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.

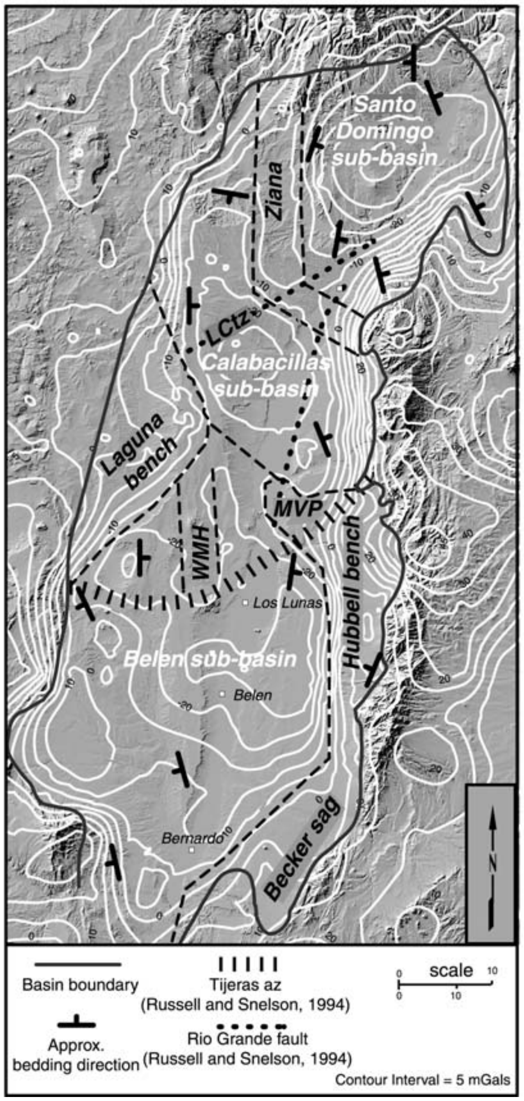

Figure 2. Shaded-relief image of the Albuquerque Basin and vicinity showing contours of the isostatic residual gravity anomaly as white contours (modified from Grauch et al., 1999). Approximate boundaries of major sub-basin depressions are shown by bold dashed lines. Major structural benches and intrabasinal 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.

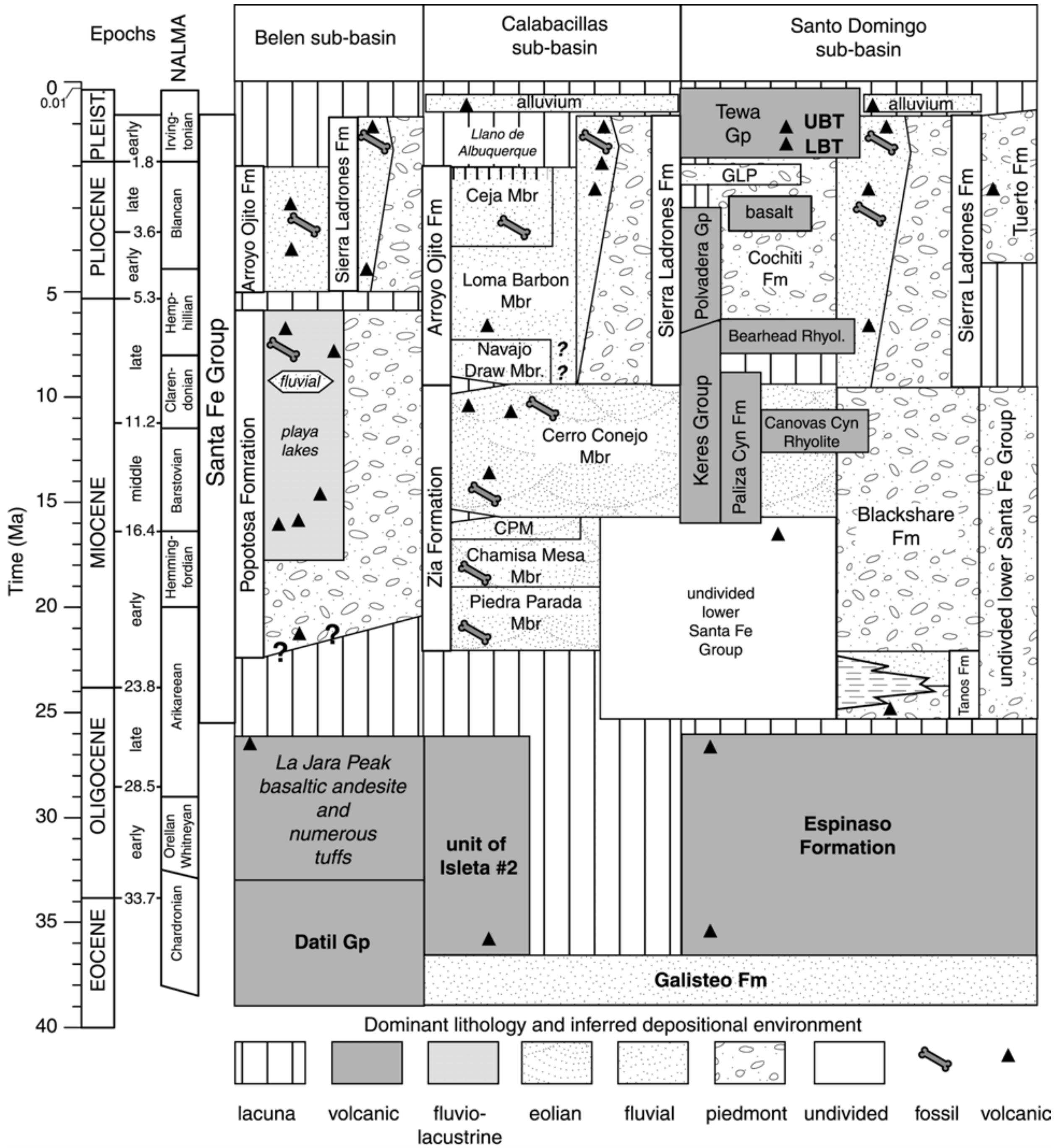

Figure 3. Schematic stratigraphic correlation diagram of the Albuquerque Basin and other basins of the Rio Grande rift, illustrating age-constraints and the North American Land Mammal "ages." Volcanic units include, the upper (UBT) and lower (LBT) Bandelier Tuff members of the Tewa Group. The Cañada 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.

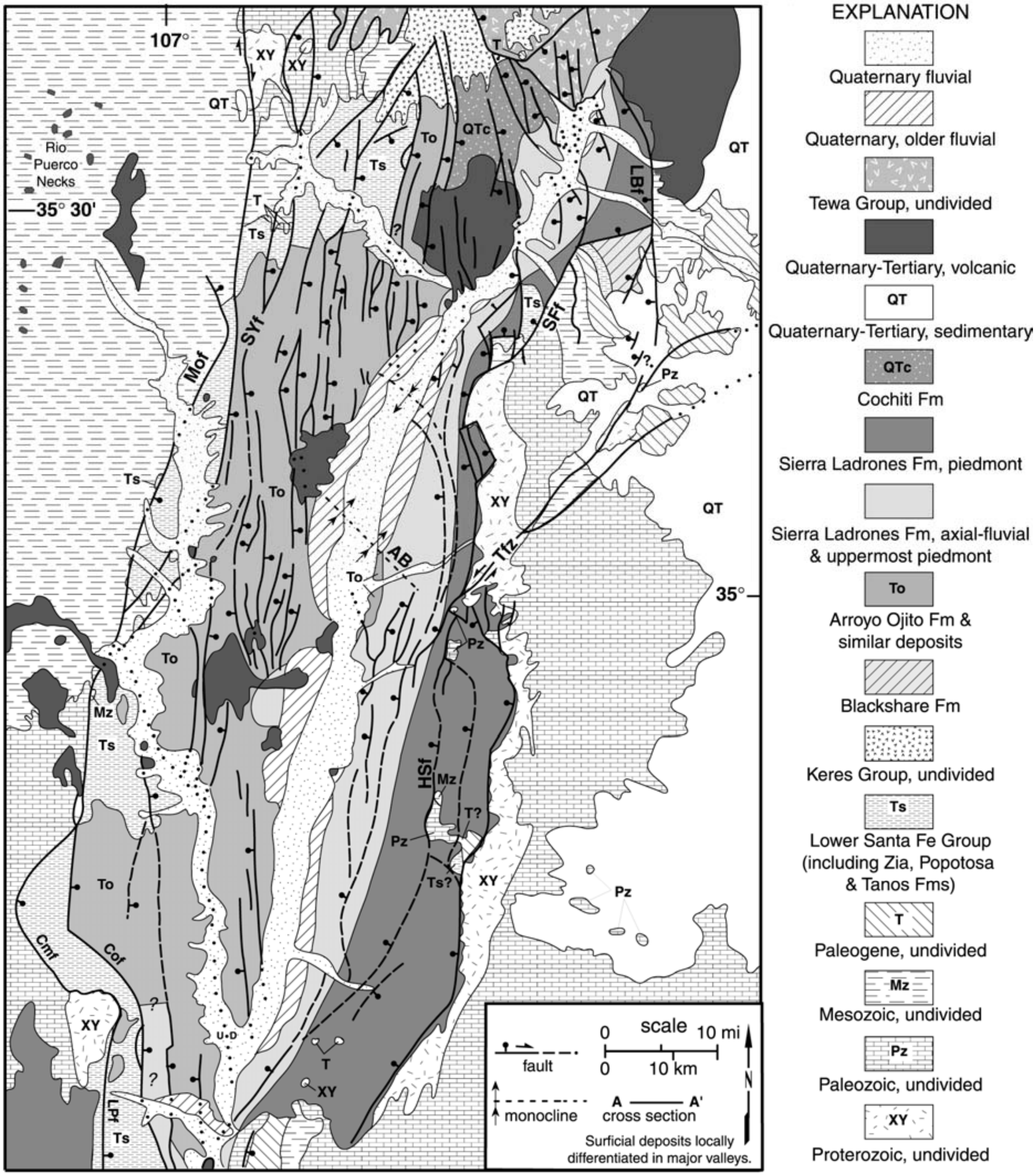

Figure 4. Generalized geologic map of the Albuquerque Basin, modified from Hawley (1996 and Hawley et al., 1995), with additional modifications from Osburn (1983), Machette et al. (1998), Maldonado et al. (1999), Connell (1997), Connell and Wells (1999), Connell et al. (1995, 1999), Cather and Connell (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.

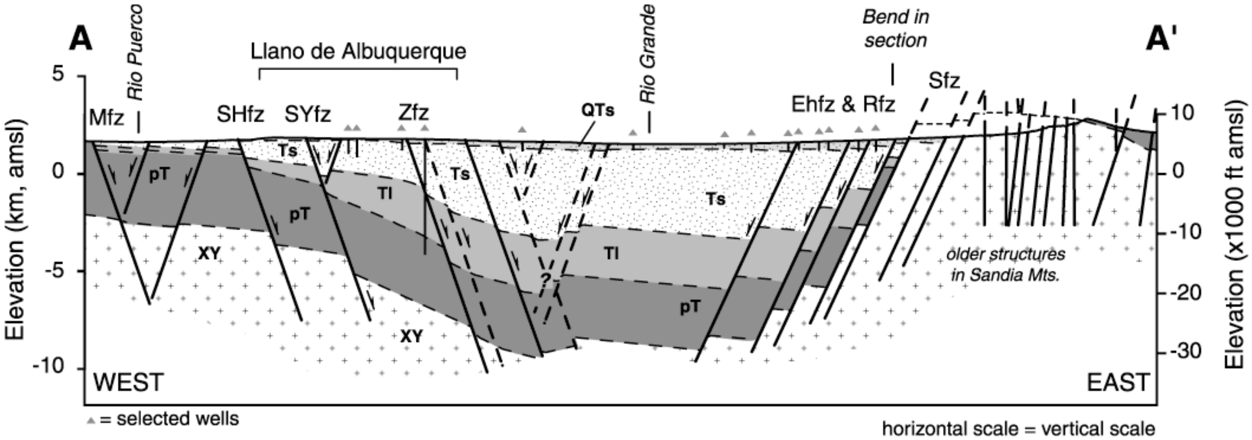

Figure 5. Generalized geologic cross section of Calabacillas sub-basin drawn at latitude of Paseo del Norte Boulevard in Albuquerque (Fig. 4). Gray triangles denote locations of selected wells that were used provide stratigraphic control for the cross section. The Llano de Albuquerque represents a broad 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).

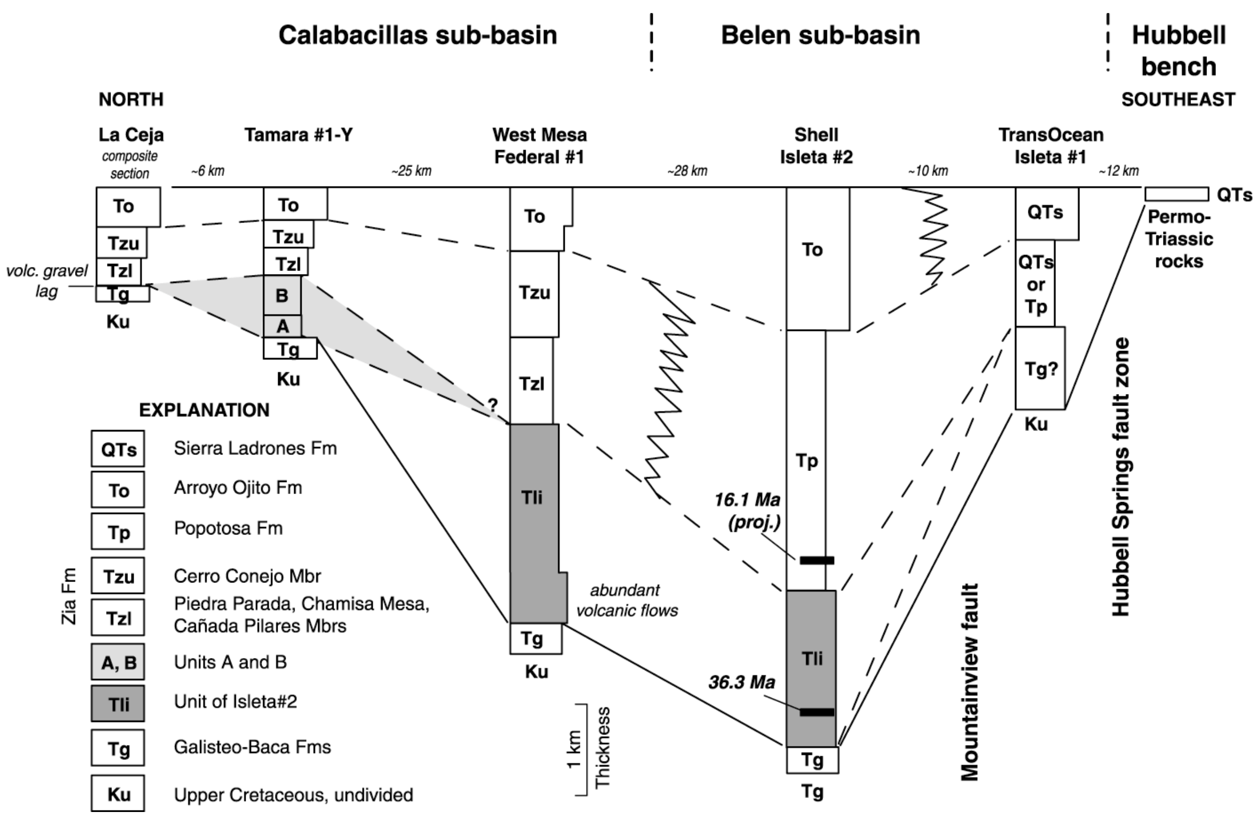

Figure 6. Stratigraphic fence of Cenozoic deposits in the Calabacillas sub-basin. Data from oil test wells (Lozinsky, 1988, 1994; Connell, Koning, and Derrick, *this volume*; Connell et al., 1999; Tedford and Barghoorn, 1999; Maldonado et al., 1999; Black and Hiss, 1974). Locations of wells and stratigraphic 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.

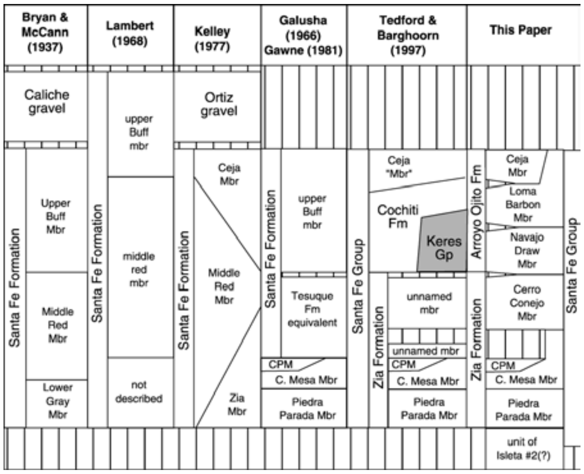

Figure 7. Summary of stratigraphic nomenclature development in the northwestern Calabacillas sub-basin. Sedimentary units include the Cañada Pilares Member (CPM) of the Zia Formation. Volcanic rocks are shaded gray.

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.

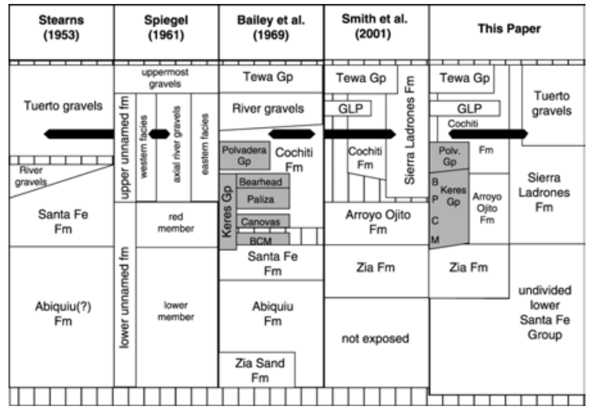

Figure 8. Summary of development of stratigraphic nomenclature in the Santo Domingo sub-basin. Shaded units are volcanic; black shading indicates the basalts of Santa Ana Mesa and Cerros del Rio. Other sedimentary units include the gravel of Lookout Park (GLP) of Smith and Kuhle (1998a, b). Volcanic 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 sub-groups 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.

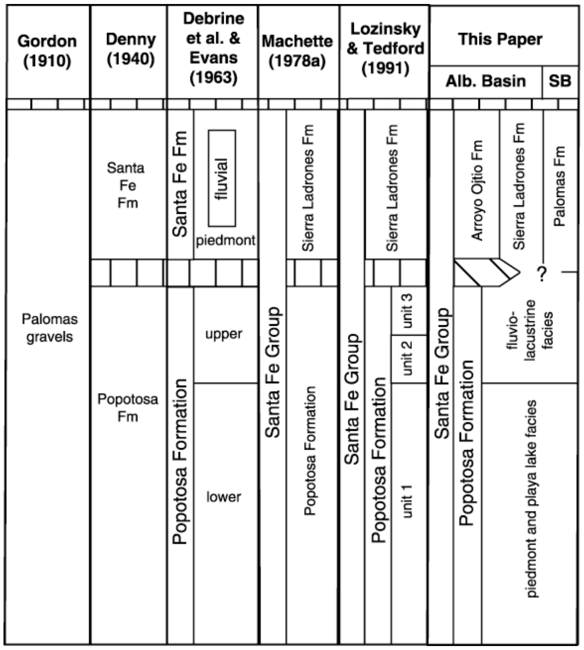

Figure 9. Summary of stratigraphic nomenclature development in the Belen sub-basin, illustrating the evolution of stratigraphic terms in the northern Socorro Basin and Belen sub-basin.

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).

ACKNOWLEDGMENTS

This study was supported by the New Mexico Bureau of Mines and Mineral Resources. Much of the data discussed for this study came 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). The author is 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 referred to in this paper. In particular, the author thanks Mr. Peter Pino for enabling access to study the geologically important localities along the Rincones de Zia, Mr. Michael Romero for facilitating access to San Felipe Pueblo lands, and Mr. Archie Chavez for access to Sandia Pueblo lands. Mr. Blane Sanchez and John Sorrell were particularly helpful in facilitating access and aiding research on Isleta Pueblo lands. This paper benefited greatly from the New Mexico Geochronological Research Laboratory. I thank Richard Chamberlin, Steve Cather, Dave Love, Gary Smith, and John Hawley for reviewing an earlier draft of this manuscript. I also thank Bill McIntosh, Steve Cather, Dave Love, and Richard Chamberlin for graciously allowing me to report some of their unpublished $^{40}Ar/^{39}Ar$ dates.

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