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GEOLOGY OF THE MORENO VALLEY, COLFAX COUNTY, NEW MEXICO

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Abstract—This paper presents a summary of the geology of the Moreno Valley based on published reports, geologic maps and unpublished theses. Rocks in the Moreno Valley range in age from Precambrian to Quaternary with thick sections of upper Paleozoic (Pennsylvanian-Permian) and lesser thicknesses of Mesozoic and Tertiary strata. All units have been disrupted by folding and faulting representing several periods of deformation ranging in age from Precambrian to Tertiary. Beginning in the latest Cretaceous, initial broad folding and uplift were followed during the early Tertiary by tighter folding, and thrust and reverse faulting. These structures were cut by later high-angle normal faults, some related to development of the present Sangre de Cristo Mountains and the adjacent Rio Grande rift. Minimum shortening in an east-west direction of as much as 2.7 km is estimated for the folding and thrust faulting. Stratigraphic throws on transverse normal faults reach nearly 3900 m. Displacements on valley-parallel normal faults are less than 100 m. Mid- to late Tertiary intrusive and extrusive rocks have had a major influence on the development of the present-day valley. The present Moreno Valley developed from blockage of the ancestral Coyote Creek drainage by early Pliocene (5 Ma) lava flows and the integration of Moreno and Cieneguilla Creek drainages to the present Cimarron River.

INTRODUCTION

The Moreno Valley is located in the Sangre de Cristo Mountains in southwest Colfax County. It covers 344 km^2 and is 43 km long and 8 km wide. The valley extends from Costilla Pass and Tetilla Peak on the north to Black Lake divide on the south. It is part of a north-trending topographic and structural depression that reaches southward to near Mora, New Mexico (Fig. 1).

This paper, which summarizes the current knowledge of the geology of the Moreno Valley, is based on work by Stevenson (1881), Ray and Smith (1941), Smith and Ray (1943), Young (1945), Brill (1952), Wanek and Read (1956), Baltz and Read (1960), Robinson et al. (1964), Petersen (1969), Clark (1966), Clark and Read (1972), Goodknight (1973), Casey (1980a, b), Condie (1980), Smith and Colpitts (1980), Reed et al. (1983), O'Neill and Mehnert (1988) and O'Neill (1988).

Altitudes in the valley range from 2489 m at Eagle Nest Lake to 3792 m on Baldy Mountain in the Cimarron Range. Climate in the valley is temperate-dry with average annual precipitation between 37 cm to 44 cm (Gabin and Lesperance, 1977).

GEOLOGIC SETTING

Rocks exposed in and around the Moreno Valley range in age from Precambrian to Quaternary. These units have been disrupted by folding and complex transverse, normal and thrust faulting; minor strike-slip faulting has occurred in the Eagle Nest area. Figure 2 shows the general distribution of rock units and structural features of the valley. It was compiled from Robinson et al. (1964), Petersen (1969), Clark and Read (1972), Goodknight (1973), Condie (1980), Smith and Colpitts (1980) and Casey (1980a).

Stratigraphy

Precambrian rocks

Rocks of Precambrian age crop out in the central and northern parts of the Moreno Valley, north of Saladon Creek. East of the valley, they form the core of the Cimarron Range, but are partly covered by Paleozoic and Cenozoic rocks in the Taos Mountains to the west. South of Saladon Creek, Precambrian rocks are buried beneath more than 3100 m of Paleozoic and Cenozoic rocks. North of American Creek, Precambrian rocks form a north-plunging ramp in the valley floor and are buried by more than 4700 m of Phanerozoic strata north of Eagle Nest. Precambrian rocks in the Moreno Valley can be divided into three general groups: granitic, gneissic and metasedimentary.

Granitic rocks

Granitic rocks consist of quartz monzonite and granite (Condie, 1980). They are pink to white and coarse-grained with variable amounts of orthoclase, microcline, sodic plagioclase, quartz and accessory biotite and magnetite, plus minor secondary quartz, chlorite, muscovite and epidote aggregates (Condie, 1980; Lipman and Reed, 1989). Granitic rocks are distributed across the northern and western parts of the valley and are weakly to moderately foliated (Lipman and Reed, 1989).

Gneissic rocks

Gneissic rocks comprise paragneiss and quartzofeldspathic gneiss. Paragneiss is confined to the northern part of the valley near the head-

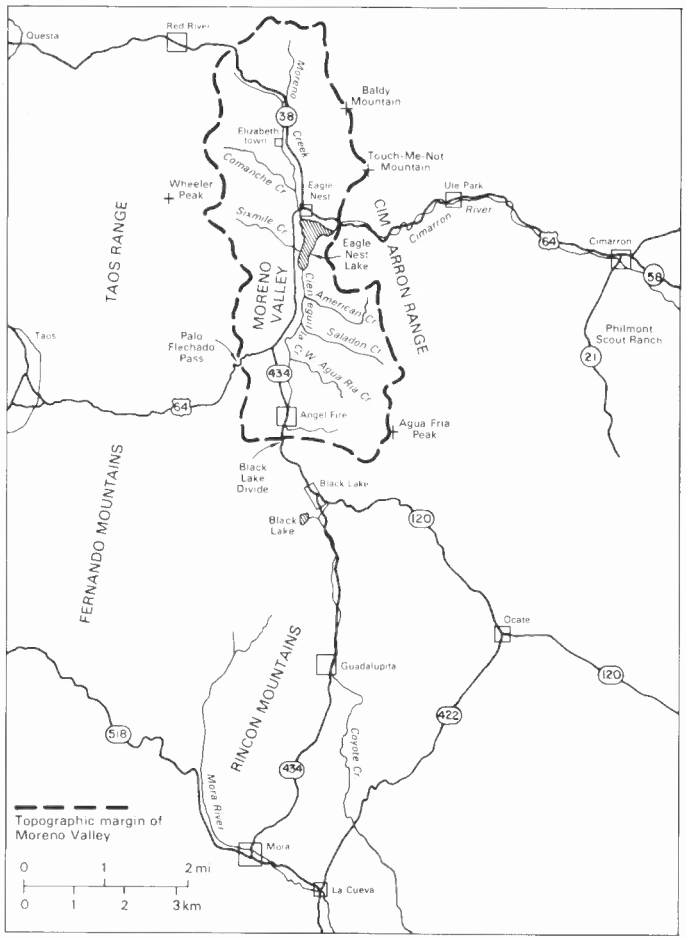
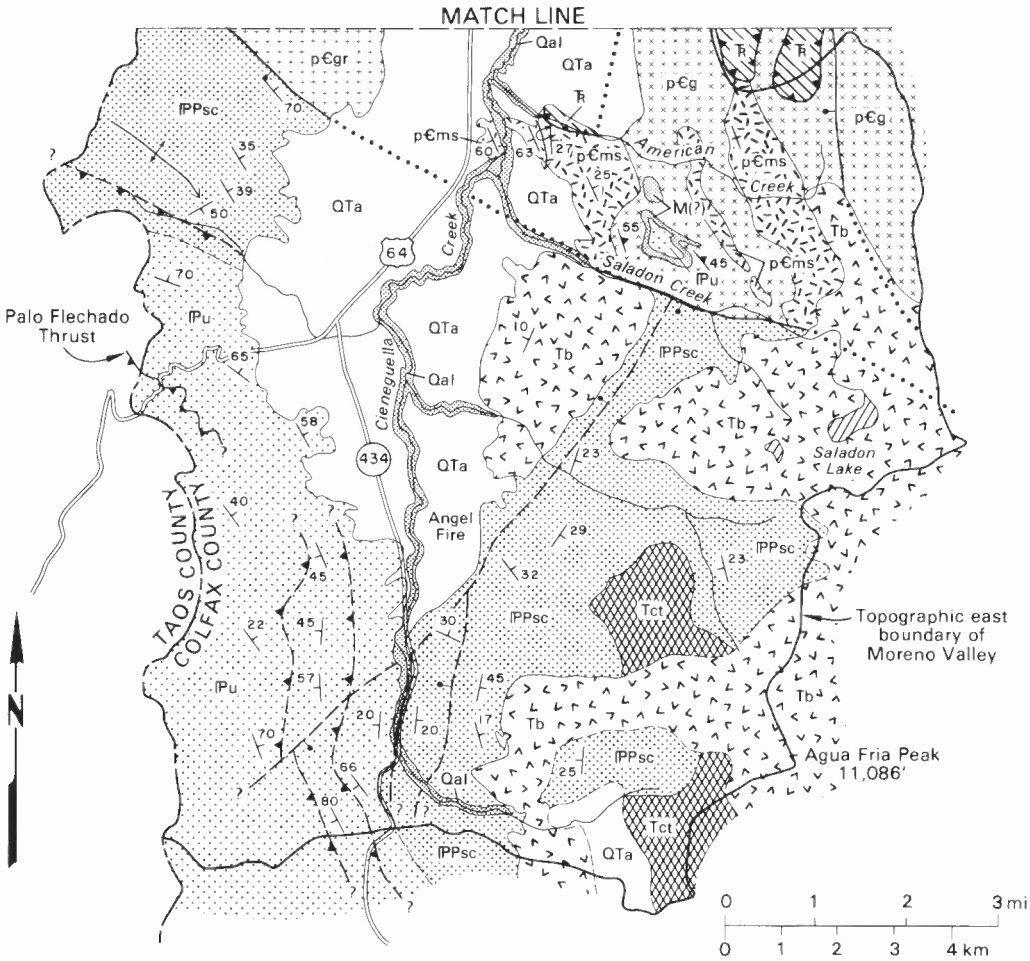


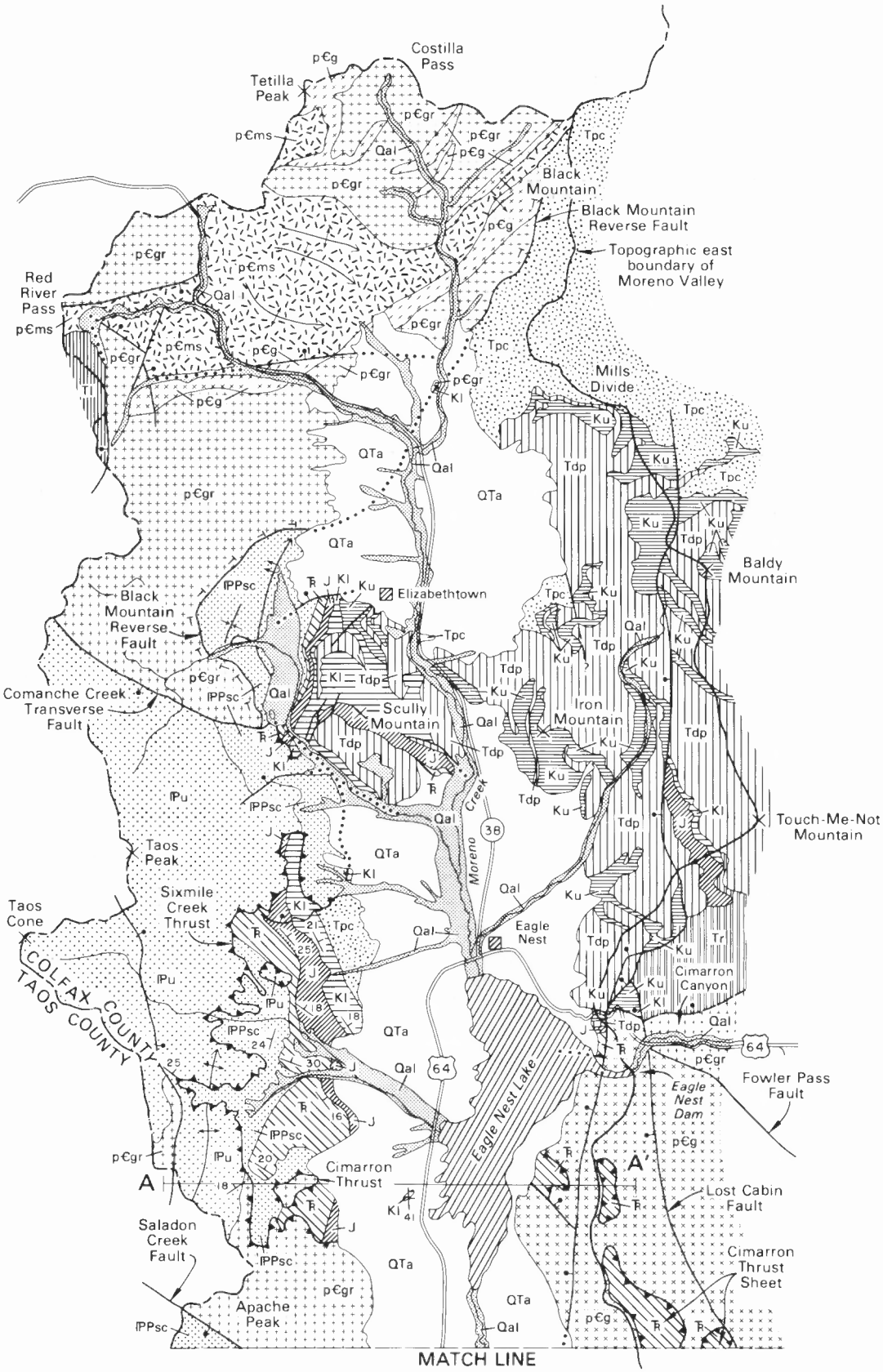
FIGURE 1. Location of the Moreno Valley in the southern Sangre de Cristo Mountains, New Mexico.



EXPLANATION

- **Holocene**—Recent stream alluvium
- **Holocene, Pliocene, and Pleistocene**—Valley fill and terrace deposits. Includes Tertiary Eagle Nest Formation
- **Pliocene**—Basalt flows of Ocate volcanic field
- **Miocene?**—Ashflow crystal tuff of Cieneguilla Mountain
- **Miocene and Oligocene**—Rhyolite sills and dikes
- **Miocene and Oligocene**—Quartz diorite porphyry sills and dikes
- **Miocene and Oligocene**—Latite flows
- **Paleocene**—Poison Canyon Formation
- **Upper and Lower Cretaceous**—Pierre, Niobrara and Carlisle Formations. Includes Trinidad, Vermejo and lower part of Raton Formations where they are present
- **Upper and Lower Cretaceous**—Greenhorn and Graneros Formations and Dakota Group
- **Upper Jurassic and Lowest Cretaceous**—Undivided Morrison, Bell Ranch and Entrada Formations
- **Upper Triassic**—Undivided Johnson Gap, Trujillo, Garita Creek and Santa Rosa Formations
- **Pennsylvanian and Lower Permian**—Sangre de Cristo Formation
- **Pennsylvanian**—Undivided Alamitos, Porvenir and Sandia Formations
- **Mississippian**—Mississippian(?) rocks
- **Proterozoic**—Granitic rocks undifferentiated
- **Proterozoic**—Metasedimentary rocks undifferentiated
- **Proterozoic**—Gneissic rocks undifferentiated
- Strike and dip of strata
- Strike and dip of foliations
- Contacts, dashed where inferred
- Normal fault; ball and bar on downthrown side, dashed where approximate, dotted where concealed
- Low angle (<45°) thrust fault; teeth toward upper plate, dashed where approximate, dotted where concealed
- High angle (>45°) reverse fault; T on upper plate, dashed where approximate, dotted where concealed
- Anticline, showing plunge
- Syncline, showing plunge
- Overturned anticline, showing plunge

FIGURE 2. Geologic map of the Moreno Valley compiled and modified from Clark and Read (1972), Condie (1980), Goodknight (1976), Petersen (1969) and Smith and Colpitts (1980).


Costilla Pass
Tetilla Peak
Black Mountain
Black Mountain Reverse Fault
Topographic east boundary of Moreno Valley
Red River Pass
Mills Divide
Baldy Mountain
Elizabethtown
Black Mountain Reverse Fault
Comanche Creek Transverse Fault
Scully Mountain
Iron Mountain
Touch-Me-Not Mountain
Taos Peak
Sixmile Creek Thrust
Taos Cone
COLFAX COUNTY
TAOS COUNTY
Moreno Creek
Eagle Nest
Cimarron Canyon
Eagle Nest Lake
Eagle Nest Dam
Fowler Pass Fault
Cimarron Thrust
Lost Cabin Fault
Saladon Creek Fault
Cimarron Thrust Sheet
Apache Peak
MATCH LINE

waters of Moreno Creek (Condie, 1980). Clark and Read (1972) mapped this unit as hematite-muscovite granulite. It is light colored with aggregates of hematite and muscovite in a matrix of medium- to finely crystalline quartz, microcline and sodic plagioclase with accessory sillimanite, biotite and zircon (Clark and Read, 1972; Condie, 1980).

Quartzofeldspathic gneiss is used here as a generic term for gneissic rocks whose protolith has not been established. Rocks in this group crop out along the eastern side of the valley from the head of Cimarron Canyon south to the headwaters of Saladon Creek. These rocks are mapped as granite by Goodknight (1976) and granite gneiss by Smith and Colpitts (1980). They form the core of the Cimarron Range (Robinson et al., 1964; Smith and Colpitts, 1980; Grambling and Dallmeyer, 1990).

Quartzofeldspathic gneiss is pink to light red and consists of medium- to finely crystalline quartz and orthoclase with accessory biotite and plagioclase (Clark and Read, 1972; Condie, 1980; Smith and Colpitts, 1980). The biotite forms segregations that impart a weak foliation to these rocks in the area of Saladon and American Creeks (Smith and Colpitts, 1980). In the Cimarron Mountains, the foliation and mineral segregations are well developed (Robinson et al., 1964).

Metasedimentary rocks

Metasedimentary rocks comprise quartzite with minor muscovite schist. The quartzites are light gray to white with accessory muscovite, magnetite and hornblende. The muscovite and magnetite form segregations along probable relict bedding planes; some quartzite beds display faint trough crossbedding. Clark and Read (1972) also reported sillimanite aggregates between quartz grains in quartzites north of Elizabethtown.

Muscovite schists form thin interbeds and interlaminations with the quartzite at American Creek. The muscovite schists are medium gray and composed of muscovite with accessory ferrotremolite, sillimanite and chlorite; neither garnet nor staurolite were observed (Smith and Colpitts, 1980).

Smith and Colpitts (1980) suspected that the quartzofeldspathic gneiss between American and Saladon Creeks had intruded the quartzite-muscovite schist sequence to the west. However, farther north and east, Grambling and Dallmeyer (1990) concluded that the quartzofeldspathic gneiss represents granulite facies metamorphic rocks.

Paleozoic rocks

Paleozoic rocks consist of unnamed Mississippian(?) rocks in the southeastern Moreno Valley; Pennsylvanian Sandia Formation (Flechado Formation of Miller et al., 1963) and Madera Group (Porvenir and Alamitos Formations of Baltz and Myers, 1984); and the Permo-Pennsylvanian Sangre de Cristo Formation. These units crop out in the southern and west-central part of the valley.

Mississippian rocks(?)

Possible Mississippian-age rocks crop out at the base of a small outlier of strata resting on Precambrian quartzofeldspathic gneiss north of Saladon Creek (Fig. 2). These rocks consist of a basal arkosic conglomerate, silicified limestone and chert. The basal arkosic conglomerate is brown and consists of pebbles and granules of quartzofeldspathic gneiss and quartzite in a matrix of poorly sorted, coarse-grained feldspathic arenite. Overlying this conglomerate is gray, red and brown bedded chert and gray to brown silicified limestone. This sequence is assigned a Mississippian age on the basis of lithology; it resembles outcrops of Espiritu Santo Formation described by Clark and Read (1972). The total thickness of Mississippian(?) rocks is 7.8 m (Smith and Colpitts, 1980).

Pennsylvanian rocks

Pennsylvanian rocks are divided into three formations: the basal Sandia Formation (Flechado Formation of Miller et al., 1963), the Porvenir Formation (Baltz and Myers, 1984) and the Alamitos Formation (restricted) (Baltz and Myers, 1984). These units represent deposition of sediments along the western flank of the Sierra Grande arch and Cimarron arch during Morrowan through Desmoinesian time. These formations are divided on the basis of biostratigraphic correlations by Baltz and Myers (1984) and are shown as undivided Pennsylvanian on Figure 2.

The Mississippian(?) rocks north of Saladon Creek are overlain by a light-gray to light-brown, medium-grained, well sorted, trough crossbedded quartzarenite that contains plant impressions. Smith and Colpitts (1980) assigned the quartzarenite to the Sandia Formation on the basis of lithology and plant fossils. The Sandia Formation consists of interbedded sandstones, conglomerates, siltstones, mudstones and limestones. The sandstones are very light- to dark-brown, medium- to very coarse-grained, moderately to poorly sorted, conglomeratic feldspathic, arkosic and quartzarenites. They are micaceous, medium to thin bedded and friable to hard, depending on the amount of calcite and clay present. Beds are lenticular and display trough crossbeds with tangential foresets. The conglomerates are light gray with medium to fine pebbles of quartzite in a very coarse-grained sandstone matrix. Lenses of conglomerate occur as channel-scour lag gravels associated with coarse sandstone channels. The siltstones are grayish brown to brownish black, slightly sandy to sandy, micaceous, locally fossiliferous and locally contorted by syndepositional, soft-sediment deformation. They are calcareous near limestone beds. The mudstones are brownish black, fissile to thickly laminated, fossiliferous and calcareous. They often display soft-sediment deformation and slump structures. This lithology is not common in the Moreno Valley but is well developed near Taos. The limestones are brownish gray to light gray and consist of sandy skeletal packstones (Casey, 1980a, b). They grade into coarse-grained conglomeratic arkosic arenites along depositional strike. Limestones are more common south and west of the Moreno Valley (Clark and Read, 1972; Casey, 1980a), but are poorly represented in this area. Casey (1980a) concluded that the Sandia Formation represents clastic and carbonate deposition along the western margin of the Sierra Grande arch in shallow marine, muddy strandplain, alluvial fan, braided stream and fan-delta environments in Morrowan and Atokan times.

Baltz and Myers (1984) proposed the name Porvenir Formation to replace the lower member of the Madera Formation (lower part of the Alamitos Formation proposed by Miller et al., 1963). They recognized a mappable lithologic unit within the Alamitos Formation of Miller et al. (1963) in roadcuts east of Mora. Based on fusulinid zonations, the Porvenir Formation represents the Desmoinesian part of the Pennsylvanian section. At Mora, it consists of interbedded limestone (oolitic, sandy and fossiliferous), gray shale, gray, red and purplish siltstones, and quartzose and arkosic sandstone. In the Moreno Valley, a similar, clastic-dominated sequence is present at Palo Flechado Pass and near Taos Peak (Clark and Read, 1972; Casey, 1980a, b; Smith and Colpitts, 1980). We feel that it is a mappable unit in the Moreno Valley on the basis of Baltz and Myers' (1984) descriptions and herein extend the name Porvenir Formation into this area.

The Porvenir Formation in the Moreno Valley and areas immediately to the west consists of interbedded sandstones, conglomerates, siltstones, mudstones and limestone. Rock types in this formation are identical to those in the underlying Sandia Formation except for the siltstones and limestones. Siltstones in the Porvenir Formation are gray, grayish brown, brownish-black, gray-green and reddish-brown, micaceous, calcareous and sandy. Siltstones and sandstones become redder up-section and grade into the overlying Alamitos Formation. The transition is so subtle that the contact is difficult to establish. Fusulinid zonations indicate that reddish siltstones and fine sandstones exposed at the foot of Palo Flechado Pass are early Desmoinesian in age (Young, 1945). The limestones are gray to light brownish-gray and range from carbonate phylloid algal wackestones and packstones to oolite grainstones. These carbonate beds are locally fossiliferous with scattered brachiopods, crinoid ossicles, bryozoans, rugose corals and rare fusulinids (Casey, 1980a, b; Colpitts and Smith, 1990). They are laterally continuous for about 5 km but cannot be correlated with carbonate beds described by Clark and Read (1972) in other parts of the valley because of structural complexity. Casey (1980a, b) states that this part of the sequence represents deposition along the eastern shelf of the Taos Trough

(Rowe-Mora Basin of other investigators) in shallow terrigenous shelf, carbonate bank, delta and braided stream environments.

The Alamitos Formation was named by Miller et al. (1963) for rocks of middle Desmoinesian to Virgilian age in the southern and southeastern Sangre de Cristo Mountains. Baltz and Myers (1984) restricted the name "Alamitos Formation" to rocks of uppermost Desmoinesian, Missourian, Virgilian and, locally, lowest Wolfcampian ages east of Mora. This alters the definition of the range of the formation as originally proposed by Miller et al. (1963). The Alamitos Formation along the Mora River consists of interbedded gray and greenish-gray silty shales, arkosic coarse sandstone and conglomerate, thin-bedded nodular limestone, thick-bedded limestone and sandy, arkosic limestones. In the Moreno Valley, the Alamitos Formation as defined by Baltz and Myers (1984) is difficult to separate from the overlying Permo-Pennsylvanian Sangre de Cristo Formation. It seems to represent a transition from the mixed marine and marginal marine strata of the Porvenir Formation to the continental fluvial and overbank deposits of the overlying Sangre de Cristo Formation. The contact with the underlying Provenir Formation is difficult to locate but can be placed approximately at the appearance of the first red siltstones and sandstones.

In the Moreno Valley, the Alamitos Formation consists of interbedded sandstones, conglomerates, siltstones and occasional rare marine limestones. The sandstones are brown, reddish brown and yellowish gray, very fine- to very coarse-grained, moderately to poorly sorted and thin to thick bedded with trough crossbeds and planar beds. Grains of quartz, orthoclase (minor) and some microcline and muscovite cemented by calcite compose the sands. The layers are laterally discontinuous over 1 or 2 km. The conglomerates occur as localized lag deposits associated with sandstone channels and consist of pebbles of quartzite and granitic rocks derived from the neighboring Sierra Grande arch to the east. The siltstones are gray, grayish green and red brown, fissile to very thin bedded, sandy, micaceous and calcareous. They increase up section toward the contact with the overlying Sangre de Cristo Formation. The limestones consist of brownish gray to light gray, carbonate mudstone to wackestone with local concentrations of crinoid ossicles and brachiopod shell fragments. Soil cover makes their lateral continuity difficult to assess.

The thickness of these units is not clear. Both Clark and Read (1972) and Roberts et al. (1976) estimated a minimum thickness of not less than 1370 m. Since this includes the lower part of the Sangre de Cristo Formation, a thickness of 914 m may be more reasonable. Casey (1980a) estimated a thickness of 2000 m for the entire Pennsylvanian section (including the lower part of the Sangre de Cristo Formation). Smith and Colpitts (1980) estimated that the Sandia (Flechado) and combined Porvenir and Alamitos Formations (mapped together as Alamitos) were about 2600 m thick. These figures may be excessive for the Moreno Valley but are reasonable for the region west of the valley where the Taos Trough was deepest during Pennsylvanian time. Structural complexity and telescoping of the section by thrust and reverse faults preclude more accurate thickness estimates in this area.

Permo-Pennsylvanian Sangre de Cristo Formation

The thick sequence of red mudstone, siltstone, sandstone and conglomerate assigned to the Sangre de Cristo Formation was deposited during middle Pennsylvanian to early Permian time. These strata are conformable(?) and gradational(?) with underlying Pennsylvanian (Desmoinesian) strata of the Madera Group in the Taos mountains. They are nonconformable with underlying Precambrian quartzofeldspathic gneiss in the Cimarron Range-Cimarron Canyon-Philmont area (Robinson et al., 1964), suggesting that the Cimarron and Sierra Grande arches were subaerially exposed east and northeast of the valley during deposition of the Sangre de Cristo Formation.

The mudstones are greenish gray to reddish brown and magenta, slightly micaceous to micaceous, slightly calcareous to non-calcareous and fissile to thin bedded with occasional thin, nodular, calcareous zones. They grade laterally into siltstone and sandstone and represent overbank muds with paleosol horizons (nodular calcareous zones) from a meandering river system (Brown, 1984).

The siltstones are gray and greenish gray to red brown and maroon, sandy to slightly sandy, slightly micaceous to micaceous and slightly calcareous. They are thin-bedded to fissile and contain brownish-gray to blood-red lenticular, thin- to medium-bedded nodular carbonate mudstones. These represent overbank silts, distal crevasse splay and lacustrine deposits (Brown, 1984) and are restricted to the lower third of the formation.

The sandstones are light gray, magenta and brown, very coarse- to fine-grained, poorly to well-sorted, arkosic to subarkosic, slightly micaceous, slightly calcareous, very thin- to thick-bedded and locally conglomeratic. They are broadly lenticular to tabular and dispaly trough-to-tabular crossbedding. These rocks represent various fluvial environments including small meandering river channels, crevasse-splay deposits, distal alluvial fan (at the base of the formation) and braided-stream deposits (Brown, 1984).

The conglomerates are light gray to brownish gray, sandy and contain granules and fine to coarse pebbles of Precambrian quartzite, schist, quartzofeldspathic gneiss, Pennsylvanian limestone and sandstone. Beds are 12 m wide and 6 m thick and lenticular. They form prominent isolated outcrops on steep hillsides. The conglomerates were formed as channel deposits in meandering streams (Brown, 1984). The thickness of the Sangre de Cristo Formation ranges from 1105 m at Coyote Creek, near Guadalupita to over 1615 m in the northern part of the Moreno Valley (Clark and Read, 1972).

Triassic rocks

Above the Sangre de Cristo Formation is a sequence of interbedded lithic and quartzose sandstone, siltstone, mudstone, limestone- and siltstone-pebble conglomerate that was deposited during late Triassic time. Clark and Read (1972) describe these strata as belonging to the Dockum Group. Recent work by Lucas et al. (1990) demonstrates that these rocks correlate with similar strata in northeastern New Mexico and southern Colorado. Based on these correlations, they recognize the Santa Rosa Formation overlain by the Garita Creek, Trujillo and Johnson Gap Formations; the Moenkopi Formation (Middle Triassic) is not present in the Moreno Valley. These units replace "Dockum Group," a term abandoned by Lucas et al. (1990). They are shown as Triassic undivided on Figure 2. Descriptions given below are based on Clark and Read (1972), Smith and Colpitts (1980) and Lucas et al. (1990).

Santa Rosa Formation

The Santa Rosa Formation is pale olive to dusky yellow and consists of very fine-grained, subrounded, well sorted, trough crossbedded quartzarenite and poorly sorted, light olive gray limestone-pebble conglomerate. The Santa Rosa Formation is 9.1 m thick along the western flank of Scully Mountain near Elizabethtown (Clark and Read, 1972; Lucas, oral commun., 1990). It lies disconformably between the underlying Sangre de Cristo Formation and overlying Garita Creek Formation (Lucas et al., 1990).

Garita Creek Formation

The Garita Creek Formation is grayish red, purple and reddish brown, micaceous and bentonitic mudstone. It is poorly exposed and forms strike valleys east of the map area. It is 9.8 m thick at Scully Mountain (Clark and Read, 1972; Lucas, oral commun., 1990). It is disconformable with the overlying Trujillo Formation.

Trujillo Formation

The Trujillo Formation is light olive gray and grayish red and consists of fine-grained, subrounded, well-sorted, micaceous, calcareous, trough crossbedded quartzarenite interbedded with yellowish gray limestone- and siltstone-pebble conglomerates. It forms prominent cliffs and is 31.7 m thick at Scully Mountain (Clark and Read, 1972; Lucas, oral commun., 1990). It is disconformably overlain by the Johnson Gap Formation.

Johnson Gap Formation

The Johnson Gap Formation consists of interbedded mudstone, sandstone and conglomerate (Clark and Read, 1972; Lucas et al., 1990). The mudstones are grayish red with pale olive reduction spots, calcareous, and have scattered mica flakes. They are often poorly exposed. The sandstones are grayish purple, fine-grained, subrounded, well-sorted, calcareous, ripple-cross-laminated litharenites. They grade locally into grayish red, calcareous, sandy siltstone and silty sandstone. The conglomerates are grayish yellow green and consist of clay and limestone pebbles in a fine- to medium-grained sandstone matrix (Clark and Read, 1972; Smith and Colpitts, 1980; Lucas et al., 1990). The Johnson Gap Formation is 53.6 m thick at Scully Mountain and is disconformable with the overlying Jurassic Entrada Sandstone (Clark and Read, 1972).

Jurassic rocks

Above the Triassic strata in the Moreno Valley is a sequence of Middle Jurassic to Upper Jurassic sandstones, siltstones, mudstones and minor limestones which comprise the Entrada Sandstone, Bell Ranch Formation and Morrison Formation. These formations crop out in a band along the west side of the valley from the south end of Eagle Nest Lake to Scully Mountain and are shown on Figure 2 as Jurassic undivided. The description of these strata is based on sections measured by Clark (Clark and Read, 1972) west of Scully Mountain.

Entrada Sandstone

The Entrada Sandstone is light gray to buff and is composed of fine- to medium-grained, subrounded, well-sorted, non-calcareous, trough crossbedded quartzarenite. It is 3.7 m thick (Clark and Read, 1972) and disconformably overlies the Triassic Johnson Gap Formation.

Bell Ranch Formation(?)

Clark and Read (1972) observed a sequence of poorly exposed red and buff mudstones and gray limestones that conformably overlie the Entrada Sandstone. Lucas et al. (1990) observed a similar poorly exposed sequence of poorly sorted sandstone and siltstone east of the valley at Cimarroncito. Clark and Read (1972) believed that this unit is correlative with the Wanakah Formation in Colorado. Lucas et al. (1990) suggest that it is correlative with the Bell Ranch Formation of northeastern New Mexico.

The mudstones are red, buff and reddish brown, have siltstone partings and are locally nodular. The limestones are light gray, finely crystalline, thin-bedded and have scattered red jasper fragments. The total thickness of the Bell Ranch Formation(?) in the Moreno Valley is 9.1 m.

Morrison Formation

The Morrison Formation consists of interbedded, poorly exposed mudstone and sandstone. The mudstones are reddish brown, maroon, purple and mottled gray and brown, slightly calcareous to non-calcareous and have occasional gray nodular limestones. The sandstones are white, light gray, buff and reddish gray, fine- to medium-grained, medium- to thin-bedded quartzarenite to quartz wacke with occasional jasper grains, siltstone partings and clay cement. The Morrison Formation is faulted at Scully Mountain, where about 76 m of this unit are exposed. The Morrison Formation is disconformably overlain by the lower Cretaceous Mesa Rica Sandstone of the Dakota Group.

Cretaceous rocks

Rocks of Cretaceous age comprise the Dakota Group, Graneros Shale, Greenhorn, Carlile and Niobrara Formations and the Pierre Shale. The Trinidad Sandstone and Vermejo Formation are truncated by late Cretaceous(?) or Tertiary erosion east of Moreno Valley near Ute Park (Pillmore, 1976). The Raton Formation grades westward into the basal part of the upper Cretaceous(?) and lower Tertiary (Paleocene) Poison Canyon Formation. In the northern and northeastern Moreno Valley, units representing upper Cretaceous and lower Tertiary strata have been intruded by middle Tertiary rhyolite and quartz diorite porphyry.

The Cretaceous rocks are arbitrarily divided into two parts on Figure 2. The lower part (K1) consists of the Dakota Group, Graneros Shale and the Greenhorn Formation. The upper part (Ku) includes the Carlile and Niobrara Formations and the Pierre Shale.

Dakota Group

The Dakota Group consists of the lower Mesa Rica Sandstone, the middle Pajarito Formation (or Pajarito Shale) and the upper Romeroville Sandstone. Depositional environments of these units are summarized by Asquith and Gilbert (1976).

The Mesa Rica Sandstone is a yellowish gray, medium-grained, medium-bedded, ledge-forming quartzarenite with thin quartz veinlets that weather to rib-like projections (Clark and Read, 1972). The unit becomes coarser-grained and more conglomeratic northward toward the Colorado state line (Pillmore and Laurie, 1976). The Mesa Rica Sandstone is 8.5 m thick at Elizabethtown.

The Pajarito Formation (formerly Pajarito Shale) consists of medium to dark gray, locally sandy and carbonaceous shales with a yellowish gray, medium-grained, medium-bedded sandstone in the middle of the sequence (Clark and Read, 1972). The shale disappears northward toward Colorado (Pillmore and Laurie, 1976) as the Dakota Group becomes sandy and conglomeratic. The Pajarito Formation is 25.9 m thick. It is disconformable with the overlying Romeroville Sandstone.

The Romeroville Sandstone is essentially the same lithology as the Mesa Rica Sandstone at the base of the Dakota Group. It is conformable with the overlying Graneros Shale. The Romeroville Sandstone is 8.2 m thick. The Dakota Group at Elizabethtown has an aggregate thickness of 43.3 m.

Graneros Shale

The Graneros Shale consists of non-calcareous dark-gray shale with occasional thin calcareous zones. It also has a few thin, grayish-orange-pink bentonite beds. Clark and Read (1972) estimated a thickness of 122 m for this unit, though Simms (1965) measured 58.5 m of Granerous Shale in the Rayado area east of the valley.

Greenhorn Formation

The Greenhorn Formation consists of dark-gray, blocky, light-gray weathering, medium- to thin-bedded carbonate mudstone interbedded with dark gray calcareous siltstone and shale. This unit is well exposed about 3 km east of Eagle Nest in a roadcut at the head of Cimarron Canyon (Clark and Read, 1972). The maximum measured thickness is 6.7 m in this area.

Carlile Formation

The Carlile Formation conformably overlies the Greenhorn Formation and consists of dark gray to black, calcareous shale with thin lenses of dark carbonate mudstone and brown-weathering septarian nodules. The lenses of carbonate correspond to the Juana Lopez Member of the Carlile Shale in central New Mexico. Clark and Read (1972) measured a minimum thickness of 90.5 m at Mills Divide in the northeastern part of the valley. This thickness may be in error because of expansion of the section by quartz-diorite-porphyry intrusions.

Niobrara Formation and Pierre Shale undifferentiated

These two units are not separated because Clark and Read (1972) were unable to recognize a consistent contact between them; numerous thick sills of quartz diorite porphyry intrude the upper part of the Cretaceous sequence in the northern part of the valley further complicating relations. In general, the Niobrara Formation consists of black, slightly calcareous, silty shales and minor limestones (carbonate mudstones). The Pierre Shale consists of non-calcareous gray shales and is locally baked to hornfelsic shale near sills and dikes of quartz-diorite porphyry (Clark and Read, 1972). The aggregate thickness has been estimated at 646 m (MacLachlan, 1976), although some section is missing because of Late Cretaceous and early Tertiary erosion (Clark and Read, 1972).

Tertiary rocks

Rocks of Tertiary age include the Poison Canyon Formation (Late Cretaceous(?) to Paleocene), latite (or andesitic) flows and rhyolite and quartz-diorite intrusives (late Oligocene(?) to middle Miocene) (Robinson et al., 1964; Clark and Read, 1972; Reed et al., 1983), ash-flow crystal tuff of Cieneguilla Mountain (Miocene?) (Smith and Colpitts, 1980), Eagle Nest Formation (late Miocene(?) to Pliocene) (Ray and Smith, 1941) and basalt flows from the Ocaté volcanic field (Pliocene) (O'Neill and Mehnert, 1988; O'Neill, 1988).

Poison Canyon Formation

The Poison Canyon Formation consists of a lower light-gray to gray interval and an upper reddish, coarse interval (Clark and Read, 1972). These intervals are not differentiated on Figure 2.

The lower interval consists of interbedded grayish white, gray and white, coarse- to very coarse-grained, moderately to poorly sorted sandstone, conglomerate and minor gray shale and siltstone (Clark and Read, 1972; Pillmore, 1976). It is 76.2 m thick and disconformably overlies the Dakota Group 3 km south of Scully Mountain. It also disconformably overlies the Carlisle Formation at Mills Divide, and the lower Niobrara Formation and Pierre Shale at Elizabethtown. Clark and Read (1972) and Pillmore (1976) suggested that this lower interval might be equivalent with the lithologically similar lower Raton(?) Formation.

The upper interval of the Poison Canyon Formation consists of pale orange to red-colored coarse-grained conglomeratic arkose, arkosic and micaceous siltstones and conglomerates with boulders of quartzite, granite, paragneiss and quartzofeldspathic gneiss. Clark and Read (1972) suggested that the boulders were derived from outcrops of Precambrian rock in the vicinity of North Moreno Creek because they observed an increase in size of the boulders and cobbles in that direction.

Clark and Read (1972) and Pillmore (1976) noted that at least 305 m of the upper interval of the Poison Canyon Formation may be present in the Moreno Valley and the neighboring Cimarron Range.

Intrusive and volcanic rocks

Clark and Read (1972), Goodknight (1976), Smith and Colpitts (1980), Reed et al. (1983), O'Neill and Mehnert (1988) and O'Neill (1988) mapped a series of intrusive and extrusive igneous rocks in the Moreno Valley and surrounding mountains that range in age from late Oligocene to late Pliocene. They consist (in ascending order) of latite, rhyolite and quartz-diorite-porphyry intrusives in the northern part of the Moreno Valley, the ash-flow crystal tuff of Cieneguilla Mountain at Angel Fire and basalt flows of the Ocaté volcanic field.

The latite intrusives are compositionally variable. They are dark- to greenish-gray, aphanitic to porphyritic with phenocrysts of plagioclase, hornblende, biotite and locally sparse olivine, sanidine and quartz (Clark and Read, 1972; Lipman and Reed, 1989). These intrusives are confined to the northwestern corner of Moreno Valley near Red River Pass (Fig. 2) and are late Oligocene to early Miocene in age based on sparse radiometric dates (Reed et al., 1983).

The rhyolite intrusives are light-colored and have a porphyritic texture with phenocrysts of quartz, orthoclase, sparse sodic sanadine, albite and biotite. At the head of Cimarron Canyon the rhyolite occurs as sills, dikes and large intrusive bodies that were emplaced in the Mesozoic section along a series of faults that separate Precambrian from Mesozoic rocks (Clark and Read, 1972). This intrusive was mapped as dacite porphyry by Robinson et al. (1964).

The age of the rhyolite intrusions was estimated as early Miocene by Clark and Read (1972). Reed et al. (1983) give a middle Miocene age for similar rocks in the Red River area based on a few radiometric dates.

Quartz-diorite porphyry occurs as thick sills and dikes north of Eagle Nest. It is gray and consists of phenocrysts of plagioclase, hornblende, biotite and quartz in a groundmass of quartz, plagioclase, orthoclase(?), hornblende and biotite (Clark and Read, 1972). Clark and Read (1972) cite a K-Ar date of 33.8 Ma for this rock. This is probably in error as this unit intrudes rhyolites that are compositionally similar to ones mapped in the Red River area (Clark and Read, 1972; Reed et al., 1983) which yield K-Ar dates of about 24 Ma (Reed et al., 1983) (upper Oligocene to lower Miocene). This suggests that the quartz-diorite porphyry is at most lower Miocene and perhaps younger.

Contact-pyrometasomatic mineralization that includes magnetite, hematite, gold and minor pyrite is associated with the quartz diorite. Fissure deposits related to this intrusive include quartz-pyrite-gold veins around Elizabethtown and Iron Mountain. These fissure deposits are the source for the placer gold deposits along Moreno Creek (Clark and Read, 1972).

The ash-flow crystal tuff of Cieneguilla Mountain is a white to light-gray, unwelded to slightly welded, lithic, crystal-rich tuff containing phenocrysts of sanidine, quartz, slightly altered euhedral biotite and lithic fragments of purplish sandstone from the Sangre de Cristo Formation and reddish Precambrian quartzofeldspathic gneiss or granite (Smith and Colpitts, 1980). The biotite content increases southward and imparts a "salt-and-pepper" appearance to the rock. The tuff has a maximum thickness of 305 m.

Smith and Colpitts (1980) and O'Neill and Mehnert (1988) suggested a Pliocene(?) age for this unit based on its relationships to overlying basalts. A second possibility is that the tuffs of Cieneguilla Mountain may be related to formation of the 26 Ma (lower Miocene) Questa caldera; their lithology is similar to tuffs erupted during formation of that caldera. They could be the distal end of an outflow sheet that has not been previously documented. A third source might be by eruption from vents that are now buried beneath Cieneguilla Mountain and/or Agua Fria Peak. This last interpretation is based on the following data: (1) aeromagnetic maps covering this region show two magnetic anomalies beneath these mountains (Cordell et al., 1976); one anomaly lies directly beneath Cieneguilla Mountain and the other lies beneath the southwestern corner of Agua Fria Peak where another accumulation of ash-flow tuff crops out (Fig. 2); (2) lithic fragments in the tuff reflect a local source; and (3) these tuffs rest directly on Permian strata with no intervening valley alluvium. The overlying Pliocene basalts have localized accumulations of alluvium along their base. This implies that the ash-flow tuff was deposited before the valley formed or during early development of the ancestral Moreno Valley in Miocene time (see discussion of valley development below).

The basalt flows of the Ocaté volcanic field cover the southeastern corner of the map area and extend eastward toward Philmont and southward toward Ocaté and Guadalupita (O'Neill and Mehnert, 1988). They consist of phenocrysts of olivine, plagioclase and clinopyroxene in an aphanitic groundmass of plagioclase laths, prismatic augite, and equant crystals of magnetite and olivine-iddingsite; rare crystals of magnetite and quartz xenocrysts also occur. O'Neill and Mehnert (1988) give dates of 4 to 5 Ma for these rocks and interpret them to be tholeiitic to alkalic basalts. The flows originated from vents in Agua Fria Mountain and flowed away from the crest in all directions (Smith and Colpitts, 1980; O'Neill and Mehnert, 1988).

Late Tertiary and Quaternary rocks

Rocks of late Tertiary and Quaternary age include the pre-basalt Eagle Nest Formation and valley-fill deposits that post-date the basalt flows from Agua Fria Mountain but pre-date the modern stream alluvium. These units are shown only as QTa on Figure 2.

Eagle Nest Formation

The Eagle Nest Formation was first applied by Ray and Smith (1941) to unconsolidated, deformed and faulted interbedded reddish clay, white tuff and coarse, white to buff sand and gravel exposed near Eagle Nest Lake. Cobbles range to 45.7 cm in diameter and were derived from outcrops of rhyolite, quartz-diorite porphyry, sandstone and contact-metamorphosed Pierre Shale; the tuff was probably eroded from outflow sheets of ash-flow tuff from the eruption of the Questa caldera.

Ray and Smith (1941) interpret this unit as valley or basin fill deposited during the development of the Moreno Valley in middle to late Tertiary time. This formation appears to pre-date the volcanic rocks, but erosion and post-eruption valley filling has obscured most outcrops except near Eagle Nest Lake. Ray and Smith (1941) suggest a Pli-

ocene(?) age for this unit although it may range from Miocene to Pliocene in age.

Valley fill

These deposits include a variety of sediments that accumulated subsequent to the eruption of Ocaté volcanic field basalts that blocked the valley in Pliocene time. These sediments include interbedded light brown to gray, slightly silty clays, gray to gray-brown micaceous silts, gray to brown, poorly sorted lithofeldspathic sands, and light brown to brownish gray gravels with cobbles of quartzite, gneiss, schist, limestone, sandstone, basalt and ash-flow tuff (Smith and Colpitts, 1980). The fill represents stream, lake and alluvial fan deposits laid down after the eruption of the basalt from Agua Fria Mountain. Smith and Colpitts (1980) estimate from drill-hole data in the southern part of the valley that the thickness of the valley fill is greater than 128 m. These rocks are probably late Pliocene to Pleistocene in age.

Quaternary rocks

Quaternary rocks include unconsolidated sand, silt and gravel deposited by modern streams. Sands are medium- to very coarse-grained, and are poorly sorted. Silts are slightly sandy and clayey. The gravels are similar to those found in the valley fill and represent re-working of those deposits. Smith and Colpitts (1980) estimated a thickness of 4.6 m for this unit. Only the sands and gravels along the stream courses are shown as Qal in Figure 2.

Structural geology

The Moreno Valley lies near the eastern margin of the Sangre de Cristo Mountains within a zone of structural complexity that has been developing since late Precambrian time. This zone begins south of Las Vegas, New Mexico and continues northward into Colorado (Dane and Bachman, 1965). It is characterized by north-trending high- and low-angle thrust faults, high-angle normal faults, and northwest-trending transverse faults in addition to asymmetric and symmetric folds (Northrop et al., 1946; Robinson et al., 1964; Clark, 1966; Baltz, 1972; Clark and Read, 1972; Goodknight, 1976; Lessard, 1976; Lessard and Bejnar, 1976; Woodward and Snyder, 1976; Baltz and O'Neill, 1980, 1984; Smith and Colpitts, 1980). These features formed primarily within the last 60 million years; the earlier periods of deformation (i.e., late Paleozoic and Precambrian) will not be discussed. See Grambling and Dallmeyer (1990) for a discussion of Precambrian tectonics and Casey (1980a, b) for a discussion of late Paleozoic structural history.

Mesozoic-Cenozoic structural geology

Broad folding and uplift in the Moreno Valley developed in latest Cretaceous to early Tertiary time. Folding and uplift were followed by a period of erosion and renewed tectonism. This later deformation took the form of folding and north-trending thrust faulting. Folding and thrust faulting were followed in middle to late Tertiary time by north- to northwest-trending normal faulting; minor northeast-trending faults also developed concurrently. The deformational history of the Moreno Valley is complex; its interpretation is greatly complicated by talus, soil and forest cover.

Folding

Folds comprise two general types. The first are broad, north-plunging open anticlines and synclines with wavelengths of approximately 3 to 6 km. These form the Moreno Valley synclinorium described by Ray and Smith (1941). This group of folds is related to uplift and erosion of the northern part of the valley. The erosion resulted in deposition of the Raton and Poison Canyon Formations during Paleocene time.

The second set of folds is associated with thrust faulting. These folds have wavelengths of less than 1 km and display a range of plunge directions. Several are overturned toward the east and are found near the leading edges of thrust faults. These features are probably fault propagation folds related to thrust fault growth. Several are shown on Figure 2.

Thrust faulting

High-angle reverse (>45° dip) and thrust faults (<45° dip) occur in a narrow zone along the west side of the valley. These faults post-date the development of the broad, open folds and associated uplift. Broad open folds and alluvial fan deposits of the Poison Canyon Formation are cut by both high- and low-angle thrust faults from west of Eagle Nest to beyond Black Mountain on the northern edge of the valley (Fig. 2).

In the northern part of the valley between Apache Peak (south) and Scully Mountain (north) Clark and Read (1972) mapped the Sixmile Creek thrust which dips 10° to the west. Pennsylvanian rocks and Permo-Pennsylvanian Sangre de Cristo Formation on the west are thrust over moderately to steeply dipping Permo-Pennsylvanian through early Tertiary (Poison Canyon) strata on the east. At Comanche Creek on the south flank of Scully Mountain, the Sixmile Creek thrust is cut by the northwest-trending Comanche Creek transverse fault. To the north the high-angle Black Mountain reverse fault extends northward passing east of Black Mountain and Costilla Pass. The Black Mountain reverse fault juxtaposes Precambrian granitic and metasedimentary rocks on the west against Permo-Pennsylvanian through Paleocene strata on the east. Although Clark and Read (1972) did not report a dip value for this fault, its topographic expression suggests that it dips at least 50°W in the northernmost part of the valley. Clark and Read (1972) argue that the Black Mountain reverse fault is an extension of the Blue Lake thrust which lies 1.6 km west of Taos Cone. A simpler interpretation is to make the Black Mountain reverse fault a northern extension of the Sixmile Creek thrust. It is easier to reconcile the change from low to high dip angles if Pennsylvanian through Tertiary strata were downdropped south of the Comanche Creek transverse fault. If the thrust has a ramp-flat-ramp geometry, raising and eroding the block north of Comanche Creek would result in exposure of a lower ramp. Clark and Read (1972) report right-lateral strike-slip displacement on the Comanche Creek transverse fault; such movement could have placed a ramp segment in the position of the Black Mountain reverse fault.

Below the Sixmile Creek thrust fault is an allochthonous sheet herein named the Cimarron Thrust sheet. This sheet rests on the Cimarron thrust fault which dips 10°W and juxtaposes Permo-Pennsylvanian through Late Cretaceous strata in the Cimarron Thrust sheet against underlying Precambrian rocks. The underlying rocks are exposed in a small window west of Eagle Nest Lake (see Figs. 2, 3).

East and south of Eagle Nest Lake are small remnants of Triassic strata resting on Precambrian quartzite and quartzofeldspathic gneiss. Clark and Read (1972) and Goodknight (1976) show these strata in nonconformable contact with the Precambrian units. However, Clark and Read (1972) observed that rocks at the base of these outliers are sheared, suggesting that the contact may be a fault. If the contact is a fault, then these allochthons are probably part of the Cimarron thrust sheet described above (Figs. 2, 3).

North of the Comanche Creek transverse fault, and west of Scully Mountain, Wanek and Read (1956) observed a second window exposing the Cimarron thrust fault, where Sangre de Cristo Formation has been thrust over Precambrian granitic rocks. Northward, the Cimarron thrust is obscured by alluvium.

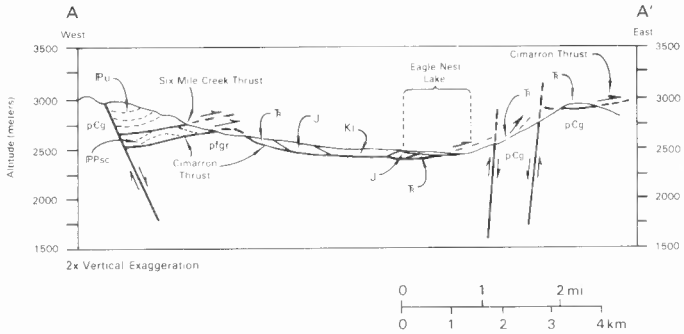


FIGURE 3. East-west geologic cross section through the central part of the Moreno Valley.

In the southern part of the valley, Smith and Colpitts (1980) observed a series of high-angle reverse faults. These faults form a zone that is easily projected along strike south toward Mora. Seismic reflection profiles by Charles B. Reynolds and Associates show these faults dip westward at angles that range from 45° to 65° (Smith and Colpitts, 1980). These faults are poorly exposed; scattered outcrops and roadcuts in the Angel Fire subdivision help delineate their approximate trends. These faults in the Angel Fire area are probably related to the Palo Flechado thrust whose strike carries it northwest of the valley. Southward, these thrusts apparently die out into a series of asymmetric, east-verging anticlines (Baltz and O'Neill, 1990). The manner in which the Palo Flechado thrust relates to the Sixmile Creek and Cimarron thrusts is not clear.

Clark and Read (1972) estimated a minimum shortening of about 2.7 km on the Sixmile Creek thrust. They did not infer any displacements on other thrust faults they mapped, but the presence of the Cimarron and Blue Lake thrust faults to the west suggests that more shortening may have occurred in this region. This amount of shortening implies that basinal facies of the Taos trough (Rowe-Mora basin) have been thrust eastward onto shelf and continental facies that were deposited along the western and southern edges of the Cimarron and Sierra Grande arches. This could also explain the sudden loss of Pennsylvanian marine strata across the axis of the Moreno Valley from west to east.

Normal faulting

Normal faults strike north, northeast and northwest throughout the valley and dip 60° to 90°. An earlier group strike roughly parallel the axis of the valley. These are cut by later transverse faults.

Valley-parallel faults dip eastward on the west side of the valley and westward on the east side (Clark and Read, 1972). They are most common on the east side of the valley where they form a stair-step pattern along the down-dropped eastern margin. On the west side, normal faults are poorly exposed; one normal fault mapped by Clark and Read (1972) east of Taos Cone and south of Taos Peak dips 60° to 65° east. It may be related to the formation of the ancestral Moreno Valley and does not seem to be associated with the Sixmile Creek thrust (Fig. 3).

Seismic reflection profiling in the southern part of the valley indicates that normal faults are widely distributed throughout the valley floor and not restricted to any one side (Smith and Colpitts, 1980). Seismic data also indicate that normal faults along the western side of the valley at Angel Fire cut folds and reverse faults. This is difficult to confirm elsewhere because exposures are so poor. Ray and Smith (1941) indicate that normal faults cut valley fill in the northern part of the valley.

Throws as large as 2740 m have been estimated on normal faults in the northern part of the valley (Clark and Read, 1972). Displacements on individual faults are difficult to assess because of slumping of the faulted beds and extreme lateral variability of the stratigraphic units. Smith and Colpitts (1980) report throws ranging from about 6 m to 91 m in the southern part of the valley. Baltz and O'Neill (1990) estimate stratigraphic throws of up to 183 m on normal faults near Guadalupita, south of Moreno Valley.

Pennsylvanian and allochthonous Triassic rocks can be used as markers to estimate displacements in Precambrian outcrops between Cimarron Canyon and Saladon Creek. Using these data, normal faults north of American Creek have throws that range from 100 m to 200 m, similar to throws observed by Baltz and O'Neill (1990), as described above.

Transverse normal faults form a dominant trend on the east side of the southern Sangre de Cristo Mountains that ranges between N40°W and N60°W. In the Moreno Valley these faults trend N58°W and divide the valley into three discrete segments. Dips on these faults are approximately 90° and displacements are typically down to the southwest.

Transverse normal faults show appreciable stratigraphic throws (Smith and Colpitts, 1980). The Saladon Creek fault juxtaposes Precambrian quartzofeldspathic gneiss, metasedimentary and granitic rocks on the north against Sangre de Cristo Formation on the south. Smith and Colpitts (1980) estimated a throw of approximately 3050 m for this fault. The Comanche Creek transverse fault juxtaposes Precambrian granitic, Paleozoic and Mesozoic rocks on the north against thrust-faulted Paleozoic, Mesozoic and Cenozoic sedimentary rocks on the south. This arrangement of Precambrian rocks faulted against Paleozoic rocks is repeated southward toward Mora in the Rincon Mountains (Baltz and O'Neill, 1990).

Some of these faults may represent reactivation of a Precambrian structural grain. Latest movements occurred prior to extrusion of the basalts from Agua Fria Peak (5 Ma) and after thrust, reverse and valley-parallel normal fault development. Actual age of latest movement is at least late Tertiary. Age of initial formation of the transverse normal faults is probably middle Proterozoic (possibly Grenville age) (Grambling, 1990, oral commun., 1990).

DEVELOPMENT OF THE MORENO VALLEY

The modern Moreno Valley is a product of a complex cycle of graben formation, sedimentation, uplift, volcanism and erosion that commenced in Miocene time (Ray and Smith, 1941; Robinson et al., 1964; O'Neill, 1988).

Following Poison Canyon deposition in the Paleocene came a period of erosion and reduction to base level by east-flowing streams. These streams carried sediment which formed the Ogallala Formation of the High Plains. During Miocene time, development of down-to-the-west valley-parallel normal faults formed a long, low, asymmetric valley. The Coyote Creek drainage system eroded headward into the area from the south, creating a low depression which subsequently filled with alluvial fan and stream deposits derived from nearby uplifted areas.

Sometime during the Miocene, Precambrian faults were reactivated and formed the transverse normal faults that segment the valley. In late Miocene time a period of uplift initiated formation of the Sangre de Cristo Mountains and the ancestral Moreno Valley. Erosion reduced the Moreno Valley area to a broad plain by late Miocene time.

In latest Miocene and early Pliocene time (4 to 5.5 Ma), a series of volcanic eruptions flooded the broad eroded plain, blocking southward flow into Coyote Creek from Moreno Valley. Lakes and swamps filled the valley and may have overflowed into drainages east of the rising Cimarron Range. Headward erosion by the Cimarron River coupled with possible down-cutting from the Moreno Valley across faulted rocks at the head of modern Cimarron Canyon led to capture of the valley drainage sometime during Pleistocene(?) time. Continued erosion coupled with uplift of the Sangre de Cristo Mountains during the Pleistocene led to complete integration of the valley flow into the Cimarron River. Continued base level reduction has produced a series of terraces that represent relict valley-fill deposits laid down before integration of Cieneguilla and Moreno Creeks with the Cimarron River.

RECOMMENDATIONS FOR STUDY

This compilation of geologic data for the Moreno Valley indicates the necessity for the investigation of several problems: (1) large-scale geologic mapping of the valley margins with southward extension down Coyote Creek to Mora; (2) continued investigation of mineral and hydrologic resources of the valley for possible future development; (3) further interpretation of depositional environments of the sedimentary rocks; (4) Cl^{36} dating of landforms in the valley floor to determine history of the valley development; and (5) determination of age of the ash-flow tuff of Cieneguilla Mountain.

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