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BRIEF SUMMARY OF THE TERTIARY GEOLOGIC HISTORY OF THE RIO GRANDE RIFT IN NORTHERN NEW MEXICO

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INTRODUCTION

The Rio Grande rift of Colorado and New Mexico can be divided into a series of structural basins. Within one of these basins (the San Luis Valley), from the New Mexico state line southward, is the Taos Plateau (Fig. 1). Adjoining it to the south is the Espanola Basin. The boundary between is marked by the termination of the Taos Plateau volcanic field and the presence of a basement ridge (Cordell, 1978) crossing the rift from the Picuris Range, through the Precambrian exposure at Cerro Azul, to the La Madera area (Fig. 2).

This paper presents an overview of the Tertiary history of the Taos Plateau, the northernmost Espanola Basin, and parts of the adjacent Tusas Mountains to the west and the Sangre de Cristo Mountains to the east.

STRUCTURE

In the Taos Plateau, the Rio Grande rift is roughly 35 km wide. The east side of the rift is defined by a major fault system along the base of the Sangre de Cristo Mountains that has a possible offset of 7-8 km (Lipman and Mehnert, 1979). The western boundary is formed by a series of north—northwest-trending normal faults that bound eastward-dipping blocks along the eastern half of the Tusas Mountains (Manley, 1982a, b). Trending parallel to the Sangre de Cristo Mountains, in the eastern half of the rift valley, are an intra-rift horst and graben. The horst brings Tertiary volcanic rocks to the surface. Pliocene basalts show about 500 m of post-emplacement offset along the eastern boundary (Lipman and Mehnert, 1979), and overlying Quaternary deposits have been faulted 17 m in the late Pleistocene (Personius and Machette, this guidebook).

The Espanola Basin is about 65 km wide. On the northeast it apparently lacks a prominent border fault; the westward-tilted blocks on the east side are bounded by down-to-the-east normal faults (Manley, 1979). The northwestern margin is formed by a major northeast-trending fault zone that places Tertiary rocks against rocks of Permian age. Displacement of 8 m.y. old basalts is about 670 m (Manley and Mehnert, 1981).

Located within the north-central Espanola Basin is a roughly 8 km wide graben, the Velarde graben. The western side of the graben is formed by a fault passing southwestward from beneath Black Mesa, across the Rio Grande, and along the southeast side of Arroyo de la Presa into the area of Clara Peak where it turns southward to become the Pajarito fault. The eastern fault border is visible from Velarde southwestward until it is lost beneath the Rio Grande floodplain northeast of Espanola (Manley, 1977). Gravity data indicate a total thickness of 2.5-5 km of basin fill in this area (Cordell, 1979). Pliocene subsidence of the graben is at least 300 m (Manley, 1979).

The boundary between the San Luis Valley and the Espanola Basin is complex. The two basins have opposite directions of major tilting, and the intra-rift grabens of each are approximately 19 km apart in a right-lateral direction. The basement ridge has a possible 3 km of structural relief (Cordell, 1979). The Velarde graben pinches out northeastward; its bounding faults appear to join each other and merge into the Embudo fault along the northwest flanks of the Picuris Range where reverse and left strike—slip motion are reported (Muehlberger, 1979). Kelley (1978) indicates a reversal of throw on the western side of the graben near the confluence of the Chama and Ojo Caliente Rivers; however, my detailed mapping in the area contradicts this interpretation. I concur with Muehlberger (1979) that the point of reversal probably overlies the basement ridge which separates the basins. Other relationships in this boundary area require further investigation.

STRATIGRAPHY

Tertiary sedimentary and volcanic rocks overlie Precambrian metamorphic rocks to the east and west of the Taos Plateau. The Precambrian rocks, uplifted in the Laramide orogeny, were almost totally stripped of later deposits. Paleozoic and Mesozoic rocks are present only locally

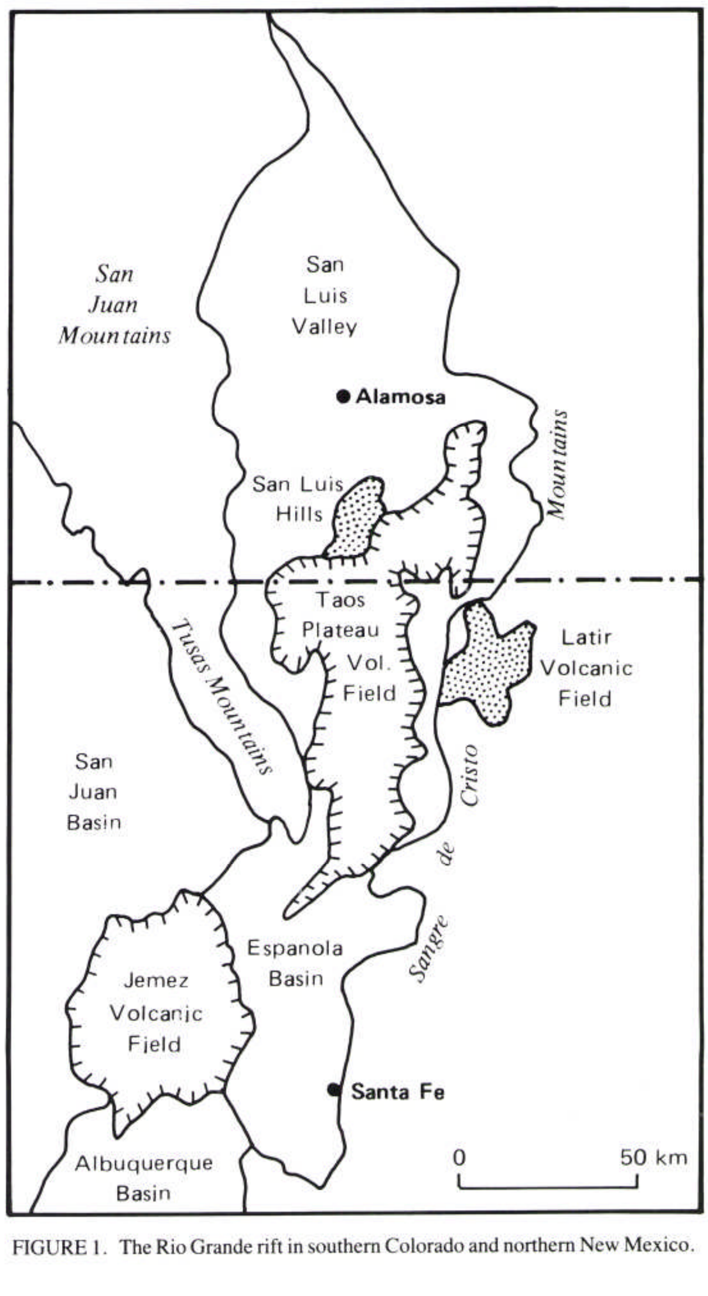

FIGURE 1. The Rio Grande rift in southern Colorado and northern New Mexico.

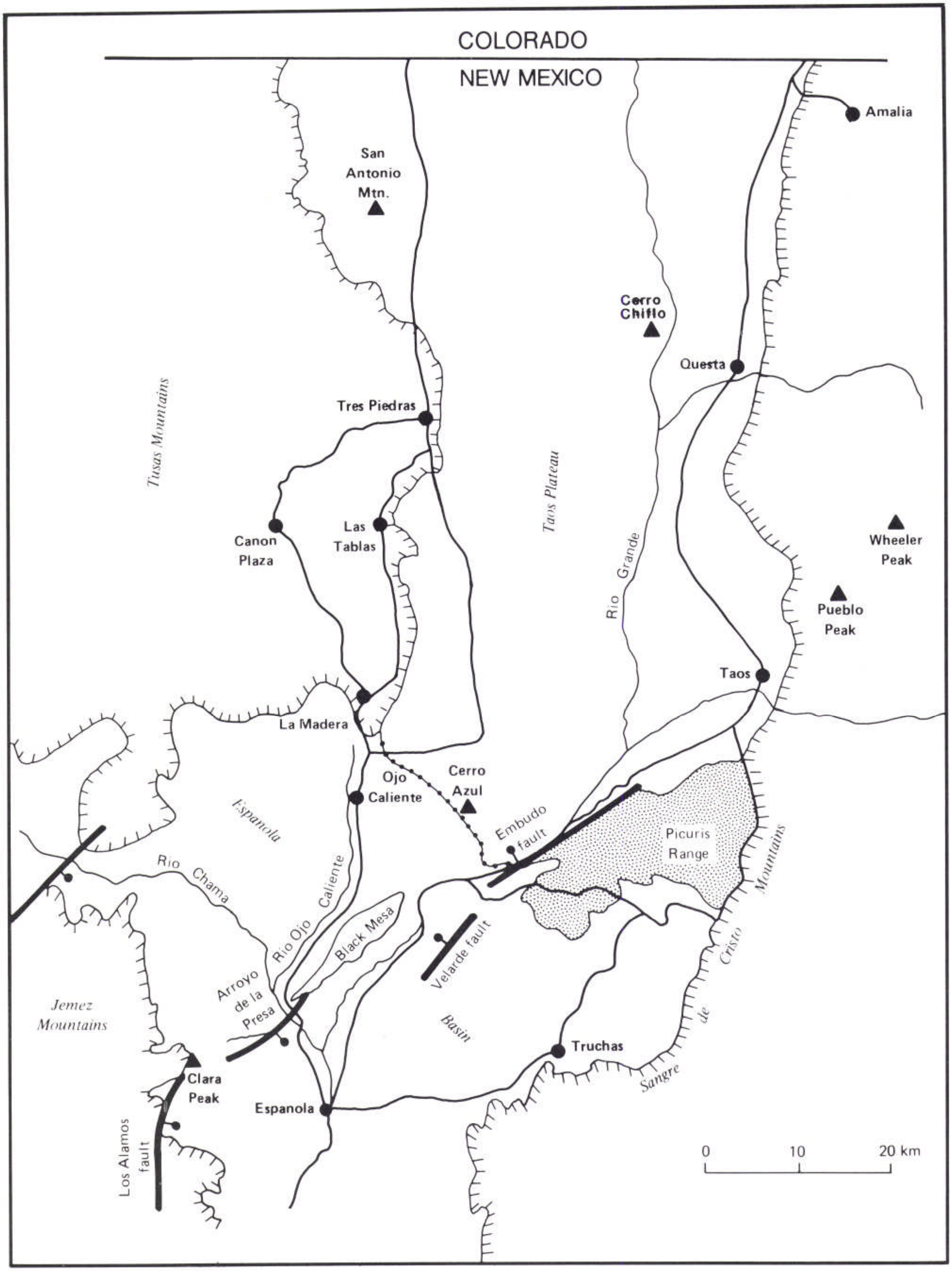

FIGURE 2. Index map for the Rio Grande rift area in north-central New Mexico.

in the westernmost and southernmost Tusas Mountains (Muehlberger, 1968; Bingler, 1968). Mississippian and Pennsylvanian rocks are exposed in the Sangre de Cristo Mountains south of Pueblo Peak (Armstrong, 1967; Dane and Bachman, 1965); structural relationships are imperfectly known.

The earliest Tertiary deposits on both sides of the rift are discontinuous coarse conglomerates and sandstones that were derived from the underlying Precambrian rocks. These units, the Eocene El Rito Formation of the Tusas Mountains and the Eocene(?) Vallejo Formation of the Sangre de Cristo Mountains, filled valleys in areas of high relief and may be locally as much as 250 m thick (Manley and Wobus, 1982a). Both are assigned an Eocene age, although no paleontological evidence is available.

Pre-rift volcanism began in the Tusas Mountains with intermediate-composition breccias that erupted from numerous local centers. In the north these are equivalent to the Conejos Formation, whose age is 35 to 31 m.y. (Lipman and others, 1970). South, near La Madera, these breccias may be as young as 26 m.y. (Manley and Wobus, I982a). Local eruptions of intermediate-composition lavas and breccias also occurred in the Sangre de Cristo Mountains between 35 and 26 m.y. (Lipman, 1983) and in the San Luis Hills (northern Taos Plateau in southern Colorado) prior to 28 m.y. (Burroughs, 1971).

The volcanic breccias, Eocene sedimentary rocks, and remnant Precambrian hills were soon buried by an influx of volcaniclastic detritus. In the Taos Plateau and the Latir Mountains, the detritus was locally derived. Elsewhere only a small percentage of the material had local sources: (1) the northern and central Tusas Mountains were covered by alluvial fans building southward from the volcanic highlands of the San Juan Mountains; (2) in the central and southern Tusas Mountains and northern Espanola Basin, detritus was primarily derived from the early Taos Plateau and Latir volcanic fields. The alluvial material comprises the Los Pinos and Abiquiu Formations (Butler, 1971), and a part (Manley, unpubl. data 1980) of the Picuris Formation of Miller (Miller and others, 1963). The Picuris Range did not create a barrier to northern-derived volcaniclastic sediment in the northeastern Espanola Basin because it was not in existence at this time.

The age of volcaniclastic deposition is fairly well known. The Los Pinos overlies the 30 m.y. old Treasure Mountain Tuff and the 28 m.y. old Masonic Park Tuff (Lipman, 1975) in the northern and central Tusas Mountains (Manley, 1982a, b; Wobus and Manley, 1982b). The termination time for Los Pinos deposition is highly variable: in the northern Tusas Mountains it continued to 5 m.y. (Lipman and Mehnert, 1975); in the northwestern Espanola Basin the transitional and interfingering upper contact appears to be 16-18 m.y. (Baldridge and others, 1980; May, 1979 and this guidebook; Manley, 1981; Manley and Mehnert, 1981; Ekas and others, this guidebook). On the east of the Taos Plateau, comparable volcaniclastic deposition ranged from about 22 to 16 m.y. (Lipman, 1983).

Sedimentary rocks of the Santa Fe Group were deposited throughout most of the area. Early deposits are volcaniclastic (Galusha and Blick, 1971; May, this guidebook) but in the northern Espanola Basin are distinguishable from the underlying Los Pinos and Abiquiu Formations because of the difference in composition and/or percentage of volcanic clasts. In the northwestern Espanola Basin, the oldest Santa Fe Group deposit, the Chama-El Rito Member of the Tesuque Formation, contains locally derived volcanic clasts (May, 1979; Manley, unpubl. data 1980). As vertical tectonism increased in the rift, Precambrian source areas became re-exposed and their contribution to the Santa Fe Group sediments came to dominate the fans spreading westward from the southern and central Sangre de Cristo Mountains. In much of the Espanola Basin these fans built across the rift, but they barely extended into the Tusas Mountains due to the eastward-tilting of the Taos Plateau. A similar shift from dominantly volcaniclastic deposition to Precambrian clasts is shown in the sedimentary rocks of the northeast edge of the rift (Lipman, 1983).

Volcanism continued in the area contemporaneously with the deposition of the Los Pinos and Abiquiu Formations. Within the Taos Plateau, rhyolite (including ash-flow tuff), rhyodacites, and andesites were erupted roughly 26 to 22 m.y. ago (Lipman and Mehnert, 1979). Similar eruptions occurred in the Sangre de Cristo Mountains and include the Tuff of Tetilla Peak (27 m.y.) and the Amalia Tuff (26 m.y.) (Lipman, 1983). Rhyolite domes (Manley and Wobus, 1982) and the Amalia Tuff, with a K-Ar age of 26 m.y. (Bingler, 1968), are interlayered with the Los Pinos Formation in the southern Tusas Mountains (Butler, 1971: Manley, 1981).

Volumentrically minor basalt flows are interlayered with the Los Pinos Formation; many now cap eastward-tilted mesas in the northern Tusas Mountains. These flows have been variously named in the past (Butler, 1971), but all of them are currently considered to belong to the Hinsdale Formation (Lipman and Mehnert, 1975). In southern Colorado these flows have a K-Ar age range of 26 to 5 m.y. (Lipman and Mehnert, 1975). In the northern Espanola Basin, K-Ar ages on three flows are 19, 21, and 22 m.y. (Baldridge and others, 1980; Manley and Mehnert, 1981).

The older Santa Fe Group deposits in the northwestern Espanola Basin contain a few scattered volcanic flows, dikes, and tuff rings. These were sporadically erupted from 13 to 7 m.y. ago (Baldridge and others, 1980; Manley and Mehnert, 1981). Included in this group is a quartz-latitic dome, Cerro Chiflo, in the central part of the Taos Plateau, which has an age of 10 m.y. (Lipman and Mehnert, 1979).

Major basaltic volcanism, beginning within the Rio Grande rift about 5 m.y. ago, constitutes the Taos Plateau volcanic field. This field covers more than 1,500 km' and consists of numerous volcanoes, flows, and several domes of compositions ranging from basalt to silicic rhyolite (Lipman and Mehnert, 1979; Dungan and others, this guidebook). The most voluminous rocks are the olivine tholeiites of the Servilleta Basalt whose K-Ar ages range from 4.5 to 2.8 m.y. (Ozima and others, 1967; Manley, 1978), but may prove to contain several older flows (Baldridge and others, 1980). The Servilleta Basalt flows were erupted from shield volcanoes to the west-central part of the Plateau.

TECTONISM

In early Miocene the Rio Grande rift was no more than a slight depression. Sedimentary rocks were deposited, from highlands to the north and northeast, across the locations of the current boundary faults. The northwest boundary faults in the Espanola Basin were active between 7 and 3 m.y. (Manley and Mehnert, 1981). In the Tusas Mountains the age control is poor; only Miocene Los Pinos and Hinsdale Formations are commonly present. Along the Rio Tusas valley Santa Fe Group deposits arc faulted, but their age is unknown (Manley and Wobus, I 982b).

In both the Espanola Basin and Taos Plateau the central grabens may have formed slightly later than the outer fault boundaries. Creation of the Velarde graben is constrained between 5 and 3 m.y. (Manley, 1979); the Taos graben is less than 4.5 m.y. (Dungan and others, this guidebook). During this interval of tectonism the Picuris Range and the Sangre de Cristo Mountains were uplifted rapidly.

Tectonism and volcanism appear to have slowed by the end of the Pliocene. The stage was set for the Quaternary erosional cycles that led to the creation of the present landscape.

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