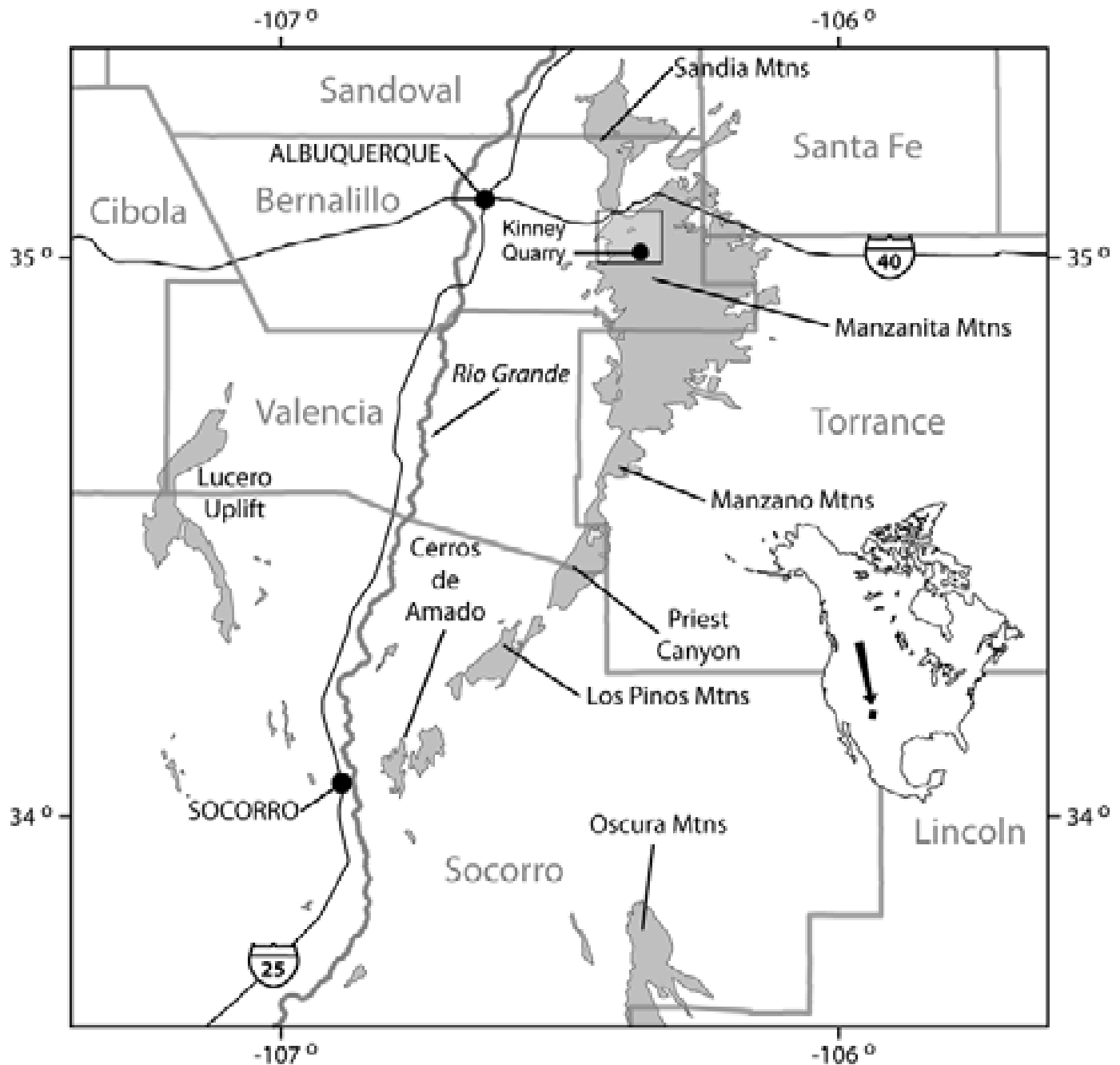


FIGURE 1. Map showing location of Kinney Brick Quarry in central New Mexico, USA (after Lucas et al., 2011).

STRATIGRAPHIC DISTRIBUTION OF FISH FOSSILS

Fish fossils at Kinney come from the lower 3 m of the outcrop, specifically a 0.5-m-thick fish-bearing interval (Fig. 2). In the section, unit 1 is a blocky black micrite that has conchoidal fractures and is very tough to break with a sledge hammer. A few isolated actinistian fragments have been found in this unit. Unit 2 is a black-banded, slightly fissile calcareous shale. The articulated actinistian and chondrichthyan remains are from this horizon. The fish are concentrated in unit 3. This unit is a fissile, banded calcareous shale, that is black with orange bands at the bottom and grades upward to orange with black bands to almost a pure orange shale. Within this 20-30 cm thick unit fish can be placed within a couple of cm of their origin within the section. While there is a concentration of fish within this 20-30 cm thick unit there are not death assemblages with many specimens on any one horizon. Unit 4 is a platy, soft

calcareous shale. It has a uniform medium gray color and is where the plant fossils are concentrated. There are a few fish from this interval. Strata higher in the section only produce a few isolated elements (for example, Huber [1992a] reported a paleoniscoid from unit 7).

FISH TAPHONOMY AND PALEOECOLOGY

The abundance of Kinney Quarry fish can be divided into three categories (Table 2). The first category is highly abundant (fish that are common in the quarry). This group includes *Acanthodes kinneyi* and *Platysomus* spp. The second group is moderately abundant; this group is about six to seven times less common than the fish in the first group. This group includes *Amphicentrum jurgenai*, *Schizolepis manzanitaensis*, types 5 and 6 of Bardack's (1992) classification of undescribed paleoniscoid morphotypes and the "small scale" morphotype. The last

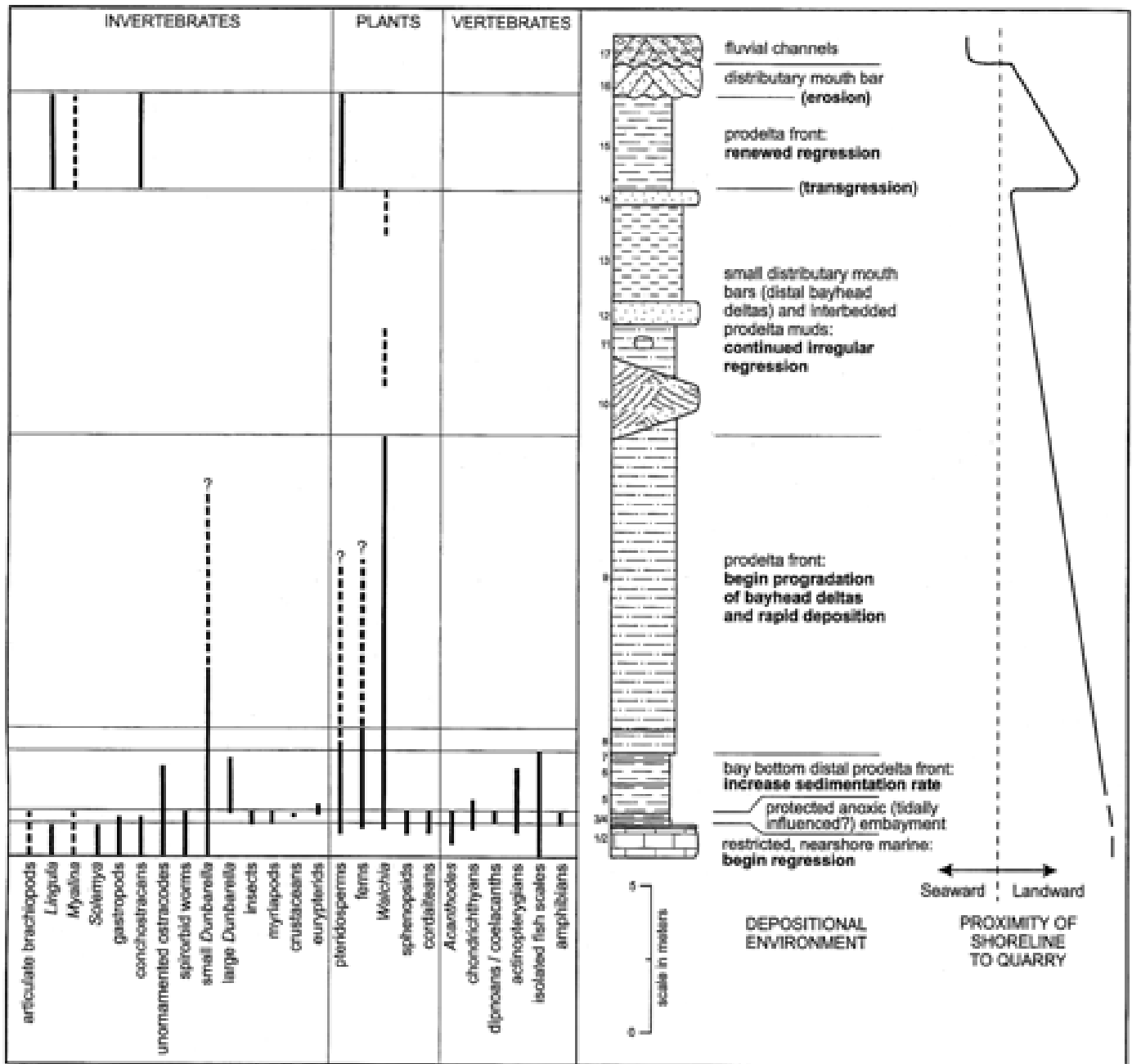


FIGURE 2. Stratigraphic section at the Kinney Brick Quarry (after Lucas et al., 1999).

group is the fish that are found as a few specimens--only one or two specimens have been recovered from the quarry. This group includes all the chondrichthyan fish, *Sagenodus (Proceratodus) hlavini*, *Pyritocephalus lowneyae*, cf. *Bourbonnella* sp., the coelacanths and types 1, 2, 3 and 4 of the undescribed palaeoniscoid morphotypes.

Whitfield (1999) describes eight classifications of estuarine residents. Figure 2 and table 2 of Whitfield provide a thorough review of the literature and consolidate the many types of life cycles into these 8 categories. We will use this classification system to analyze the paleoecology of each fish species (Table 3).

Acanthodes kinneyi Zidek 1992

At the Kinney Quarry, acanthodians are represented by one species, *Acanthodes kinneyi*. These fish are the second most abundant fish in the quarry, with 110 specimens collected (Table 2). The majority of these fossils represent fish longer than 14.5 cm; the holotype of *A.*

kinneyi is 28.0 cm long (Fig. 4A), and it is among the largest of the fish from the quarry. These fish are well preserved despite being delicately boned and scaled. *A. kinneyi* is fully squamated by 14.5 cm long (Zidek, 1975), and Zidek (1992b) notes that the endoskeleton of these fish begins to ossify unusually late, at about 14.5 cm. Because of these developmental characteristics, there is a predictable preservational bias towards larger fish, which is what we see in the Kinney collections. All of the specimens collected represent adult and sub-adult fish. There are not fry or juveniles found in the Quarry.

This species of *Acanthodes* lacks teeth. It has numerous fine gill rakers. Both of these indicate that this species was a detritivore or suspension feeder. We infer that the acanthodians were either marine migrants or freshwater migrants because they represent almost 20% of the fish found in the quarry, and they are remarkably well preserved for such delicate fish. This suggests that these fish did not travel far from the location of their death, which means they were autochthonous to the

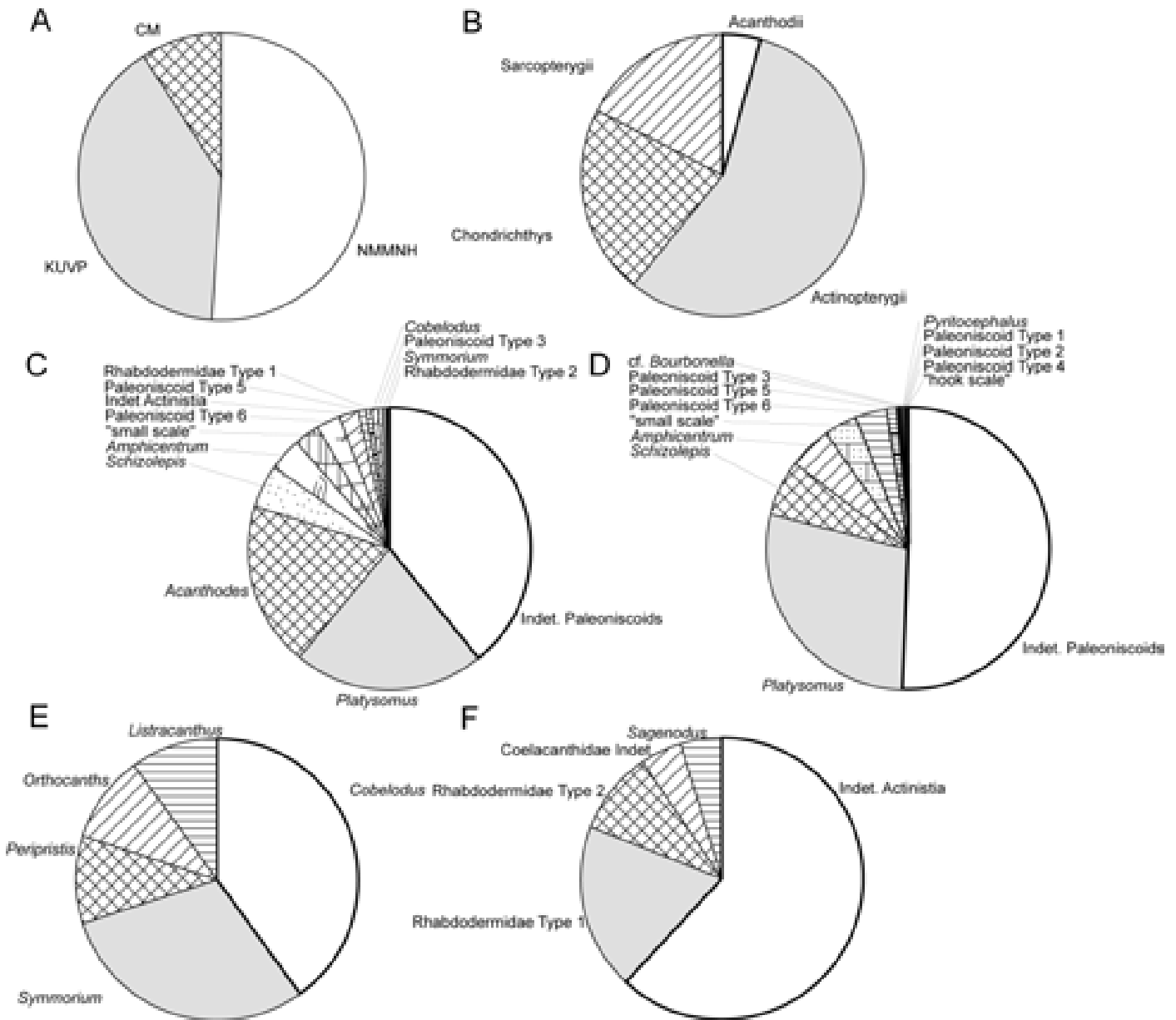


FIGURE 3. Pie diagrams showing the: **A**, distribution of specimens of Kinney Quarry fishes by collection; **B**, the overall diversity for each of the classes; **C**, the overall abundance of the fish within each class; **D**, the relative abundance of Actinopterygii; **E**, the relative abundance of Chondrichthyes; and **F**, the relative abundance of Sarcopterygii.

estuary. Zidek (1992b) inferred that the sole presence of adult and sub-adult individuals at this locality indicates it was a feeding ground for the adult fish and they migrated to spawn.

Zidek (1980, 1992b) and others (Denison, 1979; Boy, 1998, 2003; Boy and Schindler, 2000; Zajíc 2005; Heidtke, 2007, 2011) describe very different depositional environments in which acanthodians are found. They cover the spectrum from fully marine to obligate freshwater fish. Zidek (1992b) described members of *Acanthodes* as being much like modern salmonidids--anadromous and migrating to streams or lakes to spawn. This would also fit in the freshwater migrant category and would explain the lack of juvenile and fry in the estuary where the adults would feed. Juveniles of these fine-scaled, delicately boned fish are preserved at localities such as Hamilton Quarry in Kansas (Zidek, 1976, 1988). Whether the preservational bias is due to high enough energy water to disarticulate the juvenile or fry, or the juvenile and fry were living in a different environment, it is easy to infer that the adults of *Acanthodes kinneyi* were residents of the Kinney estuary

Actinopterygii

Actinopterygian fishes at Kinney are both diverse and abundant (Fig. 3D). There are five described actinopterygian taxa found in the quarry. The other 7 species are undescribed. There is a significant difference in the relative abundance of these four species. *Platysomus schultzei* and *Platysomus* sp. are the most abundant fish found in the quarry. *Amphicentrum jurgenai* is a moderately abundant species. *Tamyrhinichthys mcallisteri*, *Pyritocephalus lowneyae* and cf. *Bourbonnella* sp. are very rare. The majority of the unsubscribed species of paleoniscoids are moderately abundant, with 4-17 specimens identifiable to those morphotypes.

Platysomus Agassiz 1835

There are two species of this deep-bodied fish in the Kinney quarry, *Platysomus schultzei* Zidek 1992 and *P.* sp. indet. The main morphological differences between these two species are in the fin con-

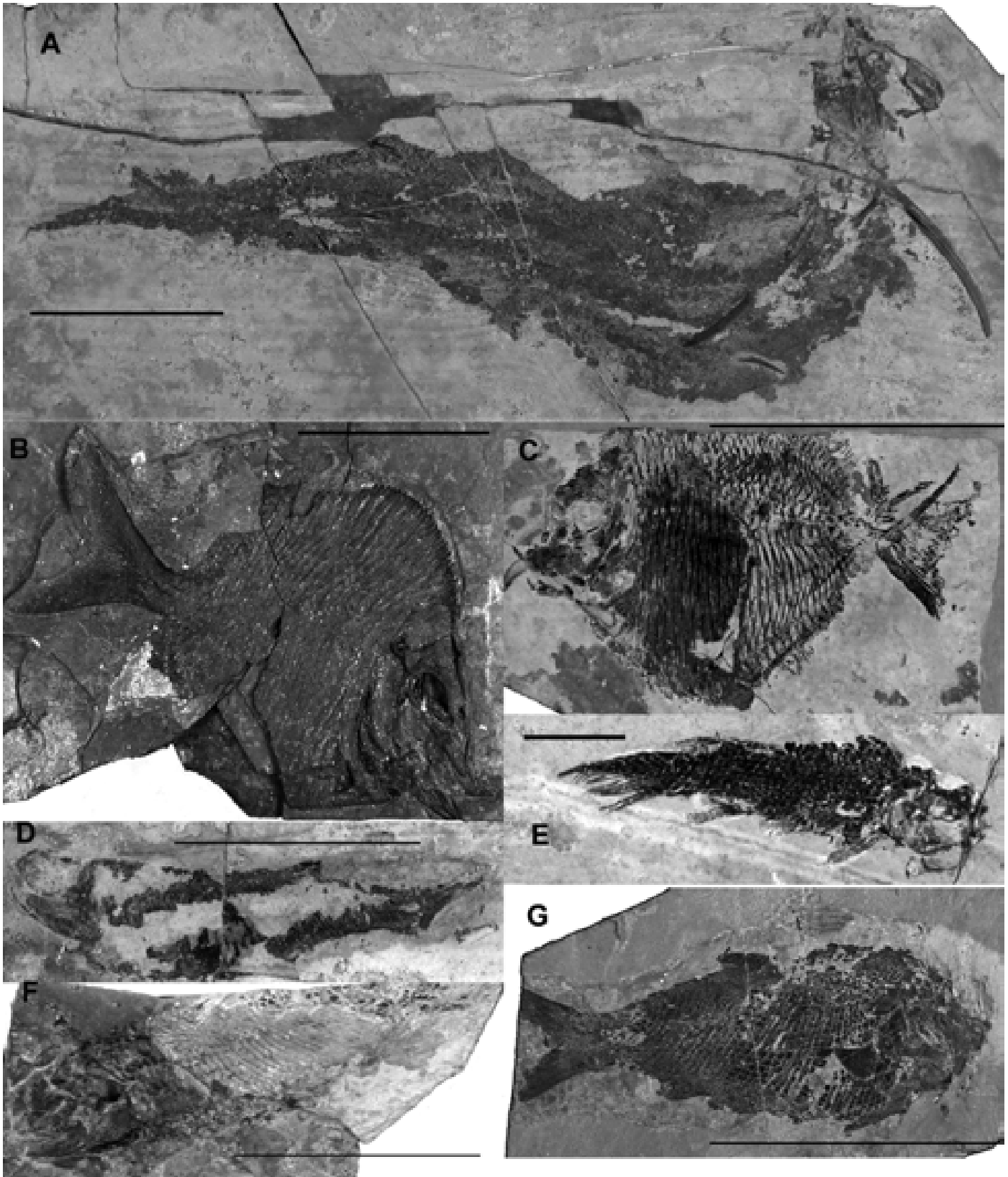


FIGURE 4. Representatives of the actinopterygian and acanthodian fish from Kinney Quarry. **A**, Holotype of *Acanthodes kinneyi*, NMMNH P-19208, as shown in Zidek (1992b, fig. 11b). **B**, The holotype of *Platyosomus schultzei*, NMMNH P-19195, as shown in Zidek (1992b, fig. 22). **C**, Paratype of *Amphicentrum jurgenai*, NMMNH P-19116, as shown in Zidek (1992b, fig. 15E). **D**, Paratype of *Schizolepis manzanitaensis*, NMMNH P-14365. **E**, Holotype of *Pyritocephalus lowneyae*, NMMNH P-12988, as shown in Huber (1992b, fig. 1). **F**, cf. *Bourbonnella*, KUVF-85296, as shown in Gottfried (1988). **G**, Palaeoniscoid type 4 of Bardack (1992), undescribed palaeoniscoid fish, NMMNH P-14368, shown in Bardack (1992, fig. 6). All scale bars are 5 cm except for E, which is 1 cm.

specimens collected are adult and sub-adult individuals. This is suggestive of the migrant hypothesis. Given the small size of this fish the juveniles could have been living in the marginal estuarine environments and thus not preserved in the Kinney sediments.

***Tanyrhynchthys mcallisteri* Gottfried 1987**

This is a rare, long-snouted palaeoniscoid predator. It ranges in size from 15 to 20 cm and has sharp, recurved teeth. It is slender and elongate, with a well-developed caudal peduncle. Gottfried (1987b) compares the morphotype of this fish with that of modern *Lepisosteus* (gar) and *Esox* (pike). The holotype of this specimen has a badly damaged skull. There is an additional specimen at NMMNH, but it is only recognizable by the hook on the scale.

Gottfried (1987b) notes that this is only the third species of long-snouted palaeonisciform to be identified in the Paleozoic. With unclear familial relations it is difficult to use presumed relatives to interpret characteristics like salinity tolerance and growth sequences. This predator is about the same size as type 5 of Bardack's morphotypes (see below). It is a rare fish, and both specimens show post-mortem gaseous deterioration of the skull prior to deposition. This sort of fish may be a lie-in-wait predator that preferred weedy stands of aquatic plants to hide in to wait for prey. Such weedy stands are often associated with the littoral zones and not the open water in the middle of the estuary. However, with only two specimens of this species it is difficult to assign it to a resident type.

***Pyritocephalus lowneyae* Huber 1992**

Pyritocephalus lowneyae is a tiny palaeoniscoid fish from the Kinney Quarry (Huber, 1992b). It is extremely rare; only one specimen has been found (Table 2), which is about 3.5 cm long and shows remarkable preservation (Fig. 4E).

This fish is of uncertain paleoecological origin. The specimen is well preserved, but only a single specimen of this small fish has been found. This occurrence of genus *Pyritocephalus* is the youngest westernmost occurrence (Gottfried, 1992). Throughout its occurrence it is found in tropical marginal marine to freshwater habitats. Westoll (1944) described this genus from Linton, Ohio, which is one of the few locations where large numbers of specimens have been found at one locality.

There are three possible paleoecological interpretations: (1)

Pyritocephalus was an estuarine resident, living around the edges of the estuary in shallow water. Its subsequent lack of abundance could have been caused by three factors. The first is that at Kinney there may have been preferential preservation of the fish that died near the deeper part of the estuary. If *Pyritocephalus* individuals were farther away from the deep water, then the distance of post-mortem travel may have been too great for such a small fish. (2) This fish was an estuarine resident but is not preserved due to predation. Indeed, these little fish are the right size to be the staple of larger fish. There are numerous specimens of gastric residue at Kinney that contain fish scales, but it is difficult to identify the species of fish being eaten from a ball of palaeoniscoid fish scales (Hunt, 1992; Hunt et al., 2012). (3) This fish was a marine straggler, and the lost fish is from a marine school that wandered into the estuary, and died.

cf. *Bourbonnella* sp.

Only one specimen of cf. *Bourbonnella* sp. has been collected from the Kinney Quarry (Fig. 4F). The fish is 8.3 cm long and shows moderate to poor preservation. It is articulated but shows post-mortem decomposition, and the pectoral fins are disarticulated (Gottfried 1987a).

Poplin (2001) describes this genus as a freshwater to euryhaline genus with a tropical distribution. The European species have freshwater origins, with the Kinney specimen being one of the few with marine associations. With only a single specimen of a species it is difficult to infer its paleoecological origin. Given that this specimen shows a great deal of disarticulation, which indicates a moderate amount of decay before burial, it is likely that this fish was a freshwater straggler into the Kinney estuary.

Indeterminate Palaeoniscoids

Bardack (1992) distinguished six morphotypes of palaeoniscoid fish in the Kinney Quarry that cannot be more precisely diagnosed due to the current taxonomy of palaeoniscoid fishes. We identified an additional morphotype, a moderately common "small-scale" morphotype. Some of these morphotypes are more abundant than others, and many of these fish are poorly preserved compared to the other Kinney fish. For example, there are numerous palaeoniscoid specimens that are little more than disarticulated scales in the shape of a fish (Fig. 3). These fish are in the intermediate range of size. The smallest are 8.0 cm long, and the largest are up to 22.0 cm long.

TABLE 3. 8 Classification of estuarine residents. Summary of Figure 2 and Table 2 of Whitfield (1999).

Resident Type	
Estuarine Resident	Fish which lives its whole life-cycle in the estuary
Estuarine Migrant	Fish that spawns in the estuary and the fry are residents of the estuary and the subadults and adults migrate to fully marine environments to grow and return to estuary to spawn
Marine Migrant	Fish that spawns and fry live in fully marine environments and the sub adults and adult fish live in the estuary to grow and then return to marine environment to spawn
Marine Straggler	Fish that lives its life cycle in the marine environment and occasionally enters estuary, but not for a prolonged period
Freshwater Migrant	Fish that spawn in the estuary and the fry are residents of the estuary and the subadults and adults migrate to fully freshwater environments to grow and return to estuary to spawn
Freshwater Straggler	Fish that lives its life cycle in the freshwater environment and occasionally enters estuary environment but not for prolonged period
Catadromous	Fish that spawns in marine environments and the larval stages pass through estuary on their way to freshwater where juvenile and subadults reach maturity and then pass through the estuary on their way back to the sea to spawn.
Anadromous	Fish that spawn in the fresh water environments and the fry and juvenile fish travel down stream passing through the estuary on their way to the fully marine environments to grow mature adults return to the fresh water environments to spawn.

Palaeoniscoid indet. "Small Scale"

A 8-12 cm long fish with small scales in relation to the size of the fish needs a full description. It is moderately abundant, being represented by 17 specimens. The specimens are well preserved. These fish were most likely living in the estuary, but it is difficult to determine if it was a migrant species or estuary resident.

Palaeoniscoid indet. "Type 5" Bardack, 1992

The largest of the indeterminate Palaeoniscoidae is "Type 5" (Bardack, 1992). The specimens of this fish are up to 22 cm long. It was most likely piscivorous. These fish have large "spike like teeth" (Bardack, 1992), as do most modern predatory fish that use teeth to grip and hold on to prey. Most likely these fish were solitary, feeding on the schools of smaller fish in the quarry. Large predatory fish typically constitute a smaller percentage of the population than do generalist feeders. This fish is uncommon in the quarry, representing 0.84 % of the specimens found. The specimens are all adult and sub-adult fish. These fish are well preserved with little to no disarticulation. The small number of these fish and their predatory behavior suggests that they were estuarine residents or marine or freshwater migrants.

Palaeoniscoid indet. "Type 6" Bardack, 1992

Type 6 is the most abundant group of palaeoniscoids--16 specimens of this type are found in the quarry. Type 6 ranges from 7 to 12 cm long. The skull morphology of this fish suggests that it was a generalist feeder, and its abundance and generally good preservation suggest it was an estuary resident, marine migrant or freshwater migrant.

Palaeoniscoid indet.

The rest of the palaeoniscoid fish are too poorly known to interpret their paleoecology with confidence. Taphonomically, these fish are very well preserved, with little to no disarticulation within the body. This indicates that these fish were dying in the estuary and were preserved with little or no post-mortem transport.

A large portion of the unidentifiable palaeoniscoids are preserved as bromalites (Hunt, 1992; Hunt et al., 2012). The partial digestion of these fish destroyed some of their identifiable characteristics. But, the large quantity of bromalites indicates that the palaeoniscoids were an important food source for other estuary residents.

Chondrichthyes

Chondrichthyan fish are uncommon in the Kinney Quarry; indeed, the quarry is almost devoid of chondrichthyan teeth and skeletal material (Fig. 3E). Most are only known from isolated teeth and spines, and we infer that most of these species represent taxa allochthonous to the estuary. If chondrichthyan fish were living in the estuary, an abundance of teeth would be seen. Nevertheless, the Kinney Quarry did marvelously preserve one cartilaginous skeleton of *Cobelodus aculeatus* (Fig. 5A).

Cobelodus aculeatus Cope 1894

Cobelodus aculeatus is the only chondrichthyan fish in the Kinney Quarry known from nearly complete skeletal material (Zidek, 1992b). This fish is known from four skeletons, all juvenile and nearly complete, with the largest being 41 cm long. One of the skeletons is complete, with the exception of the rostral area (Fig. 5A). Because of the conditions needed to preserve cartilage, these fish apparently underwent no post-mortem travel to end up in the estuary.

Given a natural mortality curve, juvenile individuals of *Cobelodus* are over-represented at Kinney if these fish were stenohaline visitors to the estuary intolerant of the salinity changes. A contrary hypothesis

would be that like the Bear Gulch Lagerstätte, the Kinney estuary was a nursery for the juveniles of these sharks (Grogan et al., 2004). This would also account for the distinct absence of adult shed teeth, which suggests that the adult fish rarely visited the estuary. We infer that these fish were marine migrants to the estuary.

Symmorium reniforme Cope, 1893

Symmorium reniforme is known at Kinney from three teeth (Fig. 5B), yet it is one of the more common chondrichthyan fish in the estuary. These teeth all show signs of damage and wear. Given the size of the teeth, Zidek (1992b) estimated that the sharks that shed them were over 2.5 m long. This fish was a marine straggler.

"*Orthacanthus huberi*" Zidek, 1992

Zidek (1992b) assigned an occipital spine and a partial braincase (Fig. 5C-D) to *Orthacanthus huberi*. Estimates by Zidek (1992b) suggest that this fish was about 1 m long. Zidek thought that this was a freshwater fish washing in from the river source of the Kinney estuary. Zidek noted that the lateral line of denticles on the spine suggests that it is from an xenacanthid, which is supported by the isolated braincase. The problem is that the spine and the braincase are isolated, so there is some question of their association as a single taxon. Also, there is much debate in the literature over either a marine versus freshwater habitus of *Orthacanthus* (Fischer et al., 2013). We infer that *Orthacanthus* at Kinney represents a freshwater fish given the smaller number and broken, isolated and worn nature of the fragments.

Peripristis aff. *P. semicircularis* Newberry and Worthen 1866

At the Kinney Quarry, *Peripristis* aff. *P. semicircularis* is represented by one incomplete tooth that shows a fair amount of post-loss wear. This specimen is in the collections at the U.S. National Museum, which is not a collection we surveyed, but we did include it in the list of species. *P. semicircularis* is a marine elasmobranch, found in shallow marine, lagoonal or deltaic deposits (Zidek, 1992b). This fish was a marine straggler in the Kinney estuary.

?*Listracanthus* Newberry and Worthen, 1870

A single spine resembling a *Listracanthus* spine was found at Kinney. This specimen is also in the collections at the U.S. National Museum, which is not a collection we surveyed but we included it in the list of species. The spine is poorly preserved, and only a tentative identification can be made. This infer that this fish was a marine straggler to the estuary.

Sarcopterygii

Sarcopterygians are not abundant in the Kinney Quarry, but the coelacanth are fairly diverse. One species of lungfish, *Sagenodus hlavini*, and three coelacanth species are present (Fig. 3F). The majority of the sarcopterygian specimens are not identifiable to species because most are disarticulated bones that can only be identified as Actinista.

Actinista

Prior to our study, only one species of coelacanth was known from the quarry. Zidek (1975) examined disarticulated pieces, noting that there were coelacanths in the quarry. Schultze (1992) looked at six specimens; one of these was a nearly complete, yet unprepared specimen that is at the Black Hills Institute. During our study six more, nearly complete skeletons have been identified, which represent two new species of coelacanth. There are three species of coelacanth known from the Kinney Quarry, but only a few specimens can be identified to species. There are 25 specimens of coelacanths from the Kinney Quarry, the majority of these consist of disarticulated bones

These fish range in preservational quality. Smaller fish tend to be

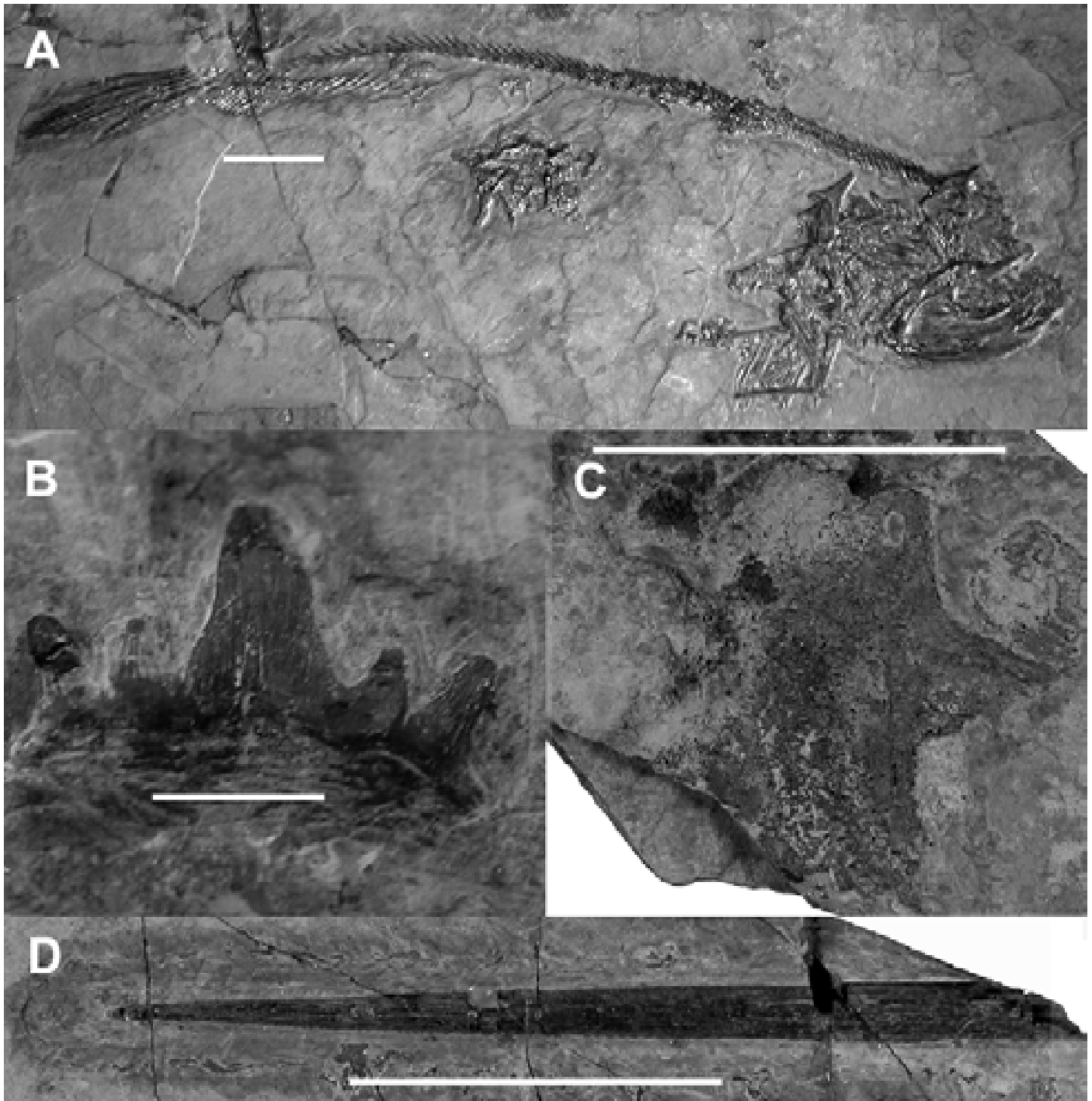


FIGURE 5. Representatives of the chondrichthyan fish from the Kinney Quarry. **A**, *Cobelodus aculeatus*, NMMNH P-19182, shown in Zidek (1992b, fig. 3A). **B**, Tooth of *Symmorium reniform*, NMMNH P-32923. **C-D**, *Orthacanthus huberi*, NMMNH P-19146, **C**, braincase shown in Zidek (1992b, fig. 2) and **D**, spine shown in Zidek (1992b, fig. 1A). All scales are 5 cm.

preserved as nearly complete skeletons, whereas the larger individuals are preserved as isolated, disarticulated bones. The bones of the larger individuals are two to three times larger than similar bones found in the smaller, nearly complete individuals. There are six specimens that are complete enough to identify: one of *Coelacanthidae* gen. et sp. indet. (Schultze, 1992) and four of *Rhabdodermidae* type 1 (Johnson and Lucas, 2001) and one of *Rhabdodermidae* Type 2. *Coelacanth*s are found throughout the fish-bearing layers; 17 of 25 specimens are from units 1 and 2, primarily from the 10 cm of shale (unit 2) immediately above the micrite that floors the quarry (Fig. 2). This is stratigraphically lower than

the majority of the other fish and is contrary to what Huber (1992a) reported, though most of these *coelacanth* specimens have been collected after his research in 1992. As a group the *coelacanth*s are juveniles to sub-adults preserved as articulated specimens, and adult specimens preserved as disarticulated bones.

***Coelacanthidae* gen. et sp. indet.**

There is only one identifiable specimen of this type (Fig 6A). Schultze (1992) commented that the Kinney *coelacanth* has affinities to *Rhabdodermidae* and *Coelacanthidae* but does not match any known

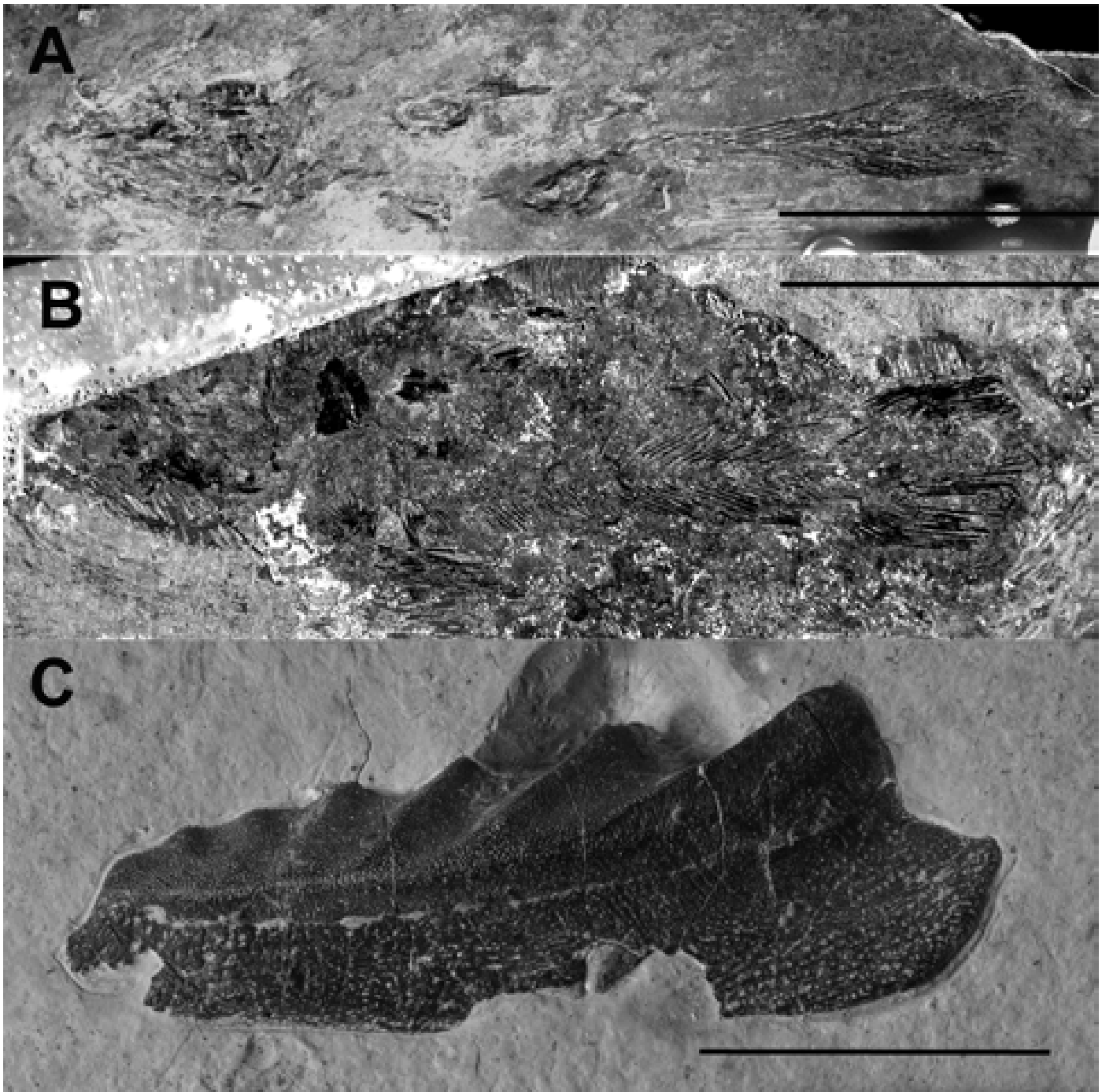


FIGURE 6. Representatives of the sarcopteygian fish from the Kinney Quarry. **A**, Schultze's (1992) coelacanth. **B**, Rhabdodermidae type 1 (Johnson and Lucas, 2001), NMMNH P-31392. **C**, *Sagenodus hlavini* tooth plate, CMNH 8311. Scale bars in **A** and **B** are 3 cm, and in **C** is 1 cm.

Paleozoic coelacanth species. A similar coelacanthid is found in a freshwater deposit in the Red Tanks Member of the Bursum Formation at Carrizo Arroyo in central New Mexico (Johnson and Lucas, 2004). This may suggest that the coelacanthid from Kinney was a freshwater straggler. This also accounts for the rarity of this fish at the Kinney Quarry. The complete specimen is from unit 2, the black shales just a few cm above the basal limestone.

Rhabdodermidae Type 1

A small coelacanth that is rare but more abundant than the Coelacanthidae gen. et sp. indet is present at the Kinney Quarry (Fig. 6B). This fish shows affinities to other late Paleozoic rhabdodermid

coelacanths, resembling *Cardiosuctor* and *Synaptotylus* (Echols, 1963; Lund and Lund, 1985; Johnson and Lucas, 2001). It is quite distinct from the other Kinney coelacanths, and its fossils derive from very low in the section, suggesting that they also represent a marine visitor in the Kinney estuary. *Cardiosuctor* is known from the marine sections of the Bear Gulch Limestone (Lund and Lund 1985). *Synaptotylus* is known from marine sections in Kansas (Schultze, 1996).

Kues and Lucas (1992) argued that units 1-4 of the Kinney Quarry were deposited in an estuary prograded over by a delta front, and that the water was becoming fresher towards the top of the stratigraphic column (Fig. 2). This may indicate that the rhabdodermid coelacanths, which are primarily found in units 1 and 2, preferred relatively marine conditions.

However, most of the specimens are isolated bones, which suggests that there was significant decomposition of these fish prior to deposition. These fish thus could have been transported post-mortem to the site of deposition. These conclusions are contrary to Forey (1991), who interpreted all Late Carboniferous coelacanths as freshwater. The differential preservation of the different sizes of these fish may indicate that the Kinney estuary was a juvenile feeding ground for a marine migrant species, which could have had some form of diadromy.

Rhabdodermidae Type 2

At Kinney, there is a third distinct morphology of coelacanths that resembles *Rhabdoderma elegans* and its relatives, and is not as abundant as type 1. *Rhabdoderma elegans* is a euryhaline species found in freshwater deposits (Forey, 1991), brackish water deposits (Mazon Creek: Charest, 2006; Hamilton and Robinson: Echols, 1964) and marine deposits (Bear Gulch: Lund and Poplin, 1999). The Kinney Rhabdodermidae type 2 are also found very low in the section. For the same reason that we interpret type 1 as marine we interpret type 2 as a marine migrant in the Kinney estuary.

Dipnoi

Sagenodus (Proceratodus) hlavini Zidek, 1975

Sagenodus hlavini (Fig. 6C) is known at Kinney from one specimen, an incomplete tooth plate and jaw fragment that is fractured and worn from post-mortem travel. This specimen is in the collections at Cleveland Museum of Natural History, which is not a collection we surveyed but we did include it in the list of species. *Sagenodus* have been interpreted as a marine fish (Schultze and Chron, 1997). This is an abundant genus during the Late Pennsylvanian (Kemp, 1996; Schultze and Chorn, 1997), so it has been interpreted as a euryhaline species that could easily move between fresh and saltwater environments, which explains its essentially worldwide distribution during the Late Pennsylvanian. The worn and fragmented nature of the tooth plate and jaw fragment from Kinney suggest postmortem transport and disarticulation. The most likely way to produce this amount of disarticulation and wear would be if it came in with the sediments being washed in from the fluvial source. Clearly, the fossil of this species was washed into the estuary and it was not member of the fauna.

Summary of Species Paleoecology

The fish estuarine residents, marine migrants, freshwater migrants are all categories of fish that complete part of their lifecycle as members of the estuary fauna. The fish from Kinney that fit that description are *Acanthodes kinneyi*, the *Platysomus* group, *Amphicentrum jurgenai*, *Schizolepis manzanitaensis*, palaeoniscoids types 5 and 6 and the "small scale" paleoniscoid, *Cobelodus aculeatus*, and Rhabdodermidae types 1 and 2 (Table 4). The less abundant and poorly preserved of these fish were likely living in the shallow, marginal waters outside the snapshot of exposed rocks at the quarry and dying and being washed into the deeper waters, where they were preserved. This group of fish encompasses a size range from some of the smaller fish in the estuary to the largest resident of the estuary. They were filling many of the niches found today in an estuarine setting.

The marine stragglers to the Kinney estuary were *Symmorium reniforme*, *Peripristis* aff. *P. semicircularis*, *Listracanthus*, adult Rhabdodermidae type 1 and type 2. These fish were coming into the estuary to feed and losing teeth or dying in the estuary, leaving evidence of their visits. The chondrichthyan taxa are found in marine deposits at other localities (Zidek, 1992b). The coelacanths are endemic to the Kinney estuary.

The few apparently freshwater species of fishes--"*Orthacanthus huberi*," *Sagenodus hlavini*, cf. *Bourbonnella* sp. and Coelacanthidae gen. et sp. indet.--were washing into the estuary post-mortem. These

fish were living and dying in the fluvial source of the estuary, and their remains were being carried by the fluvial source to the estuary, where they were deposited. Both the "*Orthacanthus*" and *Sagenodus* specimens show high levels of post-mortem wear, breakage and disarticulation that suggest transport into the estuary.

There are only a few fish of uncertain origin. These fish are *Pyritocephalus lowneyae*, *Tanyrhynchichthys mcallisteri* and most of the undescribed morphotypes of palaeoniscoids. There is not enough known about these fishes to be able to determine where they were living.

TAPHONOMY

At Kinney we are dealing with a very limited amount of outcrop. The quarry itself only exposes about 11.8 acres of the fish-bearing rocks. When compared to deltaic and estuarine settings this is a small area snapshot of a much larger system. While a prograding delta describes change over time, we are focusing on the lower 3 m of the outcrop, particularly the 0.5-m-thick fish-bearing interval. This gives us a small snapshot in time. Given our small "postage stamp" view in space and time, our taphonomic and paleoecological inferences indicate that an abundant but relatively low diversity fish fauna lived in the Kinney estuary.

When looking at the collections of the Kinney fish one thing stands out--most Kinney fishes are medium-sized. It is rare to find a fish shorter than 4 cm or longer than 25 cm. There is a definite preservational bias against smaller fish. There are two possible reasons for no small fish: (1) the Kinney Quarry is lacking in small fish because they were living in the estuary margins and had little chance to be preserved in the deeper water; or (2) there were no small fish living in the estuary, and those that are found as fossil at Kinney, such as *Pyritocephalus*, represent accidents that were transported in from another source. The absence of larger fish is evidently limited by the size of the ecospace and the complexity of the food web (see below).

PALEOECOLOGY

We can compare the Kinney fish fauna to that of some modern lagoons. For comparison, we chose a well-studied series of lagoons in the same bay in northern Queensland, Australia (Blaber et al., 1989). The fish population dynamics of these lagoons as well as their interactions with the other flora and fauna have been well studied. This system is comparable to Kinney because it is tropical, and in the estuaries studied there is both a marine and freshwater component. Various studies have examined the fish communities of the shallowest water to the deep water of these tropical estuaries. Communities tend to build in the same structures over time and space. So, by looking at the modern tropical lagoonal and estuary communities we can compare them to the community of fish at Kinney, to investigate aspects such as details of the depositional environment.

Blaber et al. (1989) studied fish in different reaches of Albatross Bay and Embley Estuary in northern Queensland, Australia to identify the fish communities in different habitats with in these estuaries. They caught 5 habitats in the bay and estuary and looked at the percentage of fish caught in the different habitats. The mud bottom areas adjacent to mangroves has a trophic composition similar to the specimens found in Kinney Quarry. The majority of the fish trapped are over 10 cm standard length. There were 29 species caught. The three dominant species of actinopterygian fish caught are *Arius proximus*, *Drepane punctata* and *Gerres abbrevias*:

Arius proximus: These fish range in length from 12 to 35 cm. They are 11.31% of the biomass. Blaber and Blaber (1980), in their survey of three Queensland estuaries, show that *Arius* is primarily a detritivore. This is a similar diet to that hypothesized for several species in from Kinney. So, we consider these fish a possible analogue of *Acanthodes kinneyi*. They represent a smaller percentage of the population than *Acanthodes* (Table 5).

TABLE 4. A breakdown of fish by estuarine residency.

Marine Migrant/Freshwater Migrant/Estuarine Resident	Freshwater/Freshwater Straggler	Marine Straggler	Indeterminant
<i>Acanthodes kinneyi</i>	" <i>Orthacanthus huberi</i> "	<i>Symmorium reniforme</i>	<i>Pyritocephalus lowneyae</i> ,
<i>Platysomus</i> group	<i>Sagenodus hlavini</i>	<i>Peripristis</i> aff. <i>P. semicircularis</i>	<i>Tanyrhynchichthys mcallisteri</i>
<i>Amphicentrum jurgenai</i>	cf. <i>Bourbonnella</i> sp.	<i>Listracanthus</i>	Palaeoniscoid Type 1
Palaeoniscoid Type 5	<i>Coelacanthidae</i> gen. et sp. indet		Palaeoniscoid Type 2
<i>Schizolepis manzanitaensis</i>			Palaeoniscoid Type 3
Palaeoniscoid Types 6			Palaeoniscoid Type 4
Palaeoniscoid "small scale"			
<i>Cobelodus aculeatus</i>			
Rhabdoremidae type 1			
Rhabdoremidae type 2			

TABLE 5. Comparison of the fish from Kinney Quarry to their ecological counterparts from the modern Alligator Creek Estuary in Australia. Comparison data come from table 5 of Blaber et al. (1989).

Trophic Level	Kinney Quarry	Percentage	Embley Estuary	Percentage
<i>Durophage</i>	<i>Platysomus</i>	21.4%	<i>Drepane punctat</i>	17.47%
<i>Detritivore</i>	<i>Acanthodes</i>	18.4%	<i>Arius proximus</i>	11.31%
<i>Generalist</i>	<i>Schizolepis</i>	4.8%	<i>Gerres abbreviatus</i>	6.69%
	<i>Amphicentrum</i>	3.6%		
	"Small Scale"	2.8%		
<i>Piscivorous</i>	Palaeoniscidae "Type 5"	0.84%	<i>Acanthopargus berda</i>	0.56%
	<i>Cobelodus</i>	0.67%	<i>Lutjanus argentimaculatus</i>	0.45%

***Drepane punctata*:** These fish range in size from 10 to 40 cm. They are 17.47% of the biomass. These fish are eating hard-shelled benthic invertebrates and small fish. Although *Drepane punctata* is a little bit larger, they may be filling niche similar to that of the *Platysomus* at Kinney. Blaber et al. (1989) note that the juvenile and subadult fish are found in the estuary.

***Gerres abbreviatus*:** These fish range in size from 3 to 8 cm. They are 6.69% of the biomass and are small generalist feeders. This is a similar diet to that hypothesized for several species from Kinney. So, we consider these fish a possible analogue of the generalist feeding palaeoniscoids and they represent the same approximate part of the sample as do the fish from Kinney Quarry (Table 5).

Sheaves (1992) studied the fish of Alligator Creek. He caught 29 species of fish using traps that only selected the fish over 5 cm long. The

result when you select for the larger fish is similar to what is seen in Kinney. The catch method in Blaber et al. (1989) similarly selected for fish over 10 cm. There is significant overlap between the two localities. Within the localities in Blaber et al. (1989) there are different proportions of fish depending on what the habitat is. Their dichotomy between shallow water (under 1 m) and deep water (over 1 m) habitats suggests that Kinney was "deep water" (more than 1 m deep at low tide). Areas with water under 1 m at low tide show a higher abundance of smaller fish. Areas with water over 1 m at low tide show a higher abundance of medium fish. This is similar to what you see in Kinney.

The habitat type of Embley estuary that matches closest with the percentage of trophic niche and estuarine residency is the mudflat associated with the mangroves. While we certainly don't see evidence of a mangrove type of community at Kinney (Feldman et al., 1992; Archer

and Clark, 1992) we do see a high abundance of plant matter washing into the estuary. The sedimentology of the fish-bearing horizons matches up with (possibly anoxic) mud bottoms (Feldman et al., 1992). In estuaries it is common to have a salt water tongue that moves in and out with the tides under the freshwater entering the estuary. This meromictic stratification can reach as many as 9 km upstream, as seen in the Sittie River in Belize (Twilley and Day, 2012). This meromictic stratification can result in anoxic bottom waters in even relatively shallow water. Wolanski et al. (1980) mention colloidal muds in Australian estuaries that move across the bottom as the tides go in and out and produce similar effects to the tidal rhythmites noted by Archer and Clark (1992).

This is also good evidence that Kinney is not an ideal Lagerstätte with respect to fish. When compared to a modern analogue it can only be compared with the deep water portions of the estuary. We know that fish diversity was not lacking in the Late Paleozoic, because we see such deposits as the Bear Gulch Limestone (Mississippian) in Montana, which has 149 fish species present (Lund and Poplin, 1999; Lund et al., 2012). Linton Ohio (Westoll 1944), Hamilton and Robinson Quarries (Zidek, 1988a, b; Schultze and Chorn, 1988) show a similar diversity to Kinney Quarry.

The second reason that may explain the lack of small fish at Kinney is that no small fish lived there, though this is highly unlikely. Fish communities build in a nested fashion from small species to large species (Kodric-Brown and Brown, 1993). A really restricted habitat or shallow water should have a low diversity and small fish. As the habitat becomes larger and/or deeper, there is more space for larger fish. Diversity increases, the small fish diversify and larger fish can be supported, because there is enough productivity in the system to support a larger size fish. The larger the ecospace becomes the larger and more diverse the community can become, and the mean size of fish increases. There are numerous examples of this in the literature.

For example, Tito de Morais and Tito de Morais (1994) studied an estuary in Africa. In water depths up to 3 m, the largest fish they caught was just under 7 cm long. When looking at the estuaries in Queensland, Australia, when the sample is biased to those fish longer than 5 cm, 80% of the diversity is lost. Thus, Kinney is not a good depositional environment for the preservation of small fish and thus not an ideal Lagerstätte for fish. It may be that Kinney only exposes outcrops from the deeper water areas and we are not seeing the laterally equivalent shallow water habitats where the smaller fish would be preserved.

The fish fauna at the Kinney Quarry closely fits that of a relatively deep-water estuarine fish community. This suggests that the water was probably no less than 1, and not more than 10 m in depth. Overall, the depth of water during deposition of the Kinney sediments were probably 5-7 m deep. When the water is deeper than about 10 m at low tide, there is another shift in the community structure, because there is a reduced effect of the tide and more space for larger predators. The mean length of all fish species increases. The size of the fish from Kinney suggests water depth between 5 and 7 m. The depth would have fluctuated with season and point in the tidal cycle. This interpretation fits with Archer and Clark's (1992) sedimentological interpretation that Kinney was a shallow water estuary feeling tidal rhythms.

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APPENDIX

Specimen data for the fish in this paper. Institutional abbreviations are: **BHI**, Black Hills Institute; **CM**, Carnegie Museum of Natural History; **CMNH**, Cleveland Museum of Natural History; **KUVP**, University of Kansas Museum of Natural History; **NMMNH**, New Mexico Museum of Natural History and Science; **USNM**, U.S. National Museum of Natural History.

ACANTHODII:

CM26555 *Acanthodes* sp., fin spine; *Acanthodes kinneyi*, complete except for tail; CM30726 *Acanthodes* sp., head and shoulder girdle; CM30740 *Acanthodes* sp., complete skeleton; CM47842 *Acanthodes kinneyi*, a (entire specimen) + b (prepectoral region only); most of a is missing & b also is missing 3/5/1999; KUVP-82988 *Acanthodes* sp., articulated posterior; KUVP-82989 *Acanthodes* sp., articulated fish part and counterpart; KUVP-82990 *Acanthodes* sp., squamation part and counterpart; KUVP-82991 *Acanthodes* sp., body, part and counterpart a-c; KUVP-82992 *Acanthodes* sp., squamation part and counterpart; KUVP-82993 *Acanthodes* sp., squamation; KUVP-82994 *Acanthodes* sp., spines, scales, branchial region; KUVP-82995 *Acanthodes* sp., spines, a-c; KUVP-82996 *Acanthodes* sp., 2 spines part and counterpart; KUVP-82997 *Acanthodes* sp., partial fish 5 pieces a-e; KUVP-82998 *Acanthodes* sp., nearly complete fish; KUVP-82999 *Acanthodes* sp., scales spine part and counterpart; KUVP-83000 *Acanthodes* sp., brachial region part and counterpart; KUVP-83001 *Acanthodes* sp., partial fish part and counterpart; KUVP-83002 *Acanthodes* sp., partial fish part and counterpart; KUVP-83003 *Acanthodes* sp., nearly complete fish part and counterpart; KUVP-83004 *Acanthodes* sp., squamation; KUVP-83005 *Acanthodes* sp., squamation part and counterpart; KUVP-83006 *Acanthodes* sp., spine; KUVP-83007 *Acanthodes* sp., squamation; KUVP-83008 *Acanthodes* sp., squamation part and counterpart; KUVP-83009 *Acanthodes* sp., partial fish; KUVP-83010 *Acanthodes* sp., spine part and counterpart; KUVP-83011 *Acanthodes* sp., squamation; KUVP-83012 *Acanthodes* sp., partial fish part and counterpart; KUVP-83013 *Acanthodes* sp., partial fish part and counterpart; KUVP-83014 *Acanthodes* sp., squamation and spine part and counterpart; KUVP-83016 *Acanthodes* sp., partial fish a-c; KUVP-83107 *Acanthodes* sp., articulated fish part and counterpart; KUVP-83015 *Acanthodes* sp., squamation and spine part and counterpart; KUVP-uncat. *Acanthodes* sp., spine; KUVP-uncat. *Acanthodes* sp., partial fish; KUVP-uncat. *Acanthodes* sp., partial fish; KUVP-uncat. *Acanthodes* sp., fin spine; KUVP-uncat. *Acanthodes* sp., spine; NMMNH P-07171 *Acanthodes kinneyi*, partial skeleton; NMMNH P-19208 *Acanthodes kinneyi*, skeleton; NMMNH P-20840 *Acanthodes* sp., part and counterpart; NMMNH P-20841 *Acanthodes* sp., part and counterpart; NMMNH P-20842 *Acanthodes* sp., partial fish; NMMNH P-20843 *Acanthodes* sp., partial fish; NMMNH P-20844 *Acanthodes* sp., partial fish; NMMNH P-20845 *Acanthodes* sp., partial fish; NMMNH P-20846 *Acanthodes* sp., partial fish; NMMNH P-20847 *Acanthodes* sp., partial fish; NMMNH P-20848 *Acanthodes* sp., part and counterpart; NMMNH P-20849 *Acanthodes* sp., partial fish; NMMNH P-20850 *Acanthodes* sp., part and counterpart; NMMNH P-29377 *Acanthodes* sp., partial skeleton; NMMNH P-31020 *Acanthodes* sp., partial fish; NMMNH P-31041 *Acanthodes* sp., scales; NMMNH P-31067 *Acanthodes* sp., scale impressions; NMMNH P-31068 *Acanthodes* sp., scale impressions; NMMNH P-31355 *Acanthodes* sp., partial skeleton; NMMNH P-31356 *Acanthodes* sp., partial skeleton; NMMNH P-31357 *Acanthodes* sp., partial skeleton; NMMNH P-31358 *Acanthodes* sp., partial skeleton; NMMNH P-31359 *Acanthodes* sp., partial skeleton; NMMNH P-31360 *Acanthodes* sp., partial skeleton; NMMNH P-31361 *Acanthodes* sp., partial skeleton; NMMNH P-31387 *Acanthodes* sp., scale impressions; NMMNH P-31388 *Acanthodes* sp., scale impressions; NMMNH P-31389 *Acanthodes* sp., scale impressions; NMMNH P-31390 *Acanthodes* sp., scale impressions; NMMNH P-31393 *Acanthodes* sp., partial skeleton; NMMNH P-31429 *Acanthodes* sp., scales; NMMNH P-31442 *Acanthodes* sp., scales; NMMNH P-32913 *Acanthodes kinneyi*, partial fish part and counterpart; NMMNH P-32914 *Acanthodes* sp., scales

same block at 32915; NMMNH P-32920 *Acanthodes* sp., scales part and counterpart; NMMNH P-32921 *Acanthodes* sp., scales; NMMNH P-32928 *Acanthodes* sp., spine; NMMNH P-32930 *Acanthodes* sp., nearly complete fish part and counterpart; NMMNH P-32932 *Acanthodes kinneyi*, nearly complete fish part and counterpart; NMMNH P-32933 *Acanthodes kinneyi*, scales part and counterpart; NMMNH P-32934 *Acanthodes* sp., spine; NMMNH P-32939 *Acanthodes* sp., scales; NMMNH P-33038 *Acanthodes* sp., scales; NMMNH P-33042 *Acanthodes* sp., scales part and counterpart; NMMNH P-33053 *Acanthodes* sp., scales part and counterpart; NMMNH P-33067 *Acanthodes* sp., scales and spine; NMMNH P-33073 *Acanthodes kinneyi* scales and pectorial girdle, part and counterpart; NMMNH P-33075 *Acanthodes* sp., scales; NMMNH P-36002 *Acanthodes kinneyi*, (1) incomplete fish part & counterpart; NMMNH P-36005 *Acanthodes kinneyi*, (1) incomplete fish part and counterpart; NMMNH P-36007 *Acanthodes kinneyi*, (5) shoulder girdle and scale pieces; NMMNH P-36011 *Acanthodes kinneyi*, (2) part/counterpart shoulder girdle and base of skull; NMMNH P-36012 *Acanthodes kinneyi*, (1) incomplete fish; NMMNH P-36013 *Acanthodes kinneyi*, (1) incomplete fish in 2 pieces; NMMNH P-36016 *Acanthodes kinneyi*, (1) piece w/scales and spine; NMMNH P-36017 *Acanthodes kinneyi*, (1) fish part and counterpart; NMMNH P-36025 *Acanthodes kinneyi*, (1) incomplete fish; NMMNH P-36026 *Acanthodes kinneyi*, (2) spines; NMMNH P-36029 *Acanthodes kinneyi*, (1) incomplete fish; NMMNH P-36032 *Acanthodes kinneyi*, (1) incomplete fish in 2 pieces; NMMNH P-36033 *Acanthodes kinneyi*, (3) incomplete fish part and counterpart in 3 pieces; NMMNH P-36110 *Acanthodes kinneyi*, (1) incomplete fish; NMMNH P-36112 *Acanthodes kinneyi*, (1) pectoral girdle and spines; NMMNH P-36116 *Acanthodes kinneyi*, (2) incomplete fish (part and counterpart); NMMNH P-36118 *Acanthodes* sp., (1) pectoral spine; NMMNH P-36124 *Acanthodes kinneyi*, (1) partial fish; NMMNH P-36126 *Acanthodes kinneyi*, (1) incomplete fish; NMMNH P-36128 *Acanthodes kinneyi*, (1) incomplete fish; NMMNH P-36129 *Acanthodes kinneyi*, (1) very large incomplete fish; NMMNH P-36130 *Acanthodes kinneyi*, (1) incomplete fish.

ACTINOPTERYGII:**PALAEONISCOIDEA:****SCHIZOLEPIS:**

KUVP-83501 *Schizolepis manzanitaensis*, HOLOTYPE; KUVP-83502 *Schizolepis manzanitaensis*, PARATYPE; KUVP-104197 *Schizolepis manzanitaensis*, PARATYPE; CM47851 *Schizolepis manzanitaensis*, patch of scales from dorsal region of body; CM47852 *Schizolepis manzanitaensis*, partial articulated fish a+b; CM47853 *Schizolepis manzanitaensis*, partial articulated fish; CM47854 *Schizolepis manzanitaensis*, partial articulated fish; KUVP-104416 *Schizolepis manzanitaensis*, partial fish; KUVP-uncat. *Schizolepis manzanitaensis*, jaw and scales; KUVP-uncat. *Schizolepis manzanitaensis*, partial fish; KUVP-uncat. *Schizolepis manzanitaensis*, scales; NMMNH P-14359 *Schizolepis manzanitaensis*, complete articulated skeleton (part and counterpart); NMMNH P-14365 *Schizolepis manzanitaensis*, (1) complete articulated skeleton; NMMNH P-14366 *Schizolepis manzanitaensis*, (1) headless articulated skeleton; NMMNH P-29383 *Schizolepis manzanitaensis*, partial fish; NMMNH P-31050 *Schizolepis manzanitaensis*, disarticulated skeleton; NMMNH P-31075 *Schizolepis manzanitaensis*, debris; NMMNH P-31432 *Schizolepis manzanitaensis*, complete skeleton; NMMNH P-32919 *Schizolepis manzanitaensis*, partial fish; NMMNH P-32924 *Schizolepis manzanitaensis*, partial fish part and counterpart; NMMNH P-32927 *Schizolepis manzanitaensis*,

regurgalite; KUVV-uncat. palaeoniscoid, broken scales; KUVV-uncat. palaeoniscoid, bones; KUVV-uncat. palaeoniscoid, disarticulated scales; KUVV-uncat. palaeoniscoid, bones *Platysomus?*; KUVV-uncat. palaeoniscoid, scales; KUVV-uncat. palaeoniscoid, scales; KUVV-uncat. palaeoniscoid, scales; KUVV-uncat. palaeoniscoid, regurgalite; KUVV-uncat. palaeoniscoid, bones; KUVV-uncat. palaeoniscoid, squished skull; KUVV-uncat. palaeoniscoid, disarticulated scales; KUVV-uncat. palaeoniscoid, opercular; KUVV-uncat. palaeoniscoid, articulated scales; KUVV-uncat. palaeoniscoid, regurgalite; KUVV-uncat. palaeoniscoid, bones; KUVV-uncat. palaeoniscoid, scales; KUVV-uncat. palaeoniscoid, regurgalite; KUVV-uncat. palaeoniscoid, scales; KUVV-uncat. palaeoniscoid, scales; NMMNH P-14360 palaeoniscoid, fish part and counterpart; NMMNH P-14361 palaeoniscoid, fish part and counterpart; NMMNH P-14362 palaeoniscoid, fish; NMMNH P-14363 palaeoniscoid, fish part and counterpart; NMMNH P-14364 palaeoniscoid, fish part and counterpart; NMMNH P-14369 palaeoniscoid, fish; NMMNH P-14371 palaeoniscoid, fish part and counterpart; NMMNH P-19167 palaeoniscoid, fragmentary skeleton; NMMNH P-19169 palaeoniscoid, fragmentary skeleton; NMMNH P-19170 palaeoniscoid, fragmentary skeleton; NMMNH P-19175 palaeoniscoid, fragmentary skeleton; NMMNH P-19176 palaeoniscoid, fragmentary skeleton; NMMNH P-19185 palaeoniscoid, fragmentary skeleton; NMMNH P-20851 palaeoniscoid, part and counterpart; NMMNH P-23981 palaeoniscoid, partial fish; NMMNH P-29375 palaeoniscoid, partial fish; NMMNH P-29380 palaeoniscoid, partial skull & body; NMMNH P-29387 palaeoniscoid, partial fish; NMMNH P-29394 palaeoniscoid, partial fish; NMMNH P-29503 palaeoniscoid, articulated scales; NMMNH P-31017 palaeoniscoid, partial fish; NMMNH P-31021 palaeoniscoid, partial fish; NMMNH P-31024 palaeoniscoid, part and counterpart; NMMNH P-31031 palaeoniscoid, fragmentary; NMMNH P-31037 palaeoniscoid, partial fish; NMMNH P-31042 palaeoniscoid, partial fish; NMMNH P-31045 palaeoniscoid, part and counterpart; NMMNH P-31047 palaeoniscoid, fin; NMMNH P-31048 palaeoniscoid, fin; NMMNH P-31049 palaeoniscoid, debris; NMMNH P-31051 palaeoniscoid, partial fish; NMMNH P-31064 palaeoniscoid, partial fish; NMMNH P-31065 palaeoniscoid, partial fish; NMMNH P-31066 palaeoniscoid, partial fish; NMMNH P-31074 palaeoniscoid, partial fish; NMMNH P-31076 palaeoniscoid, partial fish; NMMNH P-31395 palaeoniscoid, partial skeleton; NMMNH P-31396 palaeoniscoid, partial skeleton; NMMNH P-31397 palaeoniscoid, partial skeleton; NMMNH P-31398 palaeoniscoid, partial skeleton; NMMNH P-31399 palaeoniscoid, partial skeleton; NMMNH P-31400 palaeoniscoid, partial skeleton; NMMNH P-31423 palaeoniscoid, partial skeleton; NMMNH P-31424 palaeoniscoid, partial skeleton; NMMNH P-31425 palaeoniscoid, partial skeleton; NMMNH P-31426 palaeoniscoid, partial skeleton; NMMNH P-31427 palaeoniscoid, partial skeleton; NMMNH P-31428 palaeoniscoid, partial skeleton; NMMNH P-31430 palaeoniscoid, partial skeleton; NMMNH P-31431 palaeoniscoid, partial skeleton; NMMNH P-31433 palaeoniscoid, partial skeleton; NMMNH P-31435 palaeoniscoid, partial skeleton; NMMNH P-31436 palaeoniscoid, partial skeleton; NMMNH P-31437 palaeoniscoid, partial skeleton; NMMNH P-31438 palaeoniscoid, partial skeleton; NMMNH P-31439 palaeoniscoid, partial skeleton; NMMNH P-31440 palaeoniscoid, partial skeleton; NMMNH P-31441 palaeoniscoid, partial skeleton; NMMNH P-32904 palaeoniscoid, articulated scales part and counterpart; NMMNH P-32917 palaeoniscoid, disarticulated skeleton; NMMNH P-32918 palaeoniscoid, complete skeleton part and counter; NMMNH P-32922 palaeoniscoid, articulated scales; NMMNH P-32940 palaeoniscoid, scales; NMMNH P-32943 palaeoniscoid, disarticulated scales; NMMNH P-32944 palaeoniscoid, scales; NMMNH P-32945 palaeoniscoid, scales; NMMNH P-33034 palaeoniscoid, disarticulated scales; NMMNH P-33040 palaeoniscoid, on same block as 33041; NMMNH P-33045 palaeoniscoid, scales; NMMNH P-33046 palaeoniscoid, scales; NMMNH P-33049 palaeoniscoid, scales;

NMMNH P-33050 palaeoniscoid, scales; NMMNH P-33051 palaeoniscoid, articulated scales; NMMNH P-33054 palaeoniscoid, caudal region and tail part and counterpart; NMMNH P-33058 palaeoniscoid, articulated scales; NMMNH P-33059 palaeoniscoid, articulated scales; NMMNH P-33060 palaeoniscoid, articulated scales; NMMNH P-33064 palaeoniscoid, disarticulated; NMMNH P-33065 palaeoniscoid, scales; NMMNH P-33066 palaeoniscoid, head; NMMNH P-33074 palaeoniscoid, partial fish; NMMNH P-33084 palaeoniscoid, scales; NMMNH P-33085 palaeoniscoid, scales; NMMNH P-33086 palaeoniscoid, scales; NMMNH P-33088 palaeoniscoid, scales; NMMNH P-33096 palaeoniscoid, part and counterpart; NMMNH P-34859 palaeoniscoid, (2) incomplete skeleton part and counterpart; NMMNH P-34860 palaeoniscoid, (2) incomplete skeleton part and counterpart; NMMNH P-34861 palaeoniscoid, (1) incomplete skeleton; NMMNH P-36008 palaeoniscoid, (2) fin and scale pieces; NMMNH P-36010 palaeoniscoid, (1) skull bone; NMMNH P-36021 palaeoniscoid, (1) skull bone; NMMNH P-36022 palaeoniscoid, (1) incomplete fish; NMMNH P-36030 palaeoniscoid, (1) incomplete fish part & counterpart; NMMNH P-36031 palaeoniscoid, (1) macerated skull; NMMNH P-36037 palaeoniscoid, (1) skull bone; NMMNH P-36038 palaeoniscoid, (1) dorsal part of caudal fin; NMMNH P-36119 palaeoniscoid, (2) skull fragments (part and counterpart).

AMPHICENTRIDAE:

AMPHICENTRUM:

CM47845 *Amphicentrum jurgenai*; HOLOTYPE a+b whole specimen; CM30727 *Amphicentrum jurgenai*, complete fish; *Amphicentrum* sp., head and part of postcranial skeleton; CM30733 *Amphicentrum jurgenai*, complete; CM30734 *Amphicentrum jurgenai*, a+b partial; CM30735 *Amphicentrum jurgenai*, a+b partial; CM47843 *Amphicentrum jurgenai*, a+b; CM47844 *Amphicentrum jurgenai*, a+b; CM47846 *Amphicentrum jurgenai*, whole specimen nearly complete; KUVV-86162 *Amphicentrum jurgenai*, scales; KUVV-86170 *Amphicentrum jurgenai*, partial fish; KUVV-86174 *Amphicentrum jurgenai*, partial fish; KUVV-86178 *Amphicentrum jurgenai*, articulated fish lacking head; KUVV-uncat. *Amphicentrum jurgenai*, partial fish; NMMNH P-19112 *Amphicentrum* sp., part and counterpart; NMMNH P-29504 *Amphicentrum* sp., partial fish; NMMNH P-32907 *Amphicentrum jurgenai*, partial fish part and counterpart; NMMNH P-32909 *Amphicentrum jurgenai*, partial fish; NMMNH P-36015 *Amphicentrum jurgenai*, (1) skull part and counterpart; NMMNH P-36019 *Amphicentrum jurgenai*, (1) incomplete fish part and counterpart; NMMNH P-36020 *Amphicentrum jurgenai*, (1) incomplete fish; NMMNH P-36113 *Amphicentrum jurgenai*, (6) incomplete fish; NMMNH P-36114 *Amphicentrum jurgenai*, (1) incomplete fish;

PLATYSOMIDAE:

PLATYSOMUS:

CM30728 *Platysomus schultzei*, head and part of postcranial skeleton; CM30731 *Platysomus* sp., portion of postcranial skeleton; CM47831 *Platysomus schultzei*, nearly complete; CM47832 *Platysomus schultzei*, nearly complete with phyllodont tooth plates present; CM47833 *Platysomus* sp., nearly complete headless; CM47834 *Platysomus schultzei*, nearly complete, phyllodont tooth plates present; CM47835 *Platysomus schultzei*, a+b phyllodont tooth plates present; CM47836 *Platysomus schultzei*, whole fish nearly complete phyllodont tooth plates present; CM47837 *Platysomus schultzei*, nearly complete; CM47838 *Platysomus schultzei*, nearly complete, phyllodont tooth plates present; CM47839 *Platysomus schultzei*, nearly complete with phyllodont tooth plates present; CM47840 *Platysomus schultzei*, a+b whole fish phyllodont tooth plates present; CM47841 *Platysomus schultzei*, nearly complete, phyllodont tooth plates present; KUVV-115001 *Platysomus* sp., partial fish; KUVV-86161 *Platysomus schultzei*, partial fish; KUVV-

86163 *Platysomus* sp., partial fish; KUVV-86164 *Platysomus* sp., partial fish; KUVV-86165 *Platysomus* sp., partial fish; KUVV-86166 *Platysomus schultzei*, articulated fish; KUVV-86167 *Platysomus schultzei*, partial fish; KUVV-86168 *Platysomus schultzei*, articulated fish; KUVV-86169 *Platysomus schultzei*, 2 partial fish; KUVV-86171 *Platysomus schultzei*, articulated fish; KUVV-86172 *Platysomus schultzei*, partial fish; KUVV-86173 *Platysomus schultzei*, partial fish; KUVV-86175 *Platysomus schultzei*, articulated fish; KUVV-86176 *Platysomus schultzei*, articulated fish; KUVV-86177 *Platysomus schultzei*, caudal and dorsal fins; KUVV-86179 *Platysomus schultzei*, articulated fish lacking tail; KUVV-86180 *Platysomus schultzei*, articulated fish; KUVV-86181 *Platysomus schultzei*, articulated fish; KUVV-86182 *Platysomus schultzei*, articulated fish; KUVV-86183 *Platysomus schultzei*, partial articulated fish; KUVV-86184 *Platysomus schultzei*, scales; KUVV-86185 *Platysomus schultzei*, articulated fish; KUVV-86186 *Platysomus schultzei*, articulated fish; KUVV-uncat. *Platysomus* sp., fin; KUVV-uncat. *Platysomus* sp., disarticulated fish; KUVV-uncat. *Platysomus* sp., phyllodont tooth plate; KUVV-uncat. *Platysomus* sp., jaw; KUVV-uncat. *Platysomus* sp., phyllodont tooth plate; KUVV-uncat. *Platysomus* sp., regurgalite, KUVV-uncat. *Platysomus* sp., phyllodont tooth plate; KUVV-uncat. *Platysomus* sp., phyllodont tooth plate; KUVV-uncat. *Platysomus* sp., phyllodont tooth plate; KUVV-uncat. *Platysomus* sp., partial fish; KUVV-uncat. *Platysomus* sp., partial skull; KUVV-uncat. *Platysomus* sp., phyllodont tooth plate; KUVV-uncat. *Platysomus* sp., preopercular; KUVV-uncat. *Platysomus* sp., phyllodont tooth plate; KUVV-uncat. *Platysomus* sp., scales; KUVV-uncat. *Platysomus* sp., disarticulated; KUVV-uncat. *Platysomus* sp., disarticulated fish; KUVV-uncat. *Platysomus* sp., regurgalite; KUVV-uncat. *Platysomus* sp., phyllodont tooth plate; NMMNH P-14483 *Platysomus* sp., cast of fish; NMMNH P-19113 *Platysomus* sp., partial fish; NMMNH P-19114 *Platysomus* sp., partial fish; NMMNH P-19115 *Platysomus schultzei*, skeleton impression; NMMNH P-19116 *Platysomus schultzei*, skeleton impression; NMMNH P-19117 *Platysomus schultzei* skeleton impression part and counterpart; NMMNH P-19118 *Platysomus* sp., part and counterpart; NMMNH P-19145 *Platysomus* sp., skeleton; NMMNH P-19194 *Platysomus* sp., skeleton part(a) and counterpart(b); NMMNH P-19195 *Platysomus schultzei*, skeleton impression; NMMNH P-29373 *Platysomus* sp., partial fish; NMMNH P-29374 *Platysomus* sp., partial fish; NMMNH P-31046 *Platysomus* sp., partial fish; NMMNH P-31058 *Platysomus* sp., partial fish; NMMNH P-31059 *Platysomus* sp., partial fish; NMMNH P-31060 *Platysomus* sp., partial fish; NMMNH P-31061 *Platysomus* sp., partial fish; NMMNH P-31062 *Platysomus* sp., partial fish; NMMNH P-31063 *Platysomus* sp., partial fish; NMMNH P-31069 *Platysomus* sp., partial fish; NMMNH P-31070 *Platysomus* sp., partial fish; NMMNH P-31071 *Platysomus* sp., partial fish; NMMNH P-31072 *Platysomus* sp., partial fish; NMMNH P-31376 *Platysomus* sp., part and counterpart; NMMNH P-31377 *Platysomus* sp., partial skeleton; NMMNH P-31378 *Platysomus* sp., partial skeleton; NMMNH P-31379 *Platysomus* sp., part and counterpart; NMMNH P-31380 *Platysomus* sp., part and counterpart; NMMNH P-31381 *Platysomus* sp., part and counterpart; NMMNH P-31382 *Platysomus* sp., partial skeleton; NMMNH P-31383 *Platysomus* sp., part and counterpart; NMMNH P-31384 *Platysomus* sp., partial skeleton; NMMNH P-31385 *Platysomus* sp., partial skeleton; NMMNH P-31386 *Platysomus* sp., 2 fish partial skeletons; NMMNH P-32901 *Platysomus* sp., partial fish; NMMNH P-32903 *Platysomus* sp., partial fish part and counter part; NMMNH P-32905 *Platysomus* sp., fins part and counterpart; NMMNH P-32906 *Platysomus* sp., partial fish; NMMNH P-32912 *Platysomus* sp., partial fish; NMMNH P-32915 *Platysomus* sp., skull elements on same block as 32914; NMMNH P-32916 *Platysomus* sp., partial fish part and counterpart; NMMNH P-32931 *Platysomus* sp., partial fish; NMMNH P-32936 *Platysomus* sp., phyllodont tooth plate part and counterpart; NMMNH P-32938 *Platysomus* sp., skull elements; NMMNH P-32942 *Platysomus* sp., phyllodont tooth plate; NMMNH

P-32946 *Platysomus* sp., phyllodont tooth plate part and counterpart; NMMNH P-32999 *Platysomus* sp., partial fish; NMMNH P-33000 *Platysomus* sp., partial fish; NMMNH P-33036 *Platysomus* sp., disarticulated skull and phyllodont tooth plate; NMMNH P-33039 *Platysomus* sp., body; NMMNH P-33047 *Platysomus* sp., scales; NMMNH P-33048 *Platysomus* sp., scales; NMMNH P-33062 *Platysomus schultzei*, whole fish part and counterpart; NMMNH P-33069 *Platysomus* sp., disarticulated fish; NMMNH P-33078 *Platysomus* sp., phyllodont tooth plate; NMMNH P-36004 *Platysomus schultzei*, (1) incomplete fish pt/counter pt in 3 pieces; NMMNH P-36014 *Platysomus schultzei*, (7) incomplete fish in 7 pieces; NMMNH P-36027 *Platysomus* sp., (1) incomplete fish; NMMNH P-36028 *Platysomus* sp., (1) incomplete fish; NMMNH P-36036 *Platysomus* sp., (1) incomplete fish part/counterpart in 5 pieces; NMMNH P-36039 *Platysomus* sp., (1) incomplete fish in 9 pieces; NMMNH P-36105 *Platysomus* sp., (1) disarticulated fish; NMMNH P-36106 *Platysomus schultzei*, (3) incomplete fish; NMMNH P-36107 *Platysomus* sp., (1) disarticulated fish; NMMNH P-36109 *Platysomus schultzei*, (3) incomplete fish part and counterpart; NMMNH P-36121 *Platysomus* sp. (4) disarticulated fish; NMMNH P-36122 *Platysomus schultzei*, (1) incomplete fish; NMMNH P-36123 *Platysomus schultzei*, (3) incomplete fish; NMMNH P-36125 *Platysomus schultzei*, (1) incomplete fish; NMMNH P-36127 *Platysomus* sp., (1) incomplete fish; NMMNH P-36131 *Platysomus schultzei*, (4) incomplete fish; NMMNH P-36132 *Platysomus schultzei*, (1) incomplete fish.

CHONDRICHTHYES:

NMMNH P-14720 Chondrichthyes, SEM figured dermal denticle.

COBELODUS:

NMMNH P-19180 *Cobelodus* sp., partial skeleton; NMMNH P-19182 *Cobelodus aculeatus*, mostly complete skeleton (on display); NMMNH P-33090 *Cobelodus* sp., partial skeleton; NMMNH P-36024 *Cobelodus* sp., (1) incomplete skeleton in 7 pieces.

SYMMORIUM:

CM47847 *Symmorium reniforme*, tooth; NMMNH P-19206 *Symmorium reniforme*, tooth; NMMNH P-32923 *Symmorium reniforme*, tooth.

ORTHACANTHUS:

NMMNH P-19146 *Orthacanthus huberi*, braincase and spine.

LISTRACANTHUS:

USNM-187132 *Listracantus* sp., partial dermal spine.

PERIPRISTIS:

USNM-187133 *Peripristis* aff. *semicircularis* partial tooth.

SARCOPTERYGII:

DIPNOI:

CMNH-8311 *Sagenodus hlavini*, tooth plate and part of jaw.

ACTINISTA:

KUVV-82700 Actinista, parashenoid and isolated elements; KUVV-82701 Actinista, parashenoid; KUVV-82702 Actinista, articulated specimen 6 pieces; KUVV-uncat. Actinista, scales; KUVV-uncat. Actinista, broken neural arch; KUVV-uncat. Actinista, scales + bone; KUVV-uncat. Actinista, neural and hemal arches; KUVV-uncat. Actinista, scales and bone; NMMNH P-31391 Actinista, scapula; NMMNH P-32937 Actinista, neural arch; NMMNH P-33076 Actinista, bone;

NMMNH P-36117 Actinista, (3) neural arches; NMMNH P-36115 Actinista, (1) articulated neural and hemal arch.

COELACANTHIDAE:

BHI-1581 Coelacanthidae, nearly complete skeleton.

RHABDODERMIDAE:

TYPE 1: KUVU-uncat. Rhabdodermidae Type 1, partial skeleton; NMMNH P-29379 Rhabdodermidae Type 1, partial skeleton; NMMNH P-31392 Rhabdodermidae Type 1, partial skeleton; NMMNH P-31434 Rhabdodermidae Type 1, partial skeleton.

TYPE 2: KUVU-uncat. Rhabdodermidae Type 2, partial skeleton; KUVU-uncat. Rhabdodermidae Type 2, dermal armor.