MIDDLE TO LATE CENOZOIC DEVELOPMENT OF THE RIO GRANDE RIFT AND ADJACENT REGIONS IN NORTHERN NEW MEXICO

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INTRODUCTION

The middle and late Cenozoic geologic history of northern New Mexico features three, large-scale geologic processes. First, widespread volcanism in the region began at about 35 Ma and continued with varying intensity until the late Pleistocene. Second, the early Tertiary compressional Laramide orogeny gave way to Neogene extension through a yet poorly understood Oligocene transitional period. Late Cenozoic extension is best represented by the Rio Grande rift, faithfully followed by the namesake river, although normal faults are expressed across the entire width of the older Laramide tectonic welt in north-central New Mexico. Third, erosional denudation shaped the landscape as indicated not only by Quaternary incision of the Rio Grande but also by widespread removal of Mesozoic and early to middle Tertiary strata from the High Plains and Colorado Plateau provinces to the east and west of the rift, respectively.

Although emphasizing the middle and late Cenozoic evolution of the Rio Grande Valley and adjacent mountain ranges, this chapter examines all three aspects of the geologic history of northern New Mexico. A synthesis of existing data is used to reconstruct regional paleogeography since the end of Laramide deformation, and to address controversies surrounding the age and mechanisms of extensional basin formation in northern New Mexico.

BASINS OF THE RIO GRANDE RIFT

The Rio Grande rift in northern New Mexico consists of oppositely tilted half grabens flanked by discontinuous mountain ranges (Figs. 1–3). The west-tilted Española basin is separated from the east-tilted San Luis basin by the Embudo fault accommodation zone (Fig. 1). The basins are also separated by a northwest-trending basement arch traced from the Picuris Range, on the east, through Cerro Azul to the Tusas Mountains, on the west (Fig. 2; Kelley, 1978). The rift continues northward into Colorado as a linked series of narrow basins through the upper Arkansas valley

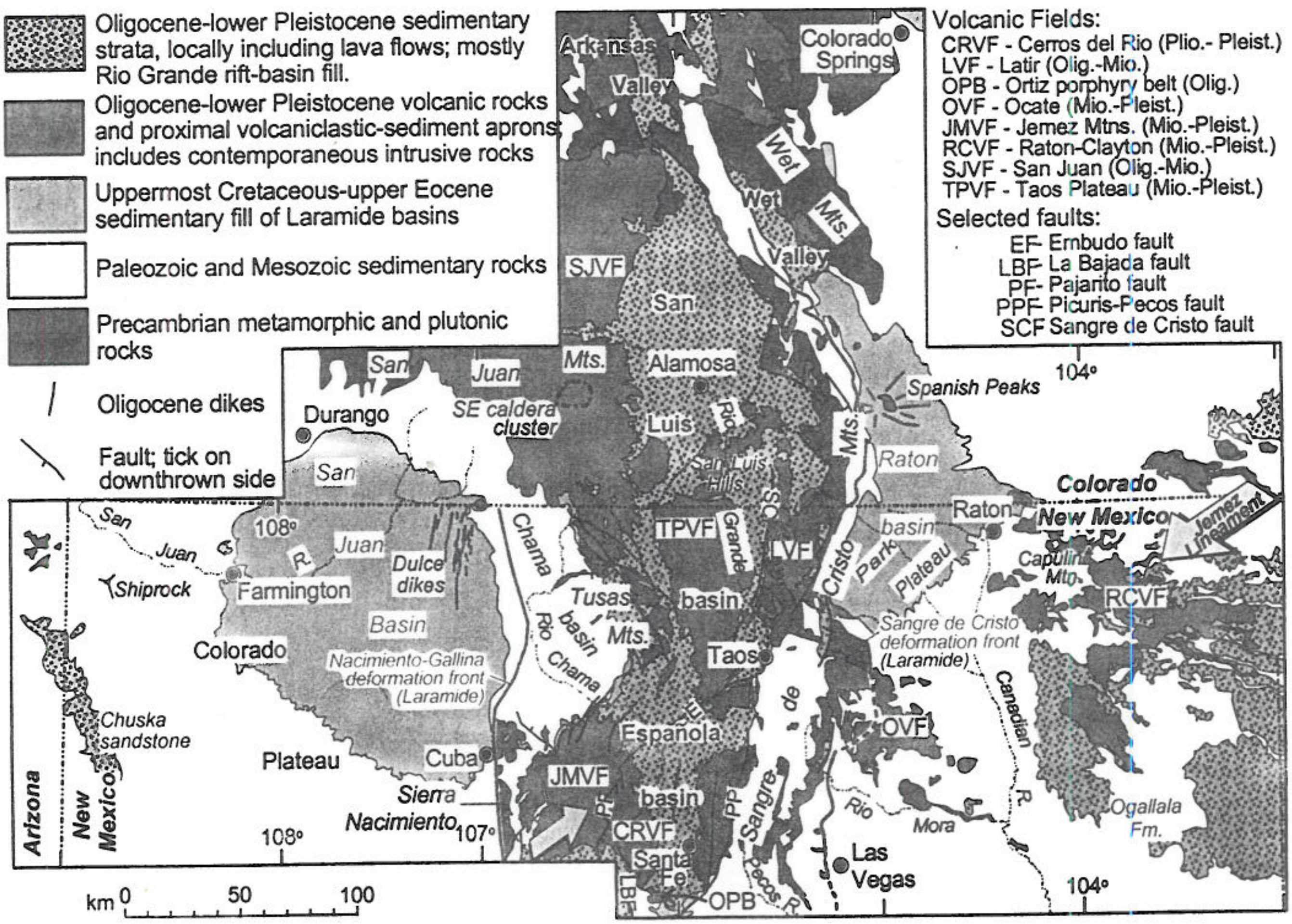

FIGURE 1. Generalized map of northern New Mexico and southern Colorado showing geologic and geographic features mentioned in text.

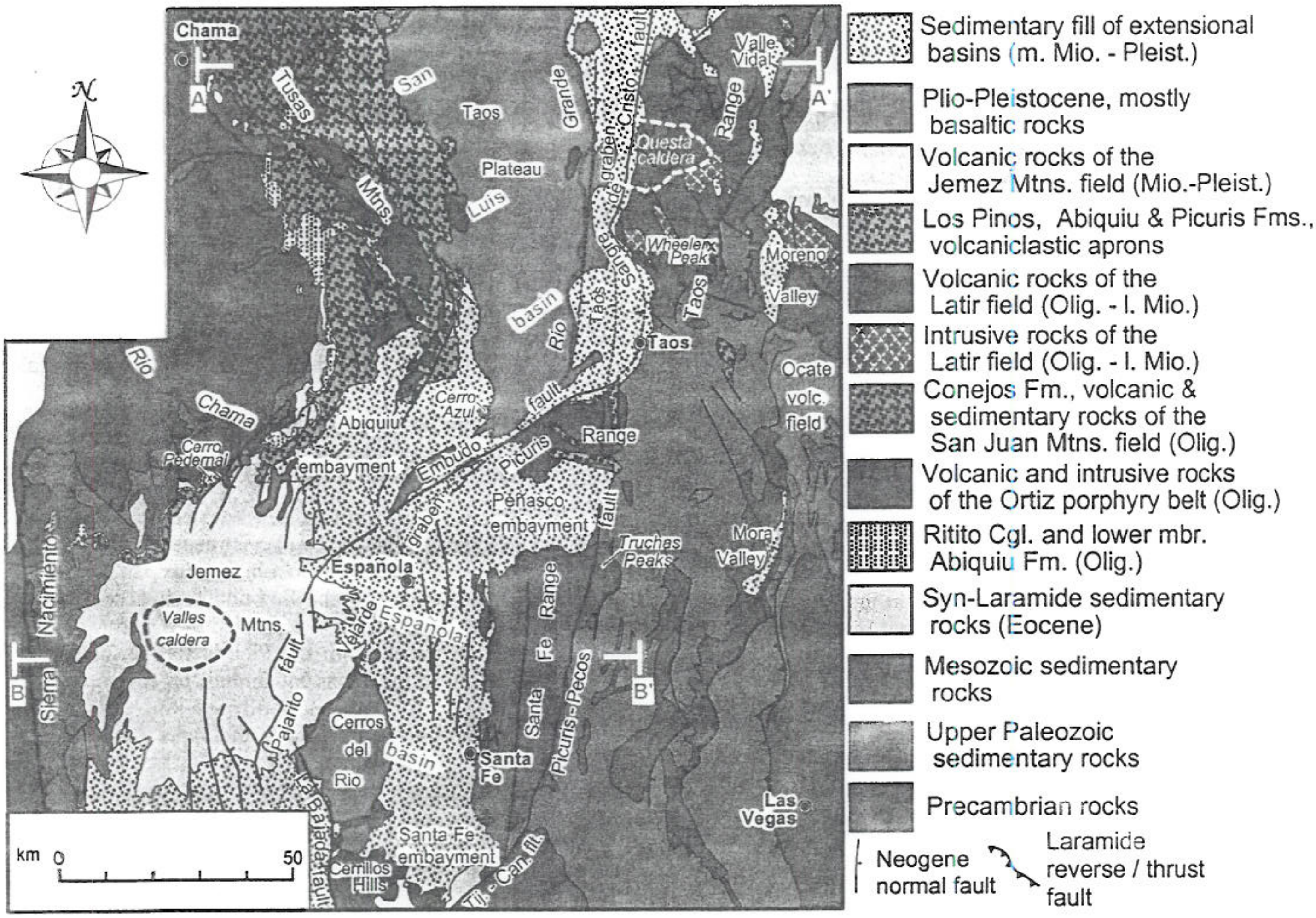

FIGURE 2. Simplified geologic map of the Rio Grande rift and adjacent mountain ranges in northern New Mexico. Cross-sections A–A' and B–B' are depicted in Figure 3.

(Fig. 1) to Leadville (Taylor, 1975; Scott, 1975; Tweto, 1979; Chapin, 1988; Chapin and Cather, 1994). Other extensional basins, such as the Wet Valley (Fig. 1), flank the rift in southern Colorado.

Deep dissection of the Española basin provides the best-exposed stratigraphic succession of rift-basin fill in the state (Fig. 4A). A long history of stratigraphic studies (e.g., Bryan, 1938; Spiegel and Baldwin, 1963; Galusha and Blick, 1971) followed by geophysical investigations (Cordell, 1978, 1979; Biehler et al., 1991; Ferguson et al., 1995) frame an unusually complete, though in places still controversial, view of the basin. In contrast, the thick fill of the San Luis basin is largely undissected (Fig. 4B) and concealed beneath Pliocene volcanic rocks of the Taos Plateau (Fig. 2).

San Luis basin

Two sub-basins comprise the San Luis basin and are separated by a northeast-striking fault along the northern margin of mid Tertiary volcanic rocks forming the San Luis Hills near the Colorado-New Mexico line (Fig. 1). Subsurface data from the northern basin reveal the western Monte Vista graben and eastern Baca graben separated by the buried Alamosa horst (Brister and Gries, 1994; Tandon et al., 1999). The Monte Vista graben contains nearly 1800 m of Eocene and Oligocene sediment that is missing from both the central horst and deep Baca graben, indicating an older Laramide and mid Tertiary western basin (Brister and Gries, 1994), which apparently terminates southward at the San Luis Hills. Approximately 6.4 km of Miocene and younger strata fill the Baca graben with 9.2 km of vertical separation along the 60° west-dipping normal fault at the base of the Sangre de Cristo Mountains, thus indicating about 8–12% extension (Kluth and Schaftenaar, 1994). This master normal fault continues steeply down to 2.6–2.8 km before flattening (Tandon et al., 1999).

The southern San Luis basin in New Mexico also encompasses a deep eastern basin, the Taos graben, containing 7–8 km of sedimentary fill (Fig. 3; Cordell, 1978). The San Luis basin shoals westward as a dip slope displaced by numerous northwest–north-northwest-striking, primarily west-side-down, faults (Lipman and Mehnert, 1979). The tilted west side of the southern San Luis basin is topographically defined by the Tusas Mountains consisting of Proterozoic rocks uplifted along the west side of the Laramide Brazos-Sangre de Cristo uplift, which are draped by Oligo-Miocene volcaniclastic rocks. The Sangre de Cristo Mountains footwall uplift also contains Proterozoic rocks that were uplifted and denuded of younger cover during Laramide deformation, and then overlain and intruded by lower Miocene igneous rocks. Although the deep San Luis basin marks the axis of the Rio Grande rift, middle Miocene to Pliocene extensional basins are present throughout the width of the Sangre de Cristo Mountains as far east

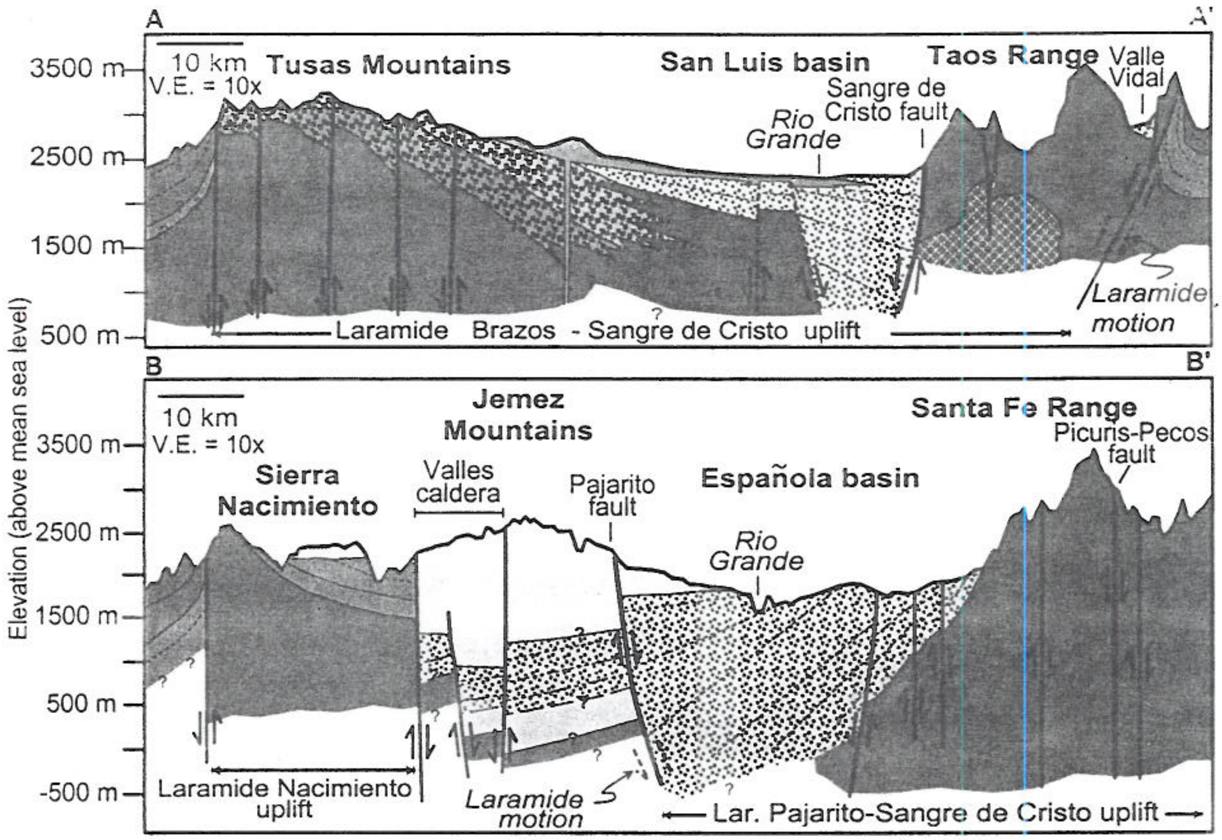

FIGURE 3. Cross-sections of the San Luis and Española basins. Lines of sections shown in Figure 2. See Figure 2 for legend.

as the reverse faults that define the Laramide deformation front facing the Raton basin (Fig. 1).

Española basin

The boundaries of the west-tilted Española basin, which has experienced about 10% extension (Golombek et al., 1983), have been treated in various fashions. Baltz (1978) and Golombek et al. (1983) place the western boundary at the Pajarito fault, whereas Manley (1979a) and Ingersoll (2001) place that margin at the eastern base of the Laramide Sierra Nacimiento uplift (Figs. 2, 3). Normal faults are prominent within the Jemez Mountains volcanic field, which partly obscures the rift margin (Smith et al., 1970; Kelley, 1978; Gardner et al., 1986), justifying the more westward placement of the basin boundary. The Española basin notably lacks a significant footwall uplift (Fig. 3).

The Pajarito and Embudo faults mark the western margin of the

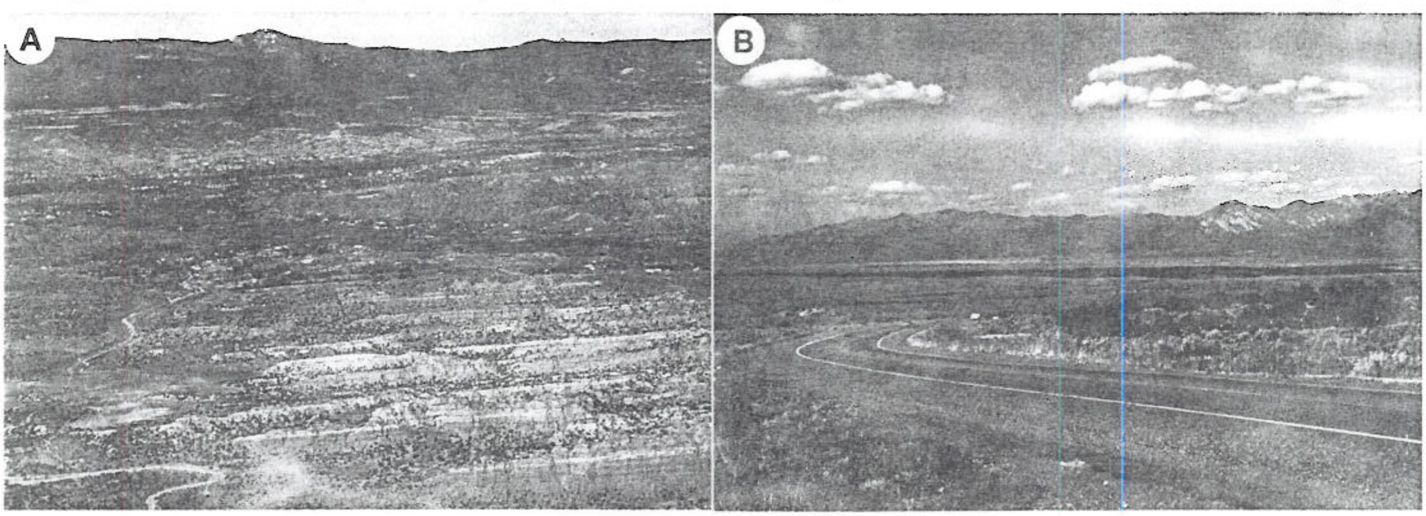

FIGURE 4. A, Oblique aerial view westward across the Española basin southeast of Española. Badlands in the foreground are eroded in west-dipping strata of the Skull Ridge Member of the Tesuque Formation with notable white ash layers. Jemez Mountains volcanic field forms the far horizon. B, View northeastward across the San Luis basin. City of Taos is in background at foot of Sangre de Cristo Mountains, which are composed of Precambrian rocks intruded by mid-Tertiary granitic bodies. Note the lack of dissection and exposure of basin-filling strata compared to the Española basin.

deep intrabasinal Velarde graben, containing more than 5 km of Tertiary sediment (Cordell, 1979; Biehler et al., 1991). The Velarde graben begins in the north as an 8 km-wide, northeast-trending trough partly defined by the Embudo fault (Manley, 1979a) and then possibly turns southward into the central Española basin. Gravity data suggest a narrow graben with an eastern boundary concealed beneath alluvial fill along the Rio Grande (Ferguson et al., 1995), although other workers have placed this boundary along or near faults that strike southward near Española into the central basin (Fig. 2. Manley, 1979a; Golombek et al., 1983; Dethier and Martin, 1984). Seismic-reflection lines, however, indicate insignificant offset of basement in this more eastern position (Biehler et al., 1991).

The Santa Fe Range of the Sangre de Cristo Mountains, consisting of Proterozoic rocks with local thin successions of upper Paleozoic strata, rise steeply above the eastern Española basin along a controversial boundary. Cabot (1938), Spiegel and Baldwin (1963), Kelley (1978), and Vernon and Riecker (1989) called attention to west-facing normal faults that locally juxtapose rift-basin fill and pre-Tertiary rocks. Baltz (1978), Manley (1979a), and Borchert (2002), however, consider these to be minor, discontinuous disruptions in a continuous dip slope, a conclusion supported by gravity (Cordell, 1979) and seismic-reflection (Biehler et al., 1991) data (Fig. 3). Galusha and Blick (1971) emphasized depositional contacts between basin-fill sediment and pre-Tertiary rocks. Denny (1940) suggested that basin-fill strata may have originally extended much farther east and subsequently stripped by erosion, an interpretation consistent with apatite-fission-track ages as old as 65 Ma (Kelley et al., 1992; Kelley and Chapin, 1995) in the Santa Fe Range. Preservation of Laramide-vintage denudation ages is consistent with recent uplift of the mountains, and possible burial beneath early-rift-basin strata until such uplift commenced.

Three pronounced topographic embayments are located adjacent to the main Española basin (Fig. 2). The Peñasco embayment forms the northeastern basin between the Santa Fe and Picuris Ranges where the contact between basin-fill strata and pre-Tertiary rocks steps abruptly eastward to the trace of the Picuris-Pecos fault (Fig. 2). The Santa Fe embayment forms the southeastern basin and consists primarily of north-dipping early Tertiary and pre-Tertiary rocks beneath a southward-tapering wedge of rift-basin fill (Grant, 1999). The pre-rift rocks culminate southward in the Cerrillos uplift east of the La Bajada fault and effectively define the southern basin boundary. The Abiquiu embayment lies west of the Embudo fault, disappears southward beneath the Jemez Mountains volcanic field, and terminates northward against the Tusas Mountains (Fig. 2). Faults along the western margin and within the Abiquiu embayment form an eastward-downstepping, structurally shallow bench between the deep rift basins on the east, and the gently folded strata of the Laramide Chama basin forming the edge of the Colorado Plateau on the west (Baldridge et al., 1994). Seismic-reflection data suggest a total of 1.1 km of dip slip across parallel, planar faults, implying about 3.5% extension (Baldridge et al., 1994). A variety of tilt directions are present in different fault blocks such that the embayment does not form a coherent structural domain (Kelley, 1978; May, 1979).

The Embudo fault accommodation zone between the San Luis and Española basins locally follows the Jemez Lineament, a northeast-striking alignment of upper Cenozoic volcanic fields and local faulting widely believed to mark a lithosphere-scale boundary inherited from Proterozoic accretionary tectonics (Fig. 1; Mayo, 1958; Lipman, 1979; Karlstrom and Humphreys, 1998; CD-ROM Working Group, 2002). The Jemez Mountains, Ocate, and Raton-Clayton volcanic fields fall along and partially define the lineament, which also delineates the southeastern margin of the Park Plateau, a deeply dissected part of the Laramide Ratón basin (Fig. 1).

STRATIGRAPHIC NOMENCLATURE

The Santa Fe Group is synonymous with Rio Grande rift basin-filling detritus throughout New Mexico and southern Colorado (e.g., Chapin, 1988). Although the group was defined in the Española basin (Spiegel and Baldwin, 1963), there is little consensus on which stratigraphic units comprise it there. The regional stratigraphy consists of Oligocene and lower Miocene sedimentary deposits of largely volcaniclastic composition, overlain by middle Miocene to Pliocene strata characterized, in most places, by increasing contributions from pre-Tertiary highlands bordering the rift (Fig. 5). Baldwin (1956) and Spiegel and Baldwin (1963) proposed raising the previously ill-defined Santa Fe Formation to group status for application as "a broad term including sedimentary and volcanic rocks related to the Rio Grande trough" (Spiegel and Baldwin, 1963, p. 39) but excluding alluvium and terraces of present drainages. Later workers (e.g., Galusha and Blick, 1971; Manley, 1979a) advocated removing from the group the Oligocene-lower Miocene strata and also Pliocene strata that locally disconformably overlie Miocene sediment. A separate stratigraphic nomenclature has also been erected for the volcanic rocks of the Jemez Mountains (Bailey et al., 1969).

It is beyond the scope of this chapter to attempt resolution of these various treatments of the boundaries of the Santa Fe Group so formation- and member-rank boundaries are emphasized (Fig. 5). Even then, ongoing mapping suggests that some members of the Tesuque Formation designated by Galusha and Blick (1971) are unmappable beyond the small areas where they were defined. The distinction of Tesuque and Chamita Formations is also questionable. The traditional stratigraphic names remain on the chronostratigraphic charts (Fig. 5) but are likely to be revised or replaced in the near future.

Middle and upper Cenozoic rocks are generally limited in lateral extent, thickness, or both, in areas east and west of the Rio Grande rift. These are not portrayed in the chronostratigraphic columns of Figure 5 but their distribution is indicated in Figure 1. In the Four Corners these include local Oligocene volcanic and intrusive rocks and the Chuska Sandstone (Wright, 1956; Smith et al., 1985; Cather et al., 2003). In the High Plains of northeastern New Mexico mostly mafic lava flows of the Ocate and Raton-Clayton volcanic fields and the Miocene alluvial and eolian strata of the Ogallala Formation unconformably overlie upper Paleozoic and Mesozoic strata.

SOURCES OF MIDDLE TO LATE CENOZOIC SEDIMENT

The provenance of detritus filling the Rio Grande rift offers insight to the paleogeographic and tectonic development of the region. Most sediment was eroded from diverse Precambrian terranes (Table 1) and syndepositional volcanic fields (Table 2) with locally important contributions from upper Paleozoic strata. Mesozoic rocks were removed from adjacent highlands by uplift and denudation during the Laramide orogeny and are not significant sediment sources to younger fill.

Precambrian detritus is notably arkosic adjacent to granitic-gneissic terranes and quartzite-rich adjacent to metasedimentary-metavolcanic terranes (Table 1; Fig. 5). Arkosic strata are derived

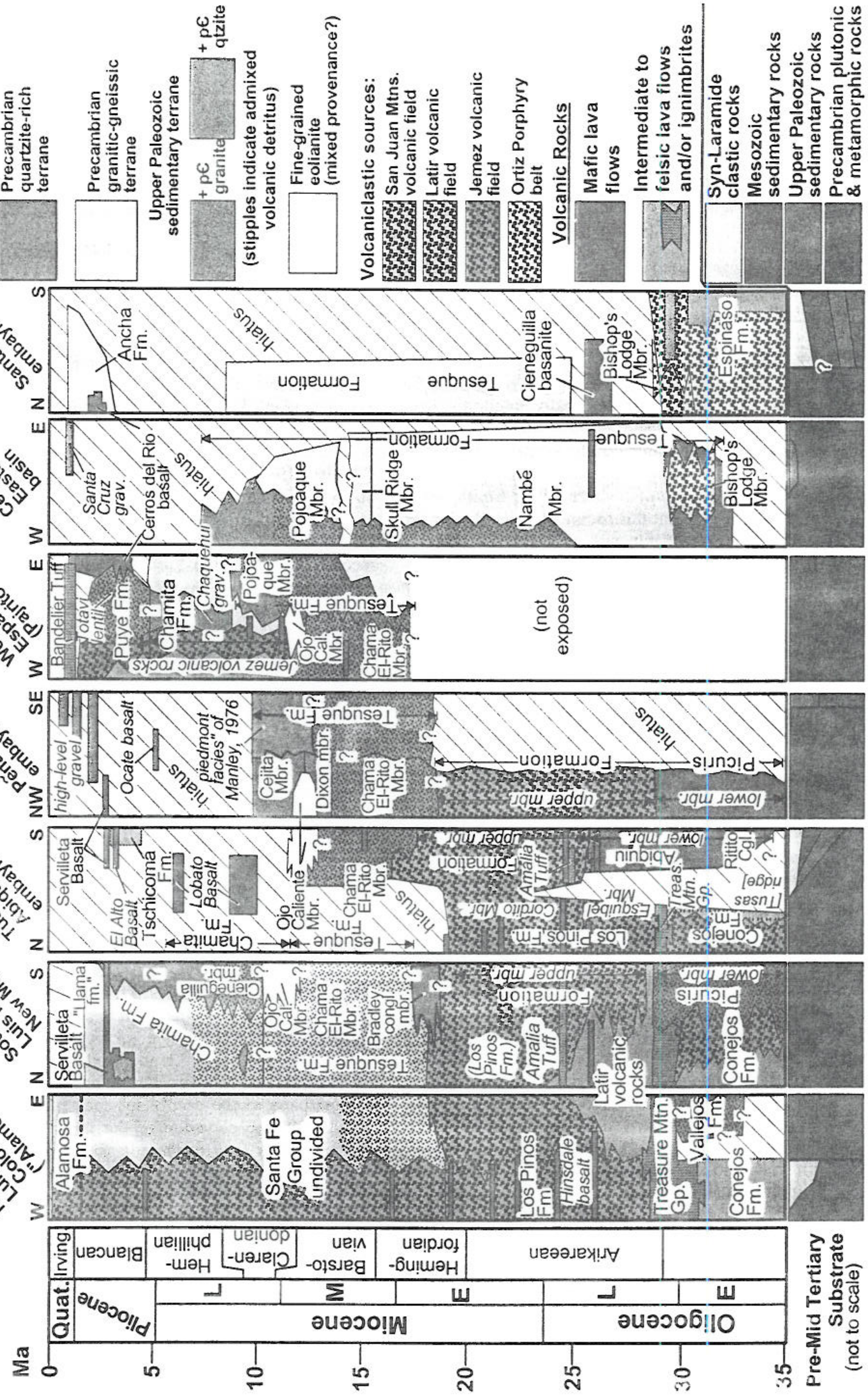

FIGURE 5. Chronostratigraphic columns for Oligocene and younger rocks along the Rio Grande rift in north-central New Mexico. See Appendix for sources of data used to construct these columns.

TABLE 1. Basement-rock types in mountain ranges of northern New Mexico

MOUNTAIN RANGE	ROCK TYPES (dominant; *subordinate*)	REFERENCES
Taos Range and adjacent Sangre de Cristo Mtns	Metavolcanic rocks, quartzite metapelitic rocks, metagranitic rocks, *metagabbro*	Condie (1979), Grambling et al. (1988)
Tusas Mountains	Quartzite, metarhyolite, *granite*	Barker (1958), Bingler (1968), Wobus 1984, Grambling et al. (1988)
Picuris Range	Quartzite, *granite, schist, metavolcanic rocks, amphibolite*	Montgomery (1953), Nielsen and Scott, Bauer (1993)
Truchas Peaks	Quartzite, metapelitic rocks, *granite, amphibolite*	Grambling (1979)
Santa Fe Range	Granite, amphibolite, *mafic gneiss*	Miller et al. (1963), Grambling et al. (1988)
Southeastern Sangre de Cristo Mtns	Metavolcanic rocks (including greenstone), metapelitic rocks, granite	Grambling et al. (1988
Sierra Nacimiento	Granite	Grambling et al. (1988)

from the Sierra Nacimiento and Santa Fe and Taos Ranges. Abundant quartzite clasts indicate sources in the Tusas and Picuris Mountains, the Truchas Peaks, and are locally important from sources east and northeast of Taos (Fig. 2).

Paleozoic-rock detritus is prominent in the Peñasco embayment (e.g., Manley, 1976a, 1977) and interbedded with arkosic strata in the central Española basin section (Cavazza, 1986; Smith, 2000; Fig. 5). In the Peñasco embayment this sediment is clearly derived from east of the Picuris-Pecos fault, and also contains volcanic clasts in mostly pre-12 Ma strata. This facies represents a persistent fluvial system that entered the basin from the northeast and produced distinctive, trough cross-bedded gravel and pebbly sand that is interbedded with thinner-bedded arkosic strata derived from the Santa Fe Range.

Coarse, limestone-clast conglomerate that is locally present in the lowest Tesuque Formation north of Santa Fe (Smith, 2000) likely has a closer source. Thin, discontinuous erosional remnants of upper Paleozoic strata are present along the east slope of Santa Fe Range and in the subsurface of the central Española basin. In most places, however, there is no Paleozoic section preserved below rift fill, and abundance of Precambrian detritus in older Laramide syntectonic strata south of Santa Fe (Gorham and Ingersoll, 1979) imply nearly complete denudation of Paleozoic rocks during the early Tertiary. Therefore, Paleozoic clasts in the lowest Tesuque Formation may originate east of the Picuris-Pecos fault at the latitude of Santa Fe.

Volcanic fields along the Rio Grande rift and Jemez Lineament were active at different times and erupted products of varying composition (Table 2). Oligocene to early Miocene magmatism in the San Juan Mountains and Latir volcanic fields (Fig. 1), and in the Ortiz porphyry belt continuing southward from the Cerrillos Hills (Fig. 2), occurred prior to and during the earliest phase of basin subsidence. The Jemez Mountains, Cerros del Rio, Taos Plateau, Ocate, and Raton-Clayton volcanic fields formed along or near the Jemez Lineament concurrently with rift basin formation and partly within the rift itself. The most voluminous volcanic outpourings occurred at the intersection of the rift and lineament in the Jemez Mountains. Ubiquitous dateable lava and pyroclastic deposits within rift-basin fill complement vertebrate paleontological data (e.g., Galusha and Blick, 1971; Tedford and Barghoorn, 1993) to establish a chronostratigraphic framework for the basins (Fig. 5).

Varying composition of detritus and isotopic ages of clasts link sedimentary deposits to specific volcanic source areas. Widespread volcaniclastic sediment of the Los Pinos Formation veneers Precambrian rocks in the Tusas Mountains (e.g., Butler, 1946, 1971; Barker, 1958), and consists of two compositionally distinct successions. Andesitic-rhydodacitic-composition detritus comparable to rocks in the San Juan volcanic field comprise the Esquibel Member (Fig. 5, column C). The overlying high-silica-rhyolite detritus of the Cordito Member, locally enclosing the upper Oligocene Amalia Tuff, is derived from the Latir volcanic field (Manley, 1981). Ingersoll and Cavazza (1991) demonstrated the utility of sandstone and conglomerate petrology to recognize the volcanic sources defining Esquibel and Cordito petrosomes of San Juan and Latir provenance, respectively, and the Plaza petrosome recording post-Amalia-Tuff volcanism in the southwestern Latir volcanic field from now-buried vents. Although, as discussed below, some intermediate-composition detritus assigned to the Esquibel petrosome is arguably derived from the Latir and Ortiz porphyry belt fields rather than the San Juan Mountains.

EARLY TO MIDDLE OLIGOCENE (36–28 MA) PALEOGEOGRAPHY

The end-Laramide landscape

A broad orogenic plateau extended north-south through north-central New Mexico at the close of the Laramide orogeny in the late Eocene. This tectonic welt was approximately 160 km wide (see labeled deformation fronts on Fig. 1) and bounded by high-angle west- and southwest-vergent reverse faults of the Nacimiento-Gallina and Brazos (Tusas Mountains) uplifts on the west, and the east-vergent reverse and thrust faults of the eastern Sangre de Cristo Mountains on the west. The opposite-verging reverse faults faced the Raton and San Juan-Chama foreland basins (Fig. 1) and narrow intra-uplift basins formed within the highland, although now largely concealed beneath the Jemez volcanic field (Cather, 1992) and fill of the northwestern San Luis basin (Brister and Gries, 1994). The Laramide plateau remains a fundamental aspect of the geologic structure in northern New Mexico, for even in most parts of the deeply subsided basins of the northern Rio Grande rift, the elevation of the top of Precambrian basement remains higher than in the adjacent Colorado Plateau and High Plains (Baltz, 1978).

Controversy surrounds understanding of when the southern Rocky Mountains reached their current high elevation, mostly between 2 and 4 km above sea level. Eaton (1986) championed Neogene epeirogenic uplift above a buoyant mantle during formation of the Rio Grande rift. Paleobotanical analyses (Gregory and Chase, 1992) suggest that the high elevation of the region was a consequence of crustal thickening during the Laramide orogeny and that current topographic relief between Precambrian-cored mountain ranges and adjacent sediment-filled basins resulted from

TABLE 2. Volcanic fields in and near the Rio Grande rift in northern New Mexico

VOLCANIC FIELD	ROCK TYPES	AGE OF VOLCANISM	REFERENCES
Ortiz porphyry belt	Latite and quartz latite and related monzonite and quartz monzonite intrusions	36–34 Ma (calc-alkaline phase) 32–28 Ma (alkaline phase)	Stearns (1953), Disbrow and Stoll (1957), Sun and Baldwin (1958), Maynard et al. (1990), Erskine and Smith (1993), Sauer (1999)
San Juan (southeast caldera cluster)	Andesitic and dacitic lava (Conjeos Fm.).	33–29.5 Ma	Lipman (1975a), Lipman and Mehnert (1975), Lipman (1989), Lipman et al. (1996)
	Dacitic, quartz latitic, and rhyolitic ignimbrites (principally Treasure Mountain Group) and associated andesitic to rhyolitic lava.	31–28.4 Ma	Lipman (1975a, 1989), Lipman and Mehnert (1975), Lipman et al (1996)
	Slightly alkaline basalt and minor rhyolite lava and tuff	26.8–22.0 Ma	Lipman (1975a)
Latir	Andesitic and rhyodacitic lava; few rhyolitic ignimbrites. Includes San Luis Hills and volcanic rocks exposed along east side of San Luis basin in Colorado	28.5–26 Ma	Lipman (1983), Lipman and Reed (1989), Johnson et al. (1989), Thompson et al. (1991), Wallace (1995)
	Peralkaline-rhyolite ignimbrites (e.g., Amalia Tuff), domes and related intrusions; formation of Questa caldera	26–24 Ma	Lipman and Reed (1989), Johnson et al. (1989), Czamanske et al. (1990)
	Andesite, rhyodacite, and rhyolitic lava; intrusion of granite plutons; some volcanic products principally known as clasts in conglomerates	22–18 Ma	Lipman and Reed (1989), Johnson et al. (1989), Czamanske et al. (1990), Ingersoll and Cavazza (1991), Smith et al. (2002)
Jemez Mountains	Tholeiitic basalt interspersed with and followed by andesite, dacitic, quartz latite and locally important rhyolite lava and tuff	14–2 Ma	Smith et al. (1970), Gardner et al. (1986); WoldeGabriel et al. (2001)
	Rhyolitic ignimbrites (formation of Valles caldera), rhyolite lava domes	1.87–0.06 Ma	Smith et al. (1970), Goff et al. (1996)
Taos Plateau	Widespread olivine tholeiitic basalt (Servilleta Basalt) with lesser basaltic andesite, alkalic basalt and large rhyodacitic volcanoes and a rhyolite dome complex	Early silicic volcanism about 10 Ma but mostly 2–5 Ma; basalts 4.8–2.8 Ma	Lipman and Mehnert (1979), Appelt (1998)
Cerros del Rio	Basalt, hawaiite, andesite, minor dacite; hydro-magmatic tuff	2.8–1.4 Ma	Aubele (1979), WoldeGabriel et al. (1996), Dethier (1997), Sawyer et al. (2002)
Raton-Clayton	Basalt with minor andesite and dacite	9.0–0.06 Ma	Stormer (1972), Calvin (1987), Stroud (1997)
Ocate	Basalt with minor basaltic andesite and dacite	8.3–0.8 Ma	O'Neill (1988), O'Neill and and Mehnert, O'Neill and Mehnert (1990), Olmsted (2000)

Neogene denudation and differential erosion (e.g., Formento-Triglio and Pazzaglia, 1998). Nonetheless, apatite-fission-track data and geomorphic analyses in some mountainous areas adjacent to the rift in northern New Mexico and southern Colorado lead to the notion that tectonic rock uplift during Laramide compression was minor compared to that which followed and was driven either by isostatic responses to erosion or mantle buoyancy (Pazzaglia and Kelley, 1998). Paleobotanical data suggest a regional increase in mean elevation from 2.5 to 4 km during the Oligocene (Chase et al., 1998) perhaps driven by thermal buoyancy accompanying widespread mid-Tertiary magmatism (Roy et al., 1999). It is likely that the early Oligocene landscape featured a broad plateau with a mean elevation between 2–3 km rising gradually above the San Juan Basin and Great Plains. On this landscape, volcanoes and their flanking aprons of coarse-grained detritus emerged in the early Oligocene, filling remnant Laramide depressions and mantling the mostly Precambrian crystalline rocks of the tectonic highlands (Fig. 6).

Volcaniclastic apron from the San Juan volcanic field

Oligocene volcanism in the north was centered in the San Juan volcanic field (Fig. 1). Andesitic lava flows and detritus of the Conejos Formation (29.5–33 Ma, Lipman and Mehnert, 1975; Lipman, 1989; Lipman et al., 1996) were followed by outflow-sheet ignimbrites from the southeast caldera cluster (31.0–28.4 Ma, Lipman, 1975a, 1989; Lipman et al., 1996; Fig. 1) that dispersed southward into New Mexico. A northwest-trending remnant Laramide, or possibly still active, highland in the Tusas Mountains is revealed by restriction of Oligocene rocks of the Conejos

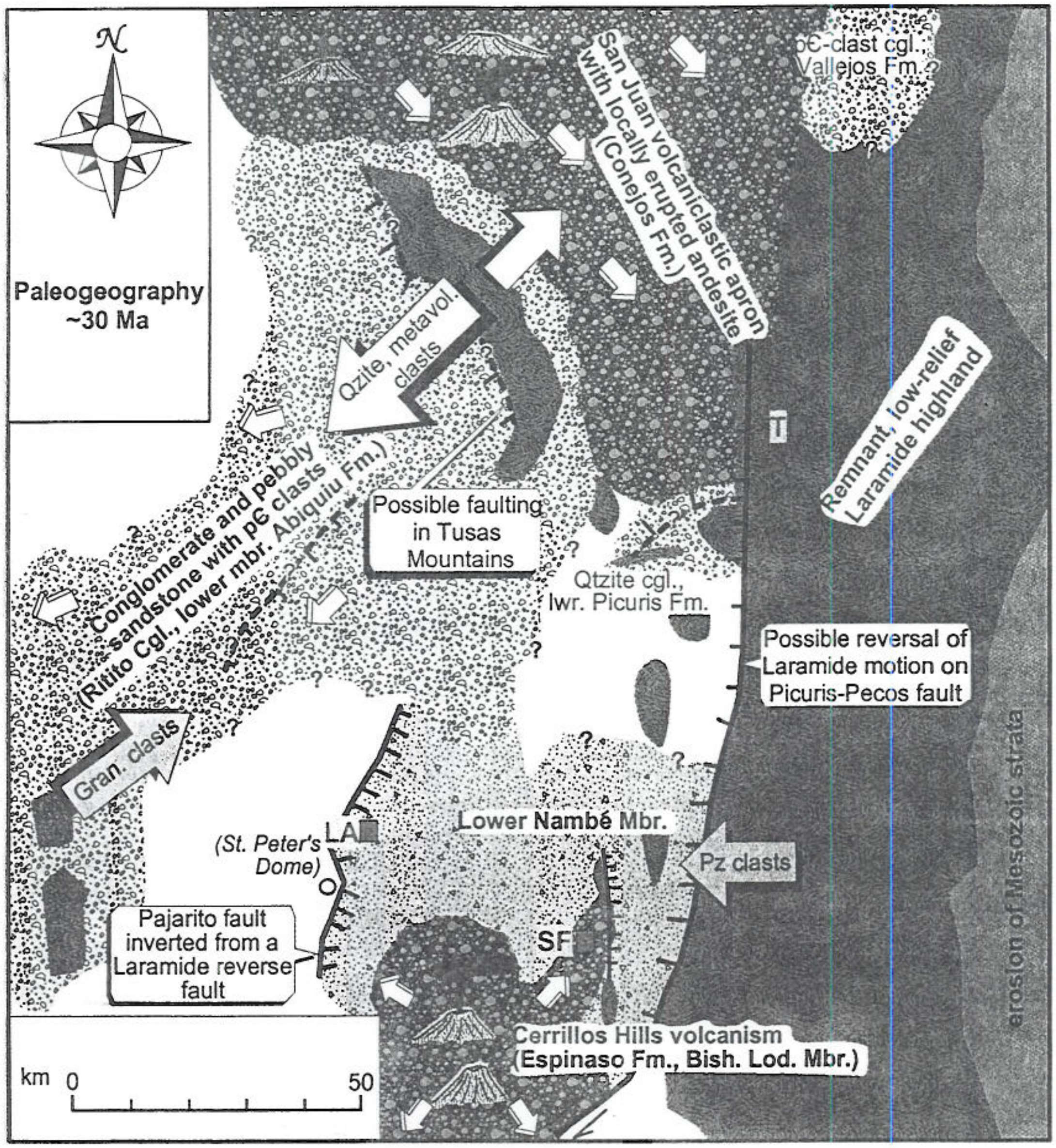

FIGURE 6. Paleogeographic map of north-central New Mexico during the late Oligocene. In this, and following figures: faults are shown with ticks on downthrown side with most likely or known-active faults illustrated as solid lines and those of more questionable nature illustrated by queried and dashed lines; arrows depict generalized sediment dispersal; white areas are not reconstructed because of lack of data; location of Taos (**T**), Santa Fe (**SF**) and Los Alamos (**LA**) are shown for reference.

Formation and overlying ignimbrites of the Treasure Mountain Group to the northeast flank of the Tusas Mountains (Fig. 5, column B; Fig. 6; Barker, 1958; Bingler, 1965, 1968; Doney, 1968; Muehlberger, 1967, 1968; Manley and Wobus, 1982a, 1982b; Wobus and Manley, 1982; Lipman et al., 1996). The Conejos volcaniclastic apron likely extended southeastward to the vicinity of the Picuris Range (Fig. 6) where the lowermost Picuris Formation, containing intermediate volcanic clasts in addition to locally derived quartzite, includes earliest Oligocene tephra (34.64 ± 0.16 Ma, Bauer et al., 1999). Quartzite-rich gravel shed from the Tusas Mountains at this same time is interbedded with volcanic rocks to the northeast, and forms a contemporaneous blanket of equivalent conglomerate (Ritito Conglomerate and lower member of the Abiquiu Formation) largely devoid of volcanic detritus to the southwest (Fig. 6; Smith, 1938; Bingler, 1968; Moore, 2000). Subsidence of the Monte Vista graben in the northwestern San Luis basin is suggested by the presence of 1200 m of Conejos Formation preserved beneath San Juan ignimbrites and younger sediment, and some seismically imaged faults offset the Conejos Formation but not younger strata (Brister and Gries, 1994).

Volcaniclastic apron from the Ortiz porphyry belt

Intrusion and extrusion of calc-alkaline and later alkaline latites of the Ortiz porphyry belt began in latest Eocene time (Maynard et al., 1990; Sauer, 1999), prior to cessation of Laramide basin-fill sedimentation south of Santa Fe (Fig. 5, column G, Table 2; Stearns, 1953; Disbrow and Stoll, 1957; Sun and Baldwin, 1958; Bachman, 1975). This volcanic field is represented by stocks and laccoliths of the Cerrillos Hills and La Cienega area (24 km southwest of Santa Fe) and underlies much of the Santa Fe embayment as well as continuing southward from the area depicted in Figure 1 for an additional 20 km. Collapse of highly porphyritic, viscous domes erupted at the north end of the chain provided abundant coarse debris for resedimentation in volcaniclastic aprons of the Espinaso Formation in the Santa Fe embayment (Fig. 5, column G) and farther south (Kautz et al., 1981; Smith et al., 1991; Erskine and Smith, 1993). Detritus and lava (~31–29 Ma) of the alkaline magmatic phase onlap Proterozoic basement south of Santa Fe (Stearns, 1953, Baldridge et al., 1980; Cather, 1992; Erskine and Smith, 1993; Read et al., 1999) indicating overfilling of the remnant Laramide basin to the south, or structural foundering of the Laramide Sangre de Cristo uplift (Fig. 6).

Volcaniclastic detritus is also found in the lower part of the Tesuque Formation within the Española basin near and north of Santa Fe. Kottlowski (in Spiegel and Baldwin, 1963) named these strata the Bishop's Lodge Member of the Tesuque Formation and showed the tuffaceous volcaniclastic sediment to be interbedded with nonvolcanogenic strata of the Nambé Member (Fig. 5, column F). Subsequent workers (Galusha and Blick, 1971; Manley, 1979a; Ingersoll and Cavazza, 1991; Ingersoll et al., 1991) assigned the Bishop's Lodge Member to the base of the formation and placed an unconformity between it and overlying Tesuque strata, which were viewed as entirely middle Miocene. More recent mapping (Smith, 2000; Read et al., 2000; Borchert, 2002) corroborates, however, the designation of at least two nearly pure volcaniclastic intervals within, not at the base of, the Tesuque Formation.

The Bishop's Lodge Member has been variously interpreted as distal equivalent of the Picuris Formation (Cabot, 1938) or Abiquiu Formation (Stearns, 1953) originating in the Latir and San Juan volcanic fields although the coarseness of clasts suggested a local source to Spiegel and Baldwin (1963). Ingersoll and Cavazza (1991) attributed derivation of these strata from the San Juan Mountains, although Large and Ingersoll (1997) noted possible erosion from the Cerrillos Hills area, consistent with paleocurrents measured by Cather (1992). Smith (2000, and in Read et al., 2000) correlate the Bishop's Lodge Member to the alkaline phase of magmatism recorded in the Espinaso Formation, because of southward coarsening of clasts of augite-biotite latite comparable to the upper Espinaso (Erskine and Smith, 1993), and a 30.45 ± 0.16 Ma pumice-lapilli fall deposit of local origin.

Oligocene deposition in the Española basin

Oligocene volcaniclastic debris of the Bishop's Lodge Member overlies and is intercalated with coarse detritus eroded from the Sangre de Cristo Mountains. Northeastward progradation of the volcaniclastic apron occasionally obstructed the westward-flowing drainages to form small, short-lived gypsiferous playas (Boyer, 1959). The lower beds of the Nambé Member north of Santa Fe are locally more than 400 m thick beneath the 30.45 Ma Bishop's Lodge tephra. These oldest Tesuque Formation strata are composed of 40–65% upper Paleozoic clasts with the remainder being primarily Proterozoic granite (Smith, 2000). Paleozoic clasts abruptly decrease above the volcaniclastic horizons where more typical arkosic Nambé strata locally enclose a ~25 Ma basalt (Baldridge et al., 1980). This lava flow is potentially correlative to the basalt and basanite flows of the Cieneguilla Limburgite of Stearns (1953) in the western Santa Fe embayment from which a 26.08 ± 0.62 Ma $^{40}Ar/^{39}Ar$ date has been recently obtained (Koning and Hallett, 2001).

Oligocene deposition upon basement rock supports the hypothesis of Cather (1992) that the Española basin east of the Pajarito fault was part of the Laramide Brazos-Sangre de Cristo uplift that was structurally inverted by reversal of motion on the Pajarito fault from west-side down to west-side up beginning in Oligocene time (Fig. 6). Subsidence of a middle Tertiary basin east of the Pajarito fault is also supported by an approximately 30° angular discordance between syn-Laramide Galisteo Formation and middle Miocene Tesuque Formation in the footwall of the Pajarito fault near St. Peter's Dome (Fig. 6, Goff et al., 1990; Cather, 1992) and the presence of Tesuque Formation as old as Oligocene within the basin to the east.

Significance of Oligocene nonvolcaniclastic sedimentation

Nonvolcaniclastic sedimentary strata are a significant component of Oligocene strata in northern New Mexico (Fig. 6). The lowermost Nambé Member in the Española basin is composed primarily of Paleozoic and Precambrian granitic detritus and the Conejos Formation and the lower members of the Abiquiu Formation (Ritito Conglomerate) and Picuris Formation contain abundant quartzite derived from the Tusas Mountains and Picuris Range. The Vallejos Formation is a thin, probably less than 100-m thick, Precambrian-clast conglomerate found locally in the Sangre de Cristo Mountains near the Colorado-New Mexico border at the base of the upper Oligocene-lower Miocene volcanic section (Fig. 5, column A; Fig. 6). Although viewed as a syn-Laramide deposit by Wallace (1995), Clark and Read (1972) reported a locally tuffaceous matrix in Vallejos conglomerate, suggesting a younger Oligocene age. A clear disconformity, and locally an angular unconformity, separates the lower member of the Abiquiu Formation from the underlying syn-Laramide strata of the El Rito Formation in the Abiquiu embayment (Doney, 1968, Smith et al., 1961, Moore, 2000). These observations suggest that Oligocene, nonvolcaniclastic conglomerates are not simply the waning phases of erosion and deposition from Laramide highlands but represent rejuvenated sedimentation from similar source areas. Furthermore, pre-21 Ma sedimentation cannot be dismissed as only volcaniclastic-apron aggradation independent of basin subsidence, as proposed by Ingersoll et al. (1990), Large and Ingersoll (1997), and Ingersoll (2001).

For the lower Abiquiu and Picuris Formations, which are only about 50–150 m thick (Smith, 1938; Bingler, 1968; Rehder, 1986; Moore, 2000), it is unclear if renewed deposition of basement-derived detritus relates to Oligocene basin subsidence or is, perhaps, a response to early Oligocene climate change (cf. Clark, 1975). Smith et al. (2002) suggest that subsidence of sufficient magnitude to accommodate these strata might result from flexural loading peripheral to the large middle Tertiary volcanic fields. Stratigraphic data (Moore, 2000; Smith et al., 2002) demonstrate, however, thinning of the lower member of the Abiquiu Formation across the Cañones fault (Fig. 6) with motion that is arguably an inversion of Laramide motion on the same structures. Faulting along the eastern margin of the rift was, therefore, underway during the late Oligocene.

In addition to evidence for inversion of Laramide motion on the Pajarito fault (Cather, 1992), faulting may also have occurred in the vicinity of the Santa Fe Range. Pointing to the abundance of Paleozoic detritus in the lower Nambé Member of the Tesuque Formation, Smith (2000) speculated that west-side down motion on the near-vertical Picuris-Pecos fault (which was a west-side-up, oblique-slip fault during the Laramide) may have occurred during the Oligocene in order to provide a source for the Paleozoic clasts (Fig. 6); Paleozoic rocks are rare today west of the fault but are common to the east (Figs. 1, 2). Oligocene subsidence and burial of the Santa Fe Range Proterozoic rocks, uplifted and denuded during the Laramide, could explain the preservation of Laramide-age apatite-fission-track cooling ages in that area (Kelley et al., 1992).

LATE OLIGOCENE TO EARLY MIOCENE (28–18 MA) PALEOGEOGRAPHY

Volcaniclastic aprons from the Latir and San Juan volcanic fields

Volcaniclastic sedimentation continued in northern New Mexico following eruption of the last Treasure Mountain Group tuffs at 28.4 Ma (Lipman et al., 1996). The intermediate-composition detritus of the Esquibel Member of the Los Pinos Formation, like older debris eroded from the San Juan volcanic field, is mostly or entirely restricted to the northeast side of the Tusas Mountains (Fig. 5, column B; Fig. 7; Barker, 1958; Bingler, 1965, 1968; Doney, 1968; Muehlberger, 1967, 1968; Manley and Wobus, 1982a, 1982b; Wobus and Manley, 1982). The absence of San Juan detritus in the Abiquiu embayment is contrary to Vazzana and Ingersoll (1981) and Ingersoll and Cavazza (1991) who correlated the volcaniclastic upper member of the Abiquiu Formation with the Esquibel Member of the Los Pinos Formation. Subsequent documentation of abundant, younger rhyolitic detritus from the Latir volcanic field (Smith, 1995; Moore, 2000) confirms, however, earlier correlations (e.g., Bingler, 1968; Manley, 1981) of the upper Abiquiu to the Cordito Member of the Los Pinos Formation, of Latir provenance, and a paucity of San Juan detritus in the Abiquiu Formation. It seems likely that San Juan-derived detritus was shed farther west into the San Juan Basin, but has subsequently been eroded from the Colorado Plateau.

The Latir volcanic field (Fig. 7) was active by 28.4 Ma (Bauer et al., 1999) with the first ignimbrites erupted at about 27 Ma (Lipman, 1983). Intermediate-composition lavas of this age are also present in the Sangre de Cristo Mountains (Lipman and Reed, 1989) and the San Luis Hills (Thompson and Machette, 1989). Rhyolitic volcanism reached a zenith with eruption of the Amalia Tuff to form the Questa caldera (Figs. 2, 5, 7). This ignimbrite, with distinctive smoky quartz and chatoyant microperthitic soda-sanidine crystals (Lipman, 1983) was dated at 25.6 ± 0.1 Ma by Czamanske et al. (1990), although higher-resolution, single-crystal-sanidine dates indicate an age of 25.1 Ma (Smith et al., 2002).

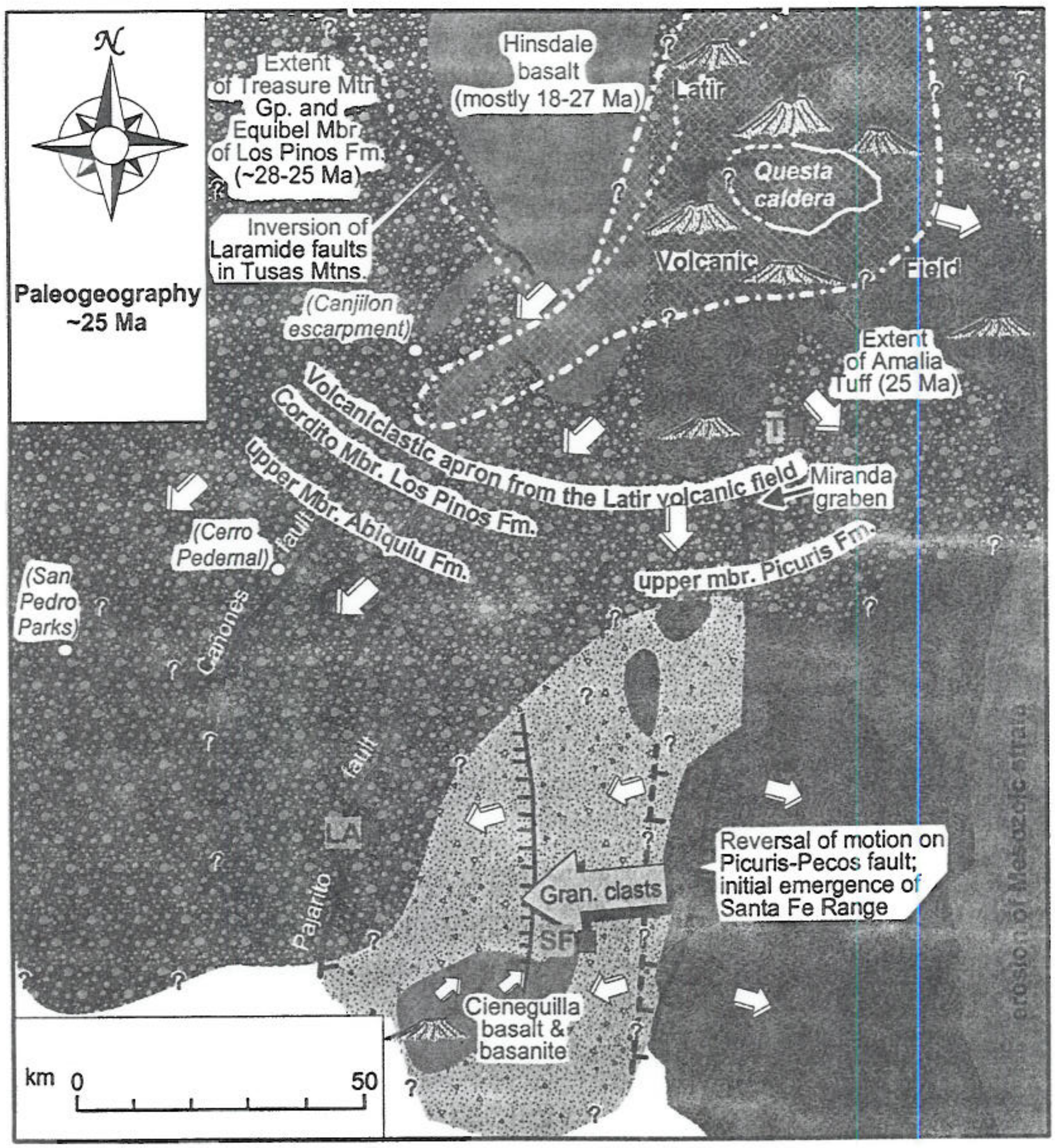

FIGURE 7. Paleogeographic map of north-central New Mexico during early Miocene.

Extrusion of rhyolitic ignimbrites and lava flows was locally accompanied by andesitic to rhyodacitic lava until 18 Ma (Ingersoll and Cavazza, 1991; Czamanske et al., 1990; Bauer et al., 1999). Most of the record of post-Amalia-Tuff volcanism comes from dated sedimentary clasts (Ekas et al., 1984; Ingersoll and Cavazza, 1991; Moore, 2000), or is inferred from dated plutons near Questa and Red River. The indirect chronology is necessary because volcanic rocks were subsequently largely stripped within uplifted parts of the Latir field in the Sangre de Cristo Mountains or buried beneath younger fill of the San Luis basin. Down faulted lower Miocene volcanic rocks locally project above the Taos Plateau at Brushy and Timber Mountains, northwest of Taos (Lipman and Mehnert, 1975; Dungan et al., 1984).

Volcanic rocks and derivative volcaniclastic detritus of the Latir volcanic field underlie the San Luis basin (Fig. 5, columns A and B; Fig. 7) as revealed by exposures around the basin margins (Manley, 1981; Wallace, 1995; Leininger, 1982; Rehder, 1986; Bauer et al., 1999). These include the upper member of the Picuris Formation on the flanks of the Picuris Range (Cabot, 1938; Montgomery, 1953; Miller et al., 1963; Leininger, 1982; Rehder, 1986), which is correlative with the Cordito Member of the Los Pinos Formation and upper member of the Abiquiu Formation. The presence of volcaniclastic detritus on both flanks of the Picuris Range indicates the lack of a significant mountain barrier at that time (Manley, 1984). Nonetheless, the mixture of Precambrian and volcanic detritus in the lower member of the Picuris Formation on the north flank of the range (Fig. 5, column B) but the absence of volcanic clasts in presumed correlative strata on the south flank (Fig. 5, column D; Rehder, 1986) implies a ridge that was progressively buried by volcaniclastic debris from the north. Subsidence of the Miranda graben (Fig. 7) likely occurred at this time (McDonald, 1999) and still preserves the most complete sections of the Picuris Formation (Rehder, 1986) across the axis of the Picuris Range. Although most of the volcanic debris in the uppermost 200–250 m of Picuris Formation contains the quartz and alkali feldspars characteristic of the Cordito petrosome of Ingersoll and Cavazza (1991), including clasts of Amalia Tuff (Rehder, 1986; Smith, *unpubl.*), the lower 150 m is mostly intermediate-composition detritus characteristic of pre-Amalia-Tuff volcanism in the Latir field (Lipman, 1983; Lipman and Reed, 1989; Table 2). Ingersoll and Cavazza (1991) mistakenly attributed these lower strata in the Sangre de Cristo Mountains and Picuris Range to a San Juan Mountains provenance (i.e., their Esquibel petrosome).

The Latir volcaniclastic apron and outflow Amalia Tuff onlapped and buried the Proterozoic-rock ridges of the Tusas Mountains (to elevations of 3500 m) and prograded across the Abiquiu embayment (Fig. 7). Some tectonic inversion of high-angle, northwest-striking Laramide reverse faults in the Tusas Mountains likely occurred at this time (Fig. 7; Muehlberger, 1967), which may have aided the deposition of the Cordito Member over the highlands that previously obstructed transport of volcaniclastic detritus into the Abiquiu embayment. Distribution of the Amalia Tuff and basalt flows of similar vintage (Fig. 7) suggest low relief, southwest-trending alluvial valleys crossing the older Tusas highland. The upper member of the Abiquiu Formation includes a variety of welded-ignimbrite and other volcanic clasts as young as 17–18 Ma (Manley and Mehnert, 1981), which are presumably cogenetic with post-Amalia-Tuff plutons near Questa (Lipman and Reed, 1979; Czmanske et al., 1990).

The time-varying input of volcanic detritus of different composition permits erection of a stratigraphy for the upper member of the Abiquiu Formation based on sandstone composition (Smith, 1995; Moore, 2000; Smith et al., 2002). These data indicate that active faulting along the western margin of the Abiquiu embayment in the earliest Miocene, accommodating thicker stratigraphic sections at within the embayment than on the adjacent Colorado Plateau (Fig. 7, Smith et al., 2002). Nonetheless, erosional remnants of volcaniclastic sedimentary strata at Cerro Pedernal and atop the Sierra Nacimiento at San Pedro Parks (Fig. 7) require that these faults did not effectively bound a basin at that time (Baltz, 1978; Manley, 1979a; May, 1984a). Likewise, eastward tilting of the San Luis basin, although likely underway as discussed below, was apparently not occurring rapidly or else a strong westward paleoslope would not have been maintained away from the Latir volcanic field.

Oligocene and early Miocene deposition of volcaniclastic debris from the San Juan Mountains and Latir volcanic fields was likely extensive across the Colorado Plateau. Ritito and Los Pinos Formations rise westward to 3200 m on the Canjilon Escarpment in the western Tusas Mountains (Fig. 7; Doney, 1968). There are no significant faults along or west of the escarpment to account for the high elevation of these outcrops suggesting that this feature is simply the eroded margin of a sedimentary apron that once extended considerably farther west. Abiquiu Formation is found up to 3200 m at the highest elevations in the northern Sierra Nacimiento (Church and Hack, 1939; Timmer, 1976). Although there is evidence for Quaternary motion on faults along the west side of the Nacimiento uplift, Formento-Triglio and Pazzaglia (1998) argue that exhumation, not faulting, accounts for present relief along the mountain front. This observation suggests that the high-elevation outcrops of Abiquiu Formation are remnants of an originally more extensive outcrop belt. Fragments of distinctive chalcedony forming the middle Pedernal chert member of the Abiquiu Formation (Church and Hack, 1939; Vazzana, 1980; Moore, 2000) are found along the continental divide west of the Sierra Nacimiento at 2400 m and may represent a lag from erosion of Abiquiu Formation that was once extensive to the west (Formento-Triglio and Pazzaglia, 1998).

Further supporting an extensive blanket of Oligocene sediment in the San Juan Basin is the presence of the Chuska Sandstone along the Arizona-New Mexico border west of the basin (Fig. 1). The Chuska is a 535-m thick eolianite with minor ash layers, and a lowermost fluvial interval (Wright, 1956; Smith et al., 1985; Cather et al., 2003). Geochronologic study of the ash layers and overlying lava flows indicate late Eocene and Oligocene deposition (Cather et al., 2003). Rising to 2900 m elevation, the Chuska Sandstone is an erosional remnant of Oligocene strata that were likely contiguous with high-elevation outcrops of Abiquiu and Los Pinos Formations east of the San Juan Basin. Upper Miocene strata in the northwestern Albuquerque basin contain clasts of Oligocene volcanic rocks (Connell et al., 1999) and may record the later denudation of widespread mid-Tertiary volcaniclastic strata in northwestern New Mexico.

Mafic magmatism

Eruption of mafic lava began in northern New Mexico in late Oligocene. This magmatism arguably marks the onset of decompression melting of lithospheric mantle related to formation of the Rio Grande rift (Lipman and Mehnert, 1975; Chapin, 1988; Ingersoll et al., 1990; Baldridge et al., 1995).

Hinsdale Formation basalt flows (Jarita basalt of early workers, see Manley, 1981) are interbedded in both members of the Los Pinos Formation, and range in age from 28 to 5 Ma (Lipman and

Mehnert, 1975, 1979; Baldridge et al., 1980; Fig. 5, columns A, B, and C). The basalt is most conspicuous in the northern Tusas Mountains (Fig. 7) where it caps east-tilted mesas projecting beneath younger fill of the San Luis basin (Manley, 1984). Strata interbedded with basalt along the western margin of the San Luis basin in Colorado onlap and rest with mild angular unconformity on underlying ignimbrites, and exhibit upward decreasing dips indicating eastward tilting of the San Luis basin beginning in late Oligocene (Lipman, 1975a; Lipman and Mehnert, 1975). The unconformity can be traced eastward in the subsurface (Brister and Gries, 1994) and clearly indicates tilting related to formation of the Baca graben.

Basanite and olivine-tholeiitic basalt of the Cieneguilla limburgite (Stearns, 1953) were erupted west-southwest of Santa Fe near La Cienega (Fig. 7) at about 25–26 Ma (Fig. 5, columns F, G; Baldridge et al., 1980; Koning and Hallett, 2001). Basalt flows in the Nambé Member of the Tesuque Formation north of Santa Fe are also about 25 Ma (Baldridge et al., 1980). Enclosing Nambé Member strata are notably richer in granitic detritus than older Tesuque Formation, suggesting reversal of the speculated west-side-down motion on the Picuris-Pecos fault and initial emergence of the Santa Fe Range (Fig. 7; Smith, 2000). Similar-age mafic lava flows are present above the Espinaso Formation south of Santa Fe at Espinaso Ridge (Kautz et al., 1981; Connell et al., 2002), and 26.5 Ma dikes (Erslev, 2001) intrude the Espinaso and older strata near the Tijeras-Cañoncito fault zone (Fig. 2) southeast of Santa Fe. Notably, the Tijeras-Cañoncito fault zone does not displace these dikes although Espinaso Formation is deformed (Lisenbee et al., 1979; Bachman, 1975; Erslev, 2001).

Strongly alkaline magmas were generated during this same time interval east and west of the rift. Most notable are the picturesque Oligocene minette necks of the Shiprock region marking exhumed diatreme conduits with K-Ar ages ranging from about 19 to 28 Ma (Laughlin et al., 1986). McDowell et al. (1986) question the validity of the youngest ages and restrict this period of volcanism to 23–28 Ma. Trachybasalt flows in the Chuska Mountains are about 25 Ma (Cather et al., 2003). Biotite-hornblende-lamprophyre dikes near Dulce (Bingler, 1968; Fig. 1) have K-Ar dates of 23.5 ± 0.9 and 27.2 ± 1.1 Ma (Aldrich et al., 1986) and are coeval with minette volcanism farther west. The north-south-striking Dulce dikes formed over a 9-km-wide zone marking the western margin of small-displacement, west-side-down Cenozoic normal faults that are found throughout the Chama basin (Smith et al., 1961). Oligocene and lower Miocene (18–37 Ma) gold-bearing alkaline intrusive rocks are also found south of Raton and east of the Sangre de Cristo Mountains (McLemore, 1995).

EARLY TO MIDDLE MIOCENE (18–12 MA) PALEOGEOGRAPHY

San Luis basin and Abiquiu embayment

As volcanism waned in northern New Mexico and south-central Colorado the older volcaniclastic aprons were dissected and exposure of Precambrian basement rocks yielded increasing arkosic and quartzitic detritus into basins of the nascent Rio Grande rift (Fig. 8; Manley, 1984). Along the southern flank of the Tusas Mountains the largely volcaniclastic Los Pinos and Abiquiu Formations grade upward into the Chama-El Rito Member of the Tesuque Formation (Galusha and Blick, 1971; May 1984b; Fig. 5, column C). The Chama-El Rito is pink lithic-arkosic sandstone and mudstone interbedded with volcanic conglomerate, and reflects mixing detritus from Precambrian and volcanic source rocks (Steinpress, 1980; Ekas et al., 1984; May 1984a). Small basanitic volcanoes (15.3 ± 0.4 Ma, Ekas et al., 1984) formed on the southward-sloping alluvial plain in the Abiquiu embayment (May, 1984b). Chama-El Rito Member is about 450 m thick in the Abiquiu embayment and is 480 m thick in exposures along the Embudo fault (Steinpress, 1980). Approximately 30 m of Chama-El Rito strata are preserved below ~8 Ma Lobato Basalt flows at Cerro Pedernal, just west of the rift-bounding Cañones fault (Fig. 8). These thicknesses suggest accumulation of strata in subsiding basins although, as during the Oligocene, this deposition continued west of the currently recognized rift margin. Oligocene through middle Miocene strata in the southwestern Abiquiu embayment dip southwestward and southeastward and are overlain unconformably by ~7–10 Ma Lobato Basalt flows that preserve northward gradients away from Jemez Mountains vents (Kelley, 1978; Moore, 2000). This angular unconformity indicates that pre-late Miocene subsidence in the embayment included a component of southward tilting.

Middle and late Miocene deposition within the central San Luis basin is not clearly understood because strata of this age are concealed beneath Taos Plateau basalt. Strata resembling the type Chama-El Rito Member are known in the subsurface (Bauer et al., 1999; Fig. 5, column B) and may comprise a substantial part of the 7–8-km-thick fill of the eastern basin, although there are no data constraining the onset of subsidence of the Taos graben. In the southern basin the Chama-El Rito Member overlies 450–650 m of Bradley conglomerate (Leininger, 1982; Dungan et al., 1984; Fig. 5; column B) composed mostly of quartzite clasts as large as 2 m derived from the adjacent Picuris Range. Chama-El Rito strata overlying the Bradley conglomerate record inundation of the area with mixed volcanic- and basement-derived detritus from the north (Leininger, 1982), possibly burying the early Picuris Range uplift implied by the presence of the Bradley conglomerate. A disconformity between the Bradley conglomerate and Chama-El Rito Member (Leininger, 1982; Dungan et al., 1984) records the reversal of paleoslope during Chama-El Rito deposition.

Middle Miocene strata in the Taos Range reveal deposition in basins east of the present Sangre de Cristo Mountains front. Poorly exposed conglomerate, sandstone, and minor mudstone as much a 500 m thick overlie volcanic and plutonic rocks of the Latir volcanic field and are interbedded with 15–16 Ma basalt flow (Lipman, 1983; Lipman and Reed, 1989). A complicated, and a yet undeciphered, history is required to account for the depositio of these basin-fill sediments upon previously unroofed 24–25 M plutons (Czamanske et al., 1990). The conglomerate, composed c both volcanic and basement clasts, is found at elevations as high a 3300 m southeast of the Questa caldera (Figs. 2, 7) and underlie among other features, the ski slopes at Red River.

The map pattern (Lipman and Reed, 1989) of middle Miocen strata in the Taos Range correlative to the Chama-El Rito Membe suggests structural accommodation in two half grabens. One nai row basin north of the Red River trends northwestward along th Rio Costillo (Cather, 1990), and the other is bounded by a fau zone striking north-northeastward along the east side of Valle Vid and passing northwest of Wheeler Peak (Figs. 2, 8). A northeas striking segment of the fault zone near Wheeler Peak coincid with a Precambrian ductile shear zone (Lipman and Reed, 1989 and footwall uplift of the Wheeler Peak area may have been unde way as early as 35 Ma, based on apatite-fission-track data (Kelle and Duncan, 1984). This middle Miocene fault zone (Fig. 8) para lels the Laramide reverse-faulted eastern margin of the Sangre d Cristo Mountains (Fig. 2), which may also have been reactivate as west-down normal faults. Inversion of Laramide reverse faul

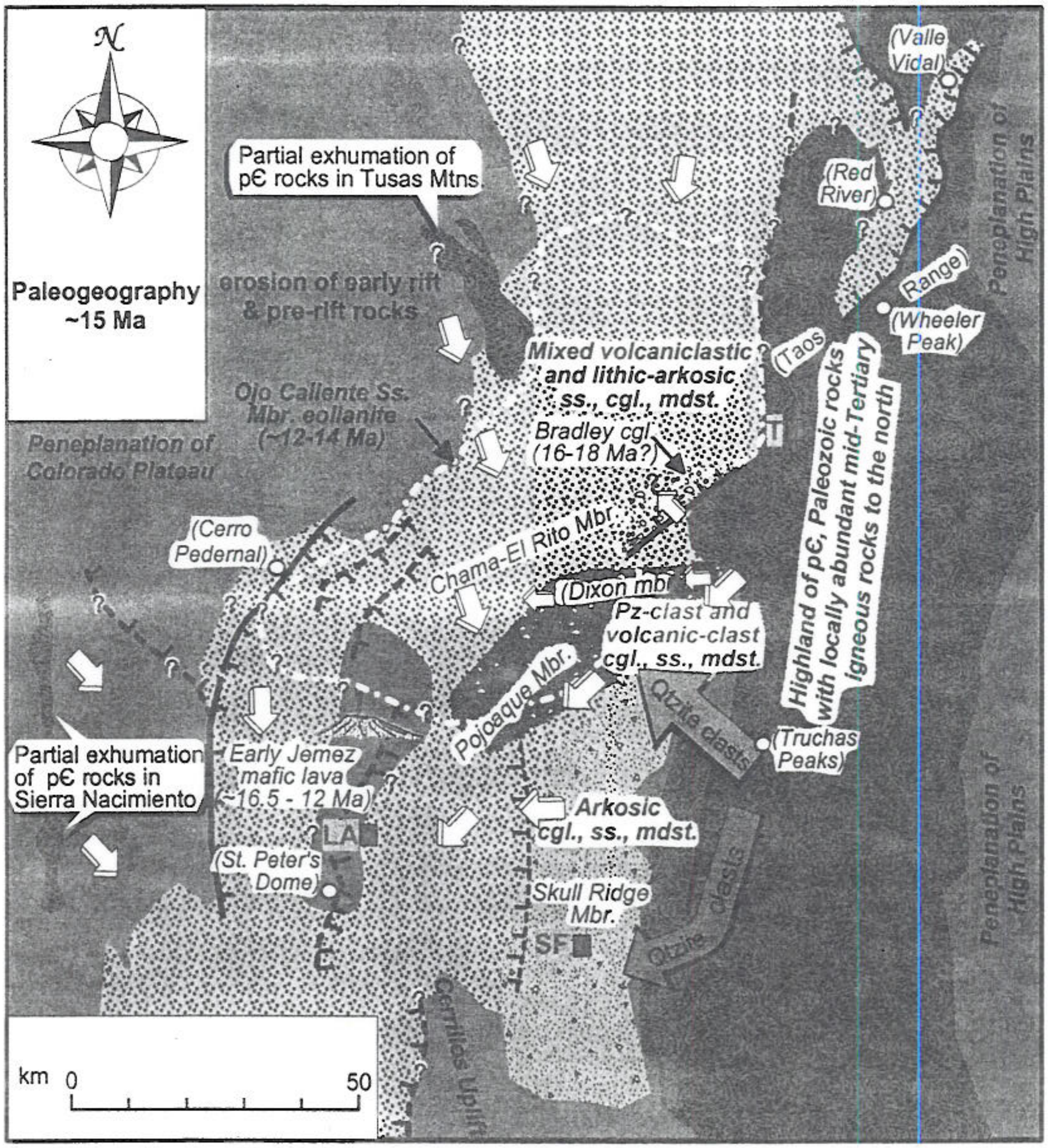

FIGURE 8. Paleogeographic map of north-central New Mexico during middle Miocene.

in Colorado may have accommodated more than 1500 m of conglomerate, of uncertain age, now exposed in the footwall of the Sangre de Cristo fault (Wallace, 1995; Fig. 1).

Mixed volcaniclastic and basement-derived alluvium of the Chama-El Rito Member is widely overlain by eolian strata of the Ojo Caliente Sandstone Member of the Tesuque Formation (Galusha and Blick, 1971; Fig. 5, columns B–E; Fig. 8). It seems unlikely that alluvial deposition was completely replaced by a middle Miocene erg, and in places Ojo Caliente Sandstone appears interbedded with fluvial facies typical of the Chama-El Rito Member or younger Chamita Formation (Dethier and Manley, 1985). Nonetheless, the widespread eolian deposition of this time period provides a useful stratigraphic marker. The Ojo Caliente Sandstone is 140 m thick in its type area west of Cerro Azul (Fig. 1; May, 1980), 200 m thick near Española (Dethier and Martin, 1984), 150 m thick on the north side of the Picuris Range (Leininger, 1982), and as thick as 260 m (Steinpress, 1980) at the west end of the Picuris Range. The quartzo-feldspathic composition of the fine-grained sandstone, which contains only about 10% lithic fragments (Steinpress, 1980), does not lend to easy provenance interpretation. Ingersoll (1990) utilized a canonical-variable analysis of point-count data to interpret a felsic-volcanic provenance for the eolianite and suggested reworking from exposed Chama-El Rito Member. This interpretation is consistent with the overwhelming dominance of volcanic fragments among the sparse lithic-sand grains (Steinpress, 1980) although the presence of chert grains not seen in the Chama-El Rito (Steinpress, 1980) may also indicate contributions from erosion of pre-Tertiary and/or syn-Laramide strata in the Chama basin.

Española basin and Jemez Mountains

Quintessential strata of the Santa Fe Group comprise the pink-hued badlands of the central and eastern Española basin (Fig. 4A). Galusha and Blick (1971) defined the Nambé, Skull Ridge, and Pojoaque Members of the Tesuque Formation as largely correlative with the Chama-El Rito Member but lacking the volcaniclastic components of the latter member (Fig. 5, column F). This distinction is blurred by the recognition of some volcanic component in the Skull Ridge and Pojoaque Members (provenance B of Cavazza, 1986), and the difficulty of distinguishing Chama-El Rito Member and the upper 600 m of Pojoaque Member west of the Rio Grande (Tedford and Barghoorn, 1993). In addition, the lower part of the Nambé Member is clearly the chronostratigraphic equivalent of pre-Tesuque strata in the Abiquiu embayment (Smith, 2000; Fig. 5, columns C, F). Cavazza (1986) and Tedford and Barghoorn (1993) criticized the lithostratigraphy of Galusha and Blick (1971) as inappropriately defined by chronological boundaries. Actually, Galusha and Blick (1971) explicitly point out the lithologic, rather than chronologic, boundaries for the members of the Tesuque Formation, although Kelley (1978) complained that these boundaries are not recognized throughout the basin.

Arkosic strata dominating the of the upper Nambé and Skull Ridge Members record a westward-inclined alluvial slope draining the Proterozoic rocks of the Santa Fe Range (Cavazza, 1989; Kuhle, 1997; Kuhle and Smith, 2001). These strata are on the order of 1300 m thick (Galusha and Blick, 1971; Smith and Battuello, 1990), and dips generally decrease upsection (Kelley, 1978; Golombek et al., 1983, fig. 3). The Skull Ridge Member is distinguished from the Nambé Member primarily by an abundance of reworked white and gray ash beds (Fig. 4A; Galusha and Blick, 1971; Izett and Obradovich, 2001) and finer grain size. These distinctive tephra layers are indistinct near Santa Fe or in the Peñasco embayment where these members are indistinguishable. The reworked ash beds (e.g., Rhoads and Smith, 1995) are in close to the Jemez Mountains but pre-date by 3–4 m.y. the oldest known silicic volcanism in that volcanic field (Gardner et al., 1986). Glass chemistry suggests that these tephra layers are distal fallout from eruptions in southwestern Nevada and along the Yellowstone hotspot track in southwestern Idaho (Perkins et al., 1998; Perkins and Nash, 2002), which is consistent with the very fine size of shards in the Española basin deposits.

The arkosic nature of Tesuque Formation in the east-central Española basin (Cavazza, 1986) indicates elimination of the Paleozoic sedimentary rock source area that contributed prominently to lowermost Nambé Member deposited in the Oligocene (Fig. 5). If the source for Paleozoic clasts was east of the Santa Fe Range, then the shift in sediment composition could represent Miocene uplift of the range (Fig. 8) and structural separation of the Española basin from the areally extensive outcrops of Pennsylvanian limestone farther east (Figs. 1, 2). Neogene uplift of the Santa Fe Range would also have led to erosion of older Tesuque strata originally deposited in that area, as first postulated by Denny (1940). Neogene uplift also explains local discrepancies in apatite-fission-track ages along the Picuris-Pecos fault implying post-Laramide, west-side-up motion (Kelley, 1995). Quartzite clasts in the Tesuque Formation at the latitude of Santa Fe (Miller et al., 1963; Cavazza, 1985; Koning and Maldonado, 2001) may reveal persistent sediment delivery across the Picuris-Pecos fault from the Truchas Peaks and possibly indicating that uplift of the Santa Fe Range progressed as a doubly north and south plunging block (Fig. 8).

Middle Miocene sediment in the Peñasco embayment was derived from erosion of highlands northeast and east of the basin. In the northwestern embayment Chama-El Rito Member alluvium, originating from the north, is separated from younger Ojo Caliente Member eolianite by 400–500 m of conglomeratic sandstone of the Dixon member, dominated by Paleozoic clasts derived from east of the Picuris-Pecos fault (Steinpress, 1980). Paleocurrent data (Cavazza, 1986) indicate that the mixed volcaniclastic and basement-derived alluvium comprising parts of the Skull Ridge and Pojoaque Members in the central Española basin was derived from the northeast by these streams exiting the Peñasco embayment (Fig. 8).

Volcanism in the Jemez Mountains area began with extrusion of middle Miocene mafic lava flows (Fig. 8). These include 14–12 Ma lava flows of the lower Lobato Basalt interbedded with Tesuque Formation west of Española (Dethier et al., 1988; Gardner et al., 1986; Aldrich and Dethier, 1990; Fig. 5, column C). A 16.5 Ma basanite lava flow is present within Tesuque Formation exposed in a strongly uplifted footwall block along a change in strike of the Pajarito fault near St. Peter's Dome (Fig. 8; Gardner et al., 1986; Goff et al., 1990). As noted above and by Cather (1992), an angular unconformity between these middle Miocene strata and underlying Eocene Galisteo Formation, along with the greater antiquity of Tesuque Formation in the central Española basin, suggest Oligocene and early Miocene motion on the Pajarito fault. Accommodation of Miocene strata and lava west of the Pajarito fault implies subsidence along the rift margin farther west (Fig. 8).

Denudation of the High Plains

Alluvial and eolian strata of the Ogallala Formation rest on a widespread unconformity cut on Permian to Cretaceous sedimentary rocks throughout the High Plains of northeastern New Mexico and adjacent Colorado, Oklahoma, and Texas (Fig. 1). Erosion of the sub-Ogallala unconformity post-dates emplacement of ~21–27 Ma intrusions of the Spanish Peaks in southern Colorado (Fig. 1; Kelley and Chapin, 1995), and pre-dates deposition of the Ogallala Formation, which is as old as 10–12 Ma in the Texas panhandle (Winkler, 1987). Apatite-fission-track ages also suggest denudation primarily between 20 Ma and 12 Ma (Kelley and Chapin, 1995). Recalling evidence above for a contiguous apron of Oligocene-lower Miocene sediment across the San Juan Basin, it seems likely that regional denudation occurred east and west of the Rio Grande rift in the early to middle Miocene, possibly removing as much as a 2–3 km thickness of older rocks. The extent to which climate change or epeirogenic processes related to but peripheral to Rio Grande rift extension drove this erosion remains unknown.

LATE MIOCENE TO PLIOCENE (12–2 MA) PALEOGEOGRAPHY

Evolution of the Rio Grande rift basins

Many workers (e.g., Baltz, 1978; Baldridge et al., 1980, Golombek et al., 1983; Ingersoll, 2001; Keller et al., 1991; Chapin and Cather, 1994) suggest that delineation of the present Rio Grande rift basins and adjacent uplifts occurred primarily during and after the middle Miocene. There are, however, insufficient data from the deep parts of the San Luis and Española basins to determine when they developed, or if subsidence and sediment-accumulation rates increased after the middle Miocene. As noted above, eastward tilting of the San Luis basin was underway by 26 Ma and a basin almost certainly had formed in the hanging wall of the Pajarito fault by 30 Ma. Nonetheless, familiar aspects of the rift-related landscape likely did not appear until the late middle to late Miocene.

Erosion of highlands peripheral to the developing rift basins in northern New Mexico yielded decreasing amounts of volcanic detritus and increasing contributions from Precambrian basement (Fig. 9). Chamita Formation alluvium, which overlies and interfingers with Ojo Caliente Sandstone Member eolianites, is distinctly less volcaniclastic than older Chama-El Rito Member (Galusha

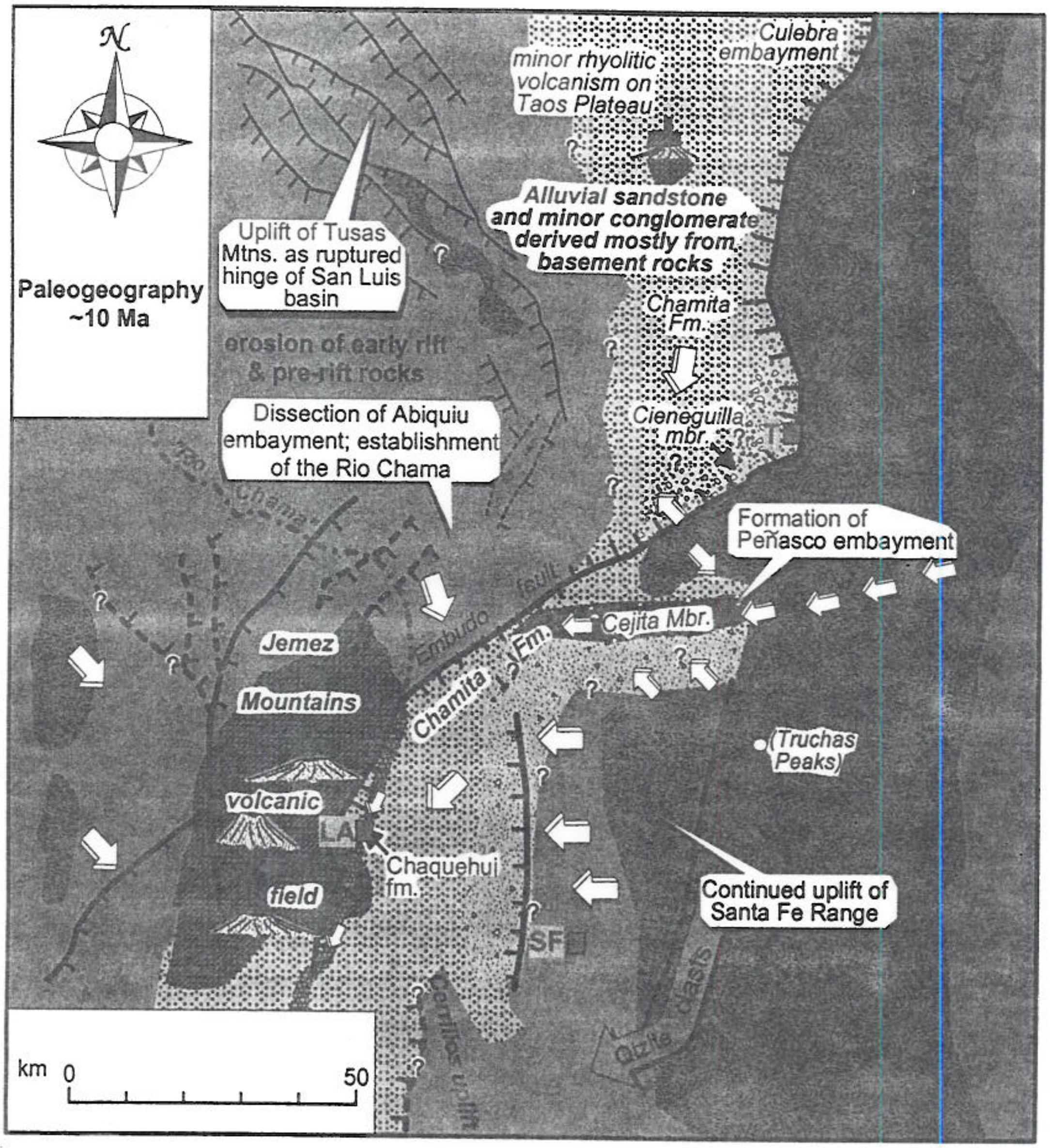

FIGURE 9. Paleogeographic map of north-central New Mexico during late Miocene.

and Blick, 1971; Aldrich and Dethier, 1990; Fig. 5, column E), with the exception of local tephra layers originating from eruptions in the Jemez Mountains (McIntosh and Quade, 1995). In some places the Chamita exhibits an upward decrease in volcaniclastic detritus (Bauer et al., 1999; Fig. 5, column B). Jemez Mountains basalt flows as old as 12.3 Ma are present in the lowermost Chamita Formation northwest of Española (Aldrich and Dethier, 1990), and deposition of the Chamita Formation persisted until after 6.7 Ma (McIntosh and Quade, 1995) and likely close to 5.3 Ma (MacFadden, 1977). The informally designated Cieneguilla member of the Tesuque Formation, subsumed into the Chamita Formation by Bauer et al. (1999), contains a lower part rich in volcanic detritus followed by quartzite clasts shed northward into the San Luis basin from the Picuris Range (Fig. 9; Leininger, 1982). By including the Llama formation of Wells et al. (1987) in the Chamita Formation, Bauer et al. (1999) extend the Chamita in the San Luis basin to upper Pliocene and at least partly correlative with the Alamosa Formation farther north (Rogers, et al., 1992: Fig. 5).

The Embudo fault zone may have been established as a topographic feature during Chamita Formation deposition. The stripping of Chama-El Rito Member and exposure of the Precambrian basement revealed in the Cieneguilla member (Leininger, 1982) represents rejuvenated uplift of the Picuris Range along the eastern trace of the Embudo fault. Except for a few outcrops just west of the Embudo fault, the Chamita Formation is absent in the Abiquiu embayment where upper Miocene Lobato Basalt flows from the Jemez Mountains unconformably overlie middle Miocene and older strata (Fig. 5, column C). Thrust faults in the Tesuque Formation north of the Picuris Range and Taos Plateau basalts are folded along northeast axes implying shortening across the eastern Embudo fault (Muehlberger, 1979), whereas only a few kilometers to the southwest extension formed the northern Velarde graben along the same trend.

Subsidence of the Abiquiu embayment probably diminished during the late Miocene and Pliocene, contrary to arguments by previous workers. For example, Manley and Mehnert (1981) noted that faults in the western embayment exhibit as much as 670-m displacement of 7.6 Ma lava flows and argued that there was no evidence of older movement. Baldridge et al. (1994) interpreted the Abiquiu embayment as an aborted rift margin with subsidence initiated around 10 Ma and ceasing shortly after 7.6 Ma. This argument is based on a presumed correlation between rift initiation and the age of Lobato Basalt volcanism and dike intrusion between 10–7.6 Ma. Several observations are, however, counter to these conclusions. First, stratigraphic study of the Abiquiu Formation demonstrates subsidence of the embayment, relative to the Chama basin, at least as early as the Oligocene (Moore, 2000; Smith et al., 2002). Secondly, approximately 1.1 km of Oligocene-middle Miocene strata accumulated in the Abiquiu embayment (May, 1980) with deposition ending by 12 Ma, suggesting that principal structural accommodation of sediment occurred prior to 10 Ma rather than being initiated at that time. Thirdly, Lobato Basalt and Tschicoma Formation volcanic rocks rest with angular unconformity on the older fill of the embayment (Kelley, 1978; Moore, 2000) demonstrating that principal subsidence with a southerly tilt component was completed prior to the late Miocene, although faulting without significant tilting continued until sometime between 7.6 and 2.8 Ma (Manley and Mehnert, 1981).

Principal uplift of the Taos Range along the Sangre de Cristo fault (Figs. 1–3) occurred after 15 Ma. Earlier basin-fill alluvium was uplifted in the footwall of the Sangre de Cristo fault beginning in the late Miocene and it is possible, though not known with certainty that the principal subsidence of the Taos graben commenced at this time. Flooding of most of the southern San Luis basin with more than 200 m of mafic lava flows of the Servilleta Basalt during the Pliocene (Table 2) effectively buries the older stratigraphic record of the basin and may reflect relatively high strain rates late in the basin history. These basalt flows are displaced by as much as 500 m by continued movement on the Sangre de Cristo fault (Lipman and Mehnert, 1979; Lipman and Reed, 1989). The steep relief of the Taos Range is one of many geomorphic indicators of Quaternary faulting along the east side of the basin (Menges, 1990).

Late Neogene subsidence of the San Luis basin was accompanied by uplift of the Tusas Mountains along steep northwest-striking faults of Laramide, and likely older (Grambling et al., 1988; Karlstrom et al., 1999), ancestry. The San Luis basin in Colorado is hinged on the west side but this hinge zone ruptured along the basin margin in New Mexico, raising the Tusas Mountains on the west as the basin tilted down to the east (Fig. 3). The western extent of this broken hinge zone may include numerous small-displacement, west-side-down normal faults of north and north-northwest strike that displace Cretaceous rocks across the entire width of the Chama basin (Smith et al., 1961). The age of Tusas Mountains uplift to accentuate the western dip slope of the southern San Luis basin is unknown but Pliocene Servilleta Basalt flows were tilted and displaced by movement on these faults (Manley and Wobus, 1982a, b). Pleistocene basalt erupted along and near the western margin of the Tusas Mountains (Fig. 2) is possibly another reflection of the ruptured hinge of the western San Luis basin.

Volcanism constructed the Jemez Mountains astride faults delineating the western margin of the Española basin, primarily between 12 and 3 Ma (Gardner et al., 1986). The near-circular plan form of the volcanic field (e.g., Fig. 2) mostly results from the radial distribution of the Pleistocene Bandelier Tuff ignimbrite outflow sheets surrounding the Valles caldera (Fig. 5, column E), and obscures the fact that most of the high-standing volcanic constructional landforms form a generally north-south ridge. It is inferred that north-south striking rift faults, and intersections of such faults with northeast-striking basement faults of the Jemez Lineament, controlled the locations of these volcanic edifices. Few faults are mapped continuously through the volcanic field (Smith et al., 1970). Nonetheless, local detailed mapping and stratigraphic study in the southeastern Jemez Mountains demonstrate the importance of rift faulting to form basins filled with upper Miocene volcanic and volcaniclastic rocks (Fig. 10; Smith, 2001). Middle and upper Miocene volcanic rocks in the southern Jemez Mountains thicken eastward (Smith et al., 1970; Gardner et al., 1986) and indicate faulting along the western margin of the Española basin beneath the volcanic field synchronous with eruptive activity (Figs. 8, 9).

The Velarde graben within the Española basin is not clearly evident until Pliocene or possibly late Miocene time (Fig. 10). Manley (1976a, 1984) documented 300 m of Pliocene subsidence in the Velarde graben, but the amount of late Miocene subsidence of this deep intrabasinal graben is unknown.

Upper Miocene and Pliocene sedimentary strata (Chamita and Puye Formations) are mostly restricted to the central and western parts of the Española basin (Fig. 10). These basin-fill strata are partly correlative to Pliocene gravel capping erosion surfaces developed in the eastern basin (Manley, 1976a; Smith and Pazzaglia, 1995) demonstrating the focus of subsidence in the hanging wall of the Pajarito fault. Pliocene Ancha Formation mapped northwest of Santa Fe and east of the Rio Grande by Spiegel and Baldwin (1963) have subsequently been assigned to the Miocene Pojoaque Member of the Tesuque Formation (Koning et al., 2002), further emphasizing the restricted nature of upper Miocene and Pliocene deposition. Focus of deposition in the western Española basin and erosion farther east are consistent with strong westward tilting of the basin and continued uplift of the Santa Fe Range (Smith and Pazzaglia, 1995) perhaps as a ruptured hinge zone analogous to the Tusas Mountains along the west side of the San Luis basin (Fig. 3). The Pojoaque fault zone (Fig. 10) may have served as the eastern margin of the Velarde graben because stratal dips increase west of these faults (Golombek et al., 1983, fig. 3) although total displacement on these faults is small (Biehler et al., 1991).

The Pliocene Ancha Formation is a 30–100-m thick alluvial blanket covering a pediment cut on older, tilted rocks of the Santa Fe embayment (Spiegel and Baldwin, 1963; Grant, 1999; Koning et al., 2002). Geomorphic expression (Spiegel and Baldwin, 1963) suggests that the Ancha in the Santa Fe area was mostly deposited by the Santa Fe River and other large streams, possibly in response to Pliocene climate change and increased sediment yield from larger watersheds in the Santa Fe Range. Ancha Formation and age-equivalent high-level strath gravel north of the Santa Fe River contain conspicuous quartzite and Paleozoic clasts foreign to the Santa Fe Range (Table 1). These clasts may record an ancestral Pecos River that entered the Española basin around the southern down plunge extent of the rising Santa Fe Range. In this speculative scenario (Fig. 10) the Santa Fe River is the lower end of a beheaded Pecos River later pirated into the High Plains.

Although the Peñasco embayment began to form in the early or middle Miocene, its present form probably appeared in the late Miocene (Fig. 10). Bedding strikes in the southern basin parallel the Pajarito fault but are more variable in the embayment, so that beds dip northward away from the Santa Fe Range and southward from the Picuris Range (Kelley, 1978; Golombek et al., 1983; fig. 3). The quartzitic gravel of the Cieneguilla member of the Tesuque Formation farther north reveals late Miocene uplift of the Picuris Range. Uplift of the northern Santa Fe Range is inferred from the development of a northwestward sloping piedmont in the southern embayment. The Peñasco embayment was drained westward by a single river, akin to the modern Rio Embudo, marked by the 120-

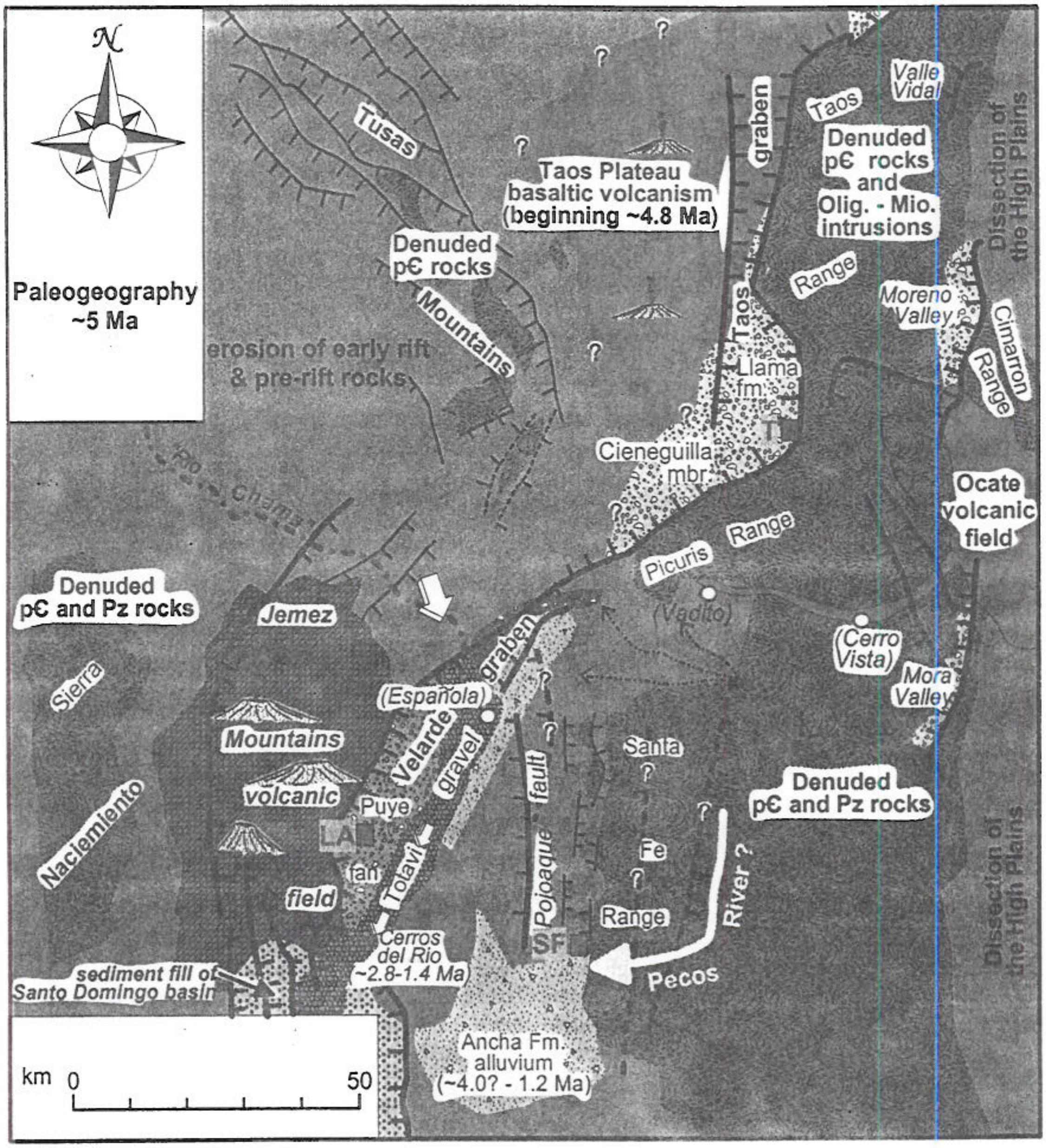

FIGURE 10. Paleogeographic map of north-central New Mexico during latest Miocene and early Pliocene.

m-thick gravel of the Cejita Member of the Tesuque Formation (Manley, 1976a, 1977, 1979b) composed mostly of Paleozoic and quartzite clasts from east of the Picuris-Pecos fault (Fig. 5, column D; Fig. 10). A forerunner of the Rio Quemado-Rio Santa Cruz drainage that currently drains the southern embayment was present by Pliocene time. This drainage is marked by a succession of high-level quartzite-rich Pliocene and early Pleistocene gravel deposits (Manley, 1976a, 1979b; Fig. 5, columns D, F; Fig. 10) delineating an oldest northwest-trending river exiting the Truchas Peaks area that pivoted to a more westerly course through time.

Basins in the Sangre de Cristo Mountains

Despite the strong definition of the San Luis and Española basins during the late Miocene and Pliocene, localized extensional basin subsidence continued farther east along and near the earlier Laramide deformation front. Although post-middle Miocene sediment in the Valle Vidal is probably not very thick (Lipman and Reed, 1989), apatite-fission-track data suggest more recent footwall uplift east of that basin than other parts of the Taos Range closer to the Sangre de Cristo fault (Pazzaglia and Kelley, 1998; Fig. 10). Topographic escarpments along the eastern side of Valle Vidal likely mark Quaternary fault rupture (Menges and Walker, 1990).

New east-tilted half grabens formed the Moreno and Mora Valleys nearly coincident with Laramide reverse faults (Fig. 10). The Moreno Valley formed in the Pliocene with normal faults cutting across older, less steep thrust and reverse faults (Colpitts and Smith, 1990). About 275 m of subsidence of the Moreno Valley occurred after eruption of 5 Ma basalt of the Ocate volcanic field that is displaced along the margin of the valley (O'Neill and Mehnert, 1988, 1990). Baltz and O'Neill (1984) infer inversion of Laramide faults along the east side of Mora Valley to accommodate 150–200 m of Quaternary basin fill.

Upper Miocene basalt erupted in the Ocate field flowed westward into the Rio Grande rift (Fig. 10). A lava flow with an $^{40}Ar/^{39}Ar$ age of 5.65 Ma caps Cerro Vista at 3017 m (Fig. 10;

Olmsted, 2000) and rests on Precambrian-clast gravel that was clearly derived from farther east (O'Neill and Mehnert, 1988). The Mora Valley intervenes between Cerro Vista and the Ocate field, so the subsidence of that minor half graben and incision of streams by more than 750 m must have occurred after 5.65 Ma. A 5.1 Ma basalt (Manley, 1976b), possibly correlative with the Cerro Vista flow, rests on Precambrian-clast conglomerate along the southern margin of the Picuris Range near Vadito (Fig. 10), and presumably marks a more westward position of a river that crossed the entire width of the Sangre de Cristo Mountains during the late Miocene. Subsidence of the Mora Valley after 4.35 Ma (Olmsted, 2000), uplift along Neogene faults within the Sangre de Cristo Mountains, or both, contributed to the defeat of this river course and was the final step in establishing a drainage divide along the entire crest of the Sangre de Cristo Mountains.

Establishment of the Rio Grande

There are no extensive lacustrine or playa facies in the exposed fill of the southern San Luis and Española basins to suggest that they were ever hydrologically closed. The persistent southward transport of volcanic detritus from the San Juan and Latir volcanic fields through the Española basin and farther south (Ingersoll and Cavazza, 1991; Large and Ingersoll, 1996) further indicates a hydrological connection of the northern basins to the Albuquerque basin where extensive playa deposits are known. Nonetheless, the broad, sheetlike character of the Abiquiu Formation, Chama El-Rito Member of the Tesuque Formation, and the Chamita Formation and the generally small dimensions (1–3 m thick) of channel bodies in these strata suggest a south-sloping alluvial plain of many parallel drainages rather than a well-integrated system of tributaries feeding a trunk river.

Rio Grande development is widely cited as beginning in the Pliocene, but actually originated earlier. Bachman and Mehert (1978) suggested that the Rio Grande originated at about 4–5 Ma, based on the apparent timing of development of through-going drainage that exited the Albuquerque basin. In the Española basin, Manley (1979a) suggested that the Rio Grande originated following deposition of the Chamita Formation and was marked by quartzite-rich gravel of the Totavi Lentil of the Puye Formation (Griggs, 1964; Waresback and Turbeville, 1990; Fig. 5, column E; Fig. 10).

Similar gravel marking an axial river of uncertain age is recognized in boreholes drilled below the Puye Formation under the Pajarito Plateau west of the Rio Grande (Purtymun, 1995; Fig. 5, column E; Fig. 9). Purtymun (1995) informally named this older gravel the Chaquehui formation with unfortunate reference to younger, unrelated, maar-crater-fill deposits in Chaquehui Canyon (a tributary to the Rio Grande south of Los Alamos; Broxton and Reneau, 1996) WoldeGabriel et al. (2001) advocate abandonment of Chaquehui formation but do not resolve the stratigraphic assignment of these strata.

Smith and Kuhle (1997) and Smith et al. (2001) corroborated the presence of a late Miocene axial river recorded in quartzite-rich gravel underlying 6.96 Ma volcanic rocks in the southeastern Jemez Mountains (Fig. 10). The upper Miocene gravel clasts are dominantly quartzite, metavolcanic rocks, and less common Pedernal chert and rare Amalia Tuff, all consistent with derivation from the Tusas Mountains and Abiquiu embayment and delivered southward by an ancestral Rio Chama (Smith et al., 2001). To date, however, the equivalent gravel train has not been found in the Chamita Formation west of Española so the details of this early axial drainage remain unclear. It seems likely, however, that an entrenched valley was established by the ancestral Rio Chama as late Miocene subsidence of the Abiquiu embayment failed to keep pace with subsidence southeast of the Embudo fault, (Fig. 10). This river conveyed coarse detritus out of the embayment and into the Española basin, flowing southward in the proximal hanging wall of the Pajarito fault where its deposits are known in the subsurface of the Pajarito Plateau (Purtymun, 1995; Broxton and Reneau, 1996).

It is less clear when the Rio Grande was established upstream of Española. There is no indication of a single, major southwestward-flowing river preserved in the type area of the Chamita Formation north of Española, although these strata have not been subject to detailed sedimentological study. A single flow of Servilleta Basalt did follow a narrow valley southwestward along the Embudo fault zone at about 2.8 Ma (Manley, 1976a; Fig. 2). Headward integration of the Rio Grande northward into the northern San Luis basin of Colorado likely did not occur until middle Pleistocene (Wells et al., 1987), and the extreme northeastern San Luis basin remains hydrologically closed.

The Puye Formation, consisting of alluvial-fan deposits shed from the Jemez Mountains and ancestral Rio Grande gravel within the Totavi Lentil (Griggs, 1964; Waresback and Turbeville, 1990), is restricted to the proximal hanging wall of the Pajarito fault. Subsidence of the western Española basin likely caused aggradation of these deposits (Waresback and Turbeville, 1990; Gonzalez and Dethier, 1991; Gonzalez, 1995). Alternatively, Reneau and Dethier (1996) suggest that Puye Formation deposition was driven by base-level rise upstream of the Cerros del Rio volcanic field, where 300 m of lava flows accumulated in the short interval between 2.5 and 2.4 Ma (Wolde Gabriel et al., 1996). While Cerros del Rio volcanism certainly affected the Rio Grande, as indicated by lacustrine mud in the Puye Formation (Waresback and Turbeville, 1990), the restriction of the Puye alluvial-fan gravel, more than 140 m thick, southeast of the Pajarito-Embudo fault system, and its absence in the Abiquiu embayment implicates subsidence in the central Española basin, as well.

There is also ambiguity regarding an ambiguous unconformity at the base of the Puye Formation (Fig. 5). Reneau and Dethier (1996) use subsurface data to interpret 255 m of relief on this unconformity, although part or all of that relief might be explained by tilting of beds rather than erosion. Although an unconformity is locally exposed between ~9 Ma Chamita Formation and 5.3 Ma Puye Formation (WoldeGabriel et al., 2001), the Chamita elsewhere is at least as young as 5.3 Ma (McFadden, 1977) and WoldeGabriel et al. (2001) assign ~9 Ma lava and volcanic gravel beneath Los Alamos to the Puye Formation. A regional hiatus between Chamita and Puye deposition is, therefore, doubtful and the exposed unconformity may be associated with intrabasin fault blocks. Regardless, the oldest Puye Formation (WoldeGabriel et al., 2001) predates Cerros del Rio volcanism by at least 2 m.y. precluding aggradation solely related to basalt extrusion downstream.

Ultimate incision of the Rio Grande and its tributaries occurred principally after eruption of the upper Bandelier Tuff at 1.2 Ma. Although the lower Bandelier Tuff (1.61 Ma) fills deep Rio Grande paleocanyons cut in Cerros del Rio basalt (Smith et al., 1970; Dethier, 1997), this is a local phenomenon where the river incised through the rapidly accumulating lava flows and the footwall uplift of the La Bajada fault (Fig. 2). Waresback (1986) describes a low-angle erosion surface at the base of the lower Bandelier Tuff (1.61 Ma), which causes progressive eastward truncation of the uppermost beds of the Puye Formation. The Puye, however, also includes latest Pliocene tephra related to eruption of the San Diego

Canyon ignimbrites in the southwestern Jemez Mountains (Turbeville et al., 1989; Turbeville, 1991). These eruptions have been variously dated at ~1.78 Ma (Spell et al., 1990) and 1.84–1.87 Ma (Smith et al., 2001), which precludes a significant hiatus for erosion prior to Bandelier Tuff eruption. There is no compelling evidence for Pliocene or early Pleistocene disconformities within the basin-fill succession anywhere else within the deepest parts of Rio Grande rift basins. It is unclear what combination of basin subsidence and volcanism, including possible magmatic tumescence of the area surrounding the Jemez Mountains volcanic field, may have contributed to producing the mild unconformity in the Española basin below the Bandelier Tuff.

Younger Pleistocene deposits, post-dating the Santa Fe Group in the sense of Spiegel and Baldwin (1963), are inset as terraces along most major perennial drainages. A correlative sequence of fill terraces along the Rio Grande and Rio Chama arguably relate to Quaternary climatic variability (Dethier et al., 1988; Gonzalez and Dethier, 1991; Gonzalez, 1995) with a long-term net incision rate ~0.1 mm/yr (Dethier et al., 1988).

Deposition, volcanism, and renewed denudation on the High Plains

Early to middle Miocene denudation of the High Plains of northeastern New Mexico was followed by deposition of the Ogallala Formation. Initial aggradation occurred in paleovalleys as much as 70 m deep (Frye et al., 1978; Gustavson, 1996). These oldest Ogallala strata contain Clarendonian vertebrates in west Texas (Winkler, 1987) and locally underlie 9 Ma lava flows in northeast New Mexico (Stroud, 1997). The Ogallala contains no fossils younger than Hemphillian, and given evidence of incision along the east slopes of the Sangre de Cristo Mountains after 5.7 Ma (O'Neill and Mehnert, 1988), it seems likely that deposition ceased in the late Miocene. The lower parts of the paleovalley fills are alluvial facies followed upward by eolian sand and silt that also form an extensive blanket on broad interfluves (Gustavson, 1996). Chapin and Cather (1994) suggested that Ogallala Formation deposition in northeastern New Mexico was initiated with increased sediment flux caused by high rates of footwall uplift in the Sangre de Cristo Mountains, although uplift during and after Ogallala deposition may have been more general throughout the Rocky Mountain region (McMillan et al., 2002). Deposition on the east flank of the Sangre de Cristo Mountains produced a graded geomorphic surface, remnants of which remain as a gangplank along the New Mexico-Colorado border extending up to 3140 m elevation in the Sangre de Cristo Mountains. Deposition ceased and denudation resumed when footwall uplift is hypothesized to have diminished in the early Pliocene (Chapin and Cather, 1994) although climatic change during the Pliocene may also have played a role in determining the geomorphic evolution of the region (Stroud, 1997).

The post-Ogallala denudation history of the High Plains of northeastern New Mexico is known from the relationship of erosion surfaces and the Ogallala Formation to lava flows of the Ocate and Raton-Clayton volcanic fields (Fig. 1; Table 2). These cinder cones and basaltic flows, with lesser volumes of more evolved lava (Stormer, 1972; Calvin, 1987; O'Neill, 1988; O'Neill and Mehnert, 1988; O'Neill and Mehnert, 1990; Stroud, 1997; Olmsted, 2000), formed along the Jemez Lineament starting at about 9 Ma and include features as young as the 60 ka Capulin Mountain cinder cone and flows (Fig. 1; Stroud, 1997). Ocate basalts are associated with three general, broad geomorphic surfaces cut on Mesozoic rocks. Lava with ages of 8.2–5.7 Ma, 5.13–4.35 Ma, and 3.36–2.85 Ma, respectively, overlie these surfaces, at progressively lower elevations (O'Neill and Mehnert, 1988; O'Neill and Mehnert, 1990; Olmsted, 2000)). Therefore, principal erosional stripping of the western Raton basin to produce the pre-Ogallala peneplain was over before 8.2 Ma and base-level lowering was underway by ~5.0 Ma with significant incision after 3.0 Ma. Raton-Clayton basalt flows cap mesas at various elevations with the oldest lava flows, ~9 Ma, resting on Ogallala Formation. Most of the post-Ogallala incision in this area has occurred since 3.6 Ma and erosional relief increases westward with 3.5 Ma basalts near present base level in the east but 490 m above base level near Raton (Stroud, 1997). Along the Canadian River (Fig. 1) incision has likely been enhanced by late Cenozoic uplift (Wisniewski, 1999; Wisniewski and Pazzaglia, 2002) driven by mantle upwelling along the Jemez Lineament (Karlstrom and Humphreys, 1998).

DISCUSSION

Age and phases of Rio Grande rifting

There is a lack of consensus about the initiation of extensional basins in northern New Mexico. Extension is viewed as beginning in late Oligocene by a number of workers (Lipman and Mehnert, 1975; Lipman, 1983; Baldridge et al., 1980; Chapin, 1988; Chapin and Cather, 1994), but Ingersoll et al. (1990) and Large and Ingersoll (1997) point out that very little is known about actual faulting and associated tilting of basin-fill sediment during the late Oligocene and early Miocene. Some documented early deformation is restricted to the close proximity of major magmatic centers in the Latir volcanic field (Lipman, 1983) and in the central Rio Grande rift near Socorro (Chamberlin, 1983), which Ingersoll (2001) dismisses as local strain associated with thermal weakening of the crust. Oligocene sedimentation in northern New Mexico clearly persisted beyond the rift boundaries and without clear designation of rift-bounding structures (Fig. 6), which further suggests the lack of widespread rift faults at that time (Baltz, 1978; Morgan and Golombek, 1984). Ingersoll et al. (1990) suggest that the widespread Oligocene volcaniclastic aprons (Los Pinos, Abiquiu and Picuris Formations) aggraded in response to volcanism in the San Juan and Latir volcanic fields and are not indicative of rift-basin formation, which may not have begun until as late as 21 Ma (Large and Ingersoll, 1997; Ingersoll, 2001). This latter view is, however, inconsistent with stratigraphic and subsurface data that reveal eastward tilting of the San Luis basin in Colorado (Lipman and Mehnert, 1975; Brister and Gries, 1994), and the commonly held view (e.g., Chapin, 1988, Baldridge et al., 1995) that widespread basaltic volcanism beginning at 25–27 Ma heralded lithospheric extension as also acknowledged by Ingersoll et al. (1990).

Smith et al. (2002) examined the role of volcanism to drive sedimentation in northern New Mexico by three mechanisms. These are (1) broad basin formation by flexure of the crust beneath the weight of the growing volcanic pile, (2) aggradation of fluvial channels and floodplains under the impact of frequent introduction of voluminous pyroclastic sediment loads, and (3) down-stream base-level rise driven by increase in headwater elevations within growing volcanic fields. Many critical variables required for rigorously assessing these three mechanisms are not readily deciphered for the mid Tertiary, but the model results of Smith et al. (2002) suggest that deposition of the Los Pinos-Abiquiu volcaniclastic apron, for example, may not require accommodation within fault-bounded basins but could be caused by broad-wavelength flexural subsidence. Nonetheless, the same study documents subsidence of the Abiquiu embayment relative to the Colorado Plateau along faults as indicated by lateral variation in the thickness of traceable intervals within the Abiquiu Formation.

The relative importance of volcanism and rift tectonics to cause Oligocene and early Miocene sedimentation remains uncertain, but a considerable body of data suggests that basin subsidence was underway in the northern Rio Grande rift by at least 30 Ma. Not all sedimentation of this vintage is volcaniclastic in composition (Figs. 5, 6). Some or all deposition of basement-derived gravel of the Ritito Conglomerate and lower members of the Abiquiu and Picuris Formation might, speculatively, be related to well-documented global climate change in the early Oligocene, accommodation in distant flexural basins loaded by volcanic fields farther north, or both. Aggradation of hundreds of meters of lower Nambé Member of the Tesuque Formation is not well accounted for by these mechanisms, however, both because of the greater thickness of the strata compared to the lower member of the Abiquiu Formation and correlatives, and because of the relatively distant position from volcanic loads. The accumulation of these strata on the older Laramide highland suggests structural inversion (Cather, 1992), although movement on faults at this time is only inferred (Fig. 6) and not unequivocally demonstrated.

The formation of an Oligocene Española basin is consistent with evidence from basins to both the north and south. More than 1200 m of Oligocene Conejos Formation is revealed by drill-hole and seismic data in the northwestern San Luis basin, and seismically imaged faults penetrate these strata but do not continue into overlying rocks (Brister and Gries, 1994). As much as 2185 m of Oligocene volcaniclastic sediment are present near the center of the northern Albuquerque basin (Lozinsky, 1994). The great thickness of this section, nearly 5 times greater than underlying syn-Laramide Galisteo Formation and nearly half as thick as overlying Santa Fe Group, which accumulated over about twice as much time, strongly imply basin subsidence to accommodate these strata. Most of the rocks associated with these early depressions in the San Luis, Española and northern Albuquerque basins are known only in the subsurface, so it is not possible to define with certainty the boundaries of the basins and the nature of the faults that constrain them. The early depressions in the San Luis and northern Albuquerque basins are coincident with Laramide-related basins and do not coincide with the deepest parts of the Neogene basins (Brister and Gries, 1994; Lozinsky, 1994). In contrast, the Oligocene Tesuque Formation accumulated atop a Laramide tectonic highland (Cather, 1992), which implies tectonic inversion and likely extension.

Post-Laramide stress regimes and basin formation

Many previous workers endorse a notion originating in the early 1980s (e.g., Baldridge et al., 1980; Morgan and Golombek, 1984; Morgan et al., 1986) that opening of the Rio Grande rift is divided into two distinct periods, associated with different stress regimes (Aldrich et al., 1986). An initial phase of deformation, from perhaps 30 Ma (or earlier) to 20–18 Ma is traditionally viewed to form broad, shallow basins, partly bounded by low-angle normal faults. This episode of tectonism is associated with volcanism and is attributed to thinning of hot lithosphere with a shallow brittle-ductile transition (Morgan and Golombek, 1984; Keller et al., 1991; Baldridge et al., 1995). Notably, such low-angle faults are only known from isolated areas near Taos and Socorro, and the nature of regional Oligocene deformation remains cryptic. A later rifting phase, perhaps beginning as late as 10 Ma (Keller et al., 1991), is associated with classic Basin-and-Range-style block faulting and the delineation of the present interconnected rift basins and flanking highlands (Morgan and Golombek, 1984; Chapin and Cather, 1994; Baldridge et al., 1995; Ingersoll, 2001).

Aldrich et al. (1986) defined the stress regimes associated with these two phases of rifting. Many faults and dikes are associated with reactivated structures and their orientations alone, in the absence of kinematic data on fault surfaces, are insufficient to resolve stress orientations. Aldrich et al. (1986) focused on dike orientations where isotopic dates also provide an assessment of age and argued, despite the complexity of intrusion along inherited trends, that the late Oligocene-early Miocene extension direction was oriented northeast-southwest and that the latter phase of deformation was associated with west-northwest/east-southeast extension. Aldrich et al. (1986) and Chapin and Cather (1994) point to early Miocene dikes in the roofs of plutons as being the best indications of regional stress orientation, but these are only known at two locations in the rift and do not comprise a robust regional data set.

The current understanding of the stratigraphy and structure of the northern New Mexico part of the rift does not clearly indicate two distinct phases of deformation. Although this traditional notion may well be correct, paleogeographic reconstructions (Figs. 5–9) do not obviously record pulses of deformation. In fact, the period of tectonic quiescence between ~18–10 Ma interpreted by Keller et al. (1991) is coincident with accumulation of about 1.3 km of Tesuque Formation in the central and eastern Española basin (Galusha and Blick, 1971; Smith and Battuello, 1990), presumably by subsidence along faults in the western basin. Without detailed information on the age and thickness of strata in the deeper parts of the San Luis and Española basins, it is not possible to test the hypothesis of two deformation pulses. Although local unconformities exist, the overall conformity of the lower Oligocene (in places upper Eocene) to uppermost Miocene or lower Pliocene stratigraphy, which in part accounts for the difficulty in defining boundaries for the Santa Fe Group, does not suggest periods of deformation separated by tectonic inactivity. The concept of two distinct phases of deformation has probably been biased from observation of an apparent magmatic lull between ~17–18 Ma and ~10–12 Ma (e.g., Baldridge et al., 1995) more than it can be substantiated from the stratigraphic records of basin subsidence.

Another traditional view is that opening of the Rio Grande rift involved left slip on north-striking rift-bounding faults (Kelley, 1977, 1979), possibly associated with clockwise rotation of the Colorado Plateau as a rigid block (Chapin and Cather, 1994). Kelley (1977, 1979) based his interpretation of left-oblique extension on a variety of large-scale features of the rift, including the conspicuous right-stepping nature of basins in northern New Mexico and local observations of slickenlines indicating a component of left slip on north-striking faults. Muehlberger (1979) suggested that along-strike change from compression to extension along the left-lateral Embudo fault resulted from counterclockwise rotation of a large crustal block to the south that includes the Española basin. Paleomagnetic data (Salyards et al., 1994) show that a single rigid block does not exist but that variable counterclockwise rotations are prominent in Miocene fill of the Española basin and consistent with left-oblique shear along north-striking faults. Chapin and Cather (1994) point out that there are more data supporting left-slip and southward translation of the Colorado Plateau among rocks older than 10 Ma, and that there is no clear indication of stress regime and relative motion of the Plateau for the last 5 m.y. If the stress regime has changed relatively recently, kinematic indicators on fault surfaces may reflect a confusing mix of strain, as pointed out for example by May (1979) for the Abiquiu embayment, and fail to provide convincing supporting evidence for the dominant importance of left-oblique extension in the rift.

Kinematic data from small structures in Colorado and northern New Mexico complicates traditional views of the stress regimes responsible for forming the rift. Wawrzyniec (1999) points to plate-motion studies (e.g., Engebretson et al., 1985) showing dex- · shear along the west margin of North America since the Late ·taceous. If this plate-margin shear sense is conveyed deep into the interior of the North American plate, then one could hypothesize right-lateral displacement along Laramide and rift faults with north-south strike, and northward translation of the Colorado Plateau. Data collected by Wawrzyniec (1999) and Erslev (2001) are consistent with northward translation of the Plateau during late Laramide deformation. Furthermore, Wawryzniec (1999) presents kinematic data for rocks in Colorado, ranging in age from 13–32 Ma, indicating a northwest-southeast-oriented extension direction. ··is direction, contrary to the results of Aldrich et al. (1986) focus-· on dike orientations, is consistent with continued northward, dextral translation of the Colorado Plateau, rather than sinistral motion on north-striking structures along the eastern Plateau margin as envisioned by Chapin and Cather (1994). The presence of east-west normal-fault-bounded basins in northern Colorado is, however, consistent with the Chapin and Cather model (1994) because the Wawrzyniec (1999) scenario requires north-south compression.

Erslev (2001) also offer insights into the tectonic transition ·ween Laramide compression and later extension. He presents ·idence from changing orientations in fault slickenlines cutting rocks of different ages that the principal compressive stress direction in northern New Mexico rotated in a counterclockwise sense from east-west to about N45°E during late Laramide deformation to N10°E by the mid Tertiary and finally to north-south. The minimum stress direction simultaneously rotated from vertical to east-west. In this scenario, mid-Tertiary faulting and basin formation could be a dextral transtensional phase in transition from Laramide compression to rift extension as a continuum of deformation ·Vawrzyniec, 1999; Erslev, 1999, 2001). Dextral motion on north-·uth faults is demonstrated to affect post-25 Ma strata of the upper Picuris Formation (Bauer et al., 1999; Erslev, 1999), upper Eocene and Oligocene rocks south of Santa Fe but not 26.5 Ma dikes (Erslev, 1999; G. Smith, *unpubl.*), and 24 Ma intrusions in northeastern New Mexico (Erslev, 1999). Compression across north-south faults is recorded in lower Picuris Formation strata younger than 34.5 Ma (Bauer et al., 1999).

All of these observations are inconsistent with northeast-southwest-oriented extension during late Oligocene and early Miocene ·ldrich et al., 1986) and are suggestive of right-lateral, northward ·anslation of the Colorado Plateau relative to the rift, contrary to the model of Chapin and Cather (1994). Further work is required to sort out these conflicting views or to resolve the discrepancy with better resolution of time varying, or spatially varying, stress orientations. It is possible, however, that there was a continuum of deformation from Eocene and earlier compression, through Oligocene-early Miocene transtension, to later Neogene extension. Transition from transtension to tension, along with variations in the intensity of magmatism may help explain the two-phase history of ·ift-basin evolution envisioned by some workers.

Relation of Middle and Late Cenozoic faulting to Laramide deformation

The formation of Oligocene and younger sedimentary basins in northern New Mexico shows a close correspondence to earlier Laramide deformation. Although most attention centers on the axial basins of the rift, it is notable that Neogene faulting persisted across the entire width of the Laramide compressional welt. Close correspondence of Laramide structures and younger faults suggests that rifting in northern New Mexico either reflects gravitational collapse of the older Laramide highland or was strongly focused there because of inherent pre-existing crustal weakness or heterogeneity in the mantle (e.g., Karlstrom and Humphreys, 1998).

The close spatial relationship between the two periods of deformation is particularly evident at the latitude of the San Luis basin. Neogene faults reactivated Laramide faults, themselves of older heritage, in the Tusas Mountains on the west. Half grabens near and south of Valle Vidal and forming the Moreno and Mora Valleys formed nearly coincident with the reverse faults defining the eastern margin of the Brazos-Sangre de Cristo uplift (Figs. 1, 3).

The relationship between Laramide structure and younger faulting at the latitude of the Española basin is more speculative. Cather (1992) interprets the Pajarito fault as a west-vergent reverse fault during the Laramide orogeny that was inverted as a normal fault in the Oligocene or early Miocene. That interpretation is consistent with data compiled here, although contested by some (e.g., Grant, 1999). Post-Laramide motion of the Picuris-Pecos fault is equivocal. Stratigraphic relationships in the Picuris Range preclude movement following deposition of the upper member of the Picuris Formation (Montgomery, 1953; Bauer and Ralser, 1995). Fission-track data indicate relatively recent denudation of the Santa Fe Range west of the fault, however, and in some areas imply ~400 m of post-Laramide, west-side-up displacement (Kelley, 1995). Speculative interpretation of sediment sources for the Nambé Member of the Tesuque Formation suggest a complicated history of west-side-up motion during the Laramide (likely accompanied by dextral slip; Cather, 1992, Bauer and Ralser, 1995) followed by west-down and then again west-up motion during the Oligocene and Miocene (Smith, 2000). This complicated history of reactivation, if substantiated, may well be favored by the near-vertical orientation of the fault (Bauer and Ralser, 1995) facilitating motion in a variety of stress-orientation regimes. West of the Española basin it is notable that the Sierra Nacimiento-Gallina arch trend of Laramide ancestry is marked by considerable, low-magnitude historic seismicity (Wong et al., 1996) and evidence of minor Quaternary offsets (Formento-Trigilio and Pazzaglia, 1998).

Ruptured hanging wall hinge zones

Although both the San Luis and Española basins are half grabens, they feature high-relief mountain ranges rising to elevations in excess of 3000 m on the distal hanging-wall margins of the basins (Fig. 3). In the case of the San Luis basin, a gently hinged basin floor in southern Colorado gives way southward to the Tusas Mountains, uplifted in the Neogene along northwest-striking faults of Laramide and older ancestry. These faults principally face away from the basin and represent a rupture of the hanging-wall hinge along prior weakness. Primary motion on these faults followed deposition of the Los Pinos Formation, perhaps beginning in the middle Miocene (Fig. 8) and continuing into the Pliocene.

A similar rupture of the distal hanging wall of the Española basin is proposed to cause uplift of the Santa Fe Range. This hypothesis is inferred from fission-track data implying Neogene denudation of the Santa Fe Range (Kelley et al., 1992; Kelley and Chapin, 1995) but is not as clearly evident as in the Tusas Mountains because of a lack of mid Tertiary strata in the Santa Fe Range to more precisely record the deformation history (Figs. 2, 3). If Paleozoic clasts in the lower Nambé Member were derived from east of the Pecos-Picuris fault, then the abrupt up-section near-disappearance of these clasts in lower Miocene strata could record the onset of uplift of the central Santa Fe Range, effectively isolating the central Española basin from a source of abundant Paleozoic-rock clasts and forming the Pecos River Valley (Smith, 2000; Fig. 1).

Uplift of the Santa Fe Range along the Picuris-Pecos fault (Fig. 3) may have progressed gradually northward and southward from the central part of the range (Figs. 7–10). Quartzite clasts from the Truchas Peaks area were still delivered to the central Española basin during the middle Miocene (Cavazza, 1985), and via a postulated westward-flowing Pecos River farther south (Figs. 7–9). The Peñasco embayment was not fully formed as an isolated recess of the Española basin until late Miocene (Fig. 9), and is part of the original eastern Española basin that has not been denuded of basin fill by Santa Fe Range uplift. This scenario is consistent with evidence for Neogene motion on the southern Picuris-Pecos fault (Kelley, 1995) and lack of post-early Miocene movement on the northernmost expression of the fault (Bauer and Ralser, 1995).

Rupture of hanging-wall hinges is not addressed in treatments of extensional basins. In northern New Mexico, this process may have accentuated the tilting of strata in half grabens, focusing recent deposition to narrow belts adjacent to the basin master fault (Fig. 9). Rupture of hanging-wall hinges also produces important sediment sources in mountain ranges uplifted on faults that faced away from, rather than toward, the principal basins. Rupturing of the basin hinges was probably favored by pre-existing steep faults that were favorably oriented to cause failure during tension.

CONCLUSIONS

Neogene extension in northern New Mexico was likely part of a continuum of deformation from the Laramide orogeny and exploited structures produced by previous faulting. Sedimentary records implying basin subsidence along the present trend of the axial basins of the Rio Grande rift go back to at least 30 Ma, and arguably 35 Ma, suggesting an onset of basin formation that does not significantly post-date Laramide deformation. Emphasis on rift faulting beginning much later (e.g., Large and Ingersoll, 1997; Ingersoll, 2001) does not adequately dismiss the evidence for earlier basin formation. Rather than distinct phases of early Tertiary compression and late Tertiary extension, deformation in this region may have been continuous or nearly so, with latest Eocene and Oligocene transtension being the heretofore unappreciated missing link (Erslev, 1999, 2001). Further work is needed to understand the nature of temporally, and perhaps spatially, varying stress orientations during and following the Oligocene transitional period. Although the deepest Oligocene and Neogene basins delineate the Rio Grande rift, proper, normal faults and shallower basins persist across the entire width of the older Laramide welt suggesting a close relationship between Laramide crustal thickening and subsequent extension. Faults active during the Laramide, and in many cases of even older ancestry (e.g., Karlstrom et al., 1999), were important during middle and late Cenozoic extension. The Pajarito fault may be an inverted Laramide reverse fault (Cather, 1992). Steep Laramide and older faults in the Tusas Mountains, and possibly in the southern Sangre de Cristo Mountains, ruptured the hinges of tilting half grabens and caused uplift of hanging wall ramps to emphasize the asymmetry of the deepest and most prominent half grabens.

Although depocenters shifted and narrowed through time (Figs. 6–10) there is no compelling evidence of discrete episodes of bas formation, coinciding with shifts in principal-stress orientatio during evolution of the northern New Mexico part of the R Grande rift. Oligocene and early Miocene sedimentation w strongly influenced by volcanism in the San Juan, and especia Latir, volcanic fields. Volcaniclastic-sediment aprons were r constrained by known faults (Baltz, 1978; Figs. 6, 7) and lik extended across broad expanses of the San Juan Basin and perha the High Plains. Nonetheless, evidence for syndepositional faulti (Moore, 2000) and local thick depocenters indicate that bas were forming at that time. The present arrangement of deep bas and adjacent highlands was not evident before 10–15 Ma (Fi 8–10). The older volcanic and volcaniclastic rocks were denu from the most strongly uplifted areas, leading to an increasing c tribution of basement-derived detritus in younger strata (Fig. The conformity of sedimentary sections and gradational conta between many units of middle and late Cenozoic age appear contradict earlier models for discrete phases of rift-basin format (e.g., Baldridge et al., 1980, 1995; Morgan and Golombek, 19 Morgan et al., 1986). Sedimentation patterns revealed by surf sections suggest overall narrowing of basins through time, cul nating in upper Miocene and younger strata being largely restr ed close to the master fault (Figs. 6–10) perhaps because of u of distal hanging-wall ramps.

Early Miocene and younger deposits record a gradual prog sion from a broad, southward-sloping alluvial plain to the de opment of a through-going Rio Grande drainage. The Rio Ch was the forerunner of the Rio Grande, collecting drainages f the Chama basin and Tusas Mountains and entering the Vel graben near Española. Three processes likely favored the fo tion of this drainage. First, post-10 Ma subsidence of the Abi embayment was less than that east of the Pajarito and Emb faults, as indicated by an unconformity between ~10 Ma b flows and older rocks in the embayment. This would favor inci of drainages in the Abiquiu embayment and Chama basin, and cy and elongation of drainage basins. Second, growth of the Je Mountains volcanic field, especially following 10 Ma (Gardn al., 1986), favored collection of drainages north of the vol field into a single southeastward-flowing river. Third, deep s dence of the Velarde graben at least by late Miocene, if not ea would favor focusing of drainage proximal to the footwall o Pajarito fault. A narrow gravel train delineates an ancestral Grande prior to 7 Ma beneath the Pajarito Plateau reflectin formation of this primary trunk stream. One of the principal t taries of the early Rio Grande was a drainage that can be ref to as the ancestral Rio Embudo. Marked by distinctive depos the Dixon and Cejita Members of the Tesuque Formation drainage along the south base of the Picuris Range apparentl headwaters east of the modern Sangre de Cristo crest as late a Ma when basalt flows from the Ocate volcanic field follow course into the Española basin (Fig. 10).

Incision of the Rio Grande and its tributaries, perhaps begi discontinuously in the early Pliocene and in unambiguous e in the early Pleistocene, is not the only record of impressiv sion in northern New Mexico; widespread denudation of the Plains of northeastern New Mexico happened at least twic first period of erosion is reflected by the widespread erosio face that underlies the Ogallala Formation and age-equivale flows. The unconformity beneath these rocks, in conjunctio apatite-fission-track dates recording denudation (Kelle Chapin, 1995), suggests erosional removal of as much as 2 rock primarily between 20 Ma and 12 Ma. Deposition

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APPENDIX — SOURCES OF DATA FOR CHRONOSTRATIGRAPHIC COLUMNS (FIG. 5)

Column A
- Brister and Gries, 1994
- Lipman and Mehnert, 1975
- Lipman , 1975a, b
- Lipman et al., 1996
- Rogers et al., 1992
- Wallace, 1995

Column B
- Appelt, 1998
- Bauer et al,. 1999
- Czamanske et al., 1990
- Dungan et al., 1984
- Johnson et al., 1989
- Leininger, 1982
- Lipman, 1989
- Rehder, 1986

Column C
- Aldrich and Dethier, 1986
- Bingler, 1968
- Butler, 1971
- Manley, 1981
- Manley and Mehnert, 1981
- Moore, 2000
- Smith, 1938

Column D
- Manley, 1976b, 1979
- Rehder, 1986
- Steinpress, 1980
- Tedford and Barghoorn, 1993

Column E
- Galusha and Blick, 1971
- Griggs, 1964

Column E (cont.)
- MacFadden, 1977
- McIntosh and Quade, 1995
- Purtymun, 1995
- Reneau and Dethier, 1996
- Tedford and Barghoorn, 1993
- Woldegabriel et al., 2001

Column F
- Galusha and Blick, 1971
- Koning et al., 2002
- Read et al., 2000
- Spiegel and Baldwin, 1963
- Tedford and Barghoorn, 1993

Column G
- Koning et al., 2002
- Spiegel and Baldwin, 1963
- Stearns, 1953
- Erskine and Smith, 1993