# Contents

## Part III: Travel Itineraries

CHAPTER 1

# Geologic History of the San Luis Valley

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The complex geologic history of the San Luis Valley (SLV or the Valley) has intrigued geologists for well over a century. This chapter summarizes the many publications produced by these geologists. Three "recent" geologic events are chiefly responsible for the spectacular modern landscape of the Valley. The eastern San Juan Mountains are a large erosional remnant of the middle Tertiary Southern Rocky Mountain Volcanic Field, whereas the flat valley floor and the Sangre de Cristo Mountains that rise abruptly up from it owe their origin to the still active, late Cenozoic Río Grande rift. Pleistocene glaciation and related processes sculpted the mountains and deposited sediment. The Great Sand Dunes resulted from the combined effects of rifting, glaciation, and wind. Although the rift-related Sangre de Cristo Mountains are a younger physiographic feature than the eastern San Juan Mountains, most rocks exposed in the Sangres are hundreds of millions of years older than those in the San Juans.

The SLV also had a fascinating geologic history prior to the Southern Rocky Mountain Volcanic Field. For example, much of the Valley floor was formerly a mountain range, not just once but twice. During the late Paleozoic, the towering Uncompahgre–San Luis Uplift rose up where the flat Valley floor is today; where the Sangre de Cristo Mountains now stand high above the Valley floor, there was a deep structural trough that

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accumulated thousands of meters of sediment shed from the uplift. During the Late Cretaceous–early Tertiary Laramide Orogeny a second ancient mountain range, the San Luis Uplift, occupied much of today's Valley floor. Deep structural basins existed west of the Laramide uplift: the Monte Vista Basin was between Alamosa and Del Norte, and the San Juan Sag, now concealed beneath a thick pile of volcanic rocks in the eastern San Juan Mountains, was farther west.

## GEOGRAPHIC AND PHYSIOGRAPHIC SETTING

The SLV is a large, high-altitude, intermontane valley within the Southern Rocky Mountains that stretches across south-central Colorado and north-central New Mexico. The SLV includes the broad, elongate, low-relief Valley floor and parts of two majestic mountain ranges that drain into the Valley. Plate 1 shows the extent of the Valley and its major geographic and physiographic features.

The Valley encompasses the headwaters of the Río Grande and extends from the drainage divide in the San Juan and Tusas Mountains on the west to the crest of the Sangre de Cristo Mountains on the east. The northern end of the Sangre de Cristo Mountains and southern end of the Sawatch Range bound the northern end of the Valley, while the Valley's southern end is near Pilar, New Mexico. Based on these limits, the SLV extends across nearly 25,000 km$^2$ (9,650 miles$^2$).

The Conejos and Red Rivers are major tributaries of the Río Grande. Surface water in most other streams and creeks in the northern part of the Valley, including San Luis Creek, either infiltrates into the ground, is diverted for irrigation, or flows into closed basins before reaching the Río Grande—helping to recharge both unconfined and confined aquifers below the Valley floor (Harmon, this volume).

The gradually rising San Juan Mountains on the west side of the Valley are known as the Eastern San Juans. A subset of the Eastern San Juans is the La Garita Mountains, which extend from near Creede toward Saguache. The section of the Eastern San Juans between US Highway 160 and the state line is also known as the South San Juans. Continuing south into New Mexico are the Tusas Mountains.

The narrow Sangre de Cristo Mountains rise abruptly above the eastern Valley floor. Within the SLV, the Sangres are subdivided into three ranges. The Northern Sangre de Cristo Range extends from Poncha Pass southward to the Blanca Massif, which includes four "fourteeners," peaks with an altitude of 14,000 feet (4,268 m) or more. Another group of fourteeners is found within the Northern Sangres above the town of Crestone.

At Blanca Massif the range bends sharply and trends east-northeast for ~22 km (~14 miles). At La Veta Pass the range returns to its southerly trend and continues to just beyond the state line, forming the Culebra Range. A single fourteener exists in the Culebra Range; it is the highest privately owned peak in the United States.

Near the state line the Sangres swing southwestward. The Taos Range starts there and extends south beyond Taos. It includes the highest peak in New Mexico. Both the Northern Sangres and the Taos Range rise abruptly from the Valley floor, whereas a wide section of relatively low-lying foothills borders the west side of the Culebra Range. A relatively low-altitude, west-trending spur called the Picurís Mountains splits off the south end of the Taos Range.

Upson (1939) subdivided the Valley into several physiographic areas, defined in part by the San Luis Hills that rise up between Alamosa and the state line. The large, broad, relatively flat region north of the San Luis Hills is the Alamosa Basin. Much of Alamosa Basin north of the Río Grande is an internally drained, closed, topographic basin.

Culebra Reentrant is the area where the Culebra Range is recessed eastward relative to the Northern Sangre de Cristo Range and the Taos Range. Between the Culebra Reentrant and San Luis Hills is the Costilla Plains. The Taos Plateau occupies much of the southern part of the SLV. Many young volcanoes rise above the Taos Plateau, including San Antonio Mountain and Ute Mountain. The deep, narrow Río Grande Gorge was cut by the Río Grande as it eroded through the Taos Plateau.



## GEOLOGIC HISTORY

Figure 1.1 is a geologic time chart annotated with the SLV's major geologic events. Geologic formations found in the Valley are listed in the stratigraphic chart in figure 1.2. Plate 2 is a simplified geologic map showing where formations, or groups of formations, crop out in the Valley. Only major faults that are still active are depicted on plate 2. They include the Sangre de Cristo fault zone, Embudo fault, Valley View fault zone, and Mesita fault.

Geologic time is informally subdivided into the Precambrian and Phanerozoic. The Precambrian started when the Earth formed nearly 4.6 billion years ago (4.6 Ga), and it ended about 541 million years ago (541 Ma) when a diverse assemblage of hard-shelled animals first appeared, marking the beginning of the Phanerozoic. The Precambrian consists of three time intervals: the Hadean, Archean, and Proterozoic Eons.

During the Hadean, planet Earth had just formed and was still very hot as a result of widespread volcanism, a partially molten surface, and frequent collisions with other solar system bodies. Rocks of Hadean age are not found in the Valley and are very rare on the planet. Archean rocks comprise the original North American craton, which consists of several assembled blocks, including the Wyoming Craton. The southern margin of the Archean craton lies in Wyoming and northwestern Colorado; hence Archean rocks do not exist in the Valley.

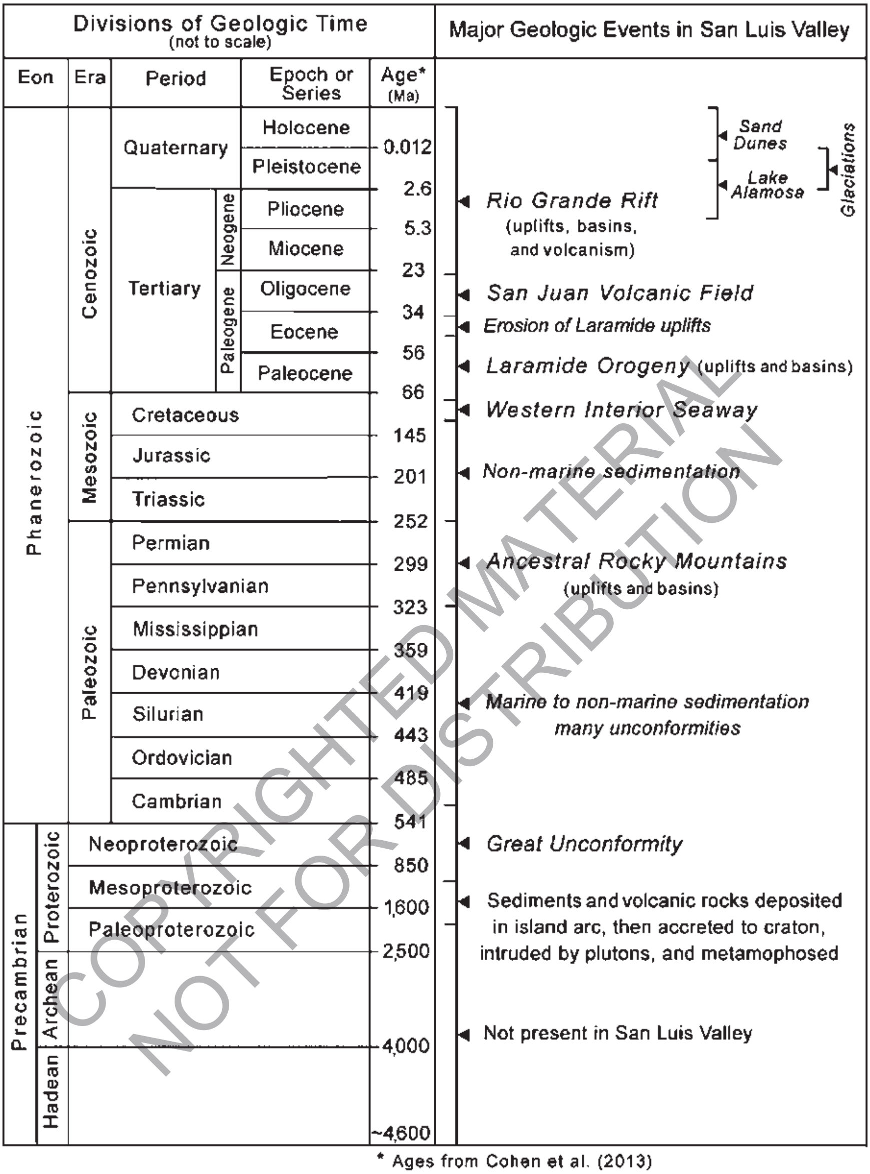



**FIGURE 1.1.** *Geologic time chart, annotated with major geologic events in San Luis Valley.*

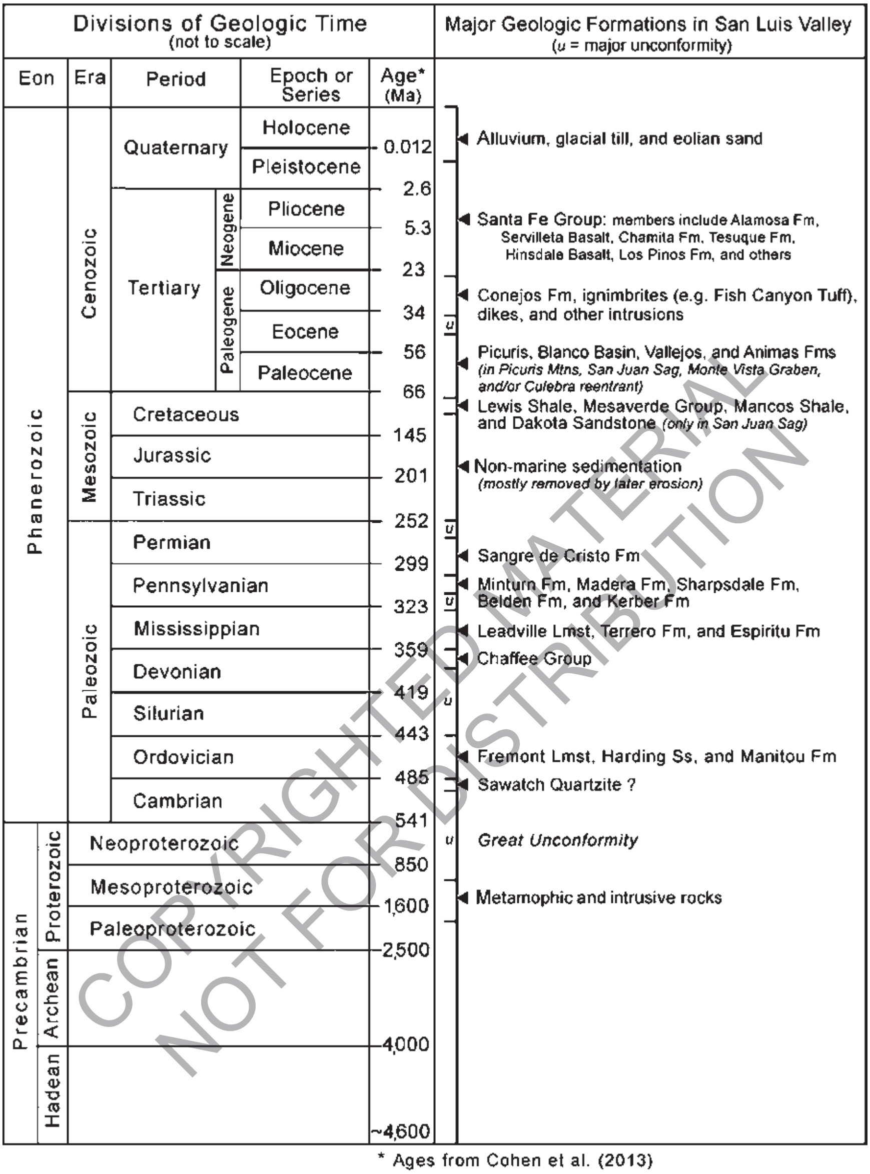


**FIGURE 1.2.** *Stratigraphic chart showing the major bedrock formations in San Luis Valley.* u; *major unconformity.*

## PROTEROZOIC EON

The SLV's oldest rocks are 1.7 to 1.8 Ga Paleoproterozoic metamorphic and igneous rocks. Thus the first ~2.8 billion years (~61%) of Earth's history are not recorded in the Valley's rocks. Many of the metamorphic rocks were originally sedimentary or volcanic rocks that were deeply buried, then metamorphosed by ~1.7 Ga intrusions that are chiefly granitic. Mafic intrusions of this age exist in some areas.

Much of the Paleoproterozoic rock exposed in the Valley consists of foliated granitic rocks and gneiss in the core of the Sangre de Cristo Mountains. Paleoproterozoic metavolcanic rocks, pegmatite dikes and lenses, and metasedimentary rocks such as quartzite, metaconglomerate, and slate are also present. Metasedimentary rocks are widespread in the Picurís Mountains (Bauer and Williams 1989).

Mesoproterozoic plutons intruded the Paleoproterozoic rocks at ~1.4 Ga (e.g., Tweto 1980; Wobus 1984; Karlstrom et al. 2004; Lindsey 2010). Mafic gabbro dikes locally intruded the Paleoproterozoic crystalline rocks in the Sangre de Cristo Mountains. Some geologists assign a Neoproterozoic age to them (e.g., NMBG&MR 2003; Lindsey et al. 2012), while others classify them as Cambrian (Reed 1984) or undifferentiated early Paleozoic/Neoproterozoic (e.g., Wallace and Lindsey 1996; Kirkham et al. 2005b; Fridrich and Kirkham 2007; Thompson et al. 2015).

Whitmeyer and Karlstrom (2007) developed a model for the Precambrian assembly and growth of the North American craton (plate 3). In this model, Proterozoic rocks in the northernmost part of the Valley were accreted to the craton as part of the Yavapai Crustal Province. Those found farther south were accreted during the Yavapai-Mazatzal Transition. (See Gonzales and Karlstrom [2011] for paleogeographic maps showing the movement and accretion of landmasses from about 1.8 to 1.68 Ga.) These rocks were originally located in the Southern Hemisphere and later transported to their modern location by plate tectonics.

Bickford and Hill (2006) and Hill and Bickford (2001) offered an alternative theory on how the ancient continental craton formed. They proposed that the existing Wyoming Craton expanded by extensional rifting and magmatism, not accretion. Another theory on the Proterozoic evolution of the southern part of the Valley involves a Mesoproterozoic orogeny recorded in rocks in the Picurís Mountains, Tusas Mountains, and nearby uplifts (Daniel et al. 1995, 2013).

Spectacular exposures of Proterozoic rocks can be seen in glacial cirques in the Blanca Massif (plate 4). Proterozoic rocks are also common in the southern end of the Sawatch Range, Picurís Mountains, and Tusas Mountains. Proterozoic rocks along the Continental Divide in the upper reaches of the Río Grande above the abandoned mining camp of Beartown are part of a large expanse of exposed Precambrian rocks in the Needles Mountains farther west. Tiny windows of Proterozoic granite poke through a thick pile of overlying Cenozoic volcanic rocks in the Conejos River canyon at Aspen Glade Campground and also along Forest

Service Road 250 above its junction with Colorado Highway 17 (Lipman 1975) but are too small to show on plate 2.

## EARLY AND MIDDLE PALEOZOIC TIME: CAMBRIAN THROUGH MISSISSIPPIAN PERIODS

A long period of erosion started during the later part of the Proterozoic and continued into the Paleozoic, reducing the Precambrian landscape to sea level. This erosion created a time gap in the rock record called "the Great Unconformity" that separates the Proterozoic rocks from overlying, younger Paleozoic sedimentary rocks. In the Northern Sangres, where Ordovician rocks rest on Proterozoic basement rocks, about 1 billion years of Earth's history is missing. More of Earth's history is missing where younger formations rest on Proterozoic rocks, as in the Culebra Range, Taos Range, and Conejos River canyon.

Major regional unconformities also characterize early and middle Paleozoic time in the Valley. Ross and Tweto (1980) pointed out that Cambrian, Ordovician, Silurian, and Devonian sedimentary formations in Colorado record only about 30 to 40 Ma of the approximately 180 Ma spanning these geologic periods, whereas the unconformities account for over three-quarters of this time interval. They comment that "one may marvel that any sediments have escaped erosion" (Ross and Tweto 1980, 47). Lower and middle Paleozoic rocks in the Valley record even less geologic time because of local emergent areas and uplifts in south-central Colorado (Ross and Tweto 1980; Lindsey 2010).

Lower Paleozoic sedimentary formations were deposited in the Colorado Sag, located between the Souixia Uplift on the north and the Uncompahgre and Sierra Grande Uplifts on the south (Myrow et al. 1999). When the ocean encroached across the sag, near-shore and marine shelf sediments were deposited. When the ocean retreated and the landmass was exposed, erosion stripped some or all of the sediment from the landmass. This sequence of events was repeated several times during the early Paleozoic, leaving behind formations composed mostly of limestone, dolomite, and sandstone.

Lindsey (2010) described evidence preserved in the Northern Sangres and Culebra Range of an Ordovician uplift that apparently persisted into the Mississippian Period (figure 1.3). The uplift extended from about the latitude of the Great Sand Dunes southward at least to the Colorado–New Mexico line. Lower Paleozoic sediment was deposited north of the uplift and progressively on-lapped the uplift from about the latitudes of Crestone to the Great Sand Dunes. During the Mississippian Period, the sediment may have eventually covered the entire uplift. Farther south, Cambrian through Devonian strata are absent, and Mississippian rocks are not present until south of Taos (Bauer et al. 2000; NMBG&MR 2003).

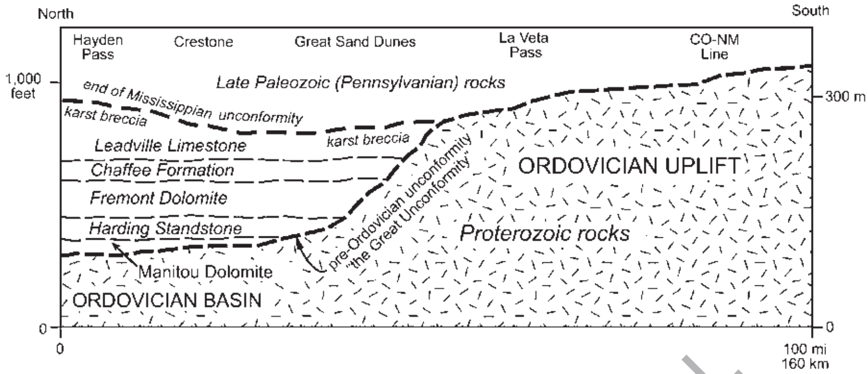


**FIGURE 1.3.** *Cross-section showing relationships between lower and middle Paleozoic formations and the Ordovician uplift along the crest of the Sangre de Cristo Mountains in Colorado. Modified from Lindsey (2010).*



The shallow sea drained away after deposition of the Mississippian limestones, and an extended period of erosion and dissolution resulted in a second major unconformity (De Voto 1980a; Lindsey 2010). The climate was warm and humid, and exposed limestone was subjected to erosion and dissolution. Sinkholes and caves formed, creating a paleo-karst topography.

Cambrian sedimentary rocks (Sawatch and Dotsero Formations) are the most widely exposed lower Paleozoic rocks in Colorado (Ross and Tweto 1980; Myrow et al. 2003), but they are not definitively recognized in the SLV. Scattered patches of quartzite found on Proterozoic rocks in the mountains west of Villa Grove may be thin remnants (centimeters thick) of Sawatch Quartzite (Burbank 1932; Cappa and Wallace 2007). Cambrian intrusive plutons are exposed in areas near the SLV, including in the Wet Mountains and at Powderhorn in Gunnison County. Cambrian plutons have not been discovered in the rocks exposed in the Valley, but they could exist in the subsurface.

The oldest known Paleozoic sedimentary rocks in the Valley are Ordovician age. They include the Manitou Dolomite, Harding Sandstone, and Fremont Dolomite (Lindsey 2010). Some of the oldest fish fossils found in the world come from the Harding Sandstone north of the Valley (Walcott 1892; Sansom et al. 1996). Overlying the Ordovician rocks are the Devonian Chaffee Formation and its members: Parting Quartzite and Dyer Dolomite. Ordovician and Devonian rocks, as well as the Mississippian Leadville Limestone, crop out only in the Northern Sangres and the southern end of the Sawatch Range. The oldest Paleozoic rocks in New Mexico's part of the Valley include the Mississippian Terrero and Espíritu Santo Formations of the Arroyo Peñasco Group, which crop out only in a small area south of Taos (Armstrong et al. 2004).

## RISE AND FALL OF THE ANCESTRAL ROCKY MOUNTAINS: LATE PALEOZOIC

The Ancestral Rocky Mountains formed during the late Paleozoic Pennsylvanian and Permian Periods (Mallory 1972a, 1972b; De Voto 1980b; Lindsey et al. 1983, 1986a; Lindsey 2010; Hoy and Ridgway 2002). Figure 1.4 outlines the major late Paleozoic uplifts and troughs in Colorado and northern New Mexico. Not all of the uplifts and troughs were active at the same times, and in some cases different parts of the same large uplift behaved differently.

Various models are invoked to explain the enigmatic formation of the intraplate, crustal-scale Ancestral Rockies and associated basins far inland from the plate boundary. Most models attribute the Ancestral Rockies to the collision of two large continental landmasses: a northern landmass called Laurasia and a southern landmass called Gondwana (e.g., Gonzales and Karlstrom 2011). The collision led to the development of the supercontinent Pangaea. Most models attribute the late Paleozoic crustal deformation to either compressional forces or strike-slip transform/wrench faults (e.g., Kues and Giles 2004; Kluth and DuChene 2009). The Appalachian Mountains formed during the early part of the collision, whereas the Ancestral Rockies rose later during the collision after tectonic stresses transferred westward across the region (Gonzales and Karlstrom 2011).



Over the years, a plethora of names were applied to the various Ancestral Rocky Mountain uplifts, troughs, and basins. Our terminology for the major late Paleozoic uplifts and troughs is provided in figure 1.4. The late Paleozoic Uncompahgre–San Luis Uplift rose up within the western part of the SLV. It extended from eastern Utah through west-central Colorado and into south-central Colorado and north-central New Mexico. Streams draining the east side of the uplift flowed into the Central Colorado Trough (or Taos Trough as it is known in New Mexico). The Picurís-Pecos fault is a major north-south–trending structure exposed in the Picurís Mountains. In some interpretations the fault extended northward into Colorado and served as the master fault along the eastern margin of the Uncompahgre–San Luis Uplift (e.g., De Voto 1980b; Kues and Giles 2004). The Picurís-Pecos fault played an important role in the structural evolution of the area throughout much of the Proterozoic and Phanerozoic (Bauer and Ralser 1995; Kelley 1995; Fankhauser and Erslev 2004).

At the latitude of the SLV, the trough was bordered on the east by the Apishapa and Sierra Grande Uplifts. They were relatively low-elevation uplands compared to the towering Uncompahgre–San Luis Uplift, and less sediment was shed from them. The northern end of the Apishapa was either taller or rose more quickly than the rest of the uplift because sediment eroded off of it was coarser-grained, leading some geologists to consider the northern end a separate uplift, the Ute Pass Uplift.

As the Uncompahgre–San Luis, Apishapa, and Sierra Grande Uplifts rose, lower and middle Paleozoic rocks were stripped off, exposing the Proterozoic core of the

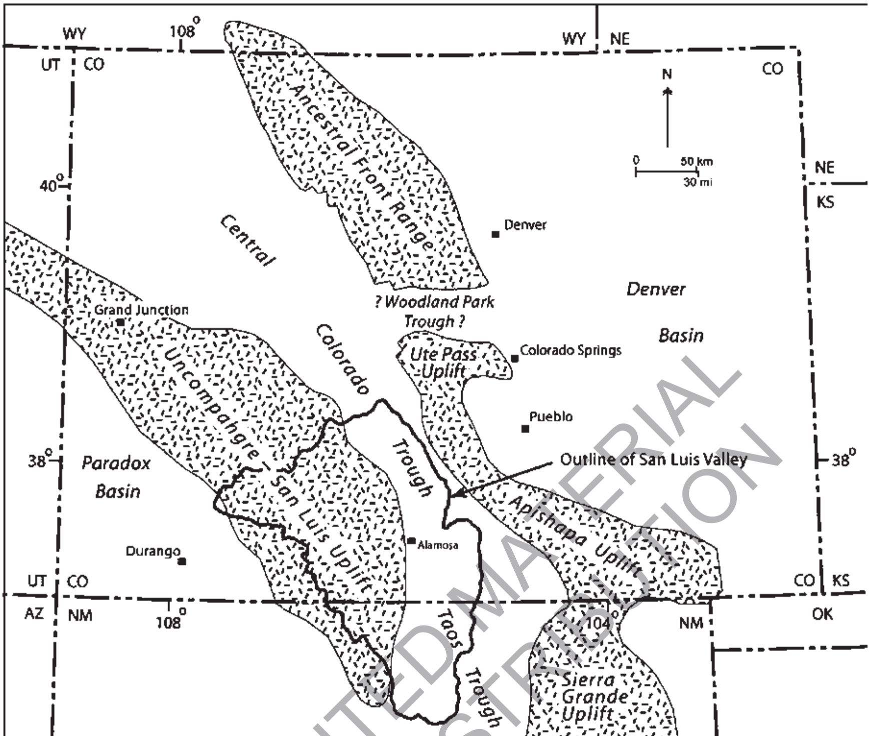


**FIGURE 1.4.** *Generalized map of the major late Paleozoic Ancestral Rocky Mountain uplifts (areas with patterns) and troughs (no patterns) of Colorado and northern New Mexico. Outline of uplifts modified from Lindsey et al. (1986a) and Barkmann et al. (2016).*



uplifts to erosion. Thick sequences of sediment, locally thousands of meters thick, eroded from the uplifts and accumulated in the troughs. Alluvial fans similar to those at the base of today's Sangre de Cristo Mountains (plate 4) formed along the eastern side of the Uncompahgre–San Luis Uplift, as evidenced by the coarse-grained conglomerates shed from it that now crop out in the Northern Sangres.

The late Paleozoic topography was somewhat similar to today's setting in that a deep fault-bounded basin lay between mountain ranges, although the ancient mountains and basins were offset to the east relative to today's topography. Another difference was that the late Paleozoic uplifts were essentially large islands surrounded by the ocean, and marine conditions existed in the deeper parts of the troughs between the landmasses.

Initially, non-marine clastic sediments were deposited in terrestrial environments adjacent to the uplifts, and shallow marine sediments were dominant in the central and deeper parts of the trough (Lindsey et al. 1986a). Sedimentary strata deposited in river deltas along the coastline of the troughs contain abrupt facies changes where beds of marine and marginal non-marine rocks are repeatedly interbedded in a stacked sequence. These cyclical sequences are interpreted as evidence of the rise and fall of the worldwide sea level as glaciers on Gondwana grew and melted during the Paleozoic (e.g., Lindsey 2010; Barkmann et al. 2016). The Ancestral Rockies were originally located near or south of the equator and later transported to its present location by plate tectonics (Lindsey 2010; Sweet and Soreghan 2010).

Near the end of the Pennsylvanian, the marine incursions into the trough ended, allowing non-marine clastic rocks to be deposited farther out into the trough, eventually blanketing the ~1,500-m-thick section of marine sediments with as much as 2,000 m of arkosic conglomerate. In the SLV, the best exposures of late Paleozoic sedimentary rocks are in the Sangres. Lindsey and colleagues (1986a) and Lindsey and Soulliere (1987) included all of the late Paleozoic strata in the Northern Sangres in the Minturn and overlying Sangre de Cristo Formations. The Minturn consists of interbedded marine and non-marine sandstone, siltstone, shale, and minor limestone, as well as turbidites, whereas the overlying Sangre de Cristo Formation consists almost entirely of continental sandstone, conglomeratic sandstone, siltstone, and shale. The Crestone Conglomerate is a prominent, well-known member of the Sangre de Cristo Formation that includes cobble and boulder conglomerates, sandstone, and minor siltstone and shale deposited in an ancient alluvial fan close to the Uncompahgre–San Luis Uplift (Mallory 1972b; Lindsey et al. 1986a). A partial section of late Paleozoic sedimentary rock also crops out in the southern end of the Sawatch Range where the marine Belden Formation and overlying Minturn Formation are preserved (Scott et al. 1978).

In the northern end of the Culebra Range, some geologists map late Paleozoic strata as the Minturn and Sangre de Cristo Formations (Vine 1974), whereas others (e.g., Wallace and Lindsey 1996; Thompson et al. 2015) correlate rocks beneath the Sangre de Cristo Formation with the Madera and Sandia Formations of Baltz and Myers (1984). At the southern end of the Taos Range, late Paleozoic formations include the Flechado, Alamitos, Sangre de Cristo, and Yeso (Kues and Giles 2004). Late Paleozoic sediments are postulated to exist in the subsurface beneath the southern end of the Taos Plateau (Bauer et al. 2016; Baltz and Meyers 1984; Bauer and Kelson 2004b). For more information on the late Paleozoic in the Valley, refer to Mallory (1972a, 1972b); De Voto (1980b); De Voto and Peel (1972); Sutherland (1972); Lindsey et al. (1983, 1986a, 1986b); Kluth (1998); Hoy and Ridgway (2002); Kues and Giles (2004); Sweet and Soreghan (2010); and Lindsey (2010).

By middle Permian time, erosion had reduced the once mighty Ancestral Rockies to a low-relief landscape that barely stood above the sedimentary fill deposited in adjacent basins. By late Permian time, the entire Valley was apparently an area of non-deposition (Maughan 1980). The Paleozoic Era ended with one of the largest mass extinction events to devastate the planet (Stanley and Luczjak 2015), called the "Great Dying," during which much of the marine and terrestrial life on Earth was wiped out by the Permian-Triassic extinction event.

## MESOZOIC ERA

The Mesozoic Era, or age of dinosaurs, consists of the Triassic, Jurassic, and Cretaceous Periods. Triassic rocks are not known to exist on either the land surface or in the subsurface within the Valley. The Valley may have continued to be an emergent area of non-deposition during the arid Triassic Period, or perhaps the rocks were eroded off during the later Laramide Orogeny. Triassic rocks crop out on the west side of the Tusas Mountains and on the east side of the Culebra Range near the state line (NMBG&MR 2003; Kirkham et al. 2005b; Fridrich and Kirkham 2007), so they may potentially exist in the subsurface in the southern part of the Valley.

By Middle to Late Jurassic time, sediments had progressively on-lapped onto the eroded remnants of the Uncompahgre–San Luis Uplift, and non-marine sediment deposition spread across the entire eroded remnant of the uplift and adjacent areas (Brister and Chapin 1994). The Morrison Formation, famous for its Jurassic dinosaur fossils, was deposited at this time. As the North American continent was driven northward by plate tectonics during the Jurassic, the climate became wetter, and rivers and lakes were widespread.



## WESTERN INTERIOR SEAWAY: LATE CRETACEOUS

During the Cretaceous, the Arctic Ocean encroached onto the continent from the north, and the Gulf of Mexico/Tethys Sea approached from the south. They met during the Late Cretaceous, creating the Western Interior Seaway and splitting the North American continent into a western landmass called Laramidia and an eastern landmass called Appalachia (plate 5). A number of geologic events, or more likely a combination of them, created the Western Interior Seaway (Mitrovica et al. 1989; Nummedal 2004). Subduction of the Farallón Plate under the North American Plate off the west coast of the continent, thrusting in the Sevier orogenic belt west of the seaway, the load of sediments deposited on the sea floor, and fluctuations in global sea levels during the Late Cretaceous are some of the proposed mechanisms that caused the land surface to subside and allowed the seas to encroach on the interior of the continent.

As the seaway grew, it transgressed (spread) across the region. Around 100 Ma, the interior seaway transgressed westward across the SLV, and by ~93 Ma the shoreline was far west of the Valley. Rivers carried large volumes of sediment eroded from mountains in Laramidia and deposited thick sequences along the shoreline and in the seaway. During this initial phase of transgression, the Dakota Sandstone was deposited in near-shore environments; clay, silt, and limestone deposited offshore became the Mancos Shale, whose members include the Graneros, Greenhorn, Carlile, and Niobrara Formations (Gries 1985).

After transgressing westward into Utah and Arizona, the shoreline regressed (retreated) back into Colorado, depositing a thick package of non-marine sediments called the Mesaverde Group over the Mancos Shale. While the retreating shoreline was located in the SLV, ocean levels rose and the seaway transgressed westward a second time, depositing another thick sequence of marine sediments called the Lewis Shale. When the Western Interior Seaway finally drained off the interior of the continent, it left behind near-shore deposits of sandstone and coastal plain sediments that followed the regressing seaway. As a result of later uplift and erosion, none of the strata deposited during the final regression exist in the Valley. Based on data from the San Juan and Ratón Basins, where Cretaceous sediments are well exposed, the shoreline regressed through the SLV about 70 Ma. This was the last time marine conditions prevailed in the Valley.

Outcrops of Jurassic and Cretaceous rocks are rare within the Valley but widespread outside of it (plate 2). A small outcrop of flat-lying Jurassic Morrison Formation and Cretaceous Dakota Sandstone rests on Proterozoic gneiss near Trickle Mountain about 25 km northwest of Saguache (Tweto et al. 1976), and two very small exposures of Morrison and Dakota strata crop out in arroyos between Crestone and the Great Sand Dunes (Watkins 1996).

Since Jurassic and Cretaceous strata are rare in the Valley, how do geologists know they once blanketed it? Evidence found in areas surrounding the Valley indicates that Jurassic and Cretaceous formations were once continuous across the region (e.g., Berman et al. 1980; Anderson and Lucas 1997; Leckie et al. 1997) but were almost entirely removed from the Valley by erosion later during the Laramide Orogeny. As described in the following section, recently drilled oil test holes encountered Jurassic and Cretaceous rocks deep in the subsurface west of Del Norte and near Crestone (Gries 1985; Watkins 1996). Also, geophysical data led Drenth and colleagues (2011) to speculate that Mesozoic strata might exist deep beneath the San Luis Hills.



## LARAMIDE OROGENY: LATE CRETACEOUS TO EARLY CENOZOIC

Starting at ~75 to 70 Ma, a new set of ancient mountains and basins began to form during a period of mountain building called the Laramide Orogeny. Figure 1.5 shows the Laramide uplifts and basins in and near the Valley. The Valley's Laramide paleogeography was almost a mirror image of its late Paleozoic paleogeography. Where thick sediments accumulated in the late Paleozoic Central Colorado Trough, the San Luis Uplift formed during the Laramide Orogeny. Large parts of the late Paleozoic Uncompahgre–San Luis Uplift became Laramide basins (San Juan Sag and Monte Vista Basin) that received sediment eroded off the Laramide uplifts.

A popular model for the Laramide Orogeny involves plate tectonics (e.g., Dickenson 1981; Dickenson and Snyder 1978). In this model, flat-slab or shallow subduction off the west coast of North America extended far inland. As the rate of subduction increased, the subducted slab became more buoyant and rose higher beneath the Rocky Mountain region, triggering crustal compression, uplift, and magmatism. More recent geophysical and geologic studies suggest an alternate model that does not involve subduction (e.g., Gonzales and Karlstrom 2011; Karlstrom et al. 2005; McCoy et al. 2005a, 2005b; Livaccari and Perry 1993). In this model, forces within the Earth reactivated old fractures and released pressures within the mantle that initiated crustal melting and compression.

Thrust faults were responsible for much of the compressional deformation along the eastern flank of the Laramide San Luis Uplift. Strike-slip faults formed in some areas, for example, the Picurís-Pecos fault at the southern end of the Valley (Bauer and Ralser 1995). Thrust faults with many kilometers of offset and associated large-amplitude folds are well exposed in the Northern Sangres (Lindsey et al. 1983, 1986b) and in the Culebra Range (Lindsey 1998).

During the Laramide, mountains rose across much of the SLV (figure 1.5). The north-south–trending San Luis Uplift extended across the eastern part of the Valley. The southern part of the Valley was occupied by the San Luis, Sangre de Cristo, and Brazos Uplifts, while the northwestern-most part was within the San Juan Uplift. In between were northeastern extensions of the San Juan Basin: the San Juan Sag and the Monte Vista Basin, which were separated by the Del Norte High. In contrast to the nearby Laramide-age San Juan, Ratón, and Chama Basins, the San Juan Sag and Monte Vista Basin lack surface expression because they are concealed by younger volcanic and sedimentary rocks (figure 1.6).

Gries (1985, 1989) described how the San Juan Sag was discovered. Kelley (1955) and Larson and Cross (1956) were the first to postulate the existence of Mesozoic sedimentary rocks in a Laramide basin concealed beneath the thick pile of middle Tertiary volcanic rocks in the eastern San Juan Mountains. Their interpretation was discredited by a geophysical study that indicated the presence of an extensive igneous batholith under the entire volcanic field, which would have obliterated any Mesozoic sedimentary rock that might have existed there (Plouff and Pakiser 1972). Then geologists found places in the Conejos River Valley where middle Tertiary volcanic rocks rested directly on Precambrian rocks, raising additional doubts about the sag. However, in 1975, oil was discovered in mineral exploration cores drilled at the eroded remnant of the Summer Coon Volcano north of Del Norte. Bob and Bill

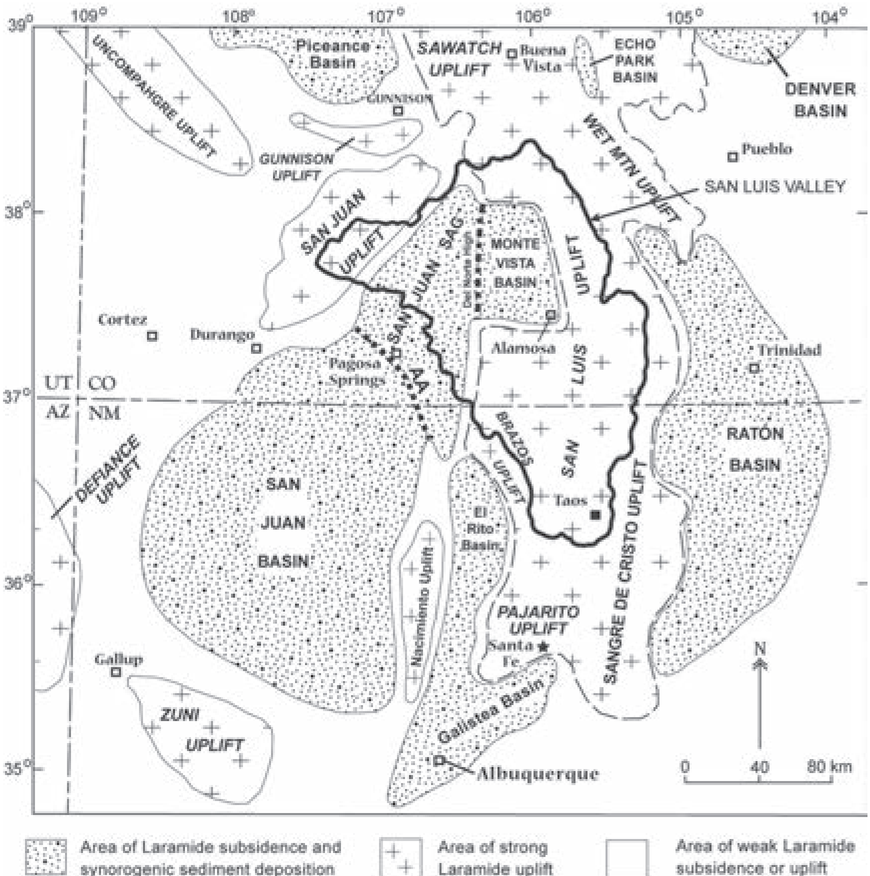


**FIGURE 1.5.** *Major Laramide uplifts and basins in the vicinity of San Luis Valley. Archuleta Anticlinorium (AA) separates San Juan Basin from San Juan Sag. Modified from Cather (2004).*

Ellithorpe told geologists about an oil seep on Hope Creek north of Big Meadows. The oil emanated from volcanic rocks but must have originated in oil-bearing sedimentary rocks concealed beneath the volcanic rocks. Oil was also found in discarded mineral exploration cores at Summitville. These clues prompted the drilling of test holes in the 1980s that proved the existence of the concealed San Juan Sag.

Subsurface relationships between Laramide structures and basins, overlying middle Tertiary volcanic rocks, and the late Cenozoic San Luis Basin are depicted in figure 1.7. The Monte Vista Basin lies under the western Valley floor. It contains nearly 700 m of arkosic sediment in the Eocene Blanco Basin Formation, is concealed beneath volcanic rocks and rift-related late Cenozoic sediments, and was modified by later rifting (Brister and Gries 1994). The San Juan Sag, which is buried beneath

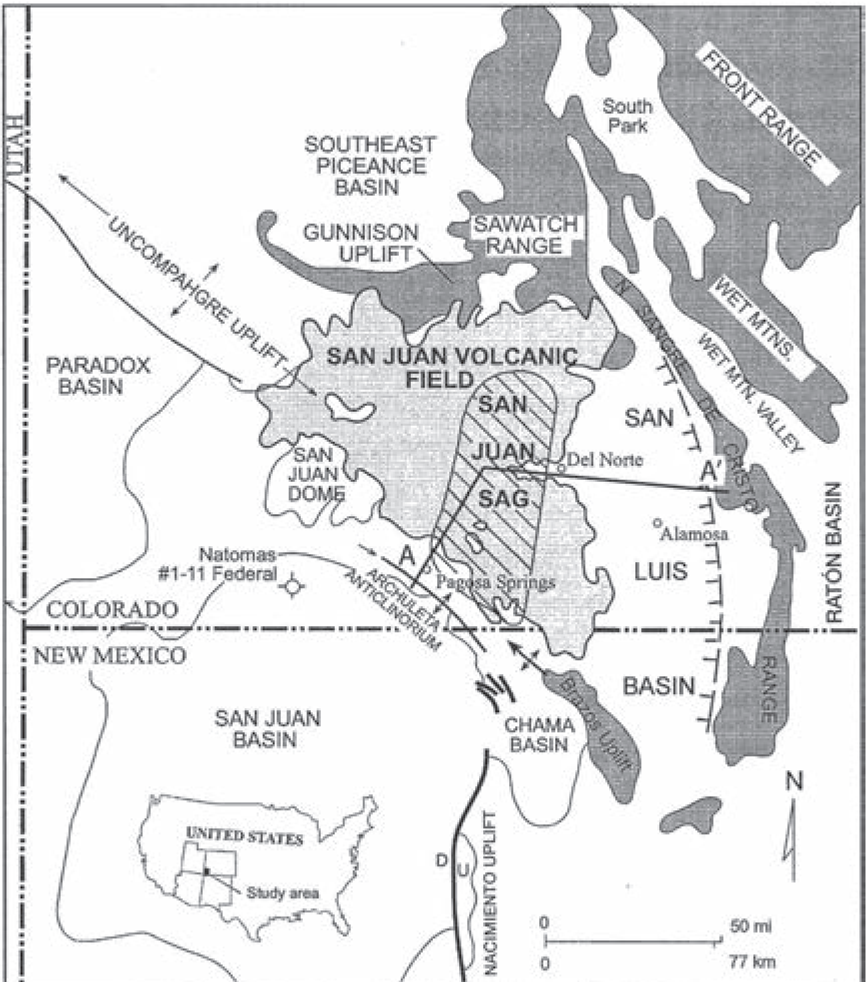


**FIGURE 1.6.** *The Laramide-age San Juan Sag (hachured area) is concealed beneath the thick sequence of volcanic rocks in the San Juan Volcanic Field. The Archuleta Anticlinorium separates the San Juan Basin from the San Juan Sag. The Del Norte High (DN High) separates the Monte Vista Basin (not labeled) from the sag. The Del Norte High and Monte Vista Basin are concealed beneath volcanic and sedimentary rocks. Cross-section A-A' is shown in figure 1.7. Modified from Gries et al. (1997).*

the thick pile of volcanic and volcaniclastic rock between Del Norte and Pagosa Springs, contains Jurassic through Eocene sediments (Brister and Chapin 1994).

Figure 1.8 is a series of cross-sections that show the evolution of the San Juan Sag. Prior to the Western Interior Seaway, late Paleozoic strata and the Triassic

Chinle Formation progressively on-lapped the western side of the late Paleozoic Uncompahgre–San Luis Uplift, and Jurassic-age Morrison strata spread across the uplift. As the Western Interior Seaway transgressed and regressed across the region, these older rocks were buried by thousands of meters of Cretaceous sediment. Intrusions and volcanism during the Late Cretaceous then formed the San Juan Uplift. Sediment eroded from it was transported southeast, depositing the initial Laramide fill in the sag (lower member of the Ánimas Formation). Regional east-northeast compression during the Paleocene created the San Luis, Sangre de Cristo, and Brazos Uplifts, and sediment eroded from them comprises the upper member of the Ánimas Formation. A third pulse of regional uplift during late Paleocene and Eocene time, this one caused by more northeasterly compression, exposed the Proterozoic core of the uplifts. Arkosic sediment shed from the uplifts comprises the Blanco Basin Formation. Tectonism during this late orogenic (mountain-building) phase developed the Del Norte High. Associated wrench faulting created the Monte Vista Basin between the Del Norte High and the San Luis Uplift, and the San Juan Sag underwent additional deformation (Brister and Chapin 1994; Gries et al. 1997).

Configuration of the San Luis Uplift in the eastern part of the Valley is more complex than shown in figure 1.5. For example, Upson (1941) described red-bed strata in the Culebra Reentrant that he called the Vallejo Formation and interpreted as Laramide sediment, suggesting that this part of the Valley was also a Laramide depositional basin. However, mapping by Kirkham and colleagues (2004) later demonstrated that the strata at Upson's type area at the forks of Vallejos Creek are Miocene Santa Fe Group, not Eocene Vallejo. Wallace (1996) and Wallace and Soulliere (1996) mapped small outcrops of the Vallejo farther north in the Culebra Reentrant. Thompson and colleagues (2015) questioned whether these were Eocene Vallejo strata and instead assigned them an Oligocene(?) age. More work is needed to assess whether Vallejo strata actually exist.

Another complexity involved small windows of Cretaceous and Jurassic rocks exposed in arroyos between Crestone and the Great Sand Dunes discovered during a gold exploration project (Watkins 1996). Core obtained from exploration test holes contained traces of Cretaceous oil. Additional deep test holes drilled farther out onto the valley floor found more evidence of Mesozoic strata with traces of oil. Were these Mesozoic rocks preserved in a structurally depressed basin adjacent to the Laramide San Luis Uplift? Or were the Mesozoic rocks part of an ancient landslide block that somehow slid off the mountains and ended up at its present location? Yet another complexity is seen at the San Luis Hills, where geophysical studies suggest that Paleozoic and Mesozoic strata may exist in the subsurface (Drenth et al. 2013).

The Valley's Mesozoic history ended at ~66 Ma when a second great mass extinction event devastated life on Earth. The Cretaceous-Tertiary extinction event (aka

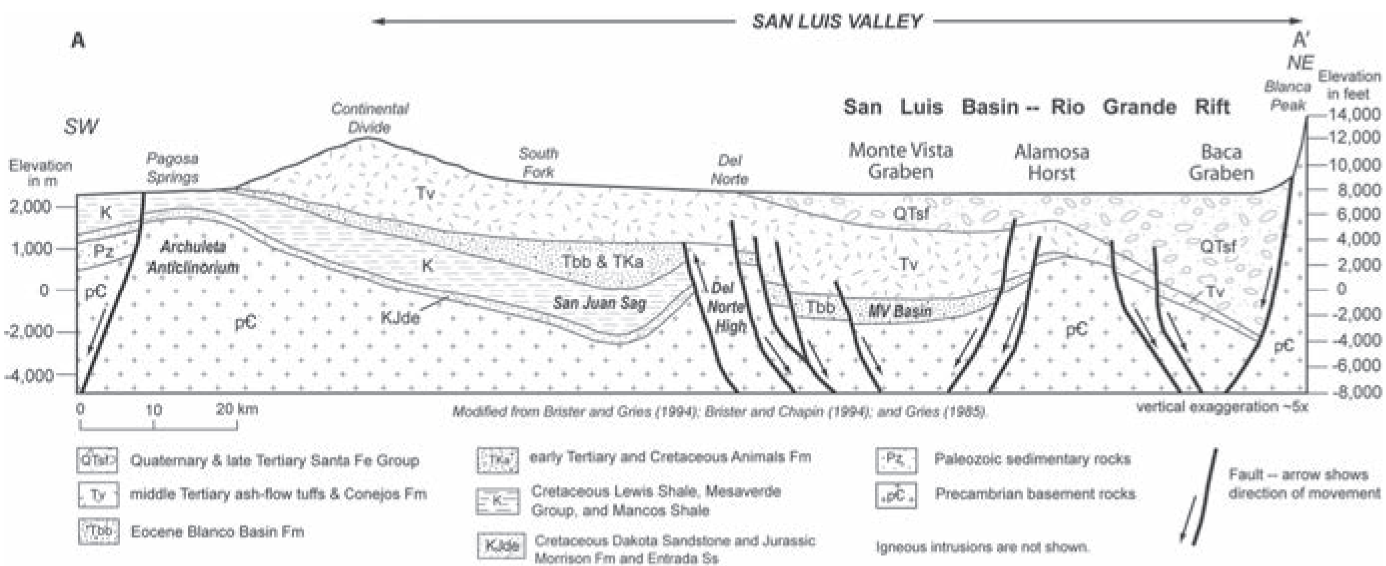


**FIGURE 1.7.** *Simplified regional cross-section that starts near Pagosa Springs, then goes through South Fork and Del Norte and across the floor of the San Luis Valley to Blanca Peak. Laramide structures (labeled with italic font) include the San Juan Basin, Archuleta Anticlinorium, San Juan Sag, Del Norte High, and Monte Vista Basin (labeled MV Basin). Middle Tertiary volcanic rocks of the San Juan Volcanic Field overlie and conceal the San Juan Sag and Del Norte High. The San Luis Basin, Alamosa Horst, and Baca Graben are rift structures. Monte Vista Graben is a Laramide structure slightly modified by rifting. See figure 1.6 for cross-section location.*

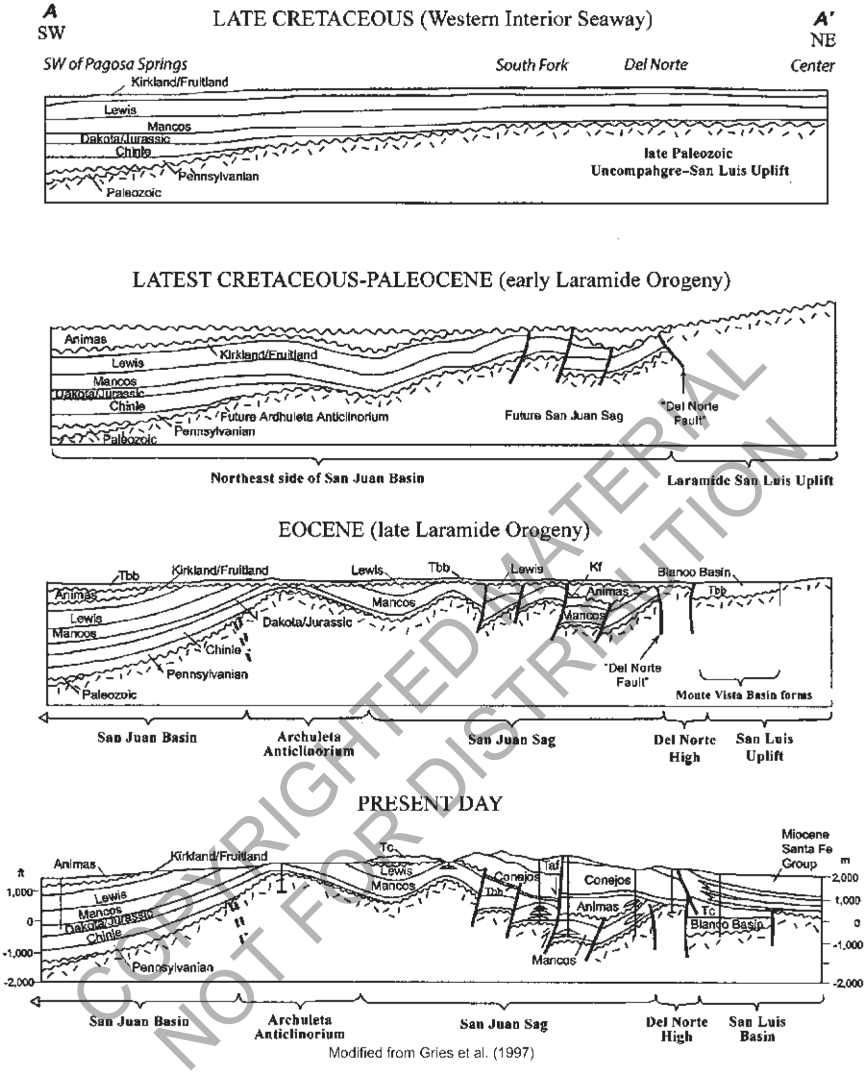


**FIGURE 1.8.** *Schematic cross-sections that illustrate the evolution of the San Juan Sag and nearby areas from Late Cretaceous to present day. The line of section extends from the San Juan Basin southwest of Pagosa Springs generally northeastward to near South Fork and Del Norte, ending in the San Luis Basin near the town of Center. Modified from Gries et al. (1997). Refer to Gries et al. (1997, figure 4) for exact line of section and well control used to construct the cross-sections.*

Cretaceous-Paleogene extinction event) is attributed to the impact of a large asteroid, although massive volcanic eruptions in India possibly contributed to it.

During and immediately following late stages of the Laramide Orogeny, a widespread erosion surface was cut across the SLV and the surrounding region (Epis and Chapin 1975). The Laramide uplifts were beveled, setting the stage for one of the major geologic events responsible for the Valley's modern landscape.

## SOUTHERN ROCKY MOUNTAIN VOLCANIC FIELD: MIDDLE CENOZOIC TIME

Eruptions from numerous volcanoes formed the composite Southern Rocky Mountain Volcanic Field during the middle Cenozoic (Steven 1975; McIntosh and Chapin 2004; Lipman 2000, 2007; Lipman and McIntosh 2011). The volcanic field extended from the Great Plains far into western Colorado and from north-central New Mexico northward beyond South Park (figure 1.9). It was constructed on the low-relief erosion surface carved across the Laramide uplifts and basins.

Volcanism started in the northern end of the volcanic field in what today is the Sawatch Range and generally migrated southward into the SLV and nearby areas. The San Juan Mountains include the largest remnant of the Southern Rocky Mountain Volcanic Field. Other remnants within the Valley crop out in the Latir volcanic area above Questa and in the San Luis Hills. Mid-Tertiary volcanic rocks also exist in the subsurface north of the San Luis Hills (Brister and McIntosh 2004). As a result of glaciation, structural tilting, and erosion and incision by rivers, spectacular exposures of these volcanoes are found in the San Juan Mountains, ranging from near pristine volcano morphology to the upper levels of sub-volcanic intrusions.

Middle Cenozoic volcanic activity initiated in the Valley near the end of the Eocene at ~35 Ma and continued through latest Oligocene time until ~25.1 Ma. The volcanism included stratovolcanoes whose lavas are intermediate in silica content—similar to those found today in the Pacific Northwest—and huge, explosive, silica-rich calderas commonly called supervolcanoes. When a supervolcano erupts, it produces a fast-moving, ground-hugging cloud of volcanic ash, blocks, and gases that spreads far across the landscape. When solidified, the erupted cloud of material forms a pyroclastic rock called ignimbrite or ash-flow tuff.

Collapsed calderas form when a large magma chamber below a volcano suddenly empties during an eruption and overlying rock layers fall into the void. A steeply dipping ring fault zone usually forms the structural boundary of a collapsed caldera. Evidence suggests that the calderas are underlain by large granitic intrusions that were probably the magmatic source of the calderas (Lipman and Bachmann 2015; Drenth et al. 2012). Continuing eruptions commonly filled the collapsed calderas with ignimbrite, which may be substantial in volume (Lipman and McIntosh 2011).

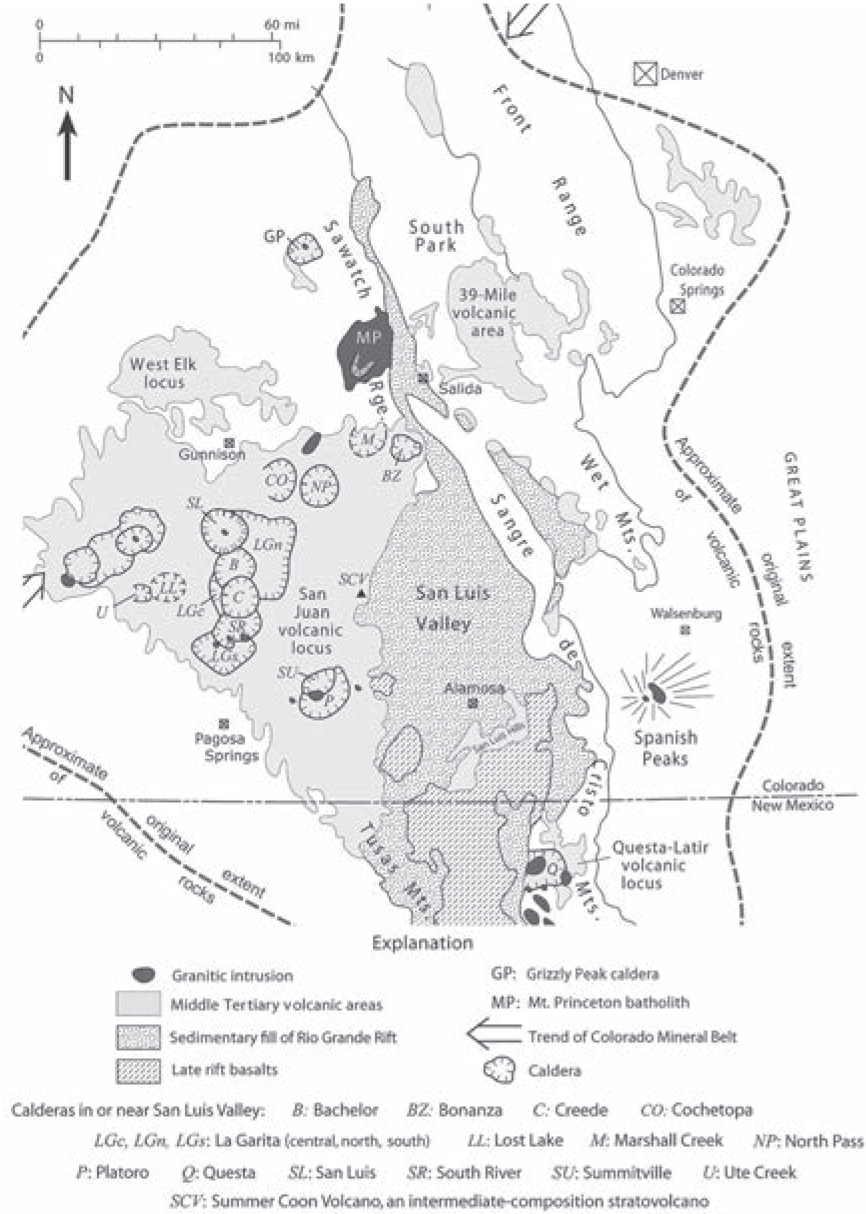


**FIGURE 1.9.** *Map of the Southern Rocky Mountain Volcanic Field showing the inferred original extent of the volcanic field, major erosional remnants, ignimbrite calderas, caldera-related igneous intrusions, and Summer Coon Volcano. Calderas within and adjacent to San Luis Valley are labeled. Modified from Lipman and McIntosh (2011) and Lipman and Bachmann (2015).*

Stratovolcanoes were initially the dominant type of volcanism. Thick aprons of andesitic lava flows and volcaniclastic rocks accumulated around the stratovolcanoes. In the Valley these rocks are called the Conejos Formation (Larson and Cross 1956; Steven and Lipman 1976). It comprises ~25,000 $km^3$ of material, which amounts to almost two-thirds of the total volume of volcanic rocks in the San Juans (Lipman and Bachmann 2015). The igneous dikes, intrusive plugs, lava flows, and volcaniclastic rocks associated with Summer Coon Volcano are spectacular eroded remnants of an early-phase stratovolcano (see figure 1.9 for location). Dikes at Summer Coon Volcano yielded K-Ar ages of 32.4 and 34.7 Ma (Lipman 1976). Conejos dikes, flows, and intrusions in the San Luis Hills have $^{40}Ar/^{39}Ar$ ages ranging from ~27.6 to 30.5 Ma (Thompson et al. 2015).

Thick sections of east-dipping Conejos volcanics also exist beneath the valley floor north of the San Luis Hills. They thin eastward. Based on drill-hole data, the Conejos is 1,047 m below the ground surface and is 1,318 m thick near the town of Center, whereas farther east, near San Luis Lake, the Conejos is 2,112 m deep and only 133 m thick (Brister and McIntosh 2004).

The volcanic field included many silica-rich supervolcanoes, fourteen of which are located in or adjacent to the Valley. Table 1.1 lists the names and areal extent of those calderas, as well as the names, ages, and volumes of ignimbrites erupted from each caldera. Some calderas produced more than one ignimbrite. Explosive silica-rich volcanism started in the Valley at 33.9 Ma with eruption of the Thorn Ranch ignimbrite from the Marshall Creek caldera. Ignimbrite eruptions continued to be focused in the northern part of the Valley, with eruption of the Bonanza ignimbrite from the Bonanza caldera (aka Bonanza–Gribbles Park caldera) at 33.25 Ma and the Saguache Creek ignimbrite from the North Pass caldera at 32.15 Ma.

After a hiatus of over 2 million years, explosive volcanism migrated to the southeastern San Juans. By this time eruptions from stratovolcanoes had diminished. Multiple ignimbrite sheets were erupted from the Platoro and Summitville calderas from 30.1 to 28.6 Ma. They include the Black Mountain, La Jara Canyon, Ojito Creek, Ra Jadero, South Fork, and Chiquito Peak ignimbrites. As activity at the Platoro and Summitville calderas waned, explosive activity began in the central San Juan area with the eruption of the Masonic Park ignimbrite at 28.7 Ma, soon followed by the 28.6 Ma Ute Ridge ignimbrite from the Ute Creek caldera and the 28.5 Ma Blue Creek ignimbrite from the Lost Lakes caldera.

No supervolcanoes were active in the Valley for the next 500,000 years, but the central San Juan area soon flared up again, starting with the tremendous eruption from the La Garita caldera at 28.02 Ma. This huge collapsed caldera was 35 km wide and 75 km long, extending from west of Wolf Creek Pass northeastward to Saguache Park (Lipman 2006; Lipman and Bachmann 2015). La Garita caldera produced the largest known volcanic eruption in the world, the ~5,000 $km^3$ of ignimbrite

**TABLE 1.1.** Names and areas of calderas in and adjacent to the San Luis Valley and the names, ages, and volume of ignimbrites erupted from them.

| | | *Ignimbrite* | | *Caldera* | |
|---|---|---|---|---|---|
| *Location* | *Name* | *Age, Ma* | *Vol., km$^3$* | *Name* | *Area, km* |
| Taos Range | Amalia | 25.1 | 500 | Questa | 14 × >15 |
| Central San Juan | Snowshoe Mountain | 26.85 | 500 | Creede | 20 × 25 |
| | Nelson Mountain | 26.90 | 500 | San Luis–Cochetopa | 9 × 9; 20 × 25 |
| | Cebolla Creek | 26.90 | 300 | San Luis complex | 14 × 16 |
| | Rat Creek | 26.90 | 150 | San Luis complex | 9 × 12 |
| | Wason Park | 27.35 | 500 | South River | 20 × 20 |
| | Blue Creek | 27.40 | 250 | [concealed] | ? |
| | Carpenter Ridge | 27.55 | 1,000 | Bachelor | 25 × 30 |
| | Fish Canyon | 28.02 | 5,000 | La Garita | 35 × 75 |
| West San Juan | Blue Mesa | 28.50 | 350 | Lost Lakes (buried) | 10 × 10 |
| | Ute Ridge | 28.60 | 350 | Ute Creek | 8 × 8 |
| Southeast San Juan | Chiquito Peak | 28.60 | 500 | Platoro | 18 × 22 |
| Central San Juan | Masonic Park | 28.70 | 500 | [concealed] | ? |
| Southeast San Juan | South Fork | 28.80 | 75 | Platoro/Summitville? | 8 × 12? |
| (Treasure Mtn Group) | Ra Jadero | 28.80 | 150 | Summitville? | 8 × 12 |
| | Ojito Creek | not dated | 100 | Summitville? | 8 × 12 |
| | La Jara Canyon | 29.90 | 1,000 | Platoro | 20 × 24 |
| | Black Mountain | 30.10 | 350 | Platoro | 20 × 24 |
| Northeast San Juan | Saguache Creek | 32.25 | 350 | North Pass | 15 × 17 |
| Southern Sawatch Range | Bonanza | 33.15 | 1,000 | Bonanza | 15 × 20 |
| | Thorn Ranch | 33.9 | 500 | Marshall Creek | 15 × 15? |

Data from Lipman and Bachmann (2015).



called the Fish Canyon Tuff. Plate 6 illustrates the tremendous size of the La Garita eruption relative to large, historic, worldwide eruptions and to well-documented, geologically recent eruptions in the United States.

The Fish Canyon Tuff spread far and wide. It forms the popular climbing cliffs at Penitente Canyon Recreation Area north of Del Norte (Elison, this volume). A short distance east, because of its eastward dip, the ignimbrite disappears into the subsurface beneath the valley floor. The Fish Canyon Tuff maintains a relatively uniform thickness in the subsurface beneath much of the valley floor, ranging from 29 m to 51 m thick where penetrated by deep test holes. Its maximum known thickness is 91 m near the base of the Sangre de Cristo Mountains east of San Luis Lakes (Brister and McIntosh 2004), suggesting that some type of topographic barrier or basin may have existed there at that time.

Subsequent caldera eruptions in the central San Juan area were mostly nested within the larger La Garita caldera, which destroyed some of the evidence of that caldera. These younger supervolcanoes include the Bachelor, South River, San Luis, and Creede calderas, which erupted during a relatively short time frame from 27.55 to 26.85 Ma. These calderas produced the Carpenter Ridge, Blue Creek, Wason Park, Rat Creek, Cebolla Creek, and Nelson Mountain ignimbrites and culminated with the Snowshoe Mountain Tuff erupted from the Creede caldera.

Topographic expression of the Creede caldera continues to be impressive. Cliffs near the north end of Creede mark the collapsed caldera's outer wall, which was locally modified by secondary landslides. Snowshoe Mountain, the prominent mountain south of Creede, is a resurgent dome that grew in the middle of the caldera. The arcuate valley of the Río Grande from Wagon Wheel Gap to Antelope Park contains tuffaceous lake sediment that was deposited in a water-filled moat between the caldera's wall and resurgent dome.

Questa caldera, located within the Questa-Latir volcanic locus, is the only supervolcano in the Sangres (Lipman and Reed 1989). It produced the youngest ignimbrite in the Valley, the ~25.1 Ma Amalia Tuff.

Most major volcanic fields form along plate boundaries. However, the Southern Rocky Mountain Volcanic Field, as well as many other mid-Cenozoic volcanic fields in the Rockies, developed far inland from the subduction zone that existed at the plate boundary on the west side of North America at that time. A flattening of the subduction zone or other changes in its geometry may have caused the Valley's mid-Cenozoic volcanism (Lipman and McIntosh 2011). Deep-seated magmatism associated with the volcanic field perhaps also caused regional domal uplift that contributed to the modern high topography in the Southern Rocky Mountains (Roy et al. 2004; Gonzales and Karlstrom 2011).



## RÍO GRANDE RIFT: LATE CENOZOIC TO TODAY

As volcanism in the Southern Rocky Mountain Volcanic Field waned, east-west–directed regional extension began to pull apart the Earth's crust, forming the structural and physiographic Río Grande rift. Extension eventually broke the crust into a series of north-south–elongated structural basins that filled with sediments locally interbedded with volcanic flows. Fractures and faults penetrated the extending crust and allowed molten mantle material to migrate upward, feeding numerous volcanoes. Volcanism evolved into a bimodal assemblage of mostly low-silica basalt and lesser volumes of high-silica dacite and rhyolite. The rift branches off the extensional Basin and Range Province near the Mexican border and extends northward through west Texas, New Mexico, and Colorado.

Rift boundaries can be defined by physiography, major rift faults, and extent of rift-related sediments and volcanics. Figure 1.10 outlines our assigned boundaries of major structural basins within the rift. One of the largest rift basins, the San Luis Basin, lies within the SLV. Rift valleys are relatively easy to recognize in Colorado and northern New Mexico but difficult to delineate in the south, where the rift overlaps with extensional valleys within the Basin and Range Province.

Bryan (1938) first described the rift. He called it the Río Grande Depression and considered the SLV the northernmost basin. Kelley (1952, 1954, 1956) documented many of the master rift faults in New Mexico and proposed that deep-seated rifting was responsible for the Río Grande Depression. In the 1960s the name Río Grande rift began to appear in the literature, and studies of the rift increased greatly as geologic mapping, particularly of volcanic fields, became widespread. Plate tectonic theory also piqued interest in major structures within the continental interior. Soon, many were studying the rift in and near the Valley (e.g., Gaca and Karig 1965; Chapin 1971, 1979; Butler 1971; Knepper 1976; Lipman and Mehnert 1975, 1979; Tweto 1978a, 1978b, 1979b; Baltz 1978; Kirkham and Rogers 1981; Burroughs 1981; McCalpin 1982; Keller et al. 1984; Kelley and Duncan 1984). Based on the physiographic tapering and narrowing of basins from south to north, the rift was initially thought to have first opened in the south and gradually migrated northward. Recent work suggests synchronous opening of the rift throughout its length (Ricketts et al. 2015).

Major rifting in the SLV began ~26–25 Ma, based on the oldest rift-related volcanic rocks in the Valley (Thompson et al. 2015) and on an angular unconformity in the San Juan Mountains where early rift basaltic rocks rest on middle Cenozoic ignimbrites (Lipman and Mehnert 1975). Lipman and Zimmerer (2016) described evidence of weak extension starting at about the same time, based on $^{40}Ar/^{39}Ar$ ages of 24.94 and 25.64 Ma on basaltic Dulce dikes west of the Valley and a maximum age of 26.53 Ma on basaltic dikes at Platoro. Others suggest that extension began elsewhere in the rift as early as ~35–30 Ma (Chapin and Cather 1994; Smith 2004; Brister and McIntosh 2004; Cather et al. 2006). Rifting continues today, based on GPS studies that document modern extension (Berglund et al. 2012; Murray 2015) and on Holocene fault ruptures (Kirkham and Rogers 1981; McCalpin 1982, 1996; Menges 1990; Crone et al 2006; Ruleman and Machette 2007; McCalpin, this volume).

Late Cenozoic rift-related faulting and bimodal volcanism occurred far beyond the major rift basins, indicating that extensional forces affected much more of the

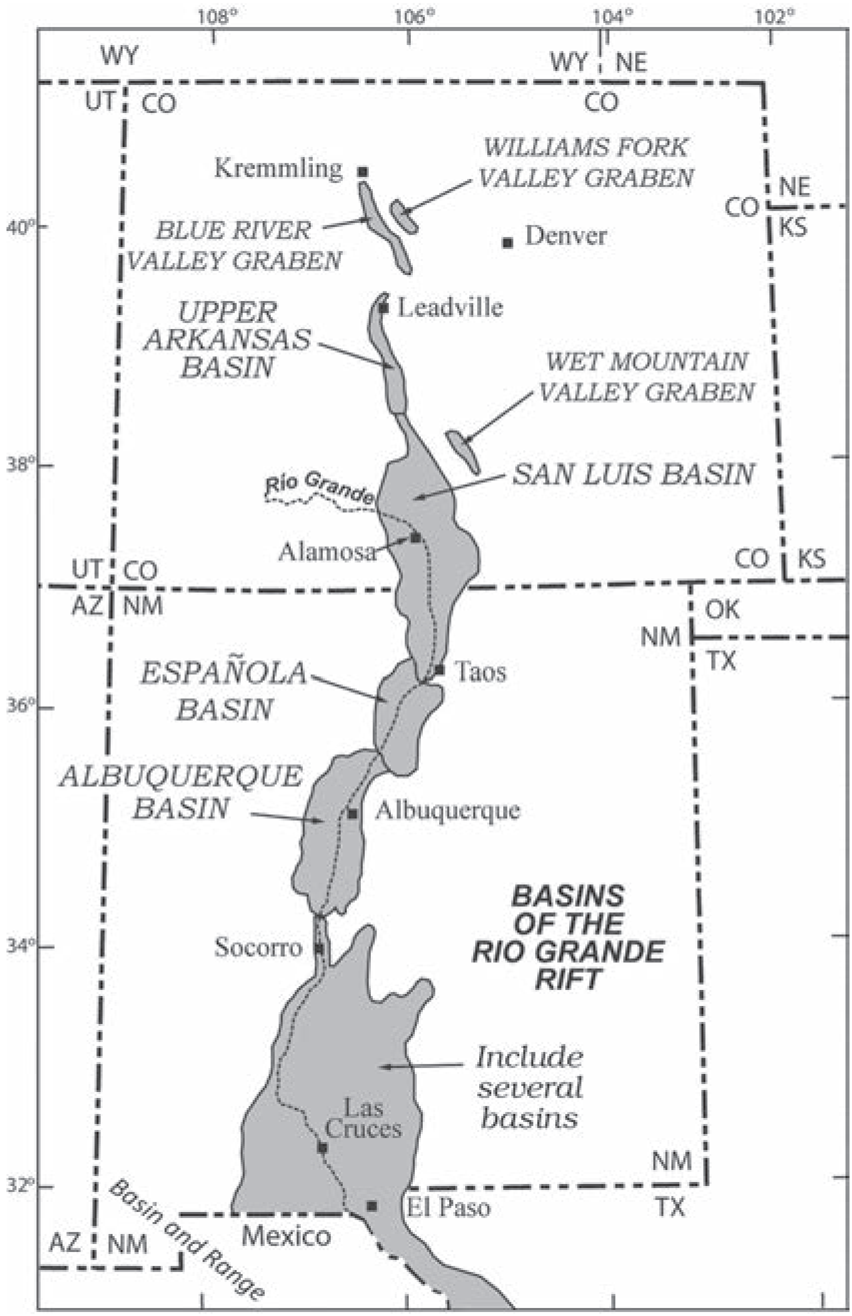


**FIGURE 1.10.** *Major structural basins of the Río Grande Rift*

crust than just the major rift basins shown in figure 1.10. Continual measurements of permanent GPS sites document that a large part of the southwestern United States is moving west but that areas east of the Valley move westward more slowly and lag behind the west side. This differential crustal pull-apart strain from the Colorado Plateau to the Great Plains created the rift. Based on a relatively short period of record and a series of east-west–oriented transects from northern Colorado to southern New Mexico, the current extensional strain rate from the Great Plains to the Colorado Plateau averages ~1.2 mm/year (Berglund et al. 2012). The highest current strain rate across the rift occurs along an east-west transect that includes the Valley (Murray 2015).

From a regional perspective, the San Luis Basin is an asymmetric, east-tilted half-graben. A graben is a down-dropped block of rock bounded on both sides by high-angle normal faults that dip toward the graben. A half-graben has a fault only on one side. The master fault of the San Luis Basin is the Sangre de Cristo fault, a north-south–striking, down-to-west normal fault located at the western base of the Sangres. The Sangres are an uplifted block called a horst. Other faults concealed beneath the valley floor were active earlier during the rift's history; they accommodated much of the early extension.

The Sangre de Cristo fault consists of three major sections. A northern section extends from near Poncha Pass to the southern side of the Blanca Massif, where the range abruptly swings eastward. Geophysical and drill-hole data indicate that faults concealed beneath the northern valley floor that subparallel the northern fault section also accommodate some of the half-graben's structural relief in this area (Watkins 1996; Grauch et al. 2012, 2013; Drenth et al. 2017). The central fault section starts near Highway 160 and ends at about the state line. Rather than being at the range front, the central section borders foothills that gradually rise from the valley floor. From the latitude of San Luis to the state line, the central and southern fault sections overlap. The southern section forms the west side of San Pedro Mesa and continues southward along the base of the Taos Range to beyond the town of Taos.

Also shown on plate 2 are three other important rift faults with late Quaternary activity: the Villa Grove fault, a complex swarm of mostly southwest-facing scarps in young alluvium that cuts obliquely across the northern end of the basin; the Mesita fault, which offsets the ~1 Ma Mesita Volcano; and the northeast-oriented Embudo fault that forms the southern end of the basin. Many other rift faults exist elsewhere in the basin, although older faults are sometimes buried by younger deposits and have no surface expression. Young rift faults are relatively rare along the hingeline on the western margin of the basin. Documented examples include a swarm of faults in the foothills west of Monte Vista and faults in the narrow northernmost end of the Valley (Lipman 1976; Kirkham and Rogers 1981; Widmann et al. 1998; Grauch and Ruleman 2013).

In contrast to east-tilted San Luis Basin, adjacent rift basins to the north and south are west-tilted half-grabens with master faults on their west sides. Kellogg (1999) attributed the asymmetry of the San Luis Basin and the opposite tilt in the Upper Arkansas Basin to inheritance from Laramide and older structural zones because master rift faults in both basins coincide with and dip in the same direction as Laramide faults. Accommodation zones allow for the structural reversals and differential subsidence between adjacent basins (Chapin and Cather 1994). At the south end of the San Luis Basin, the oblique-slip Embudo fault accommodates transfer to the Española Basin (Chapin and Cather 1994; Bauer and Kelson 2004a). Grauch and colleagues (2017) call this structure the Embudo transfer zone. The Poncha Pass block serves as the accommodation zone between the San Luis Basin and the Upper Arkansas Basin (Minor et al. 2019).

Superimposed on the regional half-graben of the San Luis Basin are second-order structural features. The basin includes three distinct structural lows separated by the centrally located San Luis Hills Horst (figure 1.11). The Conejos Formation and overlying early rift basalt flows called the Hinsdale Formation are uplifted and exposed in the horst. The adjoining northern, southern, and eastern sub-basins are structurally down-dropped hundreds to thousands of meters relative to the horst. Each sub-basin contains smaller yet still major third-order structures.

As the San Luis Basin developed, it filled with sediment and volcanic rocks, most of which are lumped into the Santa Fe Group. In the southwest part of the Valley the Santa Fe includes the volcaniclastic-rich Los Pinos Formation. In the southern sub-basin, in and adjacent to the Picurís Mountains, the Santa Fe Group includes the Tesuque and Chamita Formations and the upper part of the Picurís Formation (Aby et al. 2004, 2011; Bauer and Kelson 2004b). Since the rift is still active, overlying Quaternary deposits are considered by some geologists to be syn-rift.

Some of the best exposures of the Santa Fe Group are in the eastern sub-basin, where sediments range from arkosic to volcaniclastic, both laterally and vertically, depending on sediment source areas. Volcanic flows ranging in age from 15.2 to 10.7 Ma are interlayered with Santa Fe sediments in the eastern sub-basin (Miggins 2002; Kirkham et al. 2003, 2004, 2005a, 2005b; Thompson et al. 2015). Deposits of volcanic ash erupted from distant volcanoes are also found within the Santa Fe Group in the eastern sub-basin. They include an 11.9 Ma ash north of Ojito Creek and an 11.3 Ma deposit at the forks of Vallejos Creek (Kirkham and Heimsoth 2003; Kirkham et al. 2004). These serve as excellent local marker beds but unfortunately cannot be traced long distances, in part because of abundant rift faults that shatter this area into a confounding complex of jostled blocks (figure 1.12).

Unconformities can also complicate correlation of Santa Fe strata. Brister and Gries (1994) reported a major angular unconformity in the northern sub-basin based on interpretation of seismic data. In the eastern sub-basin, Kirkham and

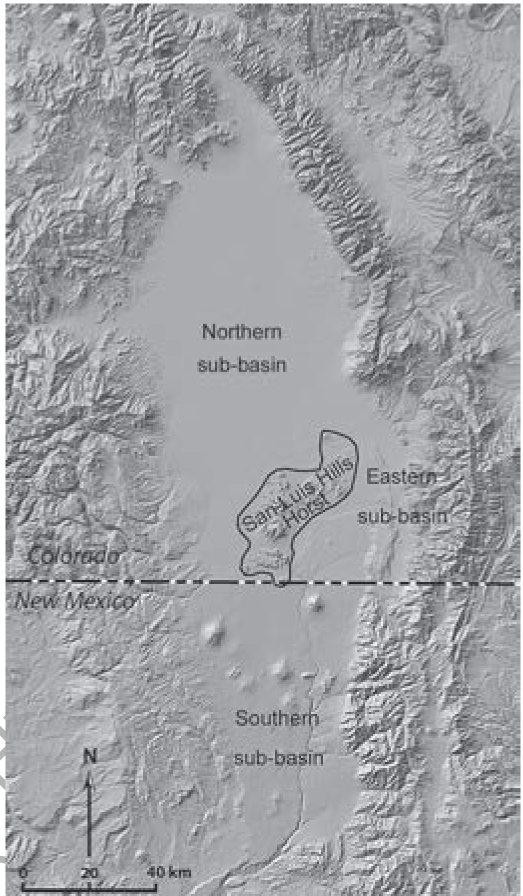


**FIGURE 1.11.** *Structural sub-basins of the San Luis Basin.*



Heimsoth (2003) postulated an angular unconformity north of Ojito Creek because of changes in dip direction between Pliocene Servilleta Basalt and underlying sediments, as well as the presence of only ~76 m of sediment between the ~3.66 Ma basalt caprock and ~11.9 Ma volcanic ash in underlying sediments.

Brister and Gries (2004) used subsurface data to subdivide the Santa Fe Group in the northern sub-basin into two units: a thick lower member containing variegated sandstone, claystone, and conglomerate and the overlying, much thinner Alamosa Formation, which was first recognized and named by Siebenthal (1910). The Alamosa Formation, or "Blue Clay" as it is known locally, consists of Pliocene to middle Pleistocene clay, silt, and fine sand deposited in and adjacent to ancient Lake Alamosa. The lake is thought to have formed about 3.5 Ma (Machette et al. 2007a; Ruleman et al. 2016). The formation is thickest and composed chiefly of clay

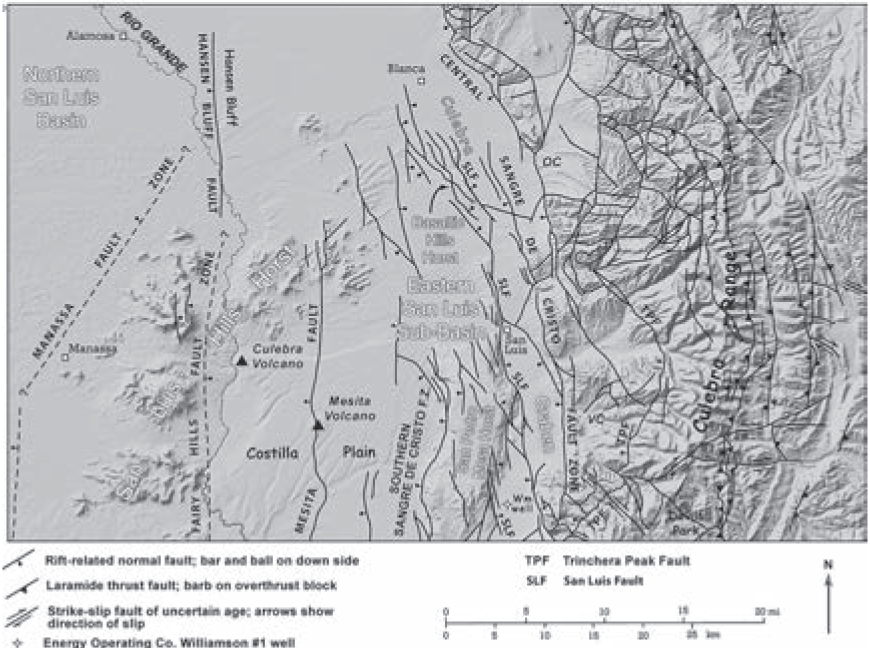


**FIGURE 1.12.** *Simplified map of the complex fault pattern in the Alamosa 30° × 60° quadrangle. OC, Ojito Creek; VC, Vallejo Creek. Modified from Thompson et al. (2015).*

and silt in the structurally deepest part of the northern sub-basin. In a cross-section by Brister and Gries (1994), the formation is about 460 m thick in the Cougar Petroleum Company Crow #1 test well near San Luis Lakes.

In contrast to the absence of rift volcanism in the northern sub-basin and minor localized volcanism in the eastern sub-basin, volcanism was widespread in the southern basin, creating the Taos Plateau Volcanic Field (figure 1.13). At least fifty-five eruptive centers exist in this volcanic field (Lipman and Mehnert 1979; Thompson et al. 2014). Pliocene Servilleta Basalt and related basalts comprise a majority of the volcanic rocks in the plateau, although silica-rich volcanoes rise locally above the plateau's tableland. Some, like San Antonio Mountain and Ute Mountain, are very large and extend several hundred meters above surrounding tablelands.

A thick sequence of Servilleta Basalt flows is exposed at the confluence of the Río Grande and the Red River, but they thin and pinch out eastward. The sequence of Servilleta flows is also thick at the Gorge Bridge on Highway 64 (aka High Bridge). Three thick packages of Servilleta lava flows separated by interbedded sediments are exposed at the Gorge Bridge. $^{40}Ar/^{39}Ar$ ages of the lower flow package (LB on plate 7) range from 4.78 to 4.50 Ma, the middle flow package (MB) is 4.11 to 4.02 Ma, and the upper flow package (UB) is 3.69 to 3.59 Ma (Cosca et al. 2014). Servilleta flows exposed at Dunn Bridge are as young as 2.97 Ma (Appelt 1998). Los Mogotes is a prominent ~4.8 Ma Servilleta-related basaltic shield volcano located in the foothills northwest of Antonito. The youngest volcano in the Taos Plateau Volcanic Field is 1.06 Ma Mesita Volcano (Thompson et al. 2015).

Several rift-related volcanoes existed on the Taos Plateau prior to eruptions of Servilleta Basalt. They include Cerro Montoso, Guadalupe Mountain, Cerro Chiflo, and Cerro de la Olla. Servilleta lavas gradually engulfed the flanks of these older volcanoes, and only the upper edifices of them now extend above the Servilleta-capped tablelands (Bauer et al. 2015). Young, post-Servilleta volcanoes, such as San Antonio Mountain, Ute Mountain, and Mesita Volcano, were constructed on top of the Servilleta flows or on Santa Fe sediments overlying the Servilleta.



Geophysical data and deep drill holes provide clues about the subsurface structure of the sub-basins. Gravity lows result from density contrasts between the sedimentary fill and Proterozoic basement rock. Plate 8 compares the modern topography of the Valley with a basin model prepared by V.J.S. Grauch based on gravity data in Keller and colleagues (1984). The model depicts what the ground surface would look like if all rift fill were removed. Intense blue colors denote the deepest parts of the basin. Aeromagnetic anomalies can identify subsurface features including buried faults and edges of buried volcanic flows (e.g., Grauch et al. 2012, 2013, 2017). Published seismic reflection lines, mostly done for oil exploration, aid subsurface interpretations in the northern sub-basin, but none are available for the eastern and southern sub-basins. Several deep drill holes are scattered across the northern sub-basin, but only one drill hole extends to Proterozoic rocks in the eastern sub-basin and none do in the southern sub-basin. Available data clearly demonstrate the existence of three deep, narrow, north-south–trending grabens in the eastern parts of all three sub-basins: the Baca Graben in the northern sub-basin, Culebra Graben in the eastern sub-basin, and Taos Graben in the southern sub-basin.

The northeast-trending Manassa fault separates the northern sub-basin from the San Luis Hills Horst (figure 1.12). Originally thought to extend across the entire valley floor from near Antonito to the Blanca Massif (e.g., Tweto 1978a, 1979a; Burroughs 1981), the Manassa fault may terminate against the Hansen Bluff fault in the Alamosa National Wildlife Refuge (Drenth et al. 2011; Thompson et al. 2015). In addition to the Baca Graben, the northern sub-basin includes the Monte Vista Graben beneath the western valley floor, which is superimposed on the Laramide Monte Vista Basin, and the buried Alamosa Horst, a concealed northward extension of the San Luis Hills Horst located between the two grabens. Figure 1.7 shows subsurface relationships between these structures.

Using gravity data, Gaca and Karig (1965) reported that the thickness of syn-rift fill in the Baca Graben could be as much as 9,140 m. Recent estimates are significantly

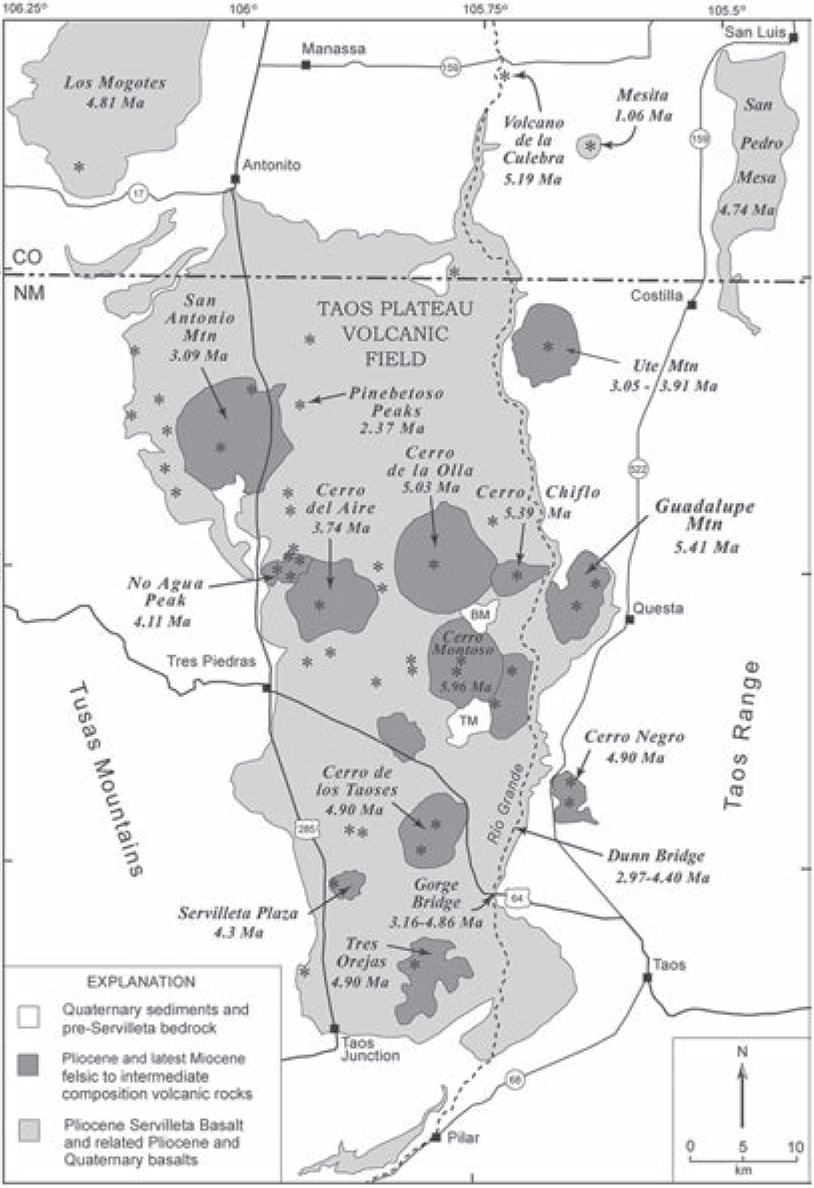


**FIGURE 1.13.** *Map of Pliocene and Quaternary volcanic rock and volcanic vents in the Taos Plateau Volcanic Field. Islands of Oligocene-Miocene volcanic rocks extend above the tableland of the plateau at Brushy Mountain (BM) and Timber Mountain (TM). Geology modified from Grauch and Keller (2004) and Thompson et al. (2015).* $^{40}Ar/^{39}Ar$ *ages from Appelt (1998), Cosca et al. (2014), and Thompson et al. (2015), all recalibrated to an age of 28.204 Ma for the monitor mineral sanidine from the Fish Canyon Tuff.*

less. Keller and colleagues (1984) described the maximum thickness as greater than 5,000 m, Brister and Gries (1994) stated it was as much as 5,600 m, and Kluth and Schaftenaar (1994) suggested 6,400 m. The most recent estimate, based on a new gravity model, describes the maximum thickness as more than 6,000 m (Drenth et al. 2011, 2016). Elevation differences between Proterozoic rocks exposed at the crest of the Sangres and the top of Proterozoic rocks in the Baca Graben indicate that nearly 8,000 m of vertical slip has occurred on faults that bound the east side of the Baca Graben during the history of the rift. Part of this displacement is accommodated by the northern Sangre de Cristo fault, while sub-parallel faults concealed beneath the valley floor also accommodated some of it (Watkins 1996; Grauch et al. 2013; Drenth et al. 2017). Other parts of the northern sub-basin have significantly less fill. The Monte Vista Graben has 1,000 to 1,100 m of rift fill (Drenth et al. 2013; Brister and McIntosh 2004), and rift fill over the Alamosa Horst was only 532 m thick in an oil test well drilled near Mosca.

At about the state line, the southern end of the San Luis Hills Horst disappears below the ground surface, swings southwest, and merges with the Tusas Mountains (Grauch and Keller 2004). Broke Off Mountain Basin, a relatively small, narrow, elongate basin between the buried horst block and the Tusas Mountains, may be a southern extension of the Monte Vista Graben because it also contains Laramide sediment, San Juan volcanics, and rift fill (Drenth et al. 2011).

Regional gravity data suggest that the deepest part of the southern sub-basin is between the Sangre de Cristo Range front and the Río Grande Gorge; there it may exceed 3 km in depth (Cordell 1978; Keller et al. 1984; Grauch and Keller 2004). This indicates that the total rift-related vertical slip on faults that form the east margin of the southern sub-basin is more than 4.9 km. The western part of the southern sub-basin is a relatively shallow, east-tilted bench broken by small-displacement normal faults as it rises toward the Tusas Mountains.

The Taos Graben occupies less than half the width of the southern sub-basin. The graben is bounded on the east by the southern Sangre de Cristo fault. The Gorge fault, which approximately coincides with the Río Grande Gorge, forms the west margin of the Taos Graben (Grauch and Keller 2004; Bauer and Kelson 2004a, 2004b). The Sunshine Valley fault zone defines the western margin of the deepest part of the Taos Graben in the area north of the Red River (Ruleman et al. 2013).

The southern Sangre de Cristo fault and Embudo fault transfer zone, as well as many concealed, now inactive fault splays found in the subsurface as much as several km outboard of the modern range front, form the southern end of the southern sub-basin. In the model of Grauch and colleagues (2017), this part of the rift began to form ~23 Ma by reactivation of strands of the old Picurís-Pecos fault. Rapid extension was under way by 18 Ma, creating depositional basins. Extension was accommodated by different faults at different times, with activity eventually evolving to the active traces of the southern Sangre de Cristo fault and Embudo fault at ~0.8 Ma.

The eastern sub-basin includes the Culebra Graben, the Servilleta-capped horsts of San Pedro Mesa and Basaltic Hills, and a structural low in the Sunshine Valley–Costilla Plains area (plate 8; figure 1.12). In contrast to other sub-basins, the eastern sub-basin includes a broad foothills belt that extends east far beyond the central Sangre de Cristo fault. The best exposures of syn-rift fill are found in these foothills. A complex zone of rift faults with down-to-west displacement forms the east flank of the Culebra Graben and separates it from uplifted Proterozoic and Paleozoic rocks in the Culebra Range. The west side of the Culebra Graben is another complex fault zone, the down-to-east San Luis fault. In 1999, Energy Operating Company drilled the Williamson #1 oil test well in the southern part of the Culebra Graben (figure 1.12). This well encountered nearly 2,000 m of Santa Fe Group sediment and volcanics resting directly on Proterozoic rocks (Kirkham et al. 2005a). Elevation differences between Proterozoic rocks in the well and at the Culebra Range crest indicate over 3,000 m of rift-related vertical displacement across the east flank of the Culebra Graben.

San Pedro Mesa is a horst with large displacement normal faults on both flanks: the southern Sangre de Cristo fault on the west and the complex San Luis fault on the east. The San Luis fault also raises the Basaltic Hills relative to the Culebra Graben, but displacement on the southern Sangre de Cristo fault either dies out or is terminated by a cross fault before reaching the Basaltic Hills. Most of San Pedro Mesa is capped by flat-lying Servilleta Basalt, although a small island of Proterozoic and Pennsylvanian rocks rises slightly above the basalt flows, and Miocene volcanic rocks comprise a large hill in the southeast part of the mesa (Kirkham et al. 2005a; Thompson et al. 2007, 2015).

On the west side of San Pedro Mesa, Servilleta Basalt is lowered about 300 m by the southern Sangre de Cristo fault, based on logs for water wells in the Costilla Plains (Zorich-Erker Engineering 1980; Winograd 1959). Drenth and colleagues (2016) estimated that syn-rift fill beneath the Costilla Plains could be as much as 2,000 m. Syn-rift fill beneath the Costilla Plains is broken locally by several young rift faults, including the Mesita fault, which displaces the 1 Ma Mesita Volcano ~13 m (Machette et al. 2007b). An aeromagnetic anomaly suggests that the Mesita fault may be a significant geologic boundary affecting the lateral extent of the Conejos Formation (Drenth et al. 2013).

The Sunshine Valley–Costilla Plain Graben is bounded on the east by the southern Sangre de Cristo fault and on the west by poorly understood faults that subparallel the Río Grande. West-side faults include the Fairy Hills fault (aka La Sauses fault) in the north (Thompson et al. 2015; Burroughs 1972; Bartlett 1984). In the south, the Gorge fault and other unnamed faults largely concealed beneath Servilleta

Basalt form the west side of the graben (Bauer and Kelson 2004a, 2004b; Ruleman et al. 2013, 2016). Ancient Lake Sunshine is found within New Mexico's part of the Sunshine Valley–Costilla Plain Graben (Winograd 1959; Ruleman et al. 2013, 2016). Although the extent of Lake Sunshine was very small compared to ancient Lake Alamosa, as much as 50 m of interstratified lacustrine clay, silt, and gravel was deposited in it, probably in a playa environment.

Most young rift faults within the Valley are high-angle normal faults. The Sangre de Cristo fault, which is high-angle at and near the ground surface (dip ~60°), may curve and become flat at a depth of about 16 km (Kluth and Schaftenaar 1994). Remnants of west-dipping, low-angle faults found along the range front between Crestone and the Great Sand Dunes and at and near the San Luis Gold Mine northeast of San Luis have been interpreted as early-rift detachment faults (Watkins 1996; Jones 1991; Benson and Jones 1996; Benson 1997). In contrast, Wallace (2004) suggested that the fault northeast of San Luis might be a Laramide fault that was later tilted by high-angle rift faults, and Caine and colleagues (2013) concluded that the low-angle Deadman fault, located between Crestone and the Great Sand Dunes, is a Laramide thrust fault.



## GLACIERS, ANCIENT LAKE ALAMOSA, AND THE GREAT SAND DUNES: QUATERNARY PERIOD

Glacial, alluvial, and eolian processes, accompanied by erosional incision and minor localized volcanism, characterize the Quaternary Period, the most recent chapter of the Valley's geologic history. At the beginning of the Quaternary (~2.6 Ma), the climate began to cool. Since then, Earth has experienced alternating cold glacial periods and warmer interglacial climates. We live in the warm Holocene Epoch, which started about 12,000 years ago. Other than the deposits left behind by the glaciers, the best evidence of these episodic climate changes, and the only evidence for the early part of the Quaternary, is from benthic foraminifera in marine sediment from deep-sea ocean cores and polar-region ice cores.

Blair and Gillam (2011) describe the early glacial periods as being relatively short and doing little to sculpt the mountains and deposit sediment, but they became longer and colder after about ~650,000 years ago (~650 ka). The Pleistocene glaciers then began to carve spectacular cirques, horns, arêtes, and U-shaped valleys in the mountains (plate 4). Terminal, end, lateral, and ground moraines left behind by glaciers contain poorly sorted, boulder-rich deposits of till, chiefly from two recent, strong pulses of glaciation. Bull Lake glaciation reached a maximum at ~130 ka, and the most recent pulse was the Pinedale glaciation, which peaked ~23 to 21 ka, based on ages of glacial deposits in the nearby Upper Arkansas Valley (Schweinsberger et al. 2016). Meltwater from Pleistocene glaciers carried sediment out of the mountains and deposited it in alluvial fans adjacent to the mountains. Some of the tallest alluvial fans in the United States formed at the mouths of valleys that drain Blanca Massif (plate 4). (Refer to Beeton and Johnson [this volume] for more details about the Valley's glacial and alluvial fan history.)

Meltwater from the glaciers also contributed water to ancient Lake Alamosa, depositing lacustrine and fluvial sediment that became the Alamosa Formation. The extent of Lake Alamosa shown on plate 1 is defined by the altitude of the highest shoreline deposits in the San Luis Hills (Machette et al. 2007a). At its maximum extent, Lake Alamosa was one of the largest high-altitude lakes in North America (Machette et al. 2013; Ruleman et al. 2016). The lake expanded and contracted depending on surface-water inflows and evaporation. As the lake shrunk, fluvial and perhaps eolian deposits were deposited over older clay-rich lake sediments, only to be buried by more clay-rich sediment when the lake grew, thus creating the interbedded lacustrine and non-lacustrine sequence of strata that characterizes much of the Alamosa Formation. The Alamosa Formation serves as the confining layer for flowing artesian wells in the northern San Luis Basin (Harmon, this volume).

Over time, the lake filled with sediment, and when it eventually overtopped it was probably not very deep. Initial age dates on the highest shoreline deposits suggested that overtopping occurred at ~440 to 430 ka (Machette et al. 2007a, 2013). Recent dating of these deposits by Ruleman and colleagues (2016) yielded ages ranging from 385 to 289 ka. They theorize that lake overtopping and draining occurred over an interval of about 100,000 years starting at ~400 ka. In their model, the draining of the lake initiated the development of the Río Grande Gorge and integration of the northern SLV with downstream reaches of the river, which agrees with the theory first proposed by Wells and colleagues (1987). Formation of the Great Sand Dunes also post-dates the draining of Lake Alamosa (Madole et al. 2016; Valdez, this volume; Madole, this volume).

The largest outcrop of Alamosa Formation strata is at the 15-m-high Hansen Bluff in the Alamosa National Wildlife Refuge. Rogers and colleagues (1992) analyzed sediment from the bluff and a 127-m-deep core hole located at the base of the bluff. It ranged in age from 2.6 to 0.67 Ma, was chiefly deposited by meandering streams that probably flowed into Lake Alamosa, and contained a wealth of paleo-climatic information. Magnetostratigraphy, index fossils, and Pleistocene tephras provided much of the age control. The sediment suggests that a large ice field first developed in the San Juan Mountains at ~800 ka and that much of the glacial modification of surrounding mountain landscapes has occurred since then (Blair and Gillam 2011). SAM Cave, located northwest of San Antonio Mountain, also yielded considerable fossil and paleo-climatological data from sediment ~1 to 0.74 Ma (Rogers et al. 2000).

The youngest volcano in the Valley is 1.06 Ma Mesita Volcano (Thompson et al. 2015). Lava flows at La Segita Peaks and Pinebetoso Peaks are 2.16 Ma and 2.37 to 2.50

Ma, respectively (Appelt 1998; recalibrated to an age of 28.204 Ma for the monitor mineral sanidine from the Fish Canyon Tuff). Pleistocene ash erupted from distant volcanoes was also deposited in the Valley. They include the 639 ka Lava Creek B tephra (from Yellowstone caldera) that is found in several outcrops near San Luis (Kirkham et al. 2003; Machette et al. 2008), the 0.74 Ma Bishop tephra (from Long Valley caldera in California) and 2.02 Ma Huckleberry Ridge tephra (from Yellowstone caldera) reported at Hansen Bluff by Rogers and colleagues (1992), and the 1.12 Ma Tsankawí Pumice Bed (from Jémez caldera) that crops out south of the Red River (Ruleman et al. 2013).

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