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USE OF EROSIONAL FEATURES FOR TECTONIC RECONSTRUCTIONS AND INTERBASIN CORRELATION: AN EXAMPLE FROM THE RIO GRANDE RIFT, NORTHERN NEW MEXICO

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Abstract—Worldwide study of continental rifts has led to the development of several models depicting the evolution of rift basins. These models generally unravel rift-basin history by analyzing rift-margin structures and syntectonic sediment. This work builds on existing models by including spatial analysis of geologic features about intrarift accommodation zones to supplement documented patterns related to rift-margin structures, and by using geomorphic and erosional data to supplement syntectonic sedimentary data. These supplemental data provide process-based information on the tectonic evolution of rift basins, and a context for regional (interbasin) correlation of erosional features. Intrarift accommodation zones, such as the Embudo fault zone, are scissors faults that reverse the polarity of asymmetric half-graben basins. This causes depositional facies and erosional features to be mirrored, occurring on opposing sides of rifts in adjacent basins. The tectonic history, constructed from syntectonic sedimentation in hanging-wall depocenters, can be augmented by erosional history on rift-margin footwall blocks and along the hinge zone of half-grabens where basin-subsidence rates are lowest or uplift occurs. Delineation of spatial patterns of geomorphic and erosional surfaces and the processes of their genesis is valuable in rift settings where syntectonic sediments are buried, and rift-margin structures are inactive and/or poorly exposed.

INTRODUCTION

Intensive study of rift basins, including the Rio Grande rift in northern New Mexico, as well as rifts worldwide, has led to the development of several models depicting their tectonic evolution, basin architecture and syntectonic sedimentation patterns (cf. Bridge and Leeder, 1979; Gibbs, 1984; Bosworth, 1985; Rosendahl et al., 1986; Leeder and Gawthorpe, 1987; Alexander and Leeder, 1987; Frostick and Reid, 1989a; Keller and Cather, 1994; others). In this paper I review similar elements of rift models with two distinctions. First, I add spatial analysis of geologic features about intrarift accommodation zones to supplement documented patterns related to rift-margin structures. Second, I use geomorphic and erosional data to supplement syntectonic sedimentary data, the staple of most other studies. The value in these supplemental data is that they provide additional means of deciphering rift basin geometry and tectonic history in basins where rift-fill sediments are buried, or rift-margin structures are inactive and/or poorly exposed. Also, these geomorphic and erosional data provide a context for regional (interbasin) correlation of erosional features.

GEOLOGIC SETTING

The following abbreviated geologic setting is drawn, in part, from Baltz (1978), Baldridge et al. (1984) and Dungan et al. (1984). The Rio Grande rift extends for over 1000 km from Leadville, central Colorado, through New Mexico to Presidio, Texas and northern Chihuahua, Mexico. The rift is characterized by east-west crustal extension, Tertiary and Quaternary faulting, volcanism and high heat flow. Rift basins have several kilometers of structural relief and sedimentary fill in their depocenters.

In northern New Mexico the rift is defined by a series of right-stepping, *en echelon* morpho-tectonically distinct basins (Fig. 1). Each basin is generally 30–60 km wide and 40–80 km long. The polarity of rift-boundary faults alternates between down-to-the-east and down-to-the-west, resulting in strongly asymmetric basins. Several northeast-trending accommodation zones traverse the rift and offset each individual basin in a dextral direction. This paper focuses on the Española and San Luis Basins in northern New Mexico, particularly the areas near the Embudo fault zone, one of the intrarift accommodation zones.

The Española Basin is bounded to the east by the Sangre de Cristo Mountains and Picuris Range, to the west by the Nacimiento uplift and Jemez volcanic field, and to the north by the Embudo fault zone. The western side of the Española Basin is actively subsiding due to down-to-the-east movement on the Pajarito fault and down-to-the-south movement on the western end of the Embudo fault.

The San Luis Basin is bound to the east by the Taos Range of the Sangre de Cristo Mountains, to the west by the Tusas Mountains and the Colorado Plateau, and to the south by the Embudo fault. The Taos Plateau and Taos graben (Dungan et al., 1984) compose the actively subsiding southeastern part of basin. The area is subsiding by down-to-the-north faulting on the east end of the Embudo fault and down-to-the-west faulting on the Sangre de Cristo fault (cf. Machette and Personius, 1984), a major mountain-front, rift-margin fault system. The structurally complex Abiquiu embayment occupies the southwestern corner of the basin. The Abiquiu embayment is a relatively shallow portion of the San Luis Basin that contains east- and southeast-dipping strata offset along numerous normal faults. The Abiquiu embayment and Taos Plateau are separated by the N30° W-trending Precambrian bedrock high that runs from the Picuris Range to the Tusas Range through Cerro Azul (Fig. 1).

STRUCTURE OF RIFT BASINS

Extensional rift tectonics, dominated by simple shear, generally produce asymmetric half-graben basins, depicted schematically in Figs. 2A and 2C (e.g., Gibbs, 1984; Bosworth, 1985; Rosendahl et al., 1986). Important tectonic and geomorphic components of asymmetric half-graben basins include:

(1) The formation of a main, rift-margin fault (Fig. 2A and 2C). Other half-graben faults are defined as synthetic or antithetic relative to this main rift-margin fault.

(2) A rift-margin footwall and a rift-interior hanging wall. The rift-margin footwall is typically back-tilted away from the rift axis (Fig. 2C).

(3) Two primary tectonic slopes—a steep footwall scarp, and a gentler, hanging-wall dip slope (Fig. 2C; Leeder and Gawthorpe, 1987).

(4) And/or a rollover and/or hinge zones on the shallow end of the hanging-wall block (Fig. 2A and 2C; Frostick and Reid, 1987).

Tilting of the half-grabens generally causes the axis of maximum subsidence to be near the footwall scarp (Fig. 2C). Sedimentary and geomorphic responses to rift tectonism can be examined in the context of where sediments accumulate, erosion occurs, and/or depositional environments migrate to relative to the main rift-margin fault.

Intrarift accommodation zones (Frostick and Reid, 1989a; also known as transfer zones [cf. Gibbs, 1984], and transform faults [cf. Muehlberger, 1979, and Bally, 1985]) provide structural boundaries between rift basins. Scissors-like motion occurs on the accommodation zones (e.g., Kelley, 1978, 1979; Muehlberger, 1979), causing reverse sense of structural displacement and direction of stratal dip between adjacent basins. Accommodation zones generally trend at acute angles to the margins of the rift. The Embudo fault zone is a scissors fault and an excellent ex-

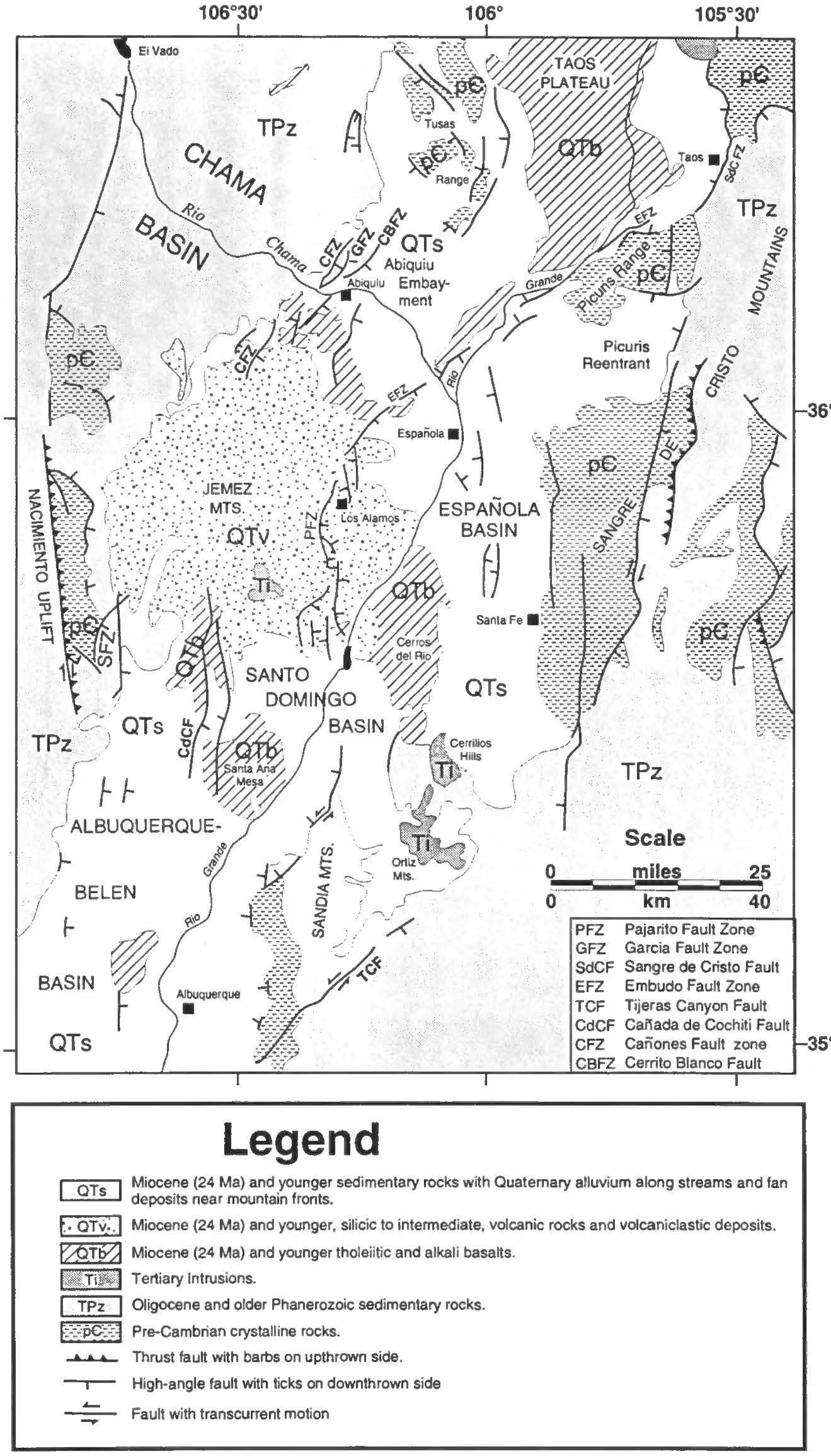


FIGURE 1. Generalized geologic map of the Rio Grande rift and its surroundings in northern New Mexico (modified from Baldridge et al., 1983).

ample of an intrarift accommodation zone. Its pivot point, where vertical displacement vectors reverse sense, is near the confluence of the Rio Chama and Rio Grande (Kelley, 1978).

SEDIMENTATION PATTERNS

Patterns of sedimentation in the Rio Grande rift of northern New Mexico are similar to patterns in other asymmetric rift basins. Tectonic movement on the main rift-margin fault tends to keep the axis of maximum basin subsidence, and correspondingly the zone of maximum sediment accumulation, close to the footwall scarp (Fig. 2C; Zones A in Figs. 3 and 4). High rates of tectonism produce rapid subsidence, causing axial through-flowing streams, lakes, playas and/or sabkhas to migrate toward the footwall scarp (e.g., Bridge and Leeder, 1979; Leeder and Gawthorpe, 1987; Alexander and Leeder, 1987). Following is a review of the regional distribution of four major rift-basin sedimentary facies: footwall-scarp (syntectonic) fan deposits; hanging-wall dip-slope fan deposits; lacustrine deposits; and axial-stream deposits.

Footwall-scarp (syntectonic) fan deposits

Fans constructed across the footwall scarp generally are fed by steep streams with limited catchments (Fig. 2B), which have relatively high stream power and transport capacity (e.g., Bull, 1979; Kelson, 1986;

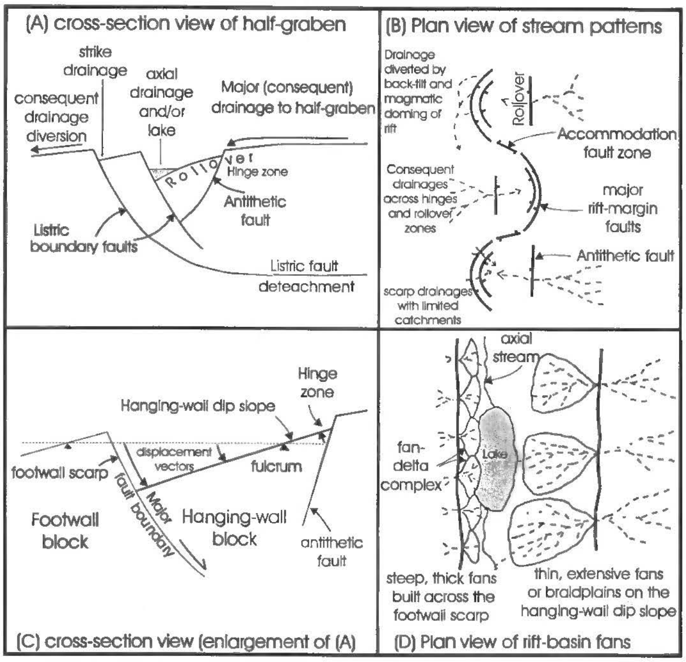


FIGURE 2. Schematic diagrams of half-graben basins in cross-section (A and C), and plan view (B, D), illustrating general structural features, geomorphic features and drainage patterns. (A) Modified from Gibbs (1984, fig. 3); (B) from Rosendahl et al. (1986, fig. 4) and Frostick and Reid (1987, fig. 3); (C) from Leeder and Gawthorpe (1987, fig.1); and (D) from Frostick and Reid (1987, fig. 2).

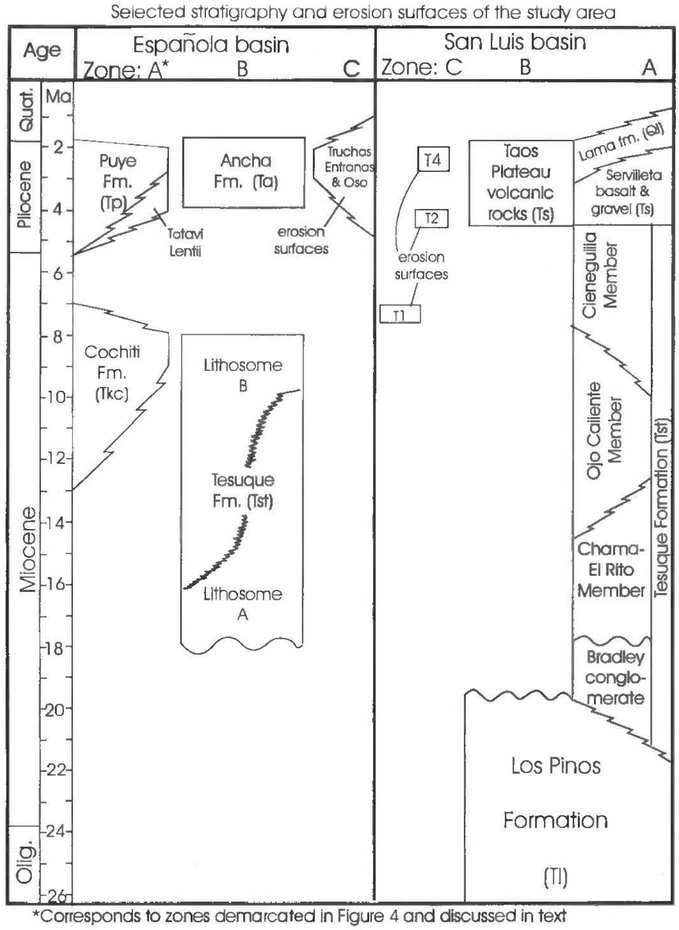


FIGURE 3. Simplified stratigraphic columns of the Española and San Luis Basins. Only those stratigraphic units discussed in the text are shown here (adopted from Galusha and Blick, 1971; Manley, 1976, 1979; Dungan et al., 1984; Gonzalez, 1993). Zones A, B and C coincide with areas indicated in Fig. 4.

Frostick and Reid, 1989a). Many investigators (e.g., Steel, 1976; Blair and Bilodeau, 1988; Leeder and Gawthorpe, 1987; Dunne and Hempton, 1984; Mack and Seager, 1990) attribute the accumulation of coarse fan sediment to the onset of tectonic activity with concomitant development of relief. However, progradation of coarse fan sediment is not necessarily diagnostic of increased tectonic uplift (cf. Blair and Bilodeau, 1988; Gordon and Heller, 1993). Nevertheless, where rates of tectonism remain high, syntectonic fans are morphologically distinct from climate-controlled fans. Syntectonic fans tend to have short, steep surfaces, and are vertically stacked next to the footwall block (Bull, 1979; Frostick and Reid 1989a).

Several characteristic syntectonic fan deposits occur in the rift basins of northern New Mexico, including the Cochiti and Puye Formations in the Española Basin, and several members of the Tesuque Formation in the southern San Luis Basin (Zones A in Fig. 3). Although the sediment of the Cochiti and Puye Formations is derived from huge volcanic piles, their location, age and geometry provide important evidence of syntectonic sedimentation patterns.

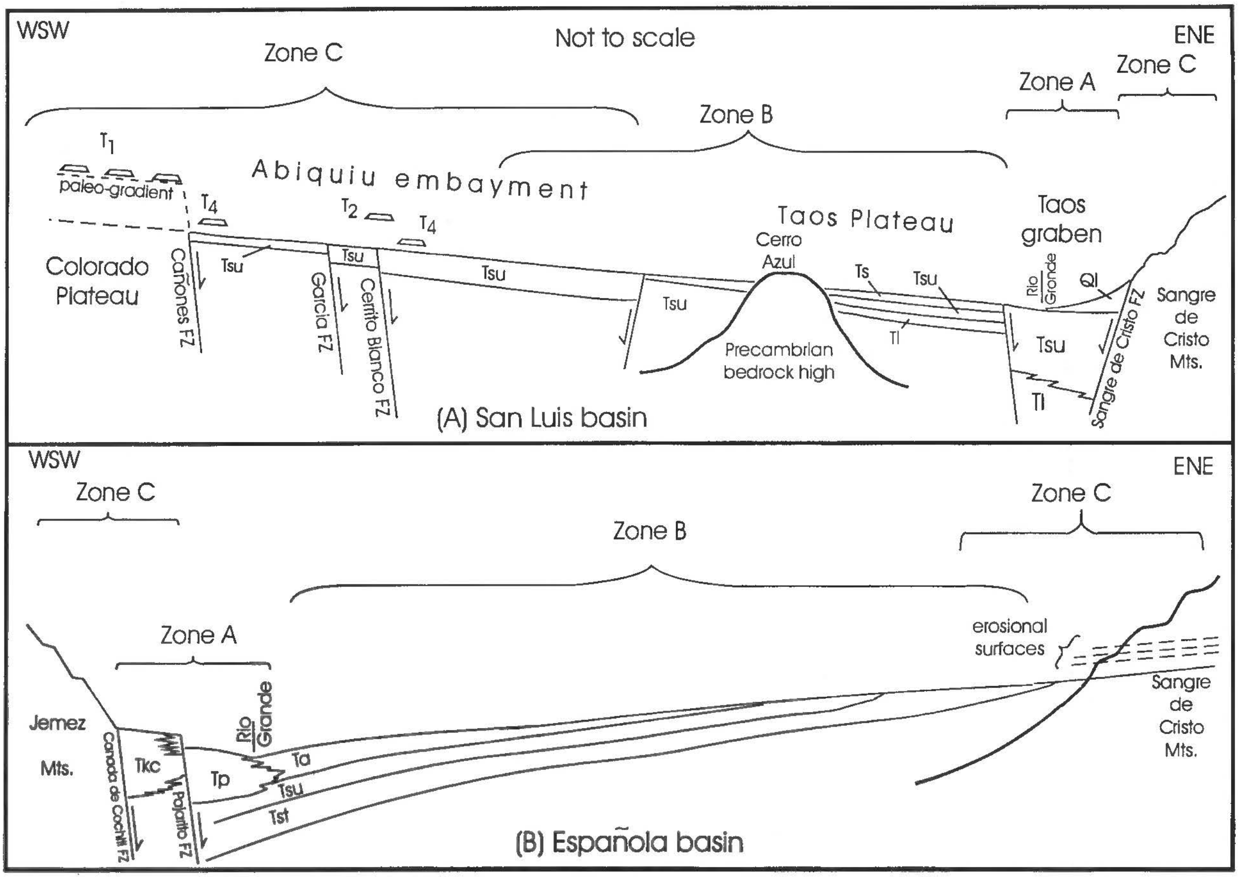
The Cochiti Formation is an immature, coarse-grained, basin-fill deposit, comprising lahar and other debris-flow deposits, fluvial deposits, minor cinder-cone debris, and numerous, thin beds of andesite, rhyolite and dacite flows and ash (Goff et al., 1990). Much of the detritus was derived from the 13-7 Ma Keres Group volcanic rocks (of Bailey et al., 1969) to the west. Drainage off the Keres outcrops transported volcaniclastic sediment toward the center of the ancestral Española Basin. The thickest portion of the Cochiti Formation was deposited between 9.5 and 8.7 Ma (Goff et al., 1990) immediately east of the Cañada de Cochiti fault zone, as concurrent down-to-the-east normal faulting created a depocenter for the volcaniclastic sediment (Smith et al., 1970; Gardner and Goff, 1984).

Another volcaniclastic fanglomerate is the Pliocene Puye Formation (Griggs, 1964), an immature, coarse-grained, basin-fill fanglomerate comprising debris flows, fluvial sediments and tephras (e.g., Bailey et al., 1969). The sediment was derived primarily from the dacitic Tschicoma Formation of the Polvadera Group. Sediment in the Puye Formation indicates that drainage systems originated on the Tschicoma domes and flowed east to southeast across a footwall scarp (the Pajarito fault zone) toward the ancestral Rio Grande.

The Puye Formation is best preserved to the east of the Pajarito fault between White Rock Canyon and the western end of the Embudo fault (Smith et al., 1970), although isolated thin and discontinuous patches of the Puye are found north of the Embudo fault in the southeastern part of the Abiquiu embayment. These thin scattered remnants may be channel-fill deposits of streams draining the northeastern flank of the Jemez Mountains (Gonzalez, 1993).

In the southern San Luis Basin, the Taos graben (cf. Lipman and Mehnert, 1979, fig. 2; Muehlberger and Muehlberger, 1982, p. 67) is a Neogene and Quaternary depocenter that has filled with up to 3 km of sediment shed from the Sangre de Cristo Mountains and Servilleta flood basalts (Figs. 3, 4). Some of the deep graben fill is believed to be the volcaniclastic Los Pinos Formation (Oligocene to early Miocene?; Lipman and Mehnert, 1979), derived from the San Juan Mountains to the west. The remainder is the Tesuque Formation of the Santa Fe Group (Muehlberger and Muehlberger, 1982; Dungan et al., 1984). Individual members of the Tesuque Formation, such as the Bradley conglomerate (informal name), and Chama-El Rito and Cieneguilla Members (Fig. 3), record episodic periods of tectonism on the Sangre de Cristo and Embudo fault zones with concomitant subsidence of the Taos graben, and uplift of the Picuris Range to the south and Sangre de Cristo Mountains to the east (Dungan et al., 1984).

Early to middle Quaternary fan sediment, derived from the Sangre de Cristo Mountains, progrades over the Tertiary fill of the Taos graben and interfingers with the Servilleta basalts (Figs. 3, 4). These fan deposits have been informally referred to as the upper Servilleta Formation (Lambert, 1966) or alternatively, the Lama formation (Pazzaglia, 1989, p. 21).



Note: Stratigraphic symbols explained in Figure 3, (Tsu is Santa Fe Group, undivided)

FIGURE 4. Simplified sketch showing the reversal in structural polarity and stratal tilt, and the spatial relations between erosion and deposition zones across the Embudo fault, an intrarift accommodation zone. The schematic cross-sections trend N60°E, parallel to the trace of the Embudo fault. Zone A demarcates basin depocenters, where rates of subsidence are greatest. Zone B is the hanging-wall dip-slope environment where relatively thin, planimetrically extensive fans, consequent stream deposits and/or braidplains form. Zone C demarcates the areas where uplift or minimum rates of subsidence occur. Geomorphic and erosional surfaces are common in this zone. Zone C is found on either the hinge side of half graben basins, or on the footwall block adjacent to the main rift-margin fault.

Hanging-wall dip-slope fan deposits

Two large, hanging-wall, dip-slope fans developed during the Neogene in the eastern Española Basin. The Miocene Tesuque Formation (specifically lithosome A of Cavazza, 1989) consists of sediment that was shed from the Sangre de Cristo Mountains, transported west by streams, and deposited in a broad braidplain that is finer and thicker toward the west.

During Plio-Pleistocene time a relatively thin (<30 m thick) but laterally extensive fan system developed along the eastern side of the Española Basin from west- to southwest-flowing streams. The fan sediment was derived from Precambrian crystalline and Paleozoic sedimentary rocks in the Sangre de Cristo Mountains, and is referred to as the Ancha Formation (Spiegel and Baldwin, 1963). At its distal limits the Ancha Formation is intercalated with the westward-transported sediment of the Puye Formation and the volcanic deposits of the Cerros del Rio volcanic field. Intercalated tephras from the Cerros del Rio volcanic field give a K-Ar date of 2.6 Ma (Dethier, in press), suggesting that the Ancha Formation is Plio-Pleistocene in age. Sediments composing the Tesuque braidplain (Cavazza, 1989) and the Ancha Formation show stratigraphic offlap, a result of sediment reworking and fan progradation westward across the Española Basin (Zone B in Figs. 3 and 4).

Lacustrine deposits

Although lacustrine and axial-stream fluvial deposits are not syntectonic sediments *sensu stricto*, their spatial distribution may be tectonically controlled by subsidence along rift-margin faults. In northern New Mexico lacustrine strata (the Culebra lake clay of Kelley, 1956) occur east of the Pajarito fault within the Puye Formation in the western and deepest part of the Española Basin. The only other local occurrence of lacustrine strata was reported by Winograd (1959), based on drill hole data and pumping tests. These lake sediments are within the Santa Fe Group in the Taos graben of the San Luis Basin.

Axial-stream deposits

Similar to lacustrine deposits, the position of axial, through-flowing stream deposits tends to be along the axis of maximum basin subsidence (e.g., Bridge and Leeder, 1979; Leeder and Gawthorpe, 1987; Alexander and Leeder, 1987; Kraus and Middleton, 1987; Blair and Bilodeau, 1988; Frostick and Reid, 1989a; Mack and Seager, 1990). The course of the modern Rio Grande illustrates strong basin asymmetry, flowing within the Taos graben in the eastern side of the San Luis Basin, then exploiting the Embudo fault zone to transfer to the western side of the Española Basin.

The Totavi Lentil (of Griggs, 1964) of the Puye Formation is an early Pliocene remnant of axial-stream deposits (Fig. 3). It, too, occurs east of the Pajarito fault zone in the deepest part of the Española Basin. The Totavi Lentil is approximately 2.6 to 5 Ma, based on paleontologic (Galusha and Blick, 1971; Tedford, 1981), paleomagnetic (MacFadden, 1977), and isotopic data (Baldridge et al., 1980; Dethier, in press).

Tectonism inhibits the formation of axial through-flowing streams in young rifts by creating structural and topographic highs, usually at accommodation zones (Frostick and Reid, 1989a). Only after sedimentation has filled basins sufficiently to allow washover from one rift lake (basin) to another can axial through-flowing streams evolve. Lithosome

B of the Tesuque Formation (Cavazza 1989) is an example of this axial stream washover from the eastern San Luis Basin toward the central and western Española Basin along an accommodation zone.

GEOMORPHIC FEATURES AND EROSIONAL SURFACES

Remnants of erosion surfaces, lava-filled paleovalleys and paleochannels, gravel-capped pediment surfaces and strath terraces are common in the rift basins of northern New Mexico. Neogene erosion surfaces are best preserved in two locations within the study area--the Abiquiu embayment (Gonzalez and Dethier, 1991; Gonzalez, 1993) and in the northeastern plateau area of the Española Basin, also referred to as the Picuris reentrant (e.g., Miller et al., 1963; Galusha and Blick, 1971; Manley, 1979; Kelley, 1979).

Lava flows and gravel deposits overlie several erosion surfaces in the Abiquiu embayment and adjacent margin of the Colorado Plateau. The gravel deposits have been identified as the *ca* 8 Ma unit T1, the *ca* 5 Ma unit T2, and the *ca* 3 Ma unit T4 (Gonzalez and Dethier, 1991; Gonzalez, 1993). These units rest on corresponding T1, T2 and T4 erosion surfaces.

The T1 erosion surface is found in three localities that straddle the Colorado Plateau/Abiquiu embayment margin: Cerro Pedernal, Mesa Escoba and Polvadera Mesa. The overlying Lobato basalt flows have a K-Ar date of 7.8 ± 0.7 Ma at Cerro Pedernal, a K-Ar date of 7.9 ± 0.5 Ma at Mesa Escoba, and a K-Ar date of 7.8 ± 0.5 Ma at Polvadera Mesa (Manley and Mehnert, 1981). The T1 erosion surfaces occur at different elevations or heights above the present channel of the Rio Chama, probably as a result of movement on the Cañones fault zone. The T1 erosion surface is approximately 1110 m above the present channel of the Rio Chama at Cerro Pedernal, 730 m at Mesa Escoba, and 540 m at Polvadera Mesa.

The T2 erosion surface occurs beneath piedmont gravel and a basalt flow on the southeast flank of Sierra Negra, a volcanic cone approximately 9 km NNE of Abiquiu. The basalt flow has a K-Ar date of 4.8 ± 0.1 Ma (Baldridge et al., 1980). At the terminus of the overlying flow, the T2 erosion surface is approximately 530 m above the present channel of the Rio Chama.

The T4 erosion surface is preserved beneath fluvial sediment and El Alto basalt flows at Cañones Mesa and Mesa de Abiquiu, and beneath fluvial sediment and Servilleta basalt at Black Mesa. These basalts have K-Ar dates of 2.8 ± 0.7 and 2.8 ± 0.5 Ma at Cañones Mesa (Manley and Mehnert, 1981), 3.2 ± 0.1 Ma at Mesa de Abiquiu (Baldridge et al., 1980), and 2.78 ± 0.44 Ma at Black Mesa (Manley, 1976). The T4 erosion surfaces occur at approximately 380 m above axial-stream grade at Cañones Mesa, 220 m above grade at Mesa de Abiquiu, and 270 m above grade at Black Mesa. The differences in height above grade have been attributed to late Neogene movement on intervening faults (Gonzalez and Dethier, 1991; Gonzalez, 1993), although this is disputed by Baldridge et al. (1994).

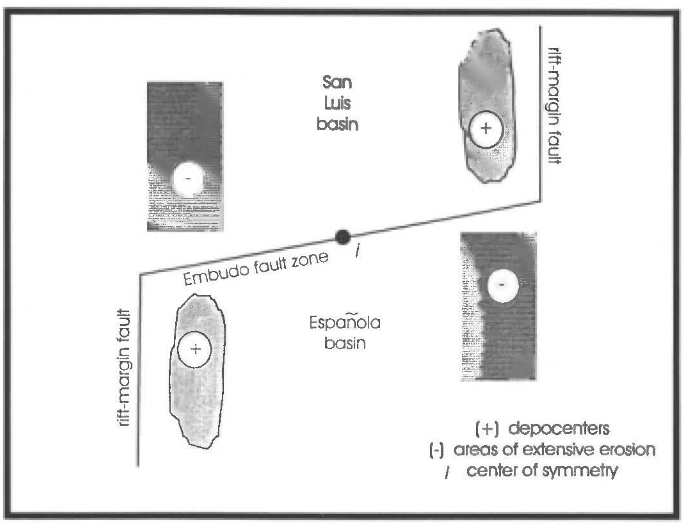


FIGURE 5. Simplified sketch showing the spatial geometry between half-graben depocenters (+) and hinges (-), where erosion surfaces are prevalent. The pivot point of an accommodation zone acts as a center of symmetry (i).

Manley (1976, 1979) studied erosional surfaces in the piedmont area of the Picuris reentrant, identifying several post-Tesuque surfaces, named the Oso, Entrañas and Truchas surfaces (oldest to youngest). The surfaces cut older valley fill and are covered by piedmont sediment, primarily sandy pebble gravel. Manley (1979) projected longitudinal profiles of these surfaces northwestward toward Black and Velarde Mesas and correlated them with stream-gravel deposits beneath these basalt-capped mesas that mark former base levels of the Rio Grande. This relation would make these surfaces time-correlative with the T4 erosion surface. Isotopic dates of basalt flows and tephra that bury the gravel deposits constrain the date of gravel deposition on the erosional surfaces at between 5 and 1.4 Ma (Manley, 1976, 1979). Stratigraphic evidence can further constrain the date between 3 and 2 Ma (Manley, 1979).

Mid- to late Quaternary erosional landforms are prevalent and inset along axial-river valleys. They formed during large-scale, rift-wide fluvial incision. Simultaneous base-level fall throughout the entire rift suggests individual basin tectonics is not the cause of these inset landforms; consequently a discussion of them is omitted here.

DISCUSSION

Spatial relations across the Embudo fault zone

Many rift investigators have noted that scissors-like motion on intrarift accommodation zones alternates structural polarity, stratal dip and the position of depocenters between adjacent half-graben basins (Figs. 4A, 4B, 5). These relations are observed across the Embudo fault zone in the Rio Grande rift of northern New Mexico, with depocenters in the eastern part of the San Luis Basin and the western part of the Española Basin. This work extends these spatial patterns to include erosion surfaces, which also occur on alternate sides of rift basins across accommodation zones. Erosion surfaces are more prevalent on the tectonically passive (or hinge) side of half-graben basins, such as the Abiquiu embayment and eastern side of the Española Basin (Figs. 4 and 5). Consequently, interbasin correlation of depocenter strata and erosional surfaces may be possible by projecting these rift features through the accommodation zone's pivot point, which serves as a center of symmetry (to borrow crystallographic parlance) to relate similar features in adjacent basins (Fig. 5).

Genesis and interbasin correlation of erosion surfaces

Tectonically controlled erosion and geomorphic surfaces are more prevalent near the tectonically more passive margin (i.e., the hinge and/or rollover zone), and on the uplifted shoulder of footwall blocks. These areas have either the lowest rates of subsidence in a rift basin or are being uplifted. Erosion occurs when the stream power, transport capacity, and gradient of hanging-wall dip-slope or footwall-scarp drainages increase because of subsidence along the footwall scarp (Bull, 1979; Kelson, 1986; Frostick and Reid, 1989a). Increased relief and steeper gradients promote stream incision and development of erosion surfaces along the half-graben hinge or uplifted shoulder of footwall blocks (Zone C, Figs. 4A, B).

Thin, planimetrically extensive fans or braidplains form in the area between basin depocenter and hinge (Figs. 2D; Zone B in Figs. 4A, B). Throughout Zone B, sediment is subject to reworking, transport down dip slope and redeposition, creating stratigraphic offlap as the fans prograde toward the basin depocenter (cf. Cavazza, 1989).

Recognition of the genesis and spatial distribution of erosion surfaces is important in reconstructing basin geometry and tectonic history, and in attempting interbasin correlation of rift features. For example, many investigators have noted prominent, widespread erosional surfaces in the rift of northern New Mexico (see Kelley, 1979 for a summary of early work); however, there has been no model to relate these surfaces to one another. Since Bryan (1938, p. 215) proposed the name "Ortiz surface" for a broad gravel-capped surface in the southeastern part of the Española embayment, numerous investigators have tried to correlate this to other broad surfaces throughout the rift with little regard to their age, genesis or position within a basin relative to rift structures. The conceptual model proposed here provides an additional means of correlating tectonically controlled erosion surfaces between adjacent basins, and relating them with time-equivalent syntectonic sediment elsewhere.

Use of geomorphic surfaces in tectonic reconstruction of rift basins

Tectonic reconstructions have been accomplished primarily through analyses of syntectonic sediment. Where the geometry, extent and age of rift-fill strata have been determined, they provide a reference to determine the amount of rift-related diastrophism that has followed or accompanied sedimentation. Structural deformation of progressively younger strata is evidence of progressively more recent tectonism.

However, in those parts of rift basin that have undergone uplift, or minimum subsidence, sediment has not accumulated and consequently cannot provide constraints on the timing, magnitude and rate of tectonism. These areas generally have experienced base-level fall, stream incision, and large-scale excavation of basin fill. In these environments, erosion surfaces, indicators of former base-level, provide supplementary information on the tectonic history and geometry of basins. If the erosion surface is offset by faulting, then one can measure the vertical separation across the fault. (Note that offset of a near horizontal plane, the erosion surface, provides vertical control on faulting, but not any measure of horizontal or oblique separation.) Offset of progressively younger erosion surfaces is evidence of correspondingly younger tectonism.

For example, the rift-fill strata exposed in the Abiquiu embayment are relatively old (Oligocene to early Miocene Abiquiu Formation). These strata provide poor age constraints on subsequent neotectonic activity. However, the T1, T2 and T4 erosion surfaces provide evidence of progressive base-level fall during the late Miocene and Pliocene. In addition, where remnants of the same erosion surface straddle a fault, vertical separation between them provides information on the magnitude and rate of offset on the fault. For example, the vertical separation between remnants of the T1 erosion surface at Cerro Pedernal and Polvadera Mesa is approximately 570 m. This value indicates the amount of faulting across the Cañones fault zone since 8 Ma. Similarly, the vertical separation of the T4 erosion surface is 170 m between Cañones Mesa and Mesa de Abiquiu, suggesting that the intervening Garcia and Cerrito Blanco faults have been active since 3 Ma (Gonzalez, 1993, p. 51-66).

Finally, it is possible to correlate erosion surfaces with time-equivalent rift-fill strata to augment basin analyses. For example, while the Pliocene T2 and T4 erosion surfaces record progressive base-level fall north of the Embudo fault zone in the Abiquiu embayment, the Pliocene Totavi Lentil and Puye Formation record concomitant aggradation of 210 m in the western Española Basin depocenter. The abrupt change between areas recording base-level fall and rise coincides with the western end of the Embudo fault zone (cf. Gonzalez, 1993, p. 157-160). Not only do the spatial patterns of aggradation and incision delineate the boundaries of structural domains, but the ages of the strata and erosion surfaces indicate times when these rift structures were active.

CONCLUSION

Examples drawn from the rift of northern New Mexico illustrate how basin analyses are augmented by: (1) relating erosional features in the relatively passive tectonic margin (hinge) with rift-fill strata in the depocenter along the tectonically active margin of half grabens; and (2) providing a spatial framework for interbasin (regional) correlation of tectonically controlled erosion surfaces. In addition, reconstruction of basin geometry and tectonic history is augmented by erosion surfaces where rift structures long have been inactive or syntectonic sediments are deeply buried. In the absence of surface exposures, subsurface exploratory data, such as seismic and borehole data, must be used to reconstruct basin history. Subsurface techniques are expensive and provide data at discrete points or lines, which is less desirable than near continuous surface exposures. Syntectonic sediment has been the staple of tectonic reconstructions. Erosional surfaces, largely ignored in such studies, provide supplemental data and augment basin analyses.

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REFERENCES

Alexander, J. and Leeder, M. R., 1987, Active control on alluvial architecture; *in* Ethridge, F. G., Flores, R. M. and Harvey, M. A., eds., Recent developments in fluvial sedimentology: Society of Economic Paleontologists and Mineralogists, Special Publication 39, p. 243-252.

Bailey, R. A., Smith, R. L. and Ross, C. S., 1969, Stratigraphic nomenclature of volcanic rocks in the Jemez Mountains, New Mexico: U.S. Geological Survey, Bulletin 1274-P, 19 p.

Baldridge, W. S., Damon, P. E., Shafiqullah, M. and Bridwell, R.J., 1980. Evolution of the central Rio Grande rift, New Mexico: new potassium–argon ages: Earth and Planetary Science Letters, v. 51, p. 309–321.

Baldridge, W. S., Bartov, Y. and Kron, A., 1983, Geologic map of the Rio Grande rift and southeastern Colorado Plateau, New Mexico and Arizona, scale 1:500,000; *supplement to* Riecker, R. E., ed., Rio Grande rift: tectonics and magmatism: American Geophysical Union, Washington, D.C.

Baldridge, W. S., Olsen, K. H. and Callender, J. F., 1984, Rio Grande rift: problems and perspectives: New Mexico Geological Society, Guidebook 35, p. 1–12.

Baldridge, W. S., Ferguson, J. F., Braile, L. W., Wang, B., Eckhardt, K., Evans, D., Schultz, C., Gilpin B., Jiracek, G. R. and Biehler, S., 1994, The western margin of the Rio Grande Rift in northern New Mexico: an aborted boundary?: Geological Society of America Bulletin, v. 105, p. 1538–1551.

Baltz, E. H., 1978, Resume of Rio Grande depression in north-central New Mexico: New Mexico Bureau of Mines and Mineral Resources, Circular 163, p. 210–228.

Bally, A. W., 1985, Structural styles and the evolution of sedimentary basins: American Association of Petroleum Geologists Education Short Course.

Blair, T. C. and Bilodeau, 1988, Development of tectonic cyclothems in rift, pull-apart, and foreland basins: sedimentary response to episodic tectonism: Geology, v. 16, p. 517–520.

Bosworth, W., 1985, Geometry of propagating continental rifts: Nature, v. 316, p. 625–627.

Bridge, J. S. and Leeder, M. R., 1979, A simulation model of alluvial stratigraphy: Sedimentology, v. 26, p. 617–644.

Bryan, K., 1938, Geology and ground-water conditions of the Rio Grande depression in Colorado and New Mexico; *in* U.S. National Resources Planning Board, the Rio Grande Joint Investigation in the upper Rio Grande basin: Washington, D.C., U.S. Government Printing Office, v. 1, part 2, p. 197–225.

Bull, W. B., 1979, Thresholds of critical power in streams: Geological Society of America Bulletin, v. 90, p. 453–464.

Cavazza, W., 1989, Sedimentation pattern of a rift-filling unit, Tesuque Formation (Miocene), Española Basin, Rio Grande rift, New Mexico: Journal of Sedimentary Petrology, v. 59, p. 287–296.

Chapin, C. E., 1971, The Rio Grande rift; part 1, modifications and additions: New Mexico Geological Society, Guidebook 22, p. 191–201.

Dethier, D.P. (in press), Geologic map of the White Rock Quadrangle: New Mexico Bureau of Mines and Mineral Resources, scale 1:24,000.

Dungan, M. A., Muehlberger, W. R., Leininger, L., Peterson, C., McMillan, N. J., Gunn, G., Lindstrom, M. and Haskin, L., 1984, Volcanic and sedimentary stratigraphy of the Rio Grande gorge and the late Cenozoic geologic evolution of the southern San Luis Valley: New Mexico Geological Society, Guidebook 35, p. 157–170.

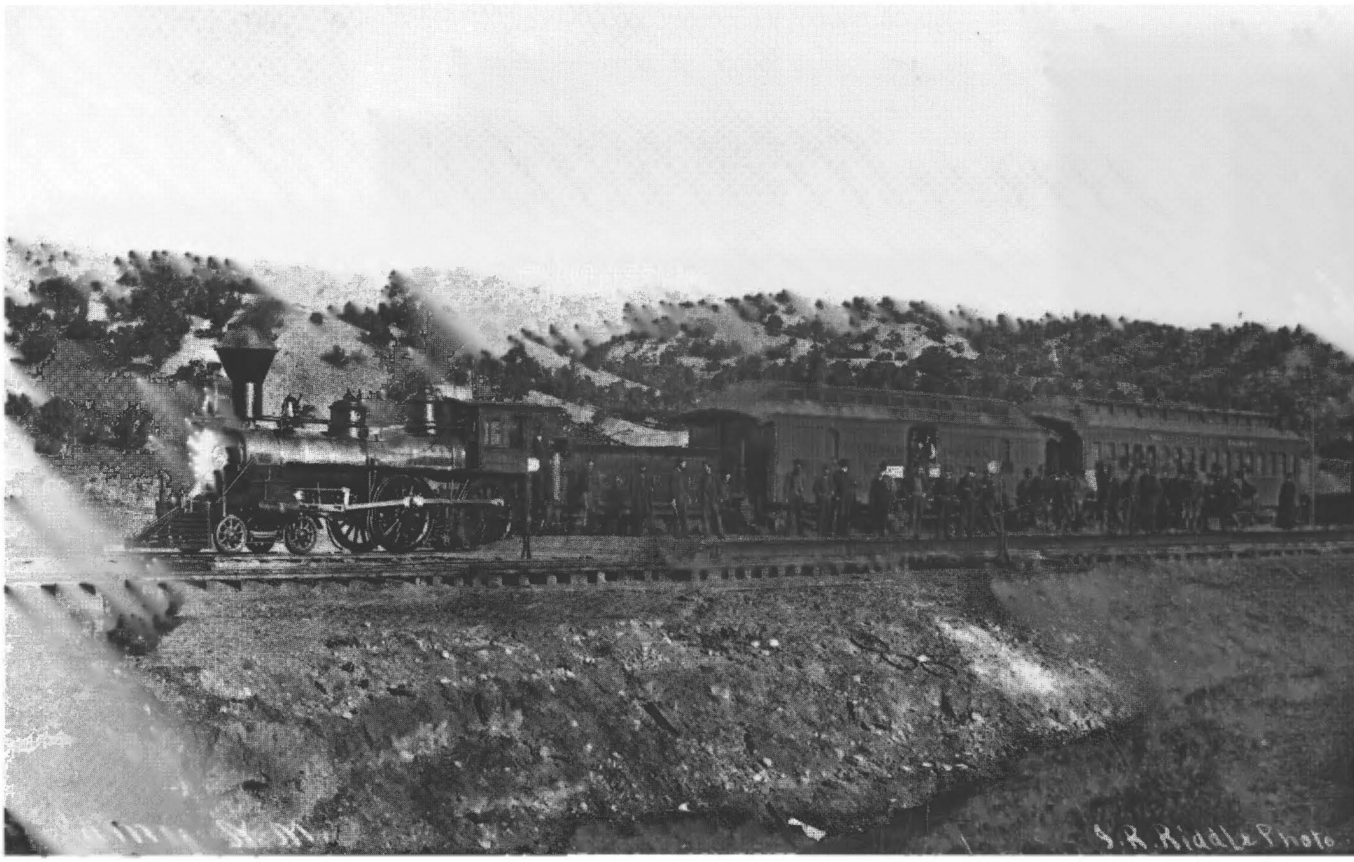
Dunne, L. A. and Hempton, M. A., 1984, Deltaic sedimentation in the Lake Hazar pull-apart basin, south-eastern Turkey: Sedimentology, v. 31, p. 401–412.

Frostick, L. E. and Reid, I., 1987, Tectonic control of desert sediment in rift basins ancient and modern; *in* Frostick, L. E. and Reid, I., eds., Desert Sediments: ancient and modern: Geological Society (of London), Special Publication 35, p. 53–68.

Frostick, L. E. and Reid, I., 1989a, Is structure the main control of river drainage and sedimentation in rifts?: Journal of African Earth Sciences, v. 11, p. 165–182.

Frostick, L. E. and Reid, I., 1989b, Climatic versus tectonic controls of fan se-

quences: lessons from the Dead Sea, Israel: Journal of the Geological Society of London, v. 146, p. 527–538.
Galusha, T. and Blick, J. C., 1971, Stratigraphy of the Santa Fe Group, New Mexico: American Museum of Natural History, Bulletin 144, 127 p.
Gardner, J. N. and Goff, F., 1984, Potassium-argon dates from the Jemez volcanic field: implications for tectonic activity in the north-central Rio Grande rift: New Mexico Geological Society, Guidebook 35, p. 75–82.
Gibbs, A. D., 1984, Structural evolution of extensional basin margins: Journal of the Geological Society of London, v. 141, p. 609–620.
Goff, F., Gardner, J. N. and Valentine, G., 1990, Geology of St. Peter's Dome area, Jemez Mountains, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Geologic Map 69, scale 1:24,000.
Gonzalez, M. A., 1993, Geomorphic and neotectonic analysis along a margin of the Colorado Plateau and Rio Grande rift in northern New Mexico [Ph.D. thesis]: Albuquerque, University of New Mexico, 302 p.
Gonzalez, M. A. and Dethier, D. P., 1991, Geomorphic and neotectonic evolution along the margin of the Colorado Plateau and Rio Grande rift, northern New Mexico: New Mexico Bureau of Mines and Mineral Resources, Bulletin 137, p. 29–45.
Gordon, I. and Heller, P.L., 1993, Evaluating major controls on basinal stratigraphy, Pine Valley, Nevada: implications for syntectonic deposition: Geological Society of America Bulletin, v. 105, p. 45–55.
Griggs, R. L., 1964, Geology and ground-water resources of the Los Alamos area, New Mexico: U.S. Geological Survey, Water-Supply Paper 1753, 107 p.
Keller, G. R. and Cather, S. M., eds., 1994, Basins of the Rio Grande rift: structure, stratigraphy and tectonic setting: Geological Society of America, Special Paper 291.
Kelley, V. C., 1956, The Rio Grande depression from Taos to Santa Fe: New Mexico Geological Society, Guidebook 7, p. 109–114.
Kelley, V. C., 1978, Geology of Española Basin, New Mexico: New Mexico Bureau of Mines and Mineral Resources Geologic Map 48, scale 1:125,000.
Kelley, V. C., 1979, Geomorphology of Española Basin: New Mexico Geological Society, Guidebook 30, p. 281–288.
Kelson, K. I., 1986. Long-term tributary adjustments to base-level lowering northern Rio Grande rift, New Mexico [M.S. thesis]: Albuquerque, University of New Mexico, 210 p.
Kraus, M. J. and Middleton, L. T., 1987, Contrasting architecture of two alluvial suites in different structural settings; *in* Ethridge, F. G., Flores, R. M. and Harvey, M. A., eds., Recent developments in fluvial sedimentology: Society of Economic Paleontologists and Mineralogists, Special Publication 39, p. 253–262.
Lambert, W., 1966, Notes on the late Cenozoic geology of the Taos-Questa area, New Mexico: New Mexico Geological Society, Guidebook 17, p. 43–50.
Leeder, M. R. and Gawthorpe, R. L., 1987, Sedimentary models for extensional tilt-block/half-graben basins; *in* Coward, M. P., Dewey, J.F. and Hancock, P. L. eds., Continental extensional tectonics: Geological Society (of London), Special Publication 28, p. 139–152.
Lipman, P. W. and Mehnert, H. H., 1979, The Taos Plateau volcanic field, northern Rio Grande rift, New Mexico; *in* Riecker, R. E. ed., Rio Grande rift: tectonics and magmatism: American Geophysical Union, Washington, D.C., p. 289–312.
MacFadden, B. J., 1977. Magnetic polarity stratigraphy of the Chamita Formation stratotype (Mio-Pliocene) of north-central New Mexico: American Journal of Science, v. 277, p. 769–800.
Machette, M. N. and Personius, S. F., 1984, Map of the Quaternary and Pliocene faults in the eastern part of the Aztec 1° X 2° quadrangle and the western part of the Raton 1° X 2° quadrangle, northern New Mexico: U.S. Geological Survey, Miscellaneous Field Studies Map MF-1465-B, scale 1:250,000 with supplemental pamphlet, 20 p.
Mack, G.H. and Seager, W. R., 1990, Tectonic control on facies distribution of the Camp Rice and Palomas Formations (Pliocene-Pleistocene) in the southern Rio Grande rift: Geological Society of America Bulletin, v. 102, p. 45–53.
Manley, K., 1976, The late Cenozoic history of the Española Basin, New Mexico [Ph.D. dissertation]: Boulder, University of Colorado, 171 p.
Manley, K., 1979, Tertiary and Quaternary stratigraphy of the northeast plateau, Española Basin, New Mexico: New Mexico Geological Society, Guidebook 30, p. 231–236.
Manley, K. and Mehnert, H. H., 1981. New K-Ar ages for Miocene and Pliocene volcanic rocks in the northwestern Española Basin and their relationships to the history of the Rio Grande rift: Isochron\West, v. 30, p. 5–8.
Miller, J. P., Montgomery, A. and Sutherland, P. K., 1963, Geology of part of the southern Sangre de Cristo Mountains, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Memoir 11, 106 p.
Muehlberger, W. R., 1979, The Embudo fault between Pilar and Arroyo Hondo, New Mexico: an active intracontinental transform fault: New Mexico Geological Society, Guidebook 30, p. 77–82.
Muehlberger, W. R. and Muehlberger, S., 1982, Española-Chama-Taos: a climb through time: Scenic Trips to the Geologic Past, No. 13, New Mexico Bureau of Mines and Mineral Resources, Socorro, New Mexico, 99 p.
Pazzaglia, F. J., 1989, Tectonic and climatic influences on the evolution of Quaternary depositional landforms along a segmented range-front fault, Sangre de Cristo Mountains, north-central New Mexico [M.S. thesis]: Albuquerque, University of New Mexico, 236 p.
Rosendahl, B. R., Reynolds, D. J., Lorber, P. M., Burgess, C. F., McGill, J., Scott, D., Lambiase, J. J. and Derksen, S. J., 1986, Structural expression of rifting: lessons from Lake Tanganyika, Africa; *in* Frostick, L. E., Renaut, R. W., Reid, I. and Tiercelin, J. J., eds., Sedimentation in the African rifts: Geological Society (of London), Special Publication 25, p. 29–43.
Smith, R. L., Bailey, R. A. and Ross, C. S., 1970, Geologic map of the Jemez Mountains, New Mexico: U.S. Geological Survey, Miscellaneous Geological Investigations Map I-571, scale 1:25,000.
Spiegel, Z. and Baldwin, B., 1963, Geology and water resources of the Santa Fe area, New Mexico: U.S. Geological Survey, Water-Supply Paper 1525, 158 p.
Steel, R. J., 1976, Devonian basins of western Norway—sedimentary response to tectonism and to varying tectonic context: Tectonophysics, v. 36, p. 207–224.
Tedford, R. H., 1981, Mammalian biochronology of the late Cenozoic basins of New Mexico: Geological Society of America Bulletin, Part I, v. 92, p. 1008–1022.
Winograd, I. J., 1959, Ground-water conditions and geology of Sunshine valley, western Taos County, New Mexico: New Mexico State Engineer Office Technical Report 12, 70 p.



Life in the fast lane—1880's style. A respectable train load of passengers and dignitaries, recently detrained from the main line *Atlantic Express* at Lamy, New Mexico, pose before boarding the north-bound 10:30 am local for the 18-mile, 1 hr, journey to Santa Fe in this ca. 1882 view. Locomotive #195, on the business end of the run, has a full head of steam and the head-end crew is anxious to depart. One of twelve such built for the AT&SF RR by the Hinckley Locomotive Works, the polished 4-4-0 "American Type" was the pride of the railroad for many years and the standard design for both passenger and freight service. NMBM&MR photo collection, #1320; photo by J. R. Riddle, courtesy Kansas State Historical Society, Topeka.