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New Mexico Geological Society Guidebook, 53rd Field Conference, Geology of White Sands, 2002, p. 141-157.

A STRATIGRAPHIC HISTORY OF THE TULAROSA BASIN AREA, SOUTH-CENTRAL NEW MEXICO

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ABSTRACT.— Precambrian through Holocene strata are exposed in tilted fault blocks surrounding the modern Tularosa Basin. Precambrian crystalline and metasedimentary rocks are unconformably overlain by a generally transgressive Late Cambrian?/Early Ordovician through Mississippian series of sandstones, carbonates, and shales deposited on the south-dipping Tobosa shelf. During Late Mississippian to Early Pennsylvanian time, Ancestral Rocky Mountain tectonism created the Pedernal uplift and conjugate Orogrande basin. Clastics and carbonates were deposited across the area in cyclic beds, recording high frequency, high amplitude glacio-eustatic sea level fluctuations modified by structural and climatic events. Late Pennsylvanian and Early Permian strata record renewed uplift of the Pedernal and rapid basin subsidence, resulting in the deposition of large volumes of clastic redbeds and carbonates that largely filled the basin and eliminated most topography. Middle Permian transgressive carbonates and evaporites filled remnant topography and record a return to shallow-marine, stable shelf environments. Minor Triassic(?) terrestrial clastics are preserved. Cretaceous units were deposited in an early southern Tethyan suite and a later northern suite connected with the Western Interior Seaway. From Late Cretaceous through Eocene time, the Laramide orogeny reorganized basin geometries, followed by Oligocene-mid Miocene and late Miocene-Holocene crustal extension and volcanism of the Rio Grande rift. The tectonics that created the rift also formed the Basin and Range extensional topography observed today in the Tularosa Basin. Tertiary through Holocene strata include terrestrial clastics derived from uplifts, playa lake deposits, volcanics, and the White Sands gypsum dunes.

INTRODUCTION

The stark beauty and well-exposed, continuous outcrops in south-central New Mexico capture the imagination of geologists and nongeologists alike. In particular, Paleozoic and Cenozoic history is displayed in panoramic detail. Numerous important tectonic and sedimentary concepts have and continue to be born and tested here, while generations of students have learned the science and art of field geology across the areas canyons and ridges. Although a large military presence restricts public access to certain areas, most of the mountain ranges, with the exception of the San Andres, are accessible.

The study area for this report includes the modern-day Tularosa Basin and surrounding Sacramento, San Andres, Jarilla, Franklin, and Hueco Mountains and Otero Mesa (Fig. 1), an area roughly bounded by the major cities of El Paso, Las Cruces and Alamogordo and covering ~20,000 km^2. Except for the Jurassic, rocks from all periods are represented (Fig. 2). The summary article that follows recounts the geologic history of this area with a stratigraphic emphasis, and relies particularly on the pioneering works of Drs. Frank Kottlowski, Lloyd Pray, and James Lee Wilson.

PRECAMBRIAN

A wide variety of igneous and metamorphic rocks are exposed in regional uplifts, and elsewhere underlie the Phanerozoic succession in southern New Mexico and West Texas (Kottlowski and Foster, 1960; Foster and Stipp, 1961; Kottlowski, 1963; Meyer, 1966; Muehlberger et al., 1966; Condie and Budding, 1979; Walker, 1992; Adams et al., 1993; Adams and Keller, 1994a, 1996). Although these rocks are best understood when placed into a tectonic context, inconsistencies in age dates make direct translations between studies difficult. Four major Precambrian tectonic events are thought to have occurred in the New Mexico-Texas region (Adams and Keller, 1996): (1) formation of the outer tectonic belt, 1.7-1.6 Ga; (2) formation of the southern granite-rhyolite province, 1.4-1.34 Ga; (3) pre-Grenville extension, 1.33-1.26 Ga; and (4) Grenville orogeny, 1.3-1.0 Ga.

Precambrian rocks exposed in the Pedernal Hills of Torrance County are part of the outer tectonic belt orogenic terrane, which extended along the Early Proterozoic southeastern margin of North America from Michigan to Arizona (Robertson et al., 1993; Adams and Keller, 1996). The Pedernal rocks consist of schists intruded by a pluton (Condie and Budding, 1979), with (from oldest to youngest): (1) medium-grained granite gneiss; (2) thinly bedded alternations of quartzite-muscovite, quartz schist-specularite, quartz schist; (3) heterogeneous metamorphic rocks including chlorite-muscovite-quartz phyllite, epidote amphibolite, quartzite, fine-grained quartz monzonitic gneiss, cataclasite, and epidiorite; and (4) intrusive granite (Woodward, 1969). This Precambrian crystalline, meta-volcanic, and meta-sedimentary complex was uplifted as part of the southern Ancestral Rocky Mountains during the Early Pennsylvanian, providing the provenance for immature Pennsylvanian arkosic sandstones deposited in the Orogrande basin.

The Pajarito Mountains and small outcrops in the northern Sacramento Mountains contain Precambrian syenite, granite, and gabbro 1.1-1.33 Ga, considered to be pre-Grenville intrusions (Bowsher, 1991; Adams and Keller, 1996). Slightly metamorphosed Precambrian fine-grained clastics and igneous intrusives are locally exposed in the southern Sacramento Mountains (Pray, 1961), dated between 1.4 Ga (Muehlberger et al., 1966) and 1.0 Ga (Condie and Budding, 1979); these metasedimentary rocks represent rare outcrops of the De Baca terrane, which underlies much of east-central New Mexico (Condie and Budding, 1979). Well data from elsewhere in the Sacramento Mountains indicate an andesitic, ultramafic, and metamorphic basement complex, whereas eastward the Pecos area contains quartzite, meta-arkose, meta-rhyolite, gneiss, and granite (Bowsher, 1991).

Extensive Precambrian outcrops occur in the San Andres and Organ Mountains, largely composed of granite intrusives into schist, quartzite, and amphibolite (Condie and Budding, 1979), and also including phyllite, talc, and dolomite (Kottlowski and

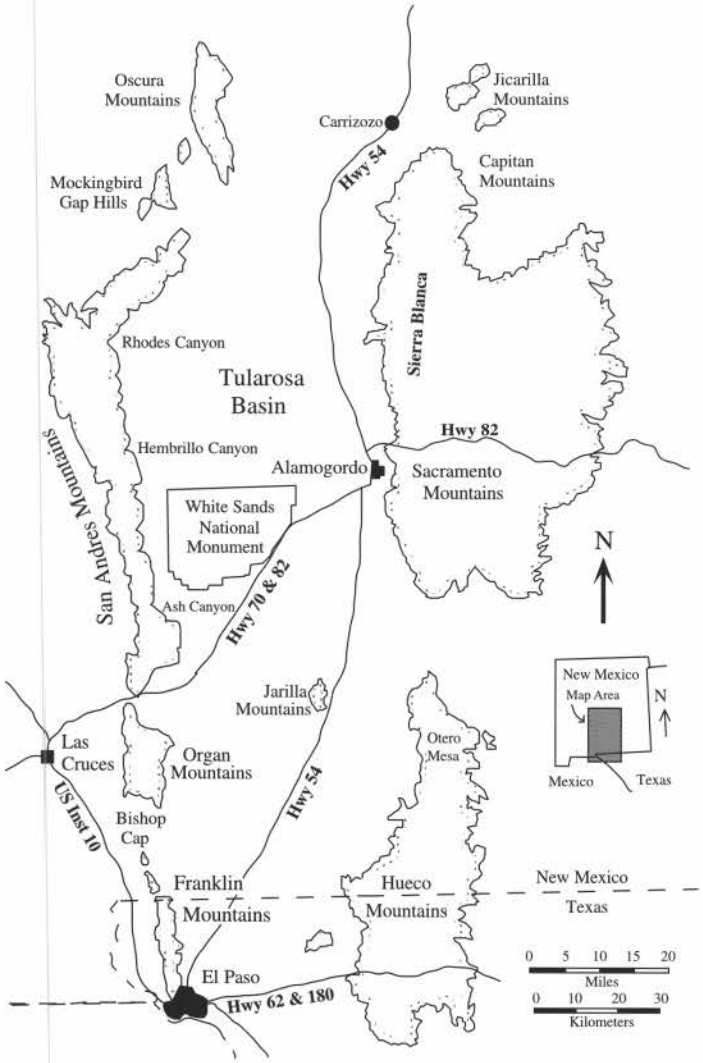


FIGURE 1. Tularosa Basin region of south-central New Mexico. Modified from Algeo et al. (1991).

LeMone, 1994). Although minor occurrences of the metasedimentary De Baca terrain may exist (Kottlowski and LeMone, 1994), most of the rocks are of the granitic-metamorphic terrane, dating from 1.0-1.7 Ga (Condie and Budding, 1979; Roths, 1991). The majority of the San Andres/Organ Mountains Precambrian rocks are interpreted as deeply eroded roots of a 1.3-1.4 Ga mountain or rift system, followed by the formation of extensional diorite-diabase dike swarms (Condie and Budding, 1979; Seager, 1981).

The Franklin Mountains contain a wide variety of crystalline and metasedimentary Precambrian lithologies. In ascending order, the Castner Marble (1.24-1.28 Ga) is interpreted as having been deposited in a low-energy carbonate ramp environment, with stromatolite mounds, flat-pebble conglomerates, ripple marks, and rhythmites (LeMone, 1996a). The overlying Mundy Breccia is a volcanic mudflow breccia, and is overlain by the Lanoria Quartzite, a metamorphosed inner-shelf sandy unit (Harbour, 1972; LeMone, 1996a). The overlying Thunderbird Group contains conglomerates, volcanics, and sandstones (Thomann, 1981). Finally, the Red Bluff Granite complex (1.13-1.15 Ga) represents intrusion of the stratigraphic succession by porphyritic alkali granite, biotite granite, and biotite-hornblende granite (Ray, 1982; LeMone, 1996a).

CAMBRIAN THROUGH MISSISSIPPIAN

Summary

From Cambrian through Devonian time, tectonics in south-central New Mexico were generally stable, forming part of the cratonic Tobosa shelf (Fig. 3). The Transcontinental Arch extended into northern New Mexico throughout this period, toeing out southwards to create a large, low-angle, south-dipping ramp. Episodic uplift of the arch and cratonic-scale transgressions and regressions created regional unconformity-bounded stratal wedges that generally thin to the north as they onlap the arch (Pray, 1961; Sloss, 1963; Kottlowski and LeMone, 1994; Bachtel and Dorobek, 1994, 1998; Cather and Harrison, this volume) (Fig. 2). Local epeirogenic warping created subtle topographic features that also exerted some control over deposition. In the Franklin Mountains area, a west-dipping component to basin physiography has been documented (LeMone, 1969, 1996b). The Mississippian system records more active tectonism than younger strata, with numerous karsted surfaces, disconformities, and angular unconformities (Pray, 1961; Kottlowski and LeMone, 1994; Bachtel and Dorobek, 1994, 1998), as well as structurally-controlled bioherm growth patterns (Bowsher, 1991; Jeffery, 1997). This more active tectonic regime was prelude to the onset of Ancestral Rocky Mountain activity.

Depositionally, a general transgression, punctuated with numerous unconformities, occurred from Cambrian through Devonian time, with relatively high sea levels persisting through the Middle Mississippian (Pray, 1961; Dott and Prothero, 1994) (Fig. 2). Paleogeographic reconstructions place south-central New Mexico within 10 degrees of the equator for the entire Paleozoic. Lower Paleozoic stratigraphy is dominated by restricted to open-marine sandstones and dolomites from stable shelf environments, with carbonates and more open-marine conditions increasing upwards, culminating in deeper-water siltstones, carbonates, and anoxic black shales in the Devonian (Pray, 1961; Kottlowski et al., 1956).

Outer ramp-to-basin nodular limestones and shales of Early and Middle Mississippian age contain abundant and occasionally large bioherms, including Waulsortian mounds (Loudon and Bowsher, 1941, 1949; Pray, 1961; Kirkby and Hunt, 1996; Jeffery, 1997; Bachtel and Dorobek, 1998; Dorobek and Bachtel, 2001). Deposition continued to occur on the south-dipping ramp, with strata forming sigmoidal wedges that thin to the north due to post Mississippian erosion and stratigraphic onlap of the Transcontinental Arch, and thin to the south due to basin sediment starvation (Wilson, 1975; Bachtel and Dorobek, 1998). Late Mississippian strata are restricted to the southern study area and may represent deposition of the oldest Orogrande basin sediments (LeMone, 1996b).

Cambro-Ordovician

Unconformably overlying the Precambrian complex is the Bliss Formation (Richardson, 1904), a Late Cambrian(?) to Early Ordovician ledge-forming, reddish-brown, glauconitic, hematitic

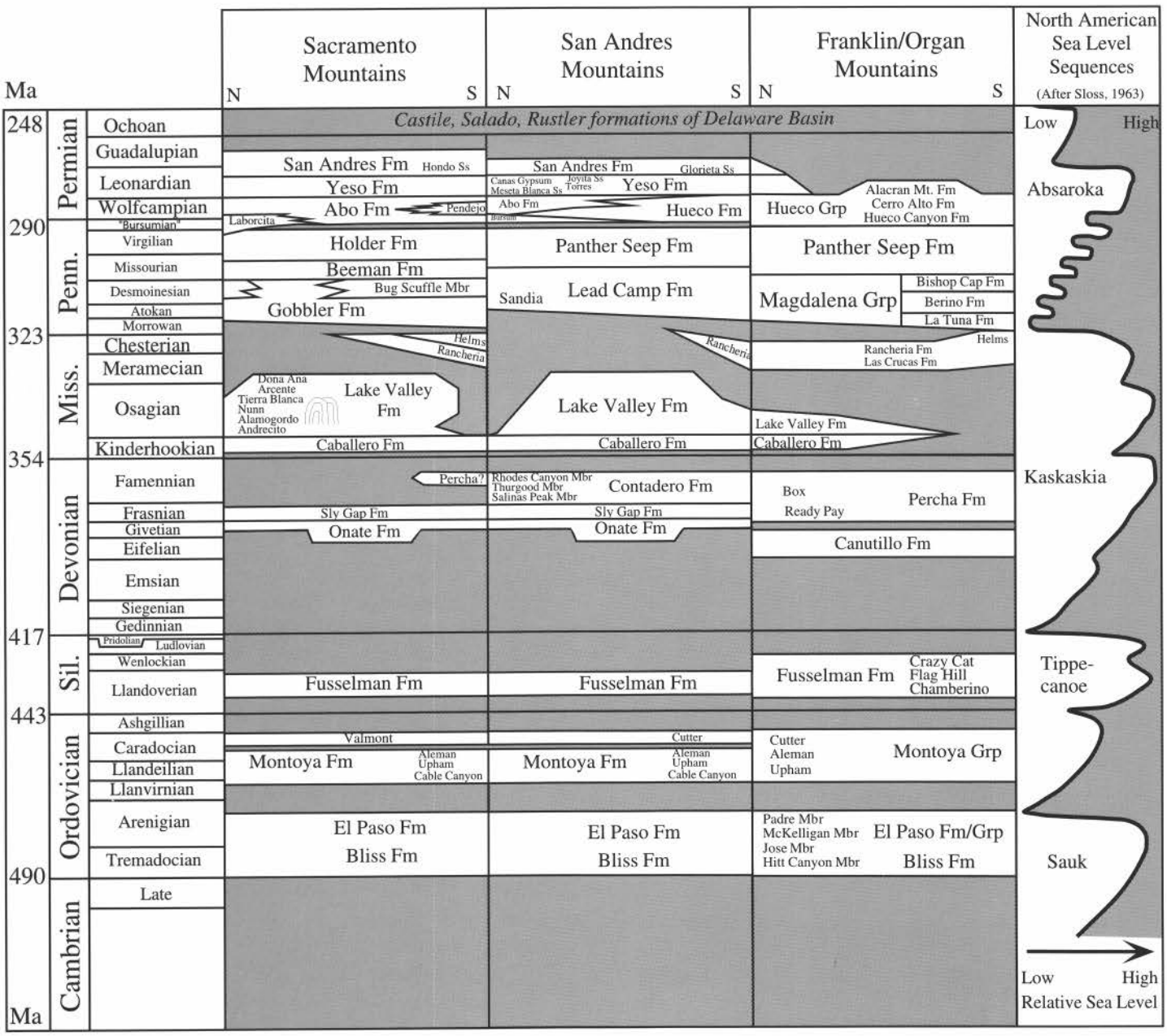


FIGURE 2. Paleozoic chronostratigraphic units in south-central New Mexico. Absolute time scale from the Geological Society of America 1999 Geologic Time Scale. Unit boundaries should be considered estimates in view of complicating factors such as diachronous contacts, conversion of series standards, and ongoing biostratigraphic studies. Global sea-level curve adapted from Dott and Prothero (1994). See text for discussions of formation versus group designations for many of the lower Paleozoic units as well as for a complete listing of Flower's (1964) El Paso units. The Pennsylvanian portion of the sea-level curve is diagramatic and not intended to represent true amplitude or frequency of cyclicity. J. Barrick and D. LeMone provided useful discussion that improved the diagram.

transgressive sandstone unit with subordinate siltstone, limestone, and dolomite. The Bliss was deposited in shoreline to shallow-marine environments (Kottlowski, 1963; LeMone, 1996b), and contains hematite ooids (Kelley, 1951). Fossils (including conodonts, lingulid brachiopods, articulate brachiopods, trilobites, gastropods, Girvanella, and Skolithos-type burrows) are generally sparse, especially in the lower portion, which creates uncertainty as to the Cambrian age designation in some areas. Late Cambrian fossils are present in the lower Bliss in the San Diego, Mud Springs, and Caballo Mountains (Kottlowski et al., 1956; Mack et al., 1998). In the Franklin, Sacramento, and San Andres Mountains the formation is probably entirely of Ordovician age (Pray, 1961; Foster, 1978; LeMone, 1996b). The unit generally thickens to the south with the addition of older beds to its base: 2-32 m in the San Andres Mountains, 33 m in the southern Sacramento Mountains, and 73 m at the Franklin Mountains type section (Richardson,

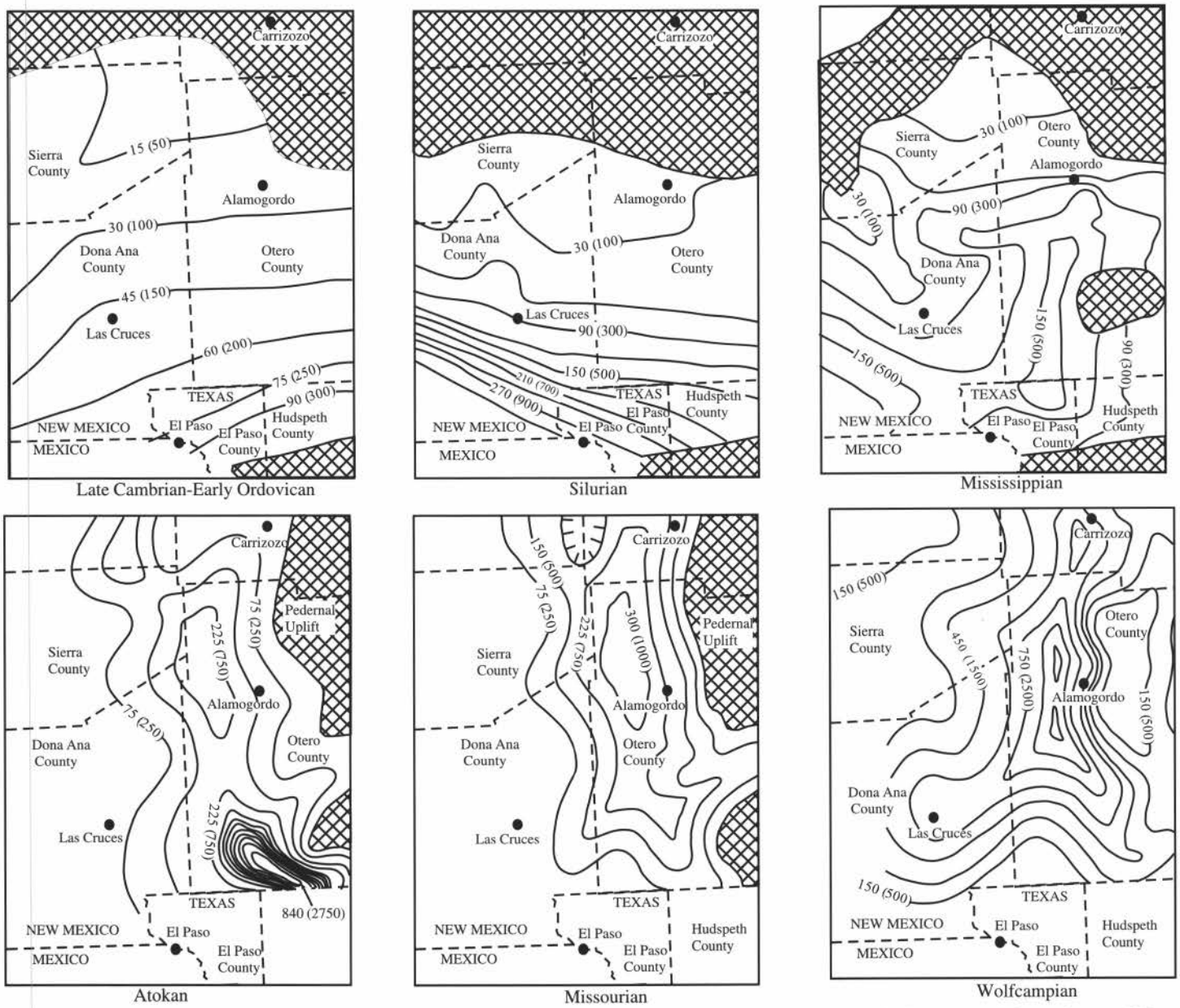


FIGURE 3. Selected isopachs illustrating the general evolution of the Tobosa shelf (top row) and Orogrande basin (bottom row) in south-central New Mexico. Contour interval varies between diagrams. Units are in meters (feet), with feet the originally published units and rounded values for meter conversions added. Eroded/depositional edges are hachured. Modified from Kottlowski (1963) and Meyer (1966). See Cather and Harrison (this volume) for a more detailed discussion of isopach trends.

1904; Kottlowski et al., 1956; Pray, 1961). The upper contact with the El Paso Formation is gradational and inconsistently correlated within a thick section of calcareous sandstones and carbonates (Kottlowski et al., 1956; Mack et al., 1998). Although likely continuous in many areas (Clemons, 1991; Kottlowski and LeMone, 1994; Mack et al., 1998), Kottlowski et al. (1956) and Pray (1953, 1961) provide discussion that the Bliss-El Paso contact may be at least locally disconformable.

The El Paso Formation or Group (Richardson, 1904) of Early Ordovician age is a skeletal limestone and dolomite unit with basal siliciclastics. It was deposited in episodically high-energy, shallow-marine environments and is generally transgressive. It thins updip to the north and east due to both depositional onlap and pre-Montoya erosion, ranging in thickness from 420 m (Franklin Mountains) to 132 m (southern Sacramento and San Andres Mountains) to 12 m (northern San Andres Mountains) (Kottlowski et al., 1956; Pray, 1961; Clemons, 1991; Mack et al., 1998). In outcrop it appears gray with thin- to medium-beds, often forming a recessive slope below the more resistant Montoya Formation. Major faunal elements include sponges, algal oncolites and laminites, echinoderms, brachiopods, trilobites, nautiloids, gastropods, and corals (Flower, 1964).

Stratigraphic nomenclature for the El Paso varies (see Flower, 1964; Clemons, 1991; LeMone, 1996). Kelley and Silver (1952)

raised the status of the El Paso Formation to Group and defined the Sierrite Limestone and Bat Cave formations. Flower (1964) divided the El Paso Group into 10 lithologic and faunal units that became recognized as formations. In ascending order these are: Sierrite, Big Hatchet, Cooks, Victorio Hills, Jose, Mud Springs Mountain, Snake Hills, McKelligon, Scenic Drive, and Florida Mountains (see detailed descriptions in Flower, 1964 and LeMone, 1982). Clemons (1991) demoted the El Paso Group to Formation and divided it into four members (in ascending order): Hitt Canyon (includes Flower's Sierrite through Victorio Hills Formations), Jose, McKelligon (includes Flower's Mud Springs Mountain through McKelligon Formations), and Padre (includes Flower's Scenic Drive and Florida Mountains Formations). Following Clemons (1991), the Hitt Canyon Member consists of a transgressive-regressive package containing basal silty limestone and dolomite, middle subtidal carbonates with numerous storm event beds, and upper restricted stromatolitic limestone; the Jose Member is a transgressive, arenaceous, bioturbated, oolitic, fossiliferous intertidal to subtidal limestone and dolomite; the McKelligon Member is a transgressive-regressive bioturbated, fossiliferous subtidal carbonate; and the Padre Member records late regression and transgression, with peritidal and subtidal arenaceous carbonates. Three major sequences containing numerous finer-scale high frequency cycles were documented in the southern Franklin Mountains, interpreted as resulting from complex interactions of glacio-eustatic sea-level oscillations and autocyclic progradation (Goldhammer et al., 1993).

The Middle to Late Ordovician Montoya Formation or Group (Richardson, 1908, 1909; Kelley and Silver, 1952; Howe, 1959) unconformably overlies the El Paso and is dominated by cherty, open-marine skeletal carbonates. The Montoya usually forms a dark-colored resistant unit, locally with abundant crinoid, coral and sponge skeletal grains (Pray, 1953). Stratigraphic status and nomenclature for the Montoya has a contorted history (reviewed in Pray, 1953, 1961; Kottlowski et al., 1956; LeMone, 1969). Mack et al. (1998) used formation rank, with members of (in ascending order): the Cable Canyon Sandstone, consisting of thin (up to ~2 m in the Sacramento Mountains and 6.5 m in the San Andres Mountains) sandstone to coarse quartz-pebble transgressive lag with some arenaceous dolomite; the Upham, composed of transgressive, bioturbated, resistant, massive dolomite to brachiopod, crinoid, trilobite, gastropod, algal, nautiloid, coral wacke-packstone, with uppermost regressive crinoidal grainstone (Kelley and Silver, 1952; Kottlowski, 1963; Herbert, 1979; LeMone, 1996b); the Aleman, composed of fossiliferous cherty limestone and dolomite (Pray, 1953; Howe, 1959; LeMone, 1996b); and the Cutter, containing thin-bedded, light gray, fine-grained dolomite with open-marine fauna capped by an unconformity surface (Pray, 1953; LeMone, 1996b). LeMone (1996b, personal commun., 2002) maintained the Montoya as having group status, and recognized the Upham, Aleman, and Cutter Formations in the Franklin Mountains. In the Sacramento Mountains, Pray (1953, 1961) defined a type section for the Valmont Dolomite that is equivalent to the Cutter. The Montoya thins to the north, with thickness up to 168 m in the Franklin Mountains, 130 m in the Hueco and southern San Andres Mountains, 88-145 m in the Sacramento Mountains, and is absent in the northern San Andres Mountains where it has been erosively removed by the Montoya-Fusselman unconformity (Kottlowski et al., 1956; Pray, 1961; Mack et al., 1998; Cather and Harrison, this volume). In the Hueco Mountains, the Montoya Formation is largely limestone and exhibits a great deal of lateral variability (Kottlowski, 1963).

Silurian

The Early Silurian Fusselman Formation (Richardson, 1908, 1909; Pray and Bowsher, 1952) unconformably overlies the Montoya Formation, and forms a regionally extensive, ledge-forming, dark gray to brown, cherty dolomite unit with only rare limestones. Although heavily dolomitized, limited fossil content (including crinoids, corals, stromatolites, and brachiopods) and lack of significant detrital material indicates deposition dominantly in an open-marine shelf environment (Pray, 1961). Above the basal unconformity, a thin, sandy, locally phosphatic transgressive lag unit exists, occasionally within scoured channels (Pray, 1961; LeMone, 1982). Three unconformity-bounded sequences have been documented within the Fusselman of the Franklin Mountains, with the uppermost unconformity extensively karsted (LeMone, 1996b). The formation thins to the north due to Pre-Devonian erosion and the depositional thinning of lower beds, ranging from 259 m in the southern Franklins, to 94 m in the southern San Andres, to 30 m in the Sacramento Mountains, to absent near Rhodes Canyon in the northern San Andres (Kottlowski et al., 1956; Pray, 1961; Kottlowski, 1963; Mack, et al., 1998; Cather and Harrison, this volume).

Devonian

The Middle to Late Devonian formations present in southern New Mexico unconformably overlie the Fusselman and underlie Mississippian units. They are typically composed of thin, recessive, locally fossiliferous gray to brown shales, siltstones, sandstones, carbonates, and barren anoxic black shales. In the San Andres and Sacramento Mountains, Devonian strata representing largely shelf environments can be grossly divided into two major unconformity-bounded packages: the Givetian clastic-dominated Oñate Formation and the upper Frasnian-Famennian mixed clastic/carbonate Sly Gap and Contadero Formations (Day, 1988). Shelf deposits from the Sacramento and San Andres Mountains grade southward into an E-W trending trough located at the approximate latitude of Las Cruces containing anoxic, barren black shales of the Percha Formation (Kottlowski, 1963; Rosado, 1970). Further south in the Franklin Mountains the Percha black shales grade into and are underlain by the Canutillo Formation, a cherty carbonate.

Interpretation of the environment of deposition responsible for the anoxia and subsequent black shale deposition ranges from shallow lagoon with algal mat covering (Seager, 1981; LeMone, 1982, 1996b; Mack et al., 1998), to "deep water" with anoxia resulting from a density-stratified seaway (Kottlowski et al., 1956; Sorauf, 1984; Day, 1988, 1998). The "deep water" model better fits the general basin physiography of the area and is the preferred interpretation. This "black shale problem" is not restricted to southern

New Mexico, but is a common interpretive conundrum throughout Devonian epeiric sea deposits in North America (e.g. Grabau, 1915; Brown and Kenig, 2001; Sageman and Arthur, 2001).

Devonian formations are well-established within individual mountain ranges, although members continue to undergo revision and correlation between ranges is not always clear (see Kottlowski et al., 1956; Seager, 1981; Sorauf, 1984; Kottlowski and LeMone, 1994). The oldest Devonian strata present in south-central New Mexico may be the Canutillo Formation (Nelson, 1940) in the Organ/Franklin Mountains. It unconformably overlies the Fusselman Formation and underlies and is a lateral facies equivalent to the lower Percha Formation, which in the southern area includes shaley facies correlative northward to the Oñate and Sly Gap Formations (Seager, 1981). The Canutillo is composed of a lower dolomitic siltstone and an upper cherty carbonate. The formation thins northward from 26 m in the Franklins to 6 m at Bishop Cap, to 1 m in the southern San Andres Mountains.

The Oñate Formation (Stevenson, 1945) of the Sacramento and San Andres Mountains is of late Givetian age and unconformably overlies the Fusselman, while to the south in the Organ/Franklin Mountains its shaley facies is incorporated within the Percha Formation. In the Sacramento Mountains the Oñate consists of open-marine shelf deposits composed of gray silty dolomite, dolomitic siltstone, and minor sandstone with bryozoans, brachiopods, and local chert (Pray, 1961). It thins from 18 m in the south-central Sacramentos to 6 m in the far northern and southern reaches of the range (Pray, 1961). In the Hueco Mountains, 32 m of sparsely fossiliferous shales, silty shales, and silty limestones may correlate to the Oñate, Sly Gap, and Percha (Kottlowski, 1963). The lower Oñate in the San Andres Mountains is similar to the Sacramento Mountain sections, augmented in the central range by nondolomitized wackestones containing corals, crinoids, brachiopods and bryozoans. The upper Oñate in the San Andres Mountains is regressive and clastic-rich, with siltstones, shale, and an upper cross-bedded sandstone unit documented (Kottlowski et al., 1956; Sorauf, 1984). The formation thins to the north and south from its maximum of 26 m in San Andres Canyon, becoming sandier to the north and shalier to the south.

The Sly Gap Formation (Stevenson, 1945), of Frasnian age, disconformably overlies the Oñate Formation (Pray, 1961; Day, 1988). It is present in the northern and central Sacramento Mountains, the entire San Andres Mountains, and is incorporated as part of the Percha Formation in the Organ/Franklin Mountains (Pray, 1961; Seager, 1981; Sorauf, 1984). It is interpreted to represent a transgressive-regressive succession deposited in shelf (Sacramento Mountains, northern San Andres Mountains) to basin (southern San Andres Mountains) environments (Day, 1988). In the Sacramento Mountains, the Sly Gap Formation contains interbeds of calcareous shale, thin-to-nodular fossiliferous lime mudstone, and lesser black shale, weathering to a distinctive yellowish color (Pray, 1961). Laudon and Bowsher (1941, 1949), Stevenson (1945), and Pray (1961) considered various upper beds of black shale in the southern Sacramento Mountains to belong to the Percha Formation, although other workers interpreted them as basin facies of the Sly Gap Formation (Kottlowski et al., 1956). In the San Andres Mountains, the Sly Gap Formation consists of nodular interbeds of fossiliferous (colonial and solitary corals, brachiopods, crinoids, ammonoids, gastropods, stromatoporoids), calcareous, silty shale, silty limestone, and calcareous siltstone (Kottlowski et al., 1956; Sorauf, 1984; Kottlowski and LeMone, 1994). In the southern San Andres Mountains, the Sly Gap is composed almost completely of dark gray to black shale deposited in anoxic environments (Kottlowski et al., 1956; Day, 1988).

The Contadero Formation (Stevenson, 1945) is recognized in the northern San Andres Mountains and originally incorporated all strata between the Sly Gap and the Mississippian, but was revised by Flower (in Kottlowski et al., 1956) to include what were originally upper Sly Gap units and also exclude upper dark shales with Fammenian fauna, which were placed in the Percha Formation. Sorauf (1984) revised Flower's member nomenclature to include: the Salinas Peak Member (sea-level highstand shale to sandstone, with upper coral-bearing nodular limestone); Thurgood Sandstone Member (regressive, fine-grained, well-indurated sandstone with calcareous cement and brachiopod fragments); and Rhodes Canyon Member (Fammenian-aged shales and burrowed siltstones with brachiopods, correlative to the Ready Pay Member of the Percha Formation). The Contadero Formation is not recognized in the Sacramento or Franklin Mountains, and may have formed in a narrow structural re-entrant largely limited to the San Andres Mountains area (Day, 1988).

In the Organ/Franklin Mountains, the Percha Formation is used to include all Middle-Late Devonian shales above the Canutillo Formation, including strata age-equivalent to the Oñate, Sly Gap, and Contadero Formations (Seager, 1981). To the north, when used at all, the Percha is constrained to only those dark shales of Fammenian age. The Percha is divided into two members: the Ready Pay (black, fissile, barren shale) and the Box (shale with nodular limestone concretions and limited fauna). As discussed, Pray (1961) considered the uppermost dark shales in the southern Sacramentos to be lower Percha rather than Sly Gap based partly on a dark shale "channel-filling" unit with angular contacts between the Sly Gap and Mississippian. Dark shales in the extreme southern Sacramento Mountains are variously interpreted as Percha or Sly Gap/Oñate basin equivalents (Kottlowski et al., 1956; Pray, 1961). Fammenian-aged shales in the San Andres Mountains are included in the Rhodes Canyon Member of the Contadero Formation (Sorauf, 1984).

Mississippian

The basal Mississippian Caballero Formation (Laudon and Bowsher, 1941) represents transgressive deposits atop the Devonian-Mississippian angular unconformity (Fig. 4). Its base contains a cherty lag, whereas the remainder of the formation consists of recessive, fossiliferous, nodular wackestones, lime mudstones, and gray calcareous shale. In the Sacramento Mountains the unit thins from 15 m in the north to 4.5 m in the south, pinching out before reaching the Hueco Mountains (Pray, 1961). In the San Andres Mountains it reaches 18 m in Ash Canyon and thins to the north and the south, wedging out completely near Hembrillo Canyon in the northern San Andres, and is not present in the Franklin Mountains (Kottlowski et al., 1956).

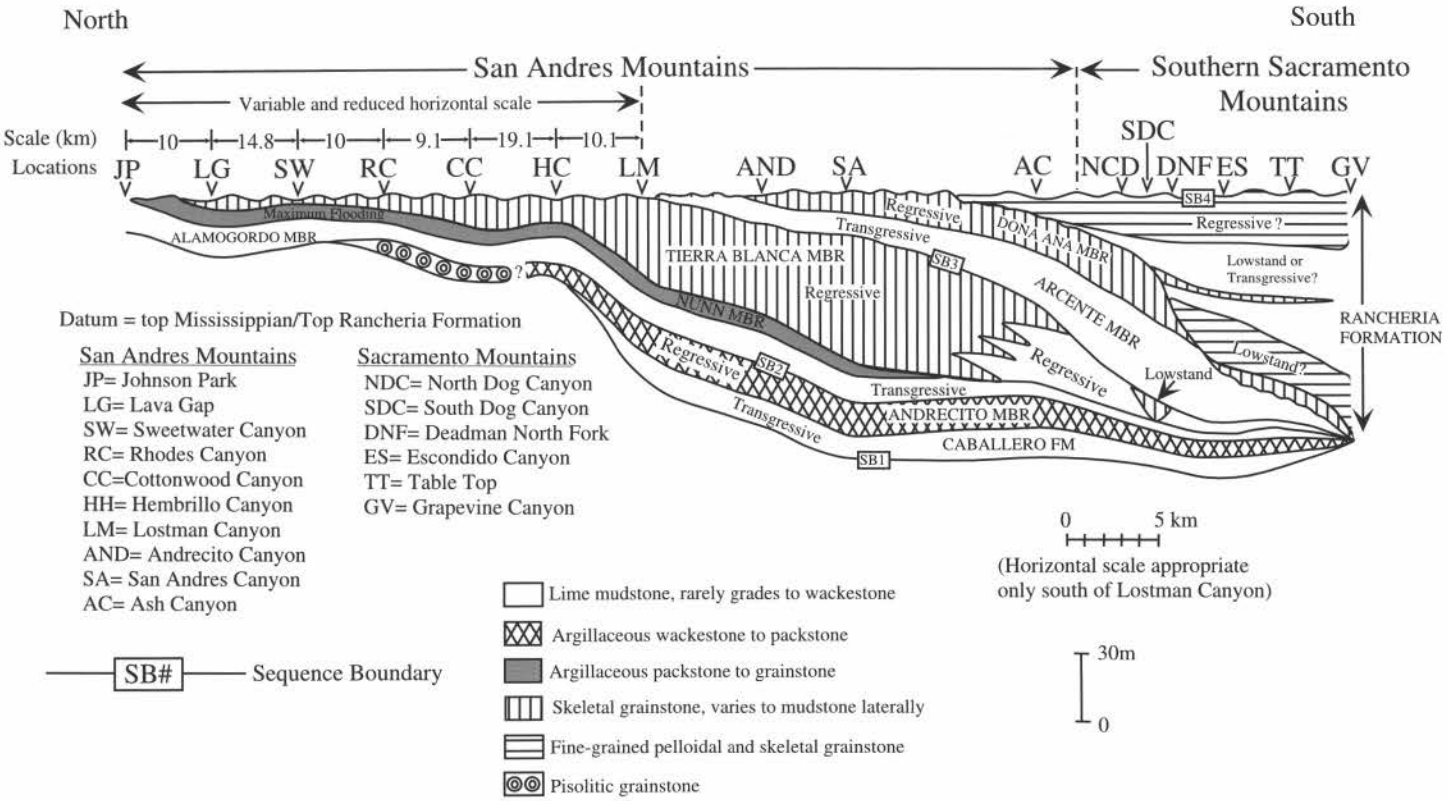


FIGURE 4. Generalized composite dip-view of the Mississippian System in the San Andres and Sacramento Mountains. Adapted from Bachtel and Dorobek (1998).

The Lake Valley Formation is composed of six members that together form a complex biohermal growth assemblage in the Sacramento and San Andres Mountains (Laudon and Bowsher, 1941, 1949; Kottlowski et al., 1956; Pray, 1961; Jeffery, 1997; Bachtel and Dorobek, 1998) (Fig. 4). The oldest member predates mound growth (Andrecito), the succeeding three members contain reef core, flank, and inter-reef facies (Alamogordo, Nunn, and Tierra Blanca), whereas the final two members postdate mound growth (Arcente and Doña Ana).

The Andrecito Member of the Lake Valley Formation is not affected by bioherm growth and is interpreted to represent progradational maximum flooding and regressive deposits of outer-to-midramp sediments following the Caballero transgression (Bachtel and Dorobek, 1994, 1998). It is composed of bryozoan, crinoid, argillaceous wackestones and lime mudstones, calcareous shale, and other minor siliciclastics. It thins to the south in the Sacramento Mountains and to the north and south of Ash Canyon in the San Andres Mountains (0-11 m in the Sacramento Mountains, 1-37 m in the San Andres Mountains) (Kottlowski et al., 1956; Pray, 1961). The unit is not present in the Franklin Mountains.

Initiation of mound growth occurred during the transgression that deposited the Alamogordo Member. This member consists of reef core (bryozoan mudstone, calcite cement), reef flank (crinoidal grainstones), and inter-reef (cherty crinoidal mudstone) facies. Thicknesses change drastically between inter-reef and reef areas, with inter-reefs typically 3.5-12 m in thickness, whereas reef cores can reach over 100 m (Pray, 1961). The unit pinches out to the south in the Sacramento and San Andres Mountains, thinning from 31 m in Ash Canyon to absent at Bishop Cap (Kottlowski et al., 1956).

The Nunn Member contains inter-reef and reef flank facies representing maximum flooding and early regressive deposits, composed largely of recessive, thinly-bedded crinoidal packstones (Bachtel and Dorobek, 1998). Reef core growth, initiated during Alamogordo deposition, continued throughout Nunn time. Thickness is dependent upon underlying and lateral reef topography, but generally reaches 36 m in the northern Sacramento Mountains and thins drastically southward to just a few meters before pinching out entirely (Pray, 1961). A clastic carbonate tongue or channel 60 m thick and 1 km wide occurs in the southern Sacramento Mountains (Bachtel and Dorobek, 1994). Maximum thickness in the San Andres Mountains reaches 20 m in San Andres Canyon, with thinning to the north and south (Kottlowski et al., 1956).

The Tierra Blanca Member contains late regressive and lowstand units, deposited as relative sea level continued to drop and middle-ramp sediments prograded across Alamogordo/Nunn outer-ramp deposits (Bachtel and Dorobek, 1994, 1998). This member, the youngest of the major mound-forming intervals, is composed of bryozoan mudstone reef cores and inter-reef/reef flank ledge-forming, cherty, crinoidal packstones and grainstones, often in lobate geometries. It reaches a maximum thickness of 60 m in reef core areas in the Sacramentos, thinning to the south until absent, whereas in the San Andres Mountains thickness varies from 30 m to absent (Kottlowski et al., 1956; Pray, 1961). It grades basinward into a dark gray, cherty, lime

mudstone to wackestone termed the Table Top Member (Lane and Ormiston, 1982).

The Arcente Member is a transgressive, dark, calcareous shale to argillaceous lime mudstone to wackestone unit that marks the end of major mound growth activity. The Arcente is largely recessive in outcrop, thickest (up to 60 m) in topographically low inter-reef areas, and thin to nonexistent over topographically high reef cores. It was erosively removed in the northern and southern Sacramento Mountains and is absent north of Andrecito Canyon in the San Andres Mountains (Kottlowski et al., 1956; Pray, 1961).

The Doña Ana Member is a regressive unit composed of cherty crinoidal grainstone through lime mudstone. Like the underlying Arcente, it is thickest in the inherited, topographically low inter-reef areas (up to 46 m thick) and thin to absent over reef cores. Well-developed clinoforms and grainstone channels attest to its regressive and progradational nature (Bachtel and Dorobek, 1994). It depositionally thinned and was erosively removed in the northern and southern Sacramentos and is only present in the southern San Andres Mountains.

The late Osagian-Mermacian Las Cruces Formation (Laudon and Bowsher, 1949) is a white weathering, fetid, dark aphanitic limestone locally broken out as a formation in the Franklin Mountains, where it reaches 18 m at its type locality. It thins to 2 m in the southern San Andres Mountains, and is not recognized as a formation in the Sacramento Mountains.

The Rancheria Formation (Fig. 4) is only present in the southern study area, where it unconformably overlies the Doña Ana Member or Las Cruces Formation. It comprises the lowstand, transgression, maximum flooding, and early regression following the Doña Ana regression (Bachtel and Dorobek, 1994, 1998), and is composed of thin- to medium-bedded lime mudstone and argillaceous, silty, cherty, peloidal grainstone with local beds of crinoid, brachiopod, shark-teeth packstone (Pray, 1961). The formation contains numerous intraformational angular unconformities and is up to 90 m thick in the southern Sacramento Mountains, thinning to absent northward in the central Sacramentos (Pray, 1961). The formation is 78 m thick in the Franklin Mountains, and up to 66 m in the Hueco Mountains. In the southern San Andres Mountains, the Rancheria reaches 59 m before pinching out northward in Ash Canyon.

The Helms Formation was deposited in the southern area during the late regression following Rancheria deposition. In the southern Sacramento Mountains it consists of up to 18 m of thinly-bedded argillaceous limestone, shale, and local oolites, thickening to 76 m southward in the Hueco Mountains. At Bishop Cap and in the Franklin Mountains it comprises about 30 m of calcareous shale, calcareous sandstone, and cherty, crinoidal, oolitic limestone (Kottlowski 1963).

PENNSYLVANIAN

Summary

The Late Mississippian and Early Pennsylvanian represent a time of major tectono-sedimentary re-adjustment throughout North America, recording the continent-continent collision of Laurasia with Gondwanaland and the ensuing Appalachian and Ouachita-Marathon orogenies, as well as the large-scale Kaskaskia-Absaroka cratonic sea-level regression (Sloss, 1963). Basin physiography in south-central New Mexico was reorganized, with structure and sedimentation affected throughout much of the remaining Paleozoic (Kottlowski, 1960; Pray, 1961; Meyer, 1966). The orogenies propagated stresses into cratonic settings to form the Ancestral Rocky Mountains as a series of positive and negative elements perpendicular to the regional stress field, probably the result of reactivated N-S Precambrian faults (Kluth and Coney, 1981). Major positive elements in the New Mexico region of Ancestral Rockies origin include: the Pedernal and Roosevelt uplifts, Florida and Central Basin platforms, Bravo dome, and Defiance-Zuni, Sierra Grande, and Matador arches; negative elements include the Orogrande, Pedregosa, Delaware, San Mateo, Lucero, Estancia, Rowe Mora, Tucumcari, and Paradox basins (Kottlowski, 1960). Paleogeographic reconstructions place southern New Mexico between 5 and 10 degrees north latitude during Pennsylvanian time.

The Pedernal uplift (Ver Wiebe, 1930; Thompson, 1942) formed the eastern margin of the Orogrande basin (Pray, 1959), an asymmetric intracratonic trough with a narrow, high-relief eastern shelf (Sacramento Shelf) and a broad, low-relief western ramp (Robledo Shelf) (Meyer, 1966; Algeo, et al., 1991). The Orogrande basin may have been periodically linked northward with the San Mateo and Lucero basins (Beck, 1993; Beck and Chapin, 1994), and perhaps at times with the Paradox basin (Fetzner, 1960). Southward, the Orogrande basin was probably linked at least sporadically with the Pedregosa basin and later with the Delaware and Marfa basins (Kottlowski, 1960, 1963).

The Pennsylvanian of south-central New Mexico, in common with much of the world, contains cyclic packages of strata (cyclothems of Weller, 1930; Wanless and Shepard, 1936) recording high frequency, large amplitude glacio-eustatic sea level fluctuations (Cline, 1959; Pray, 1961; Wilson, 1967, 1989; Goldstein, 1986; Schutter, 1989; Algeo et al., 1991; Connolly and Stanton, 1992; Soreghan, 1992; Raatz et al., 1994, Raatz and Simo, 1998); however, periods of active tectonism at times modified or obliterated the eustatic signal (Wilson, 1967; Kluth and Coney, 1981; Algeo et al., 1991; Raatz et al., 1994; Raatz and Simo, 1998). In addition, more subtle aspects of depositional control such as sediment supply and inherited topography (e.g., Gianniny, 1995), as well as climate (Goldstein, 1986; Kutzbach and Ziegler, 1989; Soreghan, 1992; Raatz and Simo, 1995, 1998) have recently been emphasized as important factors. Despite the complex interaction of depositional controls, the general dominance of eustasy created thin, unconformity-bounded cycles that may be correlative over interbasinal distances (Schutter, 1989; Algeo et al., 1991; Connolly and Stanton, 1992; Soreghan, 1992; Raatz et al., 1994; Raatz and Simo, 1998).

Control data to isopach and determine detailed Orogrande basin physiography are largely lacking; however, published attempts interpret the depocenter to have changed through time (Meyer, 1966) (Fig. 3). Starting in an extreme southern position directly west of the present day Hueco Mountains during the Morrowan and Atokan, the depocenter is interpreted to have moved significantly

northward to lie northwest of Alamogordo during Desmoinesian and Missourian time (Meyer, 1966). During rapid subsidence in the Virgilian, the basin depocenter may have shifted slightly southward to lie directly west of Alamogordo (Meyer, 1966). The Pennsylvanian as a whole reaches ~1500 m in the subsurface west of Alamogordo and thins in all directions (Meyer, 1966).

Morrowan - Desmoinesian

Uplift of the Pedernal in the early Morrowan resulted in erosion of Paleozoic strata and exposure of Precambrian crystalline rocks (Kottlowski and Foster, 1960; Meyer, 1966; Woodward, 1969; Van Wagoner, 1977; Algeo et al. 1991). Uplift of the Ancestral Rocky Mountains in the study area is marked by a regionally extensive unconformity at the Mississippian-Pennsylvanian boundary, with karsted and channeled relief and a lag of chert conglomerate (Kottlowski et al., 1956; Pray, 1961). Up to 75 m of coarse clastic material was shed onto areas flanking the Pedernal during the late Morrowan, with a facies change to limestone away from the clastic source area (Meyer, 1966). These rocks compose the lower Gobbler Formation in the Sacramento Mountains (Pray, 1961) and are generally recessive in outcrop. No deposits of Morrowan age are positively documented in the San Andres Mountains (Kottlowski and LeMone, 1994). To the south in the Organ/Franklin and Hueco Mountains, Morrowan carbonate-dominated lithologies represent open-shelf through peritidal environments of the La Tuna Formation, Magdalena Group (Harbor, 1972; Seager, 1981; Connolly and Stanton, 1986; Wilson, 1989). At Bishop Cap about 80 m of cliff-forming, cherty, skeletal limestones were deposited (Seager, 1981; Connolly and Stanton, 1986), while to the south, in both the Franklin and Hueco Mountains, 75-100 m of lower oolitic grainstone and upper Chaetetes–bearing skeletal limestones form shoaling-upward cycles (Wilson, 1989).

Coarse clastic deposition along the flanks of the Pedernal continued during Atokan time, with mixed terrigenous clastic and carbonate sediments deposited on the outer shelf and in the basin. Atokan sediments of the middle Gobbler Formation locally form over 75 m of limestone on the Sacramento Shelf and up to 850 m of terrigenous clastics in the basin center (Fig. 3) (Meyer, 1966). In the San Andres Mountains, the lower Atokan varies in thickness as it fills pre-existing topography in the karsted Mississippian, but in general is 30-100 m thick, composed of nearshore conglomeratic sandstones in the north, and more marine-influenced calcareous sandstones in the south (Kottlowski and LeMone, 1994). The upper Atokan in the San Andres is a mixture of limestones and clastics, with a general pattern suggesting a clastic source area near the modern Sierra Blanca (Kottlowski and LeMone, 1994). In the Organ Mountains, the Atokan upper La Tuna and lower Berino Formations form cyclical units of shale-skeletal limestone (Seager, 1981), while the Franklins contain 50 m of cyclical skeletal wackestones to packstones with minor grainstones (Wilson, 1989). The Atokan in the Hueco Mountains consists of ~100 m of cherty, skeletal wackestone with minor grainstone (Wilson, 1989).

Desmoinesian strata of the Gobbler (Sacramento Mountains), Lead Camp (San Andres Mountains), and upper Berino-Bishop Cap (Franklin and Hueco Mountains) Formations attain a thickness of 225 m on the Sacramento Shelf, 200 m on the Robledo Shelf, 320 m in the Franklin Mountains, 150 m in the Hueco Mountains, and over 300 m in the Orogrande basin (Meyer, 1966; Wilson, 1989). Coarse Desmoinesian clastics abut the Pedernal, while cyclic clastic-dominated deltaic deposits are spread across the Sacramento shelf. These deltaic deposits are correlative to fossiliferous, cherty, medium-bedded limestones in the San Andres and Organ Mountains, and to rugose coral, Komia lime mudstone through packstone in the Franklin Mountains (Kottlowski et al., 1956; Seager, 1981; Wilson, 1989). Early Desmoinesian strata in the Hueco Mountains contain thick (~30 m) lower shales overlain by fusulinid wackestones (Wilson, 1989).

By late Desmoinesian time, clastic influx across the Sacramento shelf had diminished. The clastic source area may have been partially drowned (Meyer, 1966), and the shelf experienced efficient sedimentary bypass through the NW-SE trending Alamo trough, resulting in clastic deposition in the trough and in the basin, but thick, massively-bedded carbonate deposits on the shelf outside the trough boundaries (Pray, 1961; Meyer, 1966; Van Wagoner, 1977; Algeo et al., 1991). These massive, cyclic, cliff-forming carbonates are termed the Bug Scuffle Member of the Gobbler Formation (Pray, 1961). The origin of the Alamo trough is debatable, with delta (Thompson, 1942; Meyer, 1966) or distributary channel axis (Pray, 1961) suggested, as well as an incisement feature due to karst and/or en echelon normal faults (Van Wagoner, 1977; Algeo et al., 1991). Late Desmoinesian deposits in the San Andres, Franklin, and Hueco Mountains are composed of lower cliff-forming, massively-bedded, skeletal limestones and upper argillaceous limestone and shale (Kottlowski et al., 1956; Wilson, 1989).

Missourian

Missourian deposits form a mixed clastic/carbonate, recessive, cyclic unit with occasional ledges. During latest Desmoinesian through early Missourian time, a pulse of tectonic activity further enhanced the relief of the Pedernal uplift and caused renewed deposition of clastics (Pray, 1961; Meyer, 1966; Raatz and Simo, 1995). Compositionally and texturally immature terrigenous clastics shed from the Pedernal formed fluvio-deltaic cycles over the northeastern (northern and central Sacramento Mountains) and northwestern (northern San Andres Mountains) shelf and basin (Kottlowski, 1959; Pray, 1961), followed in the middle and upper Missourian by a return to cyclic, carbonate-dominated deposition. The change from clastic to carbonate lithology in the middle Missourian probably represents tectonic quiescence in conjunction with rising sea levels that partially or wholly drowned the clastic source area. A cratonic-scale drying event (Phillips et al., 1985) and continental drift into higher, drier latitudes may also have reduced production of clastic sediments.

In the Sacramento Mountains, thickness of the Missourian Beeman Formation ranges from 115-160 m, thinning to the south with the loss of the lower sandstone unit (Pray, 1961; Meyer, 1966; Raatz, 1996). The carbonate-dominated upper Beeman Formation has both basin and shelf facies exposed. In updip shelf areas, the upper Beeman is composed of massive, sea-level high-

stand open-marine wackestones and packstones bounded by laminar caliche unconformity surfaces (Pray, 1961; Raatz, 1996). To the west in the Dry Canyon basin area, cyclic packages contain thin, open-marine wackestones and deep-water shales, and thick, laminated, argillaceous, pelloidal-foraminiferal grainstones, and paleosols with nodular caliche. This "basin" environment is dominated volumetrically by thick peritidal lowstand deposits correlative updip to unconformity surfaces on the shelf (Raatz and Simo, 1998). The true sea-level highstand basin deposits are represented by thin (<1 m), maximum flooding, diastemic, offshore dark shales, correlative updip to much thicker, open-marine shelf wackestones (Schutter, 1989; S.R. Schutter, personal commun., 1993; Raatz et al., 1994; Raatz and Simo, 1995, 1998). Poorly-developed to absent transgressive and regressive deposits suggest rapid sea-level changes and environmental instability, resulting in significant deposition only at stable highstand and lowstand sea-level turnaround points (Raatz and Simo, 1998). The lateral transition from open-marine carbonates to diastemic offshore dark shales over a short distance indicates a well-developed shelf break during Missourian time.

Robledo shelf deposits in the San Andres Mountains contain Missourian strata of the upper Lead Camp and lower Panther Seep Formations that thin from 115 m in the north to 50 m in the south. Thinning is due to both the depositional loss of the lower sandstone unit and also the erosion or nondeposition of upper Missourian strata (Kottlowski and LeMone, 1994). The Missourian is cyclic, with mixed clastic-carbonate shoaling-upward sequences bounded by exposure surfaces (Soreghan, 1992, 1994). Carbonate wackestones and prodelta siliciclastic mudstones rather than offshore dark shales represent the deepest water environments (Soreghan, 1994), supporting an interpretation that the western margin of the Orogrande basin was shallower and more ramp-like compared to the eastern margin's well-defined shelf break.

In the Organ/Franklin Mountains, the lower Panther Seep Formation is probably Missourian, but the exact boundary between the Missourian and the Virgilian is unknown (Seager, 1981). The lower Panther Seep Formation is composed of recessive, fine- to medium-grained terrigenous clastics and dark shales with lesser carbonates and gypsum (although the gypsum is possibly of Virgilian age; Seager, 1981).

Virgilian

Virgilian deposits are characterized by 5-70 m thick shoaling-upward sequences with large phylloid-algal bioherms, studied intensely as outcrop analogs for producing hydrocarbon fields (e.g. Wilson, 1967; Rankey et al., 1999). A tectonic episode, termed the pre-Abo event, occurred on the Sacramento Shelf during Virgilian and Wolfcampian time, resulting in numerous N-S trending faults and folds that had direct influence on sedimentation (Otte, 1959; Pray, 1961; Wilson, 1967). This sedimentary-tectonic interaction is exemplified by the shelf break in the northern Sacramento Mountains being defined by the emergence of the La Luz anticline and the basinward progradation of phylloid mounds off of its rising axis (Wilson, 1967). Phylloid-algal bioherms of early Virgilian age rim much of the shelf margin in the Orogrande basin, oriented with their long axes parallel to strike (Wilson, 1975; Toomey et al., 1977). The pre-Abo tectonic event also caused a dramatic increase in subsidence rates in the Orogrande basin, recorded in thicker Virgilian basin deposits (760 m) than during the earlier Missourian or Desmoinesian (~300 m each) (Pray, 1961; Meyer, 1966). The clastic composition of these deposits indicates that the rapidly subsiding basin received large amounts of material shed from the Pedernals. However, the carbonate-dominated lower and middle Holder Formation in the Sacramento Mountains suggests the clastics in part bypassed the eastern shelf.

In the San Andres Mountains, a definite increase in clastics is noted beginning at the Missourian-Virgilian boundary, with clastic-dominated mixed clastic-carbonate deposits documented throughout the Virgilian Panther Seep Formation (Kottlowski et al., 1956; Soreghan 1992, 1994). The Panther Seep in the southwestern Orogrande basin contains significant gypsum, with up to 31 m present in the southern San Andres Mountains (Kottlowski et al., 1956; Kottlowski, 1963; Meyer, 1966). Similar to the thick lowstand deposits documented in the Missourian basin environments, these Virgilian evaporites may represent lowstand deposition in the basin during stable sea-level turnarounds.

In the Organ/Franklin Mountains, a thick (~450 m) section of middle and upper Panther Seep Formation contains cyclic, mixed clastic-carbonate-gypsum units with complex vertical facies changes ranging from relatively deep water through exposure (Wilson et al., 1969; Seager, 1981). In the Hueco Mountains, the 150-m thick Virgilian is composed of large phylloid-algal bioherms interbedded with fine-grained clastics, shallow-water limestones, and limestone conglomerate (Toomey, 1991).

Despite active tectonism, high frequency cycles of glacio-eustatic origin remain evident throughout many Virgilian deposits (Pray, 1961; Wilson, 1967; Goldstein, 1986; Soreghan, 1992). Overall, sea level fell during the Virgilian, causing younger cycles to become more terrestrially-dominated, with associated development of thick, well-developed redbed deposits and gypsum (Kottlowski et al., 1956; Pray, 1961; Wilson, 1967; Goldstein, 1986).

"Bursumian"

Until recently, the Bursum and Laborcita Formations were designated as lowermost Permian (lower Wolfcampian). However, the newly adopted global stratotype section for the Pennsylvanian-Permian boundary (Davydov et al., 1998) and its translation to south-central New Mexico sections has necessitated an upward movement of the boundary. The Pennsylvanian-Permian contact is now recognized at the top rather than the base of the Bursum/Laborcita Formations (Wahlman and King, 2002), and the new stage unit representing these strata is tentatively termed the Bursumian (Ross and Ross, 1994; Wahlman and King, 2002). This redesignation has not eliminated the need for further work to define chronostratigraphically these complex Bursum units and their relationships to the Abo and Hueco Formations.

The Bursum and Laborcita Formations are somewhat nebulously defined as redbed/sandstone/limestone "transitional beds" marking the latest Pennsylvanian shoreline zone (Jordan, 1971).

In the northern Sacramento Mountains, the Laborcita Formation attains a maximum thickness of 330 m and was deposited during and following a major uplift of the Pedernal near the northeastern Sacramento Shelf (the Pre-Abo event), with terrestrial units near the uplift grading into marine units west and southwestward. The Laborcita commonly filled the topographic lows resulting from structural relief (Otte, 1959; Pray, 1961; Delgado, 1977). A marine facies north of Alamogordo includes large bioherms composed of phylloid-algal core facies overlain by bryozoa, crinoid, brachiopod packstone units (Otte, 1959; Rundell, 1982; Shinn et al., 1983). The Laborcita in the Jarilla Mountains contains at least 470 m of section (thermally altered by igneous intrusives), with lower clastics grading upward into carbonates (Schmidt and Craddock, 1964; Jordan, 1971). The Bursum Formation is present in the Oscura and northernmost San Andres Mountains, containing up to 72 m of mixed clastics and carbonates. Some authors recognize the Bursum south of Mockingbird Gap in the San Andres Mountains as far as Ash Canyon (Kottlowski and LeMone, 1994), while others believe these strata are more correctly referred to as Hueco Formation (Jordan, 1971). No Bursum Formation is recognized in the southern San Andres, Organ, Franklin, or Hueco Mountains (Jordan, 1971; Kottlowski and LeMone, 1994; Seager, 1981).

PERMIAN

Summary

Although some cyclicity was retained in aspects of Permian deposition (Otte, 1959; Mack and James, 1986), global tectonic forces and climate change significantly altered pre-existing environmental conditions. The formation of the megacontinent Pangaea created hot, dry, terrestrially-dominated environments with little associated deposition or preservation of marine carbonates on the craton (Dott and Prothero, 1994). Permian positive elements in West Texas and eastern New Mexico include the Pedernal uplift (Early Permian), Central Basin platform (Early Permian), and Diablo platform; negative elements include the Orogrande (Early Permian), Midland, Delaware, and Marfa basins.

The Wolfcampian succession in the Orogrande basin/northwestern shelf area represents rapid subsidence and basin infilling with mixed marine/terrestrial carbonates and clastics (Fig. 3). The final two Permian series found in the Tularosa Basin area are the transgressive Leonardian and Guadalupian, which represent a return to marine conditions within a tectonically stable shelf (Pray, 1961). Their deposition finally infilled and completely eliminated all remaining structural relief in the local area (Meyer, 1966). Leonardian restricted backreef shales, evaporites, and limestones of the Yeso Formation correlate in the Delaware basin to the deeper shelf and basin Victorio Peak Dolomite and Bone Spring Limestone. San Andres Formation massive shelf limestones of Guadalupian age form the resistant crest of the Sacramento Mountains and are exposed at the surface throughout Otero Mesa, and correlate in the Delaware basin to the Victorio Peak and Cherry Canyon Formations (Black, 1973), part of the classic Capitan Reef system exposed in the Guadalupe Mountains of southeast New Mexico and West Texas (e.g., Pray, 1988).

Wolfcampian

Active uplift and rapid subsidence created a thick, complex mosaic of terrestrial and marine environments during the Early Permian in south-central New Mexico. In general, deposits are terrestrial-dominated in the north and marine-dominated in the south (Kottlowski et al., 1956; Otte, 1959; Pray, 1961; Jordan, 1971), indicating the Orogrande basin was filling from the north with the marine depocenter moving southward. The Wolfcampian includes the Hueco and Abo Formations, ranging in thickness from 61-152 m in the Sacramento Mountains (Pray, 1961), 286-512 m in the San Andres Mountains (Kottlowski et al., 1956), 301+ m in Jarilla Mountains (Schmidt and Craddock, 1964), up to 1050 m in the basin center, 823 m in the Franklins, and ~400 m in the Hueco Mountains. The Abo thins southward, whereas the Hueco Limestone thickens southward.

The Abo Formation consists of thick (500 m near Carrizozo), terrestrial sandstone and shale redbed deposits that thin and grade southward into the marine Hueco Limestone (Otte, 1959; Pray, 1961; Jordan, 1971; Black, 1973; Mack and James, 1986). The transition occurs in the southern Sacramento and southern San Andres Mountains over a distance of ~25 km (Kottlowski, 1963). Within the transition zone in the Sacramento Mountains, a lower Abo Formation Pedernal-derived unit composed of basal conglomeratic proximal alluvial fans and upper fluvial mudstones and siltstones is termed the Danley Ranch Tongue (Bachman and Hayes, 1958; Speer, 1986). A medial interfingering Hueco marine limestone and shale unit is termed the Pendejo Tongue (Pray, 1961). Finer-grained upper Abo Formation flood basin, distributary stream, and clastic tidal flat mudstones and siltstones derived from sources in northern New Mexico and Colorado are termed the Lee Ranch Tongue (Bachman and Hayes, 1958; Jordan, 1971; Speer, 1986). In the San Andres Mountains, the lower Danley Ranch Tongue is absent, indicating the Pedernal-sourced material was only deposited locally. Here the Hueco Limestone directly overlies the Bursum (in the north) or older Pennsylvanian units (in the south), with the fine-grained Lee Ranch Tongue redbed clastics overlying the carbonates as far south as the northern Franklin Mountains (albeit thinly) (Kottlowski et al., 1956).

The Hueco Limestone, referred to variously as a formation or a group, grades from thin limestones and shales in the Sacramento Mountains to thick, massive limestones in the Hueco and Franklin Mountains. In the Sacramento Mountains, the Hueco Limestone (Pendejo Tongue) is a poorly-developed brackish limestone, dolomite, and shale unit that thickens to over 115 m in the southern part of the range. In the San Andres Mountains the Hueco is better developed, thickening from 126 m in the north to over 410 m in the south, containing calcareous shales and fossiliferous cherty limestones, locally with algal bioherms (Kottlowski and LeMone, 1994).

The Jarilla Mountains contain 301+ m of shallow-shelf carbonates and clastics (Schmidt and Craddock, 1964; Jordan, 1971, 1975), including algal bioherms. In the Hueco Mountains, the Wolfcampian is divided into four members; the basal Pow Wow Member is a conglomerate locally sourced from the southern Pedernal uplift, whereas the upper three members (Hueco Canyon, Cerro Alto, Alacran Mountain) are carbonate shelf to

shelf margin facies that include algal/tubiphytes bioherms. Recent work by Wahlman et al. (1992, 1999), Stoklosa et al. (1998), and Simo et al. (2000) have interpreted the western outliers of the Hueco Mountains to contain well-developed shelf margin facies with massive, westerly prograding phylloid-algal/tubiphytes bioherms interbedded with dark, argillaceous basin facies and shelf margin debris flow breccias.

In the Franklin Mountains, the Hueco Group reaches 823 m in thickness and is largely composed of open-marine to inner-shelf carbonates, with local algal/tubiphytes bioherms, siltstones, and shales (Wilson, 1967; Jordan, 1971, 1975). The Pow Wow Conglomerate is not present, whereas the basal Hueco Canyon Formation contains at least 20 shoaling-upward cycles ranging from open-marine wackestone to restricted grainstone (Jordan, 1971). The overlying Cerro Alto Formation records increased siliciclastic influx and reduced faunal diversity, with shallow-shelf carbonates separated by siltstones and shales. The upper, Alacran Mountain Formation contains a lower, cherty, algal-plate biostrome overlain by shallow-shelf wackstones and packstones (Jordan, 1971).

By the end of Wolfcamp time, the Orogrande basin was largely filled, and the Pedernal uplift was almost buried (Kottlowski, 1963; Meyer, 1966; Jordan, 1975), leaving the Orogrande basin area as part of the Permian Basin northwestern shelf (King, 1942).

Leonardian

Conformably overlying the Abo/Hueco Formations is the Yeso Formation, composed of a mixture of terrestrial through marine sandstone, siltstone, shale, evaporite, limestone, and dolomite. Based on subsurface and outcrop data, the Yeso is roughly 1000 m thick near Carrizozo (a local thickness maximum in the Carrizozo basin, largely caused by salt thickening) and thins to the southwest, until it is absent at Las Cruces (Foster, 1978; R. Broadhead, personal commun., 2002). Although poorly exposed in the Sacramento Mountains, it ranges from ~365-550 m in thickness, containing a lower portion dominated by gypsum with lesser limestone, shale, and rare sandstone; a middle mixed limestone, siltstone, and shale unit; and an upper limestone, dolomite, and sandstone unit (Pray, 1961). The formation has fewer redbeds, less gypsum, and becomes more carbonate-rich in the southern Sacramento Mountains and at Otero Mesa, indicating more open marine conditions southward (Pray, 1961; Black, 1973).

In the northern San Andres Mountains, the Yeso Formation is divided into four members (Wilpolt and Wanek, 1951): the basal Meseta Blanca (light red to gray silty and shaley sandstone); the Torres (interbedded gypsum, limestone, sandstone, and siltstone); the Cañas (gypsum with limestone and siltstone); and the Joyita (calcareous sandstone) (Kottlowski et al., 1956; Kottlowski and LeMone, 1994). In the San Andres Mountains, the Yeso thins from 480 m in the north to 270 m in the south, and becomes less gypsiferous southward. Limestone units are locally fossiliferous, including brachiopods, gastropods, cephalopods, and pelecypods (Kottlowski et al., 1956). In the far southern San Andres and Organ Mountains, the Abo-Yeso-San Andres Formational boundaries are difficult to differentiate (Seager, 1981). However, it is clear that the Yeso thins and contains no gypsum.

Leonardian/Guadalupian

The San Andres Formation conformably overlies the Yeso and represents a continuation of the Permian transgression, with lithologies largely composed of massive shelf limestones and dolomites. The thickest known occurrence is 285 m from a well near Three Rivers north of Alamogordo (Foster, 1978); the formation depositionally thins southwestward and is absent due to erosion over parts of the eastern Tularosa Basin and western Sacramento Mountains (Foster, 1978). A basal, generally well-rounded, fine- to medium-grained quartz arenite termed the Glorieta Sandstone often marks the base of the San Andres Formation. In the Sacramento Mountains, this unit occurs 18-36 m above the base and is incorporated in the Hondo Member (Pray, 1961), although this is probably Glorieta equivalent (Kelley, 1971).

In the Sacramento Mountains, the resistant, fossiliferous limestone and lesser dolomites of the San Andres Formation form the crest of the range, often as an erosional remnant less than 100 m thick (Pray, 1961). Kelley (1971) subdivided the formation in the Pecos Valley into the Rio Bonito (dolomite, limestone, and Glorieta Sandstone), Bonney Canyon (dolomite and limestone), and Fourmile Draw (dolomite, gypsum, redbeds, sandstone) members. Otero Mesa contains San Andres Formation at the surface, including fossiliferous Rio Bonito and Bonny Canyon carbonates (Black, 1973). Limited non-gypsiferous dolomite and sandstone of the Fourmile Draw Member are also present (Black, 1973).

In the San Andres Mountains, the San Andres Formation consists of 180 m (north) to 60 m (south) of fetid, fossiliferous limestone with local lenses of possible Glorieta Sandstone near the base (Kottlowski et al., 1956). The lower contact with the Yeso is difficult to differentiate and has been interpreted at various stratigraphic levels (Kottlowski et al., 1956; Bachman and Meyers, 1969; Seager, 1981; Kottlowski and LeMone, 1994), whereas the upper contact is an unconformity marking the end of preserved Paleozoic strata.

MESOZOIC

Summary

Relatively little Mesozoic rock is present in south-central New Mexico. The Orogrande basin was filled, eustatic sea levels were low, and the supercontinent Pangaea created hot, dry, terrestrial environments with erosion rather than deposition as the dominant force. Significant Triassic and Cretaceous strata occur north of the Sacramento Mountains in the Sierra Blanca, Capitan, and Jicarilla Mountains area, but are beyond the boundaries of this report. Interested readers are referred to Kottlowski (1963), Allen and Kottlowski (1981), Lucas (1991a), Cather (1991), and references therein. Summaries of Mesozoic strata west of the Tularosa Basin are presented in Mack et al. (1998).

Triassic

Triassic strata in the region are composed dominantly of terrestrial redbeds that thin southwestward due to Jurassic/Cretaceous erosion (Kottlowski, 1965). The only proposed Triassic rocks in

the study area are in the San Andres Mountains north of Rhodes Canyon, with 15-55 m of poorly exposed yellow, red, green, and gray calcareous siltstone and mudstone correlated to the Middle Triassic Moenkopi Formation (Kottlowski and LeMone, 1994). Bachman and Harbour (1970) and Lucas (1998), however, interpret these deposits as belonging to the Permian Artesia Formation.

Cretaceous

Marine and nonmarine Cretaceous rocks in south-central New Mexico are interpreted to have been deposited in two separate areas or depocenters. An early (largely Albian) Tethyan-related suite occurs in southwestern New Mexico and southeastern Arizona, with a northern limit trending approximately from Ash Canyon in the southern San Andres Mountains to the Jarilla Mountains south of the Sacramentos. A northern Late Cretaceous suite was deposited later as part of the Western Interior Seaway, with a local depocenter near the Caballo and Fra Cristobal Mountains west of Rhodes Canyon, and a southern limit trending from El Paso to Deming (Kottlowski, 1965; Mack et al., 1998).

In the San Andres Mountains, Cretaceous strata unconformably overlie the Triassic(?) in the north and the Permian in the south. The oldest unit, present in the southern range, is the Sarten Formation of Early Cretaceous (Albian) age, composed of 20-30 m of fossiliferous mixed shale, silty sandstone, and sandstone (Kottlowski and LeMone, 1994). Hummocky cross-stratification reported in the Sarten (Mack et al., 1998) indicates an offshore marine environment. The unit pinches out to the north near Hembrillo Canyon.

Unconformably overlying the Sarten Formation near Ash Canyon is the Late Cretaceous Dakota Group, composed of up to 56 m of unfossiliferous, quartzitic, cross-bedded sandstone (Kottlowski and LeMone, 1994). The lower portion is interpreted as fluvial, whereas the upper strata are thought to be tidal to shallow marine (Mack et al., 1998). On the crest of the northern Sacramento Mountains, a 60 m thick outlier of Dakota Group sandstone containing marine fossils was identified and described by Pray and Allen (1956) and more recently evaluated by Lucas (1991b). The unit unconformably overlies the Permian San Andres Formation limestone and contains up to 45 m of locally bioturbated and cross-bedded fine- to coarse-grained transgressive quartz arenite overlain by up to 15 m of olive shale and nodular limestone with marine fossils (Pray and Allen, 1956; Lucas, 1991b). Arenaceous foraminifers from the shale/nodular limestone unit place the upper outcrop as Early Cretaceous (late Albian) in age (Lucas, 1991b).

In the San Andres Mountains, the Dakota is conformably overlain by 200 m of dark, carbonaceous Mancos Shale, containing a lower unit (the Rio Salado Tongue) deposited in outer shoreface to offshore marine environments (Mack et al., 1998). The Mancos grades southward in the southern San Andres and Organ Mountains into a sandy, nearshore marine facies named the Eagle Ford Shale (Kottlowski et al., 1956; Kottlowski and LeMone, 1994), dated as late Turonian in age. Overlying the Mancos is 80 m of tan, cross-bedded, nearshore sandstone and shale interpreted as correlative to the regressive Gallup Sandstone (Seager, 1981; Kottlowski and LeMone, 1994).

CENOZOIC

Summary

During Late Cretaceous through Eocene time, south-central New Mexico together with much of the western U.S. experienced the Laramide orogeny, a major tectonic event that significantly reorganized crustal structure and sedimentary basin geometry. In the Oligocene, crustal extension and widespread volcanism resulted from the first stage of Rio Grande rifting, while the second stage began in the late Miocene and continues today. Discussion of the complicated and still evolving geophysical and tectonic interpretations of these major events is beyond the scope of this report, which instead will only place preserved rocks within a simplified depositional and structural framework. No Tertiary sedimentary strata are recognized in the Sacramento or northern and central San Andres Mountains, although Eocene igneous sills and dikes are locally common. The majority of preserved Tertiary deposits are located in the southern San Andres/Organ Mountains and areas southwestward.

Paleocene-Eocene

Late Cretaceous to early Tertiary Laramide deformation was generally "thick-skinned," with large, asymmetrical, basement-cored block uplifts and anticlines bounded by steep thrust faults and monoclines (Seager, 1981; Dott and Prothero, 1994). A large, northwest-trending Laramide block, the Rio Grande uplift, extends through Las Cruces and south of Truth or Consequences, with conjugate sedimentary basins formed to its north (Love Ranch basin) and south (Portillo basin) (Seager et al., 1986; Mack et al., 1998).

The Late Cretaceous to early Paleocene(?) McRae Formation (Kelley and Silver, 1952) and the Love Ranch Formation (Kottlowski et al., 1956) of late Paleocene to Eocene age represent infilling of the newly formed basins with fluvial sandstones (McRae Formation) and clastics shed in alluvial fans from Laramide highs (Love Ranch Formation). A major basal angular unconformity exists due to the burial of Laramide structures (Kelley and Silver, 1952). Love Ranch alluvium in the southern San Andres Mountains forms a succession up to 650 m thick of conglomerates, red siltstones and minor tuffs, with all material but the tuffs derived from eroded fragments of pre-Tertiary updip lithologies (Mancos-Eagle Ford down to the Hueco Formation) (Kottlowski et al., 1956). Southward, the Love Ranch thins greatly until in the Organ Mountains it comprises ~30 m of pediment veneer conglomerate resting atop and composed of clasts from the Lower Permian and Pennsylvanian (Kottlowski et al., 1956; Seager, 1981). Both the McRae and lower Love Ranch Formations contain Late Cretaceous volcanic clasts, indicating igneous activity during the early phase of deformation (Seager et al., 1997; Mack et al., 1998).

Following Love Ranch deposition, a series of Eocene to Miocene volcanic events occurred. These are mainly exposed in the Organ Mountains and farther west. The oldest volcanics in the series is the Orejon Andesite (Dunham, 1935), probably of middle Eocene age (Mack et al., 1998). The Orejon lavas contain andesite,

siliceous tuff, and pieces of granite, schist, gneiss, and limestone interbedded with breccias deposited as volcanic mudflows (Seager, 1981). The Sacramento and San Andres Mountains contain numerous igneous dikes and sills, with samples from the Sacramento Mountains dated by the argon 40-argon 39 method as middle Eocene age (42.7 ± 3.0 Ma to 45.5 ± 1.0 Ma) (Peters, 2000).

Oligocene

Volcanism in south-central New Mexico reached its peak during the Oligocene, with lavas of andesitic/basaltic and silicic composition (Mack et al., 1998). In the southern Organ Mountains, over 3000 m of rhyolite lavas and ash flow tuffs were unconformably deposited atop the Orejon andesites, followed by the intrusion and cooling of the massive Organ batholith-the granitic body that today forms the imposing and distinctive "organ pipe" peaks of the mountains (Seager, 1981). It was also during the Oligocene (~30 Ma) that the initial extension and low-angle normal faulting of the Rio Grande rift occurred, with the oldest sediments of the Santa Fe Group probably deposited during latest Oligocene time (Chapin and Seager, 1975; Lozinsky and Bauer, 1991; Adams and Keller, 1994b; Mack et al., 1998).

Miocene-Pliocene

Early stage rifting and block faulting continued into the early Miocene (until ~18 Ma), although volcanism was not common (Adams and Keller, 1994b; Mack et al., 1998). During this period, the Rio Grande rift basin was probably a broad downwarp, continually filling with complex assemblages of conglomerate, sand, silt, clay, and evaporites within alluvial fan, piedmont-slope, and playa deposits of the Santa Fe Group (Chapin and Seager, 1975). The region was volcanically and tectonically quiet from ~18 to ~10 Ma, with renewed extension beginning ~10 Ma and continuing today (Adams and Keller, 1994b). This second phase of activity involved E-W extension and resulted in the current N-S configuration of uplifts and basins (Chapin and Seager, 1975; Adams and Keller, 1994b). High-angle normal faults created horst blocks that divided the pre-existing broad downwarp into a series of smaller, segmented basins and uplifts (Chapin and Seager, 1975; Mack et al., 1998). The simplified structure of the Tularosa Basin consists either of two half grabens separated by, and dipping away from, the central Jarilla fault zone (Lozinsky and Bauer, 1991) or a nearly symmetrical horst (central basin) and graben-block (eastern and western basin) structure (Adams and Keller, 1994b). There is a maximum thickness of ~2750 m of Tertiary and Quaternary fill in the deeper basin (Healy et al., 1978). Late Pliocene to early Pleistocene lacustrine and fluvial facies are identified in the southern basin, deposited by the Rio Grande (Chapin and Seager, 1975; Mack et al. 1997).

Pleistocene-Recent

The Tularosa Basin is closed, with no external drainage. During the late Pleistocene to early Holocene, the giant Lake Otero filled the western half-graben north of the Jarilla Mountains (Herrick, 1904). The waters contained dissolved salts derived mostly from the Yeso Formation, and as the lake constricted, evaporitic minerals were concentrated and deposited. The modern remnant of Lake Otero is the intermittent Lake Lucero, containing large selenite crystals precipitated in its playa bottom. Wind blowing dominantly from the southwest began forming gypsum dunes derived from Lake Otero and Lake Lucero sediments 12,000 to 24,000 years ago, creating what is today the largest gypsum dune field in the world (LeMone, 1994). Although surrounded by restricted military lands, the dune fields are publicly accessible within the 596 km^2 White Sands National Monument.

In the northern Tularosa Basin lies one of the youngest lava fields in the state, the Malpais region containing the Carrizozo lava flows. These basaltic flows display ropey pahoehoe textures and stretch for 75 km, with an average thickness of 10-15 m, and have recently been dated as 5200 ± 700 years of age (Dunbar, 1999)

The youngest deposits in south-central New Mexico are the soils, alluvial fans, stream deposits, playas, and other similar recent sediments. Fault scarps that cut modern alluvium with as much as 25 m of relief demonstrate active movement along major mountain-bounding faults (Pray, 1961), attesting to the dynamic nature of the Rio Grande rift area and its continuing geologic evolution.

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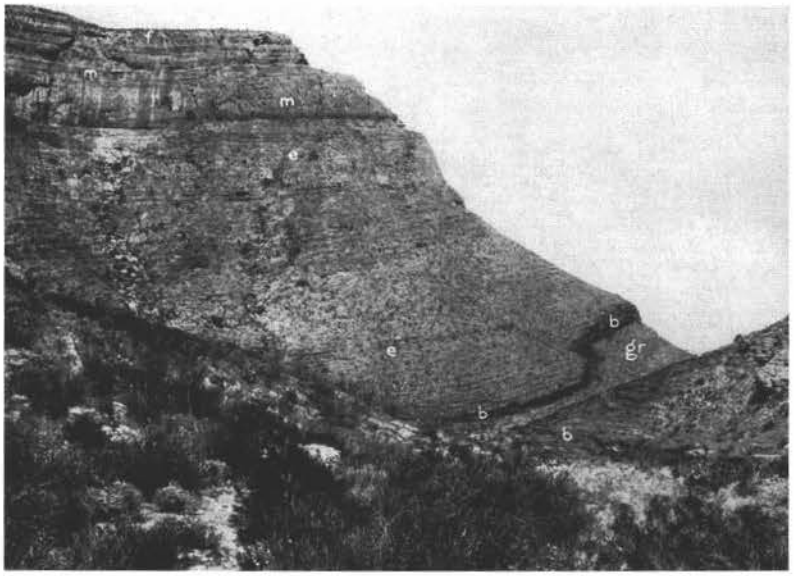
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The north wall a the mouth of Hembrillo Canyon, San Andres Mountains. Symbols: f – Fusselman Dolomite; m – Montoya Formation; e - El Paso Formation; b – Bliss Formation; gr – Proterozoic granite. Photo from Darton (1928). The dark color of the Bliss is due the high concentrations of glauconite and oolitic hematite in the unit. A minipaper by Mack (this volume) discusses the origin of these enigmatic hematite occurrences.