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THE STRATIGRAPHY OF SOUTH-CENTRAL NEW MEXICO

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Abstract—South-central New Mexico contains rocks of Precambrian, Paleozoic, Cretaceous, and Cenozoic age exposed in late Cenozoic fault-block mountains and adjacent basins. Precambrian (upper Proterozoic) metamorphic rocks and granites are unconformably overlain by a sequence of Upper Cambrian through Mississippian marine carbonates and minor sandstones and shales about 1 km thick. Late Paleozoic deformation associated with the Ancestral Rocky Mountains created the Pedernal uplift and complementary Orogrande Basin, in which were deposited approximately 2 km of Pennsylvanian and Permian mixed marine and nonmarine siliciclastic and carbonate strata. Lower Cretaceous siliciclastic and carbonate sediment approximately 500 m thick was deposited in an extensional basin (Chihuahua trough) in the southernmost part of the area, and about 1 km of Upper Cretaceous siliciclastic sediment, preserved locally in the northern part of the region, was deposited in the Western Interior foreland basin. Andesitic rocks were emplaced along the western margin of the study area, representing the easternmost spread of a Late Cretaceous continental arc in the Southwest. Latest Cretaceous (Maastrichtian) and early Tertiary (Paleocene and Eocene) compressional deformation associated with the Laramide orogeny produced a northwest-trending, basement-cored uplift and complementary basins, in which were deposited about 1 km of nonmarine siliciclastic detritus, followed by andesitic volcanism. Extension in the Rio Grande rift began in the Oligocene with minor block faulting and voluminous bimodal volcanism, including eruptions of the Organ and Doña Ana cauldrons, and was followed in the Miocene by block faulting and deposition of about 1.5 km of primarily siliciclastic detritus. Several hundred meters of sediment and minor basalt were deposited during the most recent stage of extension of the Rio Grande rift, which began in latest Miocene and Pliocene time and is responsible for the present distribution of basins and uplifts.

INTRODUCTION

The region of interest in this report covers an area of approximately 31,000 km^2 of south-central New Mexico. It ranges from Truth or Consequences in the north to the New Mexico-Mexico border and El Paso, Texas, on the south, and from the Sacramento Mountains on the east to the Good Sight and Animas Mountains on the west. This region contains numerous fault-block mountain ranges created by late Cenozoic crustal extension (Fig. 1). Exposed in these mountains are stratigraphic units ranging in age from late Precambrian to Holocene, including upper Proterozoic metamorphic and granitic rocks, approximately 3 km of Paleozoic sedimentary rocks, 2 km of Cretaceous sedimentary rocks, and 3.6 km of Cenozoic sedimentary and volcanic rocks. The purpose of this paper is to describe the stratigraphy of south-central New Mexico and to discuss the tectonic settings responsible for its development.

PRECAMBRIAN CRYSTALLINE BASEMENT

Precambrian rocks in south-central New Mexico crop out along the base of the larger fault-block ranges, generally confined between range-boundary faults and the nonconformity with lower Paleozoic strata. Outcrops range in size from barely one square kilometer in the Sacramento Mountains to several 100 km^2 in the San Andres-Organ range. Three radioisotopic dates from these outcrops indicate a late Proterozoic age for Precambrian basement in the area: 1304 ± 72 Ma (recalculated using new decay constant) from the Caballo granite (Muehlberger et al., 1966), 1655 ± 15 Ma from granite in the Black Range (Stacey and Hedlund, 1983), and 1000–1100 Ma for granite in the San Andres Mountains at Mockingbird Gap (White, 1977). McLemore (1986) has speculated that small syenite bodies in the southern Caballo Mountains may be of latest Precambrian to Cambrian or Ordovician age.

Precambrian terrane consists primarily of granitic plutons locally intruded into regionally metamorphosed rocks. In addition, small syenite bodies, diabasic dike swarms, and nearly unmetamorphosed sedimentary rocks have been described from the Caballo, Organ, and Sacramento Mountains, respectively (Pray, 1961; Seager, 1981; McLemore, 1986). The granite represents partial exposures of several distinct plutons, distinguished from one another by mineralogy, phenocryst content, presence or absence of pegmatite facies, and texture. Compositions range from granodiorite to granite. Most plutons lack a metamorphic fabric except locally at their margins. This relationship, together with sharp, concordant to locally discordant contacts with adjacent plutons or with metamorphic wall rock suggest most, if not all of the plutons are post-tectonic. Based on field

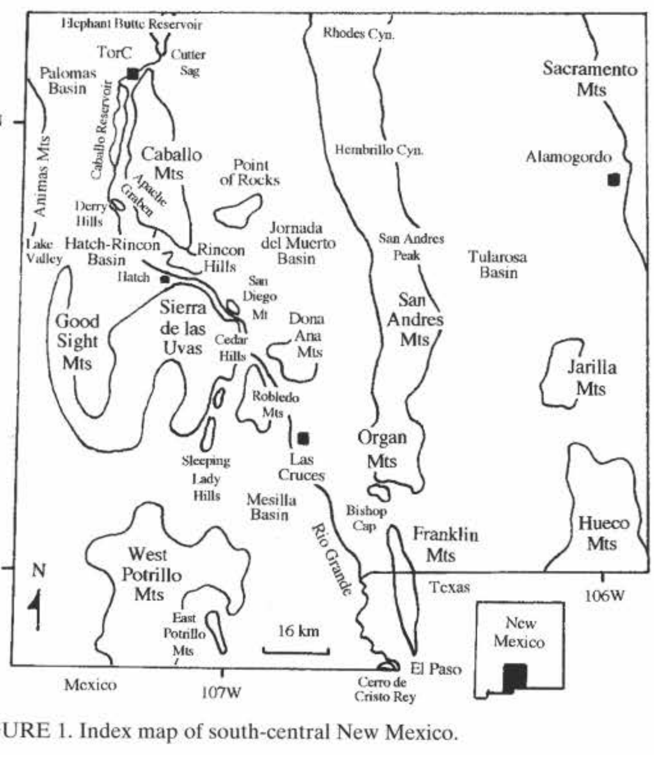

FIGURE 1. Index map of south-central New Mexico.

relationships and petrographic and compositional data, Condie and Budding (1979) considered all to be epizonal intrusions.

Metamorphic rocks form septa between plutons as well as small to large xenoliths or roof pendants within plutons. The largest tracts of metamorphic rocks are in the west-central Caballo Mountains (Bauer and Lozinsky, 1986) and in the east-central San Andres Mountains (Kottlowski et al., 1956; Condie and Budding, 1979; Seager et al., 1987). Near Hembrillo Canyon in the San Andres Mountains, the metamorphic sequence consists of interbedded metaquartzite, phyllite, and schist with lesser amphibolite, siliceous metavolcanic rocks, and talc, all metamorphosed in the greenschist to amphibolite faces (Kottlowski et al., 1956; Condie and Budding, 1979; Fitzsimmons and Kelley, 1980; Seager et al., 1987). Farther north in the range, near Salinas Peak, metadiabase and amphibolite dominate, the pelitic and quartz-rich rocks being less voluminous (Condie and Budding, 1979). Foliation strikes northwesterly at Hembrillo Canyon and northeasterly to easterly at Salinas Peak.

In the central Caballo Mountains, quartz-feldspar schist and gneiss, and red felsic gneiss are interlayered with thick and extensive sheets of amphibolite and metagabbro, metamorphosed in the amphibolite facies. Steep to moderately dipping foliation strikes easterly in the Caballo Mountains, similar to structural trends at Salinas Peak in the San Andres Mountains (Bauer and Lozinsky, 1986; Seager and Mack, in press a). Representative relationships between Precambrian granitic plutons and metamorphic rocks in the central Caballo Mountains are illustrated by outcrops between Longbottom and Burbank Canyons (Fig. 2). The area has been mapped and studied by Bauer and Lozinsky (1986) and Seager and Mack (in press a.). Three separate plutons that invaded metamorphic wall rock appear to be post-tectonic, judging from the lack of penetrative metamorphic fabrics, locally discordant contacts, and abundant xenoliths with sharp, angular, undeformed margins. However, Bauer and Lozinsky (1986) interpreted the Longbottom granodiorite to be pre- or syntectonic. Metamorphic rocks seemingly represent septa of wall rock separating the three plutons. Notable is the exceptional development of pegmatitic textures and dikes along the contact of the Caballo granite with metamorphic rocks.

The metamorphic assemblages in the Caballo Mountains and San Andres Mountains are similar enough to suggest continuity between the two areas. Protoliths may be interpreted to have been a thick succession of sandstone, argillaceous sandstone, arkosic sandstone, and shale interbedded with and/or intruded by large flows, dikes or sills of basalt and smaller sheets of ash-flow tuff or rhyolite. As noted by Condie and Budding (1979), the sequence of sedimentary rocks together with the bimodal igneous suite is suggestive of a continental rift tectonic setting, although other interpretations are possible. If the sequence does represent a rift, it was eventually deformed by primarily north-south compression, and the sedimentary and volcanic fill was metamorphosed to greenschist and amphibolite facies. Subsequent emplacement of numerous granitic plutons between 1000 and 1600 Ma ago obliterated much of the metamorphic succession.

PALEOZOIC SEDIMENTARY ROCKS

Premier outcrops of Paleozoic strata form the capping cliffs and ledges of the mountains of south-central New Mexico and record a tripartite history (Fig. 3; Kottlowski et al., 1956; Kottlowski, 1963; Kottlowski and LeMone, 1994): (1) Cambrian through Mississippian rocks, deposited in tropical seas teeming with invertebrates, thin northward toward the Transcontinental Arch; (2) northward erosional truncation was pronounced in late Mississippian and early Pennsylvanian time, followed by rise of the Pedernal uplift and deposition of mixed siliciclastic-carbonate strata in the Orogrande Basin; (3) Permian silicicilastic, carbonate and evaporite rocks buried all but the uppermost peaks of the Pedernal uplift.

Cambrian-Ordovician-Silurian strata

Early Paleozoic beds are the Cambrian–Ordovician Bliss Formation, Ordovician El Paso and Montoya Formations, and Silurian Fusselman Dolomite with type sections in the Franklin Mountains (Richardson, 1904). The Bliss Formation spans the Cambrian-Ordovician boundary. Late Cambrian fossils exist in lower beds at San Diego Mountain, Caballo Mountains, and Mud Springs Mountains, whereas earliest Ordovician faunas are widespread in upper Bliss strata. In particular, the type section in the Franklin Mountains contains Ordovician conodonts to within 15 m of its base (LeMone, 1996). In almost all sections, lower beds lack diagnostic fossils and probably are of late Cambrian age.

The Bliss Formation crops out as a conspicuous dark-brown cliff in most of the ranges. Hematite is present throughout the formation, but relatively thick beds of oolitic hematite appear to be limited to a 30 km-wide east-trending belt from Rhodes Canyon in the San Andres Mountains to north of Silver City (Kelley, 1951). This oolitic-hematite facies appears to parallel the northern Bliss shoreline (Kottlowski, 1963). In the San Andres Mountains, northward thinning of the Bliss appears to be depositional rather than erosional (Kottlowski et al., 1956).

The Bliss is dominantly medium- to coarse-grained, dark reddish-brown, glauconitic and hematitic sandstone, with sparse faunas, and minor interbeds of siltstone, limestone, and dolomite. Basal beds are commonly pebbly sandstones. Nearly everywhere, the Bliss lies on a relatively even surface that was deeply eroded in Precambrian crystalline rocks. Locally, however, there are thick channel-fill sandstones, and in some areas the thickness of the formation varies greatly in short distances. In the central Franklin Mountains, near the "Thunderbird" and Trans Mountain Highway, remnant hills of Precambrian Thunderbird Metarhyolite were overlapped by the lower El Paso Formation (Kottlowski et al., 1973).

The Bliss was deposited in shoreline and shallow-marine environments during the Late Cambrian and Early Ordovician eustatic transgression (Lewis, 1962; Kottlowski, 1963; Thompson and Potter, 1981; Chafetz et al., 1986; Stageman, 1988; Seager and Mack, 1991). The upper contact with the El Paso Formation is gradational and has been picked at several different horizons within a thick sequence of interbedded carbonates and sandstones. The Bliss has a maximum thickness of 73 m in the Franklin Mountains and thins northward to a depositional zero edge in the northern Fra Cristobal range (Nelson, 1986).

The El Paso Formation was deposited mainly as limestone, but in many areas has been dolomitized. It is of Early Ordovician age, gradationally above the Bliss Formation and unconformably beneath the Montoya Formation. The El Paso thins northward, chiefly owing to early Middle Ordovician erosion, from 414 m in the southern Franklin Mountains to the eroded edge in the southern Oscura and San Mateo Mountains. As noted by Clemons and Osburn (1986), the El Paso has been the subject of much nomenclatural confusion. Beginning with Richardson (1904), who named the formation from its type section in the Franklin Mountains, there have been numerous studies of the fossiliferous El Paso, many of which raised the El Paso to group status and applied various formation and/or member names (Cloud and Barnes, 1948; Kelley and Silver, 1952; Flower, 1953, 1959, 1965, 1969; Lucia, 1968; LeMone, 1969, 1982; Harbour, 1972; Hayes, 1975). These stratigraphic subdivisions are not mappable at the scale of 1:24,000, leading Clemons (1991) to propose that the El Paso should be restrict-

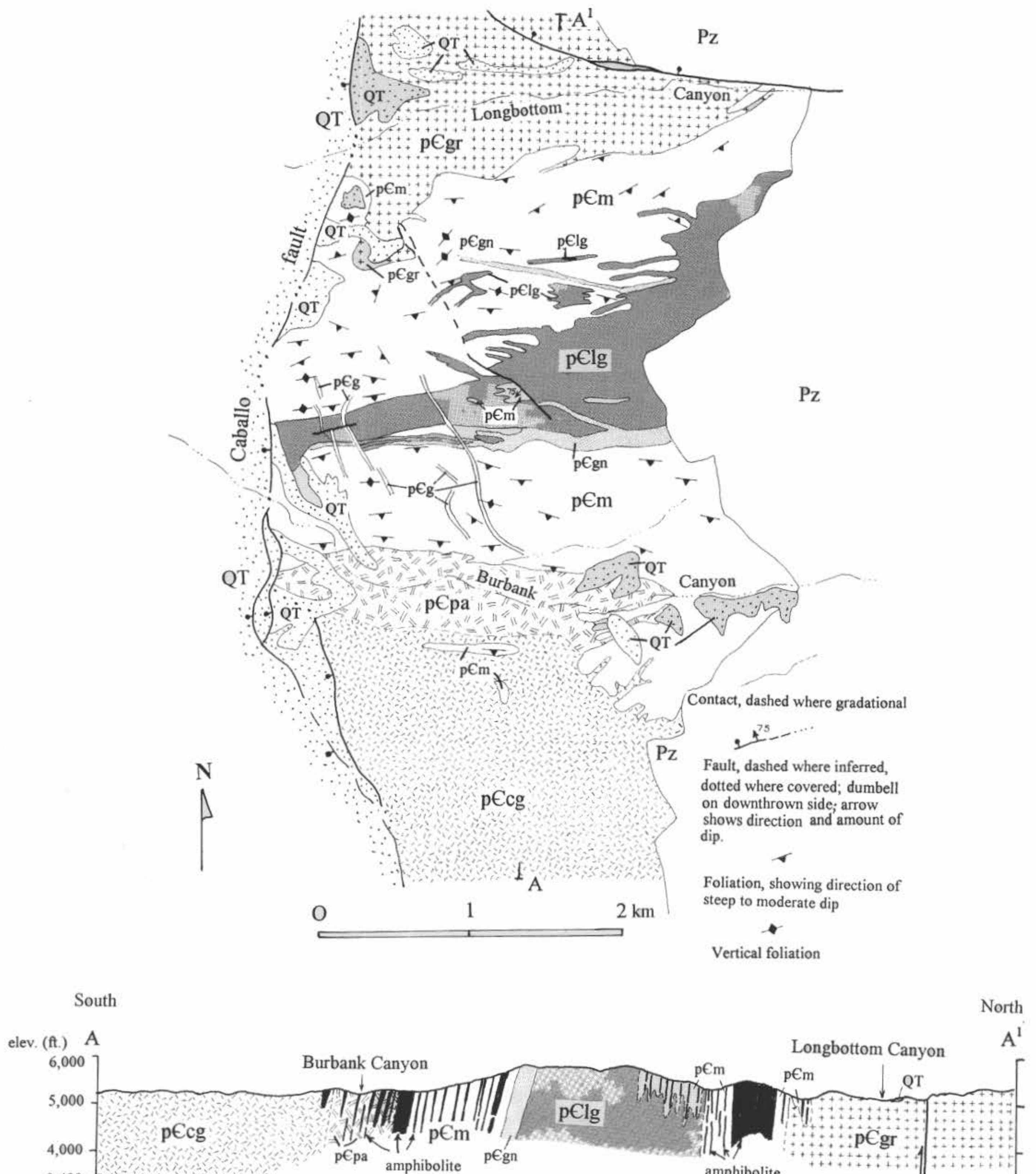

FIGURE 2. Geologic map of Precambrian rocks along the western flank of the Caballo Mountains. pЄg = Caballo granite; pЄpa = pegmatite facies of the Caballo granite interlayered with amphibolite; pЄm = amphibolite, gneiss, and schist; pЄgn = red, felsic gneiss; pЄlg = Longbottom granodiorite; pЄgr = unnamed granite; Qt = Quaternary and Tertiary fanglomerate, gravel, and sand.

ed to formation status. Using mainly the names of Hayes (1975), Clemons divided the El Paso into four members, which in ascending order are the Hitt Canyon, Jose, McKelligon, and Padre.

The Hitt Canyon Member consists of (where not dolomitized) gray silty limestone, mainly fossiliferous, with prominent irregularly reticulated laminae, lower arenaceous and silty beds, and upper stromatolitic beds (Clemons and Osburn, 1986; Clemons, 1991; Seager and Mack, 1991). About 130 m thick in the Franklin Mountains, it thins to 91 m in the southern Caballo Mountains. The distinctive Jose Member is dark-gray, sandy, oolitic limestone and bioturbated, fossiliferous limestone. The Jose Member is 6 to 14 m thick in the Caballo Mountains and 25 m thick at the type section. The McKelligon Member is light-gray, medium- to thick-bedded, bioturbated, fossiliferous limestone and intraclast limestone with sporadic sponge-Calathium stromatolitic mounds (Clemons, 1991). The McKelligon Member is 160 m thick in the Franklin Mountains, thins to 115 m in the southern San Andres Mountains and 52 m in the southern Caballo Mountains, and is only 15 m thick in the Rhodes Canyon area of the San Andres Mountains. The Padre Member is marked by thin- to medium-bedded, gray limestones, which are sandy and silty near the base. It is 105 m thick in the Franklin Mountains and is erosionally thinned to 35 m in the southern San Andres Mountains and 4 m in the Robledo Mountains, and is absent at San Diego Mountain, in the central San Andres Mountains (Hembrillo Canyon), and Caballo Mountains.

The El Paso is considered to be mostly shallow subtidal and intertidal, deposited on a cratonic platform during the latter stage of the Cambrian–Early Ordovician eustatic transgression of North America (Clemons, 1991). Goldhammer et al. (1993) discussed high-frequency cyclicity at the type section in the Franklin Mountains in the context of sequence stratigraphy.

The Montoya Formation is of late Middle to Late Ordovician age (Flower, 1965; Hayes, 1975). It overlies the El Paso Formation unconformably, and in some areas, such as the southern Franklin Mountains, karst topography was developed prior to deposition of the basal Montoya. The upper contact is also an erosional surface, which shows only slight relief beneath the Fussleman Dolomite, but abrupt, irregular relief where overlain by Devonian strata. Northward truncation of the Montoya was by pre-Devonian erosion in north-central Sierra County and by pre-Pennsylvanian erosion in southern Socorro County. The Montoya consists of four members: basal Cable Canyon Sandstone, Upham Dolomite, Aleman Dolomite, and Cutter Dolomite (Kelley and Silver, 1952). The Cutter Member is referred to as the Valmont Dolomite in the

San Andres Ls	Upper Leonardian		Absent in CM
Yeso Fm	Lower Leonardian		
Abo Fm	Middle & Upper Wolfcampian		Intertongues with Hueco
Hueco Fm	Middle & Upper Wolfcampian		Below and intertongues with Abo
Bursum/Laborcita (SAC)	Lower Wolfcampian		only Robledos, N. SAM and SAC
Pennsylvanian:	Caballo (CM)	San Andres (SAM)	Sacramentos (SAC)
Virgilian	Bar B Fm	Panther Seep Fm	Holder Fm
Missourian	Bar B Fm	Lead Camp Ls	Bursum Fm
Des Moinesian	Nakaya Fm	Lead Camp Ls	Gobbler Fm
Atokan	Red House Fm	Lead Camp Ls	Gobbler Fm
Morrowian	Red House Fm	Abs.	Gobbler Fm
Helms Fm	Upper Mississippian		only S. SAC and S. Organ Mts
Rancherio Fm	Upper Middle Mississippian		only S. SAM and S. SAC
Lake Valley Fm (ls)	Lower Middle Mississippian		
Doña Ana Ls			only S. SAM and SAC
Arcente Ls			only S. SAM and SAC
Tierra Blanca Ls			
Nunn Memb			
Alamogordo Ls			
Andrecito Memb			
Caballero Fm	Lower Mississippian		only Robledos, C. SAM and SAC
Contradero Fm	Lower Upper Upper Devonian		only SAM and SAC
Percha Sh	Upper Devonian		CM, Robledos, Organ Mts
Sly Gap Fm	Lower Upper Devonian		SAM and SAC
Oñate Fm	Middle Devonian		SAM and SAC
Fusselman Dol	Upper Silurian		eroded N. SAM
Montoya Fm (dol):	Upper Ordovician		
Cutter Memb/Valmont Dol (SAC)			
Aleman Memb			
Upham Memb			
Cable Canyon Ss Memb			
El Paso Fm (ls):	Lower Ordovician		
Padre Memb			eroded N. CM and N. SAM
McKelligan Memb			
Jose Memb			
Hitt Canyon Memb			
Bliss Fm (ss)	Upper Cambrian-Lower Ordovician		

Sacramento Mountains (Pray, 1953).

The Cable Canyon Sandstone is a dark-brown-weathering, massive to thick-bedded, ledge-forming, burrowed, dolomitic, locally crossbedded, bimodal fine- to coarse-grained sandstone. Thickness varies greatly, with the thickest section (9 m) along a southeast-trending lobe in southwestern Sierra County and northeastern Luna County, including the southern Caballo Mountains (Kottlowski, 1963). Bruno and Chafetz (1988) interpreted the Cable Canyon Sandstone to be a shallow-marine sandwave complex deposited during regional marine transgression.

The upper members of the Montoya Formation are mostly dolomite. The Upham Member is medium- to dark-gray, massive to thick-bedded, medium-grained dolomite, although limestone locally is present in the southern Franklin Mountains. Basal beds are sandy and chert nodules are scattered in the upper part. Thicknesses are relatively constant, 21–34 m, except where thinned by post-Montoya erosion. The Aleman Member gradationally overlies the Upham, and consists of medium- to dark-gray, fine-grained dolomite with numerous light-gray, elongate chert nodules that impart a distinctive ribbon-like appearance. Thicknesses vary erratically from 21 to 61 m, probably owing to varying chertification of the overlying and underlying members. The contact between the Aleman and overlying Cutter Member is marked by the disappearance of numerous chert nodules. The Cutter Member is light-gray to light-tan, fine-grained, locally thinly laminated dolomite with scattered brown chert nodules and stringers. Thicknesses are relatively uniform, 46–64 m, but the member thins abruptly northward as a result of erosion during late Silurian–middle Devonian time. Deposition of the Montoya Formation occurred during a late Ordovician transgression that followed an erosional period in early Middle Ordovician time, coinciding with a worldwide sea-level lowstand (Vail et al., 1977).

The Fusselman Dolomite is Early and Middle Silurian in age at its type section in the southern Franklin Mountains (Richardson, 1909), but northward only Middle Silurian beds were deposited (Kottlowski et al., 1956). It crops out as a prominent dark-brown-weathering dolomite cliff disconformably overlying the light-colored beds of the Cutter Member of the Montoya Formation. The Fusselman is a fine- to coarse-grained, dark-gray, massive to medium-bedded dolomite with many dark gray chert nodules in upper beds. It thins northward, unconformably underlying Devonian strata and locally basal Pennsylvanian rocks. Where the upper contact is exposed, it is commonly an undulating, channeled, silicified surface formed during late Silurian and Early and Middle Devonian time. Northward thinning is illustrated by thicknesses of 259 m in the southern Franklin Mountains, 94 m at Bishop Cap (Seager, 1981), and 19 m in the central San Andres Mountains (Hembrillo Canyon). Deposition of the Fusselman Dolomite occurred during an Early to Middle Silurian transgression that followed Early Silurian eustatic lowstand (Vail et al., 1977).

Devonian strata

Upper Middle and Upper Devonian strata contrast with the underlying Ordovician and Silurian rocks. Whereas the latter were deposited on shallow-marine platforms, the Devonian dark shales were laid down in lagoons or shallow seas with muddy bottoms and restricted circulations. In south-central New Mexico, Upper Devonian strata are commonly mapped as the Percha Formation, a sequence of dark-gray, fissile, micaceous, silty shale with thin lenses of siltstone and fine-grained sandstone. At the type section, west of Hillsboro, the lower Ready Pay Member is 40 m of dark-gray, fissile shale and is overlain by 14 m of the Box Member, which consists of gray, fossiliferous shale and nodular limestone. In the San Andres Mountains, 25–50 m of upper Middle and Upper Devonian units, consisting of fossiliferous calcareous shale and sandstone, grade southward into dark-gray shales (Sorauf, 1984). These northern units are the Middle Devonian Oñate Formation and Upper Devonian Sly Gap and Contadero Formations. In the Franklin Mountains and at Bishop Cap, the Percha is underlain by the Canutillo Formation, which thins northward from 35 to 14 m and consists of basal silty beds and upper cherty limestone and dolomite (Seager, 1981).

A northward increase in the amount of sand and silt suggests the Transcontinental Arch was a probable source for siliciclastic sediment of the Percha Formation, although siltstone in the basal part of the Canutillo Formation in the southern part of the study area may have been derived from the south. Contact with the overlying Mississippian beds appears to be one of slight erosion, but not much angular unconformity. Locally, the contact with the Early Mississippian Caballero Formation appears almost gradational. Northward, Devonian strata are erosionally truncated by Pennsylvanian rocks, and locally, such as in the Caballo Mountains, the Percha is locally missing due to deep pre-Pennsylvanian erosion.

Mississippian strata

The Early Mississippian (Kinderhookian) Caballero Formation exists only in an east-trending belt extending from the Lake Valley area to the Sacramento Mountains, including exposures in the Robledo and southern San Andres Mountains. The formation is 18–23 m thick and consists of calcareous shale and siltstone and fossiliferous, nodular limestone.

The Osagean-age Lake Valley Formation, famous for its well-preserved fossils (Kues, 1986), was subdivided by Laudon and Bowsher (1949) into six members, in ascending order, the Andrecito, Alamogordo, Nunn, Tierra Blanca, Arcente, and Doña Ana. The latter two members crop out only in the Sacramento and southern San Andres Mountains. At the type locality, the Andrecito is 20 m of thin-bedded limestone with thin calcareous shales, grading up into the Alamogordo Member, which consists of 10 m of gray, massive limestone with distinctive banded chert nodules. The fossiliferous Nunn Member is 23 m of nodular, argillaceous limestone and calcareous shale that grades up into the Tierra Blanca Member, a gray, medium-bedded to massive limestone with many light-gray chert nodules.

In the southeastern Caballo Mountains and at Lake Valley, only the four lower members are present, disconformably above the Percha Shale and erosionally overlain by basal Pennsylvanian strata. Pre-Pennsylvanian erosion removed the Mississippian from the rest of the Caballo Mountains area. To the east in the San Andres and Sacramento Mountains, the Arcente Member is as much as 60 m thick and is composed of dark-gray, calcareous siltstone, shale, and lenticular silty limestone. The overlying Doña Ana Member is similar lithologically to the Tierra Blanca Member and is up to 53 m thick. The well-known Waulsortian mounds involve the Alamogordo through Tierra Blanca Members and are overlapped by the Arcente Member. The mounds exist mainly in the Sacramento Mountains, but locally are present in the San Andres Mountains, near San Andres Peak (Bachtel and Dorobek, 1994).

In the southern San Andres and Sacramento Mountains, the Meramecian Rancheria Formation overlies the Lake Valley Formation, is up to 91 m thick, and consists primarily of black, cherty, silty, thin-bedded limestones. Calcareous shale and interbedded fossiliferous and oolitic limestones of the Chesterian-age Helms Formation overlie the Rancheria only in the southern Sacramento Mountains, Bishop Cap, and Franklin Mountains, thickening southward to 75 m in the Hueco Mountains. During Osagean and Meramecian time, most of New Mexico was covered by shallow seas (Armstrong et al., 1979), but erosion in late

Mississippian and early Pennsylvanian removed Mississippian and older rocks, generally north of the pre-Devonian erosional edge.

Pennsylvanian strata

Pennsylvanian rocks in south-central New Mexico are thick (200–920 m) and complex, chiefly of shallow-marine origin, with basal siliciclastic beds in the north and east related to pre-Pennsylvanian erosion and initial development of the Pedernal uplift (Kottlowski, 1960, 1963; Wilson, 1989). The Middle Pennsylvanian interval is primarily a thick succession of marine carbonate rocks, but in the northern Sacramento Mountains a deltaic complex is present, fed by braided streams flowing westward from the Pedernal uplift (Pray, 1961; Van Wagoner, 1977). Missourian units in most areas display an upward increase in siliciclastics, and continuing uplift of the Pedernals climaxed in Virgilian and early Wolfcampian time with a thick sequence of cyclic fluvial to shallow-marine sediments of the Panther Seep Formation, which was deposited in the Orogrande Basin (Pray, 1959; Schoderbek and Chafetz, 1988). Strata in the Fra Cristobal, Caballo, and Robledo Mountains were deposited on a marine shelf west of the Orogrande Basin. The Pedernal uplift and complementary Orogrande Basin are part of the Ancestral Rocky Mountains, which probably represent cratonic deformation inboard of the Marathon-Ouachita collisional orogen (Kluth and Coney, 1981; Kluth, 1986).

Formational names of Pennsylvanian strata differ among the various mountain ranges. Most Pennsylvanian rocks have been included in the Magdalena Group. In the Sacramento Mountains, the formations are, in ascending order, Gobbler, Beeman, and Holder (Pray, 1959); in the Franklin Mountains, they are the La Tuna, Berino, Bishop Cap, and Panther Seep Formations; in the Organ and southern San Andres Mountains, they are the Lead Camp and Panther Seep Formations (Bachman and Myers, 1963; Seager, 1981); in the northern San Andres Mountains, they are the Sandia, Lead Camp, and Panther Seep Formations (Bachman and Harbour, 1970); in the Caballo and Fra Cristobal Mountains, they are the Red House, Nakaye, and Bar B Formations (Kelley and Silver, 1952; Seager and Mack, 1991). The units are Atokan siliciclastics and limestones in the north and east, Morrowan(?) and Atokan siliciclastics and limestones in the Caballo Mountains, and Morrowan limestones in the south. Desmoinesian intervals are thick-bedded, cherty limestones, Missourian beds display an upward increase in siliciclastic detritus, and Virgilian strata range from massive limestone in the Robledo Mountains to mixed siliciclastic-limestone of the Panther Seep Formation.

The basal erosional unconformity of the Pennsylvanian rocks in most of the area is on Mississippian strata, except near the northern truncation of all of the pre-Pennsylvanian Paleozoic beds. In the Sacramento Mountains, basal Pennsylvanian channels are filled with conglomeratic sandstones; in the San Andres Mountains, chert-pebble conglomerates and chert breccia are present; and in the Caballo Mountains quartz sandstones and cherty karst breccias locally mark the contact (Kalesky, 1988). In the southwestern foothills of the Caballo Mountains, an eroded domal uplift, centered near the Red Hills, exposes Pennsylvanian strata overlying rocks ranging from the Lake Valley Formation to the El Paso Formation (Seager, 1986).

Permian strata

In the northern San Andres and Oscura Mountains, a unit of intertongued marine limestones, redbeds, and arkoses containing earliest Wolfcampian fusulinids is named the Bursum Formation (Wilpot and Wanek, 1951; Thompson, 1942, 1954). The Bursum has been traced southward in the range to the central Hembrillo Canyon area where it appears to grade into and intertongue with lower Hueco strata. The Laborcita Formation in the Sacramento Mountains is correlative to the Bursum, as is 55 m of mostly limestones bearing early Wolfcampian fusulinids in the Robledo Mountains (Thompson, 1954). In the Caballo Mountains, the uppermost part of the Bar B and lowermost Abo Formations are equivalent to the Bursum (Seager and Mack, 1991).

The Abo Formation "redbeds" consist of nonmarine siliciclastic sediments that interfinger southward with gray marine limestone and shale of the Hueco Formation (Fig. 4). The facies change takes place in a transitional zone about 25 km wide that trends northeastward across Doña Ana and Luna Counties, before turning eastward and then southeastward in Otero County (Kottlowski, 1963, 1965; Jordan, 1975; Mack and James, 1986; Mack et al., 1988a, 1995). Rocks of the transitional zone are well exposed in the Doña Ana and Robledo Mountains, where they consist of an interbedding of typical Hueco marine limestone and shale and Abo tidal-flat red sandstone and siltstone (Mack and James, 1986; Lucas et al., 1995).

The Abo Formation is 141–300 m thick in the Caballo Mountains, 254 m in the north-central San Andres Mountains near Rhodes Canyon, and almost 335 m thick in the Sacramento Mountains near Tularosa, but thins southward to 60–150 m where it interfingers with the Hueco Formation. In the Robledo Mountains, the Abo Tongue is 145 m thick, is underlain by 271 m of the Lower and Middle Members of the Hueco Formation, and is overlain by 108 m of the Upper Member of the Hueco Formation. Faunas in the Robledo Mountains date the Hueco Formation as middle and late Wolfcampian (LeMone et al., 1975; Jordan, 1975; Kues, 1995). In the vicinity of the Pedernal uplift, the Abo overlies mainly Precambrian rocks and is 60–180 m thick.

The Hueco Formation consists of argillaceous, locally cherty limestones and gray fossiliferous shales. In the San Andres Mountains, reddish shales and calcareous siltstones grade upward into the Abo redbeds. In the central and southern areas, as well as in western foothills of the northern Franklin Mountains, the basal unit is a massive biostromal limestone. Southward in the San Andres Mountains, the Hueco thickens from 98 m at Mockingbird Gap, to 127 m in Rhodes Canyon, to 413 m in the southern part of the range. In the Franklin Mountains, the Hueco Formation is 670 m thick (Harbour, 1972) with some thin persistent interbeds of yellowish siltstone, shale, and thin red siltstone in the upper part of the formation.

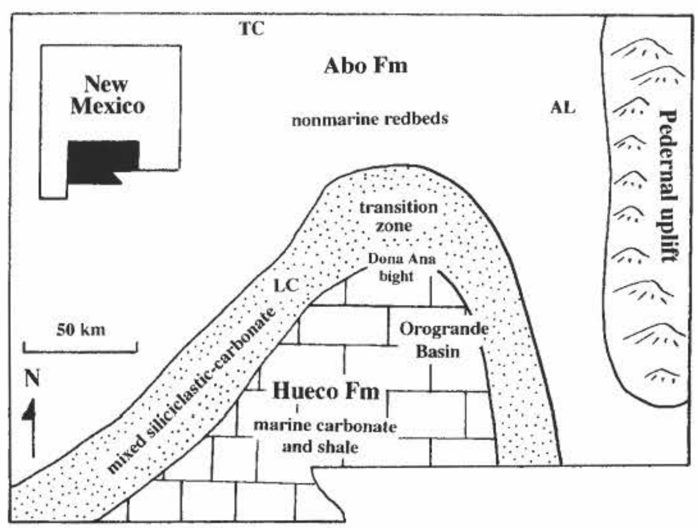

FIGURE 4. Early Permian paleogeographic map, adapted from Mack et al. (1986, 1995). LC = Las Cruces; AL = Alamogordo; TC = Truth or Consequences.

The Leonardian-age Yeso Formation conformably overlies the Abo Formation with the contact being based on the change from brick-red siltstones and shales of the Abo to orange and green sandstones of the basal member of the Yeso. At Rhodes Canyon in the north-central San Andres Mountains, the four members of the Yeso Formation named by Wilpolt and Wanek (1951) in the Oscura Mountains are well exposed. In ascending order, they are the Meseta Blanca Sandstone, Torres Member, Cañas Gypsum, and Joyita Sandstone, totaling 486 m thick. The Cañas Gypsum pinches out south of Rhodes Canyon, and no gypsum is present in the southern part of the range.

In the Caballo Mountains, the Meseta Blanca Member was recognized, but overlying members are dissimilar to those of Wilpolt and Wanek (1951). Three upper members were described in the Caballo Mountains, including in ascending order, the Red Siltstone-Dolomite, Limestone, and Sandstone-Limestone Members (Mack and Suguio, 1991). In the McLeod Tank quadrangle of the Caballo Mountains (Seager and Mack, in press b), the Yeso ranges in thickness from 214–396 m, the variation resulting from deeper pre-Late Cretaceous erosion of the thinner sections and eastward thickening of the Red Siltstone-Dolomite Member, especially among the gypsum beds. Mack and Suguio (1991) suggest that most of the sandstone beds in the Yeso Formation in the Caballo Mountains are eolian, that the carbonate rocks were deposited in tidal-flat, lagoonal, and open-marine environments, and that the gypsum is probably lagoonal in origin.

In the Sacramento Mountains, the northern sections of the Yeso Formation are about 427 m thick and contain more gypsum and pale-red siliciclastic units than the southern sections, which have a thickness of 378 m. Southward on Otero Mesa, the Yeso crops out over large areas, but complete sections are not reported. In general, the Yeso was removed by post-Permian erosion (Early Cretaceous and/or Laramide) from the area south of Las Cruces, but is locally present in the East Potrillo Mountains (Seager and Mack, 1994).

The type section of the San Andres Formation is Rhodes Canyon in the San Andres Mountains (Lee and Girty, 1909), where the formation is more than 185 m thick. Lenses of friable, yellowish-brown sandstone interbedded with limestone near the base of the formation may be correlative with the Glorietta Sandstone of central New Mexico. Above the basal arenaceous beds, the San Andres Formation consists of gray to dark-gray, medium-bedded to massive, fetid, fossiliferous limestone. The Guame Nos. 1 and 2 wells west of Hembrillo Pass encountered 138 m of San Andres beneath the Upper Cretaceous Dakota Sandstone (Kottlowski et al., 1956). In the southern part of the range, the unit labeled San Andres Formation by Kottlowski et al. (1956) includes thick beds of yellowish to pale-red sandstone and siltstone, and may be equivalent to middle calcareous beds of the Yeso Formation west of Hembrillo Pass (Bachman and Myers, 1963; Seager, 1981). Kelley and Silver (1952) mapped the San Andres Formation in the central and northern part of the Caballo Mountains, but Mack and Suguio (1991) and Seager and Mack (in press b) believe these rocks belong to the upper two members of the Yeso Formation.

CRETACEOUS SEDIMENTATION IN THE BISBEE BASIN/CHIHUAHUA TROUGH AND THE WESTERN INTERIOR FORELAND BASIN

Upper and Lower Cretaceous sedimentary rocks are exposed in south-central New Mexico, although neither series is widely distributed (Fig. 5). Lower Cretaceous rocks are well exposed in the East Potrillo Mountains and were first described by Darton (1928) and subsequently by Bowers (1960), Hoffer (1976), and Powell (1983). Recently, Seager and Mack (1994) applied the stratigraphic terminology originally developed in the Big Hatchet Mountains by Zeller (1965) to Lower Cretaceous rocks of the East Potrillo Mountains, including, in ascending order, the Hell-to-Finish, U-Bar, and Mojado Formations (Fig. 5).

The Hell-to-Finish Formation in the East Potrillo Mountains is divided into a lower Conglomerate Member and an upper Mottled Siltstone Member. Ranging from 1 to 39 m thick, the Conglomerate Member consists of grain-supported pebbles and cobbles of Paleozoic carbonate and chert. The Mottled Siltstone Member conformably overlies the conglomerate member, has a maximum thickness of 168 m, and is primarily composed of calcareous siltstone and very fine sandstone, although thin beds (<0.5 m) of pebble conglomerate are locally present. Common bioturbation, bivalve and gastropod fossils, wave oscillation ripples, and hummocky cross-stratification suggest the Mottled Siltstone Member is shallow marine in origin.

The U-Bar Formation in the East Potrillo Mountains is subdivided into six conformable members, which in ascending order are the Lower Limestone (33 m), Sandstone (66 m), Rudistid Limestone (13 m), Siltstone-Limestone (90 m), Massive Limestone (133 m), and Upper Siltstone (15 m) Members (Seager and Mack, 1994). All of the members of the U-Bar Formation appear to be shallow marine in origin, because of the presence of numerous gastropod and bivalve fossils, abundant bioturbation, and hummocky cross-stratification in the sandstones. The foraminifera *Orbitolina gracilis* in the Lower Limestone Member and *Orbitolina grossa* in the Rudistid Limestone Member indicate an early Albian age for at least the lower part of the U-Bar Formation in the East Potrillo Mountains (Douglass, 1960; Seager and Mack, 1994).

Quartz sandstones assigned by Seager and Mack (1994) to the Mojado Formation are only exposed in three low hills west of the main fault block of the East Potrillo Mountains. At this location the Mojado Formation is approximately 15 m thick and displays meter-scale trough and planar crossbeds. Because the contacts of the formation are not exposed and no age-diagnostic fossils have been found, designation as the Mojado Formation is tenuous.

Cretaceous sedimentary rocks exposed at Cerro de Cristo Rey have a cumulative thickness of 422 m and have been separated by Lovejoy (1976) into nine formations (Fig. 5). The lower five formations are Albian in age and correlate with the U-Bar and Mojado Formations farther west, while the upper four formations are early and middle Cenomanian in age and may correlate with the Dakota and part of the Mancos Formations exposed in the Caballo and San Andres Mountains (Fig. 5). Two of the Cretaceous formations at Cerro de Cristo Rey are predominantly limestone (Finlay, Buda), two others are predominantly dark gray shale (Mesilla Valley, Boquillas), and the Anapra Formation is composed primarily of sandstone with minor amounts of shale. The remaining four formations (Del Norte, Smeltertown, Muleros, Del Rio) consist of interbedded limestone and gray shale or siltstone. All of the formations are shallow marine in origin, with the possible exception of part of the Anapra Formation.

Lower Cretaceous sedimentary rocks are also exposed near Love Ranch in the southern San Andres Mountains and were assigned by Seager (1981) and Lucas and Estep (this guidebook) to the Sarten Formation. At this location the Sarten Formation is about 26 m thick, unconformably overlies Permian limestone, and is unconformably(?) overlain by the Upper Cretaceous Dakota Sandstone. At the base of the Sarten Formation is a thin (<1 m), discontinuous, locally glauconitic chert-pebble conglomerate, while the remainder of the formation is composed of poorly exposed gray shale and thin (<1 m) brown sandstones exhibiting bioturbation and hummocky stratification. Albian bivalves are present near the middle of the section (Seager, 1981; Lucas and Estep,

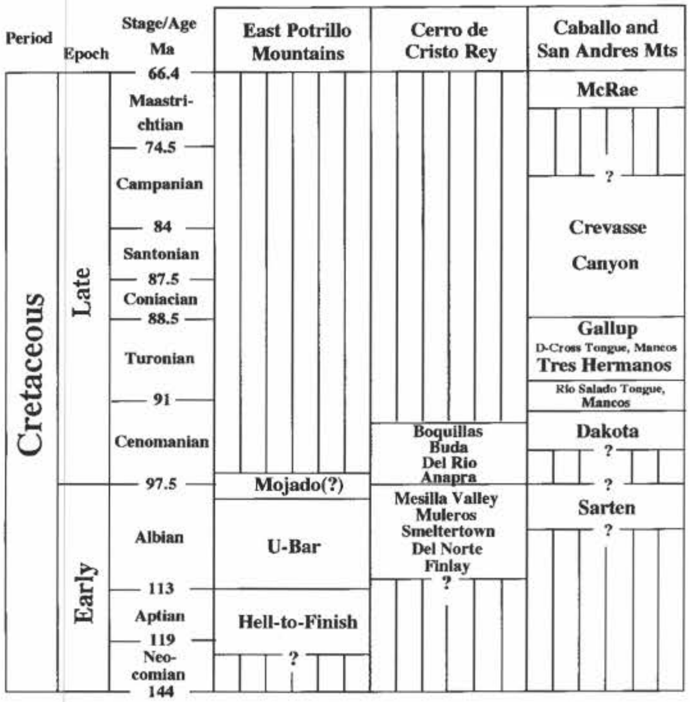

FIGURE 5. Cretaceous stratigraphy in south-central New Mexico.

this guidebook). The Sarten Formation in the southern San Andres Mountains is thinner and lacks the thick pebbly sandstone beds present in the Sarten Formation in the Cooke's Range north of Deming (Mack et al., 1988b; Lucas et al., 1988).

Upper Cretaceous sedimentary rocks are widely exposed along the northeastern flank of the Caballo Mountains, in the Cutter Sag between the Caballo Mountains and Fra Cristobal Range, in Apache Graben in the south-central part of the Caballo Mountains, and near Love Ranch in the southern San Andres Mountains (Fig. 1). Following the lead of Wallin (1983), we advocate applying stratigraphic terminology from central and northern New Mexico to these rocks, including, in ascending order, Dakota, Mancos, Tres Hermanos, D-Cross Tongue of the Mancos, Gallup, and Crevasse Canyon Formations (Fig. 5; Seager and Mack, in press b, c; Mack and Seager, in press). Only the Dakota, Mancos, and lower Tres Hermanos Formations are exposed in Apache Graben (Seager and Mack, in press b) and only the Dakota through Gallup(?) are present near Love Ranch (Seager, 1981).

The Dakota Sandstone has a maximum thickness of about 30 m and consists of interbedded quartz sandstone, locally pebbly, and gray shale. The lower part of the formation is interpreted to be fluvial in origin, whereas the upper part consists of lagoonal/flood tidal delta or shallow marine shale and sandstone. Based on outcrops in the Caballo Mountains and geophysical logs from the adjacent Jornada del Muerto Basin, Bauer (1989) suggested that the lower, fluvial part of the Dakota infilled large paleovalleys cut into the underlying Yeso Formation and is thicker in the paleovalley floors than on the interfluves. In contrast, the upper marine interval has a similar thickness throughout the area of study.

Conformably overlying the Dakota Sandstone is the late Cenomanian to early Turonian Mancos Shale. The lower interval of Mancos Shale is sometimes referred to as the Rio Salado Tongue to differentiate it from the D-Cross Tongue, which appears stratigraphically higher in the section. The Rio Salado Tongue of the Mancos consists of approximately 90 m of gray shale, siltstone, and brown very fine- to fine-grained sandstone of lower shoreface to offshore marine origin. Bioturbation is common, as are whole or fragments of bivalves. Several thin (<20 cm), light tan bentonite beds are present in the lower third of the formation. The greatest abundance and thickness of sandstone and siltstone beds exist in the basal and upper third of the Rio Salado Tongue of the Mancos Shale, suggesting the time of maximum water depth occurred during deposition of the middle of the formation. Together the upper Dakota and lower part of the Rio Salado Tongue constitute the T1 transgression of Molenaar (1983).

The Tres Hermanos Formation, which gradationally overlies the Rio Salado Tongue of the Mancos Shale, is approximately 100 m thick and is entirely Turonian in age (Hook, 1983). The basal 15 m of the formation consists of a progradational shoreline sequence of bioturbated sandstones containing crossbeds, horizontal laminae, oscillation ripple marks, and bivalve shell debris. The thick (10 m), laterally continuous shoreline sandstone, which is designated the Atarque Member, makes a prominent ridge that easily marks the base of the Tres Hermanos Formation, and is overlain by several meters of rooted, bioturbated argillaceous sandstone, locally containing coal, deposited in coastal lagoons and marshes. The middle part of the Tres Hermanos Formation, called the Carthage Member, is fluvial in origin and consists of thick (3–6 m) ledges of crossbedded fine- to medium-grained sandstone deposited in channels and interbedded olive gray and dark gray mudstone and thin (<1 m) very fine sandstones deposited on the floodplain. The stratigraphic interval from the middle of the Rio Salado Tongue of the Mancos Shale through the middle of the Tres Hermanos Formation constitutes the R1 regression of Molenaar (1983). Shallow-marine to shoreline shale and sandstone are present in the upper 15 m of the Tres Hermanos Formation. The prominent, light-colored marine sandstone at the top of the Tres Hermanos Formation is the Fite Ranch Member. The T2 transgression of Molenaar (1983) began with marine sediment of the upper Tres Hermanos Formation and continued with deposition of the lower part of the D-Cross Tongue of the Mancos Shale, which consists of about 50 m of marine gray shale and siltstone and is late Turonian in age (Hook, 1983). The D-Cross coarsens upward and grades into the Gallup Sandstone, marking the R2 regression of Molenaar (1983). The Gallup Sandstone is about 40 m thick and was deposited in delta front, distributary mouth bar, and lower delta plain environments. In central New Mexico, at Carthage, the Gallup straddles the Turonian-Coniacian boundary, although its exact age in south-central New Mexico is not known.

The Gallup Sandstone is conformably overlain by the Crevasse Canyon Formation, which consists of about 600 m of fluvial sandstone, shale, and pebble conglomerate. Olive gray shales commonly contain well-developed paleosols characterized by A, E, and argillic B horizons indicative of a well-drained floodplain in a humid to subhumid paleoclimate (Mack, 1992). One sample of pollen from the middle of the Crevasse Canyon Formation indicates a Santonian to early Campanian age (Clemons and Mack, 1988). Wallin (1983) divided the Crevasse Canyon Formation into three members, the lower Coal-Bearing Member, middle Barren Member, and upper Ash Canyon Member. This convention is difficult to apply elsewhere in the region, and Seager and Mack (in press c) and Mack and Seager (in press) generally divided the formation into Lower and Upper Members for mapping purposes. The Crevasse Canyon Formation is unconformably overlain by the uppermost Cretaceous McRae Formation, which will be discussed in the context of Laramide tectonics and sedimentation.

Cretaceous sedimentary rocks in south-central New Mexico were deposited in two different sedimentary basins. Lower Cretaceous sediment was deposited in a northwest-trending, extensional basin referred to as the Chihuahua trough in west Texas and northern Mexico and the Bisbee Basin in southeastern Arizona (Kottlowski, 1965; Greenwood et al., 1977; Bilodeau, 1982; Mack et al., 1986; Mack, 1987). The initial phase of extension is primarily represented by the Hell-to-Finish Formation and probably involved numerous sub-basins and complementary block uplifts. A subsequent phase of more regional subsidence and marine transgressions and regressions accompanied deposition of the U-Bar and Mojado Formations. Most of the area of this study was situated on a gently northward-tilted rift shoulder, as evidenced by Lower Cretaceous rocks being thin or absent and by a subcrop relationship in which in a northerly direction progressively younger stratigraphic units underlie the Upper Cretaceous Dakota Sandstone. Although the existence of the Chihuahua trough/Bisbee Basin is well established, the tectonic origin remains an open question, with suggestions including a failed arm of the Gulf of Mexico rift, back-arc extension, or extension related to strike-slip or oblique-slip faulting (Bilodeau, 1982; Dickinson et al., 1986; Saleeby and Busby-Spera, 1992; Fackler-Adams et al., 1997).

Upper Cretaceous sedimentary rocks (Dakota-Crevasse Canyon) were deposited within or on the southwestern margin of the Western Interior Seaway, which periodically covered all or a large part of the Rocky Mountain states of the USA and provinces of Canada (McGookey et al., 1972). In the central and northern Rocky Mountain region, the Western Interior Seaway is considered to have been part of a foreland basin that subsided in response to the emplacement of thrust sheets in the Sevier fold and thrust belt (Jordan, 1981; Lawton, 1985, 1994; Heller et al., 1986; DeCelles and Currie, 1996). It is not clear, however, whether this model also applies to Upper Cretaceous rocks in New Mexico, because of their overall thinness compared to farther north and because of the lack of identifiable coeval thrust faults that would have provided the tectonic load.

LARAMIDE (LATEST CRETACEOUS–EOCENE) OROGENESIS

The Laramide orogeny was a compressional tectonic event in the Western Interior of the United States that occurred from latest Cretaceous through the Eocene (Dickinson et al., 1988). The region that was the Western Interior foreland basin in Late Cretaceous time (Cenomanian through Campanian) was subsequently deformed into a series of basement-cored block uplifts separated by nonmarine intermontane basins. Andesitic arc volcanism became largely extinct west of the central segment of the Laramide broken foreland, but spread eastward into the extreme northern and southern parts of the deformed belt (i.e., Montana and New Mexico). In south-central New Mexico, a major Laramide uplift, called the Rio Grande uplift, and its complementary Love Ranch and Potrillo Basins have been identified by Seager and Mack (1986), Seager et al. (1986), Seager and Mayer (1988), and Seager et al. (1997). Andesitic volcanism both preceded and postdated activity of the Rio Grande uplift.

The Rio Grande uplift was a large, northwest-trending block uplift bounded on the northeast by southwest-dipping thrust and reverse faults and associated folds (Fig. 6; Seager and Mack, 1986; Seager et al., 1986, 1997). It consisted of the basement-cored main uplift and a structural bench on the northeastern margin that was structurally lower than the main uplift and was composed of broadly arched Paleozoic sedimentary rocks (Fig. 6; Seager et al., 1997). The Love Ranch Basin was situated north of the Rio Grande uplift and contained approximately 1460 m of syn- and post-orogenic detritus belonging to the McRae and Love Ranch Formations. The basin to the southwest of the Rio Grande uplift, the Potrillo Basin, is known primarily from the Grimm and others well, which contains 2100 m of lower Teritary siliciclastic rocks (Thompson, 1982). However, exposed on the southwestern flank of the uplift are conglomerates of the Love Ranch Formation that infilled paleovalleys

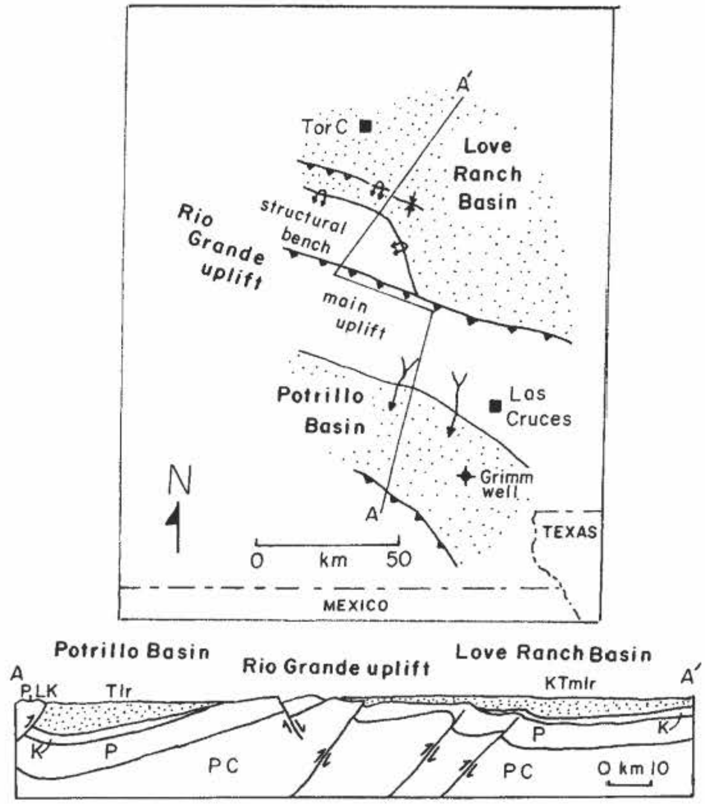

FIGURE 6. Generalized paleotectonic map and cross section of the Laramide (latest Cretaceous–Eocene) Rio Grande uplift and Potrillo and Love Ranch Basins, based on Seager and Mack (1986), Seager et al. (1986), and Seager et al. (1997).

that drained into the Potrillo Basin (Seager et al., 1986).

The McRae Formation was named by Kelley and Silver (1952) for exposures east of Elephant Butte Lake. Bushnell (1953, 1955) divided the McRae Formation into a lower Jose Creek Member (170 m) and an upper Hall Lake Member (235 m), which were used as map units by Lozinsky (1986) in the Elephant Butte quadrangle and Mack and Seager (in press) in the Engle quadrangle. The McRae Formation is considered to be latest Cretaceous (late Maastrichtian) in age, based on the presence of a Lancian dinosaur fauna (Fig. 5; Lozinsky et al., 1984; Wolberg et al., 1986; Gillette et al., 1986; Lucas et al., this guidebook). It is possible that part of the Hall Lake Member is Paleocene in age, but the K-T boundary has not been identified within the member. Both members of the McRae Formation consist primarily of fluvial sandstones, pebble to cobble conglomerates, and mudstones. Non-calcic paleosols are common in the Jose Creek Member, whereas calcic paleosols predominate in the Hall Lake Member, suggesting a change to drier climate during the late Maastrichtian (Buck and Mack, 1995). In addition, the Jose Creek Member contains from 6 to 10 thin (<1 m) siliceous beds that are interpreted to be fallout ashes (Mack and Seager, in press). The presence of only one ash bed and distinctly purple-to-red colors distinguish the Hall Lake Member from the Jose Creek Member. Petrified wood, including in situ stumps, are common in the Jose Creek Member, as are leaf fossils, particularly in the fallout ash beds (Upchurch and Mack, this guidebook).

The Love Ranch Formation was named by Kottlowski et al. (1956) for exposures of conglomerate near the abandoned Love Ranch in the southern San Andres Mountains. Exposed widely in south-central New Mexico, the Love Ranch Formation consists not only of conglomerate, but also of arkosic sandstone, red mudstone,

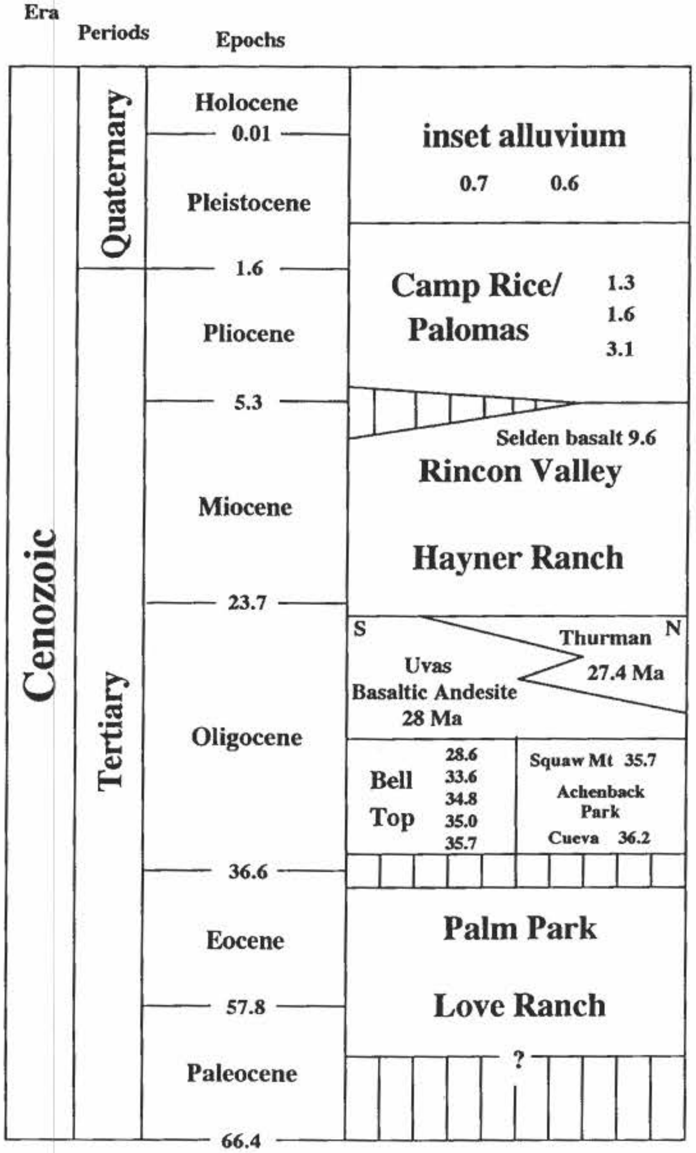

FIGURE 7. Cenozoic stratigraphy of south-central New Mexico. Numbers associated with stratigraphic units are radioisotopic ages or inferred ages in millions of years, based on Seager et al. (1975), Kortemeier (1982), Seager et al. (1984), McIntosh et al. (1991) and Mack et al. (1993, 1996).

and gypsum, depending on the paleogeographic position within the Love Ranch and Potrillo Basins. The Love Ranch Formation is considered to be Eocene in age, because of its conformable upper contact with the mid-late Eocene Palm Park Formation (Fig. 7). The age of the base is not known.

The kinematic evolution of the Rio Grande uplift and Love Ranch Basin has recently been described in detail by Seager et al. (1997). Laramide compression began in south-central New Mexico between Campanian and late Maastrichtian time, creating a broad, nearly symmetrical uplift that was capped by Upper Cretaceous arc-related intermediate to silicic volcanic rocks. With the exception of intrusive rocks exposed in the Animas Mountains (Hedlund, 1977; Kelley and Chapin, 1997), these igneous rocks were completely removed by erosion from the Rio Grande uplift and exist only as clasts in the McRae and basal Love Ranch Formations. During this early phase of development, the northeast flank of the Rio Grande uplift acted as a surface of bypass for sediment that accumulated as the McRae Formation about 40 km northeast of the main uplift in the incipient Love Ranch Basin.

Following an interruption in tectonism that may have lasted several million years, deformation resumed in early Tertiary time, resulting in deposition of up to 900 m of sediment of the Love Ranch Formation in the Love Ranch Basin. The basal conglomerate of the Love Ranch Formation extended up to 40 km from the main uplift along a northeast-tilted surface, suggesting that the uplift remained nearly symmetrical. However, the remainder of the Love Ranch Formation was deposited in an asymmetrical basin that resulted from movement on the thrust and reverse faults that bordered the main uplift and structural bench. During this time, Precambrian crystalline basement was exposed in the main uplift and the structural bench was deeply eroded. Eventually, sediment of the Love Ranch Formation onlapped the structural bench and ultimately buried the main uplift.

Immediately following deposition of the Love Ranch Formation, south-central New Mexico experienced andesitic volcanism represented by the Palm Park Formation, named by Kelley and Silver (1952) for exposures in the southern Caballo Mountains. Varying from a few hundred to 600 m thick, the Palm Park Formation is thickest in Laramide basins and thins over the Rio Grande uplift. The Palm Park Formation is correlative with the Cleofas Andesite in the Doña Ana Mountains (Seager et al., 1976) and with the Orejon Andesite in the Organ Mountains (Seager, 1981), and may be the same age as, but from a different source than the Rubio Peak Formation exposed in the Good Sight Mountains and areas west of the study area (Clemons, 1979). A middle–late Eocene age for the Palm Park Formation is based on K-Ar radioisotopic dates of 51.5 ± 2.6, 42.4 ± 1.6, and 43 Ma (Kottlowski et al., 1969; Clemons, 1979). These dates are in general agreement with those of the Rubio Peak Formation and associated rocks, which include dates of 44.7 ± 1.9, 40.7 ± 1.9, 38.8 ± 1.4, 38.0 ± 1.5, 38.0 ± 1.9, 37.6 ± 2.0, and 36.7 ± 1.4 Ma (Loring and Loring, 1980; Clemons, 1982).

The Palm Park Formation exhibits a wide variety of rock types, including bouldery lahars, crossbedded sandstones, purple and red mudstones, porphyritic andesite, latite, dacite lava flows and tuffs, and various sills and dikes. In addition, the Palm Park Formation exposed in Apache Graben in the Caballo Mountains contains thick ledges of carbonate representing hydrothermal tufa deposits (Chafetz et al., 1991). The volcanic vents responsible for the Palm Park Formation have not been located within the study area, with the possible exception of the Doña Ana Mountains (Seager et al., 1976).

OLIGOCENE BIMODAL VOLCANISM AND THE ONSET OF THE RIO GRANDE RIFT

The Oligocene epoch in southwestern New Mexico was characterized by widespread bimodal volcanism involving large, silicic cauldrons and their associated outflow sheets and volcaniclastic sedimentary rocks, and lava flows of basalt and basaltic andesite (McIntosh et al., 1991). It was also the time of initiation of crustal extension in the Rio Grande rift. New evidence in central and southern New Mexico suggest that extension may have begun as early as 35 Ma, instead of 27–28 Ma as previously thought (Seager, 1975; Chapin and Seager, 1975; Cather, 1990; Mack et al., 1994a). The principle tectonic and volcanic features of Oligocene age in south-central New Mexico were the Organ and Doña Ana cauldrons, the Goodsight-Cedar Hills depression, and the Cedar Hills vent zone (Fig. 8).

Organ and Doña Ana cauldrons

Fragments of the Organ and Doña Ana cauldrons of early Oligocene age are exposed in the Organ and Doña Ana Mountains, respectively. The size and extent of neither cauldron is known, but compositional differences between them suggest each evolved from a different magma source or from different cupolas above a large common magma chamber. Although thick intracauldron ash-flow tuffs and lavas distinguish both cauldrons, no outflow sheets have been positively identified. Most studies in the past 20 yrs have focused on the Organ cauldron (Seager, 1981; Newcomer, 1984; Beyer, 1986; Seager and McCurry, 1988; Fitzpatrick, 1989; Butcher, 1990; Yanicak, 1992; Verplanck et al., 1995).

The eastern margin of the Organ cauldron and its underlying plutonic roots, the Organ batholith, are exposed through a vertical section 5–6 km thick in the Organ Mountains. Both cauldron-fill tuffs and lavas, as well as the batholith, document emplacement of a compositionally zoned magma chamber at shallow depths between 36.2 and 35.5 Ma (Seager, 1981; Seager and McCury, 1988; McIntosh et al., 1991). As much as 3.3 km thick, intracauldron tuffs and lavas, corresponding to the Cueva, Achenback Park, and Squaw Mountain Formations (Fig. 7), are normally zoned and indicate eruption from a shallow siliceous magma chamber or cupola that ranged from aphyric rhyolite (77% SiO_2) at the top to crystal-rich rhyolite and trachyte (68–66% SiO_2) at deeper levels. Compositions of the latest tuffs and lavas overlap compositions of the oldest phase of the Organ batholith, which may be interpreted as the undrained residue of the magma chamber. This phase of the batholith is also chemically and mineralogically zoned from quartz syenite (68% SiO_2) at the highest exposures to monzodiorite (55% SiO_2), which forms a diapir-like pluton near the lowest exposed levels. Small bodies of fine-grained alkali feldspar granite at the top of the batholith may represent unerupted remnants of the siliceous cap or cupola. Three younger plutonic phases of the batholith seemingly represent three stages of silicic core magma that formed and was intruded during progressive crystallization and differentiation of the batholith. Major normal faults that transect intracauldron tuffs and are invaded by the batholith suggest the cauldron evolved in a regional extensional stress field.

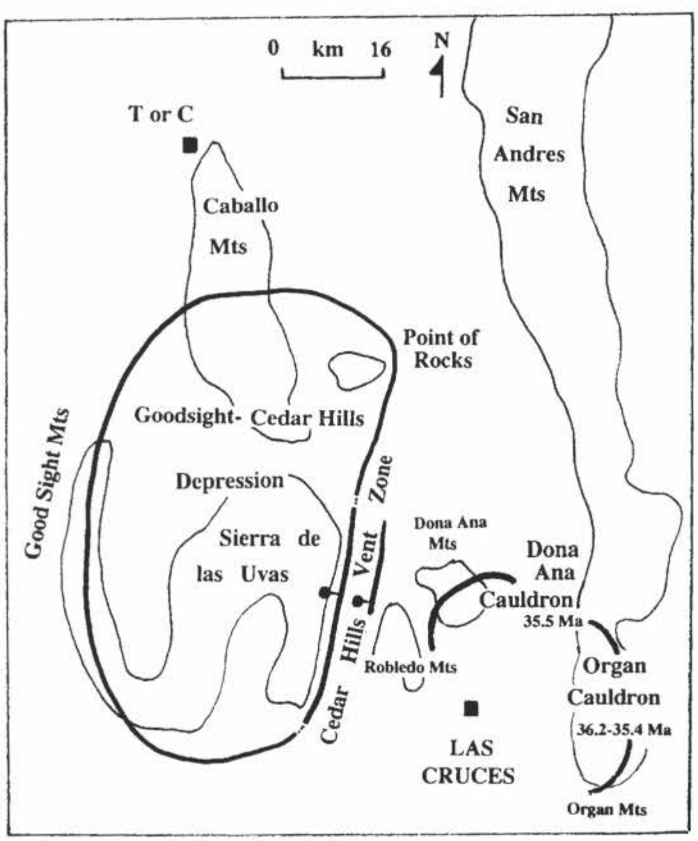

FIGURE 8. Paleotectonic features of early Oligocene age in south-central New Mexico, superimposed on the distribtution of modern fault-block mountains. Numbers are radioisotopic ages from McIntosh et al. (1991).

The Doña Ana cauldron is exposed in the southern half of the Doña Ana Mountains, where intracauldron rhyolite ash-flow tuff at least 600 m thick forms the bold, south-facing mountain front (Seager et al., 1976; Haga, 1994). Radioisotopic dates ranging from 33 to 37 Ma have been obtained from intracauldron tuffs, named Doña Ana Rhyolite, and from syenite plutons that transect the tuff (Seager et al., 1976; McIntosh et al., 1991). Although syenitic dikes and irregular bodies have invaded the cauldron fill in many places, a prominent group of thick dikes trends easterly across the central part of the range. This group separates intracauldron fill on the south side of the dikes from pre-cauldron andesite and Paleozoic rocks to the north. The dikes are interpreted to mark the northern structural margin of the Doña Ana cauldron. Locally within the cauldron, tuffaceous sedimentary rocks, landslide debris, and thin ash-flow tuffs, as much as 300 m thick, overlie the Doña Ana Rhyolite. These are also invaded by the syenitic plutons as well as by masses of flow-banded rhyolite and small bodies of andesite. Collectively, the sedimentary and intrusive rocks represent waning stages of subsidence, sedimentation, and igneous activity that followed initial voluminous eruptions of Doña Ana Rhyolite and formation of the cauldron. Because of the westward tilt of the Doña Ana Mountains, erosion of the cauldron is deeper on the eastern side of the range, extending into the floor of the cauldron and revealing plutonic intrusive rocks. In contrast, along the western flank of the range higher levels of the cauldron complex are preserved, including flow-banded rhyolite dome-flow complexes and the youngest parts of the intracauldron fill.

Bell Top Formation (Early Oligocene)

The Goodsight-Cedar Hills depression was an elliptical, subsiding trough in which was deposited from 55 to 270 m of volcanic and sedimentary rocks of the Bell Top Formation, which is exposed in the Good Sight Mountains, Sierra de las Uvas, Apache Graben, Rincon Hills, Point of Rocks, Cedar Hills, Rough and Ready Hills, and Sleeping Lady Hills (Figs. 1 and 8; Kottlowski, 1953; Seager, 1973; Clemons and Seager, 1973; Seager and Hawley, 1973; Seager et al., 1975; Clemons, 1975). The most prominent rocks in the Bell Top Formation are six mappable silicic ash flow tuffs (AFT), numbered 2 through 7, which range in thickness from 5 to 90 m. AFT 2, 3, and 4 are restricted to the Sierra de las Uvas and adjacent Cedar Hills, while AFT 5, 6, and 7 are found throughout the depression. Single-crystal sanidine $^{40}Ar/^{39}Ar$ analyses indicate the following ages for the upper five ash-flow tuffs: AFT3 = 35.7 Ma, AFT4 = 35.0 Ma, AFT5 = 34.8 Ma, AFT6 = 33.6 Ma, AFT7 = 28.6 Ma (McIntosh et al., 1991). AFT 2, 3, and 7 are vitric, while AFT4, 5, and 6 are more crystal-rich, with phenocrysts of sanidine, plagioclase, quartz, and biotite (Clemons, 1975). Major-element chemical data of Clemons (1975) indicate the ash-flow tuffs generally exceed 70 % SiO_2, with the exception of one analysis of AFT4 that has 67 % SiO_2. All of the ash-flow tuffs except AFT3 probably represent outflow sheets derived from outside the Goodsight-Cedar Hills depression. AFT3 was probably erupted from the Cedar Hills vent zone (Seager, 1973; Clemons and Seager, 1973; Seager et al., 1975), whereas AFT2 may be an outflow sheet from the Organ or Doña Ana cauldrons. AFT4 may correlate with a tuff near Winston or is perhaps a precursor of the Kneeling Nun tuff of the Emory cauldron, located approximately 120 km northwest of the Goodsight-Cedar Hills depression (McIntosh et al., 1991). The Kneeling Nun tuff is probably correlative with AFT5, although map relationships in the Good Sight Mountains suggest that AFT5 may underlie the Kneeling Nun Tuff (Clemons, 1979). AFT6 correlates with the Box

Canyon tuff, derived from the Schoolhouse Mountain cauldron located 200 km west-northwest of the Goodsight-Cedar Hills depression, and AFT7 is the Vick's Peak tuff, derived from the Nogal Canyon cauldron located 150 km north-northwest of the Goodsight-Cedar Hills depression (McIntosh et al., 1991).

In addition to ash-flow tuffs, the Bell Top Formation also contains sedimentary rocks in the Middle Sedimentary Member, positioned between AFT5 and AFT6, and in the Upper Sedimentary Member, between AFT6 and AFT7. The sedimentary rocks consist of interbedded tuffaceous sandstones and epiclastic conglomerate and sandstone. Tuffaceous sandstones represent fallout tephra reworked by shallow streams, debris flows, and perhaps by the wind. In contrast, the epiclastic conglomerates and sandstones were derived from surrounding bedrock hills and were deposited on alluvial fans and by braided streams (Mack et al., 1994a). The least abundant rock types in the Bell Top Formation are basalt and basaltic andesite. Primarily lava flows, but also including dikes and plugs, these rocks have SiO_2 contents ranging from 48.5 to 56.7 % (Clemons, 1975).

The Goodsight-Cedar Hills depression is bordered on the southeast by the Cedar Hills vent zone, a narrow (6 km), north-trending belt of intrusive and extrusive igneous rocks that are commonly included within the Bell Top Formation (Seager, 1973, 1975). The zone is primarily composed of 25 flow-banded rhyolite domes, but also contains fallout breccias, tuffaceous sandstones, and AFT3. Field relationships suggest the flow-banded rhyolite domes were emplaced after AFT3 (35.7 Ma) was erupted from the vent zone, an interpretation consistent with a recent $^{40}Ar/^{39}Ar$ single-crystal sanidine date for one of the domes of 35.4 ± 0.21 Ma (D. Ware, W. McIntosh, and N. McMillan, personal comm., 1998). The flow-banded rhyolite domes represent devolitalized magma left in the chamber after an upper gas-rich part erupted as AFT3.

The areal distribution of grain size and lithofacies, imbrication and crossbed paleocurrents, and provenance of the epiclastic conglomerates and sandstones of the Middle and Upper Sedimentary Members of the Bell Top Formation have been used by Mack et al. (1994a) to suggest that between the time of eruption of AFT5 (34.8 Ma) and AFT7 (28.6 Ma) the Goodsight-Cedar Hills depression may have been a half graben, whose footwall was the Cedar Hills vent zone and whose hanging wall was the Good Sight Mountains. This interpretation is based on: (1) a narrow band on the eastern side of the basin of westerly prograding alluvial-fan conglomerates composed almost exclusively of flow-banded rhyolite clasts, (2) a broad western belt, constituting about half of the basin, composed of easterly prograding alluvial-fan conglomerates and sandstones derived from the Palm Park/Rubio Peak and Love Ranch Formations, and (3) a narrow belt of fluvial sediment exhibiting northerly paleocurrents and composed of a mixture of clasts from both sides of the basin. This asymmetrical distribution of facies is diagnostic of a half graben (Leeder and Gawthorpe, 1987) and suggests that crustal extension in south-central New Mexico may have begun by about 35 Ma.

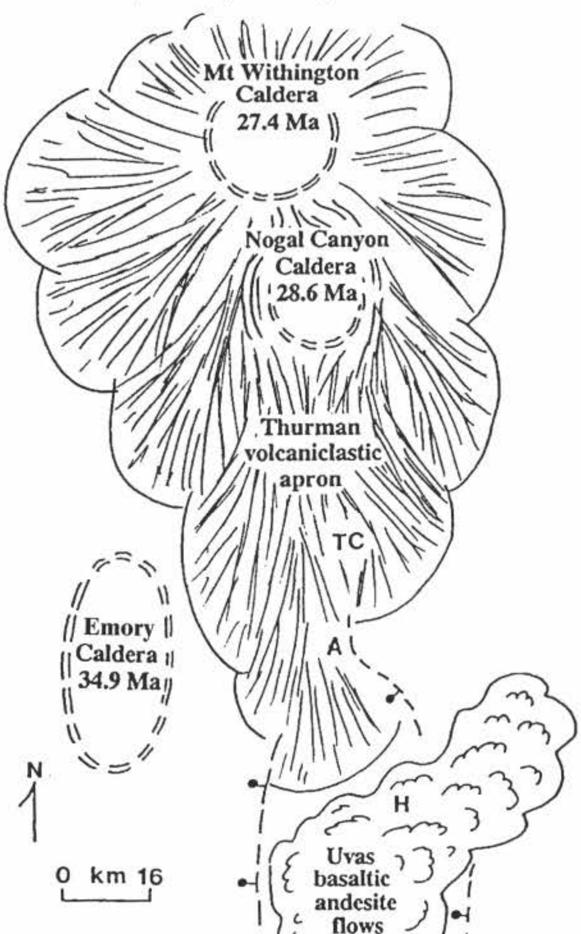

FIGURE 9. Paleogeographic reconstruction of south-central New Mexico in late Oligocene. Numbers are radioisotopic ages from McIntosh et al. (1991).

Uvas Basaltic Andesite and Thurman Formations (late Oligocene)

Deposition of the Bell Top Formation was followed by mafic volcanism of the Uvas Basaltic Andesite, named by Kottlowski (1953) for exposures in the Sierra de las Uvas (Figs. 7 and 9). The Uvas has a maximum thickness of about 250 m near the axis of the Goodsight-Cedar Hills depression/half graben, where it is composed of as many as 14 individual lava flows (Clemons and Seager, 1973). Several cinder cones within the Uvas Formation have been identified in the Sierra de las Uvas, most notably in Broad Canyon (Seager et al., 1975). K-Ar dates establish the age of the Uvas between 27 and 28 Ma and it correlates with a lava flow west of Hillsboro dated at 28.1 Ma (Seager et al., 1984). The Uvas is composed of basaltic andesite (SiO_2 = 50–60%), containing hornblende and rarely olivine phenocrysts in a matrix of plagioclase, pyroxene, and iron oxides (Clemons and Seager, 1973). Trace-element and isotopic data suggest the Uvas basaltic andesites were derived from a lithospheric mantle source (McMillan et al., 1991; McMillan, 1994). Uvas lava flows interfinger with and are overlain by the Thurman Formation in the Rincon Hills (Seager and Hawley, 1973) and the Uvas pinches out in Apache Graben, where the Thurman Formation directly overlies the Bell Top Formation.

The Thurman Formation was named for exposures in the Rincon Hills by Kelley and Silver (1952), where it is 388 m thick, but a complete section of Thurman 508 m thick is also exposed in Apache Graben in the southern Caballo Mountains. The Thurman Formation of Kelley and Silver (1952) and Seager and Hawley (1973) originally included the Uvas and Bell Top, but has subsequently been restricted to those rocks above the Uvas Basaltic Andesite in the Rincon Hills and southern Apache Graben and above the Bell Top in the northern Apache Graben, where the Uvas pinches out. The Thurman consists primarily of white to light gray, fine- to medium-grained, tuffaceous sandstones that represent fallout tephra reworked by debris flows and shallow streamflows. Much less common are red mudstones, grain-supported cobble conglomerates, and fallout tuff. A general southward dispersal of the tuffaceous sandstones is suggested by trough-crossbed paleocurrent data and by a greater abundance of mudstones and fine-grained sandstones in the Rincon Hills section compared to the section in Apache Graben (Kieling, 1993). Tuffaceous sandstones contain abundant glass shards and pumice, as well as single crystals of sanidine, quartz, plagioclase, and biotite. $^{40}Ar/^{39}Ar$ analyses of single crystals of sanidine extracted from the tuffaceous sandstones give ages of 34.9, 33.5 and 27.7 Ma, but mostly concentrate around 27.4 Ma (Boryta and McIntosh, 1994). The predominant date of 27.4 Ma suggests that most of the tephra in the Thurman Formation was derived from eruptions of the Mount Withington caldera (McIntosh et al., 1991). The radioisotopic data, coupled with the sediment dispersal data of Kieling (1993), suggest the Thurman was a volcaniclastic apron associated with the Mount Withington caldera (Fig. 9). The Thurman Formation is too poorly distributed geographically to determine if the volcanic apron was influenced by late movement on the border faults of the Goodsight-Cedar Hills half graben or by early movement on the major border faults of the Caballo Mountains block.

MIOCENE BLOCK FAULTING AND DEPOSITION OF THE HAYNER RANCH AND RINCON VALLEY FORMATIONS

The early stage of crustal extension of the Rio Grande rift culminated in Miocene time with two pulses of block faulting and concomitant sedimentation of up to 1900 m of redbeds of the Hayner Ranch and Rincon Valley Formations (Fig. 7). Despite evidence of widespread extension, there was little volcanism during this period. The Hayner Ranch Formation is devoid of volcanic rocks and the Rincon Valley Formation contains only the 9.6 Ma Selden Basalt in its upper part (Seager et al., 1984). Sedimentologic analysis of the Hayner Ranch and Rincon Valley Formations by Mack et al. (1994b) has defined Miocene paleotectonics and paleogeography, and is summarized below.

Hayner Ranch Formation

The Hayner Ranch Formation was named for exposures near San Diego Mountain, where it has a maximum thickness of 800 m (Seager et al., 1971). In addition, 500 m of sediment at San Diego Mountain that were mapped as an "unnamed transitional unit" are now included in the basal part of the Hayner Ranch Formation, bringing the total thickness to 1300 m. The Hayner Ranch Formation conformably overlies and interfingers with the Thurman Formation in Apache Graben, providing a lower limit for the age of the Hayner Ranch Formation of 27.4 Ma. At San Diego Mountain, the Hayner Ranch is conformably overlain by the Rincon Valley Formation, but near the margins of the Hayner Ranch Basin the Hayner Ranch-Rincon Valley contact is locally an angular unconformity.

The Hayner Ranch Formation was deposited in a narrow, northwest-trending northeast-tilted half graben, whose footwall was the Caballo Mountains (Fig. 10A). Asymmetry of the basin is indicat-

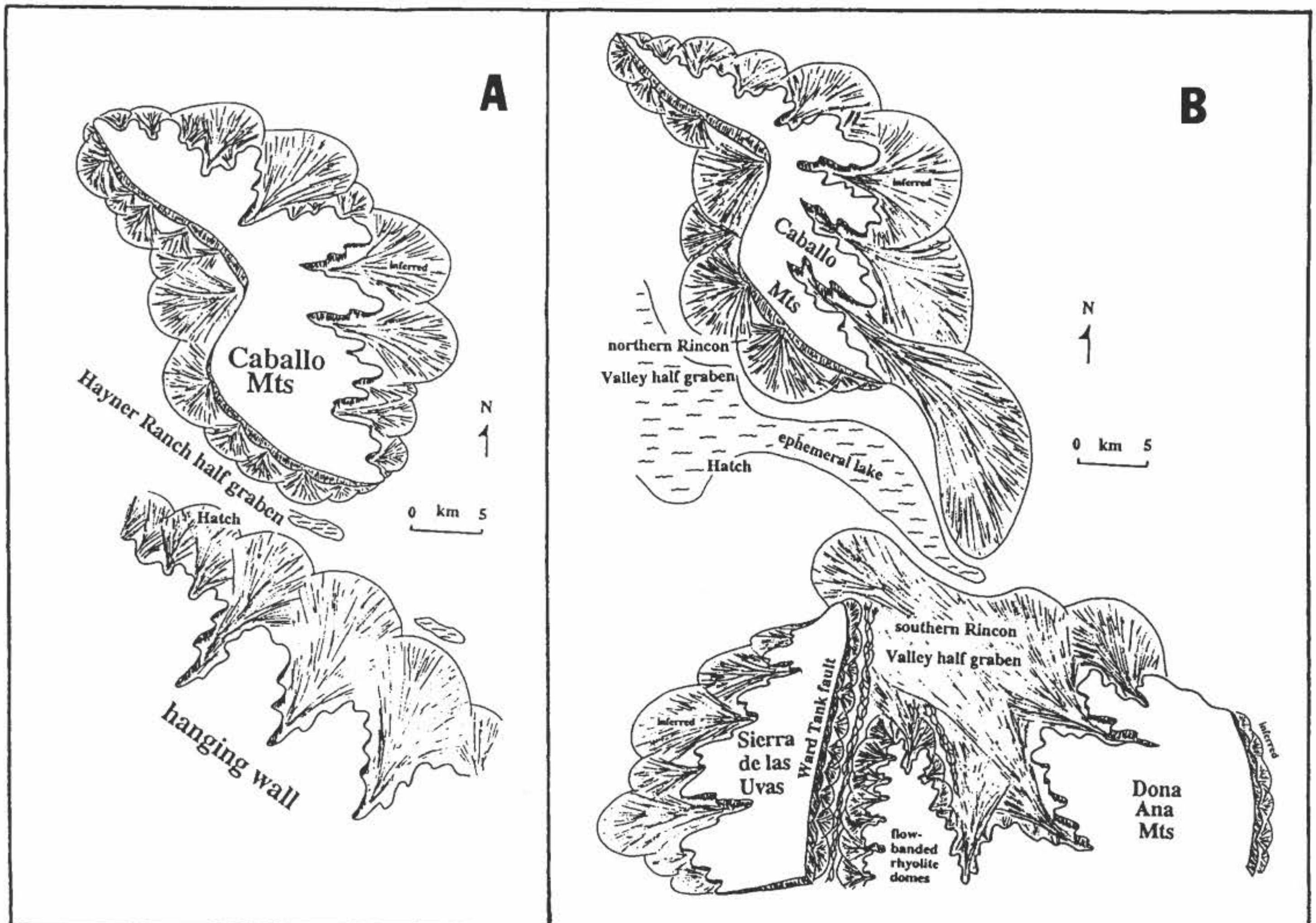

FIGURE 10. Paleogeographic reconstructions of a part of south-central New Mexico in **A,** latest Oligocene–early Miocene, corresponding to the Hayner Ranch Formation, and **B,** middle–late Miocene, corresponding to the Rincon Valley Formation, based on Mack et al. (1994b).

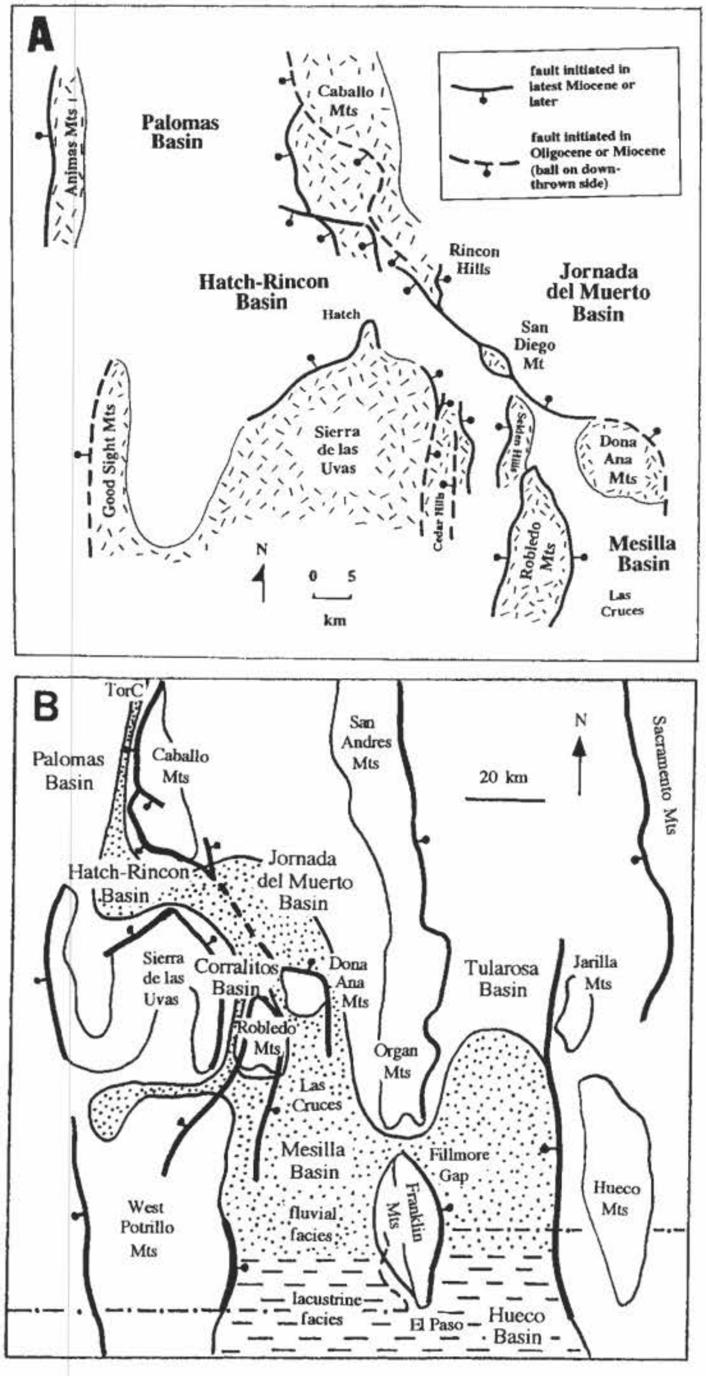

FIGURE 11. **A,** Distribution of early rift (Oligocene–Miocene) and late rift (latest Miocene or younger) normal faults in south-central New Mexico. **B,** Distribution of axial-fluvial and lacustrine lithofacies of the Palomas and Camp Rice Formations (Pliocene–early Pleistocene) in south-central New Mexico, adapted from Mack et al. (1997).

ed by a relatively narrow lithofacies belt adjacent to and derived from the footwall, a broader lithofacies belt along the southwestern margin of the basin, and the presence of lacustrine limestones within a few kilometers of the footwall block (Mack et al., 1994b). Hayner Ranch strata adjacent to the Caballo Mountains footwall display south-southwestward change from proximal to distal alluvial-fan lithofacies, as well as paleoflow directed away from the Caballo fault block. Especially diagnostic of derivation from the Caballo Mountains are proximal fan deposits directly adjacent to the border fault containing boulders in excess of 1 m in diameter. Hayner Ranch conglomerates also display an unroofing sequence commensurate with uplift and erosion of the Caballo Mountains. The lower part of the formation consists primarily of clasts of Uvas Basaltic Andesite and Bell Top AFT 5 and 6. Andesitic clasts derived from the Palm Park Formation progressively increase upsection, and well-rounded, polycyclic clasts of Precambrian granitic and metamorphic rocks and Paleozoic sedimentary rocks derived from the Love Ranch Formation appear in the upper part of the formation. In the Rincon Hills, the lower part of the Hayner Ranch Formation contains from 1 to 4 thin (<1 m) beds of lacustrine micrite limestone that contain ostracods, charophytes, and stromatolites (Kieling, 1994; Mack et al., 1994b).

The southern part of the Hayner Ranch Basin is less well constrained because of fewer outcrops. Exposed in the northern Sierra de las Uvas is a paleovalley infilled with about 30 m of boulder and cobble conglomerate of the Hayner Ranch Formation. This northeast-trending paleovalley is 200 m wide and 1 km long, and is incised into northeast-dipping lava flows of the Uvas Basaltic Andesite. Northeastward paleoflow within the paleovalley is indicated by northeastward decrease in grain size and by imbrication paleocurrent data (Mack et al., 1994b). The Hayner Ranch Formation also displays a southward onlap onto the Palm Park Formation over a distance of about 3 km in the Selden Hills (Seager et al., 1971). The paleovalley and onlap relationship, along with the absence of evidence for syn-depositional faulting, indicate that the southwestern margin of the basin was a northeastward-tilted hanging wall (Fig. 10A).

Sediment deposited on the hanging wall dip slope of the half graben is exposed in the northern Selden Hills and at San Diego Mountain. Northeastward sediment dispersal is indicated by a northeastward progression from proximal to more distal alluvial-fan and alluvial-flat lithofacies, and by northeastward-directed paleocurrents (Mack et al., 1994b). The predominance of Uvas Basaltic Andesite, flow-banded rhyolite, and Palm Park clasts and the absence of Precambrian and Paleozoic clasts of the Love Ranch Formation also suggest a southern provenance for the Hayner Ranch detritus.

Rincon Valley Formation

The type section of the Rincon Valley Formation is near San Diego Mountain, where it conformably overlies the Hayner Ranch Formation and is unconformably overlain by the Camp Rice Formation (Seager et al., 1971). The Rincon Valley Formation has a maximum thickness of about 600 m in the Rincon Hills (Seager and Hawley, 1973), and is late Miocene in age, based on the presence of the 9.6 Ma Selden Basalt lava flow in the upper part of the formation (Fig. 7; Seager et al., 1984). The age of the top of the Rincon Valley Formation is not known, but it is probably not younger than a 7.1 Ma basalt flow that erupted onto a fault block that developed after deposition of the Rincon Valley Formation (Seager et al., 1984).

Deposition of the Rincon Valley Formation took place in two half grabens (Fig. 10B; Mack et al., 1994b). Derivation of the Rincon Valley Formation from the Caballo Mountains, the footwall of the northern half graben, is based on south-southwestward-directed imbrication paleocurrents and a southwestward change from midfan to distal-fan and alluvial-flat lithofacies (Mack et al., 1994b). Basal Rincon Valley conglomerates derived from the Caballo Mountains are similar in composition to the Hayner Ranch Formation, but upper Rincon Valley conglomerates contain angular, first-cycle clasts of the Permian Abo Formation and Pennsylvanian Magdalena Group, formations that were apparently not exposed during Hayner Ranch time.

Widely exposed in the Rincon Valley Formation are red, gypsiferous mudstones deposited in an ephemeral lake. Thin (<30 cm) beds of selenite gypsum in growth position probably precipitated on the floor of the lake, whereas isolated, displacive crystals of gypsum in red mudstone probably precipitated from shallow groundwater as a result of capillary rise. Close proximity of the lacustrine lithofacies to the Caballo Mountains footwall and the relatively narrow belt of footwall-derived alluvial-fan and alluvial-flat lithofacies are evidence for asymmetry of the northern half graben.

Rincon Valley strata exposed at San Diego Mountain were also derived from the Caballo Mountains, but display a different provenance compared to other Caballo-derived sediment. In addition to clasts of Tertiary rocks, conglomerates at San Diego Mountain also contain clasts derived from the Cretaceous Dakota Formation and from the Permian Yeso Formation, which are only exposed along the eastern dip slope of the Caballo Mountains. Bedded gypsum in the Yeso Formation may have contributed to the presence of sulfate in the Rincon Valley ephemeral lake.

The footwall of the southern half graben was the Sierra de las Uvas, which was uplifted along the Ward Tank fault, whereas the hanging wall was the Doña Ana Mountains (Fig. 10B; Mack et al., 1994b). Asymmetry of the southern half graben is indicated by a narrow belt of footwall-derived alluvial-fan sediment and a broad belt of hanging wall-derived sediment. Uplift of the Sierra de las Uvas is indicated by coarse, proximal alluvial-fan conglomerates with eastward-directed paleocurrents exposed directly adjacent to the Ward Tank fault. These conglomerates consist of clasts of Uvas Basaltic Andesite and Bell Top ash-flow tuffs, one of which (AFT4) is only exposed in the Sierra de las Uvas. Uplift of the Doña Ana Mountains is indicated by a northwestward change from midfan to distal-fan lithofacies and by paleocurrents directed away from the range (Mack et al., 1994b). The Doña Ana Mountains also have distinctive rock types, which can be recognized in conglomerates deposited on the hanging wall dip slope of the southern half graben. Among these distinctive rocks are the Doña Ana Rhyolite, red siltstones of the Abo Formation, and limestones of the Hueco and Panther Seep Formations, the latter of which were locally contact metamorphosed to green marble. Provenance and paleocurrent data also suggest that flow-banded rhyolite domes were important source rocks for sediment of the southern half graben and that a drainage divide trending north across the domes may have resulted in separate southwestern and south-central source terranes (Fig. 10B; Mack et al., 1994b). Conglomerates derived from the southwestern source display northwestward-directed paleocurrents and consist primarily of flow-banded rhyolite clasts, while conglomerates from the south-central source contain Palm Park and AFT3 clasts in addition to flow-banded rhyolite. Paleoflow along the hanging wall dip slope of the southern half graben was directed northwestward, rather than westward as would be predicted along a hanging wall complementary to a north-trending border fault. This anomalous paleoflow may have resulted from northwestward tilt of a relay ramp positioned between the overlapping tips of the Jornada and Ward Tank faults. This relay ramp may also have influenced the position of the ephemeral lake, as well as allowed sediment from the southern half graben to be transported into the northern half graben.

RECENT STAGE OF EXTENSION IN THE RIO GRANDE RIFT

Deposition of the Camp Rice and Palomas Formations

The second major stage of crustal extension in the southern Rio Grande rift began in latest Miocene time (after deposition of the Rincon Valley Formation) and continues to the present (Seager, 1975; Chapin and Seager, 1975; Seager et al., 1984). This stage is responsible for the modern distribution of basins and uplifts. Many of the border faults of the early rift blocks continued to be active in the later stage, but with lower magnitudes of displacement. In general, most of the strain during the second stage was concentrated along new faults that were located within early rift basins, creating new mountain ranges such as the Rincon Hills, Selden Hills, Cedar Hills, San Diego Mountain, and the Robledo Mountains (Fig. 11A). Some faults can be shown to have experienced post-early Pleistocene movement (e.g., Jornada, Caballo, Jornada Draw faults), and Holocene offset has been demonstrated on the Caballo fault and the Organ fault (Gile, 1986, 1994; Machette, 1987; Foley et al., 1988; Seager and Mack, 1995).

The sedimentary record of the late stage of extension is called the Palomas Formation in the Palomas Basin and the Camp Rice Formation in basins to the south (Fig. 7; Strain, 1966; Hawley et al., 1969; Hawley, 1981; Lozinsky and Hawley, 1986a, b). These coeval formations have a maximum exposed thickness of about 150 m and consist of conglomerate and sandstone deposited on alluvial fans and alluvial flats, and crossbedded and horizontally laminated pebbly sand/sandstone and mudstone deposited by the ancestral Rio Grande (Seager et al., 1982, 1987). During its early history, the ancestral Rio Grande emptied into Lake Cabeza de Vaca in the Hueco Basin of west Texas and Chihuahua, Mexico (Strain, 1966). By approximately 2.2 Ma, the Hueco Basin was breached, allowing southward flow of the ancestral Rio Grande (Gustavson, 1991).

The Camp Rice and Palomas Formations are Pliocene to early Pleistocene in age, based on (1) a Blancan and lower Irvingtonian vertebrate fauna (Hawley et al., 1969; Tedford, 1981; Lucas and Oakes, 1986; Repenning and May, 1986; Vanderhill, 1986), (2) interbedded basalt lava flows dated by the K-Ar method (Bachman and Mehnert, 1978; Seager et al., 1984), (3) interbedded pumice-clast conglomerates derived from the Jemez volcanic field and dated by single-crystal sanidine $^{40}Ar/^{39}Ar$ analysis (Mack et al., 1996), and by (4) reversal magnetostratigraphy (Vanderhill, 1986; Mack et al., 1993, this guidebook; Leeder et al., 1996). The oldest exposed Palomas Formation may be older than the Sidufjall subchron of the Gilbert chron (>4.89 Ma; Berggren et al., 1995; Leeder et al., 1996) and the age of the top of the Camp Rice and Palomas Formations is at or very near the Matuyama-Brunhes chron boundary, 0.78 Ma (Mack et al., 1993, 1996, this guidebook; Berggren et al., 1995).

Axial-fluvial sediment is widely distributed in southern New Mexico having occupied six continguous basins between Truth or Consequences and Las Cruces (Fig. 11B). The spillover of the ancestral Rio Grande among these basins was controlled by movement on basin-bounding faults or by deposition to the level of a topographic gap and avulsion into the adjacent basin (Ruhe, 1962; Hawley et al., 1969; Mack et al., 1994c; Mack et al., 1997). Within an individual basin, the location, width, and relative abundance of rock types in the axial-fluvial lithofacies of the Palomas and Camp Rice Formation is controlled primarily by basin symmetry (Fig. 11B; Mack and Seager, 1990; Mack and James, 1993). The mapped belt of axial-fluvial strata is narrow (<5 km) in half grabens and positioned within a kilometer or less of the footwall block, whereas in full grabens the axial-fluvial sediment extends across almost the entire basin. Moreover, in half grabens, the axial-fluvial sediment is composed almost exclusively of multistory channel sands/sandstones, while full grabens contain interbedded channel sand/sandstone and floodplain very fine sand/silt and mudstone, some of which displays stage II calcic paleosols (Mack and James, 1992, 1993). Oxygen and carbon isotopes of the calcic paleosols suggest that throughout the time of deposition of the Camp Rice and Palomas Formations the paleoclimate became hotter, drier, and/or with more summer-seasonal precipitation (Mack et al., 1994d).

Basalt volcanism

The second stage of crustal extension in the southern Rio Grande rift was also characterized by basalt volcanism. More than 100 cinder cones, small shield volcanoes, and maars are present in and adjacent to the West Potrillo Mountains (Hoffer, 1971, 1975, 1976; Anthony et al., 1992). Recent ^{3}He surface exposure dating of basalts from the Potrillo volcanic field has yielded ages ranging from 80 to 17 ka (Anthony and Poths, 1992). The West Potrillo basalt field is located in the West Potrillo transfer zone, which exists in the overlap between the west-dipping Camel Mountain fault and the west-dipping West Robledo fault (Mack and Seager, 1995). Another basalt lava field, consisting of seven cinder cones and associated lava flows, is located in the Cutter Sag between the Fra Cristobal and Caballo Mountains. One of the flows in this field has been dated at 2.1 Ma (Bachman and Mehnert, 1978). Like the West Potrillo basalt field, the Cutter Sag field occupies a transfer zone between overlapping normal faults that dip in the same direction (Mack and Seager, 1995).

In addition to the aforementioned basalt fields, single basalt cinder cones or lava flows of Pliocene or younger age are scattered throughout the southern Rio Grande rift (Seager et al.,1982, 1984, 1987). Many of these are associated with range-boundary faults. Also present on the west mesa, southwest of Las Cruces, are three phreatomagmatic craters, Kilbourne Hole, Hunt's Hole, and Potrillo maar, the latter astride the Mexico-USA border (Seager, 1987). Kilbourne Hole and Potrillo maar are well known for their assemblage of mantle and lower crustal xenoliths (Padovani and Carter, 1977; Reid and Woods, 1978).

Trace-element and isotope geochemistry of basalts younger than 10 Ma (includes the Selden Basalt) suggests derivation from an asthenospheric mantle source (McMillan, 1994, this guidebook). This interpretation is consistent with geophysical evidence of a wedge of asthenosphere beneath the southern Rio Grande rift (Sinno et al., 1986; Daggett et al., 1986).

Post-Early Pleistocene inset alluvium

Following deposition of the Camp Rice and Palomas Formations, sedimentary basins in the Rio Grande valley began a sequence of downcutting and partial backfilling by the ancestral Rio Grande and its tributaries. This process has placed the modern Rio Grande about 100 m below its position at the end of Camp Rice and Palomas deposition 0.78 Ma, and has created as many as five mappable geomorphic surfaces, which represent the constructional tops of the backfilled sediment (Gile et al., 1981). Variability in soil development on geomorphic surfaces of different age led to the widely used four-stage classification of calcic soil morphology of Gile et al. (1966). The driving force responsible for downcutting and backfilling by the ancestral Rio Grande and its tributaries was probably climatic variations related to Northern Hemisphere glaciations (Gile et al., 1981).

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