STRATIGRAPHY AND TECTONIC DEVELOPMENT OF THE ALBUQUERQUE BASIN, CENTRAL RIO GRANDE RIFT

FIELD-TRIP GUIDEBOOK FOR THE GEOLOGICAL SOCIETY OF AMERICA
ROCKY MOUNTAIN-SOUTH CENTRAL SECTION MEETING, ALBUQUERQUE, NM
PRE-MEETINNG FIELD TRIP

MINI-PAPERS

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NEW MEXICO GEOLOGICAL SOCIETY GUIDEBOOK 50

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REVISIONS TO GUIDEBOOK AND MINI-PAPERS

This field-guide accompanied a pre-meeting field trip of the Geological Society of America Rocky Mountain and South-Central Section conference in Albuquerque, New Mexico. A limited quantity of guidebooks and mini-paper compilations were produced for participants of this field trip. A number of typographical, grammatical, and editorial errors were found in this first version of the guidebook, mainly because of logistical constraints during preparation for the field trip. In the revised version, *released on June 11, 2001*, many errors have been corrected. Many photographs, figures, and maps, shown during the field trip but not included in the first version, are included in this revision. Numerous minor editorial changes and corrections have also been made to the guidebook mini-papers.

The field-guide has been separated into two parts. Part A (open-file report 454A) contains the three-days of road logs and stop descriptions. Part B (open-file report 454B) contains a collection of mini-papers relevant to field-trip stops.

The contents of the road logs and mini-papers have been placed on open file in order to make them available to the public as soon as possible. Revision of these papers is likely because of the on-going nature of work in the region. The papers have not been edited or reviewed according to New Mexico Bureau of Mines and Mineral Resources standards. The contents of this report should not be considered final and complete until published by the New Mexico Bureau of Mines and Mineral Resources. Comments on papers in this open-file report are welcome and should be made to authors. The views and preliminary conclusions contained in this report are those of the authors and should not be interpreted as necessarily representing the official policies, either expressed or implied, of the State of New Mexico or the U.S. Government.

ACKNOWLEDGEMENTS

This field trip was supported by the New Mexico Bureau of Mines and Mineral Resources (P.A. Scholle, Director) and the New Mexico Museum of Natural History and Science. Much of the data presented during this field trip are from numerous open-file reports released by the New Mexico Bureau of Mines and Mineral Resources during the course of cooperative geologic mapping with the U.S. Geological Survey (New Mexico Statemap Project, P.W. Bauer, Program Manager). We are particularly grateful to the Pueblos of Zia, Isleta, Sandia, San Felipe, Santo Domingo, Jemez, and Santa Ana for granting access during many of the stratigraphic and mapping studies discussed during the field trip. In particular, we thank Mr. Peter Pino for enabling access to study the stratigraphically significant localities along the Rincones de Zia. We also thank Mr. Gary Nolan, Mr. Jerry Burke, and Mr. Mackie McClure for allowing access to the LaFarge and SunCountry Redimix gravel quarries near Bernalillo, New Mexico. We also thank Ms. Leanne Duree of the Ball Ranch for allowing access through their lands on Tanos Arroyo.

We thank the New Mexico Geological Society for granting permission to reprint three papers from their 1999 Guidebook 50 entitled *Albuquerque Geology* (F.J. Pazzaglia and S.G. Lucas, eds). We especially thank V.J.S. Grauch for agreeing to present summaries of recent regional geophysical surveys of the Albuquerque Basin.

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STRATIGRAPHY OF THE ALBUQUERQUE BASIN, RIO GRANDE RIFT, CENTRAL NEW MEXICO: A PROGRESS REPORT

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INTRODUCTION

The Albuquerque Basin of central New Mexico is one of the largest sedimentary basins of the Rio Grande rift, a chain of linked, predominantly asymmetric or half-graben extensional basins that extend south from central Colorado, through central New Mexico, and into western Texas and northern Mexico (Hawley, 1978; Chapin and Cather, 1994). The Albuquerque Basin is about 60 km long, and about 55 km wide and strongly faulted on nearly all sides (Fig. 1). The Albuquerque Basin also represents a transitional tectonic feature, lying between the west-tilted Española and Socorro half-graben basins. The Albuquerque Basin sits between the topographically and structurally well expressed northern Rio Grande rift of northern New Mexico and southern Colorado, and the broader Basin and Range to the south. Basins of the northern Rio Grande rift tend to step eastward (Kelley, 1982), whereas basins to the south form alternating block-faulted basins and uplifts that characterize the Basin and Range.

The Albuquerque Basin comprises a single physiographic (Fig. 2) and tectonic feature (Woodward et al., 1978) that is segmented into a number of structural sub-basins and embayments (Grauch et al., 1999). Isostatic gravity data and oil-test data (Fig. 2) indicates that the basin is segmented into three major sub-basins (Cordell, 1978, 1979; Birch, 1982; Heywood, 1992; Grauch et al., 1999; Russell and Snelson, 1994; May and Russell, 1994; Lozinsky, 1994): the northern Santo Domingo, central Calabacillas, and southern Belen sub-basins. Sub-basin boundaries are somewhat diffuse and not universally accepted (Kelley, 1977; Lozinsky, 1994; Hawley, 1996; Grauch et al., 1999). Sub-basins also contain somewhat different depositional packages of the earlier rift-basin fill, whose lateral extent may be influenced by sub-basin boundaries (Fig. 3; Cole et al., 1999). Gravity data also shows a northwest structural grain within the basin along sub-basin boundaries (Fig. 2; Grauch et al., 1999). This northwest trend is not readily apparent from surficial geologic mapping and differs from the predominantly north-trending structural grain of the basin (Fig. 4), suggesting that sub-basin boundaries are obscured by younger and less deformed basin fill. The Belen sub-basin comprises the southern half of the Albuquerque Basin, is complexly faulted, and has a westward stratal tilt. The dominantly east-tilted Calabacillas and Santo Domingo sub-basins comprise the central and northern sub-basin, respectively (Fig 5; Grauch et al., 1999). Deep oil-well data indicate that the Calabacillas sub-basin and northern part of the Belen sub-basin contain as much as 4-5 km of synrift basin fill (Lozinsky, 1994). The Santo Domingo sub-basin is a graben with a complicated subsidence history that represents a zone of accommodation between the Albuquerque and Española basins (Smith et al., 2001). The Hagan embayment is a northeast-dipping structural re-entrant between the San Francisco and La Bajada faults that contains the oldest exposed Santa Fe Group strata in the basin.

The boundaries among the major sub-basins are complicated, however, regional gravity and oil-test data can constrain their locations. The southern portion of the Belen sub-basin narrows to about 9-12 km in width near the confluence of the Rio Salado and Rio Grande. The boundary between the Belen and Calabacillas sub-basins are defined by a diffuse zone of accommodation where the direction of stratal tilts change across the Tijeras accommodation zone of Russell and Snelson (1994). Gravity data suggests that the northwest-trending Mountainview prong (Hawley, 1996; Grauch et al., 1999) probably defines the boundary between the Belen and Calabacillas sub-basins. The boundary between the Calabacillas and Santo Domingo sub-basins is quite diffuse and recognized primarily on the basis of a broad north- and northwest-trending gravity high marked by the Ziana structure (Kelley, 1977; Personius et al., 1999; Grauch et al., 1999) and Alameda structural (monoclinal) zone. Other possible boundaries between these two sub-basins is the northeast-trending Loma Colorado zone (Hawley, 1996), which is marked by a northeast-trending alignment of fault-terminations, where faults of a specific polarity of movement (i.e., east-dipping) step over into faults having the opposite sense of dip (and presumably displacement). The Loma Colorado structural feature, however, is not well expressed in the gravity data and appears to die out to the northeast. Another possible boundary between the Calabacillas and Santo Domingo sub-basins has also been proposed at the San Felipe graben (Lozinsky, 1994), between Santa Ana Mesa and the Ziana structure; however, this graben is not well expressed in the gravity data and is probably a minor feature within the Santo Domingo sub-basin.

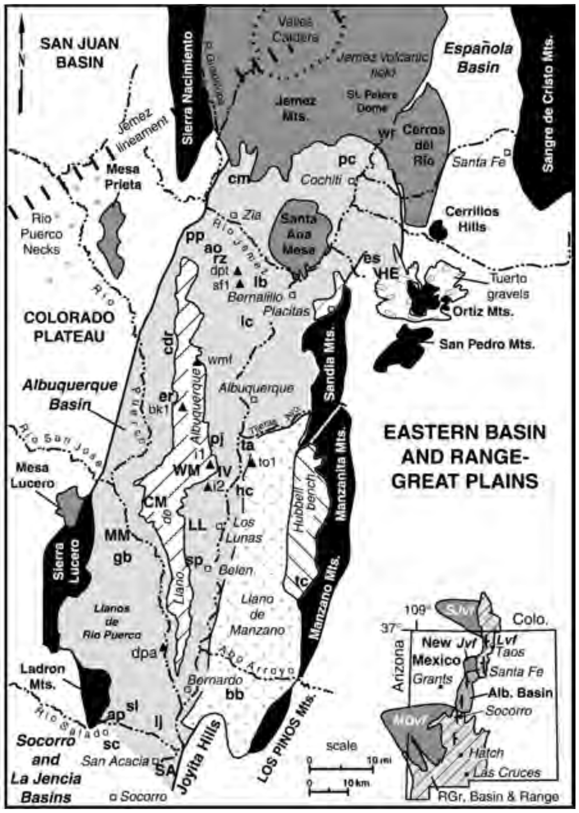

Figure 1. Albuquerque Basin and surrounding areas. Rift-flanking uplifts shown in black. Localities include: Rincones de Zia (rz), Ceja del Rio Puerco (cdr), Loma Barbon (lb), Arroyo Ojito (ao), Arroyo Piedra Parada (pp), Arroyo Popotosa (ap), Silver Creek (sc), Trigo Canyon (tc), Espinaso Ridge (es), White Rock Canyon (wr), El Rincon (er), Peralta Canyon (pc), Sierra Ladrones (sl), La Joya (lj), Chamisa Mesa (cm), Tijeras Arroyo (ta), Gabaldon badlands (gb), and Hell Canyon (hc). Volcanic features include the diabase of Mohinas Mountain (MM), trachyandesite at San Acacia (SA), Cat Mesa (CM), Wind Mesa (WM), Isleta volcano (IV), basalt at Black Butte (BB), and Los Lunas Volcano (LL). Oil-test wells (indicated by black triangles) include: Shell Santa Fe Pacific #1 (sf1), Shell Isleta #1 (i1), Davis Petroleum Tamara #1-Y (dpt), Shell Isleta #2 (i2), Burlington Resources Kachina #1 (bk1), TransOcean Isleta #1 (to1), and Davis Petroleum, Angel Eyes (dpa). Major Paleogene volcanic fields in New Mexico and southern Colorado include: Mogollon-Datil volcanic field (MDvf), San Juan volcanic field (SJvf), Jemez volcanic field (Jvf), and Latir volcanic field (Lvf).

The Albuquerque Basin was interpreted to have undergone about 17% extension in the Calabacillas and northern Belen sub-basins, near Albuquerque, and about 28% in the Belen sub-basin, near Bernardo, New Mexico (Russell and Snelson, 1994). The extension estimate for the northern part of the basin is based on the presence of the Rio Grande fault, a relatively young intrabasinal fault proposed by Russell and Snelson (1994). Their Rio Grande fault cuts the basin-bounding rift-flanking faults of the Sandia Mountains. Gravity (Grauch et al., 1999), geomorphic, and stratigraphic data (Connell and Wells, 1999; Connell et al., 1998a; Maldonado et al., 1999) questions the existence of this fault, which is buried by Quaternary alluvium. If the Rio Grande fault is not present beneath Albuquerque, then Russell and Snelson's (1994) extension estimate would also be suspect. The lack of strong structural and topographic expression of the sub-basin boundaries indicated on Figure 2 suggests a complicated history of basin development that differs from the present configuration of faults. The northwest-trending structures are obscured by younger basin fill and may represent older structural boundaries; however, some of these structures deform Plio-Pleistocene sediments.

Basin subsidence is controlled by numerous north-trending normal faults and relatively short, northeast-trending connecting faults that commonly form faulted relay ramps or transfer zones. Structural margins are typically defined by tilted footwall uplands, and basement-cored, rift-margin uplifts, such as the Sandia, Manzanita, Manzano, Los Pinos, and Ladron Mountains. These rift-bounding ranges are locally overlain by Mississippian, Pennsylvanian and Permian strata (Fig. 4) that provide a source of locally derived detritus for piedmont deposits. Other basin margins form escarpments, such as along the La Bajada fault and eastern edge of the Sierra Lucero, which form footwall uplands of moderate relief and are underlain by Pennsylvanian-Paleogene rocks. The northwestern margin is topographically subdued and defined by faults such as the Moquino fault in the Rio Puerco valley (Kelley, 1977; Tedford and Barghoorn, 1999). The eastern structural margin, near Albuquerque, New Mexico, is defined by roughly north-trending faults 1-3 km of basinward normal slip (Cordell, 1979; Russell and Snelson, 1994).

Inception of the Rio Grande rift began during late Oligocene time (Chapin and Cather, 1994; Smith, 2000; Kautz et al., 1981; Bachman and Mehnert, 1978; Galusha, 1966) as broad fault-bounded, internally drained basins began to receive sediment (Chapin and Cather, 1994). Stratal accumulation rates, calculated from scattered and sparsely dated sections indicate late Oligocene-middle Miocene stratal accumulation rates (not adjusted for compaction) of about 72-83 m/m.y. (Tedford and Barghoorn, 1999; Connell and Cather, *this volume*) for sediments near the basin margins. During late Miocene times, Lozinsky (1994) estimated an accumulation rate of about 600 m/m.y., which is considerably greater that earlier rates. During Pliocene time, the basins filled and became linked to adjoining basins with the onset of through-flowing drainages of the ancestral Rio Grande fluvial system. Stratal accumulation rates have only been estimated in a few places and suggest a much slower rate of accumulation, perhaps less than about 100 m/m.y.

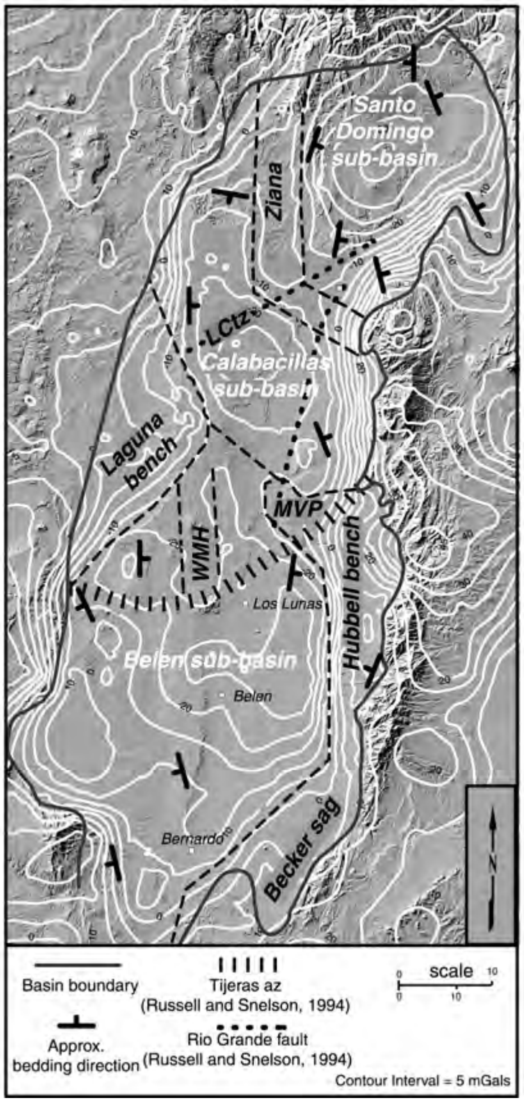

Figure 2. Shaded-relief image of the Albuquerque Basin and vicinity showing contours of the isostatic residual gravity anomaly as white contours (modified from Grauch et al., 1999). Approximate boundaries of major sub-basin depressions are shown by bold dashed lines. Major structural benches and intrabasinal positive areas include the Hubbell bench and Ziana structure (Personius et al., 2000), Mountainview Prong (MVP) and Laguna bench (terminology of Hawley, 1996), and Wind Mesa horst (WMH, Maldonado et al., 1999). Base image produced from U.S. Geological Survey National Elevation Database DEM data.

Cessation of widespread basin-fill deposition of the Santa Fe Group occurred at different times in different parts of the Albuquerque Basin, resulting in the preservation of a number of local tops to the Santa Fe Group (Connell et al., 2000). During the later part of the early Pleistocene (between 1.3-0.6 Ma), the ancestral Rio Grande began to incise deeply into Plio-Pleistocene basin fill to form the present river valley (Connell et al., 2000; Gile et al., 1981). Aggradation locally persisted into middle Pleistocene time along the front of the Manzanita and Manzano Mountains where tributary drainages were not integrated with the Rio Grande (Connell et al., 2000). The cause of this long-term entrenchment may be the result of: (1) drainage integration in the San Luis Basin of north-central New Mexico and south-central Colorado (Wells et al., 1987); (2) integration of the Rio Grande with the Gulf of Mexico (Kottlowski, 1953); (3) regional uplift (Bachman and Mehnert, 1978); or (4) shift in regional climate (Dethier et al., 1988).

Results of recent (published and unpublished) geologic mapping, stratigraphic, geomorphic, subsurface, radioisotopic, and biostratigraphic studies are reviewed in this overview of the stratigraphy of the Albuquerque Basin. This paper attempts to summarize results of mapping of over 60% of the basin that has occurred since 1994. Sedimentologic studies of basin-fill strata in the Albuquerque Basin and the Socorro region have been integrated in order to illustrate general sediment dispersal patterns (Bruning, 1973; Love and Young, 1983; Connell et al., 1999; Lozinsky and Tedford, 1991; Maldonado et al., 1999; Tedford and Barghoorn, 1999; Smith and Kuhle, 1998a; Smith et al., 2001). Geomorphic studies have delineated major constructional surfaces of the Santa Fe Group (Machette, 1985; Connell and Wells, 1999; Dethier, 1999; Maldonado et al., 1999). Subsurface data primarily involve deep oil-test and shallower water-well data (Lozinsky, 1994; Hawley, 1996; Hawley et al., 1995; Connell et al., 1998a; Cole et al., 1999), and regional gravity and aeromagnetic surveys (Grauch, 1999; Grauch et al., 1999; U.S. Geological Survey et al., 1999; Heywood, 1992). Sub-basin boundaries are defined by broad, generally discontinuous zones of high gravity that are interpreted as structurally higher intrabasinal fault blocks (Hawley, 1996, p. 12; Cole et al., 1999; Grauch et al., 1999).

Radioisotopic dates are from volcanic and volcaniclastic rocks that are interbedded with, underlie, or are overlain by, basin-fill. These dated volcanic rocks include mafic lava flows, ash-flow tuffs, fallout ashes and tuffs, and fluvially recycled pumice and tuff clasts in gravelly beds. Potassium-argon (K/Ar) dates are reported here to a precision of 0.1 Ma; $^{40}Ar/^{39}Ar$ dates are reported to a precision of 0.01 Ma, except where noted. Vertebrate fossils have been collected from numerous sites (Morgan and Lucas, 2000). Many of the fossils found in the basin have relatively long temporal ranges that limit precise stratigraphic correlation. In older deposits of the

Santa Fe Group, magnetostratigraphic studies permit correlation to other dated stratigraphic sections (Tedford and Barghoorn, 1999). Integration of various chronologic data greatly improves the chronologic resolution of basin-fill strata.

The main goal of this summary is to present an updated regional correlation and synthesis of the Santa Fe Group in the Albuquerque Basin. Recent insights on the stratigraphy and sedimentology of the basin-fill are presented in detail, primarily to clarify a rather confusing history of stratigraphic usage.

PRE-SANTA FE GROUP STRATIGRAPHY

Pre-rift strata are exposed along basin margins and in deep oil-test wells. These deposits include the Paleogene Galisteo and Diamond Tail formations, and Oligocene volcanic and volcaniclastic rocks derived from volcanic fields in New Mexico and southern Colorado, such as the Mogollon-Datil, San Juan, and Latir volcanic fields. The Galisteo and Diamond Tail formations are arkosic to subarkosic and typically lack volcanic detritus. These formations record deposition by major rivers draining Laramide uplifts during Paleocene and Eocene times (Lucas et al., 1997; Abbott et al., 1995; Ingersoll et al., 1990; Gorham and Ingersoll, 1979). Deposition of the Galisteo Formation was interrupted by widespread emplacement of intermediate to silicic volcanic rocks during late Eocene and Oligocene time; silicic volcanism was typically dominated by ignimbrite eruptions from caldera complexes and eruptive centers scattered throughout the southwestern United States and Mexico.

In central and northern New Mexico, these Oligocene eruptive centers include: the Ortiz porphyry belt (Ortiz Mountains and Cerrillos Hills), west of Santa Fe, the Mogollon-Datil volcanic field of western New Mexico, San Juan volcanic field of southern Colorado, and Latir volcanic field, just north of Taos, New Mexico. These volcanic and volcaniclastic rocks are discontinuously exposed along the southern and northeastern margins of the basin and are differentiated into three units: the Espinaso Formation, unit of Isleta #2, and volcanic and volcaniclastic units of the Datil Group and Mogollon-Datil volcanic field, including the La Jara Peak basaltic andesite The Santa Fe Group commonly overlies these Oligocene volcanic rocks, except along the northwestern part of the Calabacillas sub-basin where the Santa Fe Group overlies deposits of the upper Galisteo Formation (Lucas, 1982).

The Espinaso Formation crops out along Espinaso Ridge in the Hagan embayment, where it is about 430 m thick. The Espinaso Formation is a lithic arkose and conglomerate that formed a volcaniclastic apron around the neighboring Ortiz Mountains-Cerrillos Hills magmatic centers, which erupted between 26-37 Ma (Erskine and Smith, 1993; Kautz et al., 1981). Sandstone contains sparse to no quartz grains (Kautz et al., 1981). The Espinaso Formation conformably overlies the Galisteo Formation and is unconformably overlain by quartz-bearing lithic arkose and feldspathic arenite and volcanic-bearing conglomerate of the informally defined Tanos and Blackshare Formations of the lower Santa Fe Group (Connell and Cather, *this volume*; Cather et al., 2000).

The unit of Isleta #2 is an informal stratigraphic term applied to 1787-2185 m of upper Eocene-Oligocene strata recognized in at least six deep oil-test wells in the basin (Lozinsky, 1994; May and Russell, 1994). This volcanic-bearing succession is buried by up to 4400 m of Santa Fe Group deposits (Lozinsky, 1994). Two recent oil-test wells (Burlington Resources Kachina #1, and Davis Petroleum Tamara #1-Y) also encountered this unit in the Calabacillas sub-basin. The unit of Isleta #2 is composed of purplish-red to gray, subarkosic, volcanic-bearing sandstone with mudstone interbeds, and is therefore quite different from the composition of the Espinaso Formation. It is quite quartz rich (Q=68±9%, Lozinsky, 1994). The quartzose character and distance from known Oligocene-aged volcanic centers, and may suggest compositional maturation of instable volcanic constituents from these distant centers, which has been proposed to explain petrographic differences between the Santa Fe Group and Abiquiu Formation (Large and Ingersoll, 1997). Abundant quartz could also suggest possible contributions and mixing from other quartz-rich sources, such as on the adjacent Colorado Plateau (*see* Stone, 1979). An ash-flow tuff encountered in the unit's namesake well was K/Ar dated at 36.3±1.8 Ma (May and Russell, 1994), indicating a pre-rift heritage for the unit of Isleta #2.

Oligocene strata were not recognized on the Ziana structure (Shell Santa Fe Pacific #1; Black and Hiss, 1974). The Ziana structure is about 30 km west of Espinaso Ridge and marks the boundary between the Calabacillas and Santo Domingo sub-basins. The Davis Tamara #1-Y well, drilled about 6 km northwest of the Santa Fe Pacific #1 well, fully penetrated the Santa Fe Group section. Examination of the cuttings from the Tamara well suggests the presence of a lower 455-481-m thick interval of sand stratigraphically below the Piedra Parada Member suggests the presence of either an earlier sedimentary unit between the Piedra Parada Member and the Galisteo Formation.

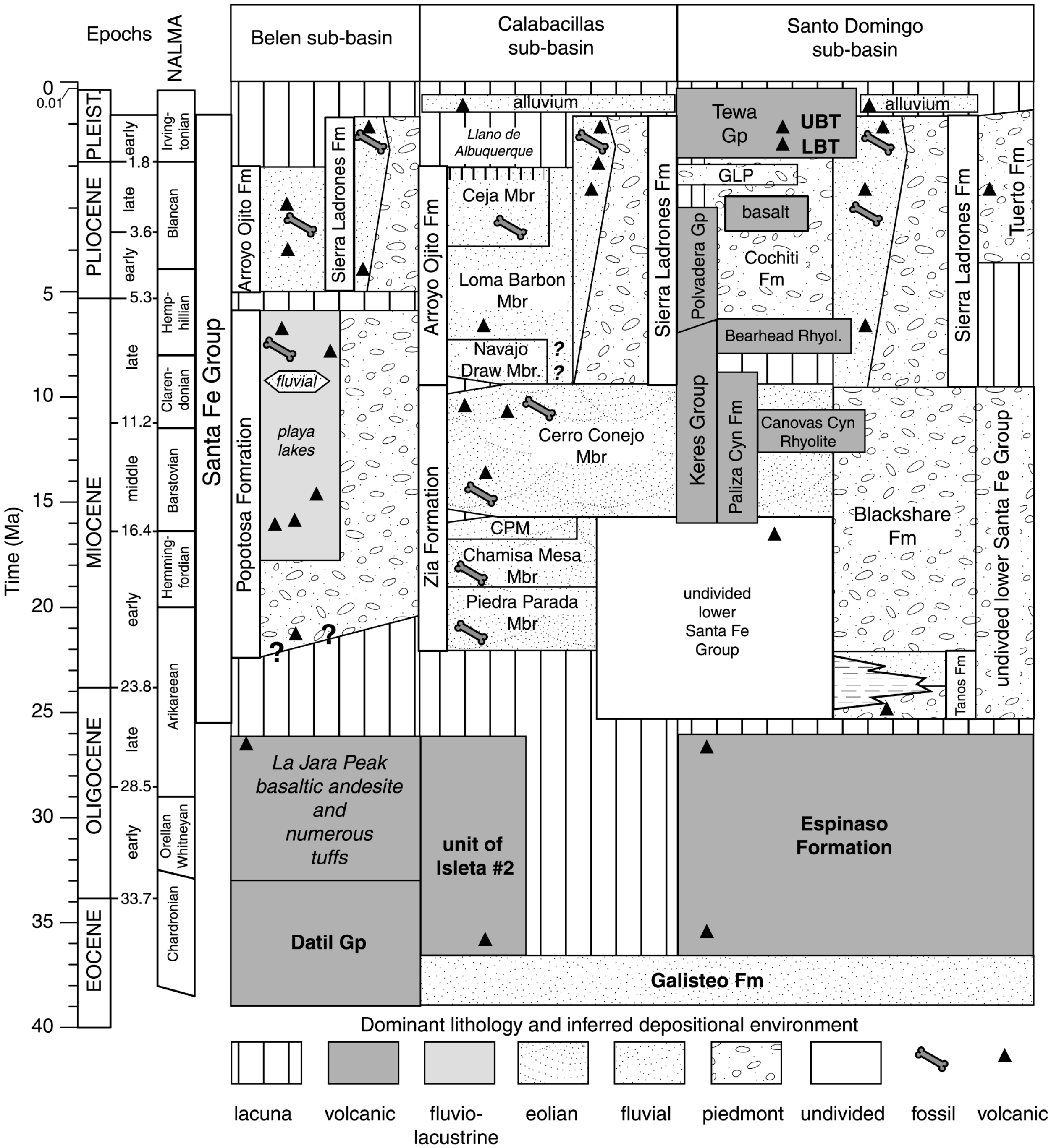

Figure 3. Schematic stratigraphic correlation diagram of the Albuquerque Basin and other basins of the Rio Grande rift, illustrating age-constraints and the North American Land Mammal "ages." Volcanic units include, the upper (UBT) and lower (LBT) Bandelier Tuff members of the Tewa Group. The Cañada Pilares Member of the Zia Formation (CPM) is locally recognized along the northwestern margin of the Calabacillas sub-basin. The gravel of Lookout Park (GLP) of Smith and Kuhle (1998a, b) is an unconformity-bounded gravel preserved on the hanging wall hinge of the Santo Domingo sub-basin.

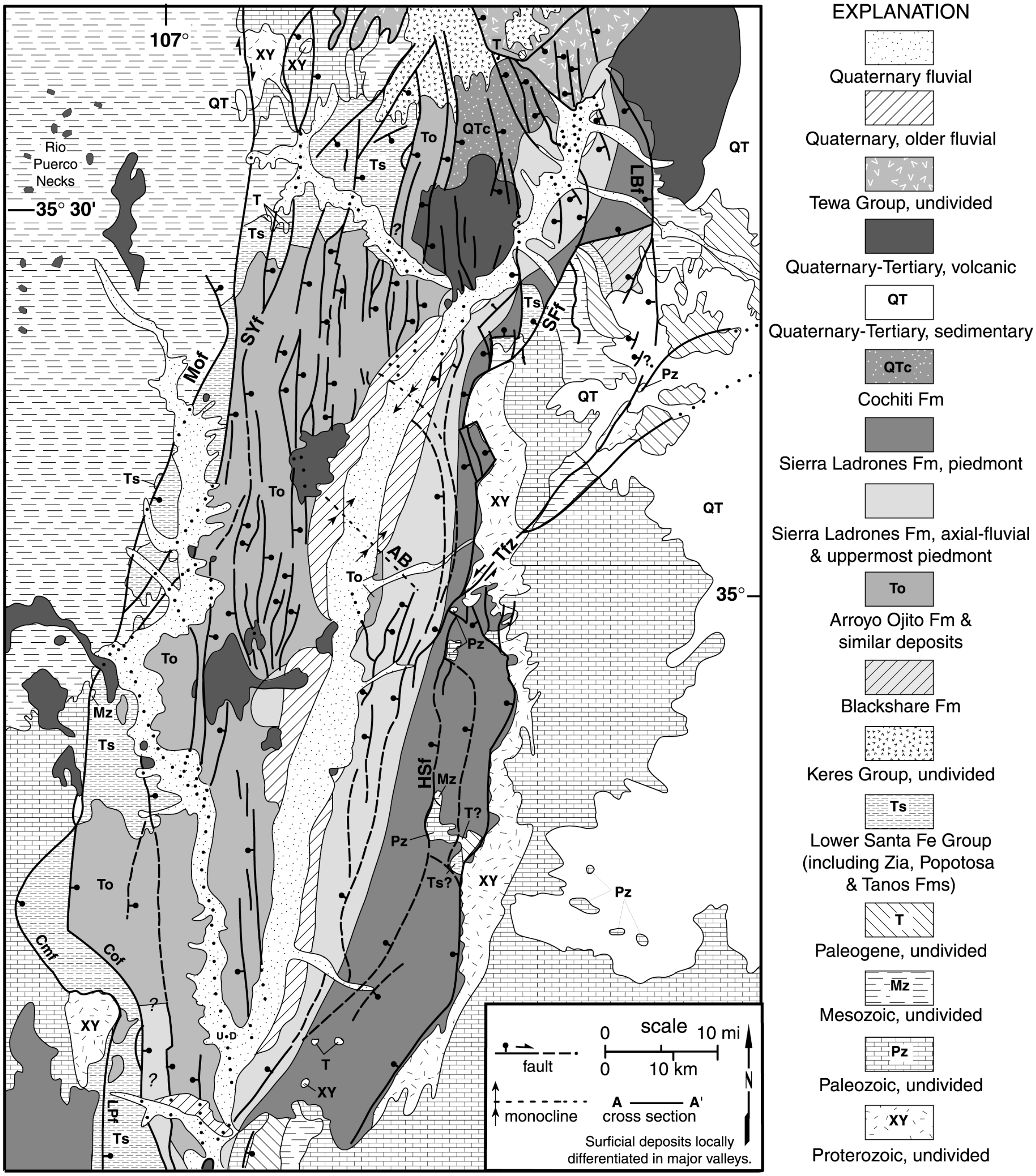

Figure 4. Generalized geologic map of the Albuquerque Basin, modified from Hawley (1996 and Hawley et al., 1995), with additional modifications from Osburn (1983), Machette et al. (1998), Maldonado et al. (1999), Connell (1997), Connell and Wells (1999), Connell et al. (1995, 1999), Cather and Connell (1998), Cather et al. (2000), Love and Young (1983), Personius et al. (2000), Smith and Kuhle (1998a, b), Lozinsky and Tedford (1991), Smith et al. (1970), and Goff et al. (1990). Line A-A' on the figure denotes the location of cross section on Figure 5. Faults include the Moquino (Mof), San Ysidro (SYf), San Francisco (SFf), Tijeras (Tfz), Hubbell Spring (HSf), Comanche (Cmf), Coyote (Cof), and Loma Peleda (LPf) faults.

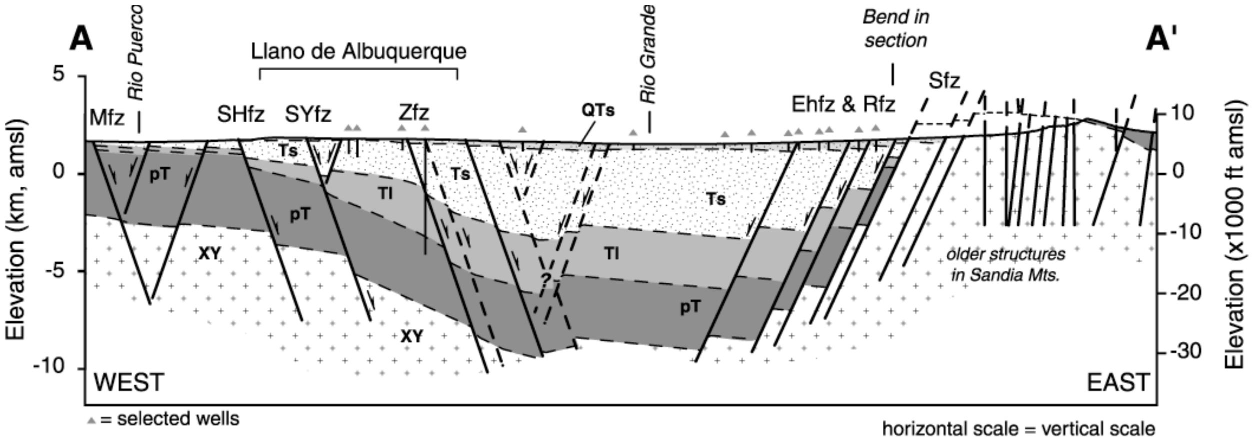

Figure 5. Generalized geologic cross section of Calabacillas sub-basin drawn at latitude of Paseo del Norte Boulevard in Albuquerque (Fig. 4). Gray triangles denote locations of selected wells that were used provide stratigraphic control for the cross section. The Llano de Albuquerque represents a broad mesa and local constructional top of the Arroyo Ojito Formation, and is the interfluve between the Rio Puerco and Rio Grande. Cross section illustrates projected depths of Proterozoic crystalline rocks (XY), pre-Tertiary (pT) sedimentary deposits, Paleogene volcanic and nonvolcanic deposits (Tl), and synrift basin fill of the Santa Fe Group (Ts, QTs). Oligo-Miocene deposits of the Santa Fe Group (Ts) include the Zia and Arroyo Ojito formations and undivided strata beneath Albuquerque, NM. Plio-Pleistocene deposits of the upper Santa Fe Group (QTs) include the upper Arroyo Ojito Formation and Sierra Ladrones Formation. Unit QTs comprises much of the aquifer used by the City of Albuquerque east of the Llano de Albuquerque. Major faults of the western margin include the Moquino (Mfz), Sand Hill (SHfz), San Ysidro (SYfz), and Zia (Zfz) fault zones. Major eastern-margin fault zones include the East Heights (EHfz), Rincon (Rfz), and Sandia (Sfz) fault zones.

Cenozoic strata in the Tamara well are petrographically distinct from the Abiquiu Formation (Connell, Koning, and Derrick, *this volume*). Additional study, however, is required to determine the spatial relationships among these possible Oligo-Miocene deposits in the northwest Calabacillas sub-basin with Abiquiu Formation sediments in the Chama sub-basin. This lower interval in the Tamara well may be correlative to the unit of Isleta #2, which is about 2.2 km thick in the Shell West Mesa Federal #1, about 25-30 km to the southeast. Correlation of this lower interval to the unit of Isleta #2 is supported by the presence of a discontinuous layer of Oligocene volcanic pebbles and cobbles at the exposed contact between the Zia Formation and subjacent strata along the western basin margin. The presence of this volcanic gravel at this contact indicates the presence of a formerly more extensive Oligocene deposit that has subsequently been eroded.

Deposits of the Mogollon-Datil volcanic field comprise an areally extensive succession of upper Eocene-Oligocene (27-34 Ma; Osburn and Chapin, 1983), ash-flow tuffs, basaltic lavas, and volcaniclastic deposits exposed in the southern Belen sub-basin. Eocene outflow tuffs were assigned to the upper Eocene Datil Group. A variety of Oligocene tuffs and cauldron-fill units overlie the Datil Group and include the 33.1 Ma Hells Mesa Tuff, 28.4 Ma Lemitar Tuff, 26-27 Ma La Jara Peak basaltic andesite and South Canyon Tuff (K/Ar dates reported in Osburn and Chapin, 1983; Bachman and Mehnert, 1978). This Oligocene volcanic succession is dominated by intermediate and silicic tuffs that are commonly densely welded. The upper part of this succession generally becomes slightly more heterolithic and contains a greater abundance of basaltic and basaltic andesite rocks (Osburn and Chapin, 1983).

An exposure of volcaniclastic sediments was recognized along the western front of the Manzano Mountains, near the mouth Trigo Canyon (Kelley, 1977). No crystalline rocks derived from the western front of the Manzano Mountains are recognized in these deposits (Karlstrom et al., 2001). A basalt flow near Trigo Canyon, at the front of the Manzano Mountains, was originally K/Ar dated at 21.2±0.8 Ma by Bachman and Mehnert (1978). Kelley (1977) considered this basalt to be a sill within the Datil Group. An $^{40}Ar/^{39}Ar$ date of 26.20±0.18 Ma (Karlstrom et al., 2001) for this flow indicates that the previous K/Ar date is too young and may have been affected by alteration. Lozinsky (1988) demonstrated the subaerial nature of this flow. On the basis of the K/Ar age and slightly heterolithic character of the volcanic gravel, he assigned these strata to the Popotosa Formation. The new date indicates that this flow is similar in age to the pre-rift Cerritos de las Minas flow (Machette, 1978a) and lies within the age range of the La Jara Peak basaltic andesite (Osburn and Chapin, 1983). The Leroy Bennett-Aguayo Comanche #1 oil-test, drilled a few kilometers north of Trigo Canyon, encountered at least 350 m of similarly described volcanic and volcaniclastic sediments (from scout ticket; Karlstrom et al., 2001). A 26 Ma date for a such a thick succession of volcanic sediments and the lack of locally derived detritus from the western front of the Manzano mountains supports correlation to subjacent Oligocene volcanic rocks, rather than the Popotosa Formation; however, additional study is needed to resolve the stratigraphic assignment of these conglomeratic beds.

SANTA FE GROUP STRATIGRAPHY AND CHRONOLOGY

Deposits of the Santa Fe Group (Spiegel and Baldwin, 1963) have been differentiated into two, and in some places three, informal sub-groups. The lower Santa Fe Group records deposition in internally drained basins (bolsons) where streams terminated onto broad alluvial plains with ephemeral or intermittent playa lakes bounded by piedmont deposits derived from emerging basin-margin uplifts. Upper Santa Fe Group strata record deposition in externally drained basins where perennial streams and rivers associated with the ancestral Rio Grande fluvial system flowed toward southern New Mexico. The middle sub-group or formation is transitional between the lower interval, representing deposition within internally drained basins, and the upper interval, representing deposition in an externally drained basin. Deposition ceased during Pleistocene time, when the Rio Grande began to incise into the earlier aggradational phase of the Santa Fe Group basin fill (Hawley et al., 1969).

Some workers (Bryan and McCann, 1937; Spiegel, 1961; Lambert, 1968; Kelley, 1977) advocated a three-part subdivision of the Santa Fe Group in the Albuquerque area, principally because of the presence of deposits that are transitional in character between the early phase of eolian, playa-lake, and fluviolacustrine sedimentation, and a later phase of fluvially dominated deposition. Unfortunately, the use of a middle Santa Fe term has been somewhat confusing, principally because of different lithostratigraphic definitions and interpretations by various workers (*see* Connell et al., 1999). Bryan and McCann (1937) proposed the term “middle red” for deposits that are mostly correlative to the Cerro Conejo Member (Connell et al., 1999). Other workers (Spiegel, 1961; Lambert, 1968; Kelley, 1977) later extended the middle red to higher stratigraphic levels than proposed by Bryan and his students (e.g., Wright, 1946; Bryan and McCann, 1937). The middle Santa Fe Group concept is useful for hydrogeologic studies (Hawley et al., 1995; Hawley and Kernodle, *in press*); however, for the purpose of this summary, this middle sub-group term is avoided in order to avoid confusion with conflicting and overlapping usage by previous workers.

Volcanic Rocks of the Jemez Mountains

The Jemez Mountains were formed by multiple volcanic eruptions since middle Miocene time. They lie on a northeast-trending zone of Quaternary and Pliocene volcanic fields called the Jemez lineament (Mayo, 1958). The volcanic rocks of the southern Jemez Mountains are placed into the Keres, Polvadera, and Tewa Groups (Figs. 3-4; Bailey et al., 1969; Smith et al., 1970). The southern Jemez Mountains are largely composed of the Miocene Keres Group. The central and northern Jemez Mountains contain the Miocene-Pliocene Polvadera Group, and the Plio-Pleistocene Tewa Group. The Keres and Polvadera groups represent volcanic events prior to the emplacement of the areally extensive Tewa Group, which covers much of the Jemez Mountains. Volcanic strata were erupted contemporaneously with subsidence in the Española Basin and Abiquiu embayment (Chama sub-basin).

The Keres Group contains basaltic, andesitic, dacitic, and rhyolitic volcanic rocks, which are subdivided into the Canovas Canyon Rhyolite (12.4-8.8 Ma; Gardner et al., 1986), Paliza Canyon Formation (13.2-7.4 Ma; Gardner et al., 1986), and Bearhead Rhyolite (7.1-6.2 Ma; Gardner et al., 1986). The Paliza Canyon Formation is lithologically variable and contains basaltic, andesitic, and dacitic rocks that extend to within 2-4 km of the eastern front of the Sierra Nacimiento (Smith et al., 1970). The 10.4±0.5 Ma basalt of Chamisa Mesa (Luedke and Smith, 1978) is included within the Paliza Canyon Formation (Gardner et al., 1986). The Bearhead Rhyolite defines the top of the Keres Group and contains the Peralta Tuff Member (6.16-6.96 Ma; Smith et al., 2001; Justet, 1999; McIntosh and Quade, 1995).

The Polvadera Group in the central Jemez Mountains contains the Tschicoma Formation (6.9-3.2 Ma; Gardner et al., 1986), which represents eruptions from a pre-Tewa Group volcanic edifice situated near the central and northeastern part of the Jemez Mountains.

The Tewa Group is a voluminous succession of rhyolitic tuff and volcanic flows that represent the most recent stage of major volcanism in the Jemez Mountains. The Tewa Group includes the Valles Rhyolite (0.1-1.0 Ma), Cerro Toledo Rhyolite (1.2-1.5 Ma), Bandelier Tuff, and Cerro Rubio quartz latite (2.2-3.6 Ma) (Gardner et al., 1986). The Bandelier Tuff and Cerro Toledo Rhyolite are locally important stratigraphic units in the Albuquerque Basin. The early Pleistocene Bandelier Tuff is the most extensive unit and is subdivided into lower (Otowi and Guaje, 1.61 Ma) and upper (Tshirege and Tsankawi, 1.22 Ma) members ($^{40}Ar/^{39}Ar$ dates of

Izett and Obradovich, 1994), which were deposited during the collapse of the Toledo and Valles calderas, respectively. Primary and fluvially recycled tephra of the Bandelier Tuff are locally common in the uppermost part of the axial-fluvial facies of the Sierra Ladrones Formation.

Lower Santa Fe Group

The lower Santa Fe sub-Group ranges from late Oligocene through late Miocene in age and records deposition in internally drained basins. These deposits are exposed along the basin margins and are either in fault contact with, or are unconformably overlain by, deposits of the upper Santa Fe Group; however, the upper/lower sub-group boundary is probably sub-basin within sub-basin depocenters (Cather et al., 1994). Lower Santa Fe Group sediments record deposition in an internally drained bolson (Hawley, 1978). The lower Santa Fe Group contains three major facies that are subdivided into four formations (Zia, Popotosa, Tanos, Blackshare formations): (1) piedmont facies consisting of stream- and debris-flow deposits derived from uplands along the basin margin piedmont slope; (2) basin-floor fluviolacustrine (playa-lake) facies consisting of ephemeral or intermittent playa lake and local fluvial deposits; and (3) eolian facies consisting of cross-bedded to massive, well sorted, fine-to medium-grained sandstone. Deposit composition reflects the lithology of upland drainages and contains sedimentary, volcanic, plutonic, and metamorphic rocks. Fluviolacustrine facies are exposed in the western and southwestern parts of the Belen sub-basin and northeastern Santo Domingo sub-basin and interfinger with piedmont facies derived from emerging rift-flank uplifts. Eolian sandstone is exposed in the western and northwestern parts of the Calabacillas sub-basin. The lateral boundary between eolian and fluviolacustrine facies is not exposed, but lies between the Burlington Resources Kachina #1 well, which encountered well sorted sandstone correlated to the Zia Formation (J.W. Hawley, 1998, oral commun.), and the Shell Isleta #2 well, where mudstone and muddy sandstone of the Popotosa Formation are recognized (Lozinsky, 1994). Thus, the lateral boundary between the Zia and Popotosa formations lies near the geophysically defined boundary of the Calabacillas and Belen sub-basins, suggesting structural control over this facies boundary (Cole et al., 1999).

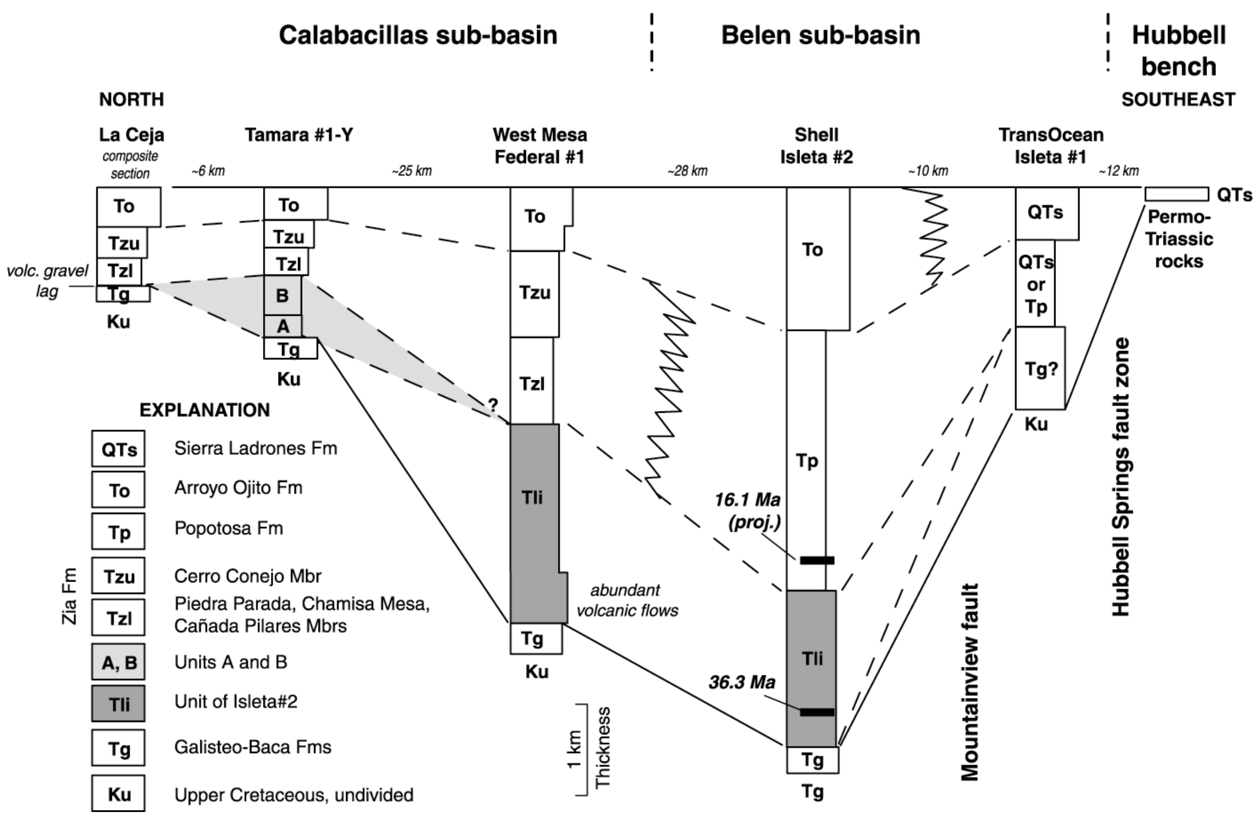

Figure 6. Stratigraphic fence of Cenozoic deposits in the Calabacillas sub-basin. Data from oil test wells (Lozinsky, 1988, 1994; Connell, Koning, and Derrick, *this volume*; Connell et al., 1999; Tedford and Barghoorn, 1999; Maldonado et al., 1999; Black and Hiss, 1974). Locations of wells and stratigraphic sections on **Figure 1**. Units A and B are interpreted as pre-Piedra Parada Member deposits encountered in the Tamara well.

Tanos and Blackshare Formations

The Tanos and Blackshare formations are newly proposed names for well-cemented, moderately tilted conglomerate, sandstone, and mudstone of the lower Santa Fe Group, exposed in the Hagan embayment (Connell and Cather, *this volume*). These informal units are unconformably overlain by the Tuerto Formation. The Tanos Formation is a 253-m thick succession of conglomerate, thinly to medium bedded mudstone and tabular sandstone that rests disconformably upon the Espinaso Formation. The age of the base of the Tanos Formation is constrained by an olivine basalt flow about 9 m above its base, which yielded a $^{40}Ar/^{39}Ar$ date of 25.41±0.32 Ma (Cather et al., 2000; Peters, 2001b), supporting an earlier K/Ar date of about 25.1±0.7 Ma (Kautz et al., 1981). Thus, the basal Santa Fe Group deposits at Espinaso Ridge are slightly older than the basal Zia Formation exposed along the western margin of the Calabacillas sub-basin. Thus, the basal Santa Fe Group deposits at Espinaso Ridge are slightly older than the basal Zia Formation exposed along the western margin of the Calabacillas sub-basin. The basal contact is sharp and scoured. A continuous dip-meter log for a nearby oil-test well indicates the presence of an angular unconformity between the Tanos and Espinaso formations.

The mapped extent of the Tanos Formation roughly coincides to strata tentatively correlated to the Abiquiu Formation by Stearns (1953) and to the Zia Formation by Kelley (1979). Stearns (1953) assigned these beds to the Abiquiu Formation, principally because of the abundance of volcanic detritus in the section. Kelley (1977) correlated them to the Zia Formation, probably on the basis of stratigraphic position, light coloration and thick tabular sandstone beds. Recent studies (Cather et al., 2000; Large and Ingersoll, 1997) indicate that these deposits were locally derived by west-northwest-flowing streams from the Ortiz Mountains, rather than from the more rhyolitic Latir eruptive center to the north near Taos, New Mexico (Ingersoll et al., 1990). Kelley (1977) interpreted these facies to be related to the Zia Formation, however the lack of large-scale crossbedding and presence of abundant mudstone suggests basin-floor deposition in basin-floor (playa-lake and mudflat) and piedmont-slope environments, rather than in an eolian dune field. These deposits are also considerably less quartz-rich than those of the Zia Formation.

The Tanos Formation is, in part, temporally equivalent to the Abiquiu Formation, but are not included in the Abiquiu Formation because they contain abundant locally derived volcanic grains and clasts that are derived from the adjacent Ortiz Mountains (Large and Ingersoll, 1997), rather than from the Latir volcanic field (Smith, 1995; Moore, 2000; Large and Ingersoll, 1997). Tanos Formation strata are not considered part of the Zia Formation, primarily because the Tanos Formation contains a thick succession of mudstone and fluvial sandstone interpreted to be deposited in a basin-floor, playa-lake/distal-piedmont setting.

The Tanos Formation is conformably overlain by a >700 m succession of sandstone and conglomerate informally called the Blackshare Formation, for the nearby Blackshare Ranch, which is in a tributary of Tanos Arroyo. The Blackshare Formation is a succession of interbedded sandstone, conglomerate and thin mudstone. Conglomerate beds are commonly lenticular and sandstone intervals commonly fine upward into thin mudstone beds that are commonly scoured by overlying lenticular conglomerate. The upper boundary of the Tanos Formation is gradational and interfingers with the overlying Blackshare Formation. An ash within the Blackshare Formation is projected to be ~670-710 m above the base. This ash yields a $^{40}Ar/^{39}Ar$ date of 11.65±0.38 Ma (Connell and Cather, *this volume*). Estimates of stratal accumulation rates (not adjusted for compaction) for much of the Tanos-Blackshare succession, based on these two dates, is about 72 m/m.y..

Zia Formation

The Zia Formation ranges from 350 m to at least 853 m in thickness and represents a predominantly eolian phase of lower Santa Fe Group deposition in the Calabacillas sub-basin. It is exposed along the eastern margin of the Rio Puerco valley (Ceja del Rio Puerco of Bryan and McCann, 1937, 1938) and along the southwestern margin of the Rio Jemez valley (Rincones de Zia, Galusha, 1966; Tedford, 1981). The southern limit of exposures of the Cerro Conejo Member are near Benavidez Ranch, about 15 km west of Rio Rancho (Morgan and Williamson, 2000). Bryan and McCann (1937) informally designated the lowermost sediments as the “lower gray” member of their Santa Fe formation.

The Zia Formation is characterized by massive to cross-stratified, weakly to moderately cemented, well to moderately sorted arkose to feldspathic arenite with scattered thin to medium bedded muddy sandstone and mudstone interbeds (Beckner, 1996; Connell et al., 1999; Tedford and Barghoorn, 1999). Concretionary zones cemented with poikilotopic calcite crystals (Beckner and Mozley, 1998) are common in the lower members, but decrease in abundance upsection (Connell et al., 1999). Paleocurrent observations indicate wind from the west (Gawne, 1981). The Zia Formation is subdivided into four members, in ascending stratigraphic order: the Piedra Parada, Chamisa Mesa, Cañada Pilares, and Cerro Conejo members. The two lowest members were defined by Galusha (1966). Gawne (1981) defined the Cañada Pilares Member, and Connell et al. (1999) proposed the Cerro Conejo Member to round out the Zia Formation stratigraphy.

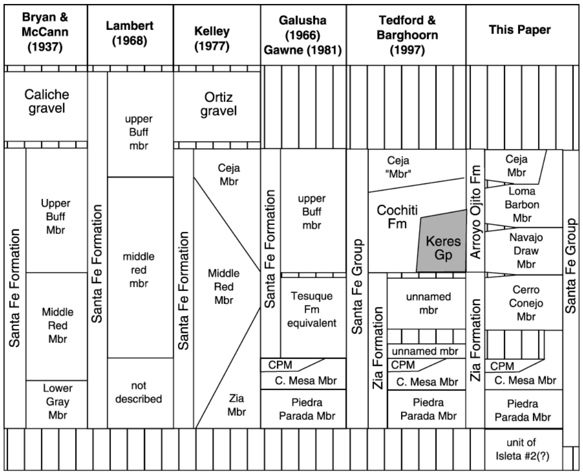

Figure 7. Summary of stratigraphic nomenclature development in the northwestern Calabacillas sub-basin. Sedimentary units include the Cañada Pilares Member (CPM) of the Zia Formation. Volcanic rocks are shaded gray.

The Piedra Parada Member is a 70-m thick eolianite succession resting upon a low relief unconformity cut onto subjacent strata (Tedford and Barghoorn, 1999). The basal contact contains a nearly continuous lag of siliceous pebbles and small cobbles derived from the subjacent Galisteo Formation and Oligocene volcanic rocks. These intermediate volcanic rocks have been shaped into ventifacts and locally lie on a calcic soil developed on older deposits (Tedford and Barghoorn, 1999). Three volcanic cobbles at this contact were dated at 31.8±1.4 Ma, 33.03±0.22, and 33.24±0.24 Ma using the $^{40}Ar/^{39}Ar$ technique on hornblende and biotite (S.M. Cather and W.C. McIntosh, written commun., 2000). The Piedra Parada Member records deposition of an eolian dune field with ephemeral interdunal ponds and sparse, widely spaced fluvial channel deposits (Gawne, 1981). A basal pebbly sandstone mostly composed of siliceous pebbles recycled from recycled Galisteo Formation and Mesozoic strata on the Colorado Plateau is present at Galusha's (1966) type Piedra Parada Member section. Paleocurrent analyses of this discontinuous basal fluviatile interval by Gawne (1981) indicate eastward paleoflow, although there is considerable scatter in her data. These clasts could have been derived from the Mogollon-Datil volcanic field to the south, the unit of Isleta #2 to the southeast, Ortiz Mountains to the east, or possibly from the San Juan volcanic field to the north; however, the proximity of these deposits to the unit of Isleta #2 in drillholes to the south suggest a probable derivation from the unit of Isleta #2.

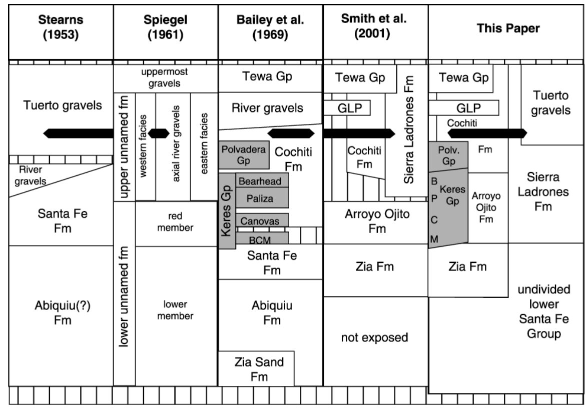

Figure 8. Summary of development of stratigraphic nomenclature in the Santo Domingo sub-basin. Shaded units are volcanic; black shading indicates the basalts of Santa Ana Mesa and Cerros del Rio. Other sedimentary units include the gravel of Lookout Park (GLP) of Smith and Kuhle (1998a, b). Volcanic units include the basalt of Chamisa Mesa (M), Canovas Canyon (CC) Formation, Paliza Canyon Formation (P), basalt at Chamisa Mesa (BCM), and Bearhead Rhyolite (B). Volcanic rocks are shaded gray. Pliocene basaltic rocks are shaded black.

Fossil mammals collected from the lower 20 m of the Piedra Parada type section and in Cañada Pilares are latest Arikareean in age (19-22 Ma, Tedford and Barghoorn, 1999). These fossils are closely correlative to fossils of the "upper Harrison beds" of Nebraska (MacFadden and Hunt, 1998), which are about 19 Ma (R.H. Tedford, 2000, written commun.). Magnetostratigraphic and biostratigraphic studies by Tedford and Barghoorn (1999) indicate that the Cañada Pilares and Cerro Conejo members accumulated at a rate of about 69-83 m/my. Extrapolation of this stratal accumulation rate to the base of the Zia Formation support an age of about 19 Ma for the base of the Piedra Parada Member (R.H. Tedford, 2000, written commun.).

The Piedra Parada Member grades upsection into the Chamisa Mesa Member (Galusha, 1966), which represents deposition of eolian sand sheets and a slight increase in fluvial and local lacustrine deposition (Tedford and Barghoorn, 1999; Gawne, 1981). Mammalian remains indicate deposition during late-early Miocene time (early to late Hemingfordian, 16-18 Ma; Tedford and Barghoorn, 1997).

The Zia Formation was further sub-divided into the late Hemingfordian (16-18 Ma; Tedford and Barghoorn, 1999) Cañada Pilares Member (Gawne, 1981), a 20- to 30-m thick succession of red and green, fluviolacustrine claystone and limestone, and thinly bedded pink sandstone, and eolian sandstone overlying the Chamisa Mesa Member (Tedford and Barghoorn, 1999; Gawne, 1981).

The Cerro Conejo Member is the highest member of the Zia Formation. The Cerro Conejo Member contains 300-320 m of very pale-brown to pink and yellowish-red, tabular to cross-bedded, moderately to well sorted sand, with minor thinly bedded mud, and rare very fine-grained pebbly sand. At the type section, the Cañada Pilares Member is missing. The top of the Cerro Conejo is conformable and Along the northern Ceja del Rio Puerco, near Navajo Draw, the contact between the Cerro Conejo and Navajo Draw Members is sharp on the footwall of the San Ysidro fault. To the east, this contact is gradational and both members interfinger (Connell et al., 1999; Koning and Personius, *in review*).

The Cerro Conejo Member locally forms prominent ledges and cliffs and is slightly redder and more thickly bedded than the more topographically subdued Piedra Parada and Chamisa Mesa members. At the type locality, over a quarter of the section contains thickly bedded, cross stratified, fine- to coarse-grained sand that locally exhibit multiple grain-fall and grain-flow laminations with local reverse grading, indicating eolian deposition. Much of the section is a mixture of massive to cross-bedded sand with subordinate, thinly to medium bedded sandy mud and mud. Mudstone beds and lenticular bedforms are more abundant in the overlying Arroyo Ojito Formation. Gravelly sand beds are rare south of the Rio Jemez valley (Connell et al., 1999), but contain a slightly greater abundance of pebbly sand north of the Rio Jemez (Chamberlin et al., 1999).

Biostratigraphic data indicate that the Cerro Conejo is late Barstovian to Clarendonian (14-8 Ma; Tedford and Barghoorn, 1999; Connell et al., 1999; Morgan and Williamson, 2000), or middle to late Miocene, in age. The Rincon quarry of Galusha (1966) contains fossils correlated to the late Barstovian land-mammal "age," which is about 12-14 Ma (Tedford and Barghoon, 1999). This quarry was re-located in the fall of 1999 and projected near the base of the type section, and not within higher units, as previously thought (*see* Connell et al., 1999). At least five altered volcanic ashes are present in the middle of this unit. Tedford and Barghoorn (1997) report a K/Ar date of 13.64±0.09 Ma on biotite from a volcanic ash near Cañada Pilares along the Ceja del Rio Puerco. A stratigraphically higher ash-bearing sequence is present just east of the Ziana structure, near US-550, where a 10.8-11.3 Ma tephras are tentatively correlated to the Trapper Creek sequence in Idaho (Personius et al., 2000; Koning and Personius, *in review*; Dunbar, 2001, oral commun., Sarna-Wojciki, 2001, written commun.). The upper part of the Cerro Conejo Member is interbedded with the 10.4 Ma basalt of Chamisa Mesa and is overlain by 9.6 Ma flows of the Paliza Canyon Formation (Chamberlin et al., 1999) along the southern flank of the Jemez Mountains. Thus, deposition of the Cerro Conejo Member occurred during part of middle to late Miocene time (*ca*. 14-10 Ma).

Magnetostratigraphic studies along the Ceja del Rio Puerco indicate the presence of a 1-1.6 m.y. hiatus in deposition near the boundary of the Cañada Pilares and Cerro Conejo members (Tedford and Barghoorn, 1999). At the type section, the basal contact with Chamisa Mesa Member sandstone is sharp. Estimates of stratal accumulation rates (not adjusted for compaction) for the Piedra Parada-Cerro Conejo succession is 79-83 m/m.y. (Tedford and Barghoorn, 1999).

The stratigraphic assignment of this unit has created debate based on the interpretation of depositional environments (Connell et al., 1999; Pazzaglia et al., 1999; Tedford and Barghoorn, 1999). The Cerro Conejo Member, originally part of Galusha's (1966) "Tesuque Formation equivalent" unit, was assigned to an upper unnamed member of the Zia Formation by Tedford and Barghoorn (1997). They subsequently included these deposits in the Arroyo Ojito Formation because of the greater proportion of fluvial sand and mud in the unit.

The Cerro Conejo Member is interpreted here to represent a transition between the lower, well sorted, sandy, eolian-dominated deposits of the Piedra Parada-Cañada Pilares succession, and the overlying, more poorly sorted, fluvially dominated units of the Arroyo Ojito Formation. Connell et al. (1999) placed the Cerro Conejo Member within the Zia Formation, based primarily on lithologic similarities to underlying members of the Zia Formation. In contrast, Tedford and Barghoorn (1999) assigned the Cerro Conejo Member to the Arroyo Ojito Formation on the basis of lithogenetic interpretations. A strictly lithologic criterion for the placement of the Cerro Conejo Member within the Zia Formation is preferred, primarily because of the sandy nature of the unit and lack of thickly bedded mudstone and conglomeratic beds, which are more abundant in the overlying fluvially dominated Arroyo Ojito Formation. Alternatively, the Cerro Conejo Member may be lithologically distinct enough to assign as its own formation, which could indicate the transitional status of this unit between the lower and upper subgroups of the Santa Fe Group. The Cerro Conejo should, however, not be included in the Arroyo Ojito Formation, because it is lithologically distinct from the fluvially dominated deposits of the overlying Arroyo Ojito Formation.

The Zia Formation is partly equivalent in age to the Oligo-Miocene Abiquiu Formation, a volcaniclastic sandstone and conglomerate derived from the Latir volcanic field in northern New Mexico. The Abiquiu Formation is exposed along the northwestern flank of the Jemez volcanic field and on the crest of the northern Sierra Nacimiento (Smith et al., 1970; Woodward, 1987; Woodward and Timmer, 1979). Petrographic studies (Beckner, 1996; Large and Ingersoll, 1997) indicate that the Zia and Abiquiu Formations are petrographically dissimilar; however, definitive evidence regarding stratigraphic relationships between these units is not known. Zia Formation sandstone is quartz-rich compared to the Abiquiu Formation and was deposited by winds from the west-southwest, with widely scattered south-southeast flowing streams (Gawne, 1981). Abiquiu Formation sandstone contains abundant feldspar and lithic fragments and was deposited by southwest-flowing streams that drained the Latir volcanic field (Smith, 1995; Moore, 2000). Sparse gravels in the Piedra Parada Member contain abundant rounded chert and quartzite with scattered intermediate volcanic rocks. The eastward transport direction of Zia Formation eolian sandstone suggests that this unit could have been recycled from arkose and subarkose of Mesozoic-Paleogene rocks exposed in the adjacent Colorado Plateau (Stone et al., 1983). Minor recycling of Abiquiu Formation strata cannot be ruled out during Zia time. The presence of Pedernal chert, a chalcedony and chert that comprises the middle member of the Abiquiu Formation (Moore, 2000; Woodward, 1987), in the overlying Arroyo Ojito Formation, demonstrates recycling of Abiquiu sediments into the Albuquerque Basin during late Miocene and Pliocene time. The presence of Pedernal Member clasts in the San Juan Basin (Love, 1997) and southeast paleoflow indicators in the Arroyo Ojito Formation, also suggest that the Abiquiu Formation probably extended west of the Sierra Nacimiento, and thus may have provided an additional source of sediment into the Albuquerque Basin. Additional study is needed to further constrain the lateral extent of the Abiquiu Formation in the San Juan Basin.

An anomalously thick succession of lower Santa Fe Group was recognized by Kelley (1977, p. 14) in the Santa Fe Pacific #1 test well, which was spudded in the Zia Formation (Black and Hiss, 1974), about 10 km east of the Zia Formation type area. This well encountered 853 m of Zia Formation strata above the Galisteo Formation. This is much thicker than the 350 m measured at the type localities (Connell et al., 1999) and indicates that the Zia Formation thickens considerably, east of the type sections on Zia Pueblo. At least 762 m of Zia Formation sandstone was recognized in the Davis Petroleum Tamara #1-Y well (Connell, Koning, and Derrick, *this volume*). Kelley (1977) speculated that the basal Zia Formation exposed to the west might be younger than the basal Zia Formation encountered in these wells. The difference in thickness between these two wells and the absence of Oligocene strata under the Ziana structure and on the exposed contact with the Zia Formation to the west suggest that erosion of older strata occurred prior to about 19 Ma in the northwestern part of the Calabacillas sub-basin.

Popotosa Formation

The Popotosa Formation comprises an >1860 m succession of moderately to well cemented, and moderately tilted, conglomerate, mudstone, and sandstone exposed along the margins of the Belen sub-basin. The Popotosa Formation was defined by Denny (1940), who considered it to be a pre-Santa Fe Group deposit. Machette (1978a) later assigned it to the lower Santa Fe Group (Fig. 9). The Popotosa Formation rests unconformably on the subjacent La Jara Peak basaltic andesite and Cerritos de las Minas (Machette, 1978a; Osburn and Chapin, 1983) and is unconformably overlain by fluvial and basin-margin deposits of the upper Santa Fe Group (Sierra Ladrones Formation; Machette, 1978a). The piedmont and fluviolacustrine members, or facies, constitute the major facies of the Popotosa Formation. Bruning (1973) designated a reference section in Silver Creek, a tributary of the Rio Salado, where he described three dominant facies: a piedmont facies; a fluviolacustrine facies; and the granite-bearing fanglomerate of Ladron Peak (Bruning, 1973; Chamberlin et al., 1982; Cather et al., 1994). The piedmont facies contain 820-1860 m of predominantly volcanic-bearing conglomerate representing deposition of coarse-grained, stream- and debris-flows deposits derived from adjacent footwall uplands along the basin margin (Bruning, 1973; Lozinsky and Tedford, 1991). These deposits interfinger with fine-grained strata of the fluviolacustrine facies, which are 240-1070 m in exposed thickness (Bruning, 1973). The fluviolacustrine facies is the most distinctive and contains light-gray and light-grayish-green to medium reddish-brown, poorly sorted, silty clay to sand with sparse pebbly beds. This facies also contains primary (bedded) and secondary (fracture fill) gypsum and numerous middle-late Miocene ash beds (Cather et al., 1994; Bruning, 1973). This facies represents deposition in a very low-gradient playa lake or alluvial flat bounded by sandy, distal alluvial fan deposits (Lozinsky and Tedford, 1991; Bruning, 1973). The fanglomerate of Ladron Peak is 150-915 m thick (Bruning, 1973), rests conformably on fluviolacustrine and piedmont facies, and is associated with the flanks of the Ladron Mountains (Bruning, 1973; Chamberlin et al., 1982). The Popotosa Formation typically dips more steeply (about 15-35º; Cather et al., 1994) and is better cemented than the overlying deposits of the upper Santa Fe Group.

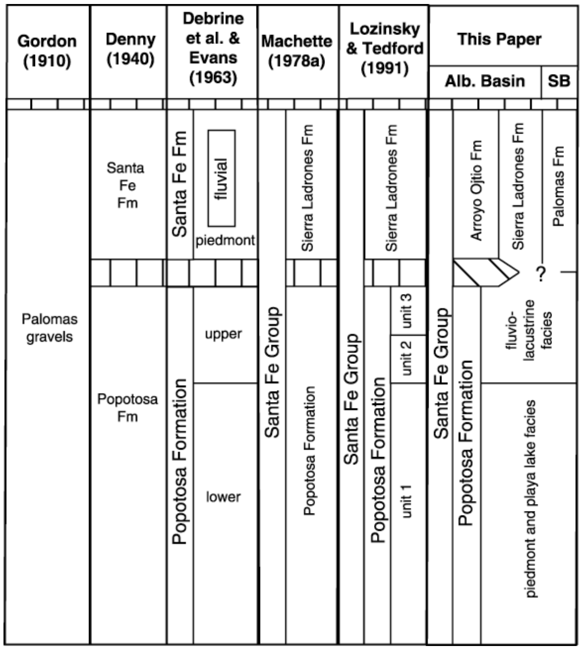

Figure 9. Summary of stratigraphic nomenclature development in the Belen sub-basin, illustrating the evolution of stratigraphic terms in the northern Socorro Basin and Belen sub-basin.

The age of the Popotosa Formation is constrained by biostratigraphic and radioisotopic data, mostly from the Socorro region. The Popotosa Formation rests unconformably on the 26.3±1.1 Ma andesite at Cerritos de las Minas (Bachman and Mehnert, 1978; Machette, 1978a). The top of the Popotosa Formation is defined by a prominent angular unconformity along the western margin of the Socorro Basin and Belen sub-basin. This unconformity probably becomes conformable near basin depocenters (Cather et al., 1994). The base of the Popotosa Formation is constrained by the 16.2±1.5 Ma Silver Creek andesite (Cather et al., 1994) in the Socorro area; however, the Popotosa is as old as 25.9±1.2 Ma unit of Arroyo Montosa in the Abbe Springs basin to the west (Osburn and Chapin, 1983). The upper age of the Popotosa Formation is constrained by a unit of the Socorro Peak Rhyolite (rhyolite of Grefco quarry; Chamberlin, 1980, 1999), about 6 km southwest of Socorro, which has been dated at 7.85±0.03 Ma (Newell, 1997, p. 13, 27). This flow is interbedded with piedmont and fluviolacustrine facies (Chamberlin, 1999). The piedmont facies at the Grefco locality contains abundant reddish-brown sandstone clasts derived from the Abo Formation, exposed along the eastern margin of the Socorro Basin (Chamberlin, 2000, oral commun.), indicating that the fluviolacustrine facies extended west of the Grefco locality by 7.9 Ma. The youngest constraint is from the 6.88±0.02 Ma (McIntosh and Chamberlin, unpubl. $^{40}Ar/^{39}Ar$ date) trachyandesite of Sedillo Hill (Chamberlin, oral commun., 2000; Osburn and Chapin, 1983), which overlies playa lake sediments (Chamberlin, 1980), about 20 km west of Socorro, New Mexico. Late Miocene (Hemphillian and possible Clarendonian) mammal fossils are recognized in the upper part of the fluviolacustrine facies in the Gabaldon badlands in the western Belen sub-basin (Lozinsky and Tedford, 1991). Deposition of the Popotosa Formation began after about 25 Ma in the Abbe Springs basin, west of Socorro, and about 15 Ma in the Socorro area (Cather et al., 1994; Osburn and Chapin, 1983). Popotosa deposition probably ended between 5-7 Ma in the northern Socorro Basin, as constrained by dates from the Socorro area. The ancestral Rio Grande began to flow through the Socorro area and into the Engle and Palomas basins by 4.5-5 Ma (Mack et al., 1996, 1993; Leeder et al., 1996).

The Popotosa Formation is temporally equivalent to the Hayner Ranch and Rincon Valley formations in the Palomas and Mesilla basins of southern New Mexico (Seager et al., 1971) and the Tesuque Formation in the Española Basin (Spiegel and Baldwin, 1963; Galusha and Blick, 1970). The Popotosa Formation is similar in age to the Zia Formation and lower part of the Arroyo Ojito Formation. The northern extent of Popotosa-equivalent fluviolacustrine mudstone extends north to near the Calabacillas-Belen sub-basin boundary (Lozinsky, 1994). Estimates of stratal accumulation (not adjusted for compaction) on the Popotosa Formation is about 600 m/m.y for the Gabaldon badlands area (Lozinsky, 1994).

Upper Santa Fe Group

Deposits of the upper Santa Fe Group are areally extensive and typically bury deformed and better cemented rocks of the lower Santa Fe Group. Upper sub-group sediments record fluvial deposition of streams and rivers through externally drained basins (Hawley, 1978). During this time, the Albuquerque Basin was a large contributory basin (Lozinsky and Hawley, 1991) where western margin tributaries merged with the ancestral Rio Grande axial-fluvial system near San Acacia, New Mexico. The ancestral Rio Grande formed a narrow (axial) trunk river in the Socorro Basin. This trunk river flowed south, near Hatch, New Mexico, where it formed a broad fluvial braid plain that was constructed during periodic avulsions into adjacent basins (Hawley et al., 1969, 1976; Mack et al., 1997; Lozinsky and Hawley, 1991).

The upper Santa Fe Group can be divided into three major lithofacies assemblages in the Albuquerque Basin, reflecting differences in deposit texture, provenance, and paleoenvironment. These lithofacies assemblages are referred to here as the western-fluvial, axial-river, and piedmont lithofacies.

Western-fluvial deposits are predominantly extrabasinal and contain locally abundant red granite, sandstone, and chert. These deposits were derived from large rivers and streams developed on the western margin of the basin. Axial-river deposits refer to detritus laid down by the ancestral Rio Grande. Composition of the fluvial facies is predominantly extrabasinal and contains a mixed assemblage of clast types (Lozinsky et al., 1991). Piedmont facies are present along the flanks of the basin, on the footwalls of major rift-margin uplifts, and contain locally derived detritus from nearby rift-border drainages.

Deposits of the upper Santa Fe Group typically have few concretionary or well cemented intervals, except locally along faults or near piedmont/axial-fluvial boundaries. Bedding is generally more lenticular than the tabular beds of the Zia Formation. Poikilotopic calcite and concretionary sandstone, common in the Zia Formation (Beckner and Mozley, 1998), are rare in stratigraphically higher deposits. Buried soils are also typically more common in the upper Santa Fe Group, and locally can be quite common and widespread near the top of the section. Upper Santa Fe Group sediments are divided into the Sierra Ladrones Formation, Cochiti Formation, Arroyo Ojito Formation, Tuerto Formation, the gravel of Lookout Park, and a number of smaller local units exposed along the structural margins of the basin.

Axial-fluvial and piedmont deposits comprise the Sierra Ladrones Formation (Machette, 1978a), which has been extended throughout much of the Albuquerque Basin (Lucas et al., 1993; Cather et al., 1994; Smith and Kuhle, 1998a; Connell and Wells, 1999). The axial-fluvial facies form a relatively narrow belt between the western fluvial and piedmont lithofacies. Piedmont deposits interfinger with western and axial-fluvial deposits near the basin margins (Machette, 1978a; Connell and Wells, 1999; Maldonado et al., 1999).

The western-fluvial lithofacies contain sandstone, conglomerate, and mudstone that were deposited by streams draining the eastern Colorado Plateau, southeastern San Juan Basin, and the Sierra Nacimiento. These western fluvial deposits comprise the Arroyo Ojito Formation (Connell et al., 1999) and stratigraphically similar facies to the south (Love and Young, 1983; and Lozinsky and Tedford, 1991). This lithofacies represents fluvial deposition of ancestral Rio Puerco, Rio Salado, Rio San Jose, and Rio Guadalupe/Jemez fluvial systems. Western fluvial lithofacies interfinger with axial-fluvial deposits of the ancestral Rio Grande near the present Rio Grande Valley (Lozinsky et al., 1991).

Western-fluvial lithofacies generally contain greater amounts of quartz than in the axial-fluvial lithofacies, which is commonly contains more volcanic detritus (Gillentine, 1996). The quartzose nature of the western-fluvial deposits indicates compositional maturity of the sandstone fraction (Large and Ingersoll, 1997), and may indicate derivation from a stable source; probably Cretaceous sediments exposed on the adjacent Colorado Plateau (Gillentine, 1996).

The Cochiti Formation interfingers with western fluvial deposits, but is composed almost entirely of volcaniclastic sediments derived from the southern Jemez Mountains.

The Sierra Ladrones Formation is herein restricted to fluvial deposits associated with the ancestral Rio Grande fluvial system and interfingering footwall-derived piedmont deposits. The Arroyo Ojito Formation is herein expanded to represent fluvial deposits derived from drainages of the western margin. The Arroyo Ojito Formation represents the most areally extensive lithofacies of the upper Santa Fe Group and can be subdivided into at least three mappable members near the northwestern margin of the Calabacillas sub-basin (Connell et al., 1999).

Relatively thin, locally derived piedmont gravels are locally preserved on hanging wall hinges and structural re-entrants in the basin. The Tuerto Formation is a volcanic-bearing gravel derived from the Ortiz Mountains and is found in the Hagan embayment. Another such deposit is the gravel of Lookout Park (Smith and Kuhle, 1998a, b), which is derived from volcanic rocks of the southeastern flank of the Jemez Mountains.

Sierra Ladrones Formation

The Sierra Ladrones Formation was defined by Machette (1978a) for slightly deformed, coarse-grained interfingering fluvial and basin-margin piedmont deposits that unconformably overlie the Popotosa Formation in the northern Socorro Basin and Belen sub-basin. No type section was measured. A composite type area was proposed on the San Acacia quadrangle, which was designated as representative of western-margin piedmont, central axial-fluvial, and eastern-margin piedmont facies tracts (Machette, 1978a); however, no stratigraphic sections were described for this widely mapped unit (Connell et al., 2001). The Sierra Ladrones Formation was deposited by a through-flowing river that marks the end of internal basin drainage represented by the Popotosa Formation. Thickness of the Sierra Ladrones Formation is greater than 470 m (estimate from cross section, Machette, 1978a) at its type area, but is over 1 km thick beneath Albuquerque (Connell et al., 1998a; Hawley, 1996). Fluvial deposits are typically light-gray to light yellowish-brown, non-cemented to locally cemented, moderately sorted, trough cross stratified sand and gravel with rare muddy interbeds that are commonly found as rip-up clasts and mud balls. Sandy and gravelly deposits typically form multilateral channels. The lack of preservation of mud suggests deposition by anastomosing or braided rivers. Piedmont deposits of the Sierra Ladrones Formation are typically better cemented and more poorly sorted than fluvial deposits. Piedmont deposits are typically light-brown to reddish-brown in color and tend to form a rather narrow belt against footwall uplands; however, the uppermost part of the piedmont facies prograded basinward by 5-10 km (up to 20 km west of the Manzano Mountains) during early Pleistocene time. Conglomeratic beds of the axial-fluvial lithofacies typically consist of well sorted, well rounded quartzite with subordinate, subrounded to subangular volcanic, hypabyssal intrusive, granite, chert, and basalt. The Pedernal chert, a locally common constituent of the Arroyo Ojito Formation, is quite rare (<1%) and is typically better rounded than in the Arroyo Ojito Formation. Piedmont lithofacies typically contain variable amounts of subangular to subrounded granite, limestone, sandstone, and metamorphic rocks derived from basin-margin drainages.

Previous workers (Debrine et al., 1966; Evans, 1966) mapped an axial-fluvial facies of the ancestral Rio Grande near Socorro, New Mexico. They traced it along the eastern margin of the Rio Grande valley to just east of San Acacia, New Mexico. A narrow, south-trending belt of axial-fluvial deposits were delineated just east of San Acacia (Cather, 1996). These fluvial deposits can be traced into Arroyo de la Parida, about 8 km northeast of Socorro, where a medial Blancan (2.7-3.7 Ma; Morgan et al., 2000) fossil assemblage is recognized in an exposed fluvial succession originally assigned to the Palomas Formation (Palomas gravels of Gordon, 1910). Machette (1978) mapped a nearly continuous, south-trending belt of axial-fluvial deposits west of San Acacia and on the footwall of the Loma Blanca fault, along the western margin of the Belen sub-basin. Interfingering piedmont deposits were assigned to the Sierra Ladrones Formation by Machette (1978a), who considered these to be derived from the eastern and western margins of the basin. The presence of basin-margin, piedmont-slope facies between two “axial-fluvial” facies indicates: 1) fluvial deposits are of different ages; 2) Machette’s (1978) eastern-margin piedmont facies (unit Tsp of Machette, 1978a) has a different origin; or 3) axial-fluvial deposits exposed near the western border was a large western-margin tributary to the Rio Grande. Paleocurrent observations and gravel composition determined from exposures just north of the Rio Salado and Rio Grande confluence indicate southeast-directed flow (Connell et al., 2001) from a volcanic-rich source area, such as the ancestral Rio Salado, which originates in volcanic rocks of the Bear Mountains. Gravel composition and paleocurrent observations indicate a western source and suggest that Machette’s (1978a) eastern-margin piedmont deposit may be part of the western-fluvial systems tract and should be reassigned to the Arroyo Ojito Formation.

Lozinsky and Tedford (1991) extended the Sierra Ladrones Formation northward into the Gabaldon badlands. They recognized that these deposits are related to fluvial systems that originated along the western margin of the basin, rather than from an ancestral Rio Grande. Paleocurrent measurements and gravel composition indicates that these deposits contain were derived from the western margin of the basin (Lozinsky and Tedford, 1991). Thus, these deposits are assigned to the Arroyo Ojito Formation.

The Sierra Ladrones Formation is broadly equivalent to the Plio-Pleistocene Camp Rice and Palomas formations (Gile et al., 1981; Lozinsky and Hawley, 1986), which record deposition of an ancestral Rio Grande beginning by around 4.5-5 Ma (Mack et al., 1993, 1996; Leeder et al., 1996). The earliest definitive evidence for an ancestral axial river the southern part of the basin is the presence of southward-directed cross-bedded fluvial sandstone underlying the 3.73±0.1 Ma basalt of Socorro Canyon, just south of Socorro, New Mexico. (R.M. Chamberlin and W.C. McIntosh, written commun., 2000). The Pliocene trachyandesite at San Acacia overlies piedmont deposits derived from the eastern basin margin (Machette, 1978a). This flow yielded a K/Ar date of 4.5±0.1 (Bachman and Mehnert, 1978), but has been dated at 4.87±0.04 Ma using the $^{40}Ar/^{39}Ar$ method (R.M. Chamberlin and W.C. McIntosh, 2000, oral communication). The presence of these basin-margin deposits only constrains the location, but not age of an ancestral axial river at the boundary of the Socorro and Albuquerque basins.

Piedmont deposits beneath the San Acacia flow contain abundant granite clasts with lesser amounts of volcanic and sedimentary detritus. The composition of piedmont deposits underlying this early Pliocene flow is contrast to the volcanic-dominated conglomerate of the Popotosa Formation mapped to the east (Cather, 1996). The presence of granite and sedimentary detritus supports Machette’s (1978a) assignment of these deposits to the Sierra Ladrones Formation, which locally constrains the age of the unconformity between the Sierra Ladrones and Popotosa formations to being older than 4.9 Ma near San Acacia. Cross-bedded fluvial sand is present near Arroyo de la Parida, which contain fossils that are indicative a medial Blancan age of about 3.6-2.7 Ma for the upper exposed part of the fluvial section there (Morgan et al., 2000).

Precise estimates of the age of the Sierra Ladrones Formation in the Belen sub-basin are problematic, principally because of the unconformable relationships with the youngest Popotosa Formation playa-lake beds at about 7-8 Ma. The oldest Sierra Ladrones piedmont deposits are older than about 4.87 Ma. Ancestral Rio Grande deposits are older than about 3.7 Ma and reports of axial-fluvial deposits entering southern New Mexico between 4.5-5 Ma suggest that the ancestral Rio Grande was flowing through the Socorro area by 4.5-5 Ma. Thus, deposition of the Sierra Ladrones Formation probably began sometime between 7-4.5 Ma.

The age of the uppermost Sierra Ladrones Formation is constrained by fallout ash from the upper Bandelier Tuff (Tshirege Member), and fluvially transported clasts of the lower Bandelier Tuff (Connell et al., 1995; Connell and Wells, 1999), early Irvingtonian (*ca*. 1.6-1.2 Ma) fossils (Lucas et al., 1993), and fallout ash from the 0.6-0.66 Ma Lava Creek B ash within inset fluvial and piedmont deposits in the Santo Domingo sub-basin (Smith and Kuhle, 1998b) and Calabacillas sub-basin (N. Dunbar, 2000, oral commun.). Thus, Sierra Ladrones Formation deposition ended between 1.3-0.6 Ma in the Albuquerque Basin. In the Socorro Basin, entrenchment of the ancestral Rio Grande began after emplacement of pumice flood deposits and fallout of the Bandelier Tuff events (Cather, 1988), which is now considered part of the upper Santa Fe Group basin-fill succession (S.M. Cather, oral commun., 2000).

Arroyo Ojito Formation

The Arroyo Ojito Formation (Connell et al., 1999) was proposed for fluvial sediments along the western margin of the Albuquerque Basin that were derived from the eastern Colorado Plateau, Sierra Nacimiento, and southern Jemez Mountains. The Arroyo Ojito Formation contains a rather diverse assemblage of volcanic, sedimentary, and plutonic clasts that can be differentiated from relatively monolithologic (i.e., volcanic) Cochiti Formation of Smith and Lavine (1996). The Arroyo Ojito Formation supercedes Manley's (1978) Cochiti Formation (Connell et al., 1999). Conglomeratic parts of the Arroyo Ojito Formation commonly contain angular to subrounded red granite, basalt, sandstone, conglomerate, and angular to subangular cobbles of the Pedernal chert, and thus differ from the redefined volcaniclastic Cochiti Formation of Smith and Lavine (1996). Gravelly beds of the Arroyo Ojito Formation, especially the Ceja Member, are distinctive because they contain locally abundant subangular red granite and Pedernal chert cobbles. Gravel beds are also poorly sorted and have a bimodal distribution of gravel, typically containing abundant pebbles and small cobbles with about 10-25% of scattered large cobbles and small boulders. The Pedernal chert of Church and Hack (1939) is a black and white chalcedony and chert of the middle member of the Abiquiu Formation (Moore, 2000). The Pedernal chert is exposed at the northern end of the Sierra Nacimiento (Woodward, 1987). It commonly forms subangular to angular blocks in gravelly beds of the upper part of the Arroyo Ojito Formation. The Pedernal chert is rarely found in ancestral Rio Grande sediments, where it is better rounded than in the Arroyo Ojito Formation.

The Arroyo Ojito Formation is 437 m thick at the type section, where it is subdivided into three members (Connell et al., 1999). The Navajo Draw Member is the lowest unit of the Arroyo Ojito Formation and overlies the Cerro Conejo Member of the Zia Formation with a fairly sharp and contact along the Ceja del Rio Puerco (Fig. 1). This contact, however, is gradational and interfingers with the Zia Formation to the east (Koning and Personius, *in review;* Connell et al., 1999).

The Navajo Draw Member is about 230 m in thickness and overlies the Cerro Conejo Member. The Navajo Draw Member marks a significant change from the mixed eolian and sand-dominated fluvial system of the Zia Formation to a more mud-gravel dominated fluvial deposition of the Arroyo Ojito Formation. This lower member is a very pale-brown to pale-yellow, lenticular, poorly to moderately sorted, fine- to coarse-grained sand and pebbly sand with minor thin to medium bedded pale-yellow mud. Gravelly beds are commonly clast supported and contain volcanic (mostly intermediate composition) pebbles and subordinate sandstone and brownish-yellow fine chert pebbles, and rare red granite and Pedernal chert clasts derived from southeast-flowing streams (Connell et al., 1999). The Navajo Draw Member is conformably overlain by the Loma Barbon Member of the Arroyo Ojito Formation, which contains fall-out lapilli and ash from the Peralta Tuff (6.8-7.3; Connell et al., 1999; Koning and Personius, *in review*).

The Loma Barbon Member is the middle unit of the Arroyo Ojito Formation and contains about 200 m of reddish-yellow to strong-brown and yellowish-brown, poorly sorted, sand, pebbly sand, and gravel at its type area. The Loma Barbon Member contains locally abundant subangular to subrounded pebbles and cobbles of red granite that is probably derived from the Sierra Nacimiento. Clast composition becomes increasingly heterolithic up section. Pedernal chert clasts also increase in abundance (Connell et al., 1999). The Loma Barbon Member is redder than the underlying Navajo Draw Member. This dominantly reddish-brown color may be the result of recycling of sandstone and mudstone of the Permo-Triassic section exposed along the flanks of Sierra Nacimiento (Woodward, 1987). A number of fallout tephra correlative to the Peralta Tuff Member (6.8-7.3 Ma, Connell et al., 1999; Koning and Personius, *in review*) are present near the middle of the unit. Rhyodacitic clasts in gravel beds having southeasterly paleoflow directions yielded dates of $^{40}Ar/^{39}Ar$ dates of 3.79-4.59 Ma (Connell, 1998), suggesting derivation from the Tschioma Formation

(Polvadera Group). Soister (1952) recognized similar deposits beneath 2.5±0.3 Ma (Bachman and Mehnert, 1978) basalt flows of Santa Ana Mesa. These deposits are likely correlative to the Loma Barbon Member. Axial-fluvial deposits of the uppermost Sierra Ladrones Formation overlie the Loma Barbon Member and similar deposits (Cather and Connell, 1998; Connell, 1998). Field relationships suggest that the Ceja Member pinches out to the east into the Loma Barbon Member near Rio Rancho and Bernalillo, New Mexico. (Connell et al., 1998; Personius et al., 2000).

The Ceja Member (Kelley, 1977) is the uppermost member of the Arroyo Ojito Formation (Connell et al., 1999). Kelley (1977) applied the term Ceja Member to Lambert's (1968, p. 271-274) upper buff member type section at El Rincon in an attempt to replace the uppermost part of the upper buff member of Bryan and McCann (1937) and Wright (1946). Later workers (Tedford, 1982; Lucas et al., 1993) restricted the Ceja Member to upper Santa Fe Group sediments derived from the western basin margin. The Ceja Member is 64 m at the type section at El Rincon (Kelley, 1977) where is forms an areally extensive pebble to small boulder conglomerate and conglomeratic sandstone beneath the Llano de Albuquerque.

The Ceja Member is poorly sorted and has a bimodal gravel distribution with abundant pebbles and scattered cobbles and boulders. The Ceja Member unconformably overlies the Navajo Draw Member on the footwall of the San Ysidro fault, but appears to conformable to the south and east. Streams of the Ceja Member were part of Bryan and McCann's (1937, 1938) Rio Chacra fluvial system, a progenitor to the Rio Puerco. Conglomeratic deposits contain rounded sandstone and sparse quartzite-bearing conglomerate that were probably recycled from older Santa Fe Group and Galisteo Formation exposed along the basin margin. The Ceja Member grades finer and thinner to the south and east, (see Maldonado et al., 1999), but retains its bimodal cobbly to bouldery character. This southward thinning and slight fining suggests that the Ceja Member may pinch out to the south-southeast, near Belen and Los Lunas; however a gravel commonly underlies the Llano de Albuquerque. Cobbles of Pedernal chert are locally common in this member. Paleocurrent observations indicate deposition by southeast-flowing streams, suggesting that the source of recycled Pedernal chert was from the Colorado Plateau, San Juan Basin, and western side of the Sierra Nacimiento. The presence of Pedernal chert (Abiquiu Formation) west of the Sierra Nacimiento is supported by the presence of Pedernal chert clasts in the southern San Juan Basin (Love, 1997); however, Miocene recycling of the Pedernal chert could have also occurred. The Ceja Member and similar deposits contain Blancan vertebrate fossils (Lucas et al., 1993; Morgan and Lucas, 1999, 2000; Wright, 1946). The Ceja Member is interbedded with 3.00±0.01 and 4.01±0.16 Ma basalt flows (Maldonado et al., 1999).

In the Belen sub-basin, fluvially transported bivalves (*Pycnodonte* and/or *Exogyra*) from the Cretaceous Dakota Formation-Mancos Shale (Greenhorn Limestone) interval are found beneath the Llano de Albuquerque, south of Los Lunas present (S.G. Lucas, written commun., 1999). Western fluvial deposits exposed beneath the southern end of the Llano de Albuquerque also contain recycled rounded obsidian clasts that were derived from the 2.8-3.3 Ma East Grants Ridge obsidian (Love and Young, 1983). Love and Young (1983) and Wright (1946) also discuss deposition by large streams draining the western margin of the basin.

Near the southern end of the Belen sub-basin, Denny (1940) and Morgan and Lucas (2000) reported Blancan fossils in Machette's (1978b) eastern margin piedmont deposits, exposed west of the Rio Grande valley and just north of the confluence with the Rio Salado (Fig. 1., lj).

Cochiti Formation

The Cochiti Formation was originally mapped and defined (Bailey et al., 1969; Smith et al., 1970) for a succession of volcanic gravel and sand derived from erosion of the Keres Group in the southern Jemez Mountains. The application of this term to subsequent geologic and stratigraphic studies has created varied and contradictory interpretations (*cf.* Manley, 1978; Smith and Lavine, 1996; Goff et al., 1990; Chamberlin et al., 1999). These wide-ranging interpretations principally arise from complications in reconciling the volcanic stratigraphy of the Jemez Mountains with the basin-fill stratigraphy of the Santa Fe Group (Smith and Lavine, 1996). The Cochiti Formation was redefined to include sedimentary strata of entirely volcanic composition that overlie Keres Group volcanic rocks and their correlative sedimentary strata south of the Jemez Mountains (Smith and Lavine, 1996). Deposition of the Cochiti Formation is partly time equivalent to the upper Arroyo Ojito Formation (Loma Barbon and Ceja members) and can be differentiated by the relative abundance of nonvolcanic clast constituents. The Cochiti Formation is very thin northwest of Santa Ana Mesa (Chamberlin et al., 1999), but thickens to about 600 m along the southeastern flank of the Jemez Mountains, in Peralta Canyon (Smith and Kuhle, 1998a, b).

The age of the Cochiti Formation is constrained by the a 6.75 Ma pyroclastic bed of the Peralta Tuff, which underlies the base at Tent Rocks, in Peralta Canyon, (Smith and Kuhle, 1998c; Smith et al., 2001). The upper Cochiti Formation interfingers with upper Pliocene basalts of Santa Ana Mesa and the lower Bandelier Tuff (Smith et al., 2001). The Plio-Pleistocene gravel of Lookout Park insets the Cochiti Formation. The Cochiti Formation records deposition of volcanic-bearing stream and piedmont sediments from about 6.8 to 1.6 Ma.

Plio-Pleistocene basin-margin deposits

A number of relatively thin conglomeratic and gravelly deposits are recognized along the faulted borders of the basin. These deposits commonly have strongly developed petrocalcic soils with Stage III to V carbonate morphology and are preserved on the footwalls of basin margin or major intrabasinal faults near basin margins (Connell and Wells, 1999; Maldonado et al., 1999).

The Tuerto Formation (gravel) was informally named for a 20-30 m thick, subhorizontal deposit of volcanic- and subvolcanic-bearing conglomerate and sandstone unconformably resting on slightly to moderately tilted older Santa Fe Group deposits (Stearns, 1953). The Tuerto Formation can easily be differentiated from underlying Santa Fe Group deposits by an abundance (about 10-25%) of green, black, and yellow hornfels (Cather et al., 2000), which are interpreted as thermally metamorphosed Mesozoic and Paleogene strata exposed along the flanks of the Ortiz Mountains (S. Maynard, 2000, oral commun.). The Tuerto Formation contain rare fine pebbles of granite, and are thus easily differentiated from the granite-bearing Ancha Formation (Spiegel and Baldwin, 1963). The basalts of Cerros del Rio (mostly emplaced between 2.5-2.8 Ma; Woldegabriel et al., 1996; Bachman and Mehnert, 1978) interfinger with the lower part of the Tuerto Formation (Stearns, 1979). The upper boundary is constrained by correlation of the upper constructional surface (Ortiz surface of Stearns, 1953) to the Plains surface formed on the Ancha Formation near Santa Fe (Spiegel and Baldwin, 1963). The top of the Ancha Formation is constrained by primary fallout ash and lapilli correlated to one of the Cerro Toledo Rhyolite tephras (*ca*. 1.48 Ma) and the presence of an ash correlated to the upper Bandelier Tuff. This ash is in deposits that are interpreted to be inset against the Ancha Formation (Koning and Hallett, 2000). Based on correlations to the Ancha Formation, the Tuerto Formation was deposited prior to 2.6 Ma. Deposition probably ceased between 1.2-1.5 Ma, however, the presence of weakly to moderately developed calcic soils (Stage II to III carbonate morphology) in the Tuerto Formation in the Hagan embayment, suggests that deposition of the Tuerto Formation may have continued into the middle Pleistocene.

The gravel of Lookout Park is an informal unit recognized along the southeastern flank of the Jemez Mountains (Smith and Kuhle, 1998a, b). This gravel unconformably overlies the Cochiti Formation, is inset against upper Pliocene basalts of Santa Ana Mesa, and is unconformably overlain by the lower member of the Bandelier Tuff. Thus, the gravel of Lookout Park was deposited between about 2.4-1.6 Ma.

Post-Santa Fe Group Deposits

The upper boundary of the Santa Fe Group of Spiegel and Baldwin (1963, p. 39) is "considered to include all but the terrace alluvium of present valleys." Most workers agree that the end of Santa Fe Group deposition occurred when the ancestral Rio Grande and major tributaries began to incise into older basin fill (Hawley et al., 1969; Gile et al., 1981; Wells et al., 1987). This definition is allostratigraphic in nature and has no strong lithologic basis, making it difficult to apply in the basin (Connell et al., 2000). Delineation of strata that post-date Santa Fe Group aggradation is ambiguous in such deposits because of lithological similarities to the underlying Santa Fe Group. Post-Santa Fe Group valley floor and piedmont deposits commonly form stepped valley border landforms inset against the Santa Fe Group. These deposits were laid down during periods of aggradation that were punctuated by climate-driven episodes of entrenchment by the ancestral Rio Grande and major tributaries (Hawley, 1978; Gile et al., 1981; Wells et al., 1987). Differentiation of post-Santa Fe Group deposits is thus locally ambiguous because the size and character of drainage basins influence entrenchment. This geomorphic-stratigraphic ambiguity is best expressed along the Manzano and Manzanita Mountains where low-order mountain-front drainages are not commonly graded to entrenched surfaces associated with the Rio Grande fluvial system. Unlike the larger drainages of Tijeras Arroyo, Hell Canyon Wash, and Abo Arroyo, streams on the western flank of the Manzanita and Manzano Mountains commonly terminate on the Llano de Manzano of Machette (1985), a broad abandoned basin-floor and piedmont slope east of the Rio Grande Valley. The Llano de Manzano forms a weakly dissected landscape (Pazzaglia and Wells, 1990; Connell and Wells, 1999) that makes differentiation of post-Santa Fe Group deposits difficult. The interaction of intrabasinal faults and competence of tributary streams both likely play a local role in defining when Santa Fe Group deposition ceased (Connell et al., 2000).

Entrenchment of the Santa Fe Group would result in a steady decline in groundwater levels as the Rio Grande and its major tributaries incise into the basin fill. Thus, deposits representing widespread basin aggradation should be relatively poorly drained with respect to their entrenched and better-drained counterparts. Such relationships are recognized in Hell Canyon Wash, where early Pleistocene pumice-bearing deposits of the ancestral Rio Grande are well cemented with sparry calcite, suggesting deposition during high groundwater. Incised deposits, however, are not well cemented and contain disseminated or micritic calcium-carbonate cements.

Pliocene-Pleistocene tectonic activity is recognized by the deposition of syntectonic depositional wedges (Smythe and Connell, 1999; colluvial wedges of Machette, 1978b) along the hanging walls of major intrabasinal normal faults.

Delineation of a single regionally correlative surface of aggradation that marks the end of Santa Fe Group deposition is problematic and should be abandoned in favor of a definition that allows for the development of multiple local tops that are diachronous. Studies of White Rock Canyon at the northern end of the Santo Domingo sub-basin indicate that the Rio Grande excavated very deep valleys into basalt of the upper Pliocene Cerros del Rio volcanic field (Reneau and Dethier, 1996). The Bandelier Tuff locally buried these deep valleys. Much of the basalt exposed along White Rock Canyon were deposited in a short time mostly between 2.8-2.3 Ma: Woldegabriel et al., 1996), resulting in the development of a constructional lava pile near the La Bajada and Pajarito faults. Evidence for a regional late Pliocene unconformity in the Española Basin in White Rock Canyon is clear; however, incision of the Rio Grande into these basalt flows (Dethier, 1999) might be a local effect caused by the river's effort to maintain a graded profile through White Rock Canyon, rather than the result of some regional unconformity.

A number of early Pleistocene constructional surfaces that locally mark the top of the Santa Fe Group are recognized south of White Rock Canyon. The early Pleistocene Sunport and Llano de Albuquerque surfaces (Albuquerque Basin), the Las Cañas surface (Socorro Basin), and the lower La Mesa surfaces (Mesilla Basin) are rather broad constructional surfaces that have clearly been entrenched by younger fluvial deposits associated with development of the Rio Grande valley. Magnetostratigraphic studies of the Camp Rice Formation in southern New Mexico, a correlative of the Sierra Ladrones Formation, indicates that widespread basin-fill deposition was mostly uninterrupted during Pliocene and early Pleistocene times (Mack et al., 1993).

West of the Rio Grande, in the Santo Domingo sub-basin, the Bandelier Tuff rests disconformably on the gravel of Lookout Park, which sits with angular unconformity on the Sierra Ladrones and Cochiti formations. Down dip and to the east, the Bandelier Tuff and a Pliocene basalt flow are part of a conformable Santa Fe Group succession on the eastern side of the Rio Grande (Smith et al., 2001; Smith and Kuhle, 1998c). Similar stratigraphic relationships are also recognized near San Felipe Pueblo, where a similarly aged conformable Santa Fe Group succession is interbedded with basalts of Santa Ana Mesa and a 1.57 Ma ash correlated to the Cerro Toledo Rhyolite (N. Dunbar, 2001, written commun; Cather and Connell, 1998).

At Tijeras Arroyo, biostratigraphic data suggest the presence of a disconformity in the section between the Arroyo Ojito Formation and overlying Bandelier-pumice-bearing fluvial deposits of the Sierra Ladrones Formation (Connell et al., 2000; Lucas et al., 1993). Biostratigraphic data (Morgan and Lucas, 1999, 2000) indicate a lack of late Blancan fossils (i.e., lack of fossils recording the Great American Interchange) in the Albuquerque Basin and suggest a hiatus in deposition occurred during late Blancan time. The Llano de Albuquerque is older than 1.2 Ma (Connell et al., 2000) and perhaps is late Pliocene in age. The probable Pliocene age of the areally extensive Llano de Albuquerque west of the Rio Grande and burial by Pleistocene deposits of the ancestral Rio Grande to the east may account for the apparent lack of late Blancan fossils, which could be buried by the younger Bandelier-pumice bearing deposits of the ancestral Rio Grande.

Another possible explanation for the lack of representative late Blancan fossils may be due to a reduction in sedimentation rate or hiatus in deposition. The disconformity at Tijeras Arroyo may be due to earlier entrenchment of the ancestral Rio Puerco fluvial system along the western margin of the basin. With cessation of Arroyo Ojito deposition along the eastern part of the basin, local unconformities would develop between the abandoned basin floor constructional surface of the Llano de Albuquerque, and continued deposition of the Sierra Ladrones Formation into the early Pleistocene. The upper boundary of the Santa Fe Group thus is time transgressive and sensitive to the competence of streams, availability of sediments, and the activity of faults (Connell et al., 2000).

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SUMMARY OF BLANCAN AND IRVINGTONIAN (PLIOCENE AND EARLY PLEISTOCENE) MAMMALIAN BIOCHRONOLOGY OF NEW MEXICO

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Significant mammalian faunas of Pliocene (latest Hemphillian and Blancan) and early Pleistocene (early and medial Irvingtonian) age are known from the Rio Grande and Gila River valleys of New Mexico. Fossiliferous exposures of the Santa Fe Group in the Rio Grande Valley, extending from the Española basin in northern New Mexico to the Mesilla basin in southernmost New Mexico, have produced 21 Blancan and six Irvingtonian vertebrate assemblages (Fig. 1). A medial Irvingtonian fauna is known from a cave deposit in the San Luis basin in northernmost New Mexico (Fig. 2). Three Blancan faunas occur in Gila Group strata in the Gila River Valley in the Mangas and Duncan basins in southwestern New Mexico (Fig. 3). More than half of these faunas contain five or more species of mammals, and many have associated radioisotopic dates and/or magnetostratigraphy, allowing for correlation with the North American land-mammal biochronology (Figs. 2-3).

Two diverse early Blancan (4.5-3.6 Ma) faunas are known from New Mexico, the Truth or Consequences Local Fauna (LF) from the Palomas basin and the Buckhorn LF from the Mangas basin. The Truth or Consequences LF contains five species of mammals indicative of the early Blancan: *Borophagus* cf. *B. hilli*, *Notolagus lepusculus*, *Neotoma quadriplicata*, *Jacobsomys* sp., and *Odocoileus brachyodontus*. Associated magnetostratigraphic data suggest correlation with either the Nunivak or Cochiti subchrons of the Gilbert Chron (between 4.6 and 4.2 Ma), which is consistent with the early Blancan age indicated by the mammalian biochronology. The Truth or Consequences LF is similar in age to the Verde LF from Arizona, and slightly older than the Rexroad 3 and Fox Canyon faunas from Kansas. The Buckhorn LF has 18 species of mammals, including two rodents typical of the early Blancan, *Mimomys poaphagus* and *Repomys panacaensis*. The Buckhorn LF also is similar in age to the Verde LF and has affinities with the Panaca LF from Nevada. Although the Buckhorn and Truth or Consequences LFs have few taxa in common, the similarities of both faunas with the Verde LF suggest they are close in age.

Eight faunas from the central and southern Rio Grande Valley are medial Blancan in age (3.6-2.7 Ma), including the Pajarito and Belen faunas from the Albuquerque basin, the Arroyo de la Parida LF from the Socorro basin, the Cuchillo Negro Creek and Elephant Butte Lake LFs from the Engle basin, the Palomas Creek LF from the Palomas basin, the Hatch LF from the Hatch-Rincon basin, and the Tonuco Mountain LF from the Jornada basin. These faunas are characterized by the presence of taxa absent from early Blancan faunas, including *Geomys* (*Nerterogeomys*) *paenebursarius*, *Equus cumminsii*, *E. scotti*, and *Camelops*, and the absence of South American immigrant mammals found in late Blancan faunas. The Pajarito LF is directly associated with a fluvially recycled pumice dated at 3.12±0.10 Ma (Maldonado et al., 1999). The Cuchillo Negro Creek and Elephant Butte Lake LFs are in close stratigraphic association with a basalt flow dated at 2.9 Ma. Magnetostratigraphy constrains the age of the Tonuco Mountain LF between 3.6 and 3.0 Ma.

The Mesilla A fauna from the Mesilla basin and the Pearson Mesa LF from the Duncan basin are late Blancan in age (2.7-2.2 Ma). Both faunas record the association of *Nannippus* with a South American immigrant, *Glyptotherium* from Mesilla A and *Glossotherium* from Pearson Mesa, restricting their age to the interval after the beginning of the Great American Interchange at about 2.7 Ma and before the extinction of *Nannippus* at about 2.2 Ma. Magnetostratigraphy further constrains the Mesilla A and Pearson Mesa faunas to the upper Gauss Chron, just prior to the Gauss/Matuyama boundary at 2.58 Ma. The Mesilla B and Virden faunas occur higher in the same stratigraphic sequences as the Mesilla A and Pearson Mesa faunas, respectively, and are latest Blancan in age (2.2-1.8 Ma). Both faunas contain taxa restricted to the Blancan, including the camels *Blancocamelus* and *Gigantocamelus* from Mesilla B, and *Canis lepophagus* from Virden. The absence of *Nannippus*, and of *Mammuthus* and other genera that first appear in the Irvingtonian, suggest an age range between 2.2 and 1.8 Ma. Magnetostratigraphic data from Mesilla B support a latest Blancan age.

The Tijeras Arroyo fauna from the Albuquerque basin and the Tortugas Mountain and Mesilla C faunas from the Mesilla basin all include *Mammuthus* and other mammals indicative of an early Irvingtonian age (1.8-1.0 Ma). The association of *Mammuthus* and *Stegomastodon* in the Tortugas Mountain LF indicates an age younger than 1.8 Ma, after the arrival of *Mammuthus* in North America from Eurasia and before the extinction of *Stegomastodon* at about 1.2 Ma. The co-occurrence of *Glyptotherium arizonae*, *Equus scotti*, and the primitive mammoth *M. meridionalis* in Tijeras Arroyo and Mesilla C is typical of southwestern early Irvingtonian faunas. Fossils of *M. meridionalis* from Tijeras Arroyo and Mesilla C are both closely associated with dates of 1.6 Ma on pumice from the lower Bandelier tuff, making them among the oldest dated mammoths in North America. San Antonio Mountain (SAM) Cave in northernmost New Mexico lacks large mammals, but the presence of the microtine rodents *Mictomys kansasensis*, an advanced species of *Allophaiomys*, *Lemmiscus curtatus*, and *Microtus* cf. *M. californicus* indicates a medial Irvingtonian age, between about 1.0 and 0.85 Ma.

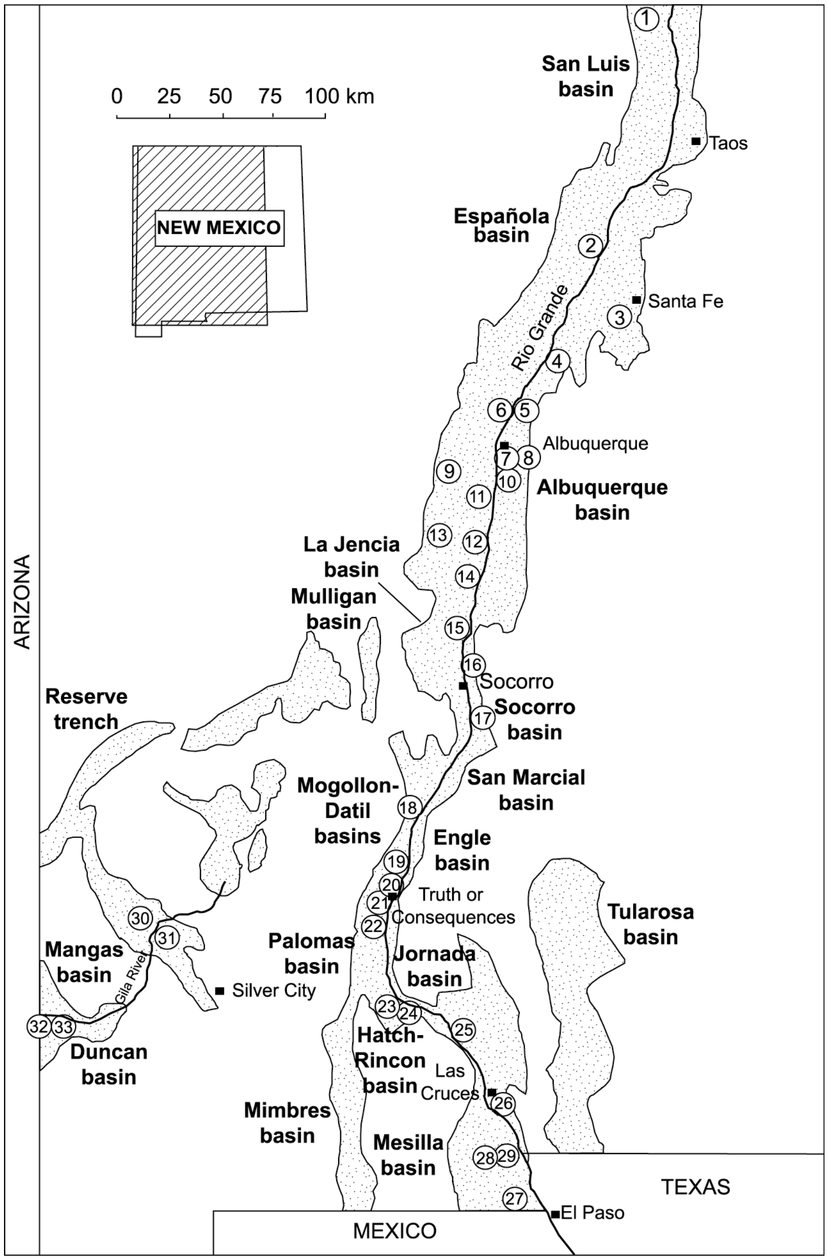

Figure 1. Map of New Mexico showing the location of late Hemphillian, Blancan, and Irvingtonian fossil sites. The structural basins are named and indicated by stippling. Sites are numbered from north to south in the Rio Grande Valley (sites 1-29), followed by sites in the Gila River Valley (sites 30-33). 1. San Antonio Mountain (SAM) Cave, medial Irvingtonian; 2. Puyé Formation site, late Hemphillian; 3. Ancha Formation sites, late Blancan; 4. Santo Domingo, late Blancan; 5. Western Mobile, early Irvingtonian; 6. Loma Colorado de Abajo, early/medial Blancan; 7. Mesa del Sol, Blancan; 8. Tijeras Arroyo, early Irvingtonian; 9. Pajarito, medial Blancan; 10. Isleta, Blancan; 11. Los Lunas, Blancan; 12. Belen, medial Blancan; 13. Mesas Mojinas, Blancan; 14. Veguita, Blancan; 15. Sevilleta, Blancan; 16. Arroyo de la Parida, medial Blancan; 17. Fite Ranch, early Irvingtonian; 18. Silver Canyon, Blancan; 19. Elephant Butte Lake, medial Blancan; 20. Cuchillo Negro Creek, medial Blancan; 21. Truth or Consequences, early Blancan; 22. Palomas Creek, medial Blancan; 23. Hatch, medial Blancan; 24. Rincon Arroyo, late Blancan/early Irvingtonian; 25. Tonuco Mountain, medial Blancan; 26. Tortugas Mountain, early Irvingtonian; 27. Mesilla A, late Blancan; 28. Mesilla B, latest Blancan; 29. Mesilla C, early Irvingtonian; 30. Buckhorn, early Blancan; 31. Walnut Canyon, latest Hemphillian; 32. Pearson Mesa, late Blancan; 33. Virden, latest Blancan.

MIOCENE MAMMALIAN FAUNAS AND BIOSTRATIGRAPHY OF THE ZIA FORMATION, NORTHERN ALBUQUERQUE BASIN, SANDOVAL COUNTY, NEW MEXICO

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Tedford (1981) reviewed the fossil mammal faunas from late Cenozoic basins in New Mexico, including the Albuquerque basin in the north-central part of the state. Early and middle Miocene mammal faunas are known from the northern third of the Albuquerque basin in Sandoval County, representing the Arikareean, Hemingfordian, Barstovian, and Clarendonian land-mammal ”ages” (Galusha, 1966; Gawne, 1975, 1976; Tedford, 1981; Tedford and Barghoorn, 1997, 1999; Morgan and Williamson, 2000). The Miocene vertebrate faunas from the northern Albuquerque basin are derived from the Zia Formation (Fig. 1). Galusha (1966) named the Zia “Sand” Formation with two members, the lower Piedra Parada Member and the upper Chamisa Mesa Member. The Cañada Pilares Member of Gawne (1981) is similar in age to the Chamisa Mesa Member, but is lithologically distinct. Connell et al. (1999) named the Cerro Conejo Member as the uppermost unit of the Zia Formation. Vertebrate fossils occur in all four members of the Zia Formation in the northern Albuquerque basin (Fig. 1).

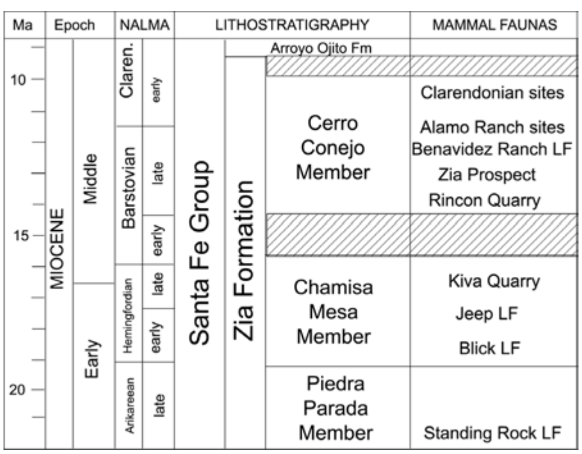

Figure 1. Lithostratigraphic and bistratigraphic correl-ation of the Zia Formation in the northern Albuquerque basin.

The Standing Rock Quarry is in the Piedra Parada Member of the Zia Formation, located in Arroyo Piedra Parada, south of San Ysidro on the Zia Reservation (Galusha, 1966). It has produced the oldest fossil mammal assemblage from the Zia Formation, the rich late Arikareean assemblage named the Standing Rock Local Fauna (LF) by Gawne (1975). Tedford (1981) assigned a late Arikareean age to the Standing Rock LF based on the association of the carnivores *Daphoenodon*, *Cephalogale*, and *Promartes* cf. *P. lepidus*, and the stenomyline camel *Stenomylus* cf. *S. gracilis*. Standing Rock Quarry is the type locality of the rodents *Proheteromys cejanus* and *Ziamys tedfordi*, named by Gawne (1975), and has also produced a nearly complete skeleton of the primitive rabbit *Archaeolagus* (Gawne, 1976). The Standing Rock LF is slightly younger than the well known late Arikareean Agate Springs Quarry from the Harrison Formation in western Nebraska.

The Blick Quarry and the stratigraphically equivalent *Cynarctoides* Quarry are in the middle of the Chamisa Mesa Member of the Zia Formation, located along Arroyo Pueblo east of Jemez Pueblo on the Jemez Reservation (Galusha, 1966). Gawne (1975) named the Blick LF for the combined fossil mammal assemblage from these two quarries. Tedford (1981) and Gawne (1975, 1976) assigned an early Hemingfordian age to the Blick LF based on the presence of the dog *Tomarctus optatus* (placed in the genus *Protomarctus* by Wang et al., 1999), the dog *Cynarctoides acridens*, the rodent *Pleurolicus* cf. *P. sulcifrons*, and the pika (ochotonid) *Oreolagus* cf. *O. nebrascensis*. The Blick Quarry is the type locality of the endemic stenomyline camel *Blickomylus galushai* (Frick and Taylor, 1968). The Blick LF is early Hemingfordian in age, and is similar to the Thomas Farm LF from Florida, the Martin Canyon LF from Colorado, and the faunas from the Runningwater Formation in Nebraska (Gawne, 1975).

The Jeep Quarry is located in the same general vicinity as the Blick Quarry in the Arroyo Pueblo drainage, but is higher stratigraphically, in the upper part of the Chamisa Mesa Member. Gawne (1975) named the Jeep LF for the mammalian assemblage from the Jeep Quarry and several nearby localities. Tedford (1981) assigned an early Hemingfordian age to the Jeep LF based on the presence of the bear dog *Amphicyon*, the mustelid *Promartes*, the camel *Protolabis*, and the mylagaulid rodent *Mesogaulus*. Other mammals from the Jeep LF (Gawne, 1975) include the bear dog *Ysengrinia*, the canids *Desmocyon thompsoni* and *Metatomarctus canavus*, the pronghorn antilocaprid *Merycodus*, and the camels *Michenia* and *Blickomylus galushai*. The Jeep Quarry is the type locality of the canid *Cynarctoides gawnae* (Wang et al., 1999). The Jeep LF is early Hemingfordian in age, slightly younger than the Blick LF, and intermediate in age between medial

Hemingfordian faunas from the Runningwater Formation and the late Hemingfordian Sheep Creek Fauna, both from Nebraska (Gawne, 1975; Tedford, 1981).

The Kiva Quarry is in the Chamisa Mesa or Piedra Parada members of the Zia Formation in the Jemez River area near Arroyo Ojito and Arroyo Piedra Parada (Cañada de Zia and Cañada Piedra Parada, respectively, of Galusha, 1966; Tedford, 1981). The presence of the borophagine dogs *Paracynarctus kelloggi* and *Microtomarctus conferta* and a primitive species of the horse genus *Protohippus* indicates a late Hemingfordian age for the Kiva Quarry (Tedford, 1981; Wang et al., 1999). The Kiva Quarry fauna is similar to mammalian faunas from the Nambé Member of the Tesuque Formation in the Española basin in northern New Mexico.

In Arroyo Ojito, and farther south along the Ceja del Rio Puerco, especially on the Alamo Ranch and Benavidez Ranch, faunas of late Barstovian age occur in the Cerro Conejo Member (usage of Connell et al., 1999) of the Zia Formation (Tedford 1981; Tedford and Barghoorn, 1999; Morgan and Williamson, 2000). The Benavidez Ranch LF is in the Cerro Conejo Member west of Rio Rancho in southern Sandoval County. The Benavidez Ranch mammalian fauna includes the rhinoceros *Peraceras*, the camels *Michenia*, *Procamelus*, and *Protolabis*, the pronghorn antilocaprid *Ramoceros*, and the proboscidean *Gomphotherium productum* (Morgan and Williamson, 2000). The Benavidez Ranch LF also has a diverse footprint fauna, including tracks made by a small wading bird, small and medium-sized camels, a rhinoceros, a horse, a large felid, a large borophagine canid, and a proboscidean (Williamson and Morgan, 2001). The most age-diagnostic taxon in the Benavidez Ranch LF is *Gomphotherium*, which first appears in southwestern faunas in the early middle Miocene at about 14.5 Ma, defining the beginning of the late Barstovian (Tedford et al., 1987; Tedford and Barghoorn, 1997, 1999). The remainder of the Benavidez Ranch LF is consistent with a late Barstovian age.

At Arroyo Ojito, the Rincon quarry of Galusha (1966) is in the lower 50 m of the Cerro Conejo type section, and the Zia prospect is near the middle of the Cerro Conejo section at Arroyo Ojito (S.D. Connell, 2000, oral commun.). The Rincon Quarry fauna includes the borophagine canids *Aelurodon ferox* and *Paratomarctus temerarius* and primitive species of the horse genera *Neohipparion* and *Pliohippus* (Tedford, 1981; Wang et al., 1999). The Rincon Quarry assemblage is similar to the late Barstovian fauna from the Santa Cruz sites in the Pojoaque Member of the Tesuque Formation in the Española basin (Tedford, 1981).

The Alamo Ranch site is another late Barstovian (Tedford, 1981, Tedford and Barghoorn, 1999). locality that is part of the Cerro Conejo Member. Faunas from the Cerro Conejo Member along the northern Ceja del Rio Puerco, from Cañada Navajo south to Cañada Pilares and Cañada Moquino, most of which are located on the Alamo Ranch, are similar to the Rincon Quarry assemblage. (Tedford, 1981) The Alamo Ranch sites are characterized by the beaver *Eucastor*, the camels *Aepycamelus*, *Michenia*, *Protolabis*, and *Procamelus*, the antilocaprid *Ramoceros*, and *Gomphotherium productum* (Tedford, 1981; Tedford and Barghoorn, 1999). The mammalian assemblages from the Cerro Conejo Member on the Alamo Ranch are typical of the late Barstovian (middle Miocene, 12-14 Ma; Tedford and Barghoorn, 1999). A late Barstovian age for the vertebrate faunas is supported by a K-Ar date of 13.64 Ma on a fallout ash in the Cerro Conejo Member (Tedford and Barghoorn, 1999).

Scattered, generally poorly documented Clarendonian mammal fossils are found in the upper part of the Cerro Conejo Member on the Zia Reservation between the San Ysidro and Zia faults of Connell et al. (1999), which are equivalent to the Jemez and Rincon faults, respectively, of Galusha (1966) and Tedford (1981). Another poorly documented Clarendonian locality includes the carnivore *Epicyon* and is near US-550 on the Santa Ana Reservation (R.H. Tedford, 1999, oral commun.), where several volcanic ashes are interbedded in the uppermost part of the Cerro Conejo Member. One of these ashes correlates to one of the Trapper Creek tephra in Idaho, which is dated at ~ 10.8 Ma (Personius et al., 2000). The age of this ash is consistent with the occurrence of the Clarendonian horses *Pliohippus* cf. *P. pernix*, *Cormohipparion* cf. *C. occidentale*, and a derived species of *Neohipparion* (Galusha, 1966; Tedford, 1981). Eastward, across the Zia fault, and in the Arroyo Arenoso drainage north of the Jemez River, rocks that are probably correlative with the Cerro Conejo Member have produced similar Clarendonian fossils (Tedford, 1981). Deposits that may be correlative with the Cerro Conejo Member are interbedded with the Chamisa Mesa basalt (Connell et al., 1999), which has a K-Ar date of about 10.4 Ma (Bailey and Smith, 1978). These two dates are consistent with a Clarendonian age for the youngest faunas from the Cerro Conejo Member.

The mammalian faunal succession from the Zia Formation in the northern Albuquerque basin begins in the late Arikareean and ends in the Clarendonian (between about 19-21 and 11 Ma), overlapping in age with much of the better known faunal sequence from the Española basin in northern New Mexico (Tedford, 1981). Late Arikareean faunas from the Aqiquiu Formation are similar in age to the Standing Rock LF in the Albuquerque basin. Early Hemingfordian sites comparable in age to the Blick and Jeep LFs appear to be absent from the Española basin. Sites from the Nambé Member of the Tesuque Formation are similar to the late Hemingfordian Kiva Quarry in the Albuquerque basin. There appears to be a hiatus in the northern Albuquerque basin sequence, equivalent to the early Barstovian, and corresponding to faunas from the Skull Ridge Member of the Tesuque Formation in the Española basin (Tedford, 1981; Tedford and Barghoorn, 1999). This hiatus is documented by magnetostratigraphy (Tedford and Barghoorn, 1999). Late Barstovian faunas from the Rincon Quarry, Alamo Ranch, and Benavidez Ranch in the northern Albuquerque basin are comparable in age to faunal assemblages in the Española basin from the Pojoaque Member of the Tesuque Formation, in particular the Santa Cruz sites (Tedford, 1981, fig. 2). The youngest faunas from the Zia Formation are Clarendonian in age, and are similar to several faunas in the Española basin, such as the Round Mountain Quarry in the Chamita Formation (Tedford, 1981).

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PLIOCENE MAMMALIAN BIOSTRATIGRAPHY AND BIOCHRONOLOGY AT LOMA COLORADO DE ABAJO, SANDOVAL COUNTY, NEW MEXICO

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Loma Colorado de Abajo is a prominent hill within the city limits of Rio Rancho in Sandoval County, about 20 km northwest of Albuquerque (Loma Machete quadrangle). Beginning in 1990 and continuing until 1996, Paul Knight collected several intriguing specimens of rodents from indurated, fine-grained reddish sandstones near the base of the exposed section on the south-facing escarpment of Loma Colorado de Abajo (New Mexico Museum of Natural History and Science [NMMNH] Site L-1462). The fossil site is located just a few hundred meters behind the recently built Rio Rancho High School, finished in the summer of 1997, although the school did not exist when the fossils were collected. The fossiliferous level is in the Loma Barbon Member in the upper part of the Arroyo Ojito Formation of Connell et al. (1999), about 8 m below the base of the overlying Ceja Member of the same formation (Fig. 1).

Morgan and Lucas (1999, 2000) described the vertebrate fossils as the Loma Colorado de Abajo local fauna (LF), which is limited in diversity, consisting of just three taxa, a small land tortoise and two rodent genera, *Spermophilus* and *Geomys*. The same stratum from which the rodent fossils were collected also contains numerous ichnofossils that appear to be rodent burrows. The Loma Colorado de Abajo LF is unique among New Mexico Blancan faunas in consisting entirely of small, burrowing vertebrates.

A ground squirrel of the genus *Spermophilus* is represented in the Loma Colorado de Abajo LF by a partial skull with P4 from a small species in the size range of living *S. tridecemlineatus*. It is considerably smaller than *Spermophilus* cf. *S. bensoni* from the Blancan of southeastern Arizona (Tomida, 1987), a species tentatively identified from the early Blancan Buckhorn LF in southwestern New Mexico (Morgan et al., 1997). The Loma Colorado *Spermophilus* skull is also smaller than *S. pattersoni* and *S. matachicensis* from the late Hemphillian Yepómera Fauna in northern Mexico (Wilson, 1949; Lindsay and Jacobs, 1985).

Three specimens from Loma Colorado de Abajo are provisionally referred to the primitive pocket gopher, *Geomys* (*Nerterogeomys*) *minor*, including a nearly complete skull, a rostrum with a complete dentition, and an edentulous left mandible. The two skulls are identified as *Geomys* on the basis of their bisulcate upper incisors, unrooted cheek teeth, and absence of enamel on the posterior surface of P4. Earlier pre-Blancan geomyids such as *Pliogeomys* have rooted cheek teeth. The fragmentary mandible lacks cheek teeth, but can be identified as a member of the extinct subgenus *Geomys* (*Nerterogeomys*) by the placement of the mental foramen ventral to the masseteric crest (Tomida, 1987). *Geomys* (*Nerterogeomys*) first appears in the early Blancan and becomes extinct in the early Irvingtonian. The Loma Colorado pocket gopher skulls are smaller than most described skulls of *Geomys* (*Nerterogeomys*), and compare most closely to the small species, *G. minor*, known from the early Blancan Rexroad Fauna in Kansas and Verde LF in Arizona, and the medial Blancan Beck Ranch LF in Texas and Benson Fauna in Arizona (Hibbard, 1967; Dalquest, 1978; Czaplewski, 1990). Repenning and May (1986) reported *G. minor* from the early Blancan Truth or Consequences LF from the Palomas Formation in Sierra County in central New Mexico. The Loma Colorado mandible is smaller than pocket gopher mandibles from the Pajarito and Belen faunas in the Albuquerque basin referred to *G.* (*Nerterogeomys*) *paenebursarius* (see Morgan and Lucas, 2000). The smaller species of *G.* (*Nerterogeomys*) that are most similar in size to the Loma Colorado *Geomys* (e.g., *G. minor*) are restricted to the Blancan, whereas the species that survive into the Irvingtonian (e.g., *G. anzensis*, *G. garbanii*, and *G. persimilis*) are larger.

The age of the Loma Colorado de Abajo LF is probably early or medial Blancan. Small species of *Geomys* (*Nerterogeomys*), such as *G. minor*, are typical of faunas of this age. Also, a medial to late Blancan fauna (older than 2.2 Ma) is known from the Ceja Member of the Arroyo Ojito Formation in Tijeras Arroyo, a unit that overlies the Loma Barbon Member. The Loma Colorado de Abajo LF is stratigraphically below and thus older than the Blancan fauna from Tijeras Arroyo. However, these two faunas have no taxa in common, so more detailed biostratigraphic comparisons are not possible.

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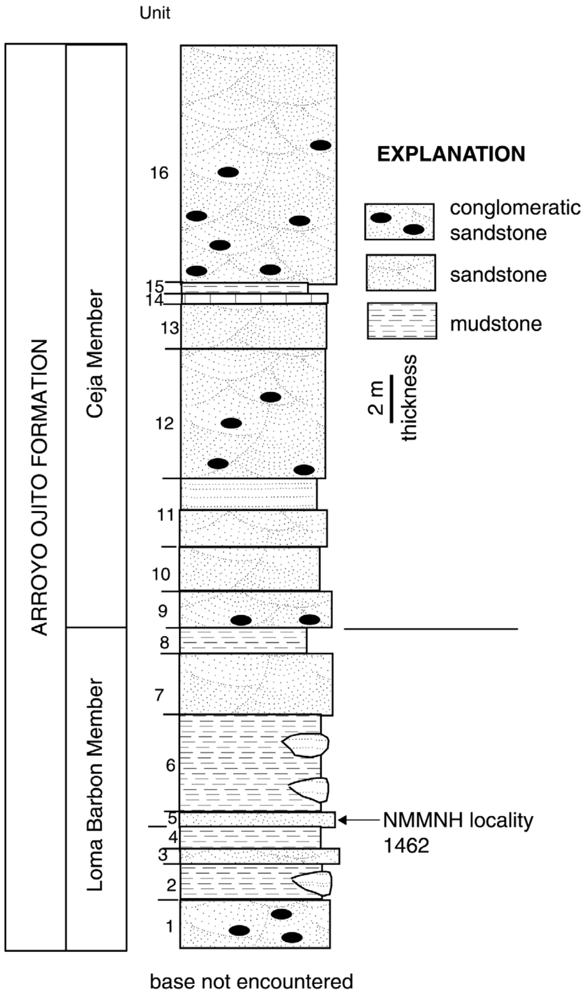

Figure 1. Stratigraphic section of the Loma Colorado de Abajo site. The top of section is at about 5530 ft (1630 m) elevation and is less than 52 m below the projected top of the Llano de Albuquerque (local top of upper Santa Fe Group).

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PLIO-PLEISTOCENE MAMMALIAN BIOSTRATIGRAPHY AND BIOCHRONOLOGY AT TIJERAS ARROYO, BERNALILLO COUNTY, NEW MEXICO

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Most of the vertebrate fossils from Tijeras Arroyo, located just south of the Albuquerque International Airport in Bernalillo County, are derived from the Sierra Ladrones Formation and are early Irvingtonian in age (Lucas et al., 1993). However, one locality (New Mexico Museum of Natural History and Science [NMMNH] site L-1458) at the base of the exposed stratigraphic section in Tijeras Arroyo (Fig. 1) has produced two species that are indicative of a Blancan age. The fossils from this site were derived from a sandstone comprising unit 1 in the stratigraphic section of Lucas et al. (1993, fig. 2). The lowermost part of the section in Tijeras Arroyo, including unit 1, was recently referred to the Ceja Member of the Arroyo Ojito Formation (Connell and Hawley, 1998; Connell et al., 1999).

Both mammals identified from site L-1458 in the Tijeras Arroyo section, *Hypolagus* cf. *H. gidleyi* and *Equus* cf. *E. cumminsii*, are typical of Blancan faunas, and do not occur in the Irvingtonian. The extinction in the late Pliocene (about 2.2 Ma) of several characteristic Blancan genera, including *Hypolagus*, *Borophagus*, *Rhynchotherium*, and *Nannippus*, is considered one of the most important biochronological events in the late Blancan (Lindsay et al., 1984). The presence of *Hypolagus* thus indicates that site L-1458 is older than 2.2 Ma. *Equus* cf. *E. cumminsii* appears to be absent from early Blancan faunas, so L-1458 is probably middle or early late Blancan in age.

Ten stratigraphically higher localities in Tijeras Arroyo have produced a significant vertebrate fauna of early Irvingtonian age (Lucas et al., 1993; Morgan and Lucas, 2000). More than 75 m of the Sierra Ladrones Formation are exposed in Tijeras Arroyo, consisting of sandstones, pumiceous sandstones, and gravels, with minor amounts of mudstone and diatomite. These sediments represent axial river deposits of an ancestral Rio Grande. The most distinctive lithologic chracteristic of these beds is the presence of reworked Guaje Pumice derived from the Bandelier Tuff, Ar/Ar dated at 1.61 Ma (Izett and Obradovich, 1994), in the units associated with an Irvingtonian fauna (units 3-8 of Lucas et al., 1993). An extensive flora of leaves and pollen from a localized volcanic ash bed was collected in the Tijeras Arroyo section (NMMNH Site L-1445). The Tijeras Arroyo flora indicates that the cottonwood forest or bosque currently found along the banks of the Rio Grande in New Mexico dates back to at least the early Pleistocene (Knight et al., 1996).

The land tortoise *Hesperotestudo* and five species of mammals, including *Glyptotherium* cf. *G. arizonae*, *Equus scotti*, *Equus* sp., *Camelops* sp., and *Mammuthus meridionalis* occur together in the Tijeras Arroyo section above the Blancan site (l-1458) discussed above (Lucas et al., 1993; Morgan and Lucas, 2000). These species constitute a fairly typical fauna of early Irvingtonian age. Three additional species of mammals, a small species of *Equus*, the llama *Hemiauchenia macrocephala* and the mammoth *Mammuthus imperator*, occur somewhat higher in the Tijeras Arroyo section than the remainder of the fauna, but probably are Irvingtonian as well.

A caudal osteoderm of a glyptodont from Tijeras Arroyo (Lucas et al., 1993) probably is not diagnostic at the species level, although this specimen almost certainly represents *Glyptotherium arizonae*. Tentative referral of this osteoderm to *G. arizonae* is reasonable as its association with *Mammuthus* rules out a Blancan age, and the Rancholabrean *G. floridanum* is restricted to the Atlantic and Gulf coastal plains (Gillette and Ray, 1981). The large horse *Equus scotti* is the most common mammal in the Tijeras Arroyo Irvingtonian fauna, represented by mandibles, isolated teeth, and postcrania (Lucas et al., 1993; Morgan and Lucas, 2000). *E. scotti* is the typical large horse in late Blancan and early Irvingtonian faunas in the southwestern United States (Hibbard and Dalquest, 1966), and occurs in medial Blancan through early Irvingtonian faunas in New Mexico (Tedford, 1981; Morgan et al., 1998). A complete equid metacarpal from Tijeras Arroyo is more slender than metacarpals of *E. scotti*, and represents a second, smaller species of *Equus* (Hibbard and Dalquest, 1966; Harris and Porter, 1980). A partial skull of a small *Equus* occurs higher in the Tijeras Arroyo section.

Lucas and Effinger (1991) and Lucas et al. (1993) referred a mandible with left and right m3 from Tijeras Arroyo to the primitive mammoth *Mammuthus meridionalis* on the basis of its low plate count and extremely thick enamel. This is one of only two records of mammoths from New Mexico referred to *M. meridionalis*, indicating that this fauna is almost certainly early Irvingtonian. The other record consists of several partial teeth, tentatively referred to *M. meridionalis*, from an early Irvingtonian fauna in the Mesilla basin (Vanderhill, 1986). Lucas et al (1993) referred a left M3 in a maxillary fragment from Tijeras Arroyo to the mammoth *Mammuthus* *imperator*. The teeth of *M. imperator* are more advanced than *M. meridonalis* in having a higher plate count, higher lamellar frequency, and thinner enamel. The *M. imperator* specimen was found about 12 m higher in the section than the remainder of the Tijeras Arroyo fauna, and thus is somewhat younger, although an Irvingtonian age is still likely (Lucas et al., 1993).

The presence of mammoths in unit 6 of Lucas et al. (1993) and above clearly establishes an Irvingtonian age for the upper part of the Tijeras Arroyo section, as *Mammuthus* is one of the defining genera of the Irvingtonian NALMA. The first appearance of *Mammuthus* in the New World occurred sometime in the early Pleistocene (early Irvingtonian) between about 1.8 and 1.6 Ma. The mammoth jaws from Tijeras Arroyo represent one of the oldest well-documented records of *Mammuthus* from North America, based on an Ar/Ar age of 1.61 Ma on Guaje Pumice from the Sierra Ladrones Formation in Tijeras Arroyo (Lucas et al., 1993; Izett and Obradovich, 1994; Lucas, 1995, 1996). Although the pumice date provides a maximum age for this site, evidence from other pumice deposits of exactly the same age farther south in the Rio Grande Valley (Mack et al., 1996, 1998) indicates that the pumice is very close in age to the fossils. The association of M. *meridionalis* with *Glyptotherium arizonae* and *Equus scotti* is indicative of an early Irvingtonian age for the Tijeras Arroyo fauna. Correlative early Irvingtonian faunas include the Tortugas Mountain LF (Lucas et al., 1999, 2000) and Mesilla Basin Fauna C (Vanderhill, 1986) from the Mesilla basin in southern New Mexico, Gilliland in Texas (Hibbard and Dalquest, 1966), and Holloman in Oklahoma (Dalquest, 1977).

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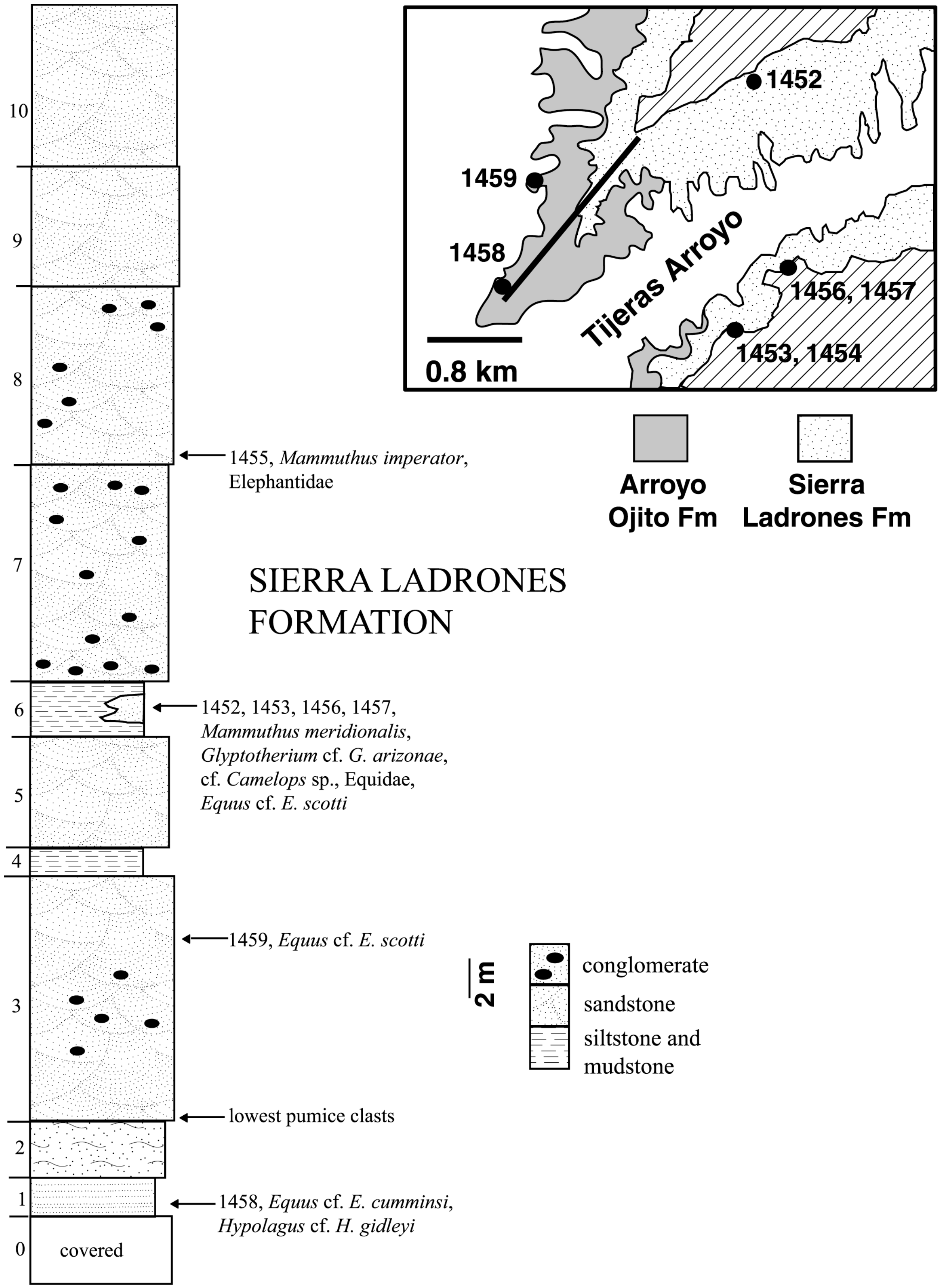

Figure 1. Stratigraphic column of Sierra Ladrones and Arroyo Ojito Formation strata at mouth of Tijeras Arroyo.

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LITHOSTRATIGRAPHY AND PLIOCENE MAMMALIAN BIOSTRATIGRAPHY AND BIOCHRONOLOGY AT BELEN, VALENCIA COUNTY, NEW MEXICO

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INTRODUCTION

Extending south of Los Lunas volcano to Belen and into northern Socorro County, badlands developed in the Arroyo Ojito Formation of Connell et al. (1999) are well exposed in an east-facing escarpment just west of Interstate Highway 25 and several km west of the Rio Grande (Fig. 1). In 1982, John Young examined numerous sections exposed on the east and west sides of the Llano de Albuquerque and described four in his master's thesis at the New Mexico Institute of Mining and Technology. The two thickest sections exposed more than 100 m of upper Santa Fe Group basin fill. As can be seen in the outcrops at the Belen site, cross-bedded gravel and gravelly sands alternate up section with finer-grained units. Weak soils and eolian deposits are also common. The gravel commonly has a suite of pebble types, including well rounded siliceous pebbles (recycled from Paleogene, Mesozoic, and upper Paleozoic units of the Colorado Plateau), basaltic, intermediate, and silicic volcanic rocks, red granitic rocks, silicified wood, sandstone concretions (recycled Mesozoic), pycnodonte shell fragments (Cretaceous; Hook and Cobban, 1977), carbonate rocks (upper Paleozoic limestones and Neogene travertines), and rare obsidian pebbles from East Grants Ridge, about 112 km northwest of here. The presence of Grants Ridge obsidian indicates that much of this section was derived from the ancestral Rio San Jose fluvial system, which is presently a tributary to the Rio Puerco.

Young (1982) compared amounts of Rb, Y, Zr and Sr in obsidian samples from East Grants Ridge to the same trace elements in the pebbles and found almost identical amounts, thereby demonstrating more than a visual match. Shackley (1998) showed that there were two similar but distinct sources of obsidian in the area of East Grants Ridge and was able to distinguish them by amounts of Zr, Y, and Nb, among other elements. Lipman and Mehnert obtained an age of 3.2 ± 0.3 Ma for the East Grants Ridge obsidian.

The obsidian is recognized in conglomeratic units as deep as 53 m below the Llano de Albuquerque surface, with a marked increase in amounts above 27 m. The presence of this obsidian constrains the age of the upper 50 m of section to less than 3 million years old.

BIOSTRATIGRAPHY

The New Mexico Museum of Natural History and Science (NMMNH) has two collections of Blancan vertebrates from southwest of Belen in Valencia County. In 1992, Bill Wood collected vertebrate fossils about 5 km southwest of Belen (NMMNH Site L-3778). Fossils from this site include lower jaws of the gomphotheriid proboscidean *Stegomastodon mirificus* and postcranial elements of the horse *Equus*. Christopher Whittle and several students collected fossils from conglomeratic sandstone and slightly indurated sandstone about 2 km southwest of Belen (NMMNH Site L-3737), about 4 km north of site L-3778 and just south of Camino del Llano Road (formerly Sosimo Padilla Road). Fossils from this site include a snake, the mole *Scalopus*, the rodent *Geomys*, the horse *Equus*, and a small antilocaprid. Because of the close proximity of sites L-3737 and 3778 southwest of Belen and their occurrence in similar strata referred to the Arroyo Ojito Formation, the fossils from these two sites are combined as the Belen Fauna (Morgan and Lucas, 2000).

The Belen Fauna (Morgan and Lucas, 2000) is composed of five species of mammals, including *Scalopus* (*Hesperoscalops*) cf. *S. blancoensis*, *Geomys* (*Neterogeomys*) cf. *G. paenebursarius*, *Equus* cf. *E. calobatus*, a small antilocaprid, and *Stegomastodon mirificus*. A dentary with m1-m3 from the Belen Fauna is the first mole (family Talpidae) ever reported from New Mexico, recent or fossil (Morgan and Lucas, 1999, 2000). This mole is referred to *Scalopus* (*Hesperoscalops*), an extinct subgenus of *Scalopus* restricted to the Blancan. Three species of *S.* (*Hesperoscalops*) have been described, *S. sewardensis* from the very early Blancan Saw Rock Canyon LF in Kansas, *S. rexroadi* from the early Blancan Rexroad and Fox Canyon faunas in Kansas and the medial Blancan Beck Ranch LF in Texas, and *S. blancoensis* from the late Blancan Blanco LF in Texas (Hibbard, 1953; Dalquest, 1975, 1978; Kurtén and Anderson, 1980). The Belen dentary is tentatively referred to *S. blancoensis* based on its similarity to that species in size and morphological features. A dentary with a complete dentition from Belen is identified as the extinct pocket gopher subgenus *Geomys* (*Nerterogeomys*).

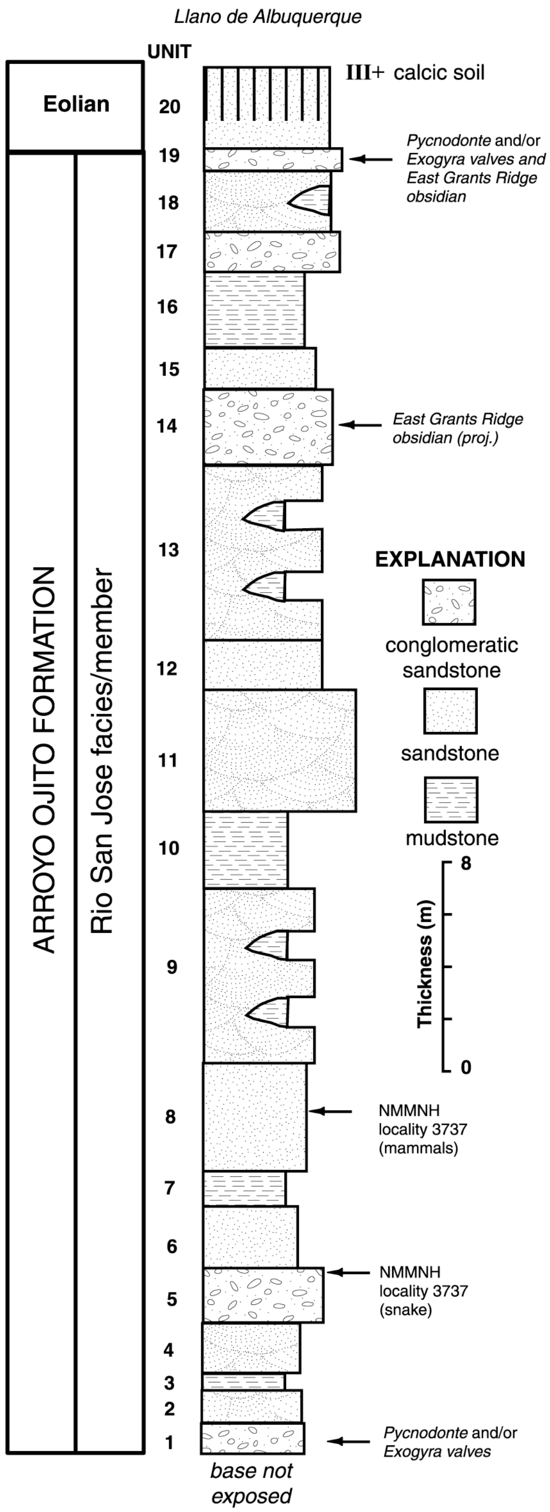

Figure 1. Stratigraphic column near Camino del Llano (formerly Sosimo Padilla Road). Modified from Morgan and Lucas (2000) with projections of *Pycnodonte* and/or *Exogyra* valves and East Grants Ridge obsidian from Young (1982).

The morphology and size of this mandible are similar to the species *G.* (*N.*) *paenebursarius*, also identified from the Pajarito LF, and first described from the late Blancan Hudspeth and Red Light LFs of southwestern Texas (Strain, 1966; Akersten, 1972).

The most common fossils in the Belen Fauna are postcranial elements of horses of the genus *Equus*, most of which are not diagnostic at the species level. A nearly complete metatarsal is tentatively referred to the large, stilt-legged horse, *E. calobatus*, a species known from the late Blancan Santo Domingo LF (Tedford, 1981) and from late Blancan and early Irvingtonian faunas in the Mesilla basin (Vanderhill, 1986). A well preserved pair of mandibles with right and left m2-m3 are referred to the gomphothere *Stegomastodon mirificus*. The presence of seven lophids on m3 separates this specimen from *Rhynchotherium* and *Cuvieronius*, and the highly complicated enamel with double trefoiling distinguishes the teeth from the more primitive species *S. rexroadensis*.

Four mammals in the Belen Fauna are age diagnostic. The extinct subgenus *Scalopus* (*Hesperoscalops*) is restricted to the Blancan, and the species *S. blancoensis* occurs in the late Blancan. *Geomys* (*Nerterogeomys*) *paenebursarius* is known from two late Blancan faunas in southwestern Texas (Strain, 1966; Akersten, 1972), and the medial Blancan Pajarito LF in the northern Albuquerque basin (Tedford, 1981; Morgan and Lucas, 2000). *Stegomastodon mirificus* is known from the medial Blancan through the early Irvingtonian, and *Equus calobatus* occurs in the late Blancan and Irvingtonian (Kurtén and Anderson, 1980). The age of the Belen Fauna thus is either medial or late Blancan. *S. blancoensis* and *E. calobatus* occur in late Blancan faunas, but are not known from the medial Blancan, whereas *G.* (*N.*) *paenebursarius* and *S. mirificus* first appear in the medial Blancan. The lack of South American immigrants in the Belen Fauna suggests a medial Blancan age, although their absence could be related to biogeographic factors. Neotropical mammals are unknown from Blancan faunas in northern New Mexico; however, *Glyptotherium* occurs in two early Irvingtonian faunas in the Albuquerque basin, Tijeras Arroyo and Western Mobile. We tentatively place the Belen Fauna in the medial Blancan based on similarities with other medial Blancan faunas (e.g., Pajarito LF) from the Arroyo Ojito Formation in the Albuquerque basin.

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PLIOCENE MAMMALIAN BIOSTRATIGRAPHY AND BIOCHRONOLOGY AT ARROYO DE LA PARIDA, SOCORRO COUNTY, NEW MEXICO

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In 1935, vertebrate fossils were first found in Arroyo de la Parida, about 6 km northeast of Socorro, Socorro County. Needham (1936) reported a complete pair of lower jaws of the gomphotheriid proboscidean *Rhynchotherium* and a lower molar of the horse *Plesippus* (now considered a subgenus of *Equus*) from an exposure of sands and gravels of the Santa Fe Group on the southern side of Arroyo de la Parida, about 2 km east of its confluence with the Rio Grande. Additional vertebrate fossils were collected from this same exposure by students from the New Mexico Institute of Mining and Technology (DeBrine et al., 1963).

Curt Teichert, a well known expatriate German invertebrate paleontologist, collected a sample of vertebrate fossils from the vicinity of Arroyo de la Parida in 1953, and donated these fossils to the American Museum of Natural History. The only locality information associated with Teichert's sample was that the fossils were collected "about four miles north of Socorro, New Mexico." Based on the general locality, preservation of the fossils, and the composition of the fauna, there is little doubt that Teichert's fossils are from the area that yields the Arroyo de la Parida local fauna (LF). The fossils collected by Teichert were summarized by Tedford (1981), and include three species of horses, *Equus simplicidens*, *E.* cf. *E. cumminsii*, and *E.* cf. *E. scotti*, the small antilocaprid *Capromeryx*, and the gomphothere *Stegomastodon*.

Lucas and Morgan (1996) described and illustrated the mandibles of *Rhynchotherium* first mentioned by Needham (1936), and referred them to the species *R. falconeri*, originally described from the Pliocene Blanco LF in Texas. Lucas and Morgan (1996) also summarized the biostratigraphy of the Arroyo de la Parida LF, including fossils collected in 1996 by two students from New Mexico Tech, Ed Frye and Mike O'Keeffe. We visited the Arroyo de la Parida area several times during 2000 and collected numerous additional fossils from 15 different sites (Morgan et al., 2000).

The Arroyo de la Parida LF is derived from a 70-m-thick sequence of sands and gravels that constitute the axial river (ancestral Rio Grande) facies of the Palomas Formation. Sandstone and conglomerate derived from the eastern basin margin interfinger with, and overlie these fluvial sediments. The strata in the vicinity of Arroyo de la Parida are located at the northern end of the Socorro basin, representing one of the northernmost occurrences of the Palomas Formation, which has its type area about 100 km farther south in Palomas Creek near Truth or Consequences in Sierra County (Lozinsky, 1986). The Arroyo de la Parida LF is composed of ten species of vertebrates: the land tortoise *Hesperotestudo*; the ground sloth *Megalonyx* cf. *M. leptostomus*; three species of horses, *Equus* cf. *E. cumminsii*, *E. scotti*, and *E. simplicidens*; two camelids, a large species of *Camelops* and a small species of *Hemiauchenia*; the small antilocaprid *Capromeryx*; and two proboscideans, *Rhynchotherium falconeri* and *Stegomastodon* sp. This is a fairly typical faunal assemblage found in New Mexico Blancan sites, mostly consisting of large grazing ungulates and dominated by horses of the genus *Equus*.

Five mammals from the Arroyo de la Parida LF are restricted to the Blancan, including *Megalonyx leptostomus*, *Equus cumminsii*, *E. simplicidens*, the large *Camelops*, and *Rhynchotherium falconeri*. The most age-diagnostic of these taxa is *Rhynchotherium*, a gomphothere that became extinct in the late Pliocene at about 2.2 Ma together with several other characteristic genera of Blancan mammals. The lower jaws of *R. falconeri* from Arroyo de la Parida were collected near the top of the local section of the Palomas Formation, suggesting that the entire fauna, most of which occurs some 40 m lower in the section, is older than 2.2 Ma. An early Blancan age for the Arroyo de la Parida LF can be ruled out by the presence of *E. scotti* and *Camelops*, both of which first appear in New Mexico faunas during the medial Blancan. The absence of South American immigrants suggests an age greater than 2.7 Ma. *Megalonyx* is the only Blancan mammal of South American origin that was not a participant in the Great American Interchange. *Megalonyx* or its progenitor arrived from South America in the late Miocene about 9 Ma. *M. leptostomus* is fairly widespread in early through late Blancan faunas. The Arroyo de la Parida LF is thus interpreted to be medial Blancan in age (3.6-2.7 Ma), and is similar to the Cuchillo Negro Creek LF from the Palomas Formation in the Engle basin near Truth or Consequences.

A Blancan fauna is known from the extreme southern end of the Albuquerque basin near San Acacia in northern Socorro County (Denny, 1940). This site is located just north of the Rio Salado on the western side of the Rio Grande, presumably from the Sierra Ladrones Formation, as this site is near the type area of the Sierra Ladrones Formation of

Machette (1978). The fauna reported by Denny (1940, p. 93) from the San Acacia site consists of the gomphothere *Stegomastodon mirificus* and an undetermined species of *Equus*. We have not examined these fossils, so the identifications are taken from Denny's paper and must be considered tentative. The San Acacia site is similar to the middle to late Blancan Arroyo de la Parida local fauna, derived from the Palomas Formation about 15 km farther south in the northern part of the Socorro basin (Tedford, 1981; Lucas and Morgan, 1996).

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STRATIGRAPHY OF THE LOWER SANTA FE GROUP, HAGAN EMBAYMENT, NORTH-CENTRAL NEW MEXICO: PRELIMINARY RESULTS

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INTRODUCTION

Geologic mapping and stratigraphic studies of upper Oligocene through middle Miocene sedimentary rocks of the Santa Fe Group exposed in the Hagan embayment constrain the initial development of the Albuquerque Basin and Rio Grande rift in north-central New Mexico. These sedimentary rocks are exposed in the Albuquerque Basin along Arroyo de la Vega de los Tanos (herein called Tanos Arroyo) at the northeastern dip-slope of Espinaso Ridge in the Hagan embayment of central New Mexico (Fig. 1). This paper presents preliminary findings of geologic studies on two formations proposed for lower Santa Fe Group strata exposed in the Hagan embayment.

STRATIGRAPHY OF TANOS ARROYO

Stratigraphic sections were measured and described on the northeastern flank of Espinaso Ridge along Tanos Arroyo (Fig. 1). The Tanos Arroyo section comprises two formation-rank units that are informally subdivided into members and lithofacies units. These deposits are composed primarily of recycled volcanic and porphyritic intrusive detritus derived from the adjacent Ortiz Mountains, on the footwall of the La Bajada fault (Fig. 2). The base of this succession is here called the Tanos Formation, which overlies the volcaniclastic Oligocene Espinaso Formation (*ca*. 36-27 Ma, Kautz et al., 1981). The Tanos Formation is a succession of moderately tilted conglomerate, thin- to medium-bedded mudstone and tabular sandstone. The Tanos Formation is 253 m thick at the type section (Figs. 2-3, TA1), where it is subdivided into a basal piedmont conglomerate member, a middle mudstone and sandstone member, and an upper tabular sandstone member. Ripple laminated sandstone beds are common in the lower part of the middle member. These lithofacies occur in a distinct stratigraphic succession at the type section and are assigned to informal member-rank terms. Mudstone beds thin to the southeast, near the mouth of Arroyo del Tuerto (Fig. 3, TA; Arroyo Pinovetito of Stearns, 1953), which is about 4 km south of the type section. The Tanos Formation contains a mudstone and fluviatile sandstone interval, suggesting deposition in a playa-lake and distal, streamflow-dominated piedmont setting. These members are associated with the transition between the piedmont-slope and the basin-floor. An olivine basalt flow, about 9 m above the base at the type section, yielded a whole-rock $^{40}Ar/^{39}Ar$ date of 25.41±0.32 Ma (W.C. McIntosh, 2000, written commun.; Cather et al., 2000), which is consistent with an earlier K/Ar date of about 25.1±0.7 Ma (Kautz et al., 1981) at the northern tip of Espinaso Ridge.

The basal contact of the Tanos Formation is sharp and slightly scoured. No angular unconformity with the underlying Espinaso Formation is apparent in outcrop. A continuous dip-meter log for the Pelto Blackshare Federal #1 well (Sec. 35, T14N, R6W, San Felipe Pueblo NE quadrangle), drilled nearly 5 km south-southeast of the Tanos type section (on file at the New Mexico Bureau of Mines and Mineral Resources in Socorro, New Mexico; Library of Subsurface Data #26,091), indicates an angular unconformity at about 460 m below land surface (bls). Strata encountered in this well are oriented about N25°E, 10-12°NW below 460 m bls, and about N60°W, 8°NE above. Thus we interpret the contact between the Espinaso and Tanos formations to be unconformable. Restoration of Tanos Formation bedding to horizontal attitude indicates that the Espinaso Formation was oriented about N10-12°W, 10-14°NE prior to deposition of the Tanos Formation. The dip-meter log does not show significant steepening in dips that would indicate the presence of a normal fault, which commonly have dips of about 60°. Thus, the dip-meter log indicates that the Espinaso Formation underwent an episode of deformation prior to deposition of the Tanos Formation.

The age of this unconformity is bracketed by a K/Ar date of 26.9±0.6 Ma reported for a nepheline latite flow about 130 m below the top of the Espinaso Formation (Kautz et al., 1981) and the basalt dated 25.41±0.32 Ma in the basal Tanos Formation. Thus, the hiatus represented by this unconformity at the type section is thus less than 1.5 m.y. in duration, and likely spans a much shorter interval of time. If there was basal onlap of Tanos Formation to the east, then this unconformity might span an even shorter period of time towards the center of the basin, which was presumably northwest of the type section.

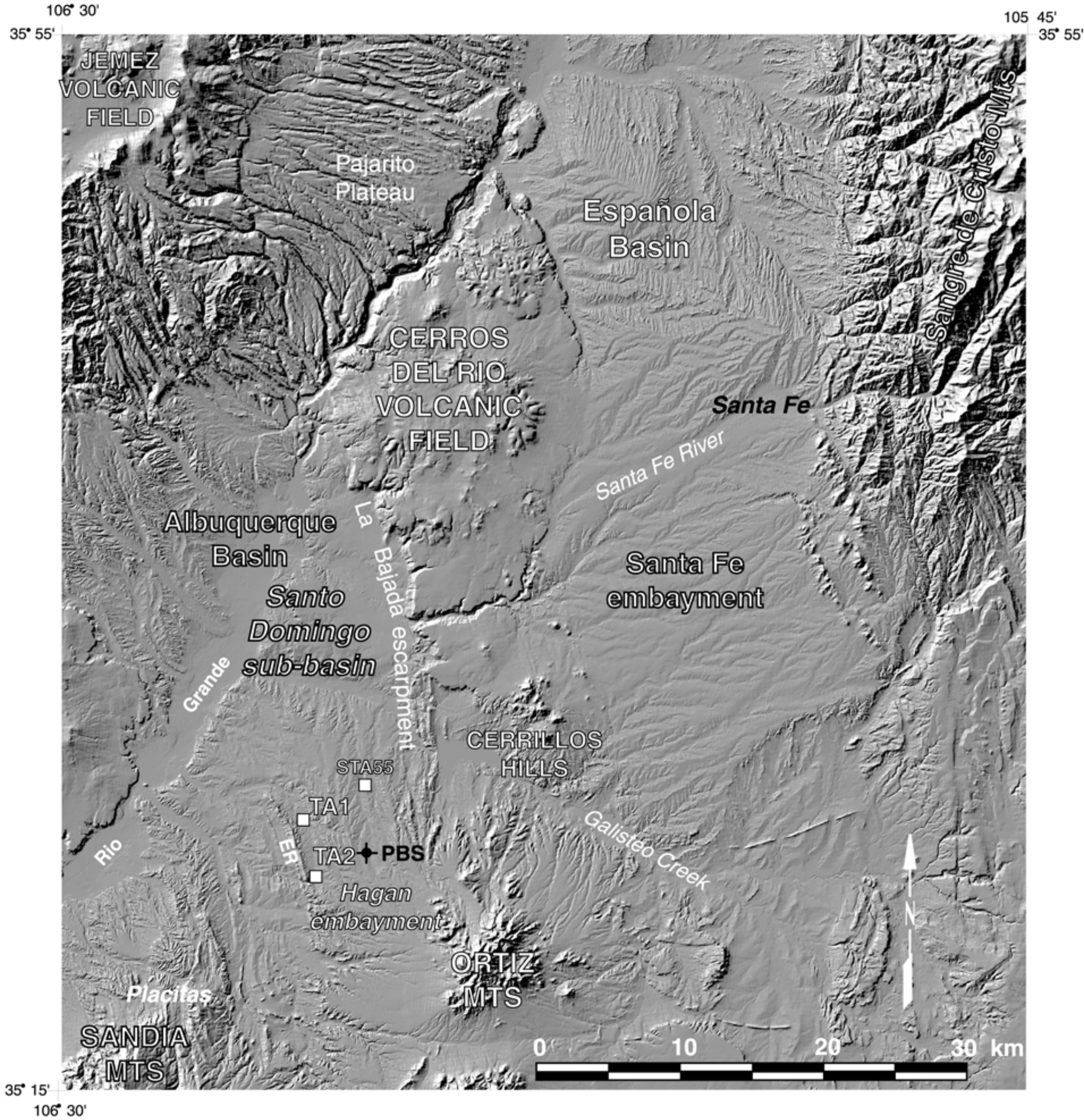

Figure 1. Shaded relief map, illustrating the locations of major geographic features and the study area. Base produced from U.S. Geological Survey 30-m DEM data. Localities include the Pelto Blackshare Federal #1 (PBS), Tanos Arroyo sections (TA1, TA2), and selected localities mentioned in text.

The mapped extent of the Tanos Formation corresponds approximately with strata tentatively assigned to the Abiquiu Formation by Stearns (1953), and to strata Kelley (1979) correlated to the Zia Formation. The Tanos Formation is in part, temporally equivalent to the Abiquiu Formation (Tedford, 1981; Moore, 2000). The Tanos Formation, however, is lithologically dissimilar to the Abiquiu Formation because it contains abundant locally derived volcanic detritus derived from the adjacent Ortiz Mountains (Large and Ingersoll, 1997). In contrast, the Abiquiu Formation in the Abiquiu embayment, about 70 km northwest of the study area (Smith, 1995; Moore, 2000), consists largely of epiclastic sediments derived from the Latir volcanic field of northern New Mexico. Paleocurrent measurements from the Tanos and Blackshare formations indicate flow to the west-northwest, away from the highlands of the Ortiz Mountains (Fig. 4) and support the petrographic interpretations of Large and Ingersoll (1997) that these deposits were derived from the Ortiz Mountains.

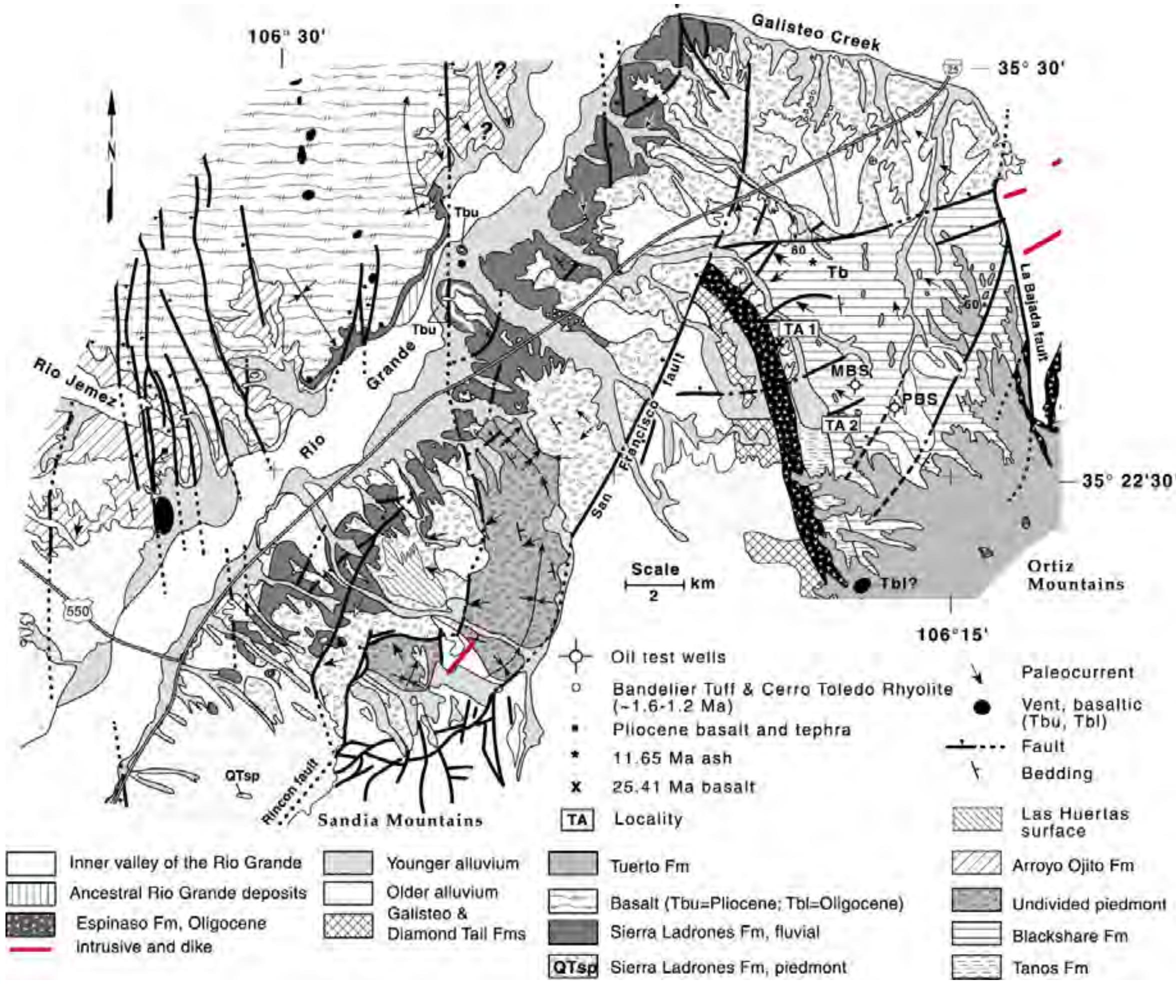

Figure 2. Simplified geologic map of the southeastern Santo Domingo sub-basin, illustrating locations stratigraphic sections along Tanos Arroyo (TA1 and TA2). Compiled from Cather and Connell (1998), Cather et al. (2000), Connell, (1998), Connell et al. (1995), and unpublished mapping.

The basal Santa Fe Group strata in the Hagan embayment are older than the Zia Formation (Fig. 5) and are not directly correlative as originally suggested by Kelley (1977). The Zia Formation is exposed 30-45 km to the west on the northwestern margin of the Albuquerque Basin. The Hagan embayment contains a thick succession of mudstone and fluvial sandstone derived from local sources to the east, whereas the eolian-dominated lower Zia Formation was deposited by westerly winds and sparse, widely spaced southeast-flowing streams (Beckner and Mozley, 1998; Gawne, 1981).

The Tanos Formation is conformably overlain by a >700-m thick succession of sandstone, conglomerate, and minor mudstone herein called the Blackshare Formation, for the nearby Blackshare Ranch, located in a tributary of Tanos Arroyo. The Blackshare Formation is interpreted as stream-flow and hyperconcentrated-flow deposits laid down by streams that originated form the Ortiz Mountains and eastern margin of the Hagan embayment. Conglomerate beds are commonly lenticular and sandstone intervals commonly fine upward into thinly bedded mudstone, which have upper contacts that are commonly scoured by lenticular conglomerate of an overlying fining-upward sequence. The upper boundary of the Tanos Formation is gradational and interfingers with the overlying Blackshare Formation. The contact is placed at the lowest lenticular pebbly to cobbly sandstone in this tabular sandstone/conglomeratic-sandstone transition. This contact was chosen on the basis of measured sections and differs slightly from the mapped contact (Cather et al., 2000), which was placed at the top of the highest, thickly bedded, tabular sandstone.

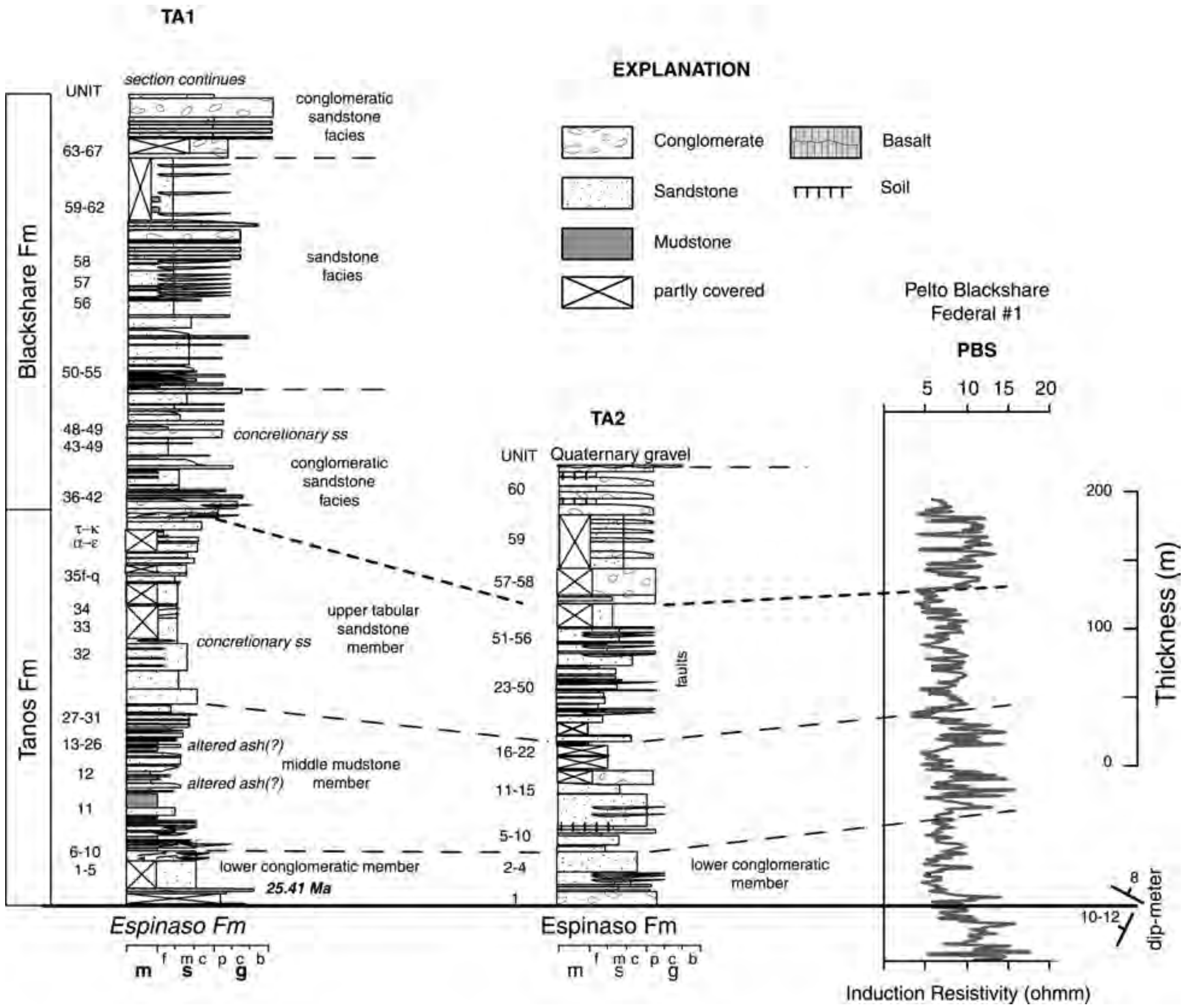

Figure 3. Composite stratigraphic section of the type locality (TA1) and Arroyo del Tuerto reference section (TA2) of the Tanos and Blackshare formations. Horizontal scale indicates approximate maximum grain size. The Pelto Blackshare Federal #1 (PBS), drilled about 6 km to the east, encountered similar deposits as interpreted from borehole geophysics and a continuous dip-meter log.

The type section of the Blackshare Formation is about 312 m above the top of the Tanos Formation type section. A complete section of the Blackshare Formation was not measured because the top is not recognized in the study area and exposures are commonly quite poor northeast of Tanos Arroyo. Discontinuous outcrops of the Blackshare Formation extend 6 km east to the La Bajada fault.

The Blackshare Formation is locally differentiated into three mappable textural lithofacies (Cather et al., 2000), following methods proposed by Cather (1997). These units interfinger, are not superposed, and do not necessarily occur in any particular stratigraphic order. The conglomeratic piedmont lithofacies consists of well cemented conglomerate and subordinate sandstone. The conglomeratic sandstone lithofacies consists of subequal amounts of sandstone and conglomerate. The sandstone member contains sandstone with subordinate conglomerate and mudstone.

An ash within the upper exposures of the Blackshare Formation was projected into the type section, where it is between 670-710 m (estimated from geologic map of Cather et al., 2000) above the base. This ash yielded a single-crystal (on sanidine) $^{40}Ar/^{39}Ar$ date of 11.65±0.38 Ma (W.C. McIntosh, 2000, written commun., Cather et al., 2000). Other fluvially recycled ashes, up to 3 m in thickness, occupy similar stratigraphic positions to the dated ash (Cather et al., 2000; Stearns, 1953); however, they are too fine grained to be dated using the $^{40}Ar/^{39}Ar$ technique.

The Plio-Pleistocene Tuerto Formation overlies the Blackshare and Tanos formations with angular unconformity. The subhorizontally bedded Tuerto Formation overlies beds of the Tanos Formation that tilt 27-36°NE. Dips in the Blackshare Formation progressively decrease upsection, where stratal tilts of 4-16°NE are observed stratigraphically above the 11.65 Ma ash in the Blackshare Formation; higher stratal tilts are commonly near faults. The top of the Blackshare Formation is cut by the La Bajada fault or is unconformably overlain by the subhorizontally bedded Plio-Pleistocene Tuerto Formation.

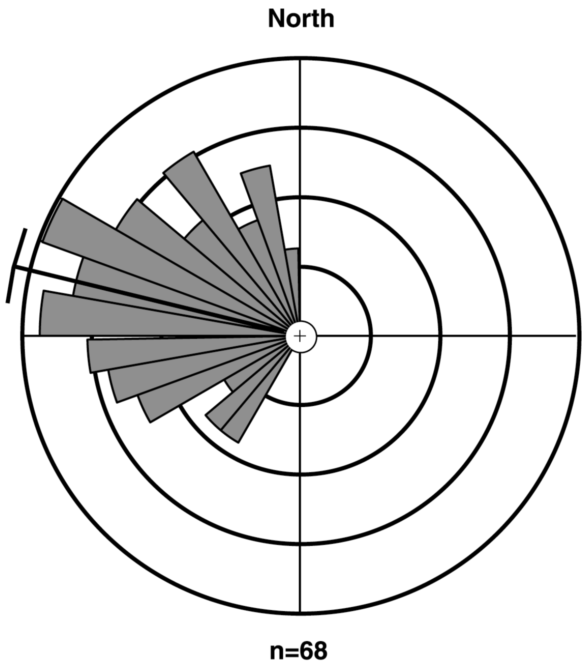

Figure 4. Rose diagram of paleocurrent data determined from gravel imbrication, channel orientation and cross stratification, indicating westward paleoflow from the Ortiz Mountains. Eight measurements were made in the basal Tanos Formation, which are not significantly different from paleocurrent directions measured in the overlying Blackshare Formation. Data compiled from geologic map of the San Felipe Pueblo NE quadrangle and measured sections (Cather et al., 2000). Data is combined into 10° intervals and the correlation coefficient (r) is 0.85.

Gravel in the Tanos and Blackshare formations are predominantly composed of monzanite and andesite porphyry with sparse (<2%) rounded quartzite, petrified wood, iron-stained sandstone, and hornfels (Fig. 6). The hornfels clasts are interpreted to be thermally metamorphosed sandstone and shale from the Cretaceous Mesaverde Group or Mancos Shale, which was intruded by the Oligocene Ortiz porphyry in the footwall of the La Bajada fault (S. Maynard, oral commun., 2000). Hornfels pebbles increase in abundance upsection in the interval above the measured section (Fig. 6). Sand in the Tanos and Blackshare formations is mostly lithic arkose and feldspathic litharenite, and differs from the nonquartzose lithic arkose of the subjacent Espinaso Formation (Large and Ingersoll, 1997; Kautz et al., 1981).

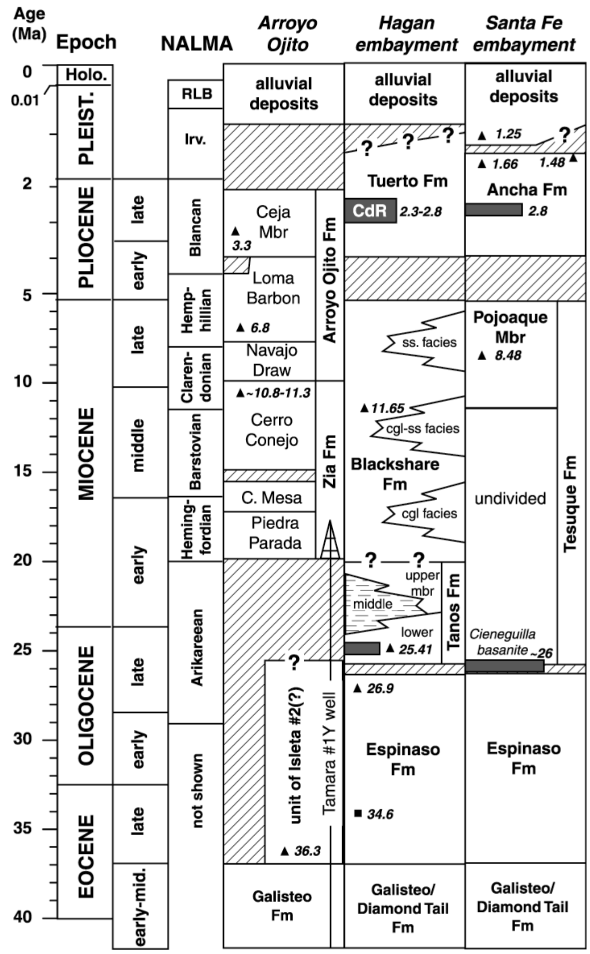

Figure 5. Correlation chart, illustrating correlations of selected Santa Fe Group units at Arroyo Ojito in the northwestern Calabacillas sub-basin (Connell et al., 1999), Hagan embayment (this study), and Santa Fe embayment (Koning et al., *this volume*). The Cerros del Rio volcanic field is denoted by CdR. Triangles are dates (in Ma) from primary volcanic units; boxes are recycled volcanic deposits; and shaded boxes are basaltic flows.

The stratigraphically lower Galisteo and Diamond Tail formations are arkosic to subarkosic and contain abundant quartz (Fig. 7). The abrupt increase in quartz content of the Tanos Formation, relative to the subjacent Espinaso Formation, suggest that older quartzose rocks were rapidly exposed on the footwall of an emerging La Bajada fault. The composition of the Tanos-Blackshare deposits relative to the Espinaso and Galisteo formations do not suggest a simple mixing of the Espinaso and Galisteo and Diamond Tail formations, principally because of the greater abundance of lithic fragments in Tanos-Blackshare succession. These data suggest contributions from other lithic sources, or possibly differences in grain size of the components analyzed among the various studies compiled for Figure 7 (Ingersoll et al., 1984); however, compositional differences are probably too great to be accounted for by grain size alone. The rather sharp increase in quartz content across the Espinaso-Tanos contact indicate a rather abrupt change in the composition of upland drainages, rather than progressive unroofing of the formerly extensive volcanic cover of the Espinaso Formation. Fairly rapid exhumation of the basin border along major faults, such as the nearby La Bajada fault, could account for this abrupt change in source lithology. Oligo-Miocene movement along this fault might have also resulted in the development of the angular unconformity recognized on dip-meter log of the Pelto Blackshare Federal #1.

IMPLICATIONS

The base of the Tanos Formation is younger than the >30.48 Ma onset of deposition of the Nambé Member of the Tesuque Formation reported by Smith (2000) in the Española basin. The Nambé Member is one of the oldest basin-fill units of the Santa Fe Group in the Española basin. The Tanos Formation, however, is older than the eolianites of the Piedra Parada Member of the Zia Formation, which overlie Eocene and Upper Cretaceous strata along the western margin of the Albuquerque Basin. The basal contact of the Piedra Parada Member contains scattered Oligocene volcanic clasts, indicating the presence of formerly extensive, but probably thin, Oligocene deposits prior to deposition of the Zia Formation. Many of these volcanic cobbles and pebbles have been sculpted into ventifacts (Tedford and Barghoorn, 1999), suggesting that this boundary was subjected to prolonged exposure and erosion on the hangingwall dip slope of the Calabacillas sub-basin. The presence of playa-lake mudstone and distal-piedmont sandstone on the hanging wall of the La Bajada fault in the Hagan embayment suggests that basin subsidence started with extensional block faulting, probably along the La Bajada fault. Definitive constraints on the onset of movement of the La Bajada fault are not available at this time, however the abrupt change in sand composition across the Espinaso-Tanos boundary and the lack of playa-lake deposits in the lower part of the Tesuque Formation in the Santa Fe embayment and southeastern Española basin suggests that the La Bajada fault was probably active since late Oligocene time.

Oligo-Miocene activity on the La Bajada fault does not support the two-stage model of development of the Albuquerque Basin (Large and Ingersoll, 1997; Ingersoll and Yin, 1993), which proposes that the northern portion of the Albuquerque Basin was a part of the Española basin (their Tesuque basin) during early Miocene time. In their model, the western margin of the basin was the depocenter until middle or late Miocene time, when they propose that a younger La Bajada fault and the range-bounding faults of the Sandia Mountains (Sandia-Rincon faults) began to move and establish the generally east-tilted character of the northern Albuquerque Basin.

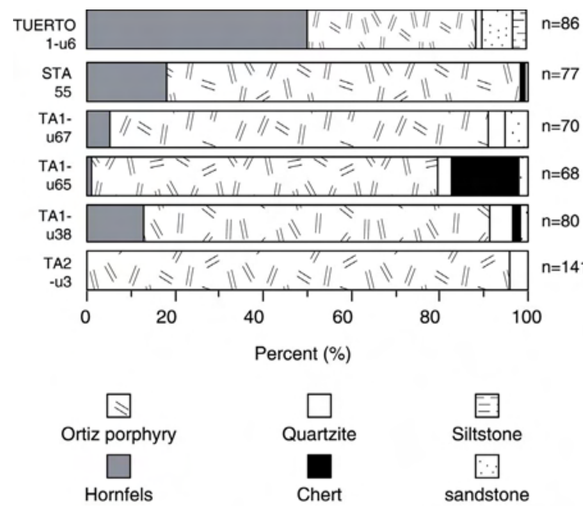

Figure 6. Stacked bar graph illustrating upsection variations (from bottom to top) in gravel composition in the Tanos, Blackshare, and Tuerto formations. Porphyritic hypabyssal intrusive and volcanic rocks derived from the Ortiz Mountains (Ortiz porphyry) dominate the basal Tanos Formation (TA2-u3). Gravel within the Blackshare Formation (TA1-u38, u65, u67, and STA 55) tends to become more diverse upsection. The Tuerto Formation (Tuerto 1-u6) is typically more heterolithic and contains a greater abundance of hornfels gravel than the underlying Blackshare Formation.

The eastward thickening of the Miocene Zia Formation (Connell et al., 1999) and preservation of probable Oligocene sedimentary rocks in the Tamara #1-Y well (Connell, Koning and Derrick, *this volume*), indicates that local stripping of Oligocene volcanic rocks occurred during late Oligocene or early Miocene time along the western margin of the basin. During this time, the Hagan embayment was receiving sediment. The unconformity between Tanos and Espinaso formations in the Pelto Blackshare Federal #1 indicates late Oligocene deformation in the Hagan embayment. The progressive decrease in stratal tilts upsection in the Tanos-Blackshare section indicates that deformation and concomitant sedimentation occurred after 25.4 Ma. Deformation of the Tanos-Blackshare succession is partially constrained by a 2.8 Ma (K/Ar date) on a basalt flow of the Cerros del Rio volcanic field (Bachman and Mehnert, 1978) that interfingers with hypabssyal-intrusive- and volcanic-bearing conglomerate correlated to the sub-horizontally bedded Tuerto Formation. The presence of late Pliocene basalt flows interbedded with the Tuerto Formation indicates that much of the stratal tilting in the Hagan embayment occurred prior to about 2.8 Ma. A paleomagnetic study of a 30.9 Ma mafic dike near the northern flank of the Sandia Mountains also indicates that much of the deformation and stratal tilt at the southern end of the Santo Domingo sub-basin occurred after 30.9 Ma (Lundahl and Geissman, 1999; see also Salyards et al., 1994; Brown and Golombek, 1985, 1986).

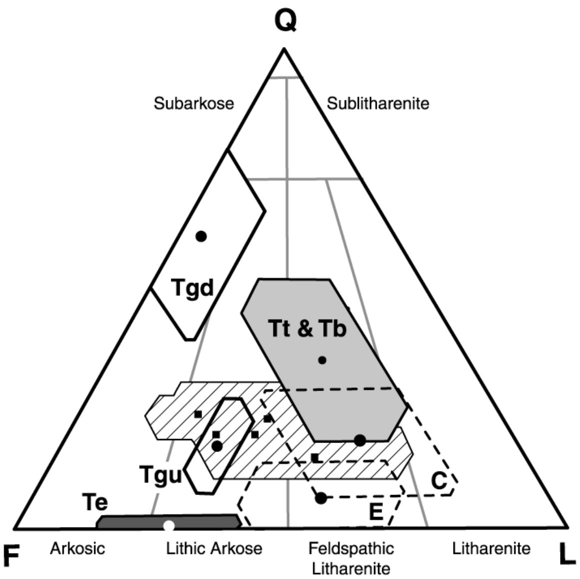

Figure 7. Sandstone petrographic data (means and fields of variations based on one standard deviation) for undivided Galisteo-Diamond Tail Formations (Tgd), upper Galisteo Formation (Tgu), Espinaso Formation (Te), Cordito (C) and Esquibel (E) petrofacies (Abiquiu Formation correlatives, see Large and Ingersoll, 1997), and undivided Tanos and Blackshare formations (Tt, Tb, shaded). The hachured area denotes the Abiquiu Formation and sub-unit lithofacies in the Abiquiu embayment (Moore, 2000). The Tanos and Blackshare formations contain more quartz than the underlying Espinaso Formation, but contain more lithic fragments than would be expected from mixing of Te and Tgd only. Data are summarized from Gorham (1979), Kautz et al. (1981), Large and Ingersoll (1997), and Moore (2000).

Estimates of stratal accumulation rates (not corrected for compaction) suggest that the basal Santa Fe Group accumulated between 69-83 m/m.y. along the western margin (Tedford and Barghoorn, 1999) during early through middle Miocene time, and about 72 m/m.y. along the eastern margin in the Hagan embayment during late Oligocene through Miocene time. These estimates are significantly lower than estimates of 600 m/m.y. for stratigraphically higher, late Miocene, playa-lake deposits of the Popotosa Formation in the southern part of the Albuquerque Basin (Lozinsky, 1988).

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STRATIGRAPHY OF THE TUERTO AND ANCHA FORMATIONS (UPPER SANTA FE GROUP), HAGAN AND SANTA FE EMBAYMENTS, NORTH-CENTRAL NEW MEXICO

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INTRODUCTION

Geologic studies and $^{40}Ar/^{39}Ar$ dating of subhorizontally bedded strata of the upper Santa Fe Group in the vicinity of the Santa Fe and Hagan embayments (Fig. 1) indicate that revision of the Ancha and Tuerto formations are necessary. The Ancha and Tuerto formations are included in the youngest strata of the Santa Fe Group, as defined by Spiegel and Baldwin (1963), and consist of broad, thin alluvial aprons of Plio-Pleistocene age derived from local uplands along the eastern margins of the Albuquerque and Española basins, Rio Grande rift, north-central New Mexico (Fig. 2). The Ancha Formation is composed mostly of granitic alluvium derived from the southeastern flank of the Sangre de Cristo Mountains and is located in the Santa Fe embayment, a west-sloping piedmont associated with the southwestern flank of the Sangre de Cristo Mountains. The Tuerto formation is composed mostly of porphryitic intrusive, volcanic, and hornfels rocks derived from eroding Oligocene, volcanic edifices of the Ortiz Mountains and Cerrillos Hills and is recognized mainly in the Hagan embayment (Fig. 1 and 2).

ANCHA FORMATION

The Ancha Formation was defined by Spiegel and Baldwin (1963, p. 45-50) for arkosic gravel, sand, and silt, inferred to be late Pliocene to Pleistocene in age, that lie with angular unconformity upon moderately tilted Tesuque Formation near Santa Fe, New Mexico. They established a partial type section for the Ancha Formation in Cañada Ancha, just north of the Santa Fe embayment (section CA, Fig. 3). The lower 3/5 of their type section, however, contains an 8.48±0.14 Ma tephra and is lithologically similar to the Pojoaque Member of the Tesuque Fm, which we correlate to most of their type section. The upper quarter of their type Ancha section contains basalt flows and basaltic tephra of the Cerros del Rio volcanic field, which was emplaced between 2.8 and 1.4 Ma (David Sawyer, personal commun., 2001), with the most voluminous activity occurring between 2.3-2.8 Ma (Woldegabriel et al., 1996; Bachman and Mehnert, 1978; Sawyer et al., 2001). Beneath the upper volcanic flows and volcaniclastics is 12-17(?) m of strata, containing 1-5% quartzite clasts, that is similar to a Pliocene deposit (unit Ta) mapped by Dethier (1997) that interfingers with Pliocene basalt tephra of the Cerros del Rio volcanic field.

The lower Cañada Ancha section contains hard, poorly sorted, grayish to brownish, pumiceous beds (Fig. 3). Although subhorizontal at the type section, these beds belong to a stratigraphic interval that continues 10 km along-strike to the north, where they are overlain by younger strata dipping up to 5° to the west (Fig. 2) (Koning and Maldonado, *in preparation*). Considering that the Ancha Formation is typically subhorizontal, the correlation of these pumiceous beds to strata that locally have been appreciably deformed supports our interpretation that the lower type Ancha section should be assigned to the subjacent Tesuque Formation. We do not assign these granite-bearing deposits (commonly >90% granitic clasts) to the Chamita Formation because paleocurrent data indicates general derivation from the east. In contrast, the more heterolithic, quartzite-bearing deposits of the Chamita Fm were derived from the north and northeast (*cf*. Galusha and Blick, 1971; Tedford and Barghoorn, 1993). Based on these interpretations and the presence of 8.48 Ma tephra, we propose that most of the Ancha Formation partial type section of Spiegel and Baldwin (1963) at Cañada Ancha is correlative to the Pojoaque Member of the Tesuque Formation.

STRATIGRAPHY OF MIDDLE AND UPPER PLEISTOCENE FLUVIAL DEPOSITS OF THE RIO GRANDE (POST-SANTA FE GROUP) AND THE GEOMORPHIC DEVELOPMENT OF THE RIO GRANDE VALLEY, NORTHERN ALBUQUERQUE BASIN, CENTRAL NEW MEXICO

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INTRODUCTION

Alluvial and fluvial deposits inset against Plio-Pleistocene deposits of the upper Santa Fe Group (Sierra Ladrones and Arroyo Ojito formations) record the development of the Rio Grande valley (Fig. 1) in the northern part of the Albuquerque basin since early Pleistocene time. These fluvial terrace deposits contain pebbly to cobbly sand and gravel with abundant rounded quartzite, subordinate volcanic, and sparse plutonic clasts derived from northern New Mexico. Although the composition of the gravel in these deposits is similar, they can be differentiated into distinct and mappable formation- and member-rank units on the basis of landscape-topographic position, inset relationships, soil morphology, and height of the basal contact above the Rio Grande as determined from outcrop and drillhole data (Table 1; Connell and Love, 2000). These fluvial deposits overlie, and locally interfinger with, alluvial deposits derived from paleo-valley margins and basin margin uplands (Fig. 2). Constructional terrace treads are not commonly preserved in older deposits, but are locally well preserved in younger deposits.

Kirk Bryan (1909) recognized two distinct types of ancestral Rio Grande deposits, his older Rio Grande beds (now called upper Santa Fe Group), and his younger, inset Rio Grande gravels (post Santa-Fe Group). Lambert (1968) completed the first detailed geologic mapping of the Albuquerque area and proposed the terms Los Duranes, Edith, and Menaul formations for prominent fluvial terrace deposits associated with the ancestral Rio Grande, however, these terms were not formally defined. Lambert (1968) correctly suggested that a higher and older unit (his Qu(?)g) may be an inset fluvial deposit of the ancestral Rio Grande (Tercero alto terrace of Machette, 1985).

We informally adopt three additional lithostratigraphic terms to clarify and extend Lambert's inset Rio Grande stratigraphy. We propose lithostratigraphic terms to these fluvial deposits principally to avoid confusion in the use of geomorphic terms, such as the primero, segundo, and tercero alto surfaces (Lambert, 1968), for lithologic units. Furthermore, these geomorphic (i.e., "-alto") terms were imported by Lambert (1968) for geomorphic surfaces described by Bryan and McCann (1936, 1938) in the upper Rio Puerco valley without careful comparison of soil-morphologic and geomorphic character of deposits within each drainage basin. Thus, these geomorphic terms may not be applicable in the Rio Grande valley without additional work to establish surface correlations across the Llano de Albuquerque, the interfluve between the Rio Grande and Rio Puerco valleys. Fluvial deposits discussed in this paper are, in increasing order of age, the Los Padillas, Arenal, Los Duranes, Menaul, Edith, and Lomatas Negras formations.

Although these inset ancestral Rio Grande units may be classified and differentiated allostratigraphically, we consider them as lithologic units of formation- and member rank that can be differentiated on the basis of bounding unconformities, stratigraphic position, and lithologic character.

Recent geologic mapping of the Albuquerque area (Cather and Connell, 1998; Connell, 1997, 1998; Connell et al., 1998; Love, 1997; Love et al., 1998; Smith and Kuhle, 1998; Personius et al., 2000) delineate a suite of inset fluvial deposits associated with the axial-fluvial ancestral Rio Grande. Inset terrace deposits record episodic incision and partial aggradation of the ancestral Rio Grande during Pleistocene and Holocene time. Lack of exposure and preservation of terrace deposits between Galisteo Creek and Las Huertas Creek hampers correlation to partially dated terrace successions at the northern margin of the basin and in White Rock Canyon (Dethier, 1999; Smith and Kuhle, 1998), southward into Albuquerque; however, correlation of these units using soil-morphology, landscape position, and stratigraphic relationships provide at least limited local constraints on the Rio Grande terrace stratigraphy.

Soil-morphologic information derived from profiles for fluvial and piedmont deposits are described on well preserved parts of constructional geomorphic surfaces (Connell, 1996). Carbonate morphology follows the morphogenetic classification system of Gile et al. (1966).

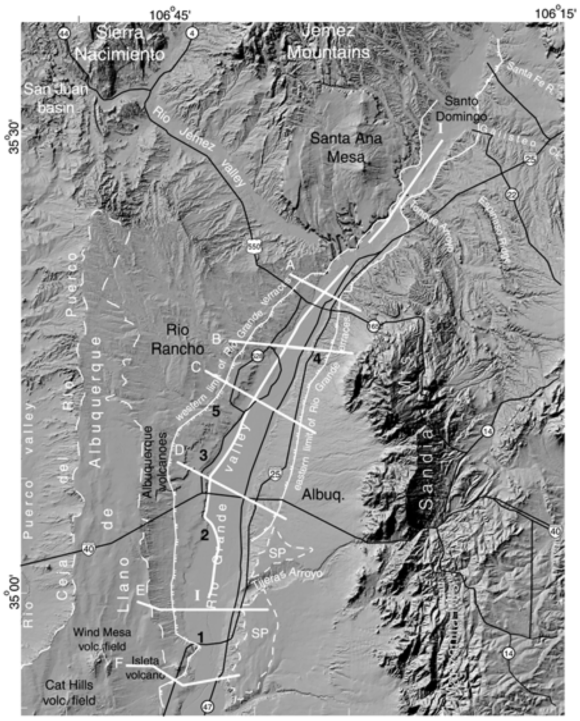

Figure 1. Shaded relief image of the northern part of the Albuquerque Basin (derived from U.S. Geological Survey 10-m DEM data) illustrating the approximate locations of terrace risers (hachured lines), the Sunport surface (SP), stratigraphic sections (1-5), and cross section lines (A-F).

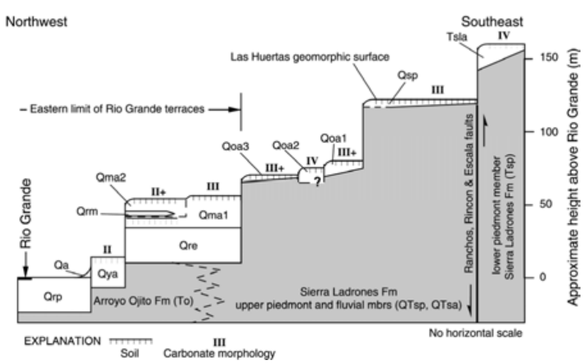

Figure 2. Block diagram of geomorphic relationships among entrenched post-Santa Fe Group deposits along the western piedmont of the Sandia Mountains and east of the Rio Grande valley (from Connell and Wells, 1999).

Lomatas Negras Formation

The highest and presumably oldest preserved Rio Grande terrace deposit in the Albuquerque-Rio Rancho area is informally called the Lomatas Negras Formation for Arroyo Lomatas Negras, where a buttress unconformity between this deposit and the underlying Arroyo Ojito Formation is exposed in the Loma Machete quadrangle (unit Qtag, Personius et al., 2000). The Lomatas Negras Formation is typically less than 16 ft (5 m) thick and consists of moderately consolidated and weakly cemented sandy pebble to cobble gravel primarily composed of subrounded to rounded quartzite, volcanic rocks, granite and sparse basalt (Fig. 3). This unit is discontinuously exposed along the western margin of the Rio Grande valley, where it is recognized as a lag of rounded quartzite-bearing gravel typically between about 215-245 ft (65-75 m) above the Rio Grande floodplain, which is underlain by the Los Padillas Formation (Fig. 4). The basal contact forms a low-relief strath cut onto slightly tilted deposits of the Arroyo Ojito Formation. The top is commonly eroded and is commonly overlain by middle Pleistocene alluvium derived from drainages heading in the Llano de Albuquerque. Projections of the base suggest that it is inset against early Pleistocene aggradational surfaces that define local tops of the Santa Fe Group, such as the Las Huertas and Sunport geomorphic surfaces (Connell et al., 1995, 1998; Connell and Wells, 1999; Lambert, 1968).

Correlative deposits to the south (Qg(?) of Lambert, 1968) underlie the late-middle Pleistocene (156±20 ka, Peate et al., 1996) Albuquerque Volcanoes basalt (Figs. 3-4). Projections of the Lomatas Negras Formation north of Bernalillo are limited by the lack of preserved terraces, so, we provisionally correlate these highest gravel deposits with the Lomatas Negras Formation, recognizing the possibility that additional unrecognized terrace levels and deposits may be present along the valley margins. Similar deposits are recognized near Santo Domingo (Qta1 of Smith and Kuhle, 1998), which contain the *ca*. 0.66 Ma Lava Creek B ash from the Yellowstone area of Wyoming. A gravel quarry in the Pajarito Grant (Isleta quadrangle) along the western margin of the Rio Grande valley exposes an ash within an aggradation succession of fluvial sand and gravel. This ash has been geochemically correlated to the Lava Creek B (N. Dunbar, 2000, personal commun.) It lies within pebbly to cobbly sand and gravels that grade upward into a succession of sand with lenses of pebbly sand. This unit is slightly lower, at ~46 m above the Rio Grande, than Lomatas Negras deposits to the north, suggesting the presence of additional unrecognized middle Pleistocene fluvial units, or intrabasinal faulting has down-dropped the Pajarito Grant exposures. The Lomatas Negras Formation is interpreted to be inset against the Sunport surface, which contains a 1.26 Ma ash near the top of this Santa Fe Group section in Tijeras Arroyo. These stratigraphic and geomorphic relationships indicate that the Lomatas Negras Formation was deposited between about 1.3 and 0.7 Ma.

Table 1. Summary of geomorphic, soil-morphologic, and lithologic data for ancestral Rio Grande fluvial, piedmont and valley border deposits, listed in increasing order of age.

Unit	Height above Rio Grande (m)	Thickness (m)	Carbonate Morphology	Geomorphic/stratigraphic position
Qrp	0	15-24	0	Lowest inset deposit; inner valley floodplain.
Qay	0-3	<21	0, I	Inset against Qpm; grades to Qrp.
Qra	15	3-6	II+	Primero alto surface, inset against Qrd.
Qam, Qpm	~65, eroded top	45	III	Alluvial deposits west of Rio Grande valley; Overlies Qrd.
Qrd	44-48	6-52	II+	Segundo alto surface, inset against Qre
Qpm	8-30	15-51	II+, III+	Piedmont deposits of Sandia Mts; east of Rio Grande valley; interfingers with Qrm.
Qrm	26-36	3	II+	Overlies Qpm and Qre; may be correlative to part of Qrd.
Qre	12-24, eroded top	3-12	not determined	Inset against Qrl, inset by Qrd; underlies Qpm with stage III + carbonate morphology.
Qao, Qpo	~100, eroded top	<30	III to IV	Overlies Qrl; inset by Qpm and Qre.
Qrl	~46-75, eroded top	5-20	III, eroded	Inset against Sunport surface. Contains ash correlated to the Lava Creek B.
Las Huertas	~120	---	III+	Local top of Sierra Ladrones Formation
SP	~95	---	III+	Sunport surface of Lambert (1968): youngest Santa Fe Group constructional basin-floor surface.

Edith Formation

The Edith Formation is a 10-40 ft (3-12 m) thick deposit that typically comprises a single upward fining sequence of basal gravel and overlying sandy to muddy floodplain deposits. The Edith Formation serves as a useful and longitudinally extensive marker along the eastern margin of the Rio Grande valley, between Albuquerque and San Felipe Pueblo, New Mexico. This fluvial deposit can be physically correlated across 33 km, from its type area in Albuquerque (Lambert, 1968, p. 264-266 and p. 277-280), to near Algodones, New Mexico (Lambert, 1968; Connell et al., 1995; Connell, 1998, 1997; and Cather and Connell, 1998). The Edith Formation is a poorly to moderately consolidated, locally cemented deposits of pale-brown to yellowish-brown gravel, sand and sandy clay that forms laterally extensive outcrops along the inner valley escarpment of the Rio Grande. Commonly recognized as an upward-fining succession of a 7-26 ft (2-8 m) thick, basal quartzite-rich, cobble gravel that grades up-section into a 13-32 ft (4-10 m) thick succession of yellowish-brown sand and reddish-brown mud. The upper contact is locally marked by a thin, white diatomite between Sandia Wash and Bernalillo. Gravel contains ~30% rounded quartzite and ~40% volcanic rocks with subordinate granite, metamorphic, and sandstone clasts, and sparse, rounded and densely welded Bandelier Tuff (Connell, 1996). The Edith Formation unconformably overlies tilted sandstone of the Arroyo Ojito and Sierra Ladrones formations and is overlain by piedmont alluvium derived from the Sandia Mountains (Fig. 5). Where the top of the Edith Formation is preserved, it typically contains weakly developed soils with Stage I carbonate morphology. This weak degree of soil development suggests that deposition of piedmont and valley border fan sediments occurred shortly after deposition of the Edith Formation.

The Edith Formation contains Rancholabrean fossils, most notably *Bison*, *Mastodon*, *Camelops*, and *Equus* (Lucas et al., 1988). Lambert (1968) considered the Edith Formation to represent a late Pleistocene terrace deposited during the latest Pleistocene glacial events. Soils developed in these piedmont deposits exhibit moderately developed Bt and Btk horizons with moderately thick clay films and Stage III+ carbonate morphology, suggesting a middle Pleistocene age for these deposits (Connell, 1996; Connell and Wells, 1999).

The base of the Edith Formation forms a prominent strath that lies about 40-80 ft (12-24 m) above the Rio Grande floodplain and is about 30 m higher than the base of the Los Duranes Formation (Connell, 1998). The elevation of this basal strath is lower than the base of the Lomatas Negras Formation suggesting that the Edith Formation is inset against

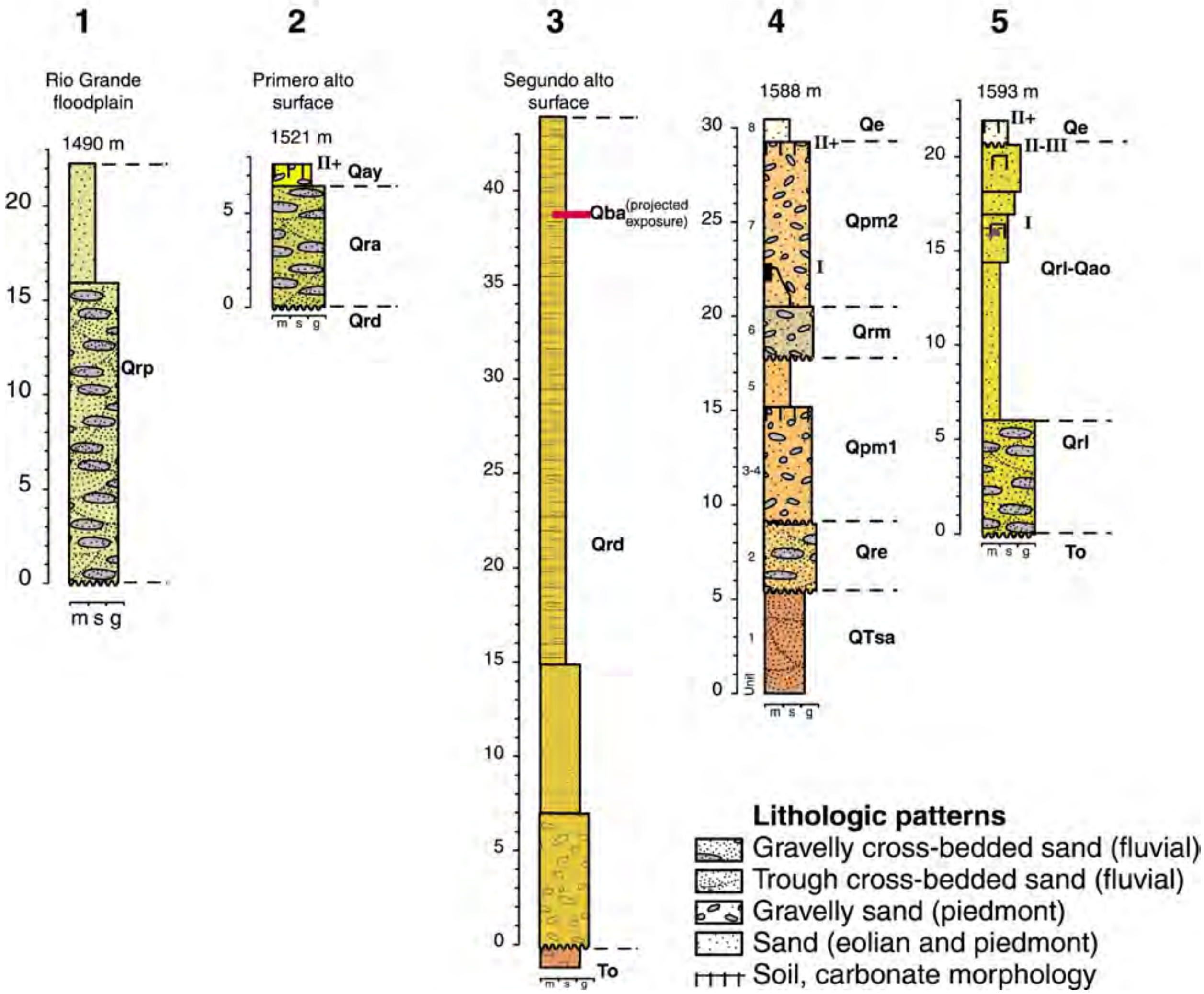

Figure 3. Stratigraphic and drillhole sections of Pleistocene fluvial deposits of the ancestral and modern Rio Grande along the Rio Grande valley: 1) Los Padillas Formation at the Black Mesa-Isleta Drain piezometer nest; 2) Arenal Formation at Efren quarry (modified from Lambert, 1968; Machette et al., 1997); 3) Los Duranes Formation at the Sierra Vista West piezometer nest (data from Chamberlin et al., 1998); 4) Edith and Menaul formations at Sandia Wash (Connell, 1996); and 5) Lomatas Negras Formation at Arroyo de las Calabacillas and Arroyo de las Lomatas Negras.

the Lomatas Negras Formation. A partially exposed buttress unconformity between eastern-margin piedmont alluvium and upper Santa Fe Group deposits marks the eastern extent of this unit. This unconformity is locally exposed in arroyos between Algodones and Bernalillo, New Mexico.

Lambert (1968) recognized the unpaired nature of terraces in Albuquerque, but assigned the Edith Formation to the topographically lower primero alto terrace, which is underlain by the Los Duranes Formation in SW Albuquerque. Lambert (1968) correlated the Edith Formation with the primero alto terrace, and therefore interpreted it to be younger than the Los Duranes Formation. The primero alto terrace is the lowest fluvial-terrace tread in SW Albuquerque and is underlain by rounded pebbly sandstone that is inset against the Los Duranes Formation. Soils on the primero alto terrace are weakly developed (stage I to II+ carbonate morphology, Machette et al., 1997) compared to piedmont deposits overlying the Edith Formation. Therefore, it is likely that the gravels underlying the primero alto terrace are probably much younger than the Edith Formation. Therefore, if the Edith Formation is older than the Los Duranes Formation (see below), it was deposited prior to about 100-160 ka.

The Edith Formation may correlate to fluvial terrace deposits near Santo Domingo Pueblo (Smith and Kuhle, 1998). Deposits at Santo Domingo Pueblo are approximately 30-m thick and about 30-35 m above the Rio Grande (Qta3 of Smith and Kuhle, 1998). The lack of strongly developed soils between the Edith Formation and interfingering middle Pleistocene piedmont alluvium suggests that the Edith Formation was deposited closer in time to the Los Duranes Formation. Thus, the Edith Formation was deposited between 0.66 and 0.16 Ma, and was probably laid down during the later part of the middle Pleistocene.

Los Duranes Formation

The Los Duranes Formation of Lambert (1968) is a 40-52 m fill terrace consisting of poorly to moderately consolidated deposits of light reddish-brown, pale-brown to yellowish-brown gravel, sand, and minor sandy clay derived from the ancestral Rio Grande and tributary streams. The base typically buried by deposits of the Rio Grande floodplain (Los Padillas Formation) in the Albuquerque. The basal contact forms a low-relief strath approximately 20 ft (6 m) above the Rio Grande floodplain near Bernalillo, New Mexico (Figs. 3-4), where the Los Duranes Formation is eroded by numerous arroyos and is about 20-23 ft (6-7 m) thick. The basal contact is approximately 100 ft (30 m) lower than the base of the Edith Formation. The terrace tread on top of the Los Duranes Formation (~42-48 m above the Rio Grande) is about 12-32 m higher than the top of the Edith Formation. Geologic mapping and comparison of subsurface data indicate that the base of the Edith Formation is about 20-25 m higher than the base of Los Duranes Formation, suggesting that the Los Duranes is inset against the Edith. Just north of Bernalillo, New Mexico, deposits correlated to the Los Duranes Formation (Connell, 1998) contain the Rancholabrean mammal *Bison latifrons* (Smartt et al., 1991, SW1/4, NE1/4, Section 19, T13N, R4E), which supports a middle Pleistocene age. The Los Duranes Formation is also overlain by the 98-110 ka Cat Hills basalt (Maldonado et al., 1999), and locally buries flows of the 156±20 ka (Peate et al., 1996) Albuquerque volcanoes basalt. Thus deposition of the Los Duranes Formation ended between 160-100 ka, near the end of the marine oxygen isotope stage 6 at about 128 ka (Morrison, 1991).

Near Bernalillo, the basal contact of the Los Duranes(?) Formation, exposed along the western margin of the of the Rio Grande valley, is approximately 30 m lower than the basal contact of the Edith Formation, which is well exposed along the eastern margin of the valley. This western valley-margin fluvial deposit was originally assigned to the Edith Formation by Smartt et al. (1991), however, these are interpreted to be younger inset deposits that are likely correlative to the Los Duranes Formation (Connell, 1998; Connell and Wells, 1999).

The terrace tread (top) of the Los Duranes Formation is locally called the segundo alto surface in the Albuquerque area (Lambert, 1968; Hawley, 1996), where it forms a broad constructional surface west of the Rio Grande. Kelley and Kudo (1978) called this terrace the Los Lunas terrace, near Isleta Pueblo, however, we support the term Los Duranes Formation as defined earlier by Lambert (1968). The Los Duranes Formation represents a major aggradational episode that may have locally buried the Edith Formation; however, the Edith Formation could also possibly mark the base of the aggrading Los Duranes fluvial succession.

Menaul Formation(?)

The Menaul Formation of Lambert (1968) is generally less than 10 ft (3 m) thick and overlies interfingering piedmont deposits that overlie the Edith Formation. The Menaul Formation consists of poorly consolidated deposits of yellowish-brown pebble gravel and pebbly sand derived from the ancestral Rio Grande. Rounded quartzite pebbles that are generally smaller in size than pebbles and cobbles in the Edith Formation. The Menaul gravel forms discontinuous, lensoidal exposures along the eastern margin o the Rio Grande valley. The basal contact is approximately 85-118 ft (26-36 m) above the Rio Grande floodplain. The Menaul Formation is conformably overlain by younger, eastern-margin piedmont alluvium exhibiting Stage II+ carbonate morphology, and is inset by younger stream alluvium that exhibits weakly developed soils, suggesting a late Pleistocene age of deposition.

Soils on piedmont deposits overlying the Menaul are generally similar to the Los Duranes Formation; however, differences in parent material texture make soil-based correlations somewhat ambiguous. Similarities in height above the Rio Grande and soil development on the Los Duranes Formation and the Menaul Formation suggest that these two units may be correlative. Thus, the Menaul Formation may be temporally correlative to the Los Duranes Formation, and is likely a member of this unit. These units may be associated with an aggradational episode, possibly associated with aggradation of the Los Duranes, middle Pleistocene piedmont alluvium. The Edith Formation may represent the base of a Los Duranes-Menaul aggradational episode during the late-middle Pleistocene. The base Edith Formation is consistently higher than the base of the Los Duranes Formation, suggesting that the Edith is older; however, definitive crosscutting relationships have not been demonstrated.

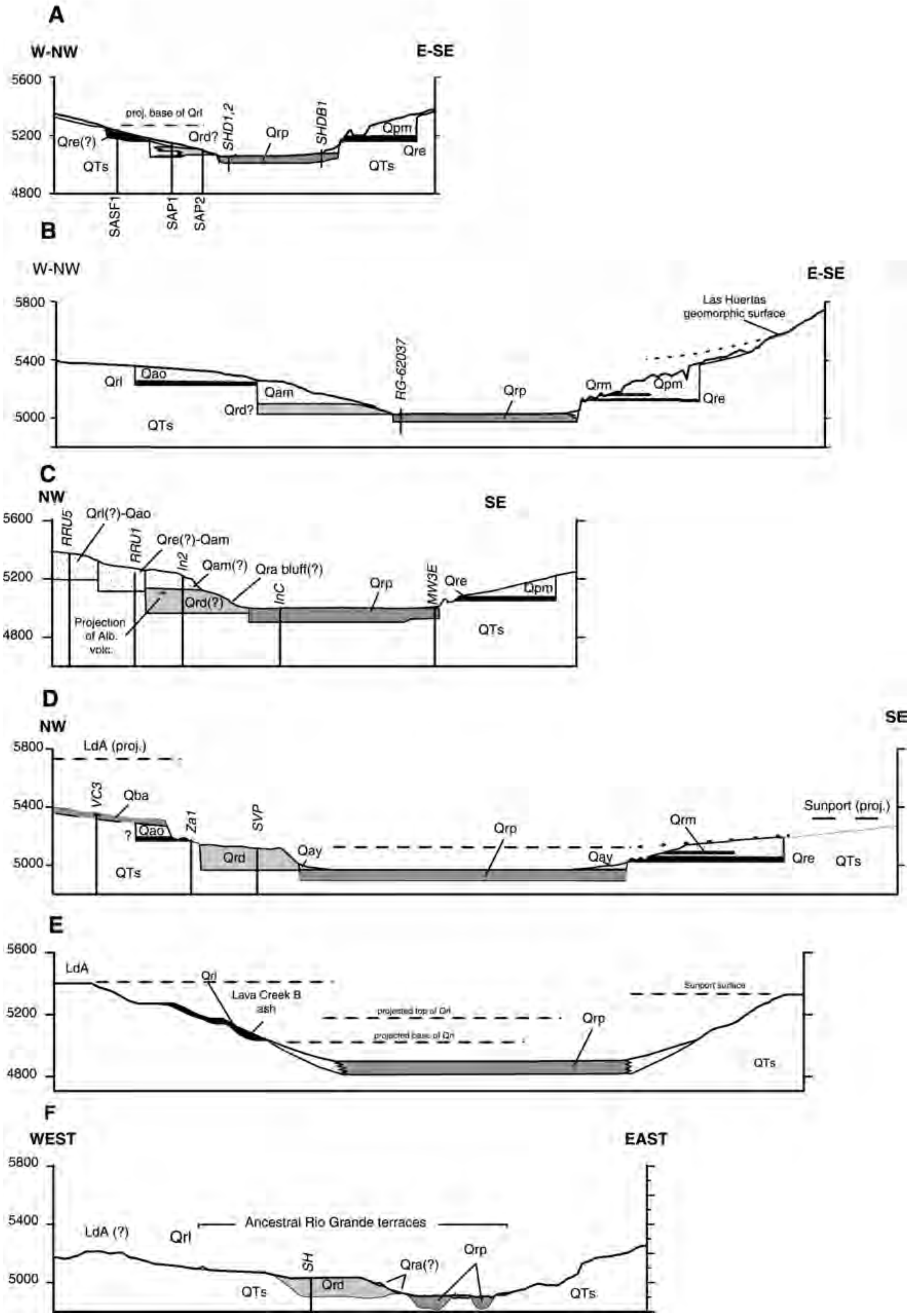

Figure 4. Simplified geologic cross sections across the Rio Grande valley, illustrating inset relationships among progressively lower fluvial deposits. Letters indicate location of profiles on Figure 1 and elevations of cross sections are in feet above mean sea level. See Table 1 for description of symbols. Unit QTs denotes upper Santa Fe Group deposits.

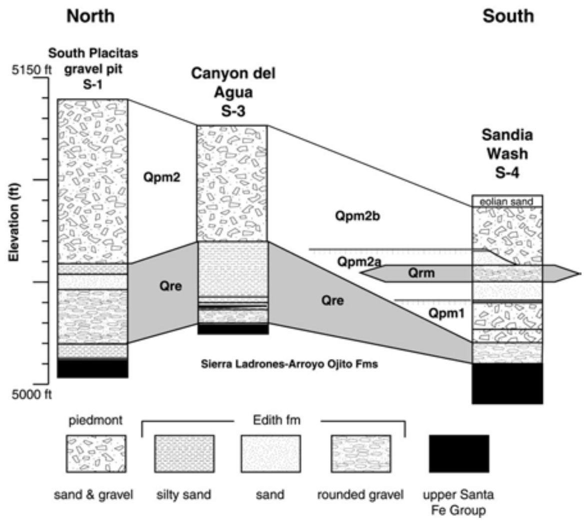

Figure 5. Stratigraphic fence of Edith Formation and piedmont deposits exposed along eastern margin of the Rio Grande valley, between Sandia Wash and highway NM-165, illustrating stratigraphic relationships among fluvial-terrace and piedmont deposits.

Arenal Formation

The lowest preserved terrace deposit is the Arenal Formation, which was named for exposures just west of the Arenal Main Canal in SW Albuquerque (Connell et al., 1998). The Arenal Formation is 3-6 m thick and is inset against the Los Duranes Formation. The Arenal Formation consists of poorly consolidated deposits of very pale-brown to yellow sandy pebble to cobble gravel recognized along the northwestern margin of the Rio Grande inner valley. Gravel clasts are primarily rounded quartzite and subrounded volcanic rocks (welded tuff and rare pumice) with minor granite. Soil development is very weak, with Stage I to II+ carbonate morphology (Machette et al., 1997; Machette, 1985). The top of the Arenal Formation is the primero alto surface of Lambert (1968), which is 15-21 m above the Rio Grande. This deposit is not correlative to the Edith Formation as originally interpreted by Lambert. This unit is interpreted to have been deposited during late Pleistocene time, probably between about 71-28 ka.

Los Padillas Formation

The Las Padillas Formation underlies the modern Rio Grande valley and floodplain and is interpreted to represent the latest incision/aggradation phase of the Rio Grande, which was probably deposited during latest Pleistocene-Holocene time. The Rio Grande floodplain (inner valley) ranges 3-8 km in width in most places and occupies only a portion of the 10-12 km maximum width of the entire ancestral Rio Grande systems tract of the Sierra Ladrones Formation (Connell, 1997, 1998; Connell et al., 1995; Maldonado et al., 1999; Smith and Kuhle, 1998). The top comprises the modern floodplain and channel of the Rio Grande. The Los Padillas Formation is 15-29 m thick and consists of unconsolidated to poorly consolidated, pale-brown, fine- to coarse-grained sand and rounded gravel with subordinate, discontinuous, lensoidal interbeds of fine-grained sand, silt, and clay derived from the Rio Grande. This unit is recognized in drillholes and named for deposits underlying the broad inner valley floodplain near the community of Los Padillas in SW Albuquerque (Connell et al., 1998; Connell and Love, 2000). Drillhole data indicate that the Los Padillas Formation commonly has a gravelly base and unconformably overlies the Arroyo Ojito Formation. This basal contact is locally cemented with calcium carbonate. The Los Padillas Formation is overlain, and interfingers with, late Pleistocene to Holocene valley border alluvial deposits derived from major tributary drainages.

Because this unit has not been entrenched by the Rio Grande, no age direct constraints are available for the base of the alluvium of the inner valley in the study area. This deposit underlies a continuous and relatively broad valley floor that extends south from the Albuquerque basin through southern New Mexico, where radiocarbon dates indicate aggradation of the inner valley by early Holocene time (Hawley and Kottlowski, 1969; Hawley et al., 1976). The base of the Los Padillas Formation was probably cut during the last glacial maximum, which is constrained at ~15-22 ka in the neighboring Estancia basin, just east of the Manzano Mountains. (Allen and Anderson, 2000). Thus, the inner valley alluvium was probably incised during the latest Pleistocene and aggraded during much of Holocene time. Near the mouth of Tijeras Arroyo, charcoal was recovered from about 2-3 m below the top of a valley border fan that prograded across the Los Padillas Formation and forms a broad valley border fan than has pushed the modern Rio Grande to the western edge of its modern (inner) valley. This sample yielded a radiocarbon date of about 4550 yrs. BP (Connell et al., 1998), which constrains the bulk of deposition of the Los Padillas Formation to middle Holocene and earlier.

EVOLUTION OF THE RIO GRANDE VALLEY

Santa Fe Group basin-fill deposits of the ancestral Rio Grande generally differ in the scale and thickness relative to younger inset deposits, which were deposited in well defined valley. During widespread aggradation of the basin (Santa Fe Group time), the ancestral Rio Grande intimately interfingered with piedmont deposits derived from rift-margin uplifts, such as the Sandia Mountains

(Connell and Wells, 1999; Maldonado et al., 1999). Field and age relationships in the near Santa Ana Mesa also indicate that the ancestral Rio Grande also interfingered with fluvial deposits correlated with the Arroyo Ojito Formation (Cather and Connell, 1998). During development of the Rio Grande valley (post-Santa Fe Group time), the Rio Grande cut deeply into older basin-fill, typically leaving large buttress unconformities between inset deposits and older basin fill of the upper Santa Fe Group (Fig. 8).

Younger late Pleistocene-Holocene alluvial deposits are commonly confined in arroyo channels cut into older piedmont deposits east of the Rio Grande valley. These deposits commonly form valley border alluvial fans along bluffs cut by a meandering Rio Grande. These fans commonly prograde across floodplain and channel deposits in the inner valley. The present discharge is inadequate to transport sediment out of the valley. The presence of progressively inset fluvial deposits along the margins of the modern valley indicates that episodes of prolonged higher discharge were necessary to flush sediment and erode the valley. Such episodes must have occurred prior to aggradation of valley fills, such as these fluvial terrace deposits.

Progradation of middle Holocene tributary valley border fans across the modern Rio Grande floodplain suggests that deposition of tributary and piedmont facies occurred during drier (interglacial) conditions. Deposition of fluvial terraces in semi-arid regions probably occurred during the transition from wetter to drier climates (Schumm, 1965; Bull, 1991). The lack of strong soils between the terrace deposits of the ancestral Rio Grande and piedmont and valley border deposits suggests that piedmont and valley border deposition occurred soon after the development of major fluvial terrace deposits.

Age constraints for the Los Duranes Formation indicate that aggradation of fluvial deposits occurred near the end of glacial periods. If we extrapolate ages based on this model of terrace development, then we can provide at least a first order approximation for ages of other poorly dated terrace deposits throughout the study area (Fig. 6). The age of the Edith Formation is still rather poorly constrained. The Edith Formation is Rancholabrean in age and older than the Los Duranes Formation, suggesting that the Edith may have been deposited sometime during MOIS 8, 10, or 12. The lack of strongly developed soils on the top of the Edith Formation suggests that deposition of this unit occurred closer in time to the Los Duranes Formation.

Correlation of these deposits and provisional age constraints indicate that the ancestral positions of the Rio Grande have been modified by tectonic activity (Fig. 7). Most notably, the Edith Formation, which forms a nearly continuous outcrop band from Albuquerque just south of San Felipe, New Mexico, is faulted. The Bernalillo fault displaced this deposit by about 7 m down to the west near Bernalillo (Connell, 1996). Between cross sections B-B' and C-C' of Figure 4, the basal contact of the Edith Formation is down-dropped to the south by about 15 m by the northwest-trending Alameda structural zone. This decrease in height above local base level is also recognized by a change in stratigraphic positions relative to piedmont deposits to the east. Younger piedmont alluvium (Qay, Fig. 2) is typically found overlying the Edith Formation south of the Alameda structural zone (East Heights fault zone), a zone of flexure or normal faults that displace the Edith Formation in a down-to-the-southwest sense. North of the Alameda zone, tributary stream deposits are inset against the Edith Formation and are found in well defined valleys (*see map by* Connell, 1997).

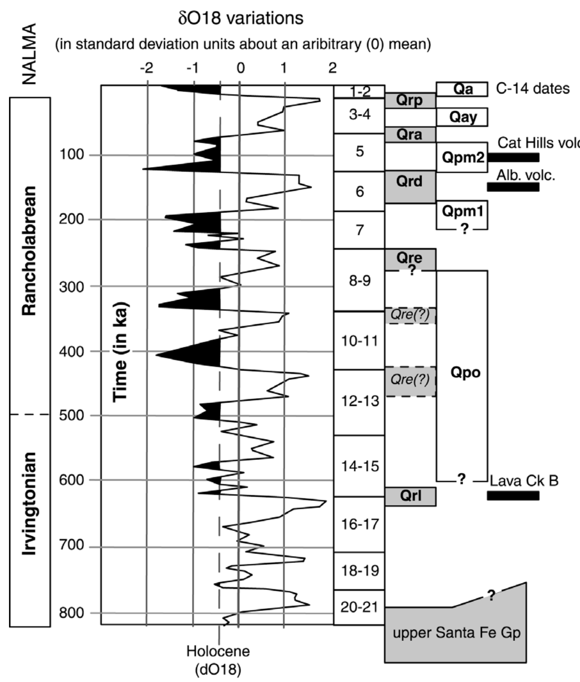

Figure 6. Correlation of fluvial deposits inferred ages. The age of the top of the Los Duranes Formation is constrained by middle and late Pleistocene basalt flows. The Lomatas Negras Formation contains the middle Pleistocene Lava Creek B ash. The Edith Formation contains middle-late Pleistocene Rancholabrean fossils and is older than the Los Duranes Formation, however, its precise age is not well constrained. Younger deposits are constrained by a radiocarbon date of 4550 yr. BP. The Edith Formation is interpreted to be older than the Los Duranes Formation and precise than the Lomatas Negras Formation. More precise age control has not been established and the Edith Formation could have been deposited during different climatic episodes.

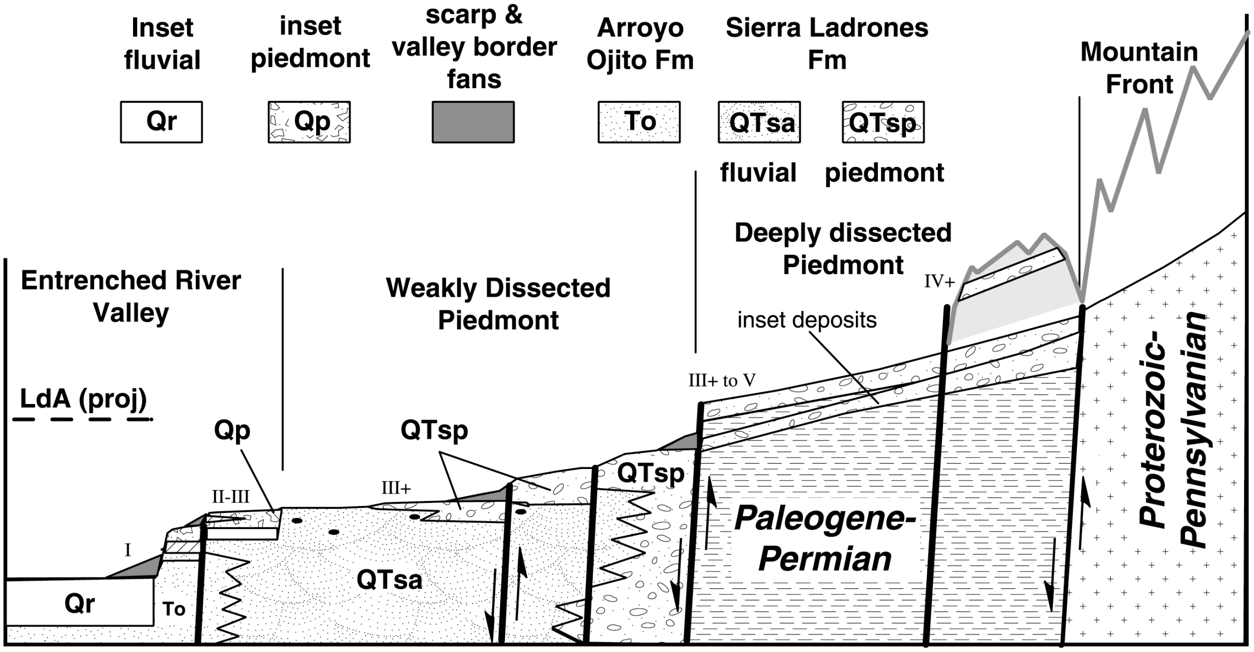

Figure 7. Generalized cross section across part of the piedmont of the Sandia Mountains, illustrating interfingering relationships among aggrading sediments of the upper Santa Fe Group, and inset post-Santa Fe Group deposits. Pedogenic carbonate morphology of constructional deposit surfaces is indicated by roman numerals that indicate the morphogenetic stage of soil development.

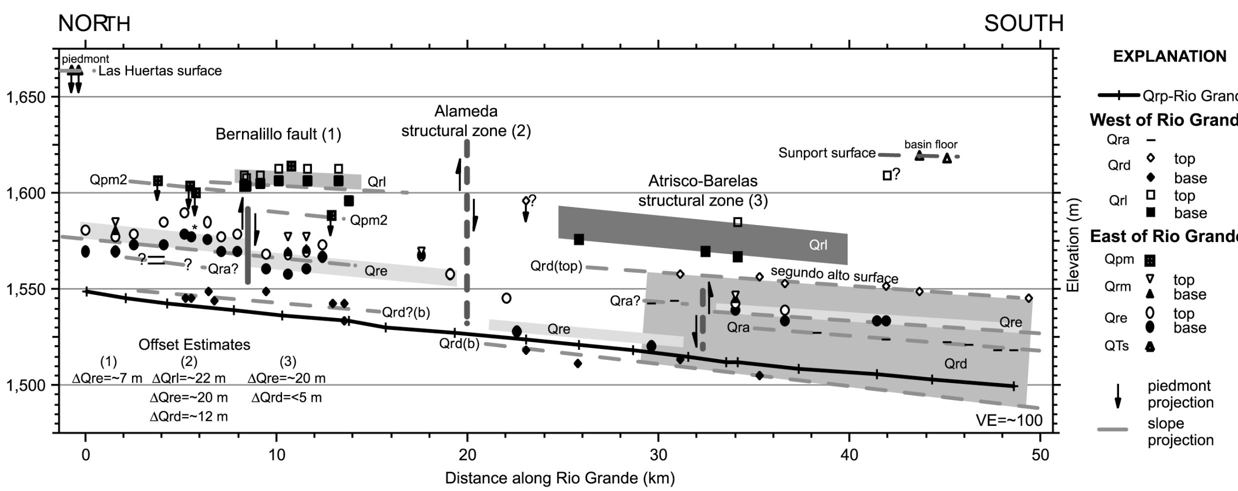

Figure 8. Longitudinal profile along Rio Grande, illustrating inset relationships among ancestral Rio Grande terraces and early Pleistocene aged constructional surfaces that locally mark the end of Santa Fe Group deposition (Las Huertas and Sunport geomorphic surfaces). The Edith and Los Duranes formations are deformed by northwest-trending faults that alter the elevation of the basal contact of these two units.

During late Pliocene time, the ancestral Rio Grande formed an axial-river that flowed within a few kilometers of the western front of the Sandia Mountains (Fig. 9a). During early Pleistocene time, between about 1.3-0.7 Ma, the Rio Grande began to entrench into the basin fill, just west of the modern valley. Piedmont deposits prograded across much of the piedmont-slope of the Sandia Mountains and buried these basin-fill fluvial deposits (Fig. 9b). During middle Pleistocene time, the Rio Grande episodically entrenched into older terrace deposits and basin-fill of the Santa Fe Group. These episodes of entrenchment were followed by periods of partial backfilling of the valley and progradation of piedmont and valley border deposits (Figs. 9c and 9d). The latest episode of entrenchment and partial backfilling occurred during the latest Pleistocene, when middle Pleistocene tributary deposits were abandoned during entrenchment, and valleys partially aggraded later during latest Pleistocene and Holocene time (Fig. 9e).

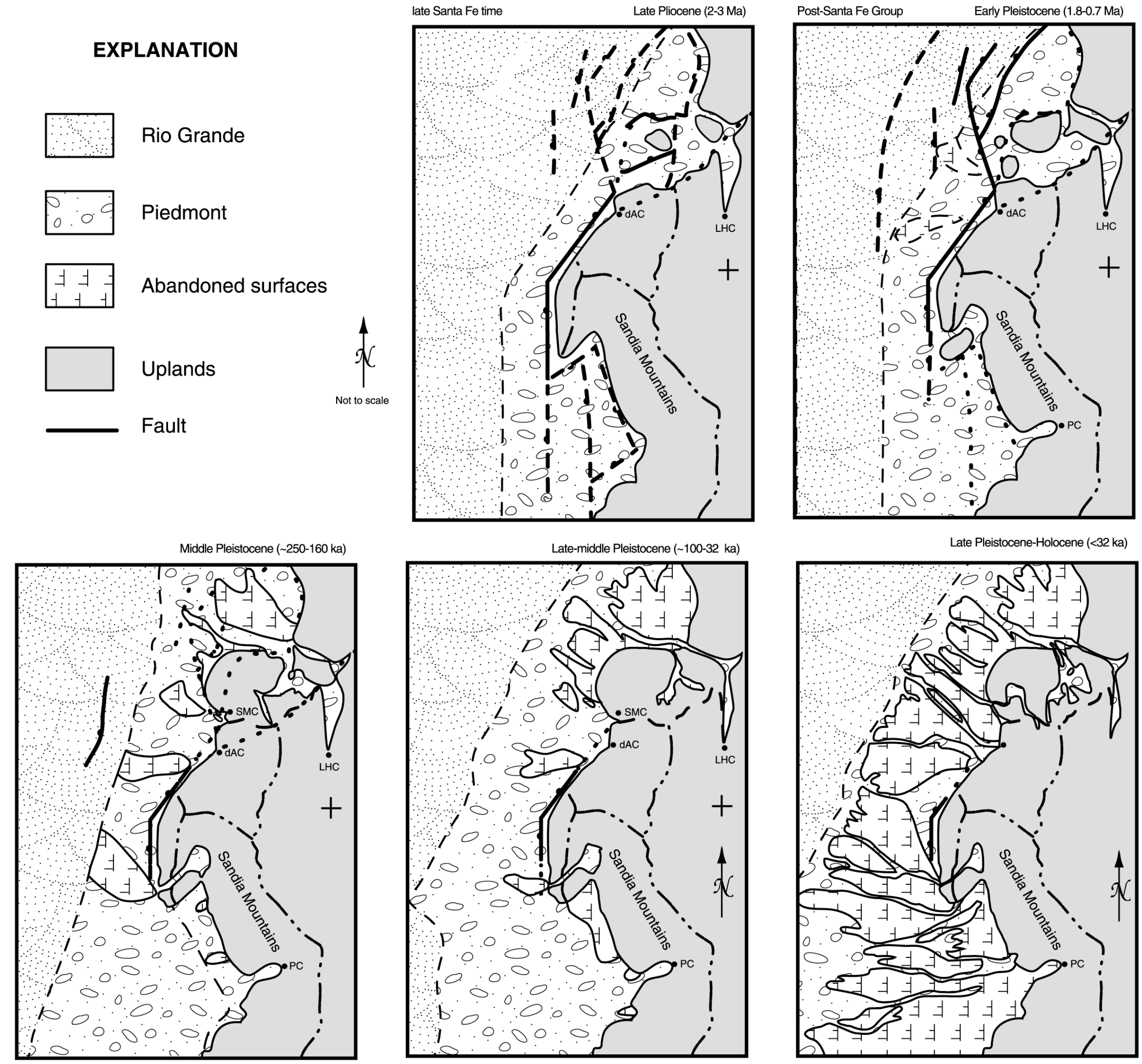

Figure 9. Paleogeographic maps of the latest phase of basin filling of the Santa Fe Group, and Pleistocene development of the Rio Grande valley (modified from Connell, 1996). Las Huertas Creek (LHC), Pino Canyon (PC), and del Agua Canyon (dAC) are shown for reference.

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PRELIMINARY INTERPRETATION OF CENOZOIC STRATA IN THE TAMARA NO. 1-Y WELL, SANDOVAL COUNTY, NORTH-CENTRAL NEW MEXICO

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INTRODUCTION

The Tamara #1-Y well (API 30-043-20934) is a wildcat oil-test that was drilled in northwest of Rio Rancho, New Mexico (Sec. 3, T13N. R2E., Bernalillo NW quadrangle; UTM: N: 3,916,580 m, E: 344,615 m, Zone 13, NAD83) in 1995 by Davis Petroleum Co. The Tamara well was spudded into the Ceja Member (upper Arroyo Ojito Formation of Connell et al., 1999), near the northern edge of the Llano de Albuquerque, at an elevation of about 1865 m (6120 ft) above mean sea level. The well was drilled between December 1, 1995 and January 16, 1996. According to the scout ticket, the well stopped in the Triassic Chinle Group at a depth of 8723 ft (2659 m) below land surface (bls); however, this correlation was not confirmed in this study.

Washed cuttings from this well are archived at the New Mexico Bureau of Mines and Mineral Resources (NMBMMR Library #46,891), in Socorro, New Mexico. The well was cased to 329 m bls and cuttings were collected below 360 m bls; a number of intervals were not sampled, probably due to loss of circulation during drilling. Cuttings were visually evaluated using a sample preparation microscope on available intervals between 360 and 2015 m bls. Detrital modes of sand were determined on medium-grained sand (400 points per sample) at eight sample intervals in the Cenozoic section (Appendix 1) and normalized to the modified Gazzi-Dickinson method (Table 1).

Borehole geophysical logs, archived at the NMBMMR, are available below 2103 m (6900 ft) bls. Digital borehole geophysical logs of natural gamma ray and induction resistivity of the entire well were obtained from the Denver Earth Resources Library and the U.S. Geological Survey.

The Tamara well was spudded into the Santa Fe Group near its local top on the Llano de Albuquerque, near La Ceja (Rincones de Zia of Galusha, 1966; and Tedford, 1981) and fully penetrated the Cenozoic section, thus providing an opportunity to document variations in thickness of the Santa Fe Group across intrabasinal faults of the northwestern Albuquerque Basin. This area has been mapped in detail (Fig. 1) and provides additional stratigraphic control for this well.

Table 1. Recalculated detrital mode parameters, normalized to percent, of point counts (Appendix 1) for medium-grained sand from the Tamara well using the modified Gazzi-Dickinson method (Dickinson, 1970). Volcanic grains comprise nearly all of the lithic parameters. Units are the Cerro Conejo (Tzc) and undivided Chamisa Mesa-Piedra Parada (Tzm) members of the Zia Formation, unit A and B, and Galisteo Formation (Tg). The Galisteo Formation contains volcanic grains, probably from contamination by caving of upper volcanic-bearing units.

Interval ft, bls	Unit	Modified Gazzi-Dickinson Method		
		%Q	%F	%L
1390-1420	Tzc	68	18	15
2620-2650	Tzm	67	17	16
3970-4000	B	64	22	15
4150-4180	B	72	18	10
5020-5050	A	68	19	14
5230-5260	A	70	19	11
5290-5320	Tg	63	25	12
5410-5440	Tg	67	21	12

Another purpose in studying the Tamara well was to document the presence of older or pre-Santa Fe Group Cenozoic strata near the basin margin, such as the Abiquiu, Popotosa, or Tanos formations, or the unit of Isleta #2. The Abiquiu Formation contains mostly epiclastic sediments derived from the rhyolitic Latir volcanic field in northern New Mexico (Smith, 1995; Moore, 2000). Much of the Abiquiu Formation was deposited between *ca*. 18-27 Ma (Moore, 2000; Tedford and Barghoorn, 1993) and is temporally correlative to the Piedra Parada and Chamisa Mesa members of the Zia Formation (Fig. 2). The southernmost mapped location of the Abiquiu Formation is near the town of Gilman, New Mexico,

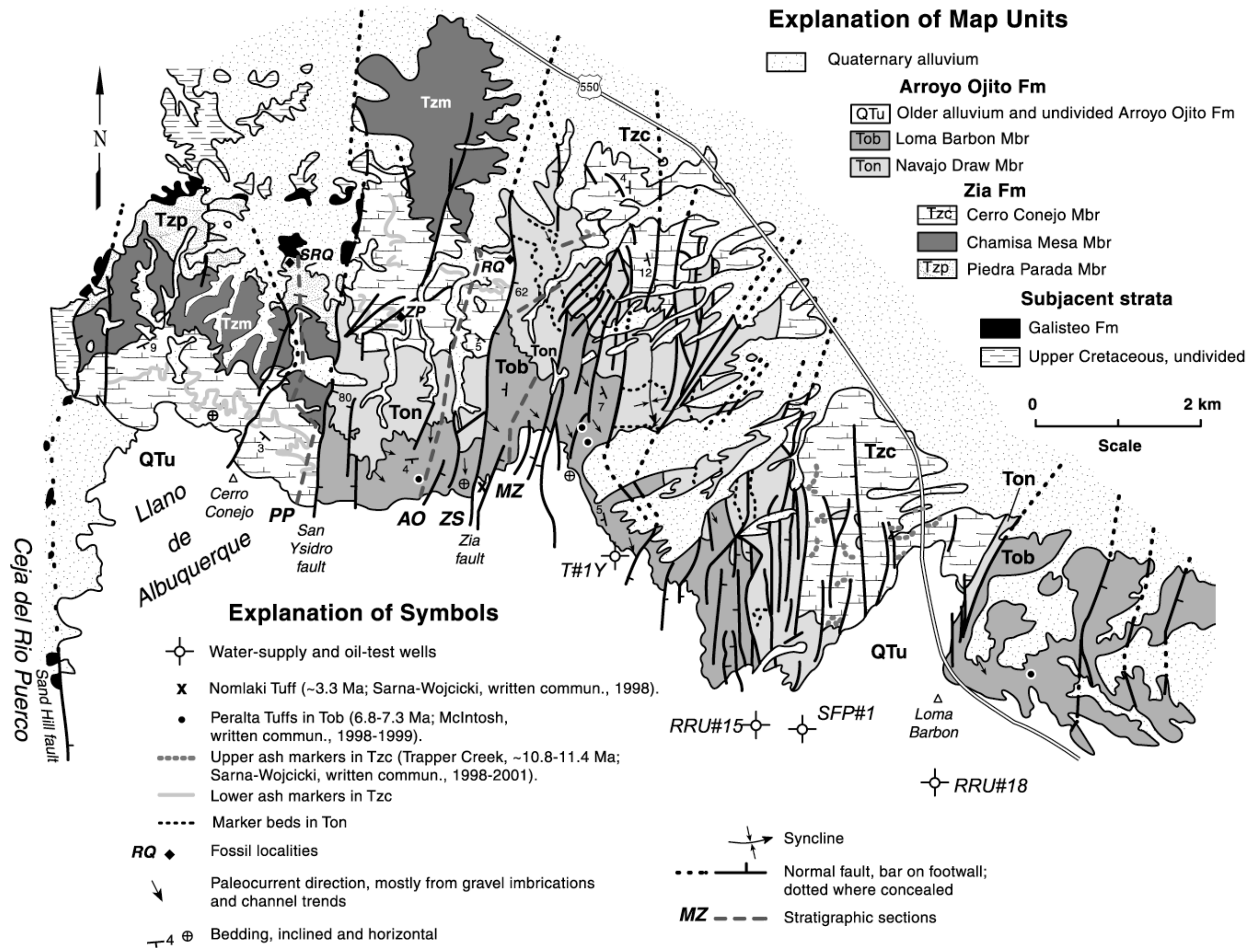

Figure 1. Generalized geologic map of the northwestern margin of the Calabacillas sub-basin (Albuquerque Basin), modified from the Cerro Conejo, Bernalillo NW, Loma Machette, and Bernalillo quadrangles (Connell, 1998; Koning and Personius, *in review*; Koning et al., 1998; and Personius et al., 2000). Stratigraphic study sites include Arroyo Piedra Parada (PP), Arroyo Ojito (AO), Zia fault (ZS), and the Marillo-Zia (MZ) sections. Unit QTu includes the Ceja Member of the Arroyo Ojito Formation of Connell et al. (1999). The Ceja Member unconformably overlies the Navajo Draw Member (not shown) on the footwall of the San Ysidro fault, but overlies the Loma Barbon Member to the east. Fossil localities of Galusha (1966; Tedford, 1981) indicated by black diamonds include the: Sanding Rock Quarry (SRQ; late Arikareean, 19-22 Ma), Rincon Quarry (RQ; late Barstovian, 12-14 Ma) and Zia Prospect (ZP; late Barstovian, 12-14 Ma). Volcanic ashes in the upper Cerro Conejo Member are correlated to the middle to late Miocene Trapper Creek tephra (Personius et al., 2000; Koning and Personius, *in review*). Water-supply and oil-test wells include the Tamara well (T#1Y), Santa Fe Pacific #1 (SFP#1), Rio Rancho Utilities #15 and #18 (RRU#15 and RRU#18, respectively).

about 40 km north of the drill site (Duchene et al., 1981). Other temporally correlative units to the Abiquiu and Zia formations include the Tanos and Blackshare formations, exposed in the Hagan embayment, along the eastern margin of the Albuquerque Basin. The Tanos Formation is as old as 25.4 Ma (Connell and Cather, *this volume*) and contains volcanic-bearing sediments derived from the Ortiz Mts., along the eastern rift margin in the Hagan embayment. To the south are exposed of the Popotosa Formation. The Popotosa Formation is a thick succession of mudstone and sandstone unit that is at least 15 Ma in the Belen sub-basin to the south (Lozinsky, 1994) and may be as old as *ca*. 25 Ma in the Abbe Springs basin, west of Socorro, New Mexico (Osburn and Chapin, 1983).

The Zia Formation (Galusha, 1966) comprises the basal part of the lower Santa Fe Group in the Calabacillas sub-basin (Fig. 2). The Zia Formation is dominated by eolian sandstone; fluviatile sandstone and mudstone beds tend to become more common upsection. The Zia Formation has been subdivided into the Piedra Parada, Chamisa Mesa, Cañada Pilares, and Cerro Conejo members (Galusha, 1966;

Gawne, 1981; Connell et al., 1999). The Piedra Parada Member unconformably overlies the pre-rift Galisteo Formation (Eocene) and Menefee Formation (Cretaceous). The age of the lower Piedra Parada Member is constrained by mammalian fossils at Standing Rock quarry (Fig. 1; Galusha, 1966), which indicate a late Arikareean North American land mammal "age." Correlations of these fossils to well dated localities in the Great Plains indicate an age of 19-22 Ma (Tedford and Barghoorn, 1999), although the biostratigraphy of the Standing Rock quarry suggests an age of *ca*. 19 Ma (R.H. Tedford, 2000, written commun.).

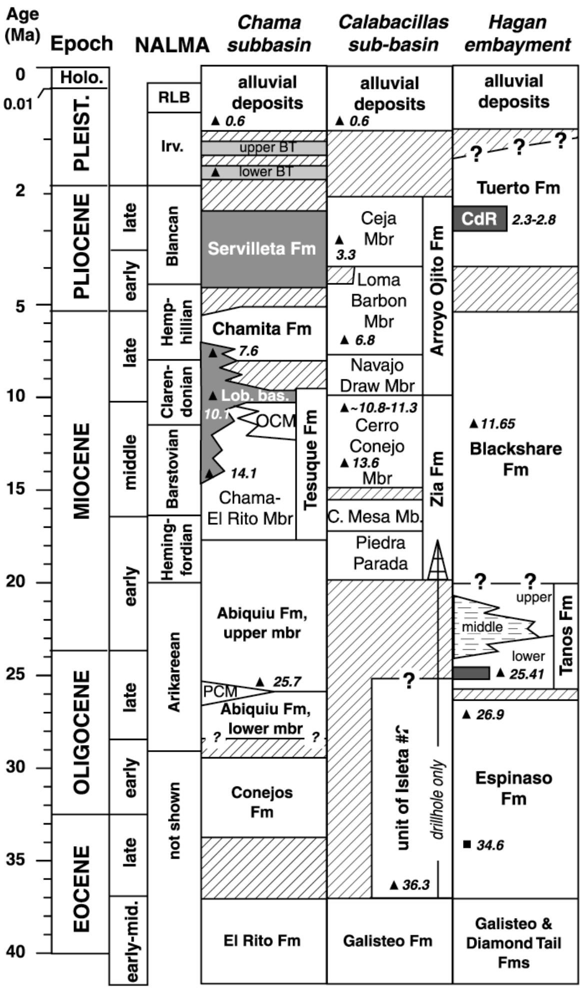

Figure 2. Correlation chart of selected Santa Fe Group units in the northwestern Calabacillas and Chama sub-basins (Connell et al., 1999; Tedford and Barghoorn, 1993; Moore, 2000; Lozinsky, 1994), and the Hagen embayment (Connell and Cather, *this volume*). Triangles denote dates (in Ma) of primary volcanic units. Shaded boxes denote basaltic flows. Abbreviations include, Lobato basalt (Lob. bas.), Ojo Caliente Member of the Tesuque Formation (OCM), and Pedernal chert Member of the Abiquiu Formation (PCM). North American Land Mammal "Ages" (NMLMA).

The Arroyo Ojito Formation (Connell et al., 1999) overlies the Zia Formation and is locally subdivided into three member units, in descending stratigraphic order: the Ceja, Loma Barbon, and Navajo Draw members. These units contain sand, gravel, and mud deposits by S-SE flowing rivers during late Miocene, Pliocene, and earliest Pleistocene times (Connell et al., 1999).

The unit of Isleta #2 (late Eocene-Oligocene) was proposed by Lozinsky (1988, 1994) for a 2 km thick succession of purplish-red to gray volcanic-bearing sandstone and mudstone recognized in deep oil test wells 25-30 km to the south; however, it is not exposed in the basin. The sand is arkose, lithic arkose, and subarkose (Lozinsky, 1994). This unit is temporally correlative to Oligocene volcanic and volcaniclastic units of the Espinaso Formation, Mogollon-Datil volcanic field, and San Juan volcanic field.

LITHOLOGY OF THE TAMARA #1-Y WELL

Cenozoic sediments examined in the Tamara well are predominantly fine- to coarse-grained sand with interbedded mud and sparse fine gravelly sand. Sand composition ranges from subarkose, lithic arkose, and feldspathic arenite. The stratigraphy of the upper part of the Tamara well is constrained by excellent exposures of the Arroyo Ojito and Zia Formations along the southern margin of the Rio Jemez valley that have been mapped by Koning (Koning and Personius, *in review*; Koning et al., 1998; Connell et al., 1999).

Geologic mapping (Koning and Personius, *in review*; Koning et al., 1998; Connell et al., 1999; Personius et al., 2000) indicates that neighboring strata of the Arroyo Ojito and Zia Formations generally dip about 3-10ºSW. A dip-meter log of strata below 2103 m (6900 ft) bls indicates that Cretaceous rocks dip as much as 20ºSW; however, it is not known whether stratal tilts in the upper part of the drillhole section progressively increase downhole, or whether they increase across unconformities. The thickness of Zia and Arroyo Ojito Formations were trigonometrically corrected using a 7º dip because of similar stratal tilts in exposures to the north. Deposit thickness was adjusted for 7º and 20º dips in lower units (Appendix 2). Lag times are not known for the samples and may contribute up to several meters of error in estimating stratigraphic boundaries, probably resulting in a slight increase in estimating apparent unit thickness. Dip-adjusted thickness of deposits correlated to the Zia and Arroyo Ojito Formations in the Tamara well is about 1138 m (Appendix 2). This is slightly thicker than estimates of about 1060 m for a composite Santa Fe Group section exposed to the west on the footwalls of the Zia and San Ysidro faults (Connell et al., 1999), and is considerably thicker than the 410 m of Zia section exposed on the footwall of the Sand Hill fault (Tedford and Barghoorn, 1999), near the structural boundary of the basin.

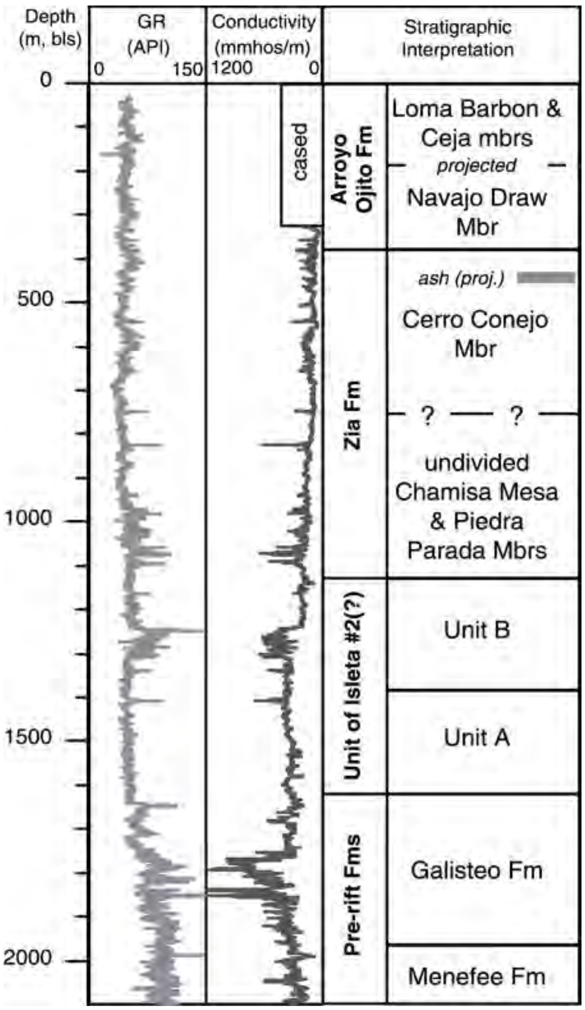

Figure 3. Interpreted stratigraphic column of Cenozoic sediments for the Tamara #1-Y well, including natural gamma ray (GR) and electrical conductivity logs (calculated from induction resistivity log) for comparison purposes. Stratigraphic interpretations are based on evaluation of cuttings and projection of contacts from geologic mapping.

Projections of mapped contacts on the Bernalillo NW quadrangle (Koning and Personius, *in review*) indicate that the base of the Loma Barbon Member of the Arroyo Ojito Formation is at 183 to 378 m bls (Fig. 3). Deposits correlated to the Ceja Member crop out along the northern rim of the Llano de Albuquerque (La Ceja) and are 30-85 m in thickness, of which ~15 m are penetrated in this well (Koning and Personius, *in review*; Personius et al., 2000). At about 180 m bls, deposits of very pale brown, fine-to medium-grained, quartz-rich sand are correlated to the Navajo Draw Member on the basis on comparisons with nearby geologic mapping (Koning and Personius, *in review*; Koning et al., 1998). At 405-424 m (1330-1390 ft) bls, traces of a gray altered tephra are recognized. This tephra-rich zone is in a similar stratigraphic position relative to ashes recognized in the upper part of the Cerro Conejo Member of the Zia Formation (usage of Connell et al., 1999) on the Bernalillo NW and Loma Machette quadrangles (Koning and Personius, *in review*; Personius et al., 2000). Some of these exposed ashes have been geochemically correlated to some of the middle to late Miocene (*ca*. 11-10 Ma) Trapper Creek tephra from Idaho (Personius et al., 2000; A. Sarna-Wojcicki, written commun., 2001; N. Dunbar, 2001, written commun., 2001). The lower part of the Cerro Conejo Member contains fossils that indicate a middle Miocene age (Tedford, 1981; Tedford and Barghoorn, 1999). The Cerro Conejo Member is 369-m thick in the Tamara well. The base of this unit is gradational with a 393-m thick succession of generally very pale brown to light gray, medium- to coarse-grained, subrounded to rounded, quartz-rich sandstone with abundant frosted quartz grains. This thick unit is correlated to the Chamisa Mesa and Piedra Parada members of the Zia Formation. The Zia Formation is composed of lithic arkose to feldspathic arenite (Beckner, 1996) and cemented zones are commonly recognized in this interval (Beckner and Mozley, 1998; Mozley and Davis, 1996).

The basal 0.5-3 m of the Zia section exposed in Arroyo Piedra Parada contains fluviatile gravel composed mostly of rounded chert pebbles derived from the Galisteo Formation with scattered cobbles of rounded, intermediate volcanic rocks (Fig. 4) deposited by southeast-flowing streams (Gawne, 1981). Elsewhere, these cobbles form a discontinuous stone pavement, where many of the volcanic clasts have been sculpted by the wind into ventifacts (Tedford and Barghoorn, 1999; Gawne, 1981). Volcanic clasts have been $^{40}Ar/^{39}Ar$ dated between 32 to 33 Ma (three dates: 31.8±1.8 Ma, 33.03±0.02 Ma, 33.24±0.24 Ma; S.M. Cather, W.C. McIntosh, unpubl. data), indicating that they were once part of the subjacent middle Tertiary volcaniclastic succession. These deposits unconformably rest upon upper Eocene sandstone and mudstone of the upper Galisteo Formation (Lucas, 1982). Thus, this gravel-bearing interval represents an unconformity that probably ranges between about 10-14 m.y. in duration between the Galisteo and Zia Formations at the northwestern margin of the Albuquerque Basin.

Between 1146-1393 m bls is unit B, an interval of pink to very pale-brown, mostly fine- to medium-grained, quartz-rich feldspathic arenite and lithic arkose recognized below strata correlated to the Piedra Parada Member in the Tamara well. Traces of a white ash and sparse scattered volcanic grains are observed between 1283-1292 m (4210-4240 ft) and 1366-1375 m (4480-4510 ft) bls, respectively. The lower 27 m of this interval contains trace amounts of purplish-gray intermediate volcanic rocks. The stratigraphic position of this unit, below the Piedra Parada Member, suggests that it might be correlative to the Abiquiu Formation. Petrographic analysis of this interval indicates that the Cenozoic portion of the Tamara well is distinct from the Abiquiu Formation, which contains considerably less quartz than Cenozoic deposits studied in the Tamara well (Fig. 5).

GUIDE TO THE GEOLOGY OF THE EASTERN SIDE OF THE RIO GRANDE VALLEY ALONG SOUTHBOUND I-25 FROM RIO BRAVO BOULEVARD TO BOSQUE FARMS, BERNALILLO AND VALENCIA COUNTIES, NEW MEXICO

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INTRODUCTION

The Plio-Pleistocene geology of exposures east of the Rio Grande floodplain to the top of the Sunport surface and equivalent surfaces from Rio Bravo Boulevard southward to Bosque Farms looks deceptively simple from a distance, but is complex at local outcrop scale. We interpret the exposures of the uppermost Santa Fe Group (Arroyo Ojito and Sierra Ladrones Fms) along the Rio Grande Valley from Rio Bravo Boulevard in Albuquerque, south to Isleta Pueblo, Bosque Farms, and Los Lunas, New Mexico, using several basic geological concepts pertinent to rift basins. The first concept (1) relates the fill of a half graben to a combination of an axial river and lengthy hanging wall tributaries and short footwall fans that are transverse to the axial river (Leeder and Gawthorpe, 1987; Fig. 1). The second concept (2) is that normal faults in the northern Albuquerque basin change scarp-face direction (Chamberlin, 1999; Fig. 2). The third concept (3) combines the first two, as a half-graben axial stream slaloms back and forth, following the changing position of the lowest points among half-graben basins and sub-basins (Fig. 3). The fourth concept (4) is that of fluvial fans, which are fluvial deposits that spread laterally and longitudinally along a basin floor from a large feeder trunk stream, typically developed on hanging walls, not from the shorter transverse streamflow-dominated piedmont deposits derived from footwall uplifts along the eastern basin margin (Love and Seager, 1996; Fig. 4). The fifth concept (5) is that of spillovers as fluvial systems extend fluvial fans into adjacent basins (Mack et al., 1997; Fig. 5). The sixth concept (6) combines concepts of fluvial fans and spillovers to a basin where three fluvial systems compete for axial position along a major half graben, but two of the fans enter the basin from the broad hanging wall (Fig. 6). The seventh concept (7) is the breakup the simple half-graben with an axial-fluvial system and large hanging-wall tributaries (concept 1) into a series of smaller sub-parallel half grabens influenced by changes in fault-dip polarity across intrabasinal normal faults (Fig. 7). Finally, real world complications make geology less simple (8) wherein (a) local volcanoes disrupt the fluvial systems, (b) water and sediment discharge change through time for all of the fluvial and alluvial systems (perennial and ephemeral streams), (c) stream gradients change due to local tectonic perturbations, (d) some uplifted blocks are beveled by a laterally swinging and aggrading combined axial stream, and (e) piedmont deposits interfinger with axial stream deposits along the margin of the active half graben (Fig. 8).

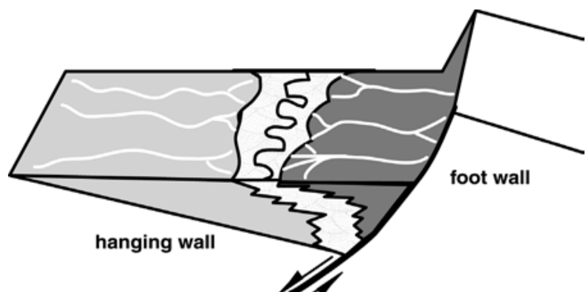

Figure 1. Axial stream in a half graben. Note short footwall fans and longer tributaries descending the hanging wall.

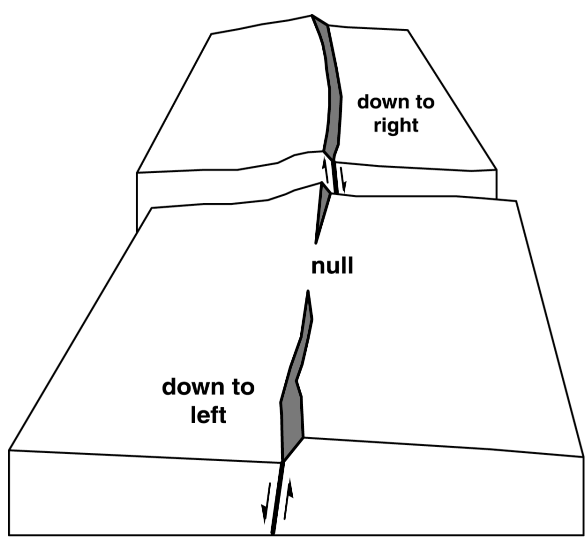

Figure 2. Block diagram showing steep fault with alternating scarps and null points along it.

GEOLOGY AND GEOMORPHOLOGY OF THE ISLETA AREA

If these concepts are to be applied to the geology of the Isleta area, the scale of the concepts must match local conditions. Applying concept 1 (the axial stream in a half graben with broad hanging wall) to this area, the major half-graben fault in Plio-Pleistocene time is the Hubbell Spring fault zone, 13 km east of the west edge of the Sunport surface. The axial-fluvial systems tract of the early Pleistocene ancestral Rio Grande is also about 13 km wide beneath the Sunport and Llano de Manzano surfaces. The hanging wall with tributaries extended ~ 26-32 km east from the valley of the Rio Puerco to the axial ancestral Rio Grande near the latitude of southern Albuquerque. These fluvial fans had headwaters beyond the hanging wall and crossed the basin diagonally from the north and northwest extending 48-161 km north-south. They deposited sediments over hundreds of square km up-gradient from the axial Rio Grande. High sediment delivery to the basin from the major tributaries probably overwhelmed small tectonic disruptions on the hanging wall.

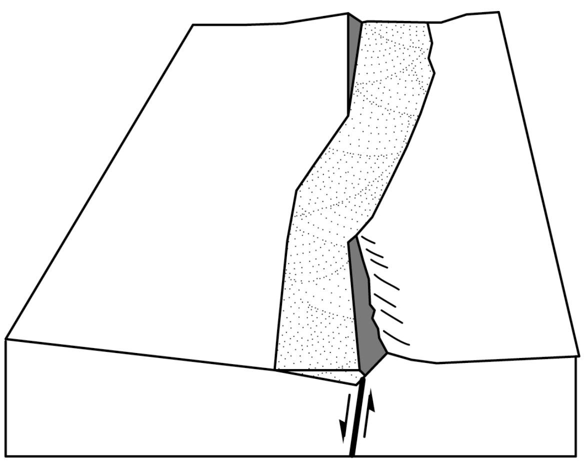

Figure 3. Block diagram illustrating axial stream that slaloms between half grabens along alternating-scarp fault blocks.

Working downsection from the Sunport surface, at the top of the exposures beneath eolian sheet sands and a strong stage III to local stage IV calcic soil are sand and gravel of the ancestral Rio Grande. This >10-m thick gravelly sand is a mixture of resistant, well-rounded clasts of extrabasinal origin that include boulders (up to 4 m in diameter) of upper Bandelier Tuff (UBT, 1.22 Ma). Also included are locally preserved Tschirege Ash (1.22 Ma; see below), pebbles of Tewa-Group pumice and obsidian from the Jemez Mountains (beginning with the 1.8-Ma San Diego Canyon Ignimbrite; *cf.* Self et al., 1996). The boulders were probably deposited as a result of a breakout flood from a breached lake in the Valles caldera that formed soon after caldera collapse and emplacement of the UBT. The breakout flood likely swept through the Jemez River canyon, picked up boulders of Tertiary basalts and Precambrian crystalline rocks, and spread out across the Sunport surface. The Rio Grande reworked these flood deposits shortly after this flood and prior to entrenchment of the present valley. A water-reworked fine-grained ash from within the upper part of the section yielded sanidine crystals with peaks in the age spectra from 1.05 to 1.7 Ma; however, the younger age is associated with fairly low concentrations of potassium, which suggests that part of this ash was altered. Thus, this 1.05 Ma date is too young. An ash bed, recognized below the Sunport surface along the southern margin of Tijeras Arroyo, yielded a $^{40}Ar/^{39}Ar$ date of 1.26±0.02 Ma, which is consistent with an upper Bandelier Tuff age. A fluvial terrace deposit of the ancestral Rio Grande west of the Rio Grande is lower than the Sunport, suggesting that it is inset against the Sunport surface. This terrace deposit contains a fluvially recycled ash that has been chemically correlated to the *ca*. 0.60-0.66 Ma Lava Creek B ash from the Yellowstone hotspot in Wyoming. These two tephra constrain the age of the Sunport surface to between 1.2-0.6 Ma. Paleomagnetic studies of fine-grained deposits near the local top of the section, between Hell Canyon Wash and Tijeras Arroyo, are pending.

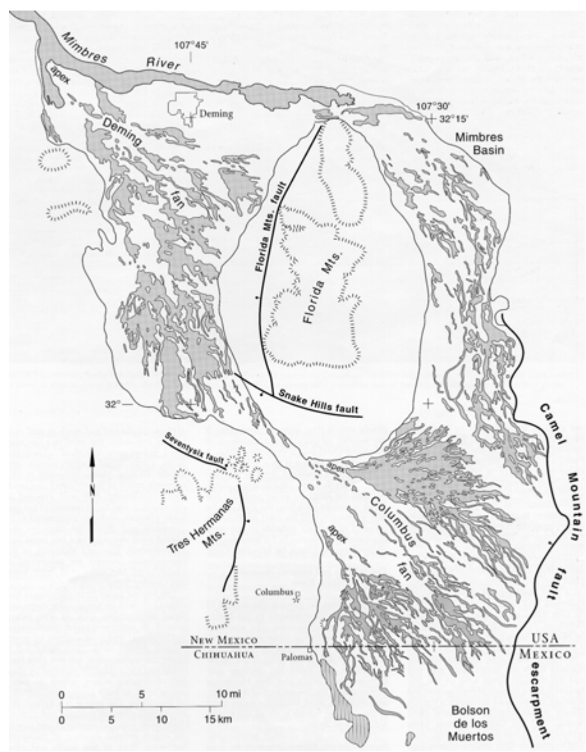

Figure 4. Fluvial fans of the Rio Mimbres system (from Love and Seager, 1996). Note that the Mimbres fluvial system is already out in a basin before it spreads into fan shapes.

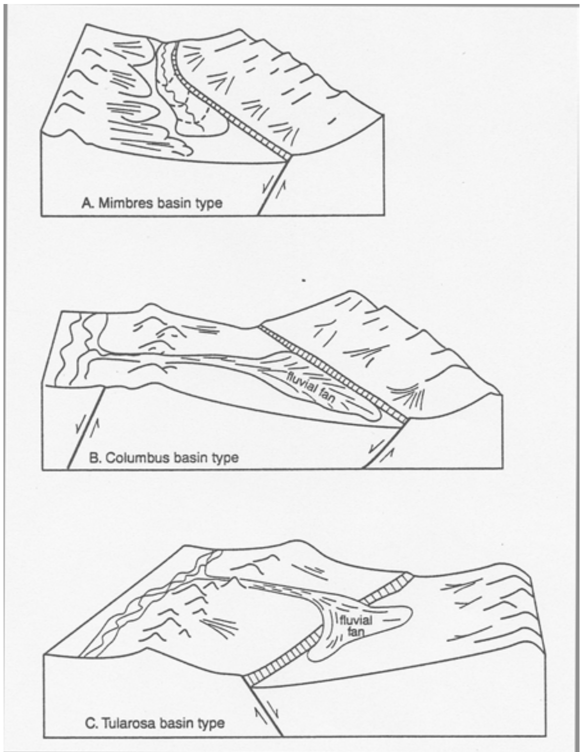

Figure 5. Spillover fluvial systems in different basinal situations (from Mack et al., 1997). Mimbres type (A), where the fluvial system enters and flows through the basin nearly parallel to the footwall scarp and axial valley. Columbus type (B), in which the fluvial system flows down the hanging wall dip slope of the spillover basin and builds a fluvial fan perpendicular to the footwall scarp. Tularosa type (C), in which the fluvial system moves across the footwall scarp into the spillover basin and builds a fluvial fan on the hanging wall dip slope perpendicular to the basin axis.

The axial Rio Grande deposits associated with the Sunport surface and Llano de Manzano are cut by numerous normal faults with separation of up to 15 m (Fig. 9). This faulted surface is partially buried by a piedmont alluvial apron prograding west from the Manzanita and Manzano Mountains.

Beneath the upper 10 m of sediment, the valley-margin geology is complicated. Locally, ancestral Rio Grande fluvial deposits, containing Tewa-Group pumice and obsidian, extend 30 m below the Sunport surface where they rest upon Pliocene deposits of the Arroyo Ojito Formation. Biostratigraphic data indicate the presence of a disconformity between the Arroyo Ojito Fm and pumice-bearing fluvial deposits of the Sierra Ladrones Formation. To the south, near Isleta Pueblo, this disconformity is more pronounced and occurs within 10 m of the Sunport surface, where it is an angular unconformity. Ancestral Rio Grande deposits tend to thicken to the east, where they are exposed in the walls of both Tijeras Canyon and Hell Canyon. At a monitoring well drilled on Mesa del Sol well, 6.5 km southeast of the Rio Bravo interchange, at least 500 m of fluvial sediments were penetrated.

Valley-margin exposures between Tijeras Canyon and Hell Canyon are cut by three major, and numerous minor faults in a north-northwest-trending zone called the Palace-Pipeline fault zone. Two of the major faults are normal, down to the west. The third is down to the east. There are hints of strike-slip motion as well, but no definitive piercing points to demonstrate horizontal movement. Minor faults are subparallel to this zone, but tend to bend to the east or west.

Although the topographic relief of the present valley is about 130 m, the total exposed stratigraphic section is more than 135 m thick because of faulting and local stratal tilts. Beds dip as much as 7ºSE; apparent dips along the outcrop belt are roughly 1º to the south. The southeastward tilt has preserved the middle and upper parts of the stratigraphic section above that seen on the highest block and below the youngest fluvial deposits of the Sunport surface (Fig. 10). The highest exposed structural block is stripped of at least 73 m of section seen on other blocks. Critical stratigraphic markers for the Plio-Pleistocene section include <5 cm-thick basaltic tephra geochemically correlated to Isleta volcano (2.7-2.8 Ma), Hawaiite tephra (unknown age), fluvially recycled Pliocene pumice and Bandelier Tuff, and thick (~24 m) reddish-brown clay, silty clay, and fine sand.

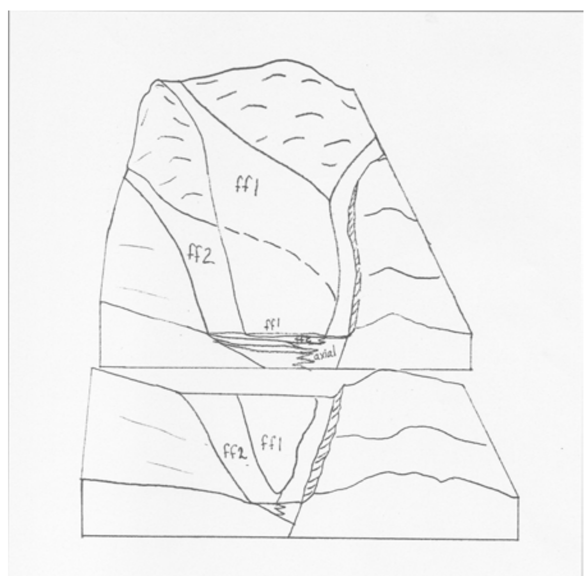

Figure 6. Schematic half-graben basin with axial fluvial system and two fluvial fans (ff1 and ff2) descending the hanging wall.

At the southern end of the exposures (Fig. 10), about 20 m of section is exposed between the top of the axial gravel and an exposure of lower(?) Bandelier ash that has been dated at about 1.55 Ma.

An $^{40}Ar/^{39}Ar$ date on this ash sampled north of Hell Canyon was interpreted to include crystals of the 1.05 Ma Valles dome rhyolite (Love et al., 2001), but these crystals have been recently reinterpreted to be fluvially reworked Bandelier ash (N. Dunbar and W. McIntosh, personal commun., 2001). Beneath this ash north of Hell Canyon is about 30 m of cross-bedded, pumice-bearing, loose pebbly sand of the ancestral Rio Grande. Beneath them are pale, cemented, fine-grained deposits that indicate a high local water table (spring-related or krenegenic deposits). Below these deposits are 24 m of fine-grained reddish-brown beds and 49 m of crossbedded, planar-bedded sand, and cross-bedded pebbly-to-cobbly sand. The sandy units locally include at least 4 different pumice-bearing beds. Below is basaltic tephra correlated to the eruptions of the Isleta tuff ring, base surge, lava flows and other cinder eruptions. These tephra are found on at least four different structural blocks at various elevations. Maximum offset across at least two faults is about 100 m.

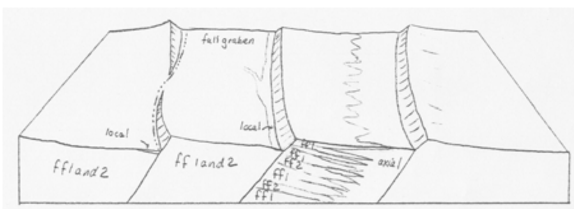

Figure 7. Block diagram showing the breakup of the half-graben hanging wall into several segments and the underlying stratigraphy resulting from fluvial fans and axial system of Figure 6.

Continuing downward in the section, beneath the Isleta tephra on the central uplifted block are another 21 m of cross-bedded pebbly sand, sand, and silty-clay planar beds. Locally, at least two more pumice-bearing beds crop out below the Isleta tephra.

North of Isleta, beneath Mesa del Sol are Pliocene sections exposed on uplifted fault blocks. One measured section includes 17 m of concretionary sandstone, silt, and clay with a 3-cm thick Hawaiite ash beneath coarser crossbedded loose sand. Another 35-m section has a pumice bed at its base that correlates geochemically to a pumice bed beneath Los Lunas Volcano to the southwest and Rio Rancho to the northwest. This pumice bed is in the same part of the section at Los Lunas volcano that contains an $^{40}Ar/^{39}Ar$-dated 3.12 Ma pumice (Maldonado et al., 1999).

The Pliocene crossbedded sandstones, pebbly sandstones, pumice and basaltic tephra-bearing sandstones beneath the thick reddish-brown fine-grained marker are part of a thick, laterally extensive package of transverse fluvial deposits. This package, deposited by major western-margin rivers and streams, is called the Arroyo Ojito Formation (Connell et al., 1999). From Rio Bravo Boulevard south, the units include volcanic clasts from both the western side of the Jemez Mountains (paleo-Jemez River to the north) and clasts from the Rio Puerco (coming into the Albuquerque basin from the northwest). Both types of deposits were spread across the northern Albuquerque basin as low-gradient fluvial fans that interfingered with each other. Reworked, water-rounded pumice from Jemez eruptions spread laterally at least 24 km east-west, and at least 113 km north-south. These fluvial fans descended southeastward to join the axial Rio Grande, which was the axial stream along the Hubbell Spring fault zone. The presence of relatively monolithologic pumice-bearing units 113 km south of their source in the Jemez Mountains suggests that these streams may have followed local alternating half-graben sub-basins (west of here, closer to Wind Mesa). The reddish-brown marker unit with spring-related units at the top, followed by deposition of ancestral Rio Grande along the Isleta valley-border transect, may signal a rearrangement of the structure of the hanging-wall and a subsequent adjustment of the northern and western fluvial fans.

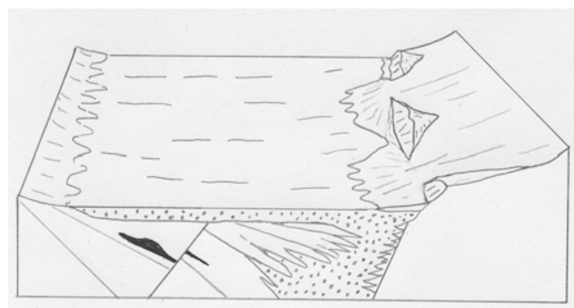

Figure 8. Block diagram showing complications of lateral erosion of axial stream across fault blocks, buried volcano, and advancement of piedmont across part of floodplain.

Exposures in Tijeras and Hell Canyons as well as along the Rio Grande Valley near Isleta show that more than 30 m of axial Rio Grande sand and gravel aggraded before the stream shifted westward. It beveled and buried at least two uplifted fault blocks, received breakout flood debris from the Jemez River, and finally began to entrench the present valley west of its former course (Fig. 10).

Across the valley southwest of this stop are Black Mesa and Isleta volcano, two Pliocene basaltic eruptive units with $^{40}Ar/^{39}Ar$ dates ranging from 2.7 to 2.8 Ma (Maldonado et al., 1999). In erosional contact with, above, and inset below the basalts, are several levels of inset terrace deposits. The highest terrace is 79 m above the Rio Grande. In the gravel pit north of the Black Mesa basalt flow is an exposure of Lava Creek B ash, about 46 m above the river.

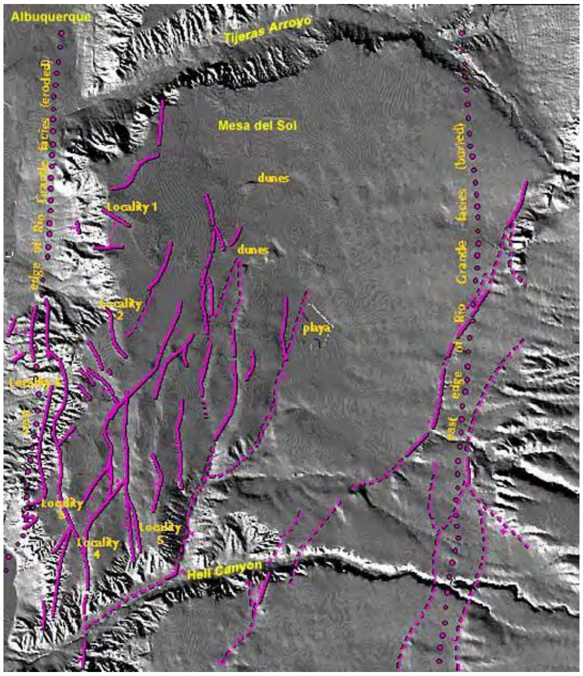

Figure 9. Digital elevation model of Sunport surface between Tijeras Canyon and Hell Canyon showing faulted blocks and encroachment of piedmont from east.

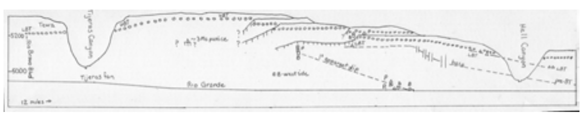

Figure 10. Schematic north-south sketch of about 18 km of exposures of Santa Fe-Group sediments from Rio Bravo Boulevard (I-25, exit 220) to bluffs east of Bosque Farms, New Mexico.

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