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Synopsis of the Pennsylvanian System in the Sandia Mountains, New Mexico, USA

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SYNOPSIS OF THE PENNSYLVANIAN SYSTEM IN THE SANDIA MOUNTAINS, NEW MEXICO, USA

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Abstract—Pennsylvanian sedimentary rocks in the Sandia Mountains, New Mexico, form an approximately 586 m thick composite stratigraphic section. The succession includes the Atokan–Desmoinesian Sandia Formation, which consists primarily of siliciclastic strata that include coals deposited in a fluviodeltaic to shallow-marine environment, the Desmoinesian Gray Mesa Formation, which is composed of marine carbonate rocks, and the Desmoinesian and Virgilian Atrasado and Bursum Formations, which are siliciclastic and carbonate strata representing nonmarine and shallow-marine environments. Age determinations are based on fusulinid and conodont faunas. Animal and plant fossils occur throughout the Pennsylvanian succession. Plant fossils record a significant climatic change across the Middle–Late Pennsylvanian boundary from humid/subhumid to semiarid conditions, which were drier than the concurrent changes in coal basins of central Pangea. The Sandia Mountains section records three tectonic pulses of the Ancestral Rocky Mountain orogeny: Atokan, Missourian, and late Virgilian–Wolfcampian. We contend that Late Paleozoic glacioeustasy, long considered the primary driver of Pennsylvanian sedimentation, was overprinted by local tectonic and microclimatic changes in this area.

INTRODUCTION

The Sandia Mountains are a late Cenozoic uplift of the Basin and Range physiographic province located immediately east of Albuquerque (Fig. 1). Extending north–south for nearly 27 km (17 mi) and east–west for approximately 13 km (8 mi), the Sandia Mountains are a basement-cored, fault-block mountain range that rose during Neogene time. The core of the range is primarily the Mesoproterozoic Sandia granite, overlain by upper Paleozoic strata that make up most of the eastern dip-slope of the range.

The purpose of this paper is to summarize the stratigraphy, petrography, paleontology, depositional systems, biostratigraphy, and Pennsylvanian geologic history of the Sandia Mountains, which is presented in detail by Lucas et al. (2026). Pennsylvanian marine and nonmarine sediments were deposited when the area was in the western part of the Pangean paleoequatorial belt. The studied stratigraphic sections extend from the southernmost end of the Sandia Mountains at Carlito Spring to a complete section measured on fault blocks near Placitas at the northern end of the range (Fig. 1). The current synopsis of the detailed study by Lucas et al. (2026) presented here focuses on: (1) a summary for each lithostratigraphic unit (brief lithologic description, depositional environment, and paleontology/age); and (2) a brief review of the paleobotany, paleogeography, and controls on deposition of Pennsylvanian strata. Sandstone classification follows Pettijohn et al. (1987). In this paper, the term fossil assemblage refers to fossils from a particular lithostratigraphic unit or to fossils seen in thin section and implies no detailed paleoecological or biochronological significance.

PENNSYLVANIAN LITHOSTRATIGRAPHIC NOMENCLATURE

The Pennsylvanian strata in the Sandia Mountains are the (ascending) Sandia, Gray Mesa, Atrasado, and Bursum Formations (Figs. 2 and 7). The Gray Mesa Formation can be locally divided into the Elephant Butte, Whiskey Canyon, and Garcia Members (in ascending stratigraphic order). The Atrasado Formation can be divided into the (ascending) Bartolo, Amado, Tinajas, Council Spring, Burrego, Story, Del Cuerto, and Moya Members. For each formation, we describe a characteristic stratigraphic section in well-exposed areas. Unless noted otherwise, all descriptions apply to Pennsylvanian strata exposed in the Sandia Mountains.

SANDIA FORMATION

Lithostratigraphy

The Sandia Formation is a mixed siliciclastic-carbonate unit lying at the base of the Pennsylvanian section across much of northern and central New Mexico. The Sandia Formation represents the oldest synorogenic deposits of the Ancestral Rocky Mountain (ARM) orogeny in this portion of the state (Krainer and Lucas, 2013a). The Sandia Formation ranges in thickness from 15 to 124 m (49–407 ft). It overlies either Proterozoic basement nonconformably or Mississippian strata disconformably, and is conformably overlain by the Gray Mesa Formation. The Sandia Formation is composed dominantly of siliciclastics and is characterized by distinct lateral changes in thickness and facies (Toomey, 1953; Catacosinos, 1962; Krainer and Lucas, 2013a). The thinnest section of the Sandia Formation is exposed at Caballo Canyon, where it is only 15 m (49 ft) in thickness and rests on Proterozoic basement. The thickest section is the reference section at Doc Long Picnic Area, with

a thickness of 124 m (407 ft; Catcosino, 1962; Krainer and Lucas, 2013a; Toomey, 1953).

The Sandia Formation comprises the following lithofacies: (1) structureless conglomerate with mostly quartzite clasts; (2) conglomeratic sandstone displaying trough crossbedding; (3) sandstone exhibiting different bedform types; (4) shale/siltstone, gray to dark gray, some coal-bearing and containing plant-fossil fragments (including *Cordaites* and *Neuropteris*);

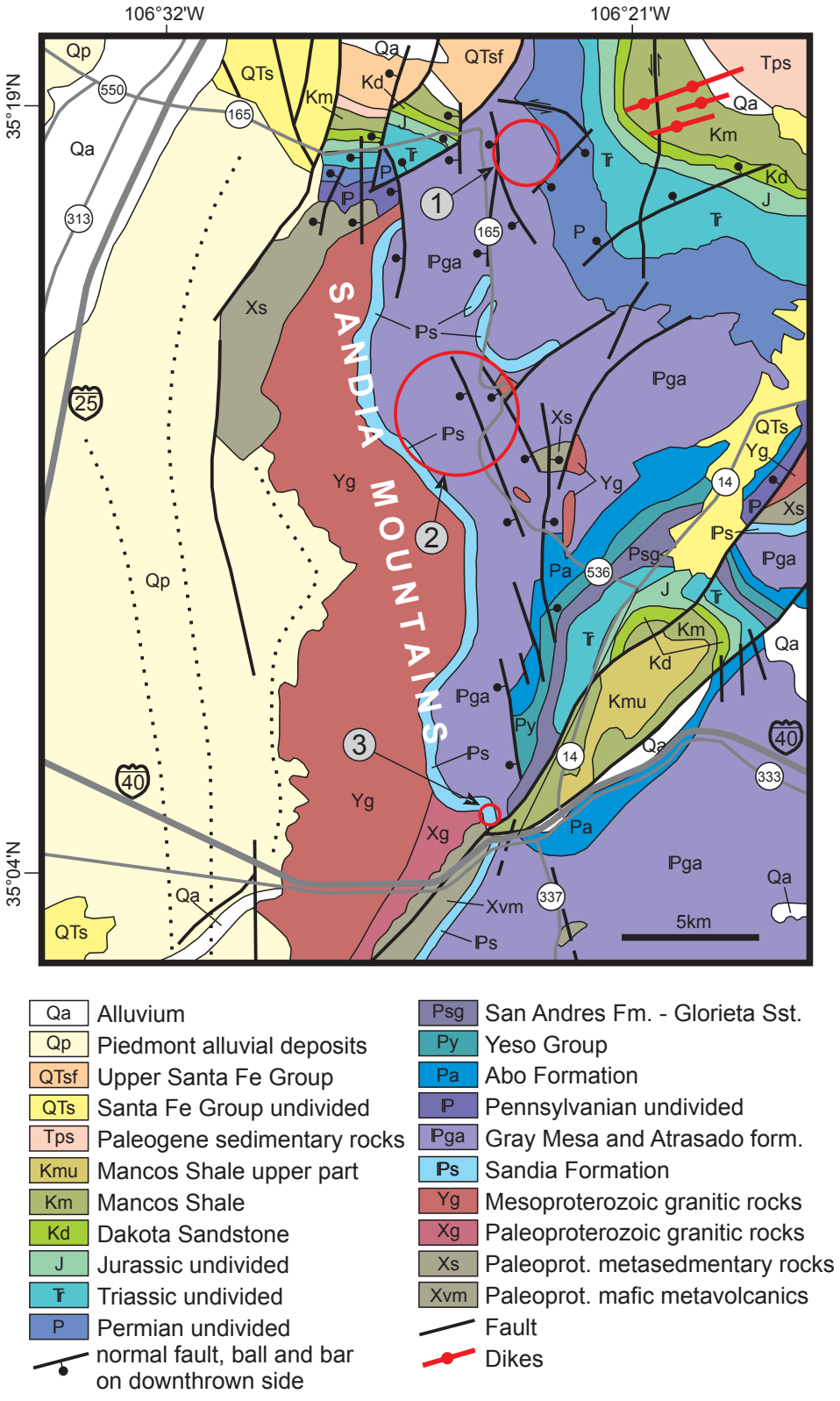


FIGURE 1. Map of Pennsylvanian outcrop area in Sandia Mountains. Locations of Pennsylvanian sections measured and described by Lucas et al. (2026) are in three numbered (red) circles: 1) Near Placitas (Tecolote, Montezuma, Placitas, Tunnel Springs); 2) Along Sandia crestline and dipslope to Doc Long (La Luz Trail, Palomas Peak, Sandia Crest, Sandia reference, Tejano Highway, and Tejano Canyon); and 3) Carlito Spring. Geology after NMBGMR (2003).

and (5) thin- to medium-bedded fossiliferous limestone containing crinoid fragments, brachiopods, and bryozoans.

Pennsylvanian outcrops across State Road 536 (Sandia Crest Highway) from the Doc Long Picnic Area are well known as one of the best exposed and most easily accessible outcrops of the Great Unconformity in the Sandia Mountains (Fig. 3). Here, the base of the Sandia Formation rests nonconformably on Proterozoic Sandia granite and was designated the lectostratotype (reference section) of the Sandia Formation by Krainer et al. (2011).

At this location (Fig. 3), the Sandia Formation comprises three parts: (1) a lower, heterolithic unit (69-m [226-ft]-thick) and composed of alternating shale/siltstone, sandstone, pebbly sandstone, sandy (quartz sand) limestone and limestone; (2) a middle, limestone unit (21-m [69-ft]-thick) and composed of bedded limestone; and (3) an upper sandstone unit (34-m [112-ft]-thick) and composed of coarse-grained, crossbedded sandstone, rare limestone and covered intervals. These subdivisions are not recognizable in other studied sections in the Sandia Mountains.

Sedimentary Petrography

The Sandia Formation is dominated by quartz arenite in the lower part and sublitharenite, subarkose, and lithic arenite in the upper part. Three sandstone types are distinguished, all containing moderate to relatively high proportions of detrital feldspar grains, (1) mixed siliciclastic-carbonate sandstone containing some sedimentary rock fragments, fossils and calcite cement, (2) sandstone, well-sorted, cemented by calcite, but lacking fossils, and (3) sandstone that is cemented by quartz overgrowths and contains small amounts of matrix; calcite cement and fossils are absent.

Typical limestone microfacies are bioclastic wackestone to floatstone, phylloid algal wackestone to floatstone, crinoidal wackestone, bioclastic grainstone, oolitic grainstone and wackestone, and rare crinoidal rudstone and brachiopod rudstone. The fossil assemblages are diverse (see below, and Krainer et al., 2011; Lucas et al., 2026).

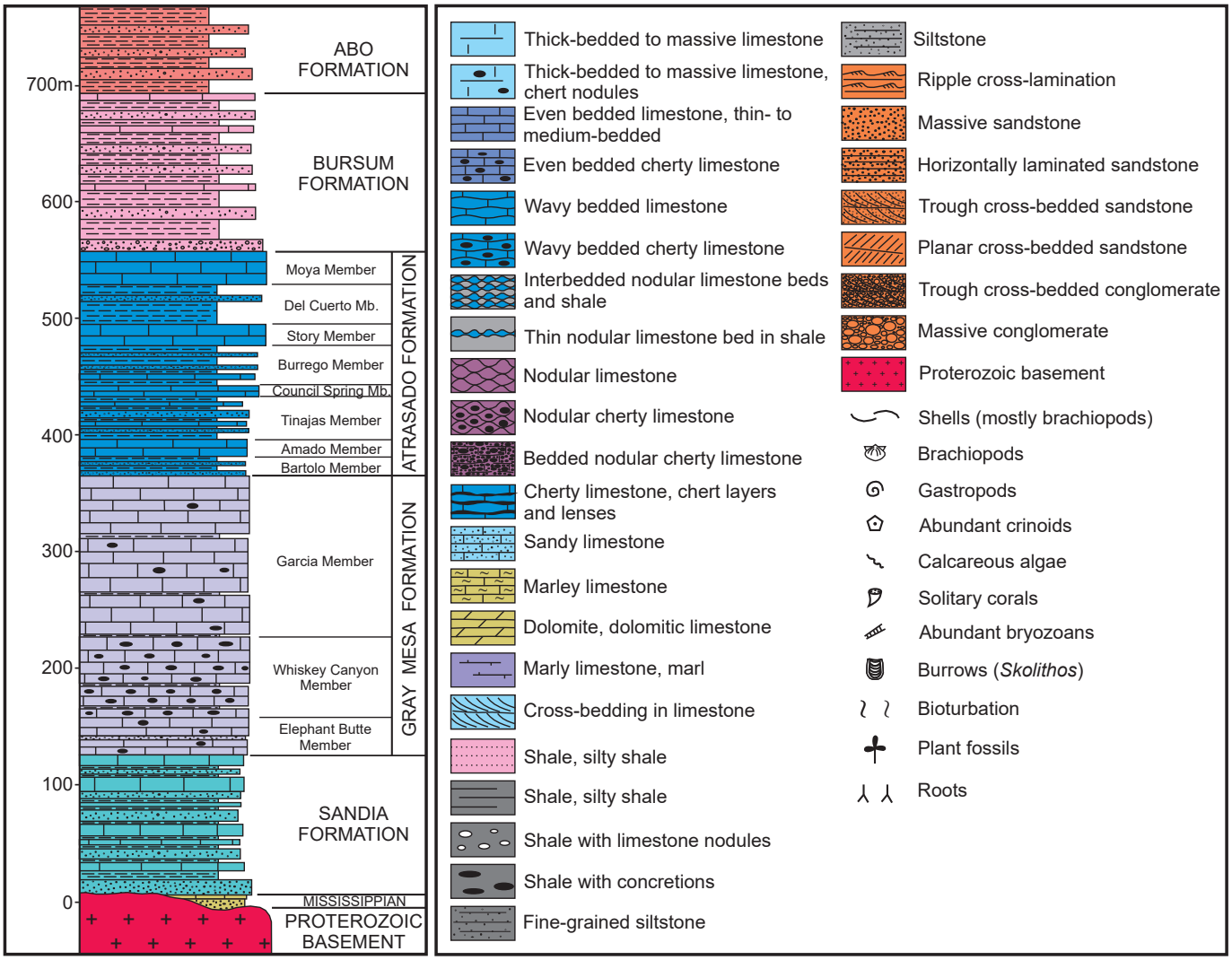


FIGURE 2. Generalized summary of Pennsylvanian lithostratigraphic units in Sandia Mountains (left) and legend for lithologic symbols in measured stratigraphic sections (right). In these sections (but not shown on this figure), the black dots with acronyms (example: THE 30) are levels at which petrographic and/or microfossil samples were collected. Also, in the following sections, the colored boxes with acronyms and numbers indicate fossil localities.

Paleontology and Age

In outcrop, brachiopods and crinoid columnals are common macrofossils (see Northrop, 1961, and Kelley and Northrop, 1975, for lists of taxa). All Sandia Formation microfacies contain fossil assemblages of moderate to high diversity; the most abundant fossils are fragments of crinoids, bryozoans, and brachiopods with less abundant smaller foraminifers, fusulinids, ostracods, gastropods, bivalves, brachiopod spines, and rare trilobites. Locally, phylloid algae and sponge spicules are abundant. Macrofossil plants are also locally abundant. These plant fossils indicate a Middle Pennsylvanian age, and consist mostly of taxa attributed to wetland habitats or paleosols developed in areas with high moisture content. Microfossils–palynomorphs, algae, foraminiferans, and conodonts–indicate that most of the formation is of Atokan age, and locally Desmoinesian in age in the uppermost part.

Depositional Systems

At the Doc Long section, the Sandia Formation is composed of alternating shale, sandstone, pebbly sandstone, fossiliferous limestone, and covered (probably shale) intervals that indicate deposition in mostly marine environments ranging from high-energy shoreface to low-energy deeper shelf (Krainer et al., 2011). Pebbly sandstone and sandstone are interpreted as fluvial/delta plain to upper shoreface deposits (Krainer et al., 2011; Lucas et al., 2026). Muddy limestone units (wackestone and floatstone) are interpreted as deposited in a deeper, open-marine shelf environment below wave base; grainstone and rudstone with marine fossils indicate shallow, high-energy, open-marine environments (Krainer et al., 2011; Krainer and Lucas, 2013a).

North of Sandia Peak (Caballo Canyon and Tecolote sections), the Sandia Formation is entirely to dominantly siliciclastic with few limestone intervals. From Sandia Peak towards the south, there are many covered intervals, limestone, rare sandstone, and very rare conglomerate. In the southernmost section at Carlito Spring, siliciclastic sedimentary rocks dominate, and conglomerate is much more abundant than farther north.

Shale and siltstone were deposited in a wide range of environments, from nonmarine fluviodeltaic (commonly containing fossil plant fragments) to brackish (containing the inarticulate brachiopod *Lingula*) to shallow-marine (containing brachiopods and crinoid fragments) and middle-outer shelf settings below wave base (see Lucas et al. 2026).

GRAY MESA FORMATION

Lithostratigraphy

The Gray Mesa Formation is the most extensively exposed and lithologically consistent Pennsylvanian stratigraphic unit in the Sandia Mountains. It is essentially a succession of limestone beds that caps the crestline of the mountains and is exposed at diverse outcrops from Placitas to Carlito Spring. The thickest section of the Gray Mesa Formation is exposed near Placitas, with a total thickness of 233 m (765 ft). At the Tecolote B section (Fig. 4), the exposed thickness is 170 m (558 ft), and it is mostly composed of limestone with some covered intervals, rare conglomerate and thin sandstone interbeds. Covered intervals (shale/siltstone) are 0.3–5.1 m (1–17 ft) in thickness and are more abundant and thicker in the upper part. Lithologic subdivision into Elephant Butte, Whiskey Canyon, and Garcia Members is possible.

Sedimentary Petrography

Limestones of the Gray Mesa Formation are dominated by microfacies with muddy (micrite) textures, ranging from wackestone to floatstone. Poorly to well-washed grainstone, packstone, and rudstone are rare. All limestone units contain a diverse assemblage of echinoderm (crinoid) fragments, bryozoans, and brachiopods, the most common fossils.

The most common microfacies at Palomas Peak, where the upper part of the Gray Mesa Formation contains a thick, algal bioherm, is bioturbated bioclastic wackestone with a diverse fossil assemblage. Other microfacies include crinoidal packstone, grainstone, floatstone, and rudstone. Typical microfacies of the mound facies at Palomas Peak are phylloid algal bafflestone to floatstone, with some cementstone. Thin, subarkosic sandstone beds occur in the upper part of the formation at sections Tecolote B and C (Fig. 4; for details see Lucas et al., 2026).

Paleontology and Age

Fossils observed in outcrop include crinoid fragments, brachiopods, bryozoans, and calcareous algae. The macrofossil assemblages thus mostly include representatives of the typical late Paleozoic shelly benthos. The general composition of the assemblages is very similar to those of the Gray Mesa Formation to the south (e.g., Manzano, Los Pinos, Oscura, Little San Pascual, Fra Cristobal, and Caballo Mountains). All Gray Mesa microfacies display a diverse fossil assemblage. The most abundant fossils are echinoderm (crinoid), bryozoan, and brachiopod fragments. Recrystallized phylloid algal fragments are locally abundant. Less abundant are smaller foraminifers, fusulinids, ostracods, gastropods, bivalves, and trilobites. All microfacies contain abundant, undeterminable recrystallized skeletons. Biostratigraphic age determinations based on fusulinids and conodonts indicate that the formation is primarily early Desmoinesian (see Lucas et al. 2026).

Depositional Systems

Gray Mesa Formation facies are uniform over more than 300 km (186 mi), extending from the Nacimiento and Jemez Mountains and southern Sangre de Cristo Mountains in the north to the Mud Springs and Caballo Mountains in southern New Mexico (Nelson et al., 2013b; Lucas et al., 2021a, 2022, 2023a, 2023b, and 2025). Grainstone, packstone, and rudstone represent high-energy deposits, some of the shoreface

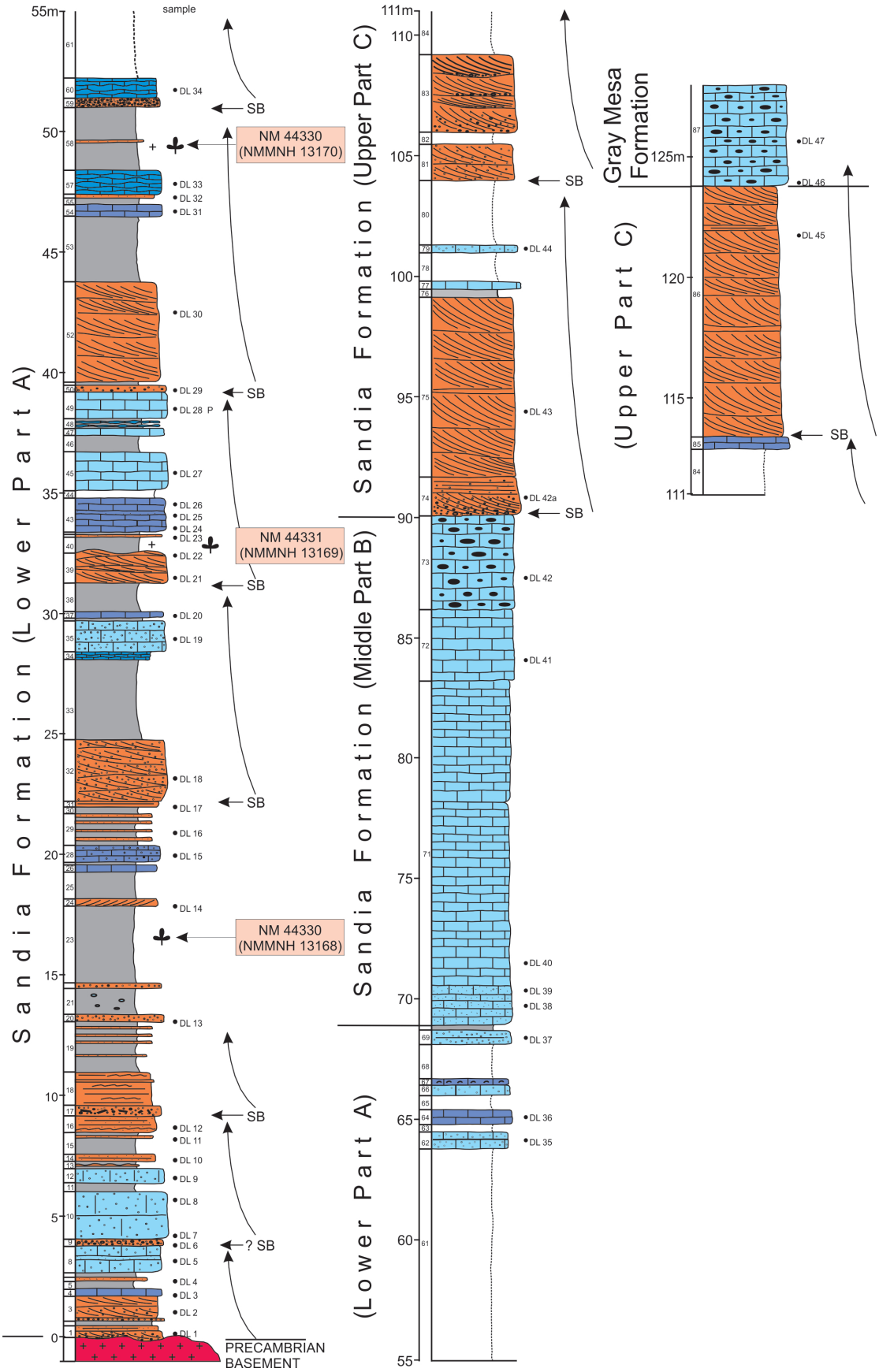


FIGURE 3. Reference section (lectostratotype) of Sandia Formation near Doc Long Picnic Area. See Figure 2 for the legend to lithologic symbols. Base of section at 374,655 m and 3,893,220 m, and top at 374,755 m and 3,892,938 m (UTM Zone 13, datum NAD 83). Note that the width of lithologic units in the diagram is schematic and largely reflects less resistant units (narrower) and more resistant (wider) lithologies.

and others formed during storms (Nelson et al., 2013b; Lucas et al., 2021a). As in other Gray Mesa Formation sections of New Mexico, the most common type of grain association of Gray Mesa limestone is "bryonoderm extended" (Beauchamp 1994; bryozoans, echinoderms [crinoids], and brachiopods), with subordinate corals and smaller and larger benthic foraminifers. A detailed microfacies analysis of the Pennsylvanian stratigraphy (i.e., Sandia, Gray Mesa, and Atrasado Formations) in the Socorro area is presented in Krainer et al. (2017) and provides a good representation of the microfacies of these units across their outcrop areas. From this, we interpret the Gray Mesa limestone to represent deposition in low-energy, deeper shelf environments (below wave base), as indicated by the predominance of muddy textures (e.g., bioclastic mudstone, wackestone, and floatstone) and the local abundance of chert, in part derived from siliceous sponges. Sediments that formed in high-energy settings, such as grainstone, packstone, and rudstone, are rare, as are subaerial exposure surfaces (see details in Lucas et al., 2026; Lucas et al., 2014, 2021a, 2022, 2023a, and 2023b).

Previous studies of the Gray Mesa Formation in the Sandia Mountains (e.g., Wiberg and Smith, 1994) documented multiple 2–12 m (7–39 ft) thick shoaling-upward successions interpreted as meter-scale, glacioeustatic transgressive–regressive (T–R) cycles capped by subaerial exposure surfaces. However, Lucas et al. (2026) emphasize that heterogeneous facies relationships, strong lateral variability, and discontinuous stratigraphic geometries across the Sandia Mountains complicate this interpretation. In this view, tectonic and depositional processes are interpreted to exert a primary control on stratigraphic preservation and expression, potentially obscuring or overprinting any glacioeustatic signal (see Lucas et al., 2026).

Smith (1999) measured and described stratigraphic sections along the Tejano Canyon roadcuts, assigning the exposed units to the Sandia and Madera Formations. He interpreted meter-scale stratigraphic alternations within the succession as depositional cycles driven by long-eccentricity orbital forcing. Lucas et al. (2026) reexamined these exposures and interpreted them as fault-bounded blocks of the Sandia Formation and the Gray Mesa Formation, concluding that the Atrasado Formation is neither fully nor adequately represented at this locality. Biostratigraphic data, including fusulinid and conodont assemblages, indicate predominantly early Desmoinesian ages for the Tejano Canyon sections, suggesting a more limited temporal span than inferred by Smith and colleagues in the absence of sufficiently precise age control to assign durations to possible cycles (see Lucas et al., 2026).

ATRASADO FORMATION

The Sandia Mountains expose the northernmost outcrops of the Atrasado Formation. The southernmost outcrops are in the Oscura Mountains of Socorro County, about 140 km (87 mi) to the south. Across this outcrop belt, the Atrasado Formation is a mixed siliciclastic and carbonate unit divisible into eight persistent members, in ascending stratigraphic order: Bartolo, Amado, Tinajas, Council Spring, Burrego, Story, Del Cuerto, and Moya (Nelson et al., 2013a).

Lithostratigraphy

No single, well-exposed, continuous, or nearly continuous stratigraphic section of the Atrasado Formation has yet been identified in the Sandia Mountains. However, in the Placitas area, on the very northern end of the range, a complete stratigraphic section of the Atrasado Formation can be pieced together across two fault-block sections, each containing overlapping portions of the formation (Fig. 5). At the Montezuma fault section, the upper part (Council Spring–Moya Members) is exposed, and the Tecolote D section exposes the lower-middle part (Bartolo–lower part of the Burrego Member). This results in a total Atrasado Formation thickness of approximately 160 m (525 ft). At these stratigraphic sections, members of the Atrasado Formation are represented in ascending stratigraphic order:

1. Bartolo Member (33-m [108-ft]-thick) is a mixed siliciclastic and carbonate unit, extensively covered but with some prominent limestone and sandstone beds. The base contains a mixed siliciclastic and carbonate sandstone (Coyote Sandstone of Herrick, 1900), which is common at many outcrops (Lucas et al., 2021a).
2. Amado Member (12-m [39-ft]-thick) is composed primarily of micritic limestone and phylloid algal limestone.
3. Tinajas Member (19-m [62-ft]-thick) is a succession of covered intervals and limestone. The Tinajas Member is 59-m (194-ft)-thick at the Tejano Canyon B section. Covered intervals, which are 1–5 m (3–16 ft) in thickness, most likely represent shale or siltstone.
4. Council Spring Member (17-m [56-ft]-thick) is in the lower part, mostly limestone—bedded muddy limestone (wackestone), and nodular limestone (wackestone) intercalated with shale, and overlain by phylloid algal limestone.
5. Burrego Member (43-m [141-ft]-thick) mostly contains shale/siltstone that is partly covered. Thin sandstone and limestone beds are intercalated.
6. Story Member (19-m [62-ft]-thick) is composed of thin-bedded (at the base), thick-bedded (above), and massive algal limestone (phylloid algal wackestone to floatstone).
7. Del Cuerto Member (10-m [33-ft]-thick) is composed of three trough-crossbedded sandstone units that are separated by covered intervals and red mudstone/siltstone.
8. Moya Member (7-m [23-ft]-thick) is composed of crinoidal limestone, cherty micritic limestone (wackestone), and medium-bedded phylloid algal limestone, a phylloid algal wackestone to floatstone that is 3 m (10 ft) in thickness.

The base of the overlying Bursum Formation is red mudstone, overlain by trough-crossbedded sandstone.

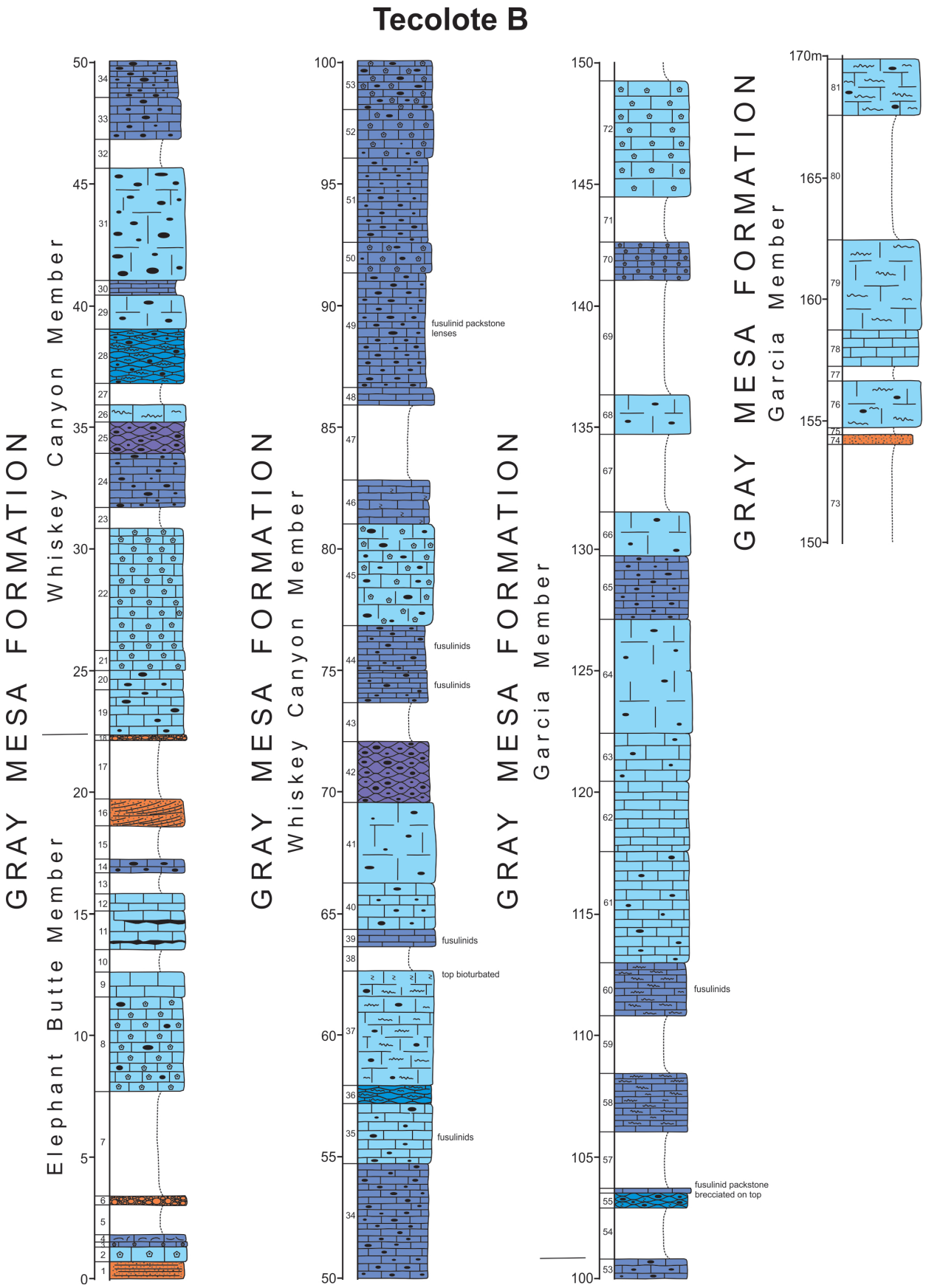


FIGURE 4. Tecolote B section. See Figure 2 for the legend to lithologic symbols. Base of section at 372,444 m and 3,908,553 m, and top at 372,746 m and 3,908,564 m (UTM Zone 13, datum NAD 83). Note that the width of lithologic units in the diagram is schematic and largely reflects less resistant units (narrower) and more resistant (wider) lithologies.

Sedimentary Petrography

In sections of the Atrasado Formation from the southern Sandia Mountains to the Manzano Mountains (Cedro Peak, Tejano Canyon, Sol se Mete, Priest Canyon), Los Pinos Mountains, Cerros de Amado, Sierra Ladrones, and at the type section in the Lucero uplift, the limestone facies is uniform and composed mainly of microfacies types that are micrite rich, including bioclastic wackestone to floatstone, fusulinid wackestone to floatstone, crinoidal wackestone to packstone, and phylloid algal floatstone; grainstone is rare. All microfacies, except bindstone, which is very rare, contain diverse fossil assemblages that are generally similar to those of the limestone facies of the Gray Mesa Formation (see Lucas et al., 2026). Sandstones of the Atrasado Formation are subarkose and sublitharenite; very rare quartzarenite and lithic arenite are also present.

Paleontology and Age

Atrasado Formation strata in the Sandia Mountains yield diverse and abundant fossils. Most marine fossils are typical late Paleozoic shelly benthos—brachiopods, bryozoans, and crinoids—that have been relatively little studied in the Sandia Mountains and not interpreted for their biostratigraphic significance herein. The nonmarine fossils (i.e., macroscopic and microscopic plant remains) are of limited use because of their sparseness and preservational problems, partly reflecting the coarseness of the enclosing deposits, and thus are of little biostratigraphic value. However, macro- and micro-flora show a distinct floristic change within the Atrasado Formation across the Middle–Late Pennsylvanian boundary in the Amado Member, particularly an increased abundance and diversity of xeromorphic plants (e.g., conifers), concurrent with the disappearance of lycospores and the rise of marattialean tree-fern spores. The age of the Atrasado Formation strata is based on foraminifers (especially fusulinids) and conodonts that range from late Desmoinesian through Virgilian (see Lucas et al., 2026).

Depositional Systems

Atrasado Formation deposition was mostly marine, although some nonmarine strata with sedimentary features typical of fluvial environments are present, lacking marine fossils, and containing terrestrial plant fossils. The siliciclastic-dominated members vary in thickness laterally, and the facies indicate that their spatial distribution and lateral variability were controlled by both tectonically created accommodation and climate. Sediment transport and deposition likely occurred under the more subhumid to semiarid climatic intervals of glacial cycles (Cecil and Dulong, 2003).

Texture and fossil assemblages (dominantly bioclastic wackestone to floatstone, with subordinate fusulinid wackestone to floatstone, phylloid algal floatstone and crinoidal wackestone to packstone and rare peloid-intraclast grainstone, bioclastic grainstone to packstone and rudstone) indicate that Atrasado Formation limestones were deposited under dominantly low-energy conditions in shallow, open to (rarely) restricted (bindstone, bioclastic mudstone) normal marine depositional environments as indicated primarily by fossil content (see Lucas et al. 2026). Pedogenic overprinting of some limestones, indicating subaerial exposure, is not observed. There is little lateral variation in thickness and facies of limestone-dominated members, indicating deposition during periods when sea level rose and flooded lowland areas, permitting wider, spatially continuous space for carbonate formation on a uniform, wide shelf. This occurred during drier parts of glacial cycles, when the effects of tectonic activity were apparently reduced, which were more apparent during intervals of greater moisture availability associated with increased siliciclastic deposition (Cecil and Dulong, 2003).

BURSUM FORMATION

Lithostratigraphy

Across much of New Mexico, the Bursum Formation is generally less than 100 m (328 ft) in thickness and composed of alternating nonmarine red beds and shallow-marine shale, sandstone, conglomerate, and limestone. This unit is characterized by distinct lateral variations in thickness and facies, which we consider to represent four members: (1) Oso Ridge Member, (2) Red Tanks Member, (3) Bruton Member, and (4) Laborcita Member (Lucas and Krainer, 2004; Krainer and Lucas, 2009, 2013b). We assign the Bursum Formation in the Sandia Mountains to the Red Tanks Member, which is mostly composed of alternating nonmarine red beds (shale, sandstone, conglomerate) and a lesser amount of shallow-marine shale, sandstone, conglomerate, marl, and limestone that locally form well-developed cycles.

We measured four complete stratigraphic sections of the Bursum Formation in the Sandia Mountains, two near Placitas, one just east of the range near Sandia Park, and one in Tijeras Canyon at Los Pinos. Lucas et al. (1999) published data on the Placitas sections, which they therein referred to as the "Wild Cow Formation." Lucas et al. (2014) presented data on the section near San Pedro. The Bursum Formation measures 139 m (456 ft) at the Crest of Montezuma (Fig. 6), the thickest Red Tanks Member identified to date. At Sandia Park, the Bursum Formation is 90-m (295-ft)-thick, and at Los Pinos, it is 125 m (410 ft) in thickness. The Bursum Formation is considerably thinner (12 m [39 ft]) at Placitas, just a few kilometers from the Montezuma stratigraphic section.

An unconformity, locally angular, is developed at the base of the Bursum Formation across its outcrop belt, resulting from a tectonic pulse of the ARM orogeny. Carbonate conglomerates at Placitas and Montezuma B contain clasts of reworked bioclastic mudstone and wackestone, indicating that older carbonate sedimentary rocks were reworked, either from the underlying Atrasado Formation or from the Bursum Formation (i.e., intra-Bursum tectonics). The nonmarine Abo Formation unconformably overlies the Bursum Formation here and elsewhere.

At the Montezuma B section (Fig. 6), nodular micritic and cherty wackestone at the top of the Moya Member of the Atrasado Formation are overlain by mostly covered grayish-red shale-siltstone (4.3 m [14 ft]) and grayish-red, trough-crossbedded sandstone (1 m [3 ft]), assigned to the Bursum Formation. Here, the Bursum Formation is 139.2 m (456.7 ft) in thickness. We draw the boundary with the overlying Abo Formation on top of the highest marine limestone interval of the Bursum Formation (Fig. 6, unit 34). Mudstone-shale/siltstone is purplish red with green bands or color mottling. Shale in the upper part of the section contains calcrete nodules. Covered (shale) intervals are 0.8–26.5 m (2.6–86.9 ft) in thickness. Conglomerate is present as a 1.2 m (3.9 ft) thick interval near the base that is a limestone-pebble conglomerate with tabular beds 0.2–0.3 m (0.7–1.0 ft) in thickness.

Sandstone consists of: (1) trough-crossbedded units (0.8-4.5 m [2.6–14.8 ft] thick), some bioturbated (*Skolithos*); (2) horizontally laminated beds, 0.5–1.1 m (1.6–3.6 ft) in thickness; (3) thin beds and lenses intercalated in red mudstone, 1.4 m (4.6 ft) in thickness; and (4) ripple-laminated, fine-grained beds in the uppermost part, that partly display climbing ripples (1.1 m [3.3 ft]), and partly occur as thin, ripple-laminated, fine-grained beds intercalated with thin red mudstone beds.

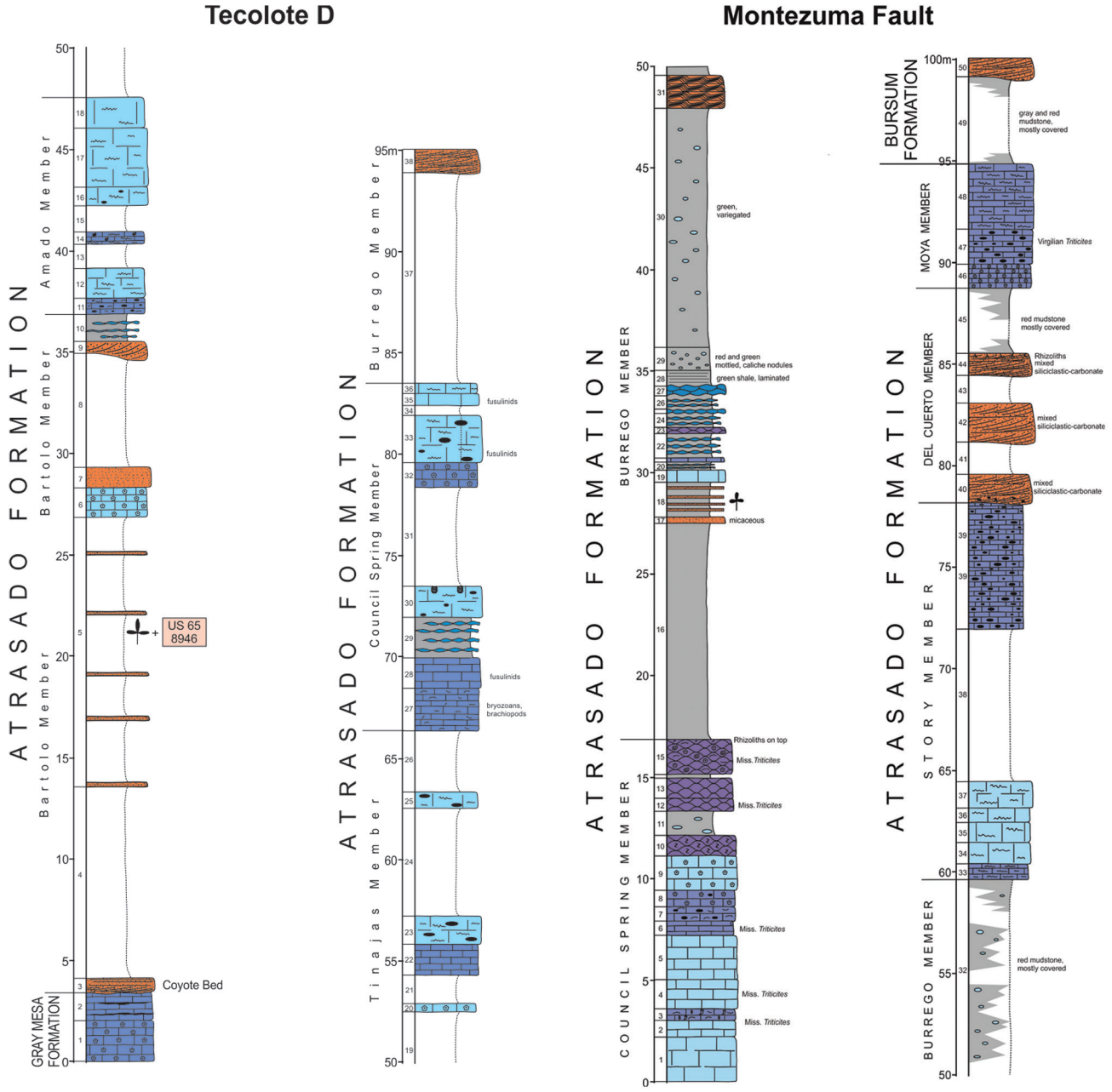


FIGURE 5. Montezuma fault Atrasado section (right) and Tecolote D Atrasado section (left). Montezuma fault section: base of section at UTM zone 13, 372851, 3,909,446 and top at 373,089, 3,909,487 (datum NAD 83). Tecolote D Atrasado section (right). Tecolote D section base at 372,623 m and 3,908,689 m, and top at 372,830 m and 3,908,454 m (UTM Zone 13, datum NAD 83). See Figure 2 for the legend to lithologic symbols. Note that the width of lithologic units in the diagram is schematic and largely reflects less resistant units (narrower) and more resistant (wider) lithologies.

Limestone is rare and restricted to the upper 60 m (197 ft) of the formation. The following lithotypes are observed: (1) thin (0.5-m [1.6-ft]-thick), nodular wackestone units; (2) thin (0.2 and 0.5 m [0.7 and 1.6 ft] in thickness), nodular calcretes, based on their texture and content, including siliciclastic grains; and (3) thin (2.1-m [6.9-ft]), nodular bioclastic wackestone beds interbedded with shale.

Sedimentary Petrography

At Placitas, fine-grained sandy conglomerates are clast-supported, moderately sorted, well-washed, and composed of subangular to subrounded grains that include various types of carbonate rocks. A coquina bed is composed of bioclastic wackestone, grainstone, and rudstone. Marine limestone nodules are composed of micrite that contains few marine fossils. Thin beds composed of bioclastic wackestone to packstone to rudstone contain a diverse fossil assemblage.

Sandstone is mostly subarkose, and rarely sublitharenite. However, sandstones at Los Pinos and Sandia Park are arkosic (Lucas et al., 2014). At Sandia Park, the dominant limestone microfacies is bioclastic wackestone to packstone, and subordinately packstone to rudstone, all containing diverse marine fossil assemblages (Krainer and Lucas, 2013b; Lucas et al., 2026). At the Los Pinos section, bioclastic wackestone (locally grading into packstone and rudstone) with a diverse fossil assemblage is the most abundant microfacies. Two thin limestone beds in the uppermost part are composed of ostracodal wackestone and wackestone to floatstone containing a low-diversity fossil assemblage (see Lucas et al., 2014).

Paleontology and Age

Given its mixed marine–nonmarine origin, Bursum Formation strata in the Sandia Mountains yield diverse and abundant fossils, mostly marine invertebrates, including bryozoans, echinoderms (crinoids), brachiopods, ostracods, smaller foraminifers, spicules, rare fusulinids, and trilobites; plant fossils are uncommon. The most significant biostratigraphic determination is based on palynology (see below), indicating that Bursum strata at the northern end of the Sandia Mountains contain a Virgilian-age flora, consistent with previous age interpretations. The macroflora collected nearby is not age-diagnostic and bears a strong environmental overprint of seasonal climate. Some vertebrate fossils were reported at Placitas. Thin section examination of Bursum Formation limestone reveal abundant algae and smaller foraminifers; locally, ostracods and fusulinids are abundant.

The Bursum Formation is generally considered to be of Wolfcampian age based on fusulinids. However, as earlier noted by Lucas and Krainer (2004), north of about the latitude of Belen, New Mexico, the Bursum Formation is partly or wholly of Virgilian age. This is borne out by fusulinid data in the Sandia Mountains. The late Garner Wilde identified Virgilian fusulinids in both the Sandia Park and Los Pinos sections (Lucas et al., 2014). This is in agreement with our palynological analyses.

Depositional Systems

The Bursum Formation is an uncomformity-bounded, mixed marine and nonmarine unit between dominantly shallow-marine strata below (Atrasado Formation) and nonmarine red beds (Abo Formation) above. The distinct lateral variations in thickness and facies, particularly at the northern end of the Sandia Mountains, at least in part reflect regional synsedimentary tectonic movements of the ARM (see Krainer and Lucas, 2009, 2013b).

At Placitas, mudstone-siltstone intervals are difficult to interpret because diagnostic fossils are absent. Limestone nodules contain a few fossils, indicating a shallow, restricted-marine setting. These nodules are pedogenically overprinted. Rhizoliths on bed tops also indicate a nonmarine environment.

Fine-grained conglomerates that are moderately sorted and composed of subangular to subrounded grains are interpreted as fluvial channel-fill deposits that formed on a coastal plain; the few invertebrate fossil fragments present are probably reworked from older carbonate rocks. The two coquina beds, composed of bioclastic wackestone, packstone, and rudstone, represent storm layers that accumulated during two short marine flooding events under open, shallow, normal-marine conditions. The muddy limestone bed contains a low-diversity invertebrate fossil assemblage (bioclastic mudstone containing smaller foraminifers, many ostracods, and rare *Efluegelia*), indicating deposition in a shallow, restricted, low-energy environment. The topmost limestone bed is composed of bioclastic wackestone to packstone with a moderately diverse fossil assemblage, indicating deposition in a shallow, open marine environment of normal salinity and low to moderate water turbulence (see Lucas et al., 2026).

As at Placitas, the Bursum Formation at the Crest of Montezuma is dominantly nonmarine with a few thin marine limestone intercalations whose microfacies and fossils indicate deposition in a normal, shallow-marine environment. Locally, as at the type section or sections east of Socorro, upward-fining depositional sequences are well developed and commonly capped by marine limestone flooding events (Krainer and Lucas, 2009, 2013b). Such depositional sequences are absent at Placitas and Montezuma, and are locally present at Sandia Park and Los Pinos.

The significant difference in thickness between the Bursum Formation at Placitas (12 m [39 ft]) and Montezuma (139 m [456 ft]), which are just a few kilometers apart, indicates that the locus of sedimentation was strongly influenced by tectonic creation of accommodation, and that, at Placitas, considerable parts of the Bursum Formation may also have been eroded prior to deposition of the Permian Abo Formation (e.g., Krainer and Lucas, 2009b; Lucas et al., 2026). Such influence on sedimentation is also documented by lateral facies variations among the Montezuma, Sandia Park, and Los Pinos sections (Lucas et al., 2014).

Locally, allocyclic glacioeustatic sea-level fluctuations and linked climatic changes likely were responsible for the formation of fining-upward (transgressive) depositional sequences in the Bursum Formation. Autocyclic processes, such as

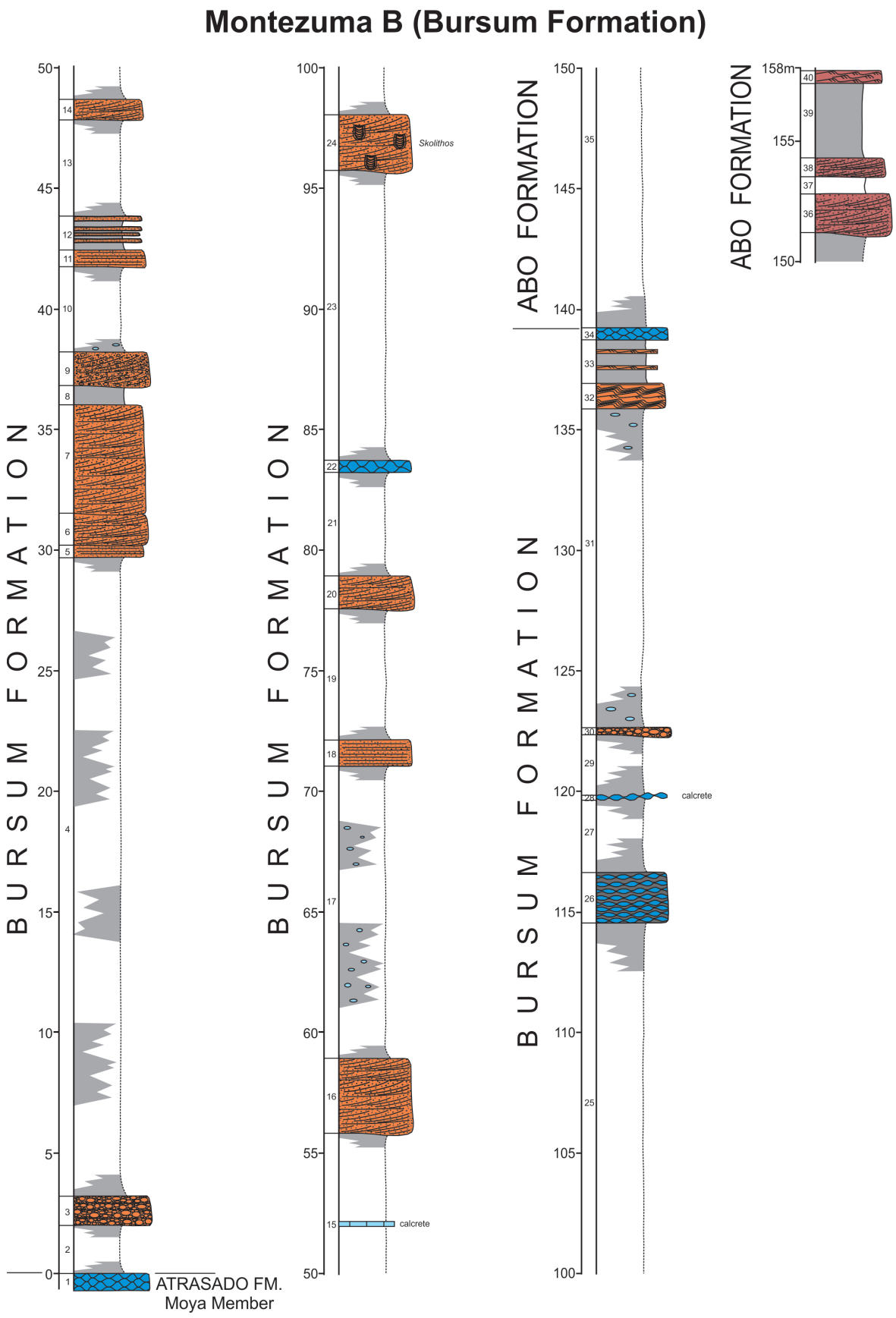


FIGURE 6. Montezuma B Bursum section: base at 373,064 m and 3,909,412 m, and top at 373,558 m and 3,909,036 (UTM Zone 13, datum NAD 83). See Figure 2 for the legend to lithologic symbols. Note that the width of lithologic units in the diagram is schematic and largely reflects less resistant units (narrower) and more resistant (wider) lithologies.

differential subsidence, may have also driven Bursum sedimentation. In general, the pronounced lateral variation in thickness and facies of the Bursum Formation is likely, at least in part, a response to ARM tectonism.

PALEOBOTANY

We undertook a detailed study of paleofloras from the Sandia Mountains, primarily those of the Middle Pennsylvanian Sandia Formation (see Lucas et al., 2026). The composition of the flora during the Middle Pennsylvanian, dominated by hydrophilic taxa, is typical of wet soils (e.g., DiMichele et al., 2023; Lucas et al., 2026), although reflective of a seasonal moisture regime, probably subhumid (in the terms of Cecil, 2003, 6–8 months of the year during which rainfall exceeds evapotranspiration). At times, the regional climate, which was humid enough for organic matter to accumulate over relatively long time spans, and are represented today by thin coal beds or organic-rich shale, mostly 4–5 cm (1.6–2.0 in) in thickness, and of uncertain lateral continuity.

The described floras are almost certainly linked to the wettest phases of glacial cycles, and thus are contemporaneous with peat/coal formation in central Pangea, forming at times of lower sea-level when coastal plains were broadly exposed (Cecil et al., 1985, 2003). In contrast to the coal-basin areas of central Pangea, even the wettest Desmoinesian climates of western Pangea were seasonally dry, likely subhumid (Cecil, 2003), rather than the contemporaneous humid, even occasionally perhumid, conditions that existed in central Pangea. This hypothesis is supported by the character of the sediments (gravels, sandstones, and siltstones, with abundant mica), the sedimentary dynamics (channels with evidence of flash flooding), and the composition of the floras (containing many species uncommon or rarely encountered in central Pangean wetlands).

The most important aspect of this temporal series of plant-fossil assemblages is the marked floristic difference between the Middle and Late Pennsylvanian collections. The older assemblages contain no conifers, in particular, but also no callipterids or noeggerathialeans. These plants appear in the Late Pennsylvanian, where conifer fossils are well preserved, though not abundant, and are consistent with floristic changes across this stage boundary in the Euramerican equatorial region (e.g., DiMichele et al., 2023). Conifers, however, appear in the Middle Pennsylvanian farther to the south in the Desmoinesian floras (Bartolo Formation) in Socorro County. These specimens are rare but indicate the earlier presence of conifers. Quantitative analysis of the Socorro County floras (Schachat et al., 2023) also illustrates an abrupt shift toward drought-tolerant composition across the Middle–Late Pennsylvanian boundary.

The Missourian-age Kinney Quarry flora in the Manzano Mountains contains an abundance of conifers, mixed with wetland elements, such as marattialean tree ferns. Interestingly, no positively identifiable tree-fern remains were recovered from the two Late Pennsylvanian Sandia Mountains floras, which are not far from the Kinney Quarry assemblage. Marattialean foliage occurs in Late Pennsylvanian floras from Socorro County, but is rare in Middle Pennsylvanian floras there, and is similarly uncommon in the Late Pennsylvanian floras from southern New Mexico. This general rarity may reflect a regional climate hostile to these tree ferns, which, although broadly distributed across Euramerica in the Pennsylvanian and early Permian, are still generally associated with environments with high soil moisture.

These Middle Pennsylvanian Sandia Mountains floras are also of interest because many species are uncommon in typical central Pangean coal-bed roof shale floras. Included are such species as *Linopteris neuropteroides*, *Neuralethopteris lindahlii*, and *Sphenopteridium germanicum*, the latter earlier than expected for that species.

Also found in Atokan assemblages are *Lesleya* and *Taeniopteris* at about the same temporal level as in the United States midcontinent region. These indicate a climatic shift that affected western and central equatorial Pangea. Where found in Early and Middle Pennsylvanian strata of the United States, they are typically associated with environments characterized by strong moisture seasonality and possibly thin soils (Leary and Pfefferkorn, 1977; Bashforth et al., 2016).

The Atokan is considered to be a period of generally humid climate at the wettest points of Pennsylvanian glacial cycles, terminated by an abrupt change to more strongly seasonal climates, at all cycle stages in the Desmoinesian, based on coal-basin studies in Europe (Van Hoof et al., 2013), the Appalachian Mountains (Cecil et al., 1985), and the Illinois Basin and midcontinent region (DiMichele et al., 2023). A similar pattern is detected in western Pangea, although muted by the significantly lower abundance of Atokan coal beds and their absence in the Desmoinesian. As with other climatic markers, western Pangea appears more seasonally dry than in the central parts of the supercontinent at all glacial cycle phases. In summary, the Sandia Mountains floras are a significant addition to our growing understanding of the similarities and differences of vegetation across the Pangean tropics and the climatic patterns to which plants are so strongly tied.

PALEOGEOGRAPHY

Many paleogeographic maps of Pennsylvanian New Mexico show the Sandia Mountains located just west of a westward-projecting portion of the Pedernal uplift (e.g., Kues and Giles, 2004; Nelson and Lucas, 2011). Given its proximity, the Pedernal uplift was likely the principal source of clastic sediments transported towards and deposited in the Sandia Mountains during the Pennsylvanian. However, paleocurrent and more petrographic data are needed to establish this with certainty.

Some workers place the area of the Sandia Mountains during the Pennsylvanian in the northwestern part of a late Paleozoic Estancia basin that would have been west of the Pedernal uplift (i.e., Wiberg and Smith, 1994; Smith, 1999). However, as circumscribed, the Estancia basin is well south and east of the Sandia Mountains. Thus, we do not posit that the Pennsylvanian strata in the Sandia Mountains were deposited in a

late Paleozoic Estancia basin. Instead, we contend they were deposited on the northeastern end of a broad shelf of the Orogrande basin (McKee and Crosby, 1975; Nelson and Lucas, 2011; Lucas et al., 2021a).

DEPOSITIONAL CONTROLS

Pennsylvanian sedimentation in the Sandia Mountains reflects the interaction of tectonics, climate, and glacioeustatic sea-level change, consistent with patterns recognized across late Paleozoic basins of western Pangea. Previous local and regional studies have emphasized the role of glacioeustasy in driving cyclic sedimentation (e.g., Soreghan, 1994; Wiberg and Smith, 1994; Raatz and Simo, 1998; Smith, 1999; Wiberg and Smith; Elrick and Scott, 2010). In the Sandia Mountains, field relationships indicate that the stratigraphic expression of cyclicity is variable and locally obscured. Measured sections display substantial lateral variability in thickness and facies, suggesting strong tectonic controls on deposition (see Lucas et al., 2026).

For example, in the Lucero uplift of Valencia County, southeast of the Sandia Mountains, Elrick and Scott (2010) interpreted Gray Mesa Formation strata as recording glacioeustatic cyclicity based on facies and isotopic data. Although similar lithofacies are present in the Sandia Mountains, the stratigraphic sections examined there do not exhibit

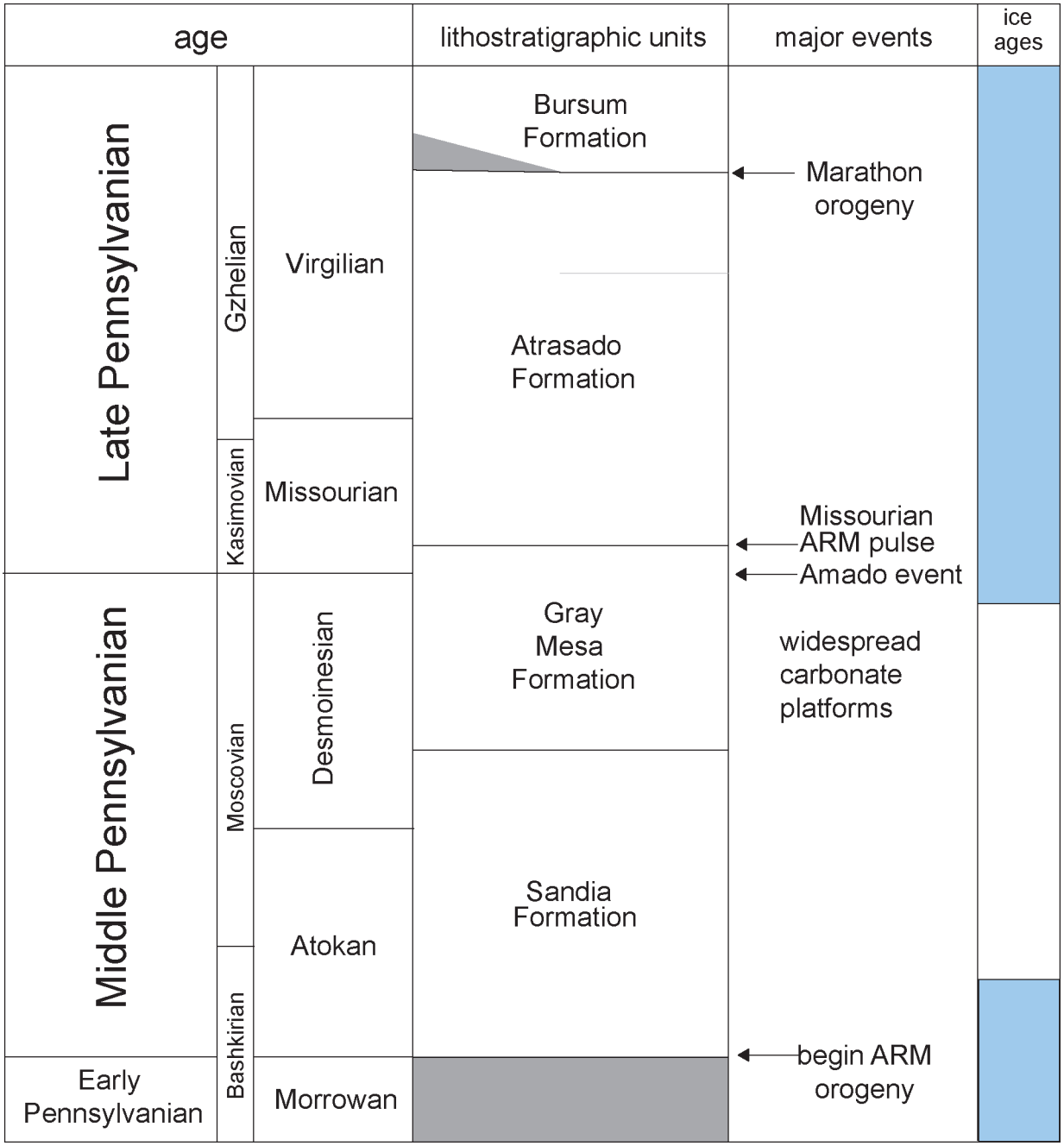


FIGURE 7. Summary of lithostratigraphy, ages, and major Pennsylvanian geologic events in the Sandia Mountains.

laterally continuous, systematically organized cycles of the type described in that study, suggesting a stronger influence of local tectonics and depositional variability.

The only direct analysis of Pennsylvanian sedimentation in the Sandia Mountains is that focused on identifying cycles supposedly forced by glacioeustasy (i.e., Wiberg, 1993; Wiberg and Smith, 1994; Smith, 1999). Of these studies, Wiberg and Smith (1994) examined what they called the Madera Limestone—corresponding approximately to the Gray Mesa Formation—at five localities (Montezuma Ridge, Tunnel Spring, Sandia Crest, Kiawanis Cabin, and Tijeras Canyon), with a maximum thickness of 90 m (295 ft) at Kiwanis Cabin, within which they recognized 15 Transgressive-Regressive (T–R) 4th order cycles. On the basis of a correlation with the Cedro Peak section of Myers (1973) in the nearby Manzanita Mountains, they inferred that their Madera Limestone ranges between the *Beedeina arizonensis* and *Beedeina girtyi* fusulinid zones of Desmoinesian age. Based on this age range, they estimated T–R cycle durations of 200,000 to 473,000 years. From this, they concluded that these cycles are mainly caused by eustatic sea-level changes and that ARM tectonism was a lesser factor in driving deposition.

In contrast to the work of Wiberg and Smith (1994), our measured Gray Mesa Formation sections show either disorganized facies stacking or weakly expressed upward-thickening trends that could be ascribed to depositional cycles. Correlations among measured sections at the bed or bedset level are not repeatable, primarily because marker beds lack continuity. In the Sandia Mountains, we see little evidence for glacioeustatically driven cycles in Pennsylvanian sections, although such cycles are expected and may be masked by local tectonic, erosional, and sedimentary complexities.

We were unable to verify Wiberg and Smith's (1994) correlations of their Gray Mesa Formation sections. Marker beds are not identified, their correlations could not be verified, and are not likely lithostratigraphic because in some instances sandstone beds were correlated with limestones (see Lucas et al., 2026). Inferred cycles were correlated with one another, even though the thicknesses and lithologic composition of beds in any given cycle differ from section to section. Given the methodological uncertainties, we regard their correlations as speculative and question their interpretation of glacioeustasy as the primary control on sedimentation in the Sandia Mountains.

GEOLOGIC HISTORY

Pennsylvanian strata in the Sandia Mountains record deposition during the ARM orogeny along the western margin of equatorial Pangea. The stratigraphic succession reflects the interplay of onlap of the Great Unconformity, ARM tectonics, sediment supply, and regional-to-global climatic influences, including glacioeustatic sea-level fluctuations. As noted above, previous studies have interpreted Pennsylvanian successions in New Mexico as recording glacioeustatically driven cyclicity. In the Sandia Mountains, however, substantial lateral variability and discontinuity complicate recognition of laterally persistent, systematically organized cycles. As a result, we contend that regional signals are difficult to isolate and are interpreted here as variably expressed within a stratigraphic framework that was strongly influenced by local tectonics and sediment supply.

Deposition of Pennsylvanian strata in the Sandia Mountains took place at the northern end of the Orogrande basin on what Lucas et al. (2022) called the Bernalillo–Placitas shelf of that basin. To the east, the Pedernal uplift was the nearest sediment source. That uplift extended north–south from a northern end very close (a few tens of kilometers at most) to a northern terminus of the present Sandia Mountains to a southern terminus south of Alamogordo in Otero County, southern New Mexico. Pennsylvanian-age strata west of the Pedernal uplift exhibit a similar lithostratigraphic architecture over a 270 km (168 mi) transect. These are the Pennsylvanian strata in the Sacramento Mountains, Oscura Mountains, Cerros de Amado, Sierra Ladrones, Los Pinos Mountains, Manzano Mountains, and Sandia Mountains (Lucas et al., 2026). That similarity indicates continuity of some events across the transect (Fig. 7).

Atokan

Mississippian–Lower Pennsylvanian sedimentary rocks were partly eroded prior to deposition of the Sandia Formation, caused by a tectonic pulse that marks the beginning of the ARM orogeny (Fig. 7). The presence of considerable amounts of carbonate sedimentary rock fragments in many Sandia Formation sandstones indicates that besides Proterozoic basement rocks (mostly granite), carbonate rocks of pre-Sandia Formation age were reworked (Mississippian–Lower Pennsylvanian Arroyo Peñasco, Log Springs, and Osha Canyon Formations). These older sedimentary rocks were originally more widespread in the Sandia Mountains before deposition of the Sandia Formation.

In the Sandia Mountains, the Sandia Formation is characterized by pronounced lateral variations in thickness and facies. North of Sandia Peak, the thickness of the Sandia Formation is less than 25 m (82 ft), and the lithofacies are dominantly to entirely siliciclastic. Southward from Sandia Peak, the Sandia Formation is much thicker (53–91 m [174–299 ft]) and contains a higher proportion of limestone and less sandstone and conglomerate.

Unlike the stratigraphic sections along the crest of the Sanida Mountains, outcrops along Tejano Canyon and near the Doc Long Picnic Area are well exposed. There, the thickness of the Sandia Formation varies from 88 m (289 ft), 104 m (341 ft), and 125 m (410 ft), and is composed of alternating siliciclastic (siltstone, sandstone, conglomerate) and carbonate rocks. These distinct lateral variations in thickness and facies over relatively short distances suggest that the locus of sediment deposition was mainly controlled by variable accommodation created by an irregular Proterozoic bedrock surface, and the gradational and possibly intertonguing nature of the Sandia–Gray Mesa Formation contact is probably also in part responsible for these variations in thickness.

Thus, both tectonism and climate were important drivers of deposition in the Sandia Formation. To what extent

glacioeustasy drove Sandia Formation deposition is more difficult to evaluate, although polar glaciation would have concurrently affected both sea level and climate, the latter differently at different latitudes and modified by the topography of the local-to-regional terrestrial environment.

Desmoinesian

Most of the Gray Mesa Formation is early Desmoinesian in age and represents a major change of depositional regime to platform-carbonate-dominated deposition. The most striking feature of the Gray Mesa Formation is the distinct variation in thickness, ranging from 54 m (177 ft) at the Tejano Canyon stratigraphic section 4A, to more than 111 m (364 ft) at Sandia Crest, 122 m (400 ft) at Palomas Peak, and 233 m (764 ft) at Tecolote at the northern end of the Sandia Mountains. These thickness variations likely reflect local differences in crustal subsidence, unconformities in the section, and the irregular topography on which these sediments were deposited.

Microfacies types and biota of the Gray Mesa Formation limestones do not differ significantly from those of Gray Mesa sections elsewhere in New Mexico, indicating a broadly similar depositional setting. The microfacies, as well as the high diversity and composition of the fossil assemblage, indicate limestone formation in an open-marine shelf environment with normal salinity under mostly low-energy conditions below wave base. One exception is a phylloid algal mound near the top of the Gray Mesa Formation at Palomas Peak, which is approximately 20 m (66 ft) in thickness and extends laterally over several hundred meters.

The thickness variations and the presence of thin intercalated sandstone and conglomerate beds in some Gray Mesa Formation sections indicate that tectonic processes influenced sedimentation, resulting in local differences in the rate and extent of subsidence. Tectonics and climate also affected the location and abundance of coarse-grained siliciclastic influx.

Deposition of the Atrasado Formation began during late Desmoinesian time with a shift from dominantly carbonate deposition (Gray Mesa Formation) to mixed clastic-carbonate deposition (Bartolo Member). Particularly striking is the Gray Mesa–Bartolo contact, which is commonly marked by sandstone (i.e., Coyote Sandstone bed) and a sequence boundary possibly reflecting a tectonic event (Lucas et al., 2021a, 2026). This suggests that a Missourian pulse of the ARM (Lucas et al., 2021b) may, in fact, have begun in late Desmoinesian time.

Missourian–Virgilian

The Amado Member of the Atrasado Formation is of earliest Missourian age in the Sandia Mountains. It appears to represent a sedimentary signal of an eustatic event (a rise in sea level) across the Middle–Late Pennsylvanian (Desmoinesian–Missourian) boundary. It was termed the "Amado event" by Lucas et al. (2021b), and is interpreted as a substantial glacioeustatic event correlative with the Swope cyclothem of the U.S. midcontinent region. It marks the onset of a major global environmental event during the Kasimovian/Missourian interval (Heckel, 2008; Rygel et al., 2008), possibly a large-scale deglaciation event in Gondwana. This was a major event that would have affected sea level and climate across the Pangean tropics, including the area where Pennsylvanian strata of the Sandia Mountains were deposited.

Subsequent deposition of the Tinajas Member during the Missourian encompassed nonmarine fluvial, deltaic, shallow-marine, and marginal-marine (estuarine) deposition. These strata likely reflect subhumid climates, which, in combination with differential tectonic uplift/subsidence along the western flank of part of the Pedernal uplift, would have enhanced sediment entrainment and dispersal. This tectonic pulse is seen elsewhere across central New Mexico in Missourian clastic strata in the Manzano and the Sacramento Mountains (Lucas et al., 2021a, b).

The upper part of the Atrasado Formation across the Missourian–Virgilian boundary is dominantly shallow-marine shelfal deposits. A third ARM pulse is marked by deposition of the Bursum Formation and took place across the Virgilian–Wolfcampian boundary. This was a local or regional peak of ARM deformation (Dickinson and Lawton, 2003).

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Sandia alumroot *(Heuchera pulchella)* on Gray Mesa Formation, Sandia Mountains, New Mexico. This rare perennial species is endemic to the mountains of central New Mexico and typically occurs at high elevations above 3,000 m (~9,850 ft). Photograph courtesy of Celestyn Brożek.



Gray Mesa Formation limestone near the top of the La Luz trail. Photograph courtesy of Stan Davis.