A shifting rift—Geophysical insights into the evolution of Rio Grande rift margins and the Embudo transfer zone near Taos, New Mexico

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ABSTRACT

We present a detailed example of how a subbasin develops adjacent to a transfer zone in the Rio Grande rift. The Embudo transfer zone in the Rio Grande rift is considered one of the classic examples and has been used as the inspiration for several theoretical models. Despite this attention, the history of its development into a major rift structure is poorly known along its northern extent near Taos, New Mexico. Geologic evidence for all but its young rift history is concealed under Quaternary cover. We focus on understanding the pre-Quaternary evolution of the subbasin by integrating diverse pieces of geologic and geophysical information. As a result, we present a substantively new understanding of the tectonic configuration and evolution of the northern extent of the Embudo fault and its adjacent subbasin.

We integrate geophysical, borehole, and geologic information to interpret the subsurface configuration of the rift margins formed by the Embudo and Sangre de Cristo faults and the geometry of the subbasin within the Taos emplacement. Key features interpreted include (1) an imperfect D-shaped subbasin that slopes to the east and southeast, with the deepest point ~2 km below the valley floor located northwest of Taos at ~36° 26′ N latitude and 105° 37′ W longitude; (2) a concealed Embudo fault system that extends as much as 7 km wider than is mapped at the surface, wherein fault strands disrupt or truncate flows of Pliocene Servilleta Basalt and step down into the subbasin with a minimum of 1.8 km of vertical displacement; and (3) a similar, wider than is mapped at the surface, wherein fault strands disrupt or truncate flows of Pliocene Servilleta Basalt and step down into the subbasin with a minimum of 1.8 km of vertical displacement.

From the geophysical interpretations and subsurface models, we infer relations between faulting and flows of Pliocene Servilleta Basalt and older, buried basaltic rocks that, combined with geologic mapping, suggest a revised rift history involving shifts in the locus of fault activity as the Taos subbasin developed. We speculate that faults related to north-striking grabens at the end of Laramide time formed the first west-down master faults. The Embudo fault may have initiated in early Miocene southwest of the Taos region. Normal-oblique slip on these early fault strands likely transitioned in space and time to dominantly left-lateral slip as the Embudo fault propagated to the northeast. During and shortly after eruption of Servilleta Basalt, proto-Embudo fault strands were active along and parallel to the modern, NE-aligned Rio Pueblo de Taos, ~4–7 km basinward of the modern, mapped Embudo fault zone. Faults along the northeastern subbasin margin had northwest strikes for most of the period of subbasin formation and were located ~5–7 km basinward of the modern Sangre de Cristo fault. The locus of fault activity shifted to more northerly striking faults within 2 km of the modern range front sometime after Servilleta volcanism had ceased. The northerly faults may have linked with the northeasterly proto-Embudo faults at this time, concurrent with the development of N-striking Los Cordovas normal faults within the interior of the subbasin. By middle Pleistocene(?), time, the Los Cordovas faults had become inactive, and the linked Embudo-Sangre de Cristo fault system migrated to the south, to the modern range front.

INTRODUCTION

The Neogene Rio Grande rift forms a series of north-south elongated structural basins that extend from Mexico to northern Colorado (inset, Fig. 1). The basins are characterized as one or more asymmetric half-grabens that generally tilt toward northerly striking master faults, with the polarities of the tilts varying between basins of the rift (Chapin and Cather, 1994). In northern New Mexico, the rift basins make significant right steps. In addition, tilt directions and associated master faults of the half-grabens vary from north to south, from east-tilted in the Albuquerque Basin, to west- to northwest-tilted in the Española Basin, and again east-tilted in the San Luis Basin.

The northeast-striking, left-oblique Embudo and related faults have long been recognized as the mechanism for accommodating transfer of strain between the oppositely tilted half-grabens of the Española and San Luis Basins...
Evidence for early Miocene initiation of the Embudo transfer zone comes mainly from the southern segment of the zone in the northern Española Basin, where Miocene basin-fill units (Santa Fe Group) are well exposed (Manley, 1978; Muehlberger, 1979; Dungan et al., 1984; Aldrich, 1986; Ingersoll et al., 1990; Aby and Koning, 2004; Bauer and Kelson, 2004a; Koning et al., 2004; Koning et al., 2013). Sedimentological evidence for the late Miocene uplift of the flanking Picuris Mountains has been used to signal the onset of activity on the northern Embudo fault (Manley, 1978; Muehlberger, 1979; Dungan et al., 1984; Ingersoll et al., 1990; Aby and Koning, 2004; Bauer and Kelson, 2004a; Koning et al., 2004; Koning et al., 2013). In any case, a common assumption is that the northern Embudo fault has been active since early Miocene, based on inferences that it had already formed a linked system between the Sangre de Cristo fault on the north or the Pajarito fault on the south (Bauer and Kelson, 2004a; Koning et al., 2004).
is that the northern Embudo fault has been active near its present location throughout its history.

Direct evidence of the pre-Quaternary history of the northern Embudo fault is buried beneath extensive alluvial cover. Subsurface data are required to investigate the geometry and structure of the subbasin formed by the linked Embudo–Sangre de Cristo fault system. Fortunately, a series of hydrologic and hydrogeologic studies conducted and ongoing in the Taos area over the past three decades have resulted in the collection of an unprecedented amount of deep well information and geophysical data that can be used for indirect investigation of subsurface structure (Reynolds, 1986, 1992; Bauer et al., 1999; Benson, 2004; Drakos et al., 2004; Grauch et al., 2004; Bauer et al., 2014; Bauer et al., 2016; Johnson et al., 2016). Additional geophysical studies have focused on understanding the three-dimensional framework of the Rio Grande rift (Bauer et al., 2004; Grauch and Keller, 2004; Bankey et al., 2007). The Taos area is also remarkable as the setting for geological and geophysical training of astronauts (Dickerson, 2004; Muehlberger, 2004), whose data also contributed to this study.

Despite the numerous subsurface investigations, several geophysical observations have remained puzzling. First, gravity data suggest that the basin margin associated with the Embudo fault zone has a more gradual slope and greater structural relief than expected for common models of antithetic or convergent transfer zones (Grauch and Keller, 2004). Second, both correlation of basaltic rocks in wells and analysis of aeromagnetic data suggest that a >200 m deep structural graben exists west of the Sangre de Cristo mountain front (Drakos et al., 2004; Grauch et al., 2004); yet analysis of the broad features of gravity and aeromagnetic data suggest that no such graben exists at the basin floor (Grauch and Keller, 2004; Grauch et al., 2004). Finally, several aeromagnetically inferred structural domains of differing orientations, which converge north of the Town of Taos, are difficult to understand in light of the locations of the present-day rift margins (Grauch et al., 2004).

Thus, we were motivated to integrate all the subsurface information with detailed geologic mapping to resolve the puzzling observations and develop a clearer understanding of the subsurface geology and Cenozoic evolution of the northern Embudo and Sangre de Cristo faults. From this synthesis, we propose several revisions to the fault history and development of the associated rift basin. Importantly, we suggest that fault activity has shifted to different locations throughout rift development. We speculate that the shifts have only partially followed a systematic migration away or toward the basin depocenter.

To develop the arguments for this and other hypotheses, we first present background information for all the evidence and data that were considered during the synthesis. The information comes from geologic mapping, and gravity, aeromagnetic, borehole, and physical-property data. After describing the geophysical methods we used, we present key geophysical interpretations of the subsurface geology and the reasoning behind the conclusions. We discuss age implications for the geophysical interpretations and then, finally, propose a revised history of the structural evolution of this corner of the rift, illustrated by a set of generalized paleogeographic maps.

### REGIONAL SETTING

The study area is located in the southeastern corner of the San Luis Basin, surrounding the Town of Taos, New Mexico (Fig. 1). The San Luis Basin is one of a series of north-south–elongated structural basins that compose the Cenozoic Rio Grande rift, which extends from Mexico to Colorado (Fig. 1 inset). As the basins developed in response to rifting of the crust, they filled mostly with clastic sediments and some lava flows. The poorly to moderately consolidated sediments, known as the Santa Fe Group, accumulated in the San Luis Basin mainly during the period 26–2 million yr B.P. (Ma) (Ingersoll et al., 1990; Brister and Gries, 1994; Bauer and Kelson, 2004a; Smith, 2004).

Proterozoic rocks, which form the basement of the San Luis Basin and much of the surrounding mountains, are composed of a diversity of rock types that are a product of continental accretion, plutonism, and regional metamorphism (Karlstrom et al., 2004). Both the Picuris-Pecos fault system and the Jemez lineament (Fig. 1) are considered to be major crustal boundaries that first developed in Proterozoic time, remaining episodically active ever since (Alrich, 1988; Bauer and Raiser, 1995). During the late Paleozoic, highlands located in the Tusas and Sangre de Cristo Mountains (Fig. 1; Kluth, 1986) shed sediments into intervening basins that were occasionally covered by shallow seas (Baltz and Myers, 1999). During the Late Cretaceous to Eocene Laramide orogeny, the Tusas and Sangre de Cristo Mountains were elevated again, shedding sediments into local basins east of the San Juan and Tusas Mountains (Manley, 1981; Brister and Gries, 1994).

From late Eocene to late Oligocene time, voluminous lava flows and ash-flows erupted primarily from the San Juan volcanic field (Fig. 1), blanketing most of the surrounding area (Southern Rocky Mountain volcanic field of Lipman, 2007). The related Latir volcanic field, located just north of the study area (Fig. 1), was active toward the end of this period with the eruption of the 25.4 Ma Amalia Tuff and accompanying collapse of the Questa caldera (Lipman et al., 1986; Zimmerer and McIntosh, 2012). Oligocene volcanic rocks are exposed in the central Taos Plateau (Thompson et al., 1986), indicating that such rocks may be widespread beneath the Taos Plateau volcanic field.

The transition from widespread volcanism to rift sedimentation is commonly considered to be ca. 26–25 Ma, marked by eruption of the Amalia Tuff and the onset of regional basaltic volcanism (Lipman and Meinert, 1975; Thompson et al., 1991). In reality, both tectonic styles overlap in age by several million years (Smith, 2004; Zimmerer and McIntosh, 2012), and some workers have suggested that basin subsidence did not initiate until 21 Ma (Ingersoll et al., 1990). The pre-rift stage of volcanism and plutonism persisted as late as ca. 23–22 Ma (Zimmerer and McIntosh, 2012), while erosion of volcanic highlands was widespread (Ingersoll et al., 1990; Smith, 2004).

Basin subsidence and accumulation of Santa Fe Group sediments in the San Luis Basin was at its peak during middle Miocene time (Bauer and Kelson, 2004a; Smith, 2004). Basaltic lavas erupted locally near the Colorado–New Mexico border during 15–11 Ma (Miggins et al., 2002; Thompson et al., 2004).
Geologic Units

Detailed geologic mapping recently compiled and updated by P.W. Bauer and K.J. Kelson from five 75 min quadrangles (Bauer et al., 1997; Kelson et al., 1998a; Bauer et al., 2000; Bauer and Kelson, 2001; Kelson and Bauer, 2003) plus regional mapping from a sixth quadrangle in the northwest corner of the study area (R.A. Thompson, U.S. Geological Survey (USGS), 2004, written commun.) are generalized for the study area in Figure 3. A map representing a subset of the updated, detailed geologic mapping, focused on the Embudo fault zone, is presented in Bauer et al. (2016). The generalized units are described below, noting only aspects that are important to the Cenozoic geologic history and geophysical interpretation.

Pre-Rift Rocks

Proterozoic and Paleozoic rocks are exposed in the Sangre de Cristo and Picuris Mountains (Figs. 2 and 3). Proterozoic rocks (Xu) include a diversity of rock types. Paleoproterozoic granite and granitic gneiss are exposed in the eastern Picuris Mountains and in the northeastern study area (Lipman and Reed, 1988; Bauer et al., 1997). Paleoproterozoic quartzite and schist of the Hondo Group are exposed in the Picuris Mountains, where they are folded and overlie the mixed felsic and mafic metavolcanic-metasedimentary sequence of the Paleoproterozoic Vadito Group (Bauer, 1993).

Paleozoic rocks (Pzu) exposed east of the Town of Taos represent strata that were deposited in a trough (Taos trough), which may be as much as 1800 m thick in the Sangre de Cristo Mountains (Baltz and Myers, 1999). The strata are mostly fine-grained siliciclastic rocks with minor limestone (Miller et al., 1963). The Paleozoic rocks are fault-bounded on the south and north, where they terminate against Proterozoic rocks (Fig. 3). How far they extend beneath the rift basin west of the Sangre de Cristo fault is uncertain. Baltz and Myers (1999) argued that they extend west of the Rio Grande and underlie the southern Taos Plateau. Woodward et al. (1999) argued that the Taos trough was bounded to the west by an active Picuris-Pecos fault during the Pennsylvanian, thus limiting the western extent of deposition.

Pre-Rift and Early Rift Deposits

Although early rift volcanic rocks of the Latir volcanic field (Fig. 1) are not exposed in the study area, they may exist on the basin floor. If so, their compositions may be similar to the intermediate-composition lavas and Amalia Tuff that are exposed in two small hills in the central Taos Plateau (Thompson et al., 1986).

The Picuris Tuff of Cabot (1938) is composed of distal volcanioclastie deposits and proximal sediments (Rehder, 1986; Aby et al., 2004; McDonald and Nielsen, 2004); thus, Picuris Formation more appropriately reflects the varied
Figure 2. Tectonic setting of the study area showing mapped faults compiled from geologic 7.5-min quadrangle maps (Bauer et al., 1997; Kelson et al., 1998a; Bauer et al., 2000; Bauer and Kelson, 2001; Kelson and Bauer, 2003). The Sangre de Cristo fault is divided into the Cañon and Hondo sections (Kelson et al., 1998b, 1998c). The short-dashed red line marks the limits of the Town of Taos. Locations of Figures 8 and 13 are shown by the solid red outlines.
Figure 3. Geology of the study area, showing profile locations and selected wells. The geologic map is simplified from recent compilation and revision of previous 1:24,000-scale geologic mapping by Bauer and Kelson (see text). Label "Pmag site" along the Rio Grande is where Brown et al. (1993) analyzed paleomagnetic measurements for a vertical section of Servilleta Basalt. Limit of shallow Servilleta Basalt [dark-gray, long-dashed line] is interpreted from aeromagnetic data.

EXPLANATION

Surficial Deposits
- Qa — Quaternary alluvium, stream terrace, colluvium, landslide, and eolian deposits

Synrift Deposits
- QTI — Upper part of Santa Fe Group (Lama formation), Deposits interbedded with and overlying Servilleta Basalt
- Tbs — Pliocene basalt, primarily Servilleta Basalt
- Tv — Pliocene intermediate to felsic rocks of Taos Plateau volcanic field
- Tsf — Lower part of Santa Fe Group (Miocene), including Tesuque and Chamita Formations

Pre-rift to Early Rift Deposits
- Tp — Tertiary Picuris Formation

Pre-rift Rocks
- Pzu — Paleozoic sedimentary rocks, dominantly clastic
- Xu — Proterozoic rocks, undivided, including granitic rocks, felsic and mafic metamorphic rocks, and quartzite

Structure
- Mapped fault, dashed where approximately located, dotted where concealed. Bar on down-thrown side of normal fault; arrows show strike-slip motion. Upthrown (U) and downthrown (D) sides of high-angle fault; arrows show strike-slip motion.
lithologies and is used here. The deposits range in age from greater than 34.5 Ma to less than 18.6 Ma (Aby et al., 2004), thus preserving an important record of the erosion of highlands during the transition from Oligocene volcanism to early rift basin subsidence.

Aby et al. (2004) divided the Picuris Formation into three informal members. The lower member (>34.5–28.3 Ma) is composed of fine-grained material with interbedded coarse conglomerate and is 300–400 m thick just southwest of the study area. Conglomerates contain Proterozoic, Paleozoic, and volcanic clasts. The volcanic clasts must be older than the Latir volcanic field; so they likely were transported southward from the San Juan volcanic field or derived from a local source. The middle tuffaceous sandstone member (28.3 to <23 Ma) records the influx of debris from the Latir volcanic field to the north, with progressively greater input from basement highlands through time. It is <125 m thick in the study area (Aby et al., 2004). The upper member (<23 Ma to <18.6 Ma) is composed of volcaniclastic pebble-conglomerate to mudstone and is gradational into the overlying Santa Fe Group units. Thickness is ~200–250 m in the study area. Paleoflow measurements and clast compositions indicate that the main source was the Latir volcanic field to the north, with minor contributions from Proterozoic basement and San Juan volcanic field lavas (Rehder, 1986; Aby et al., 2004).

**Synrift Deposits**

The Miocene section of the Santa Fe Group (Tsf on Fig. 3) is poorly exposed in the study area but has been well characterized to the southwest in the Española Basin (Fig. 1). From studies of exposures in the northern Picuris Mountains and examination of cuttings from wells, workers generally accept that correlative formations of Miocene age occur in the subsurface of the study area, including the Tesuque and overlying Chamita Formations (Bauer and Kelson, 2004a; Drakos et al., 2004; Bauer et al., 2016). The Tesuque and Chamita Formations are both composed of siliciclastic sediment, primarily sand (Koning et al., 2013). Only the lower part of the Tesuque Formation (Chama–El Rito Member) is exposed in the study area in the northern Picuris Mountains. A distinctive eolian sand unit (Ojo Caliente Sandstone Member), which overlies the Chama–El Rito Member, is exposed in the southwestern-most study area and is recognized as a marker unit in well logs (Drakos et al., 2004). By recognizing this marker unit, a minimum of 840 m of Miocene Santa Fe Group sediments are documented within the basin in the RP3200 well near the center of the study area (Fig. 3) (Glorieta Geoscience, Inc., 2007).

Limited exposures of intermediate-composition volcanic rocks (Tv) of the Taos Plateau volcanic field are present on the west side of the study area (Fig. 3). Servilleta Basalt (Tbs) is exposed more extensively (Fig. 3) and is commonly encountered in water wells drilld throughout the northwestern part of the study area (Benson, 2004; Drakos et al., 2004; Bauer et al., 2016). Servilleta Basalt refers to multiple flows of olivine tholeiitic basalt with a distinctive diktytaxitic texture, ranging in age from ca. 5.5 to 3 Ma in New Mexico (Lipman and Mehnert, 1979; Appelt, 1998; Read et al., 2004; Cosca et al., 2014; Kelson et al., 2015). Where they are exposed in the Rio Grande gorge, basalts form a 150-m-thick section composed of individual flows that are discontinuous, interfered, and variable in thickness (Dungan et al., 1984). Intervening sedimentary intervals generally are 0.35–4.5 m thick (Leininger, 1982) but have thicknesses as much as 50 m in water wells near Taos (Glorieta Geoscience, Inc., 2007) and 70 m in river gorges northwest of the study area (Dungan et al., 1984; Read et al., 2004).

Previous workers have relied on the division of the Servilleta Basalt into Upper, Middle, and Lower units developed by Dungan et al. (1984) in the Rio Grande gorge to correlate basalts across distances and between wells (e.g., Drakos et al., 2004). However, paleomagnetic studies of Servilleta Basalt at several sites within the gorge demonstrated that the magnetostatigraphy of flow packages do not correlate with these divisions (Brown et al., 1993). Moreover, recent age dating and chemical studies of flows in the Rio Grande gorge reveal much greater discontinuity of flows and complexities due to paleo-topography than previously thought (M.A. Cosca and R.A. Thompson, USGS, 2015, oral commun.). These results warn against adhering to a strict basalt stratigraphy when correlating basaltic intervals between wells.

Nomenclature for rift sediments that are contemporaneous with eruption of Servilleta Basalt is not consistent in the literature. In this study, we follow the criteria established by Kelson et al. (2015), who informally define the Lama formation (QTI) as Santa Fe Group sediments that are interbedded with flows of Servilleta Basalt and were deposited after the first Servilleta Basalt flow (ca. 5.5 Ma) and before incision of the Rio Grande (early[?] to middle[?] Pleistocene). The Lama formation consists of poorly sorted sand, pebbles, and cobbles, commonly in a fine-grained matrix (Bauer and Kelson, 2004a; Kelson et al., 2015). The sediments are highly oxidized and weathered, suggesting they were saturated by a high water table before the water table lowered in association with regional stream incision (Bauer and Kelson, 2004a). Clast compositions and paleoflow indicators suggest sources came from Proterozoic, Paleozoic, older volcanic, and Servilleta Basalt terranes from a variety of directions (Kelson et al., 2015). Bordering the northeastern part of the study area, large thicknesses (>25 m) of Lama formation record the advancement of massive alluvial fans from the mountain front to the east (Kelson and Bauer, 2006).

**Surficial Deposits**

Surficial deposits (Qa) cover most of the study area (Fig. 3) and generally lie unconformably on the Lama formation and equivalent units. This unconformity marks the beginning of regional stream incision (Kelson et al., 2015). Although the deposits are not differentiated on Figure 3, detailed mapping of the surficial deposits has provided key information for understanding the timing and nature of exposed fault scarps (e.g., Kelson et al., 2004a, 2004b) and helps constrain the age of past activity of faults that are interpreted mainly from geophysical evidence.
The mountain fronts are dominated by coalescing alluvial fans, ranging in age from early (?) Pleistocene to Holocene (Bauer and Kelson, 2004a; Bauer et al., 2016). They interdigitate with stream terrace deposits that flank the major streams (Kelson and Wells, 1989). Landslide deposits cover the deep gorges in the Rio Grande and Rio Pueblo de Taos, and colluvium is prevalent on steep slopes at the base of the Sangre de Cristo Mountains. Eolian deposits form a thin blanket over much of the southern Taos Plateau.

Structure

**Picuris-Pecos Fault System**

The north-south Picuris-Pecos fault system is a major crustal structure that has been episodically active since Proterozoic time (Montgomery, 1953; Miller et al., 1963; Bauer and Raiser, 1995). It consists of an 8-km-wide zone of north-striking faults that can be traced for ~84 km south of the study area (Fig. 1). The major strands juxtapose Proterozoic metasedimentary rocks, Proterozoic granite, Paleozoic sedimentary rocks, and Tertiary strata in various combinations (Bauer and Kelson, 2004a). During the Late Cretaceous Laramide orogeny, west-up, right-lateral displacement formed north-trending grabens that were later filled with sediments of the Picuris Formation (Bauer and Kelson, 2004a; McDonald and Nielsen, 2004). Kinematic evidence of west-down and right-lateral slip involving the youngest part of the Picuris Formation and increasing throw to the north suggests that early rift development included dextral-oblique movement on north-south faults of the Picuris-Pecos fault system (Bauer and Kelson, 2004a; McDonald and Nielsen, 2004). Displacement of a 5.7-Ma basalt southwest of the study area indicates fault activity as young as late Miocene (Bauer et al., 2005).

Because the Picuris-Pecos fault system is a major crustal boundary that is interrupted at the Embudo fault zone, it likely has a northward extension into the southern San Luis Basin. Bauer and Kelson (2004a) speculated that the northward extension underlies and controls the much younger Los Cordovas faults within the Cañon section at the large bedrock salient in the range front where the Cañon section meets the Sangre de Cristo fault near the intersection with the Picuris-Pecos fault system (Fig. 2; Kelson et al., 2004a).

The exposed Embudo fault zone is several kilometers in width (Bauer and Kelson, 2004a, 2004b; Kelson et al., 2004a). The complex patterns in the styles of deformation reflect large variability in local displacements and net slips along strike (Kelson et al., 2004a). Reverse thrust relations are well documented in the southwestern part of the study area where the Embudo fault zone makes a bend from northeast to east (Muehlberger, 1979; Machette and Personius, 1984; Kelson et al., 2004a; Koning et al., 2004). Kelson et al. (2004a) demonstrated that these are local features associated with a flower structure and that the overall sense of movement is left lateral.

The timing and amount of lateral displacement along the fault zone is poorly known. On the basis of sedimentological evidence, previous workers inferred that the Embudo fault zone formed in concert with the uplift of the Picuris Mountains in late Miocene (Manley, 1978; Muehlberger, 1979; Dungan et al., 1984; Ingersoll et al., 1990; McDonald and Nielsen, 2004) or early Miocene (Bauer and Kelson, 2004a; Koning et al., 2013). Kinematic observations from areas southwest of the study area show that left-lateral slip has dominated since ca. 12–11 Ma (Steinpress, 1981; Aby and Koning, 2004; Kelson et al., 2004a). Within the study area, Muehlberger (1979) approximated 3 km of total vertical offset between Proterozoic rocks uplifted in the Picuris Mountains and the deep basin to the north using preliminary gravity-data estimates for basin thickness. Bauer and Kelson (2004b) estimated 105 m of vertical throw on a 3-Ma basalt just southwest of the study area. Combined with detailed measurements of rake, they estimate 35 m per million years (m/m.y.) and 102 m/m.y. for the vertical and horizontal components of slip, respectively. The youngest fault activity displaces fan deposits that are latest Pleistocene to Holocene in age (Kelson et al., 2004a).

The coincidence of the Embudo fault system with the Jemez lineament, a more regional alignment of structural zones and Pliocene volcanism (Fig. 1), suggests that its development may have been controlled, at least in part, by a preexisting crustal fabric (Aldrich, 1986; Koning et al., 2004). The Jemez lineament is considered to be the manifestation of a crustal weakness along which tectonic and magmatic activity has concentrated since 15 Ma (Lipman, 1980; Aldrich and Laughlin, 1984). The crustal weakness may be a suture of accreted crust that formed during the Proterozoic (Magnani et al., 2004).

**Embudo Fault Zone**

The northeast-striking Embudo fault zone is ~64 km long and links the west-dipping Sangre de Cristo fault system in the southern San Luis Basin to the southeast-dipping Pajarito fault in the northern Española Basin (Fig. 1). The southwest-to-northeast transition between the northeast-striking Embudo fault zone to the north-striking Sangre de Cristo fault system appears gradual (Kelson, 1997; Bauer and Kelson, 2004a). Slip on the Embudo fault is mainly left-oblique on steep fault strands, with dominantly lateral slip transitioning to dominantly northwest-down normal slip as it approaches the Sangre de Cristo fault near the intersection with the Picuris-Pecos fault system (Fig. 2; Kelson et al., 2004a).

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**Sangre de Cristo Fault System**

The Sangre de Cristo fault system refers to a series of west-dipping normal faults that form the ~200-km-long eastern rift border of the San Luis Basin in Colorado and New Mexico. It is divided into sections on the basis of differences in geomorphic expression, including the Hondo and Cañon sections in the study area (Fig. 2; Kelson et al., 1998b, 1998c). The Hondo section meets the Cañon section at the large bedrock salient in the range front where the
Rio Pueblo de Taos emerges from the mountains. The Hondo section extends northwestern another 15 km, out of the study area. The 14-km-long Cañon section merges with the Embudo fault zone at the intersection of the Picuris–Pecos fault system, where Rio Grande del Rancho emerges from the mountains (Fig. 2; Bauer and Kelso, 2004a).

North of the bedrock salient at Rio Pueblo de Taos, fault strands in the Hondo section form branching patterns as much as 2 km wide that wrap around a bedrock protrusion in the mountain front. The pattern diverges locally from the overall strike of N 30° W of the Hondo section north of the study area (Fig. 1). The range-front morphology is steep and linear, with prominent fault scarps developed on Pleistocene to Holocene fan material within the study area (Machette and Personius, 1984; Menges, 1990; Kelso et al., 1999b).

The Cañon section is also composed of a series of branching fault strands that extend as much as 2 km in width. Unlike the Hondo section, the Cañon section strikes about N 20° E. Prominent scarps developed on alluvial fan surfaces suggest multiple events have occurred since late Pleistocene (Bauer and Kelso, 2004a; Kelso et al., 2004b). Fault planes typically dip steeply (>60°) west to northwest. Slickenlines are dominantly dip slip with increasing left-oblique slip in the transition zone to the Embudo fault zone.

**Los Cordovas Faults**

The Los Cordovas faults are generally north-striking, west-down normal faults that are exposed within a 5–6-km-wide zone interior to the Taos embayment (Fig. 2). A wider zone (5–9 km) extending farther east is suggested from photo reconnaissance and aeromagnetic data (Machette and Personius, 1984; Grauch et al., 2004; S. Personius, USGS, 2015, written commun.).

Several of the mapped Los Cordovas faults form west-facing, eroded scarps 15–30 m high that juxtapose Servilleta Basalt against downthrown Quaternary sedimentary units (Fig. 3; Bauer and Kelso, 2004a). Some faults likely have greater throw in the subsurface (Machette and Personius, 1984). The most recent fault movement may be as old as early(?!) Pleistocene and no younger than middle Pleistocene (Machette and Personius, 1984; Bauer and Kelso, 2004a). Fault planes are poorly exposed, although fault dips of 89° and 45° have been reported (Lambert, 1966 and Bauer and Kelso, 2004a, respectively).

**DATA SOURCES**

**Well Data**

Lithologic information is available from a number of deep (>150 m) water wells east of the Rio Grande within the study area (Fig. 3; Table 1). In addition to wells drilled for domestic and municipal supply, many of the deep wells were drilled by Federal agencies to understand and monitor groundwater as part of water-rights settlements. Several hydrogeologic studies have been undertaken to compile and synthesize the lithologic information from the deep wells, which also provide a subsurface stratigraphic framework for the Taos area (Bauer et al., 1999; Benson, 2004; Drakos et al., 2004). Information from these and additional water wells was compiled in recent hydrogeologic studies undertaken by New Mexico Bureau of Geology and Mineral Resources (Bauer et al., 2014; Johnson et al., 2016). As part of these studies, well locations were verified on the ground and lithologic information was reassessed from driller’s logs.

For the current study, we used well information from the previous studies, supplemented by data provided courtesy of Glorieta Geoscience, Inc. (GGI). Locations for a subset of these wells are shown on Figure 3 and listed in Table 1, along with information about basal interval penetrated by the wells. During the course of the current study, lithologic picks for the Town Yard well were revised from previous work, which reported the well had encountered Paleozoic bedrock at 219-m depth (e.g., Bauer and Kelso, 2004a; Drakos et al., 2004).

We concluded that the well did not reach bedrock, based on comparisons of well logs and lithologic cuttings with those of the nearby BOR-3 well, analysis of gravity data between these wells, and estimates of physical properties from borehole logs. Moreover, detailed examination of archived well cuttings by D. Koning and S. Aby (New Mexico Bureau of Geology and Mineral Resources [NMBGMR], 2013, written commun.) suggested that both the BOR-3 and Town Yard wells penetrated a gravel sequence within the Miocene Tesuque Formation at the interval formerly considered as Paleozoic rocks.

**Gravity Data**

Gravity data for the study area were compiled from several sources. Vintage regional data, which are archived from data collected by various workers since the 1960s, were retrieved from the PACES online database (http://gis.utep.edu, accessed November 2006). In 1999 and 2000, NMBGMR collected a large number of gravity stations at close spacing along a network of profiles on Taos Pueblo lands. Problems with the positioning information and data processing of the data were later resolved (B. Drenth, USGS, 2011, proprietary data). Additional gravity data were collected to fill critical gaps in coverage within Taos Pueblo and the Town of Taos in 2010 and 2015 (Drenth, 2016). Some of the new stations reoccupied and ultimately replaced data from the vintage database.

Standard Bouguer corrections (Blakely, 1995) were applied for each gravity station using a reduction density of 2670 kg per cubic meter (kg/m³). Terrain corrections were applied within a 167-km radius to the data at each station using digital terrain data with a resolution as fine as 10 m. The station data were then interpolated onto a grid at a 500-m interval. In order to focus on density variations within the upper crust, an isostatic regional field was removed from the terrain-corrected Bouguer gravity grid using parameters established for New Mexico by Heywood (1992). The resulting “isostatic residual gravity map” (Fig. 4) generally isolates the gravity effects produced by sources within the upper 10 km of the crust (Simpson et al., 1986).
### TABLE 1. SELECTED WELLS FOR THE STUDY AREA

<table>
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<th>Well identifier*</th>
<th>Easting† (m)</th>
<th>Northing† (m)</th>
<th>Surface elevation† (m)</th>
<th>Total depth (m)</th>
<th>Elevation at total depth (m)</th>
<th>Basalt depth intervals§ (m)</th>
<th>Elevation, top of basalt (m)</th>
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**Notes:** Well data are from Bauer et al. (2016) and Johnson et al. (2016). All wells except Alcon, Ruckendorfer, and TV-230 penetrate only basin fill with or without interbedded basalt. Alcon reaches Proterozoic quartzite at total depth; Ruckendorfer reaches Paleozoic shale at 180 m depth; TV-230 is entirely within Proterozoic basement.

*Well identifiers are used to label well locations on figures.

†Easting, northing, and elevation refer to Universal Transverse Mercator projection, zone 13, World Geodetic System 1984 (WGS84).

§TD—total depth.
Figure 4. Gravity map of the study area after removal of the isostatic regional field of Heywood (1992), showing station locations. Note the nonlinear color display to enhance variations. Limit of shallow Servilleta Basalt (dark-gray, long-dashed line) is interpreted from aeromagnetic data (Fig. 5). A few artificial data points were constructed to constrain the 3D modeling, which extended outside the study area (see text). Only one of these constraint points is located within the study area (southwest corner).
Magnetic Data

Magnetic data used for this study come from three aeromagnetic surveys, a regional compilation of aeromagnetic data, and a ground magnetic traverse. High-resolution aeromagnetic data are available over the basin area of the study area (yellow dashed outline on Fig. 5). The eastern, western, and far northern portions of this area are covered by the Taos (Bankey et al., 2004), Taos West (Bankey et al., 2007), and central San Luis (Bankey et al., 2005) aeromagnetic surveys, respectively. The latter two surveys were flown by fixed-wing aircraft; the Taos survey was flown by helicopter. All three surveys were flown along traverse lines oriented east-west, spaced 200 m apart, and nominally 150 m above ground. Orthogonal lines were flown north-south at 1000-m spacing. After flight-line data from the surveys were processed to remove known Earth’s field variations and reduce ordinary data-acquisition errors, they were interpolated onto a 50-m grid. The gridded data were analytically continued from the variable observation surface to a surface consistently draped 100 m above ground. To extend the aeromagnetic data coverage outside of the area of high-resolution survey coverage and over the mountain flanks, gridded data from a compilation of regional aeromagnetic surveys (Kucks et al., 2001) for the State of New Mexico were merged. A noticeable reduction in resolution of anomalies occurs outside the boundary of high-resolution survey coverage (Fig. 5).

A standard reduction-to-pole (RTP) transformation was then applied to the gridded data using a declination of 10° and inclination of 64°, which is the general orientation of the Earth’s field in the Taos area. Reduction-to-pole transformations correct for shifts of anomalies away from the centers of their magnetic sources; these shifts are an effect of the oblique orientation of the measured magnetic field at high latitudes with respect to Earth’s surface (Blakely, 1995). This type of correction is useful for Oligocene and younger volcanic fields, where remanent magnetizations are expected to have orientations that are within ~25° of the Earth’s field direction (Bath, 1968). Figure 5 is a color shaded-relief image of the final, RTP aeromagnetic data.

In addition to aeromagnetic data, a 6-km ground magnetic traverse was acquired across the Embudo fault zone, where high-resolution aeromagnetic coverage was not available (Fig. 5). The data were acquired as continuous total-field measurements from a cesium-vapor magnetometer mounted on a pole held at head height as the operator traversed the ground on foot. Measurements were acquired continuously except for one interruption to cross a major road (State Highway 68) that was bordered by berms and barbed-wire fences. After collection, data spikes were removed and data were interpolated across the 110-m gap in measurements. The data were then upward continued by 100 m, gridded, and merged with the total-field aeromagnetic grid so that a straight profile could be extracted for modeling. Some loss in data integrity is expected as a result of this procedure, because (1) accurate upward continuation relies on knowledge of data in a broader area than was sampled along the ground (Blakely, 1995), and (2) bends in the ground traverse forced the data extraction to pull data from extrapolated gridded data. To avoid compounding the problem of using extrapolated or interpolated data, RTP transformation was not applied, because this operation also relies on knowledge of data in a broader area to be accurate.

Key Geologic Units for Gravity Interpretation

Physical properties of geologic units are important for geophysical interpretation because they provide the tie between lithology and geophysical fields. For gravity data, the applicable physical property is bulk density, which is the overall mass per unit volume of rocks, sediments, and their pore spaces. For magnetic data, the applicable physical property is total magnetization, which is determined by the quantity of naturally occurring magnetic minerals in rocks and sediments and the nature of their permanent magnetizations.

Figure 6 shows a stratigraphic column of generalized geologic units expected to lie in the subsurface of the study area in comparison to graphs of density and magnetic properties of specific geologic units or rock types. The graphs summarize physical properties measured or estimated from rocks or geologic units within the San Luis or neighboring basins, tabulated in Tables A1 and A2 in the Appendix. The graphical portrayal of the physical properties facilitates the identification of the key geologic units for gravity and magnetic interpretation, that is, those units to which each geophysical method is most sensitive.

Physical Properties and Geologic Units

Physical properties of geologic units are important for geophysical interpretation because they provide the tie between lithology and geophysical fields. For gravity data, the applicable physical property is bulk density, which is the overall mass per unit volume of rocks, sediments, and their pore spaces. For magnetic data, the applicable physical property is total magnetization, which is determined by the quantity of naturally occurring magnetic minerals in rocks and sediments and the nature of their permanent magnetizations.

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Key Geologic Units for Gravity Interpretation

The key geologic units for gravity interpretation are the combined sedimentary deposits within the basin, encompassing the Picuris Formation, Santa Fe Group (including the Lama formation) and surficial deposits. Many previous workers have recognized that the low bulk densities of poorly consolidated sedimentary deposits that infill basins composed of moderate- to high-density bedrock produce gravity lows with amplitudes that are generally proportional to basin-fill thickness (e.g., Cordell, 1978, 1979; Birch, 1982; Keller et al., 1984; Bott and Hinze, 1995).

Logs from various boreholes within the Albuquerque and Española Basins show that densities of the Santa Fe Group are systematically much lower (~400 kg/m³ difference) than the older rock types that compose the basin floor and rift flanks (Grauch and Connell, 2013). Because lithologic types appear fairly similar for the Santa Fe Group in the Albuquerque, Española, and San Luis Basins, we use the simplified density-depth function derived by Grauch and Connell (2013) to a depth of 2.25 km in the top portion of Figure 6B to compare densities. Density information for the Picuris Formation is limited to estimates from a sonic log in one well (Bauer, 2016). Using a common empirical relation (Gardner et al., 1974), densities ranging from 2200–2300 kg/m³ were estimated for the Picuris Formation from the sonic log in the depth range of 100–366 m. No systematic variation of density with depth was observed. These densities and shallow depths are only slightly higher than the density-depth function for the Santa Fe Group (Fig. 6B). Thus, the Picuris Formation cannot be reliably distinguished from the Santa Fe Group on the basis of density.
Figure 5. Color shaded-relief image of merged aeromagnetic data after application of the reduction-to-pole transformation. Note the nonlinear color display to enhance variations and the shading to enhance linear features (illumination is from the east). The white, long-dashed line delimits the inferred area of shallow Servilleta Basalt, indicated by the high-frequency aeromagnetic patterns. The dashed yellow line is the limit of high-resolution aeromagnetic surveys. Location of magnetic ground traverse is indicated by the white-outlined blue line along profile A–A′. Label “Pmag site” along the Rio Grande is where Brown et al. (1993) analyzed paleomagnetic measurements for a vertical section of Servilleta Basalt.
Figure 6. Densities and total magnetizations for geologic units of the study area. (A) Conceptual stratigraphic column for geologic units and their time relations to rift formation. TPVF—Taos Plateau volcanic field; SJVF—San Juan volcanic field. (B) Graphical representations of ranges of densities for specific units or rock types from Table A1; depicted as a density-depth function for the synrift deposits. Values mark median values, considered to be the best representation of the overall density of the unit. Densities are compared to the average crustal density of 2670 kg/m³ (red dashed line). (C) Graphical representations of potential ranges of total magnetizations for specific units or rock types from Table A2. Note the log scale, which allows for better comparisons at low magnetizations but deemphasizes the large differences between strongly and weakly magnetic units. Qualitative categories of total magnetizations for aeromagnetic interpretation, from effectively non-magnetic to strongly magnetic for consideration, are from Bath and Jahren (1984).
Due to the variability of thicknesses and extents of multiple flows of Servilleta Basalt, we decided to estimate density for the combined package of Servilleta Basalt and intervening Lama formation rather than consider the effects of high-density (~2700 kg/m³) individual flows. To do so, we assumed that the variability of densities within the Servilleta-Lama package would be similar to that found within Guadalupe Mountain, a hill composed of trachytes and interbedded sediments at the east-central part of the Taos Plateau, northwest of our study area. An average density of 2450 kg/m³ for the hill (total relief of ~1200 m) was estimated from geophysical analysis (Grauch et al., 2015). Even if Guadalupe Mountain is not a realistic analog for the Servilleta-Lama package, the range of depths of the package is fairly small compared to the overall thickness of the Santa Fe Group within the interior of the basin (Fig. 6B). Thus, the gravity effects caused by any density errors are small compared to the gravity effects of the larger volume of sedimentary fill.

Figure 6B shows that median density values for pre-rift geologic units (except for the pre-rift lower member of the Picuris Formation) have a significant density contrast with the Santa Fe Group (>150 kg/m³) at all depths. Although the range of densities for the San Juan volcanic field rocks overlaps with Santa Fe Group densities, the range was determined by considering typical densities for two dominant rock types within the volcanic field: ignimbrites (2200 kg/m³) and andesites (2500 kg/m³) (Table A1; Drenth et al., 2012). Because the andesites have much greater volume than the ignimbrites (Drenth et al., 2012), the density of the greatest volume of San Juan volcanic rocks that may be present within the study area likely fall within the higher end of the density range (>2450 kg/m³).

**Key Geologic Units for Magnetic Interpretation**

Total magnetization is the vector sum of two components: remanent and induced (Blakeley, 1995; Hansen et al., 2005). The induced component is a function of the quantity of magnetic minerals (commonly magnetite) and is a vector that is always oriented parallel to the present-day Earth's magnetic field. The quantity of magnetic minerals is proportional to magnetic susceptibility, a property that can be measured from hand samples or outcrops. Typical magnetic susceptibilities for geologic units in the study area are tabulated in Table A2 of the Appendix. Induced components in Table A2 are obtained by multiplication (with proper unit conversion) of the magnetic susceptibilities with the amplitude of the Earth's magnetic field in the study area (assumed to be 51,700 nanoteslas).

The remanent component represents the vector sum of all permanent magnetizations held by the magnetic minerals, which have fixed directions irrespective of the ambient magnetic field (McElhinny, 1973). Remanent components that are generally aligned with or opposite to the present-day Earth's field are considered to have normal or reversed polarity, respectively. The remanent component is determined from paleomagnetic laboratory measurements of oriented samples. Measurements of remanent magnetization of rocks in the study area are limited to the Servilleta Basalt (Brown et al., 1993), requiring different approaches to estimating remanent components or total magnetizations for other geologic units, as explained in Table A2. For example, remanent components for sedimentary units are considered negligible based on rock type. For volcanic units, observed aeromagnetic field data over hills composed of the volcanic unit can be compared to data computed from magnetic models of the same hills to estimate total magnetization directly. Estimates of minimum and maximum total magnetizations (which we call the potential range because vector components were mostly unknown) were then tabulated for particular geologic units in the study area (Table A2) and are summarized graphically on Figure 6C.

The key geologic units for magnetic interpretation are Tertiary volcanic rocks, which are considered to be magnetic to strongly magnetic for purposes of aeromagnetic interpretation (Fig. 6C). Of these, Servilleta Basalt is the most important, because it is the most widespread in the study area. Remanent components are more than an order of magnitude greater than the induced components for Servilleta Basalt and Middle Tertiary volcanic rocks and for some of the intermediate-composition volcanic rocks (Table A2). Considering that remanent polarities of these rocks can be normal or reversed (Brown et al., 1993; M.R. Hudson, USGS, 2012, written commun.; Walker et al., 2012) they are expected to produce both positive and negative, large-amplitude aeromagnetic anomalies in the study area. Aeromagnetic anomalies reflecting remanent polarities set at the time of cooling have been recognized in the region previously (Grauch and Keller, 2004; Thompson et al., 2006; Drenth et al., 2011; Grauch et al., 2015; M.R. Hudson, USGS, 2015, written commun.).

We have assumed that the induced component is the dominant component for Proterozoic rocks (Table A2) because (1) the long time since their formation increases the likelihood that remanence has been reduced significantly due to heat in the crust (McElhinny, 1973), and (2) large-amplitude aeromagnetic anomalies over them are mostly positive in the region (Cordell 1976; Grauch and Keller, 2004). Aeromagnetic anomalies associated with these rocks are mostly regional and located outside the study area. For these reasons and from experience after aeromagnetic analysis during the course of this study, we consider the effects of the most magnetic Proterozoic rocks to be only locally important within our study area.

Magnetic susceptibilities of the Santa Fe Group and Picuris Formation are at the high end of common ranges of magnetic susceptibilities for sediments (Hudson et al., 2008; Grauch and Hudson, 2011), likely due to the large percentage of volcaniclastic content. Faults that juxtapose units of the Santa Fe Group with differing magnetic susceptibilities are easily detected by high-resolution aeromagnetic surveys where no volcanic rocks are present (Grauch and Hudson, 2007). However, aeromagnetic anomalies are not produced where there is no contrast in magnetic properties between juxtaposed units, even if both units have high magnetic susceptibilities. Because most of the synrift sediments in the study area have high magnetic susceptibilities, it is likely that this latter scenario is common.
METHODS

3D Basin Model from Gravity Inversion

Gravity data are particularly effective in determining the subsurface configuration of structural basins in the Rio Grande rift, owing to the generally large contrasts between low-density sedimentary fill and surrounding high-density crystalline bedrock (Cordell, 1979; Keller et al., 1984). We use a gravity inversion developed specifically for sedimentary basins by Jachens and Moring (1990) that takes advantage of these large density contrasts. The technique has been used effectively in several other basins of the Rio Grande rift (Grauch et al., 2008; Drenth et al., 2013; Grauch and Connell, 2013). It is based on separation of observed data into regional and residual components, where the residual component represents the gravity effects of the thickness of basin fill, and the regional component represents the effects of density variations in bedrock underlying and adjacent to the basin (Blakely, 1995). Important advantages of this technique are that it (1) incorporates variations of density with depth known from boreholes, (2) satisfies independent geological and geophysical constraints, (3) treats the rift basins separately from the complications of density variations in pre-rift rocks as much as possible, and (4) accommodates 3D aspects of basin shape.

The 3D gravity inversion of Jachens and Moring (1990) was applied to isostatic residual gravity data for the San Luis Basin using computer program DEPTH2BS, version 1.6.8 (B.A. Chuchel, U.S. Geological Survey). Input included the density-depth function for basin fill that was developed for the Albuquerque Basin (Grauch and Connell, 2013; Fig. 6B), wells that reached below basin fill, and estimated basin thicknesses from seismic data (none of which are in the study area; Drenth et al., 2013). A few additional, empirical constraints were applied in the extreme southwestern part of the study area to regulate how the method dealt with the large gravity and topographic gradients in the area (green, circled X in the lower left corner of Fig. 4).

We constructed a structural elevation grid representing the basin floor (Fig. 7) by subtracting thickness values resulting from the 3D inversion from a smoothed digital terrain model. The terrain was smoothed so that artifacts of topography were not introduced during the subtraction. In order to smooth extreme topographic relief in the basin areas (e.g., the Rio Grande gorge) without overly smoothing the relief in the mountainous rift flanks, we separated the digital topography grid into basin and non-basin areas. Lowpass filters were applied to the separated grids, but the filter was more extreme for the basin area grid. The smoothed grids were then merged together before subtraction of the thickness grid.

General limitations of the 3D gravity inversion approach are discussed in detail in Grauch and Connell (2013, their appendix E). The most significant limitations for the study area are that (1) abrupt thickness variations across normal faults are extremely smoothed, and (2) thicknesses are somewhat underestimated where high-density volcanic rocks are present. Smooth rather than abrupt thickness variations across normal faults arise from the reliance on grids to model thickness. Small-offset faults may not be recognizable in the model, and multiple faults may appear as a sloping surface. Despite these problems, major basin boundaries are still readily identifiable, albeit somewhat ambiguous regarding dip and presence of a single versus multiple faults. Modeled thicknesses are underestimated where Servilleta Basalt overlies the sedimentary fill. Attempts to remove the gravity effects of the basalt before inversion suffered from uncertainty in the extent and thickness of the Servilleta-Lama package across the study area and appeared to introduce error rather than correct for it. Instead, we qualitatively recognize from 2D modeling experiments that the 3D basin model surface can be as much as 150 m too shallow under a 200-m thickness of Servilleta Basalt. Because the thickest basin layers occur mostly where the basin is >1 km deep, this error is less than 15%—within the errors expected to arise from smoothing of normal faults at the basin margins.

The pre-rift age of the lower part of the Picuris Formation and its similar density to that of rift-age basin fill introduce some error in the assertion that the 3D model represents a rift subsurface. Maximum estimates of the thickness of the lower Picuris Formation range from 300 to 400 m (Aby et al., 2004; McDonald and Neilsen, 2004). If the lower Picuris Formation evenly blanketed the study area before 25 Ma, the 3D model may overestimate rift fill thickness by as much as 400 m. However, well data from the northern flanks of the Picuris Mountains suggest that some erosion of the deposits occurred at higher structural levels (Bauer et al., 2016). Combined with the limitations of modeling Servilleta Basalt that underestimate basin-fill thickness, both errors may approximately cancel each other within the deep parts of the 3D model.

Aeromagnetic Interpretation Methods

Interpretation of the aeromagnetic map centered on recognition of patterns that correspond to the key geologic units, mainly Servilleta Basalt. Patterns of strongly negative versus strongly positive magnetic-field values can be used to help distinguish basalt with reversed or normal polarity, respectively. Comparison of aeromagnetic patterns to topographic patterns also helps determine the polarity of the total magnetization. Aeromagnetic patterns that have similar shape to topographic features indicate normal polarity. Aeromagnetic patterns that are similar to topographic features but are inverted (e.g., an aeromagnetic low is located over a topographic high) indicate reversed polarity.

Patterns with high-frequency variations and large ranges in amplitude are indicative of magnetic rocks that are shallow. In contrast, broad variations with small ranges in amplitude are indicative of magnetic rocks that are deep. Aeromagnetic lineaments, or alignments of gradients or elongated anomalies, generally correspond to steeply dipping, linear geologic contacts, such as faults.

Computing the horizontal gradient magnitude (HGM) of the RTP aeromagnetic grid facilitated recognition of aeromagnetic lineaments. The method produces a map that forms ridges over steep gradients, intuitively similar to finding the inflection point of a curve (Cordell and Grauch, 1985). However, interpretation of these maps is not straightforward for stacked magnetic layers (Grauch and Hudson, 2011), such as the multiple basalt flows represented by Servilleta Basalt, and requires an experienced interpreter.
Figure 7. Structural elevation of the 3D gravity model of the Taos subbasin. The elevations represent the basin floor and rift flanks with basin fill removed. The limit of shallow Servilleta Basalt is inferred from Figure 5.
2D Profile Models

Approach

Three profile models were located across the study area (Figs. 3–5) to examine basin features. For each profile, data were extracted from the isostatic residual gravity and RTP aeromagnetic grids. Locations were chosen to optimize constraints from wells and to demonstrate certain concepts. Geologic relations and geophysical representations were developed synergistically, using constraints from geologic mapping, well data, 3D gravity modeling, and magnetic modeling. Structure of the basin floor was constrained by gravity modeling, using the 3D basin model as a general guide. The only limited seismic data that were available provide constraints on B–B'. The configuration of shallow basalts is constrained by well data, with help from magnetic modeling for A–A'. Numerous unknown variables on B–B' and C–C' led us to use magnetic modeling only as a general guide to configuring the shallow basalts, explained in the Modeling Limitations section below.

Regional Field Removal

Following standard practice, a regional field was removed from the profile data before 2D profile modeling. For consistency between profiles, the regional fields were extracted from grids. For the gravity profiles, the computed grid resulting from the final step of the gravity inversion was used as the regional field. This choice allows the modeling to focus on aspects of the basin shape. As a result, density contrasts within the basement or caused by the Paleozoic sedimentary section above the basement need not be included in the modeling. For this reason, model bodies representing basement or basement overlain by Paleozoic rocks were all assigned a density of 2670 kg/m³.

In all profiles, except for A–A', the regional field removed from the RTP aeromagnetic data was the longest wavelength output of a matched filtering process (Syberg, 1972; Phillips, 2001). Matched filtering uses the frequency spectrum of the data to design filters that match different parts of the spectrum. In this case, the regional field was developed from matched filtering of the longest wavelengths from an RTP aeromagnetic compilation of surveys that cover the San Luis Basin from Taos, New Mexico, to Alamosa, Colorado. The regional field is assumed to reflect broad variations in basement magnetization; so its subtraction from the observed data serves to remove these effects. Thus, model bodies representing basement or basement overlain by Paleozoic rocks were all assigned a magnetic susceptibility of zero. On the other hand, some basement-related magnetic anomalies remain in the profile data in areas where the basement is fairly shallow because the wavelengths of the shallow anomalies are too short to be included in the long-wavelength component of the data. This was the case for C–C', where an intrabasement magnetic body was required to fit the data at the mountain front.

The matched-filtered regional field was not removed from the data for A–A', because these data were not RTP, as discussed in the section on Geophysical Data. In this case, magnetic contrasts in the basement were included in the model.

Magnetic Depth Estimation

Using the principle that shallow sources generate anomalies with steeper gradients than those produced by similar deep sources, aeromagnetic data are commonly analyzed to estimate depths to tops of magnetic sources. Grauch et al. (2004) applied several different depth-estimate methods to gridded aeromagnetic data for the Taos survey, providing an overall view of depths to magnetic sources that is consistent with depths to basalts found in wells. For this study, we used the computer program PDEPTH on profile data (Phillips, 1997), which incorporates a variety of methods so that results can be examined together. One of these, the multiple-source Werner method (Hansen and Simmonds, 1993), gave the most robust results.

The multiple-source Werner method analyzes profile data in a window that slides across the profile data. Window size can be varied by the user to focus on shallow or deep sources (small or large window sizes, respectively). The solutions resulting from the analysis of each overlapping window of data cluster at the top of a source, giving a good estimate of its depth where the source is isolated and the data are free of noise. We tested solutions for both contact and sheet source assumptions. A contact source is a vertical step with infinite depth extent. A magnetic sheet source has zero thickness and extends to infinity in three directions. From fundamental principles, contact sources are always estimated at shallower levels than are sheet sources. Because most magnetic sources are better represented by a shape that is somewhere between a contact and a sheet, a pragmatic approach is to consider that estimates using contact and sheet sources provide minimum and maximum estimates of depth, respectively (Hansen and Simmonds, 1993).

The large variability caused by layered magnetic basalts that are stacked over each other in the study area presents a challenge for any magnetic depth estimation method (Grauch and Hudson, 2007). Thus, the results of the multi-source Werner method were used only as a general guide for modeling purposes. Solutions resulting from a variety of window size and contact versus sheet source assumptions are shown. The sheet source assumption tended to provide more solutions than for the contact source assumption.

Modeling Limitations

The 3D gravity model provides the fundamental building block of the subsurface interpretations, because it is a good representation of the overall basin geometry. However, it cannot distinguish between steeply sloping versus faulted basin margins. Therefore, the 3D model was used only as a general
guide for constructing basement faults on the 2D models. Faults on the 2D models were located using mapped faults or local gradients in the observed data; then fault displacements were constrained by the fit of the calculated gravity curve to the observed curve.

Modeling of shallow Servilleta Basalt for the profiles proved difficult because of the high sensitivity of the magnetic modeling to multiple, unpredictable unknowns: (1) variations in thickness and total magnetization of individual layers, and (2) variations in thickness, number of layers, and total magnetizations between layers within a stack of multiple layers. Only small variations in any one of these variables produce large differences in the calculated magnetic curve. One could fit the magnetic curve by arbitrarily choosing to vary only one of these variables, but the result may not be meaningful. Moreover, three-dimensional lateral variations in the basalts are only poorly captured from a profile across the aeromagnetic data; 2D modeling may not properly represent the lateral variations. Instead, we rely mostly on well data to constrain thickness and depth to basalts for the profile models. Magnetic depth estimates provided some general guidance on depth to the bottom of stacked basalts layers and lateral terminations of the basalts. Gravity modeling provided little help because the data are fairly insensitive to variations in depth and thicknesses of the basalts in comparison to variations in thickness of the basin fill.

### GEOPHYSICAL INTERPRETATIONS

In this section, the geophysical data are integrated with geologic mapping and well data to build different aspects of the picture of the buried geometry of the Taos subbasin. We begin by describing the aeromagnetic expression of Servilleta Basalt, because it is a prerequisite for understanding the geophysical expression of faults at the subbasin margins. We then address the subsurface configurations of subbasin margins associated with the major rift fault systems and their relations to intrabasin basaltic layers. We end with a view of the overall subbasin geometry compared to shallow fault patterns.

#### Aeromagnetic Expression of Servilleta Basalt

The highly varied, “busy” anomaly pattern observed in the aeromagnetic image in the central and northwestern region of the study area (Fig. 5) is interpreted as the expression of near-surface volcanic rocks related to the Taos Plateau volcanic field, primarily Servilleta Basalt. This interpretation is supported by (1) exposures of Servilleta Basalt scattered throughout the area and in the Rio Grande gorge (Fig. 3); (2) their strong magnetizations (Fig. 6C, Table A2 in the Appendix); and (3) presence and absence of basalt in shallow water wells that match the differences in the aeromagnetic patterns (Grauch et al., 2004). The eastern and southern limit of the area interpreted as shallow Servilleta Basalt is outlined by a long-dashed line (Fig. 5). Within the interpreted area of shallow Servilleta Basalt, well data show evidence of multiple flows interbedded with sediments, where the shallowest flows are mostly within 100 m of the surface, or at elevations of 203–210 m above sea level (Table 1). The deepest basalt intervals in these wells generally extend downward to elevations of ~1900 m, giving an overall interval of Servilleta Basalt and intervening Lama formation of ~200 m.

The patterns and commonly low aeromagnetic values within the area interpreted as shallow Servilleta Basalt indicate that total magnetizations of the multiple layers are dominantly reversed polarity. This inference is supported by comparison of aeromagnetic patterns to topographic patterns and paleomagnetic measurements from the Rio Grande gorge. Topographic relief developed on Servilleta Basalt inversely correlates to the shapes of anomalies in several places, indicating that the total magnetizations of layered Servilleta Basalt in those areas are dominated by strong, reversed-polarity remanence. For example, the Rio Grande gorge, which is a topographic low incised into layers of Servilleta Basalt, correlates inversely to a strong linear aeromagnetic high across most of the gorge’s extent across the study area (Fig. 5). Paleomagnetic measurements (Brown et al., 1993) from a 132-m vertical stack of layers of Servilleta Basalt in the Rio Grande gorge (“Pmag site” on Fig. 5) show polarities from the lower 38–132 m of the section are reversed and for the top 38 m are normal.

#### Southern Subbasin Margin and the Embudo Fault Zone

A zone of northeasterly trends is evident in the aeromagnetic image between the mapped Embudo fault system and the Rio Pueblo de Taos, an area that is mostly covered by surficial deposits (L1 and L2 on Fig. 8). The northeasterly zone, and lineament L1 in particular, is remarkably straight for almost 11 km. Lineament L1 corresponds to the southern limit of the interpreted area of shallow Servilleta Basalt (long-dashed lines on Figs. 3, 5, and 7). The region of very smooth, relatively high magnetic values to the southeast of this line is interpreted as the expression of the sedimentary section without the presence of Servilleta Basalt. This interpretation is supported by deep water wells in the area (cf. Fig. 8 and Table 1). Lineament L2 generally follows Rio Pueblo de Taos (Fig. 8). Close inspection of the anomalies compared to the stream channel suggests the sources of the aeromagnetic highs are contained within the southern bank of the stream rather than a topographic effect (Fig. 8). Previous workers have speculated that a structure underlies this stream on the basis of its linearity (Muehlberger, 1979; Machette and Personius, 1984).

The orientation and linearity of anomalies within the northeasterly zone suggest that they are associated with the Embudo fault zone. The abrupt limit of the aeromagnetic patterns associated with Servilleta Basalt at aeromagnetic lineament L1 suggests that the basalt terminates against a fault. This interpretation is supported where the lineament coincides with a fault mapped for a short distance across the basalt-rimmed, Hondo Canyon in the southwestern corner of the study area (Fig. 8). The basin is not present south of the mapped fault.
To examine the interpretations of this zone of lineaments more closely, we constructed a 2D model along profile A–A′ (Fig. 9). This profile location utilizes the aeromagnetic data across the lineament zone on the north and the ground magnetic data across the mapped Embudo fault system on the south (Fig. 5). Lithologic picks from a number of water wells, which fall within 2 km of the profile location (Fig. 8), were used to help construct the geologic cross section and geophysical model. Geologic relations for this section were also guided by a number of additional parallel and crossing, geophysically constrained, geologic cross sections (not shown), which were constructed south of Rio Pueblo de Taos for a hydrogeology study (Bauer et al., 2016).

Section A–A′ (Fig. 9) shows several important features that are well constrained by the data and geologic mapping. First, wells and 2D magnetic modeling support the interpretation of aeromagnetic L1 as the southern fault termination of Servilleta Basalt. A number of other faults are inferred from the aeromagnetic data. Well data suggest that the number of flows and their thicknesses increase to the north within the fault blocks.

Second, constraints from the 3D basin model and geologic mapping suggest that the Embudo fault zone is 7 km wide along profile A–A′, with significant normal displacement (at least 1.8 km). The zone extends from mapped faults 0.5 km north of well TV-230 to the Rio Pueblo de Taos (distance 8 km to 1 km, respectively, on Fig. 9). Almost 1 km of the normal displacement is accommodated between the L1 fault and the northernmost mapped strand of the Embudo fault (between 3.5 and 6 km distance on Fig. 9).

Finally, prominent anomalies from the ground magnetic traverse (shown after upward continuation on Fig. 9 between distance 5.5 and 7.5 km) are likely caused by basaltic rocks reported deep in the Alcon well. The best fit from...
Figure 9. Geophysical model and geologic cross section along profile A–A’ (located on Fig. 8). Geologic codes from Figure 3. Modeled basalt layers are color coded by total magnetization polarity and intensity in amperes per meter (A/m). Circled A and T indicate strike-slip movement away and toward the viewer, respectively. The fault lines are wavy to indicate that they represent complex fault zones. Regional gravity field was removed before modeling.
magnetic modeling was a sheet-like source with strong, reversed-polarity total magnetization, consistent with a volcanic layer at the elevation found in the well. Total magnetization with reversed polarity is more indicative of a volcanic rather than Proterozoic source, for reasons given in the section on magnetic properties. In its position near the base of the Picuris Formation, the basaltic rocks in the Alcon well must be older than 34.5 Ma (Aby et al., 2004). Total magnetization of 5.5 A/m, derived from magnetic modeling, is consistent with magnetizations estimated for middle Tertiary volcanic rocks (Fig. 6C). Although we have not tried to depict its presence deeper than 1200 m elevation on Figure 9, our discovery of likely late Eocene lava flows at the southern part of this profile suggests that deep pre-rift volcanic rocks may exist elsewhere, deeper within the subsurface.

The details of the fault shown beneath Rio Pueblo de Taos in A’A’ are not well constrained. No fault is exposed, yet a fault is suggested by aeromagnetic lineament L2. L2 on A’A’ is represented by the prominent high-low anomaly pair between 0.5 and 1.5 distance (Fig. 9). The high of the pair follows the southern bank of the stream. We found it difficult to model the anomaly pair as faulted and displaced basalt layers while maintaining a semblance of correlation between basalts in wells along the Rio Pueblo de Taos off the profile (e.g., Bauer et al., 2016). Instead, we model the anomaly pair as an individual, normal-polarity flow restricted to the south bank of Rio Pueblo de Taos and extending westward of the stream. To explain the linearity of the high along L2, we argue that this model represents a late-stage Servilleta lava that flowed along a structurally controlled channel. Overall, correlations in the wells and magnetic modeling are difficult along the Rio Pueblo de Taos; these difficulties allow for alternate models depicting the details of this area. Stratigraphic relations that are difficult to correlate between wells are consistent with growth faulting, which is commonly observed in the area (Bauer and Kelson, 2004a).

**Eastern Subbasin Margin and the Sangre de Cristo Fault System**

Previous workers have interpreted a deep (>200 m) graben along the eastern subbasin margin (Drakos et al., 2004) by correlating basalts between wells. For example, the shallowest basalts in BOR-5 and Torreon wells, which are only 1–3 km east of the area interpreted as shallow Servilleta Basalt (Fig. 5), are more than 300 m below the elevation of 1900 m expected for the lowest Servilleta Basalt. The shallowest basalts in the BOR-3, BOR-6, and Town Yard wells are only slightly higher than 1900 m elevation (Table 1). Magnetic depth estimates also delineate this zone of deep magnetic basalt east and southeast of the interpreted area of shallow Servilleta Basalt (Grauch et al., 2004).

Three different concepts to explain the deep basaltic intervals in wells on the east are illustrated schematically in Figure 10. Previous workers’ interpretation of a graben within a half-graben (Fig. 10A) is the simplest explanation and assumes that the basalts are all Servilleta Basalt. A major, east-down normal fault between the two eastern wells is required to fit this concept, but conflicts with the gravity constraints, which show a progressively westward-deepening basin floor. A concept of a low-angle normal fault with rotated blocks also assumes all basaltic intervals are Servilleta Basalt but uses significant back tilting on rotated blocks to explain the basalts at different depths (Fig. 10B). This concept is contraindicated by seismic sections, which show subhorizontal or slightly west-tilted horizons instead of east-tilted blocks within 4–5 km of the mountain front (Fig. 10B). Moreover, there is no geologic evidence of low-angle faulting at the mountain front farther east. The third concept (Fig. 10C) is a half-graben, which assumes that the basalts in the wells do not correlate from east to west. The eastern basaltic rocks are assumed to be older than Servilleta Basalt, implying that Servilleta Basalt was never deposited on or was eroded off the structural block on the east. Making this assumption allows for a half-graben geometry that fits all the independent evidence. Implications of this scenario are that the master fault of the half-graben was previously located between the two eastern wells and then later shifted to the modern location next to the range front.

Note that all three conceptual models in Figure 10 depict listric faults that join a detachment at the brittle-ductile transition below the subsurface floor. Such a detachment is required to provide a tectonic mechanism for development of the left-lateral slip observed for the Embudo fault zone (Morley et al., 1990; Faulds and Varga, 1998; Goteti et al., 2013).

The utility of the third concept of a half-graben with different-age basaltic rocks for satisfying all the known evidence is demonstrated by geophysical models for profiles B–B′ and C–C′ (Figs. 11 and 12). Profile B–B′ crosses the subsurface near its deepest spot and utilizes several deep wells for constraints (Fig. 7). The lowest basaltic interval penetrated in the K3 well on B–B′ is vertically separated by 162 m (530 ft) from the interval intersected by the Torreon well; yet the wells are less than 1.5 km apart. If the basalt in the Torreon well is older than those in the K3 well, following the concept in Figure 10C, the modeled faults fit constraints from gravity data and a seismic-reflection section while explaining the relations hypothesized for the basaltic rocks found in the wells. A 6-km-wide zone of down-to-the-west basement faults, accommodating the cumulative displacement along the eastern subsurface margin, is modeled west of the range front.

Similar relations among faulted basalts and subbasin geometry that developed from modeling B–B′ also result from modeling profile C–C′ (Fig. 12). C–C′ crosses the subsurface margin just north of where the Embudo and Sangre de Cristo fault systems join (Fig. 2) and ties to B–B′ at the Taos Airport well. As in B–B′, hypothesizing that basaltic rocks that are found deep (180 m) in the BOR-3 and Town Yard wells are older than those found much shallower (30 m) in the Elston well only 2 km to the northwest allows a model that is compatible with both the well data and the gravity constraints.

Both profile models B–B′ and C–C′ depict basement faults stepping down into the subsurface away from the range front and then transitioning to a gentler incline on the basin floor that slopes up toward the Rio Grande. This basin-floor slope transition occurs underneath the area of shallow basalt in both models, just west of well K3 in B–B′ and near the Rio Pueblo de Taos in C–C′ (Figs. 11 and 12).
Figure 10. Schematic representations of three different concepts to explain abrupt eastward deepening of basaltic intervals observed in wells relative to the Sangre de Cristo fault zone. A checklist indicates whether independent evidence supports or permits (green check) or conflicts with (red X) each concept. Gravity data constraints are represented by the dotted magenta line. An interpretation of horizons constrained by seismic-reflection data (Reynolds, 1986) is shown for comparison at the appropriate position at the range front. Small-displacement, west-down, Los Cordovas faults are not shown. Assuming basaltic rocks are different ages (C) is the only concept where all known evidence is satisfied.
Figure 11. Geophysical section along profile B–B’. Structure of the basin floor was constrained by gravity modeling, using the 3D basin model as a general guide, and seismic data near the range front (Reynolds, 1986). Faults are implied by steps and abrupt offsets of the modeled bodies. Modeled densities and magnetic susceptibilities are given except for basalt layers, which are color coded by total magnetization polarity and intensity in amperes per meter (A/m). Older basalt (dark blue) is conceptually required to reconcile geophysical and well data. The regional fields shown were removed from the data before modeling. Magnetic depth estimates are depicted as solutions from the multi-source Werner method (Hansen and Simmonds, 1993), where the circle shows the top of a contact or sheet, as indicated. RTP—reduced-to-pole.
Figure 12. Geophysical section along profile C–C'. Description of modeling, symbols, and annotations as in Figure 11. A magnetic, intrabasement contact is indicated by the cluster of magnetic depth estimates at about distance 19 km. The modeled body has narrow extent, and may represent an intrusion within the basement. Its top may represent the contact between magnetic crystalline rocks and overlying, much more weakly magnetic, Paleozoic sedimentary rocks. RTP—reduced-to-pole.
This overall depiction of the basin floor in both models is well constrained by the gravity data. The 3D basin model shows a smoothed version of this overall geometry (purple dotted line on Figs. 11 and 12). The profile modeling builds on the 3D model depiction, yielding a more geologically reasonable depiction of basement faulting that provides a good model fit and is guided by mapped faults and by breaks indicated by the aeromagnetic and seismic data (where available). Although the details of offsets and dips of basement faults are still poorly resolved, we have confidence in the general depiction of the basin floor.

Models for both B–B′ and C–C′ show that the deep, older basalt terminates west of the range front and has little to no displacement overlying faulted basement to the west, where the older basalt is modeled underneath the younger basalt (between 16 and 18 km distance on Fig. 11 and 15–17 km distance on Fig. 12). These relations derive from several independent constraints. First, the easternmost presence of modeled older basalt in B–B′ is constrained by the proprietary seismic line that nearly coincides with the eastern end of B–B′ (Fig. 11). A strong reflection observed at the western end of the line is the only strong reflection along the line with a travel time that places it well above the gravity-constrained basement. This observation suggests that the older basalt is present at ~300 m depth in this location but is not present to the east. Correlating this interpreted basalt layer to basalt evidence in the Torreon well to the west suggests only small displacements due to basement faulting. Finally, subhorizontal older basaltic layers extending southeastward from BOR-3 well in C–C′ are supported by the aeromagnetic modeling, aeromagnetic map interpretation, and magnetic depth estimates. Magnetic depth estimates indicate a significant magnetic contrast occurs somewhere within the top 200 m of the surface at about distance 17 km (Fig. 12), which is consistent with the truncation of magnetic basaltic rocks against sedimentary basin fill.

**Northeastern Subbasin Margin**

Trends of contours in the 3D model show that the northeastern subbasin margin changes orientation going northward, from north to northwest. It is most obvious along contours that follow the locations of the Torreon, Sheepskin, BOR-5, and BOR-7 wells (Fig. 13B). Even though the north-striking part of the mapped Hondo section of the Sangre de Cristo fault deviates from the orientation of magnetic basaltic rocks against sedimentary basin fill.
entation of the subbasin margin within the study area, the northwest trend of the subbasin margin parallels the overall northwest strike of the Hondo section to the north, outside of the study area (Fig. 1; Machette and Personius, 1984; Kelson et al., 1998b; Bauer and Kelson, 2004a). A subtle gradient in the aeromagnetic image (lineament L4) has a similar northwest trend (Fig. 13A). (Note that several high-frequency anomalies in this area are caused by buildings located along NE-SW roads). In fact, L4 is located where the slope of the subbasin margin steepens somewhat (not easily apparent from the color contour maps). Thus, although the aeromagnetic expression of L4 is caused by basalt that was emplaced late in rift history, its coincidence with the major subbasin boundary suggests that it is associated with a fault system that has had a long history since early rift time.

An area of smooth aeromagnetic texture extends for ~4 km west of the Hondo section of the southern Sangre de Cristo fault (Fig. 13A). It is bounded on both east and west sides by northerly trending lineaments L3 and L5 and is centered over the northeastern subbasin margin. We interpret the broad low as evidence that the deep older basaltic rocks modeled in B–B center over the northeastern subbasin margin. We interpret the broad low (Fig. 13A) that was emplaced late in rift history, its coincidence with the major subbasin boundary suggests that it is associated with a fault system that has had a long history since early rift time. An area of smooth aeromagnetic texture extends for ~4 km west of the Hondo section of the southern Sangre de Cristo fault (Fig. 13A). It is bounded on both east and west sides by northerly trending lineaments L3 and L5 and is centered over the northeastern subbasin margin. We interpret the broad low as evidence that the deep older basaltic rocks modeled in B–B centered over the northeastern subbasin margin. We interpret the broad low (Fig. 13A) that was emplaced late in rift history, its coincidence with the major subbasin boundary suggests that it is associated with a fault system that has had a long history since early rift time.

The lineaments L3 and L5 support the hypothesis that major normal faults have been active on either side of the region of the broad low, where we interpret deep basaltic rocks overlain by sediments. On the eastern side, L5 is a curvilinear, narrow high-low aeromagnetic anomaly pair that is located between and generally parallel to the mapped range-front faults and the Rio Lucero (Fig. 13). The narrow low consistently corresponds to a distinctive zone of incoherent reflections in the proprietary seismic data (Reynolds, 1986, 1992) and to a zone where stratigraphic correlations break down between time-domain electromagnetic soundings within the top 300 m of the surface (L. Ball, USGS, 2015, proprietary data). Synthesis of these proprietary data with the aeromagnetic and gravity data suggests that the narrow aeromagnetic low coincides with (1) a major-displacement fault at the basin margin associated with the Sangre de Cristo fault system and (2) a buried paleochannel incised at the fault zone that separates basalt on the west from basement rocks on the east (Bauer et al., 2014).

On the western side of the broad aeromagnetic low, L3 is a prominent NNW, linear aeromagnetic high that lies near the eastern limit of the interpreted area of shallow Servilleta Basalt (white long-dashed line on Fig. 13). Two other similar linear aeromagnetic highs parallel lineament L3; one is 1 km and the other is 2 km farther west. The prominent linear magnetic highs can be explained by faults that disrupt layers of primarily reversely polarized basaltic layers, with negative anomalies over the layers and highs over the faulted edges. The linear anomalies parallel mapped Los Cordovas faults to the west (Fig. 2), suggesting the linear anomalies mark the easternmost extent of Los Cordovas–style faulting. These linear anomalies are more prominent and well-defined than the aeromagnetic expression of mapped Los Cordovas faults to the west (cf. Figs. 5 and 12). We explain the difference by the absence of the youngest Servilleta Basalt flows on the east and interpret this area to be outside the limit of inferred shallow Servilleta Basalt (Fig. 13A).

**Overall Subbasin Geometry and Shallow Fault Patterns**

The geophysical models show an elongated and asymmetric subbasin within the Taos embayment (Fig. 7). The subbasin is crescent-shaped, similar to the outline of the Taos embayment, and reaches a maximum depth of ~2 km below the valley floor near the maximum curvature of the crescent (X on Fig. 7), or at ~36° 26’ N latitude and 105° 37’ W longitude. Closed contours on Figure 7 suggest an imperfect D shape that is elongated on the north and southwest.

These overall subbasin shapes are the result of cumulative fault activity at the rift margins since the inception of rifting, at ca. 25 Ma. In contrast, lineaments in the aeromagnetic data represent faults expressed by offsets in shallow, magnetic volcanic rocks, mainly 5.5–3 Ma Servilleta Basalt within the interior part of the subbasin (Fig. 5). Thus, comparing the 3D model to faults inferred from aeromagnetic data can provide insights into the structure and perhaps age of shallow versus deep parts of the subbasin.

To develop a fault map from the aeromagnetic data for the study area, we computed the horizontal gradient magnitude (HGM) of the RTP aeromagnetic data (Fig. 14A). Lines drawn where faults are interpreted from the HGM map are combined with mapped faults in Fig. 14B. The overall pattern of faulting shows (1) northeasterly trending faults paralleling Río Pueblo de Taos that extend across the whole study area, (2) a northwesterly trend that intersects the northeasterly pattern near the confluence of the Ríos Pueblo de Taos and Lucero, (3) a zone of faults (both mapped and aeromagnetically inferred) that follow the entire mountain front surrounding the Taos embayment, and (4) a wide zone of northerly trending faults throughout the interior of the subbasin that is bounded on the east by the northeasterly trend. The fault patterns from geologic mapping and aeromagnetic interpretation are overlain on the 3D basin model from the gravity data on Figure 15. The northeasterly and northwesterly trends in the fault patterns generally follow the crescent-shaped eastern margin of the subbasin (compare inferred faults to the zone of green colors that parallel the mountain front on Fig. 15), suggesting that these younger, shallow faults followed the same general pattern as the older, deeper faults. The northeasterly trend is aligned with the more regional Jemez lineament (Fig. 1), suggesting a more pervasive influence of this crustal boundary than previously thought. The shallow faults that ring the mountain extend across the whole study area to the zone of green colors that parallel the mountain front on Fig. 15, suggesting that these younger, shallow faults followed the same general pattern as the older, deeper faults. The northeasterly trend is aligned with the more regional Jemez lineament (Fig. 1), suggesting a more pervasive influence of this crustal boundary than previously thought. The shallow faults that ring the mountain front deviate from the overall subbasin margin only in the northeastern part of the study area, where the deep subbasin margin follows the narrow zone of northwesterly striking shallow faults instead (e.g., L4 on Fig. 15). The wide zone of northerly trending faults in the central part of the study area occurs within the deep part of the basin. The easternmost of these
faults crosses the northeastern subbasin margin obliquely. The divergence of strikes of the shallow faults at the northeastern subbasin margin suggests that northwest-striking margin faults represent an early phase of rift-margin development, which later shifted to more northerly strikes. Alternatively, the northwest-trending rift margin could be the result of parallel, north-striking faults that (a) systematically increase in strike length to the west, and (b) consistently lose vertical throw toward their northern tips.

## DISCUSSION

Details of the relations among faulting and basaltic layers resulting from the geophysical interpretations and modeling have important implications for the evolution of the four major fault systems surrounding the Taos embayment. In this section, we discuss several of these implications and how well they are constrained.
Figure 15. Mapped (black) and interpreted (yellow) faults from Figure 14 overlain on the 3D model of Figure 7. Interpreted faults associated with aeromagnetic lineaments of Figures 8 and 13 are labeled. Limit of shallow Servilleta Basalt (white long-dashed line) is interpreted from aeromagnetic data.
Ages of Deep Basalts

An important hypothesis resulting from the modeling done to reconcile geophysical and well data is the presence of faulted basaltic rock in the eastern and southeastern subbasin margins that is older than Servilleta Basalt (Figs. 9–12). The presence of older basaltic rock is compelling at A–A′ because it is found in the Alcon well near the base of the Picuris Formation. The minimum age of the lower member of the Picuris Formation is constrained by overlying ash dated at ca. 34.5 Ma (Aby et al., 2004). These authors speculated that clasts of Tertiary intermediate to mafic volcanic rocks found in the lower member along the northern flanks of the Picuris Mountains were derived from intermediate-composition volcanic rocks from the San Juan volcanic field to the north, or from a buried, more local source.

The hypothesis that deep basaltic rocks in other wells (BOR-3 and Town Yard, BOR-5, and Torreon) are older than Servilleta Basalt has some support from preliminary microprobe examination of basalt samples from the Torreon, Town Yard, and BOR-5 wells (N. Dunbar, NMBGMR, 2014, written commun.). These basalts have dissimilar chemistry and texture to known samples of Servilleta Basalt. Moreover, basalts in the Torreon and BOR-5 wells appear similar in chemistry and texture to each other. Preliminary examination of the Torreon well basalt by laser-induced breakdown spectroscopy (LIBS) also suggests that this basalt is dissimilar from typical Servilleta Basalt, although alteration of the sample may have affected the outcome (N. McMillan, New Mexico State University, 2014, written commun.). Basalt that is penetrated by the BOR-6 well is at an elevation of 1965 m (Table 1), close to the lowest elevation expected for Servilleta Basalt. Thus, these rocks may also be older than Servilleta Basalt. Alternatively, it may represent the oldest, 5.5-Ma Servilleta Basalt. Additional laboratory analyses and comparisons of basaltic samples from wells might provide clarification on the correlations of basalts in our cross sections.

From relations developed during construction of geologic cross sections in the area of these two wells, basaltic rocks from the BOR-5 and Torreon wells appear to be within the lower part of the Santa Fe Group, or middle Miocene in age (Bauer et al., 2014). If so, the magmatism may be correlative with the episodic basaltic eruptions that occurred during 15–11 Ma in the central San Luis Basin (Miggins et al., 2002; Thompson et al., 2007, 2015) and 14–8 Ma in the northern Jemez volcanic field (Goff and Gardner, 2004; Thompson et al., 2008; WoldeGabriel et al., 2013).

Implications for the Embudo Fault Zone

Northwest-down, dominantly left-lateral slip is well documented along the Embudo fault zone in the southwestern corner of the study area (Bauer and Kelson, 2004b; Kelson et al., 2004a), yet a surprising amount of overall vertical displacement is evident from the geophysical models and geologic cross sections (Figs. 7 and 9; Bauer et al., 2016). Considering a 3.3 km elevation for the top of the Picuris Mountains near profile A–A′, the total structural relief across the Embudo fault zone is ~2.6 km, somewhat less than the 3.0 km originally estimated by Muehlberger (1979). Using this estimate, an average vertical-slip rate of 87 m per million years (m/m.y.) is required since rift initiation, at ca. 25 Ma. Alternatively, considering the 1.8 km thickness of low-density fill indicated by gravity modeling as a minimum estimate of vertical throw since rift initiation, an average vertical-slip rate greater than 72 m/m.y. is required. Using a displaced Servilleta Basalt flow as a marker, Bauer and Kelson (2004b) estimated average vertical and horizontal components of slip of the Embudo fault zone since 3 Ma to be 35 m/m.y. and 96 m/m.y., respectively. Clearly, a vertical-slip rate of 35 m/m.y. cannot account for the observed vertical relief over 25 m.y. Either vertical slip rates were considerably greater before 3 Ma, the Embudo fault has been active for much longer than 25 m.y., the estimates of vertical-slip rates since 3 Ma are too low, or a combination of these.

Aeromagnetic lineaments L1 and L2 (Fig. 8) are interpreted to represent earlier fault activity related to the modern Embudo fault zone. The continuity of these lineaments over ~11 km distance suggests that buried Servilleta Basalt flows are disrupted at, or terminate against, segments of the Embudo fault. There are two plausible explanations for the terminated basalt, both of which require significant throw: (1) lava flows ponded against an existing fault scarp, or (2) lavas flowed across the fault and were later vertically displaced and eroded off the footwall. The first explanation suggests that a very long, topographically prominent fault scarp or paleochannel existed along the fault at the time of eruption. From relations at well TV-115 along profile A–A′ (Fig. 9), ~6 m thickness of lava would have accumulated against Santa Fe Group sediments (Table 1). Because 6 m seems unreasonably high for a topographic scarp developed in Santa Fe Group sediments, this thickness of lava suggests that multiple flows might have erupted while the fault was active. The second explanation suggests that faulting with at least 6 m of vertical displacement must have occurred after the basalt was emplaced but became inactive prior to middle Pleistocene, the oldest age of the materials that cover it today (Bauer et al., 1997, 2000, 2016). From inspection of the geologic and geophysical relations observed for the model of profile A–A′, we prefer the second explanation. Thickening and increasing number of basalt layers north of TV-115 along A–A′ suggest that faulting was successively progressing southward toward the mountain front.

Aeromagnetic lineament L2 (Fig. 8) follows the banks of the Rio Pueblo de Taos. The linearity of the stream prompted previous workers to infer an underlying structural origin (Muehlberger, 1979; Machette and Personius, 1984). The extensive length and northeast trend of L2 are generally parallel to lineament L1; so we similarly infer that it is caused by oblique-slip faulting that has disrupted Servilleta Basalt. Multiple basalt flows are present in numerous shallow water wells on either side of the stream (Bauer et al., 2016). The wide variations in thickness of basalt layers and intervening sediments in these wells make correlations difficult across distances as short as 1 km. We interpret this stratigraphic variability as an indication that faulting, volcanism, erosion, and
sedimentation were all episodically active in this area during the age range of the basalts (5.5–3 Ma). The extreme variability might also be explained by basalt that flowed along paleochannels of Rio Pueblo de Taos, as hypothesized in the geophysical model for section A–A’ (Fig. 9).

Because the aeromagnetic anomalies that make up the zone of north-easterly lineaments arise from Servilleta Basalt, these inferred faults must have been active during and/or shortly after the period of time when Servilleta Basalt flows were erupted (5.5–3 Ma). Their strong alignment with the Jemez lineament (Fig. 1) suggests an underlying crustal control on their initiation. The absence of fault scarps offsetting late Quaternary sediments within this zone (Fig. 3) suggests that the faults are older than middle Pleistocene, the age of the oldest alluvium that covers the area (Bauer et al., 1997, 2000).

Implications for the Sangre de Cristo Fault System

Several relations from modeling profiles B–B’ and C–C’ have implications for the development of the Sangre de Cristo fault system and the eastern and northeastern subbasin margins. First, the models support a concept of a half-graben with Servilleta Basalt on the west and older basaltic rocks on the east that are structurally higher, as illustrated in Figure 10C. Important implications are that the master fault of the half-graben was earlier located 4–5 km west of the modern mountain front and then later shifted to the east. Moreover, Servilleta Basalt is absent on the footwall of the earlier half-graben, where it either was not deposited or was eroded soon after deposition.

Second, flows of Servilleta Basalt abruptly terminate against some basement faults but are not obviously offset by others, and where the flows terminate, the topmost flow(s) extend farther toward the range front than do the flows below (Figs. 11 and 12). These relations are supported by the profile magnetic modeling and aeromagnetic map interpretation. They were guided by the results from modeling of A–A’ across the Embudo fault zone (Fig. 9), where these relations are well constrained. In both cases, the relations observed suggest growth faulting and Servilleta volcanism were coeval.

Third, the models show that the older basaltic layers do not extend as far east as the range front and have little to no displacement overlying faulted basement to the west, where the older basalt flows are modeled underneath the Servilleta Basalt (between 16 and 18 km distance on Fig. 11 and 15–17 km distance on Fig. 12). These relations imply that, during or shortly after eruption of the older flows, normal faulting was concentrated where the older basaltic layers terminate. In addition, some normal faulting involving the basement to the west of this point occurred before the eruption of the older lava, possibly even prior to rift initiation.

Finally, the depiction of large fault displacements on the older basaltic layers beneath Servilleta Basalt has implications for the ages of faulting but modeling uncertainties make these implications speculative. Constraints from the K3 well in B–B’ and the Elston well in C–C’ suggest that the older basaltic rocks must be below the total depths of these wells, if present at all. Although we conclude that the older basaltic layers are thus displaced to much lower levels under these wells, it is also possible that the older lavas did not flow into this area originally or were removed by erosion before faulting.

The patterns of shallow faults inferred from geologic mapping and aeromagnetic data compared to the overall shape of the northeastern subbasin margin (Fig. 15) also have implications for fault history. The northeast margin must have had a prolonged, early period of margin formation along north-west-striking faults that later shifted to activity along more northerly faults near the present-day mountain front. This conclusion results from the northwest trend of the subbasin margin, which diverges from the northerly trends of faults near the range front, and is crossed obliquely by northerly trending faults in the interior of the basin. Moreover, as discussed earlier, we interpret the northwest-trending, aeromagnetic lineament L4 as a boundary between oldest, Pliocene Servilleta Basalt on the west and an extensive area of subhorizontal, deep, possibly middle Miocene basaltic layers on the east. The deeper basaltic layers overlie the northwest-trending part of the subbasin margin but show little displacement (e.g., Figs. 11 and 12), except at lineament L4. Therefore, we hypothesize that fault activity on the northeastern subbasin margin shifted in time, from NW-striking faults east of L4 before eruption of the oldest basaltic flows, to a focus at L4 during eruption of Servilleta Basalt, then eastward to a focus at north-striking aeromagnetic lineament L5, and finally even farther eastward to the modern range front.

Relations between Picuris-Pecos and Los Cordovas Fault Strands

Strands of the Picuris-Pecos fault system are cut by the younger Embudo fault zone but likely extend northward at the basin floor (Bauer and Kelson, 2004a). Using the patterns of mapped and aeromagnetically inferred faults (Figs. 13 and 14), we can speculate on how the strands extend northward and possibly influenced younger faulting. Within the wide zone of north-erly trending faults associated with Los Cordovas faults, individual faults appear to bend to the southeast near and across the northeasternly trending zone of faults that includes L1 and L2 (Fig. 14B). The bends suggest there may be local vertical-axis rotation of fault blocks adjacent to the Embudo fault zone, consistent with structural variability observed in the field (Kelson et al., 2004a). South of lineament L1, the northerly faults are not well defined aeromagnetically, because shallow Servilleta Basalt is generally absent. Nevertheless, one can envision connections with fault strands of the Picuris-Pecos fault system in the mountain flanks (Fig. 15) that include possible left-lateral displacements at the northeasterly zone. Thus, the patterns are compatible with previous hypotheses that the Los Cordovas faults are inherited from remnant strands of the Picuris-Pecos fault system north of the Embudo fault zone.
**23–18 Ma (early Miocene)**
Proto-Sangre de Cristo Fault

- Upper Picuris formation and possibly earliest Santa Fe Group deposition of volcaniclastic detritus from the north.
- Buried remnants of an early Eocene (?) lava flow, possibly from the northwest, that is now preserved in the Alcon well.

Reorganization of Picuris-Pecos fault system from post-Laramide graben formation to west-down normal-oblique slip, increasing in throw in the central part of the study area, indicated by larger hachures on normal faults. Intrusion of plutons in Latir volcanic field (north of study area).

**18–11 Ma (early and middle Miocene)**
Rapid Extension and Basin Formation

- Santa Fe Group deposition of volcaniclastic detritus from the north and lithoclastics from the emerging, low-relief Sangre de Cristo Mountains.
- Basaltic lava flow, possibly from a local source, that is now preserved in wells (e.g., BOR-5) in the eastern part of the study area.
- Period of rapid extension, basin formation, and west-down activity on normal faults at the eastern basin margin with Sangre de Cristo Mountains beginning to emerge. On the south, activity on the Picuris-Pecos fault system is waning (dashed lines). Oblique slip on the proto-Embudo fault develops by 11 Ma, with most activity southwest of the study area. Santa Fe Group includes widespread eolian deposits from ~13–11 Ma.

**11–6 Ma (middle and late Miocene)**
Embudo Fault Extends to Northeast

- Santa Fe Group deposition of volcaniclastic detritus from the north and lithoclastics from the emergent Sangre de Cristo and Picuris Mountains.
- Rapid extension continues, with lesser activity on faults shown with dashed lines. The Embudo fault propagates to the northeast, following the present-day trend of the Rio Pueblo de Taos (lineament L2). Left-oblique slip dominates the southwestern Embudo fault. Dip slip dominates on the northeast, focused on lineament L4. En echelon faults develop between the Embudo and the Sangre de Cristo fault systems, where multiple faults may be active and possibly linked. Older lava is covered by Santa Fe Group and displaced by normal faults.
6–3 Ma (late Miocene and Pliocene)
Eruptions of Taos Plateau Volcanic Field (TPVF)

- Alluvial fans of Lama formation deposited between and after multiple eruptions of Servilleta Basalt.
- Servilleta Basalt
- Intermediate-composition lava flows from TPVF, less widespread than Servilleta Basalt
- Servilleta Basalt
- Paleozoic sedimentary rocks and Proterozoic basement composing the mountains, with Picuris formation preserved in grabens.

Multiple eruptions of Servilleta Basalt while normal faults at the eastern basin margin begin to focus along lineament L5. Parallel, NE-striking, oblique faults likely flank the southeastern basin margin (e.g., L1 and L2), causing gentle southeastward tilt of basalt layers. Hard links between oblique and normal fault strands are possible (queried). Basalt may have ponded against L4 (shown) or was later faulted at L4 and the footwall eroded.

3–0.8 Ma (Pliocene and early Pleistocene)
Los Cordovas Faults

- Pliocene alluvial fans of Lama formation and early Pleistocene alluvium and piedmont deposits
- Servilleta Basalt
- Paleozoic sedimentary rocks and Proterozoic basement composing the mountains, with Picuris formation preserved in grabens.

End of Taos Plateau volcanism. Continued deposition of Lama formation in the Pliocene, with large fans advancing across the northeastern part of the study area. Embudo and Sangre de Cristo faults are linked, following lineament L1 and the present-day range front on the north. Servilleta Basalt is eroded from footwalls of normal faults at southeastern margin. N-S Los Cordovas faults develop in a wide zone west of L3, perhaps following old Picuris-Pecos fault strands within the basement.

<0.8 Ma (middle Pleistocene to Present)
Modern Linked Embudo-Sangre de Cristo Faults

- Surficial deposits associated with modern drainages and slopes. Eolian deposits on the Taos Plateau
- Servilleta Basalt
- Paleozoic sedimentary rocks and Proterozoic basement composing the high-relief mountains, with Picuris formation preserved in grabens.

Los Cordovas and strands of Embudo and Sangre de Cristo faults within the basin interior become inactive. The linked Embudo and Sangre de Cristo faults have shifted away from the basin interior to the present location. The Sangre de Cristo and Picuris Mountains have high relief. Rio Grande and Rio Pueblo de Taos have cut deep gorges in Servilleta Basalt layers. Surficial deposits cover most of the study area.
A PROPOSED RIFT HISTORY FOR THE TAOS AREA

Implications from the integrated geophysical interpretations and models allow us to revise previous scenarios of the Cenozoic structural development of the Taos area (Bauer and Kelson, 2004a; Smith, 2004; Smith et al., 2004). Although some aspects of the history are well constrained, the complete picture through time is not. Nevertheless, we have developed a set of six paleogeographic maps that depict the evolution of the Rio Grande rift within the study area (Fig. 16). These maps draw heavily on the previous work of many researchers.

The first paleogeographic map (Fig. 16A) depicts early stages of rifting in early Miocene time (23–18 Ma). Volcanism related to the San Juan volcanic field had ended. A lava flow related to the early volcanism, which is now preserved in the Alcon well (Fig. 9), had been buried by deposits of the lower part of the Picuris Formation. During the 23–18 Ma interval, grabens along high-angle, north-striking faults formed as the Picuris-Pecos fault system was transitioning from Laramide-style to rift-style kinematics. Abundant detritus from the Latir volcanic field was being transported southward into the study area and deposited as the upper part of the Picuris Formation, coeval with deposition of earliest Santa Fe Group elsewhere.

By early and middle Miocene (18–11 Ma), graben formation had transitioned to intense rift-basin formation and accumulation of the Santa Fe Group (Fig. 16B). Uplift of the Sangre de Cristo Mountains accompanied basin subsidence, providing additional detritus of Proterozoic and Paleozoic rocks to the mix of sediments shedding from the north. Eruption of basaltic rocks that are now deeply buried on the east side of the study area would have occurred sometime during the latter part of this time period. Possible sources are mafic lavas that erupted from the Jemez volcanic field to the southwest (Fig. 16B). Oblique slip began to develop on a proto-Embudo fault aligned with the regional Jemez lineament, with activity mostly to the southwest of the study area.

From middle to late Miocene (11–6 Ma), the principal slip on the Embudo fault was along the area of the modern Rio Pueblo de Taos (lineament L2), with mostly normal slip at the southern subbasin margin within the study area (Fig. 16C). The dominance of normal slip at this time explains the large amount of vertical displacement observed across the Embudo fault zone today. Based on the strong northwest trend in the 3D model and displacements on the older basalts inferred from the cross-section models, we infer that most of the activity along the northeastern rift margin was concentrated on the north-west-striking fault (lineament L4) that is basinward of the present-day mountain front. Perhaps the northwest strike developed as the northward-propagating faults from the study area started to interact with the southward-propagating faults from north of the study area, following concepts described by Rosendahl (1987). Likewise, the proto-Embudo and proto-Sangre de Cristo fault systems were interacting with each other, and they may have formed a link at this time. The Picuris Mountains continued to emerge, with detritus shed to the north and northwest.

During latest Miocene and early Pliocene time (6–3 Ma), Servilleta Basalt was periodically eroding from multiple vents and spreading across the area (Fig. 16D). Between eruptive cycles, the clastic sediments of the Lama formation were being shed from the nearby mountains. Intermediate-composition volcanoes were erupting northwest and west of the study area, including Tres Orejas at the western edge of the study area. Deeper layers of possible Servilleta Basalt as old as 5.5 Ma (or an earlier, unrecognized eruptive sequence) may have flowed as far north as the north-central part of the study area, which may lie at depth in well BOR-6 today. These lavas flowed into the rapidly subsiding subbasin, where they ponded against fault scarps or lapped over them and were subsequently faulted and eroded from the footwalls shortly after deposition. The principal segment of the Embudo fault continued to occupy the area of the present-day Rio Pueblo de Taos (lineament L2), although other segments formed to the south, which were also aligned with the regional Jemez lineament. Slip was probably left-oblique, with increasing dip slip along the north-central strands of the fault. The northwest-striking fault at the northeastern subbasin margin (lineament L4) was still active, because the Servilleta Basalt was unable to cross it or was being eroded from its footwall.

During late Pliocene to middle Pleistocene time (3–0.8 Ma), eruption of Servilleta Basalt had ceased within the study area and the linked Embudo–Sangre de Cristo fault had shifted closer to the present-day mountain front, following aeromagnetic lineament L1 (Fig. 16E). Slip on the fault was oblique, but with enough dip slip to allow for erosion of the basalt off the footwall. It is likely that other normal and oblique-slip faults were active between the main fault and the mountain front during this time. Southeast tilting of the basalt layers in concert with growth faulting developed toward the end of this time. The Los Cordovas faults also became active, possibly controlled by buried strands of the Picuris-Pecos fault system.

From middle Pleistocene to present (<0.8 Ma), the Los Cordovas faults and the northeast-trending faults within the interior of the basin became inactive and were partially buried beneath middle Pleistocene and younger fan material (Fig. 16F). Activity became concentrated at the present-day, high-relief mountain front, along the still linked Embudo–Sangre de Cristo fault, which developed greater concave curvature surrounding the subbasin. The Rio Grande and Rio Pueblo de Taos became incised into layers of Servilleta Basalt, allowing throughgoing drainage from the San Luis Basin to the south.

SUMMARY

We present a detailed example of how a subbasin develops adjacent to a transfer zone in the Rio Grande rift. The Embudo transfer zone in the Rio Grande rift is considered one of the classic examples and has been used as the inspiration for several theoretical models. Despite this attention, the
history of its development into a major rift structure is poorly known along its northern extent, near Taos, New Mexico. Geologic evidence for all but its young rift history is buried under Quaternary cover. We focus on understanding the pre-Quaternary evidence that is in the subsurface by integrating diverse pieces of geologic and geophysical information. As a result, we present a substantively new understanding of the tectonic configuration and evolution of the northern extent of the Embudo fault and its adjacent subbasin.

The ~64-km-long, Embudo fault zone forms the link between the east-down western border of the northern Española Basin and the west-down eastern border of the southern San Luis Basin. At the northern end of the fault zone, the rift margins follow an eastward curving reentrant in the mountain front known as the Taos embayment, which partially surrounds the Town of Taos, New Mexico, on the south and east. The NE- to E-striking, left-oblique Embudo fault zone transitions to the N-striking, west-down, normal Sangre de Cristo fault along the edges of the embayment, forming the structural margins of the Taos subbasin.

To better understand the subsurface geology, we synthesize aeromagnetic and gravity data, borehole and physical-property information, and geologic mapping. We rely heavily on a 3D basin model derived from gravity data, patterns on the aeromagnetic map, and 2D profile modeling, to develop the following new interpretations about subsurface geology:

1. The subbasin within the Taos embayment has an imperfect D shape in map view, with the deepest point (2 km depth) at 36° 26′ N latitude and 105° 37′ W longitude, northwest of Taos. The basin floor slopes gently east and southeast toward this point.
2. Multiple Servilleta Basalt flows and intervening sediments, generally 200 m thick, lie at depths of 0–100 m over a limited area within the interior of the subbasin.
3. The Embudo fault zone, along the southern subbasin margin, extends wider in the subsurface than its mapped width at the surface, encompassing a zone as much as 9 km wide that extends northward from the range front to the Rio Pueblo de Taos. Aeromagnetic lineaments indicate that flows of Servilleta Basalt, concealed by middle Pleistocene (?) alluvium, are disrupted or truncated by faults in this zone.
4. The basin floor gradually steps down across the wide Embudo fault zone, with a vertical displacement of ~1.8 km and a total structural relief of ~2.6 km. Average vertical slip rates of 72–96 m/m.y. are required to accommodate these estimates of vertical displacement since rift initiation at ca. 25 Ma. These rates are much higher than the average vertical slip of 35 m/m.y. previously estimated for the past 3 Ma.
5. Along the eastern subbasin margin, a 5–7-km-wide zone of normal, west-down, stepped faults extends basinward from the north-striking Sangre de Cristo range front fault. We hypothesize that Servilleta Basalt is absent within this zone and that basaltic rocks found deep in wells within the zone are older, perhaps middle Miocene in age. This hypothesis resolves apparent conflicts between previous structural models resulting from well correlations and constraints from gravity data.
6. Overall patterns of inferred and mapped shallow faults show general correspondence compared to the overall shape of the deep subbasin margins, as expressed by the 3D gravity model. Northeasterly trends align with the Jemez lineament, a northeasterly regional crustal boundary. Exceptions to the correspondence are north of Taos, where north-striking faults related to the Honda section of the Sangre de Cristo fault diverge from the northwest-trending, northeastern subbasin margin.

7. North-striking, early to middle Pleistocene Los Cordovas faults in the interior of the basin extend -2–4 km farther east than mapped from surface exposures. They also obliquely cross the northwest trend of the northeastern subbasin margin.

Using these interpretations, we infer relations between faulting and flows of Pliocene Servilleta Basalt and older, buried basaltic rocks that reveal previously unrecognized aspects of the history of faulting and subbasin formation. Combined with evidence from geologic mapping, we improve the understanding of the Cenozoic evolution of the Taos subbasin and its faulted margins. The history involves shifts in the locus of fault activity at the rift margins as the Taos subbasin developed, as illustrated by a set of paleogeographic maps for the study area.

During 23–18 Ma (early Miocene), we speculate that early rift extension was concentrated on strands of the Picuris-Pecos fault system, a northerly trending crustal boundary that today intersects the Embudo and Sangre de Cristo faults where they join. We suggest that some of the graben-forming faults that were active within this fault system at the end of Laramide time formed the first west-down master faults of the future subbasin (proto–Sangre de Cristo fault system).

During 18–11 Ma (early and middle Miocene), rapid extension and basin formation increased activity at the eastern subbasin margin, while fault strands of the Picuris-Pecos fault system to the south in the Picuris Mountains became less active. Lava flows related to the basaltic rocks presently buried in wells in the eastern part of the study area might have erupted along the eastern subbasin margin at this time.

During 11–6 Ma (middle and late Miocene), proto-Embudo fault strands were likely aligned with the Jemez lineament and the modern, NE-aligned Rio Pueblo de Taos, as much as 7 km basinward of the modern Embudo fault zone. Left-oblique slip had developed at the southwest end of the fault zone, out of the study area, but likely transitioned to mostly normal slip to the northeast, within the study area. Northwest-striking normal faults formed the margin of the northeastern part of the subbasin, an orientation divergent from the modern north-striking Sangre de Cristo faults. The faults at all the subbasin margins were likely interacting with each other and may have started to link.
During 6–3 Ma (late Miocene and Pliocene), volcanism of the Taos Plateau volcanic field was well under way. Proto–Embudo fault strands at the southern subbasin margin remained active during eruption of Servilleta Basalt along and parallel to the modern, NE-aligned Rio Pueblo de Taos. Servilleta Basalt likely did not extend across the northeastern subbasin margin, where north-west-striking faults were still active 1–4 km west of the modern, north-striking range front.

During 3–0.8 Ma (Pliocene and early Pleistocene), volcanism of the Taos Plateau was ending, and north-striking Los Cordovas normal faults became active in the interior of the subbasin. Their locations may have been controlled by relict Picuris-Pecos fault strands located on the basin floor. Faulting at the northeastern subbasin margin changed orientation and location to more northerly striking faults within 2 km of the current range front. The Sangre de Cristo fault and Embudo faults were likely linked at this time.

After middle Pleistocene(?) time (<0.8 Ma), the Los Cordovas faults and strands of the extended Embudo fault zone that were along and parallel to the current Rio Pueblo de Taos became inactive. The locus of activity along the linked Embudo–Sangre de Cristo fault system shifted southward and eastward to the present-day mountain front. The modern landscape of high-relief mountains surrounding the Taos embayment was well established, and the Rio Grande and Rio Pueblo de Taos incised deep gorges in the basin floor.

APPENDIX

TABLE A1. DENSITIES BY GEOLOGIC UNIT

<table>
<thead>
<tr>
<th>Geologic unit*</th>
<th>Data type†</th>
<th>Range of bulk densities (kg/m³)</th>
<th>Estimated mean density§ (kg/m³)</th>
<th>Rationale and data source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pliocene volcanic rocks</td>
<td>Geophysical estimate</td>
<td>2450</td>
<td>Assume similar density as that estimated for the bulk of volcanic rocks and interbedded sediments composing Guadalupe Mountain, east-central Taos Plateau (Grauch et al., 2015).</td>
<td></td>
</tr>
<tr>
<td>Servilleta Basalt (Tbs) plus Lama formation (QTL)</td>
<td>Geophysical estimate</td>
<td>2450</td>
<td>Assume similar density as that estimated for the bulk of volcanic rocks and interbedded sediments composing Guadalupe Mountain, east-central Taos Plateau (Grauch et al., 2015).</td>
<td></td>
</tr>
<tr>
<td>Intermediate volcanic rocks (Tv)</td>
<td>Geophysical estimate</td>
<td>2450</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sedimentary basin fill</td>
<td>From Albuquerque Basin</td>
<td>2100–2250</td>
<td>2170</td>
<td>Following the density-depth function developed from borehole density logs for Albuquerque Basin (Grauch and Connell, 2013).</td>
</tr>
<tr>
<td>Santa Fe Group above 1.25 km depth (QTL, Tsf)</td>
<td>From Albuquerque Basin</td>
<td>2250–2400</td>
<td>2300</td>
<td>Following the density-depth function developed from borehole density logs for Albuquerque Basin (Grauch and Connell, 2013).</td>
</tr>
<tr>
<td>Picuris Formation (Tp)</td>
<td>Sonic log</td>
<td>2200–2300</td>
<td>2250</td>
<td>Estimates from a sonic log of one well at depth range of 100–366 m (Bauer, 2016).</td>
</tr>
<tr>
<td>Rocks from San Juan volcanic field (SJVF)</td>
<td>Weighted estimate</td>
<td>2200–2500</td>
<td>2450</td>
<td>Weighted estimate based on ignimbrites (density of 2200) and andesites (density of 2500) composing 1/6 and 5/6 of the volume of the field, respectively, from Drenth et al. (2012).</td>
</tr>
<tr>
<td>Igneibrites and andesites</td>
<td>From</td>
<td>2460–2660</td>
<td>2540</td>
<td>Average density from measurements of saturated samples from Taos area (Grauch and Drenth, 2016).</td>
</tr>
<tr>
<td>Paleozoic rocks (Pzu)</td>
<td>Samples</td>
<td>2670–2680</td>
<td>2670</td>
<td>Average density from measurements of saturated samples from Taos area (Grauch and Drenth, 2016).</td>
</tr>
<tr>
<td>Shale and sandstone</td>
<td>Samples</td>
<td>2550–2680</td>
<td>2600</td>
<td>Bulk density measurements from Tusas and Picuris Mountains (Grauch and Drenth, 2016).</td>
</tr>
<tr>
<td>Limestone</td>
<td>Samples</td>
<td>2590–2860</td>
<td>2880</td>
<td>Bulk density measurements from Tusas and Picuris Mountains (Grauch and Drenth, 2016).</td>
</tr>
<tr>
<td>Proterozoic rocks (Xu)</td>
<td>Samples</td>
<td>2590–3060</td>
<td>2700</td>
<td>Bulk density measurements of Hondo Group from Tusas and Picuris Mountains (Grauch and Drenth, 2016).</td>
</tr>
<tr>
<td>Granite and granitic gneiss</td>
<td>Samples</td>
<td>2770–3050</td>
<td>2910</td>
<td>Bulk density measurements from Tusas Mountains (Grauch and Drenth, 2016).</td>
</tr>
<tr>
<td>Metasedimentary rocks</td>
<td>Samples</td>
<td>2590–3060</td>
<td>2700</td>
<td>Bulk density measurements of Hondo Group from Tusas and Picuris Mountains (Grauch and Drenth, 2016).</td>
</tr>
<tr>
<td>Metasedimentary and metavolcanic sequence</td>
<td>Samples</td>
<td>2770–3050</td>
<td>2910</td>
<td>Bulk density measurements from Tusas Mountains (Grauch and Drenth, 2016).</td>
</tr>
</tbody>
</table>

*Geologic unit codes explained on Figure 3.  
†From Albuquerque Basin—Compilation of various density logs from deep boreholes in the Albuquerque Basin, described in Grauch and Connell (2013). Geophysical estimate—Estimated by finding the Bouguer reduction density that gives least correlation to terrain, described as Nettleton’s method (Telford et al., 1990). Samples—Measurements taken on hand samples collected in the field from one or more outcrops. Sonic log—Density range and median value estimated using empirical relations for sedimentary basins (Gardner et al., 1974) and inspection of borehole sonic log. Weighted estimate—Representative density of whole volcanic field estimated based on borehole density estimates and weighted by volume within the volcanic field, from Drenth et al. (2012).  
§Density estimated from median values of sample measurements (Grauch and Drenth, 2016) or as noted.
TABLE A2. MAGNETIC PROPERTIES BY GEOLOGIC UNIT

<table>
<thead>
<tr>
<th>Geologic unit*</th>
<th>Typical magnetic susceptibility $\times 10^3$ (SI)†</th>
<th>Induced component‡ (A/m)</th>
<th>Estimated total magnetization§ (A/m)</th>
<th>Polarity indicated from anomalies</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Minimum</td>
<td>Maximum</td>
<td>Minimum</td>
<td>Maximum</td>
</tr>
<tr>
<td>Surficial deposits (Qa)</td>
<td>0.41</td>
<td>0.67</td>
<td>0.02</td>
<td>0.03</td>
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<tr>
<td>Servilleta Basalt (Tbs)</td>
<td>2.00</td>
<td>5.32</td>
<td>0.08</td>
<td>0.22</td>
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<tr>
<td>Intermediate-composition volcanic rocks (mostly out of study area)</td>
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<td></td>
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<tr>
<td>Cerrito de los Taoses andesite</td>
<td>12.1</td>
<td>14.1</td>
<td>0.50</td>
<td>0.58</td>
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<tr>
<td>Cerro Montoso andesite</td>
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<td></td>
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<tr>
<td>Cerro Negro dacite</td>
<td>15.0</td>
<td>15.0</td>
<td>0.62</td>
<td>0.62</td>
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<tr>
<td>Tres Orejas dacite</td>
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<td></td>
<td></td>
<td></td>
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<tr>
<td>Sedimentary basin fill</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Santa Fe Group, Lama formation (QTl)</td>
<td>1.50</td>
<td>3.50</td>
<td>0.06</td>
<td>0.14</td>
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<tr>
<td>Santa Fe Group (Tsf)</td>
<td>1.20</td>
<td>3.69</td>
<td>0.05</td>
<td>0.15</td>
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<tr>
<td>Picuris Formation (Tp)</td>
<td>3.26</td>
<td>4.58</td>
<td>0.13</td>
<td>0.19</td>
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<tr>
<td>Middle Tertiary volcanic rocks (out of study area)</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Basalt of Hinsdale Formation (ca. 21 Ma), Tusas Mountains</td>
<td>4.39</td>
<td>22.0</td>
<td>0.18</td>
<td>0.91</td>
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<tr>
<td>Basalt of Hinsdale Formation (ca. 26 Ma), San Luis Hills</td>
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<td></td>
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<tr>
<td>Andesite of upper Conejos Formation (ca. 30 Ma), San Luis Hills</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Paleozoic rocks (Pzu)</td>
<td>0.10</td>
<td>0.15</td>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>Proterozoic rocks (Xu)</td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Granite and granitic gneiss</td>
<td>0.28</td>
<td>1.84</td>
<td>0.01</td>
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<tr>
<td>Metasedimentary rocks (Hondo Group)</td>
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<td>1.00</td>
<td>0.00</td>
<td>0.04</td>
</tr>
<tr>
<td>Metasedimentary and metavolcanic sequence (Vadito Group)</td>
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<td>0.50</td>
<td>0.00</td>
<td>0.02</td>
</tr>
<tr>
<td>Metavolcanic rocks from Tusas Mountains</td>
<td>1.00</td>
<td>6.42</td>
<td>0.04</td>
<td>0.26</td>
</tr>
</tbody>
</table>

*Geologic unit codes explained on Figure 3. Ages and locations of geologic units from Tusas Mountains and San Luis Hills are from Drenth et al. (2011) and Thompson et al. (2015), respectively.
†Magnetic susceptibility values are from field and sample measurements by Grauch and Drenth (2016). Minimum and maximum median values from multiple sites are reported for all geologic units except for Cerrito de los Taoses andesite (two samples from one site) and Cerro Negro dacite (one sample), where median values could not be computed.
‡Induced component computed by multiplying the typical minimum and maximum values of magnetic susceptibility by Earth’s magnetic field of 51,700 nanoteslas, following Hansen et al. (2005) to convert units to amperes per meter (A/m).
§Induced component computed by multiplying the typical minimum and maximum values of magnetic susceptibility by Earth’s magnetic field of 51,700 nanoteslas, following Hansen et al. (2005) to convert units to amperes per meter (A/m).
# Total magnetizations estimated as follows. Values in square brackets assume negligible remanent magnetization based on rock type, resulting in total magnetizations that are essentially the same as the induced components. Minimum and maximum values of total magnetization for Servilleta Basalt were estimated from the range of natural remanent magnetizations (NRM) of 2.00 to 4.15 A/m measured by Brown et al. (1993) minus or plus the maximum induced component, respectively. Values for the ca. 21 Ma basalt of the Hinsdale Formation were estimated from geophysical modeling (Drenth et al., 2011). All other values for intermediate-composition volcanic rocks and Middle Tertiary volcanic rocks were estimated by qualitative assessment of observed aeromagnetic anomalies compared to magnetic terrain models of hills composed of the unit listed. A maximum 25% contribution of NRM to the total magnetization is assumed for Santa Fe Group units, following findings from Albuquerque Basin (Hudson et al., 2008).

ACKNOWLEDGMENTS

We are delighted to acknowledge the assistance of National Aeronautics and Space Administration (NASA) astronaut Stan Love in the collection of magnetic-susceptibility, gravity, and ground-based magnetic data in 2011. We are grateful to Glosita Geoscience, Inc. (GGI) for generously sharing their well data and interpretations with us. Mike Powers, U.S. Geological Survey (USGS), guided us in understanding the shallow seismic-reflection sections in the area. Discussions with Mike Cosca, Ren Thompson, and Kenzie Turner of the USGS were helpful for understanding the volcanic geology. Dan Koning and Scott Aby of the New Mexico Bureau of Geology and Mineral Resources (NMBGMR) provided valuable advice on stratigraphic relations in several key wells, as well as enlightening discussion. We appreciate thoughtful consultations and review of an early draft by Mark Hudson, USGS, and journal reviews by an anonymous reviewer and Patricia Dickerson (University of Texas at Austin).

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