GEOLOGIC MAP OF THE LUIS LOPEZ 7.5 MINUTE QUADRANGLE, SOCORRO COUNTY, NEW MEXICO

New Mexico Bureau of Mines and Mineral Resources
Open-file Report 421

Richard M. Chamberlin and Ted L. Eggleston

1996
Revised August, 1999

CONTENTS

Page

ABSTRACT

The Luis Lopez quadrangle lies within the southern section of the Rio Grande rift, where two pulses of WSW directed lithospheric extension have broken a subduction-related Oligocene caldera complex into a north-trending array of tilted fault-block ranges and alluvial basins. Topographic elements of the quadrangle include the narrow north-trending Chupadera Mountains, and adjacent piedmont slopes of the Broken Tank and Socorro basins. The latter is represented by a dissected and faulted bajada-slope that descends to the floodplain of the Rio Grande at the east margin of the quadrangle.

The strongly east-tilted Chupadera block, mostly of Miocene age, locally provides a cross-section like view of the southeastern sector of the 31.9 Ma Socorro caldera. As much as 2 km of lithic-rich caldera facies Hells Mesa Tuff exposed at Red Canyon locally defines the collapsed core of the Socorro caldera. Bedded autoclastic ignimbrite breccias and intercalated fall deposits at the top of the intracaldera Hells Mesa probably reflect a brief resurgent phase. Autoclasts in the upper ignimbrite are texturally and compositionally identical to a coarsely porphyritic, spherulitic lava flow of Hells Mesa age, which is partially exposed in the southern moat-like area, near Nogal Canyon.

Sedimentary and volcanic units that back filled the Socorro caldera, prior to the next major ignimbrite eruption (28.7 Ma La Jencia Tuff), are collectively assigned to the Luis Lopez Formation. Distribution patterns of rhyolitic to andesitic conglomeratic sandstones in the lower Luis Lopez are consistent with derivation from a resurgent dome and the southern caldera wall.

Clasts of pre-caldera rhyolite (33.7 Ma) occur in the upper wall-derived facies of this moat-fill unit, which was deposited during a lull in volcanism (31.9-30.2 Ma). Volcanic strata of the Luis Lopez formation consist of early trachybasalt lavas, medial lithic-rich rhyolitic ignimbrites and andesitic porphyry lavas, and late-stage flow-banded rhyolite lava domes and tuffs. Ages of Luis Lopez volcanic units, from about 30.2 to 28.7 Ma, imply that they are premonitory events related to a developing magma system that culminated in eruption of the 28.7 Ma La Jencia Tuff from the adjacent Sawmill Canyon caldera. The eastern topographic wall of the Sawmill Canyon caldera is partially exposed near Black Canyon; here breccias and conglomerates derived from the Luis Lopez formation are overlapped by 28.0 Ma Lemitar Tuff at the caldera wall.

Moderately tilted volcanic-rich conglomerates and playa claystones of the Miocene Popotosa Formation are preserved in tilt-block depressions within and adjacent to the Chupadera range. Intercalated dacite and trachybasalt flows of late Miocene age help define local unconformities associated with fault-block highlands in the early rift basins. The 8.42 Ma basalt of Broken Tank flowed westward across the Chupadera block on inset piedmont gravels of the Popotosa Formation near Walnut Creek and Broken Tank; it then flowed northward onto playa muds near Bear Canyon.

Episodes of hydrothermal alteration and mineralization were locally associated with shallow silicic magmatism and volcanism of Oligocene and late Miocene age. Potassium-metasomatism and slightly younger manganese-oxide vein deposits in the northern Chupadera Mountains appear to be temporally and spatially linked to late Miocene rhyolitic magmatism in the Socorro Peak area, about 6 km north of the quadrangle.

Pliocene and Pleistocene sedimentary deposits of the Sierra Ladrones Formation record the transition from early-rift closed basins to late-rift, valley-fill deposits of the ancestral Rio Grande and its tributary drainages. Important aquifers within the quadrangle include a newly recognized fluvial-fan deposit of early Pliocene age (ca. 5 Ma) apparently emanating from the eastern Magdalena Mountains, and well known sands and gravels of the ancestral Rio Grande (axial facies of Sierra Ladrones Formation). Aquifer units, possibly as much as 300 m thick, are locally preserved on subsiding, high-angle fault blocks near the eastern margin of the quadrangle.

INTRODUCTION

The Luis Lopez 7.5 minute quadrangle, here after referred to as "the quadrangle," is located in central New Mexico, along the western border of the Rio Grande valley about 6 miles southwest of the town of Socorro (Fig. 1). The village of Luis Lopez lies in the northeast corner of the quadrangle adjacent to the floodplain of the Rio Grande.

Dominant topographic elements of the quadrangle include the narrow north-trending Chupadera Mountain block, along its western flank, and a complexly eroded and faulted bajada, which generally slopes eastward from the range front into the Rio Grande Valley (Socorro Basin, Fig. 2). Large canyons descending from near the alpine crest of the Magdalena Mountains, such as Red Canyon (Fig. 1), cut across the east-tilted Chupadera Mountain block. These superimposed (let down) drainages apparently reflect a time in the early Pliocene when alluvial fans emanating from the Magdalena Mountains sloped uniformly eastward to the ancestral Rio Grande (ie.Chupadera block was largely buried). The highest point in the quadrangle is an unnamed peak on the western edge of the Chupadera block (elev. 6410 feet, near Broken Tank Mesa) and the lowest point is on the Rio Grande floodplain in the southeast corner (elev. 4540 feet).

Access to the quadrangle is provided by graded county roads and old mining roads that connect with frontage roads and exits along Interstate 25, or US Highway 60. At the east margin of the quadrangle, low roadcuts along I-25 locally expose ancestral Rio Grande deposits disconformably overlain by middle Pleistocene alluvial-fan deposits.

Climate of the quadrangle is temperate and semiarid; most of the 9–11 inches of annual precipitation in the Socorro region comes from monsoonal thunderstorms in August and September. Snowmelt on the crest of the Magdalena Mountains also provides a major source of groundwater recharge to the western flank of the Socorro Basin (Roybal, 1991, fig. 5). Vegetation in the quadrangle consists mostly of creosote bush, mesquite, various grasses, cacti, and very sparse juniper on rocky slopes in the Chupadera Mountains. Cottonwoods and walnut trees occur locally along Nogal Canyon. The principal land use is cattle grazing; most of the land is leased to local ranchers by the U.S. Bureau of Land Management. There is no active mining in the quadrangle; recently, however mine-waste rocks have been used as a source of riprap to line U.S. Bureau of Reclamation irrigation and conveyance channels on the Rio Grande floodplain. The "Chupadera Mountains" may have been named for sucking insects (biting gnats) that swarm over these low ranges near Socorro during the summer monsoons (Pearce, 1965, p. 34).

Previous geologic studies of the Chupadera Mountains (Miesch, 1956; Willard, 1973) emphasized the distribution and chemistry of small manganese deposits in the Luis Lopez district. These earlier studies predated establishment of a regional stratigraphic and structural framework for Cenozoic volcanic rocks of the Socorro-Magdalena region (e.g. Osburn and Chapin, 1983a, 1983b; Chapin and others, 1978). As presented here (Sheet 1, Sheet 2, Table 1, Table 2), the volcanic geology of the quadrangle is compiled with some revisions from the dissertation map of Chamberlin (1980) and thesis map of Eggleston (1982). New high-precision $^{40}Ar/^{39}Ar$ age dates of volcanic rocks in the quadrangle (Table 3) allow a more confident revised interpretation of the local magmatic and eruptive history of the quadrangle, which is integrated with a precise regional chronostratigraphic framework for the Mogollon-Datil volcanic field (McIntosh and others, 1991, 1992). Recent geologic mapping of late Cenozoic basin fill deposits in the eastern two-thirds of the quadrangle, by the senior author, provides new control on the lateral extent of ancestral Rio Grande deposits (QTsf), in addition to new constraints on the Quaternary history of the Socorro Canyon fault zone. Ancestral Rio Grande deposits form the primary aquifer for the town of Socorro and adjacent farm communities along the Socorro basin. The Socorro Canyon fault zone, an active element of the Rio Grande rift, represents a significant potential earthquake hazard for residents of central New Mexico.

General descriptions of stratigraphic units and associated intrusions within the quadrangle are listed in Table 2. Age relationships and correlation of these maps units are listed in Figure 3.

Formal stratigraphic nomenclature used here is based on the stratigraphic chart of Osburn and Chapin, 1983a, with some additional nomenclature from Cather and others, 1994. This report was revised in August 1997, primarily to incorporate new age data on the basalt of Broken Tank and the basalt of Olney Ranch (Table 3).

CENOZOIC TECTONIC SETTING

The Luis Lopez quadrangle lies within the southern Rio Grande rift, where 50 to 100 percent extensional strain over the last 29 million years has broken a middle Tertiary caldera complex and surrounding volcanic plateau into a north-trending array of tilted fault block ranges and intervening alluvial basins (Chapin and Cather, 1994; Chapin 1989). In the Socorro-Magdalena region, the dominant mechanism of extension in the brittle upper crust has been progressive dip slip and concurrent rotation of imbricate north-trending normal fault blocks, originally termed domino-style crustal extension (Chamberlin, 1978, 1983). The process of extension has been episodic, apparently in concert with zeniths of regional magmatic pulses at 28 and 11 Ma (Cather and others, 1994). Large displacement (1–3 km) low-angle normal faults locally outline range fronts and control major north-trending mountain canyons. They define first-order tilt blocks that are internally deformed like a deck of cards by numerous second order imbricate faults (eg. Angelier and Colletta, 1983). Amounts and rates of extensional strain in the Socorro-Magdalena area of the rift appear to have varied widely in space and time (Chamberlin and Osburn, 1984). Approximately 50–100 percent Neogene extension in the Lemitar Mountains (revised after Chamberlin, 1983) is distinctly greater than the 25–40 percent of Neogene extension in the Nogal Canyon area of the Chupadera Mountains (E-E, Sheet 2). Late-stage high-angle rift faults of Pliocene to Pleistocene age locally truncate low-angle early rift faults (Chamberlin, 1983), and apparently represent a minor component of total regional extension.

The strongly east-tilted first-order fault block of the Chupadera range provides a longitudinal profile across the southeastern margin and collapsed core of the Socorro caldera (Osburn and Chapin, 1983a, 1983b). We note that virtually all *large* circular volcanic depressions, termed calderas, are underlain by subsided piston-like structures, termed cauldrons (Lipman, 1984, Smith and Bailey, 1968). Significant structural and topographic relief in the Socorro region reveals all levels of caldera features, including thick intracaldera tuffs, exhumed unconformities at caldera walls, and ring-fracture zones bounding subsided cauldron blocks. Following the example of Lipman, 1984, we use the more general term "caldera."

The Socorro caldera is the largest and oldest of an Oligocene caldera complex that trends west-southwest from Socorro over a distance of 75 km (Fig. 2). These calderas are most likely the surface expression of shallow overlapping and nested silicic plutons (c.f. Lipman, 1984) that were periodically emplaced along a deeply penetrating lateral shear zone known as the Morenci lineament (Chapin and others, 1978; Chapin, 1989; Ratte, 1989).

Between Morenci, Arizona and Socorro, New Mexico, a distance of 230 km, the Morenci lineament is marked by a belt of Oligocene and Miocene volcanic centers about 25 to 40 km wide. Analysis of seismic refraction lines and gravity data by Schnieder and Keller (1994) indicate that the crust-mantle boundary deepens rapidly to the north, near O-Bar-O Mountain, about 50 km south of Datil. Schnieder and Keller (1994) interpret this east-northeast trending inclination of the Moho (10° north-northwest), as the southeastern structural margin of the Colorado Plateau. O-Bar-O Mountain represents one of 5 middle Tertiary central volcanoes that define the Morenci lineament along the southern margin of the plains of San Agustin (Ratté, 1989). We suggest that Schnieder's southeastern structural margin of the plateau is largely coincident with the Morenci lineament.

Based on regional seismic reflection data from northern Catron county, Garmezy (1990) interpreted the southeast margin of the Colorado Plateau as a major right-lateral shear zone of Laramide ancestry. The recently compiled geologic map of New Mexico (Anderson and Jones, 1994) shows a distinct oroclinal bend in regional fault trends coincident with the Morenci lineament in west-central New Mexico. Regional trends strike northwest on the south side of the lineament and northeast on the north side. The northeast trend (060) of the unextended western part Morenci lineament is roughly parallel to that of Precambrian orogenic belts in basement rocks of the southwestern United States (Condie, 1982), thus a Precambrian ancestry is also possible.

Within the strongly extended caldera complex southwest of Socorro, a segment of the Morenci lineament is expressed as a transverse tilt-block domain boundary, referred to as the Socorro accommodation zone (SAZ, Chapin, 1989) and previously called a transverse shear zone (Chapin and others, 1978). Chapin's comprehensive summary (1989), points out that the SAZ has periodically "leaked" magmas about every 1–3 million years for the last 32 million years, except in the last 4 million years. However, a geophysically defined mid-crustal magma body below the Socorro region (Sanford and others, 1977) represents a potential source for future volcanic eruptions (Chapin, 1989).

A similar accommodation zone across the Colorado River extensional corridor (Nevada-Arizona line) has been analyzed by Faulds and others (1990). Several structural models involving transverse boundaries between tilt-block domains presumed to lie above low-angle detachment faults were tested by Faulds and others (1990). They conclude that opposing dip directions for underlying detachments subjacent to an accommodation zone would produce significant strike slip offset parallel to the domain boundary. Evidence of strike slip along the Nevada-Arizona zone and the SAZ is lacking (Chapin, 1989), thus detachment faults, as commonly envisioned (Lister and others, 1986) are probably not needed to explain patterns of extensional strain in the Socorro region.

Gravity is clearly the maximum compressive stress, and one of the major driving forces of extensional tectonics (Wernicke, 1992). Jones and others (1996) have recently shown that present day differences in regional gravitational potential energy are sufficient to drive active crustal extension in the Rio Grande rift and the northern Basin and Range. We suggest that the direction of fault block rotation in extensional domains may be initially determined by the preexisting structural grain of basement rocks and the inherently unstable gravitational potential energy of inclined imbricate blocks (like slightly inclined books on a shelf).

The SAZ may be the surface expression of a major strike-slip fault zone of Laramide ancestry that juxtaposed large basement terranes of opposing structural grain. Presumably prior to regional extension, north-trending basement zones of weakness (eg. old faults) in favorable orientations dipped mostly to the east on the north side of the Morenci lineament, and mostly to the west on the south side

STRUCTURAL EVOLUTION

Understanding the structural evolution of the Luis Lopez quadrangle requires a general temporal and spatial classification of faults, folds, dikes, veins and associated features such as local angular unconformities, volcanic vents and alteration zones. Structures in the quadrangle are grouped into three major time periods of apparent activity: 1) middle to late Oligocene faults and dikes associated with the Socorro and Sawmill Canyon calderas, 2) late Oligocene to late Miocene *rotated* normal fault blocks of the early Rio Grande rift, and 3) mostly non-rotated high-angle normal faults of Pliocene to Pleistocene age associated with the late stage Rio Grande rift. In addition, east-northeast striking dikes, scissors faults and associated flexures in the northern Chupadera Mountains are here assigned to the transverse tilt block domain boundary of the SAZ, which has been active since Oligocene time.

Generally west-southwest directed extension in the Socorro region of the Rio Grande rift has been episodic (Cather and others, 1994) and is strongly domainal, that is, variable in degree and style in both space and time (eg. Chapin and others, 1978; Chamberlin and Osburn, 1984;

Chapin, 1989; Chamberlin and others, 1994). Contrasts between the northern and southern Chupadera Mountains provide an example of domcinal extension. Northwest tilted blocks in the northern Chupadera Mountains, north of the Tower Mine, lie on the southern margin of a broad strongly extended domain of west tilted "domino" blocks. The west tilted domain is spatially associated with a widespread pile of late Oligocene basaltic andesite lavas (La Jara Peak Basaltic Andesite, Tlp) as much as 500 m thick. Less tilted and extended blocks (east tilted dominos) in the southern Chupadera Mountains near Nogal Canyon are not associated with late Oligocene basaltic andesite lavas, but rather only minor (0–30 m) thicknesses of locally derived rhyolitic sandstones (Tjus, Tvs1, Tvs2) stratigraphically equivalent to the La Jara Peak lavas.

Two major classes of structures are spatially defined; they are: 1) mostly NNW to N striking longitudinal normal faults of the Rio Grande rift that displace Miocene to Pleistocene basin fill deposits in addition to older volcanic "basement" (A-A' to F-F'; sheet 2), and 2) mostly ENE to E striking transverse faults and dikes of the Socorro and Sawmill Canyon calderas that are locally associated with unconformities and "growth faults" of Oligocene age (G-G', sheet 2).

The overall fault pattern is, however, a complex interlocking mosaic locally showing both major spatial trends. Relatively short transverse "bends" (or deflections) in longitudinal fault traces are best understood if viewed as block corners formed by the intersection of rift faults with preexisting planes of weakness, such as caldera ring fractures or older brittle faults in crystalline basement rocks. Because internal body forces of gravity and buoyancy locally drive extensional fault block rotation (eg. Reiter and others,1992; Jones and others, 1996), the preexisting transverse structures can act as *local* stress domain boundaries. Thus these transverse discontinuities tend to inhibit the along strike propagation of the longitudinal gravity shears (normal faults). The degree of segmentation of these longitudinal faults can also be influenced by lateral variations in the degree of plasticity (eg. thermal regime) of "hot" middle crust or asthenospheric mantle that is presumably being "flattened" by gravity flow beneath the brittle extended terranes of the Rio Grande rift (eg. Wernicke, 1992).

Faults in the quadrangle are also classified in terms of their apparent sense of shear, based primarily on stratigraphic separation combined with limited observations of striations on exposed fault planes (sheet 1). Apparent offset, or lack of offset, of older dikes by longitudinal faults also provides control on sense of shear.

Since late Oligocene time (ca 27.4 Ma), the first order Chupadera Mountains block has been tilted 35 to 45 degrees to the east northeast by rotational normal faulting within the rift. This rotation was probably episodic, occurring mostly in late Oligocene and middle to late Miocene time. Caldera structures, early rift longitudinal faults, and Oligocene to Miocene strata appear to have been rotated, at least in part, concurrently with faulting.

In terms of their present position, originally high-angle transverse caldera-related faults bounding the "piston-like" Red Canyon horst (G-G', sheet 2) now have the appearance of sinistral and dextral oblique-slip faults. Removing 45 degrees of local ENE tilt from the exposed segment of the Chupadera Spring ring-fracture fault (sheet 1; G-G', sheet 2) indicates that its original strike was more northerly (ca. 040) and its original dip was steep (75–85°NW).

The predominant down-to-the west early rift faults are now low-angle (20–45° dip) dip-slip normal faults that show large stratigraphic separations of Oligocene and Miocene strata. In order of decreasing stratigraphic throw, examples of large low-angle normal faults are the range bounding Chupadera fault (eg. B-B' sheet 2) and intrarange faults such as the Bianchi, Nancy, Sinuoso, Extraño, and Shrine Valley. Extrapolation of map data for the eastern Magdalena Mountains (Molino Peak quadrangle of Osburn and others, 1981, Fig. 1) suggest the maximum stratigraphic throw on the Chupadera fault, since 27.4 Ma, is about 1.5 to 2 km. This estimate takes into account the likelihood of mostly concealed second-order low-angle faults hidden by Pliocene and Pleistocene piedmont gravels under the Broken Tank Basin (Fig. 1; see Osburn and others, 1981, plate 2). Total Neogene extension to the WSW near the latitude of Nogal Canyon is estimated to be about 40 percent (E-E' sheet 2, shows 23 percent extension, which does not include the large low-angle Chupadera fault). Antithetic Miocene early rift faults and older caldera related antithetic normal faults, which exhibit moderate to minor down-to-the-east displacements, now have the appearance of high angle-reverse faults that dip about 55 to 90° west. Because they are interpreted as originally high-angle normal faults, rotated passively to their present position, these antithetic faults are termed "pseudo-reverse faults." In combination with low-angle normal faults, these pseudo-reverse faults appear to outline easterly rotated horsts and grabens (E-E' sheet 2). A vein filled pseudo- reverse fault of Miocene age is locally exposed east of the Red Hill mine (C-C' sheet 2). This minor fault is considered to be a third-order rift structure rotated within the dominant first-order Chupadera Mountains block.

Where Oligocene caldera structures strike northerly, they may be difficult to discriminate from rotated early rift structures, especially since they may be reactivated by rifting. For example the north-trending margin of the Sawmill Canyon caldera near the Tower Mine is largely masked by Miocene displacement (post Tpfm) on the Nancy fault. Only a local erosional unconformity of late Oligocene age (base of Txx; ca 28.7–28.0 Ma) and local drill hole data (MC-1; A-A' sheet 2) supports its caldera related origin.

Structural relationships of the Esperanza fault (eg. E-E' sheet 2) suggest that this pseudo-reverse fault may be caldera related. It may have formed by magmatic uplift of the hanging wall (west side) during local premonitory eruptions (Tzt; Tzbr) that lead to eruption of the La Jenica Tuff (Tj) from the Sawmill Canyon caldera.

The most obvious late rift structures in the quadrangle are low north-trending fault scarps, that offset graded piedmont-slopes and underlying volcanic-rich gravels of Pliocene to late Pleistocene age in the Socorro Basin. Rare exposures in alluvial deposits and in older Miocene to Oligocene "bedrock" north of the quadrangle, (near Socorro Canyon, Chamberlin, 1980, 1996) indicate that these active, or recently active, normal faults dip steeply to the east or west at angles of 60 to 75 degrees. There is no apparent indication of significant lateral slip on these north-trending high-angle faults. However, the Quaternary fault traces do make abrupt bends, splay progressively to the south, and outline local horst-graben blocks (Chamberlin, 1996). The overall pattern of active faults near Socorro (Chamberlin, mapping in progress) suggests transverse discontinuities, such as the SAZ, and Oligocene ring fractures may continue to influence the geometry of progressively extending fault blocks.

Stratigraphic and structural relationships of hydrothermal manganese oxide and calcite veins in and adjacent to the Chupadera Mountains indicate they are of latest Miocene to early Pliocene age (ca. 4–6 Ma). These manganese veins formed at about the same time as that when the eastern Magdalena Mountains became a significant source of piedmont gravels. About 2 km NNE of the Red Hill Mine, disseminated manganese oxides, silica and calcite form the cement of a 0.6 m thick conglomerate bed within the early piedmont gravels (Tsp) shed mostly from the Magdalena Mountains. Banded manganese-calcite veins in older Oligocene to Miocene "bedrock" (eg. Tlu, Thm, Tpfm) typically occupy minor north- trending high-angle faults and tension fractures (eg. Tower Mine vein). The relative youth of steep manganese-bearing breccia zones and veins at the Red Hill Mine is indicated by the unusually small angle between faults and "bedding" (bedding here is compaction foliation; C-C' sheet 2). The east flank of the Chupadera range block mostly north of Nogal Canyon, is locally bound by high-angle down-to-the-east normal faults of minor displacement and probable Pliocene to Pleistocene age (A-A', B-B' D-D', sheet 2). Pleistocene displacement of this range-bounding fault zone is evident about 2 km south of the Red Hill Mine (NW¼ sec. 28, T4S, R1W).

Map-scale folds are locally exposed in Miocene basin-fill deposits along the Shrine Valley and just north of the mouth of Nogal Canyon (sheet 1). These third or forth-order local structures may be interpreted as drape folds in more flexiable sedimentary strata over and adjacent to sharp edges of brittle fault blocks. The sinuous axis of the gentle synclinal fold within

Shrine Valley appears to coincide with block corners and scissors faults of the SAZ. This fold is locally defined by inward dipping cuestas of basalt (Tbt), which is interbedded in ductile playa muds of the Popotosa Formation. Associated brittle deformation in the basalt is likely but has not been observed in the field. The small asymmetric anticline north of Nogal Canyon is tentatively correlated with draping over the southerly projected trace of the Shrine Valley fault. Calcite and manganese oxide locally filled tension fractures in basaltic lavas that occur along the crest of this anticline (sheet 1). A very poorly exposed anticlinal drape fold is also inferred above the West Valley fault in section A-A'(sheet 2). This inferred fold helps explain rapid flattening of easterly dips in the upper Popotosa Formation (Tpf) as locally exposed on the north wall of "South" Blue Canyon. The largest and most subtle drape fold occurs above the crest on the Chupadera footwall block. The crest of this broad anticlinal fold is expressed by subhorizontal lavas and conglomerates of late Miocene age at Table Mountain (Tbt, Tpf, Tbo, sheet 1).

Narrow zones of anomalous stratal dip, generally to the north or south, are locally associated with transverse fault block corners. Examples are south dipping strata along the Olney Ranch zone in the area west of the Esperanza fault. In the northwest corner of the quadrangle, late Miocene lavas (Tsd, Tbt) and associated beds of the Popotosa Formation dip anomalously to the north-northwest at dips of 5–10 degrees. These anomalous dip zones are interpreted as faulted drape folds associated with uplifted or downwarped block corners.

Angular unconformities in the quadrangle range from distinct local discontinuities at formation boundaries, to subtle intraformational dip variations or "fanning" within depositional units. A 10–20 degree angular unconformity occurs at the top of caldera facies Hells Mesa Tuff (Thf, Thuf) everywhere observed, except south of the ring fracture zone (Chupadera Spring fault) where basal caldera-fill alluvium (Tzs1) appears to rest conformably on the inner wall facies of the Hells Mesa (Thw). An angular unconformity of 10–15 degrees is also common between Oligocene volcanic strata and Miocene fanglomerates (Tpfm, Tpf) in the northern Chupadera Mountains. In larger exposures, Popotosa fanglomerates (Tpf, Tpfm) commonly show upsection decrease in dip; examples are found west of the Torres Mine (B-B' sheet 2) and east of Nogal Canyon (E-E' sheet 2). These intraformational unconformities are almost certainly related to progressive domino-style block rotation in Miocene time.

South of Nogal Canyon, the unconformity at the base of the Popotosa Formation exhibits as much as 300 m of paleotopographic relief associated with west trending paleovalleys, but no obvious local angular disordance is present. However deeper levels of paleovalley fills are more steeply tilted than shallow levels, again implying concurrent sedimentation and block tilting. Relatively minor eastward tilting (3–5°) may have occurred in this area in late Oligocene time (ca. 30.1–28.0 Ma), most of the 20–45 degree eastward tilt, however, must have occurred in middle to late Miocene time (ca. 16–7 Ma). Pliocene and Pleistocene piedmont gravels (Tsp, QTsp and Qvo) are angularly discordant above and inset against the Miocene Popotosa Formation and older volcanic rocks (A-A' and E-E' sheet 2). Alteration zones and mineralization are discussed in a later section.

SOCORRO AND SAWMILL CANYON CALDERAS

The distribution and thickness of Oligocene volcanic rocks in the quadrangle is dominated by volcano-tectonic structures (growth faults) of the southeastern sector of the Socorro caldera, source of the 31.9 Ma Hells Mesa Tuff (Osburn and Chapin, 1983a, 1983b, Table 3). The Sawmill Canyon caldera, source of the 28.7 Ma La Jencia Tuff, is "nested" within the Socorro caldera (Fig. 2). New $^{40}Ar/^{39}Ar$ ages of rhyolitic ignimbrites and lava domes in the Luis Lopez Formation (Table 3) indicate most of this locally erupted caldera fill represents premonitory events leading to eruption of the La Jencia Tuff.

However, a brief episode (<0.1 Ma) of resurgent uplift of the Socorro caldera core, shortly after the main eruptions of Hells Mesa ignimbrite also is indicated by the following observations: 1) a minor late-stage Hells Mesa lava dome (Thr), 2) slightly older "bedded" ignimbrites (Thf), and 3) Hells Mesa-derived alluvial deposits in the basal Luis Lopez Formation (northern facies of Tzs1). Profile G-G' (sheet 2) illustrates the composite volcano-tectonic structures of the southeastern Socorro caldera in late Oligocene time at about 28 Ma. The profile line was carefully located to minimize effects of younger longitudinal rift faults.

The regional volcano-tectonic history of the SAZ and its control of the Socorro-Magdalena caldera complex has been well documented by Chapin, 1989 (also Osburn and Chapin, 1983 a, 1983b, McIntosh and others, 1991, 1992). Based on map data (Sheet 1) and preferred age data (Table 3), the Oligocene volcano-tectonic events recorded in the quadrangle are:

1) at **31.9** Ma, large volume eruption of crystal-rich quartz-rich ignimbrite and

contemporaneous collapse of the Socorro caldera; slumping and mass wasting of over-steepened southern wall during collapse yields large slabs of andesite porphyry (Thb) and rock avalanche breccias interspersed with thick "puddle" of intracaldera tuff (Thm); "dirty" lithic-rich ignimbrite grades upward into relatively clean lithic-poor ignimbrite (Thu) deposited after the main collapse phase; relatively thin upper ignimbrite (Thw) fills a shallow embayment in caldera wall south of main ring fracture zone;

2) at **31.9** Ma, following a brief lull, initially explosive eruption of small volume crystal-rich quartz-rich ignimbrite blasts "coagulated" magma from vent to produce an autoclastic lag-fall deposit (lower Thf); rapidly oscillating magma pressures then form a series of thin ignimbrites and interbedded ash falls (upper Thf); shortly thereafter the final Hells Mesa magmatic event produces slow effusion of at least one small crystal-rich lava dome (Thr) and moderate resurgent doming (10–15°) of core area (mostly west of the quadrangle);

3) from **31.9** to **30.2** Ma, during an extended lull in volcanism, stream deposits locally derived from andesitic caldera wall and rhyolitic resurgent core, fill shallow moat-like trough at Nogal Canyon (south and north facies of Tzs1) and in north sector (Tzs); a slightly younger heterolithic alluviual deposit (upper Tzs1 at Nogal Canyon) derived mostly from a 33.6 Ma pre-caldera rhyolite probably represents enlargement or breaching of the southeast caldera rim by a northwest flowing drainage;

4) beginning shortly before **30.1** Ma, eruptions of fine grained mafic lavas (Tza1) in

northeast sector and southern sector (Tza) mark onset of Sawmill Canyon caldera magmatic cycle; basaltic lavas (Tzb) may have flowed into the caldera from a vent area southeast of the caldera.

5) from **30.1** to about **29** Ma, eruption of moderate volume crystal-poor lithic-rich rhyolite ignimbrites (Tzt, Tzt1, Tzt2) occurs penecontemporaneously with andesitic lavas (Tza2, Tza3, Tzap, Tzas) that become progressively coarser grained; local unconformities along Black Canyon fault and local avalanche breccia zones (Thuf clasts) in Tzt2 and upper Tzt suggest initial uplift of Red Canyon horst during this period; andesitic to rhyolitic paleovalley fills inset into top of Tzt near Nogal Canyon probably represent lowered base level of erosion associated with a shallow depression not far to the west (source of Tzt);

6) from **28.8–28.7** Ma, magmatic uplift of Red Canyon horst was presumably accelerated contemporaneous with injection of crystal-poor rhyolite (Tirj) into Torres and Black Canyon fault zones; these dikes probably feed local rhyolitic eruptions near Black Canyon (Tzc, Tzct); central magmatic uplift reopened "old" ring fracture zone of Socorro caldera leading to emplacement of crystal-poor rhyolite tephra (Tzbrt, Tzct) and equivalent lava domes (Tzbr1, Tzbr2, Tzc) in southern and northern sectors (Nogal Canyon and near Socorro Peak; Chamberlin, 1980);

7) just prior to **28.7** Ma, rapid down cutting and back filling of paleovalleys (Tzs2) occurs on south flank of lava dome complex (Tzbr1) near Nogal Canyon; this supports presence of preexisting paleocaynon in southeast rim of Socorro caldera (see event 3);

8) at **28.7** Ma, large volume eruption of crystal-poor rhyolite ignimbrite and contemporaneous collapse of the Sawmill Canyon caldera occurs just west of the Tower Mine area; proximal outflow rapidly buried topography of southeastern "moat" and rim of Socorro caldera in Nogal Canyon-Chupadera Mountain area (Eggleston, 1982);

9) from **28.7** to **28.1** Ma, slumping, mass-wasting and erosion of eastern topographic wall of Sawmill Canyon caldera occurs near Tower mine; strongly brecciated block of Tzc and Tza3 intersected in bottom of Tower Mine drill hole (MC-1, A-A', sheet 2) probably represents ancient landslide block derived from upper Luis Lopez Formation; coarse breccia and conglomerates (Txx) are locally shed from caldera wall formed by Tzc, Tza, Thu and Thf; younger rhyolitic sandstone (Txs) may be partly eolian;

10) at **28.3** Ma, moderately crystal rich rhyolite dike is injected along preexisting andesite dike north of Black Canyon, probably representing premonitory and distal magmatism leading to eruption of Lemitar Tuff from the Hardy Ridge caldera (Fig. 2);

11) and finally at **28.0** Ma, moderate volume eruption of compositionally zoned crystal poor to crystal rich ignimbrite (Tll and Tlu) during subsidence of Hardy Ridge caldera, outflow locally overlaps east wall of Sawmill Canyon caldera near Tower Mine.

The Socorro caldera was most likely formed on the north flank of an andesitic stratovolcano complex of Eocene age (ca. 37–32 Ma; Spears Formation of Osburn and Chapin, 1983b). Indirect evidence of these older andesitic volcanoes is provided by large andesite porphyry slabs (Thb) in caldera facies tuff that were apparently derived from an andesitic terrane to the south. Also, a proximal vent area is suggested by a modertely thick pile of coarsely porphyritic andesite lavas (150 m) that underlie *caldera facies* Hells Mesa Tuff near North Baldy in the Magdalena Mountains (Krewedl, 1974). Finally, the absence of Hells Mesa outflow to the south of the Socorro caldera (eg. Caballo Mountains), as noted by McIntosh and others, 1991, also supports the presence of a precaldera volcanic highland on the south margin of the caldera. These and previously listed observations suggest that mafic to intermediate composition magmas are temporally and spatially linked to generation of the large volume silicic magma chambers that formed the Socorro-Magdalena caldera complex.

LATE OLIGOCENE ONSET OF RIFTING

Late Oligocene ignimbrites erupted from calderas to the west and southwest of the Luis Lopez quadrangle (Fig. 2; McIntosh and others, 1991 fig. 1), flowed into the quadrangle and locally buried structures associated with the Socorro caldera. From oldest to youngest these outflow ignimbrites include the La Jencia Tuff (Tj), Vicks Peak Tuff (Tvp), Lemitar Tuff (Tll, Tlu), and South Canyon Tuff (Tsc). All were erupted during the so called regional "ignimbrite flare up" (eg. Chapin, 1989) at about 28 ± 1 Ma; in this case specifically between 28.7 and 27.4 Ma (McIntosh and others, 1991, 1992). North of the SAZ, in the Lemitar Mountains, a thick pile of basaltic andesite lavas erupted from the Colorado Plateau margin (Tlp) and the younger ignimbrite sheets filled in early rift extensional fault block topography to form multiple westward thickening prisms exhibiting as much as 300 m of structural relief (Chamberlin, 1983). A thin, westward tapering wedge of La Jara Peak near the Tower Mine (A-A' sheet 2) implies relatively minor late Oligocene (ca 28.0–27.4) eastward block tilting in this area. Similar tilted fault blocks of late Oligocene age are widespread north of the SAZ, from the Joyita Hills to the eastern San Mateo Mountain (Beck, 1993; Chamberlin, 1983; Brown, 1972; and Ferguson, 1991).

Contemporaneous regional extension of the brittle upper crust, beginning at about 28.5 Ma, widespread basaltic andesite volcanism, and the ignimbrite "flare up" are most likely linked together by a regional midcrustal thermal anomaly that peaked at about this time (McIntosh and others, 1992). Discontinuous flow, expressed as closely spaced normal-faulting of the upper crust was most likely the expression of plastic flow (flattening) of the unusually hot middle crust during this time (eg. Wernicke, 1992). If this is true, then linear zones of more rapidly extending upper crust may have migrated laterally with time in response to waxing and waning of deep seated thermal anomalies.

La Jara Peak basaltic andesite lavas are not present in the southern Chupadera Mountains (Eggleston, 1982). Instead, thin (0–30 m), tabular to slightly wedge shaped locally derived rhyolitic sandstones and conglomerates (Tvs1, Tvs2, Tjus) occur between the late Oligocene ignimbrite sheets near Nogal Canyon. Local distribution patterns and thin wedge-shaped geometries of these sandstone units (eg. sec. 4 and 9, T5S, R1W) would permit the interpretation of normal faulting and *minor* block tilting contemporaneous with their deposition. Also, preferential preservation of the 28.5 Ma Vicks Peak Tuff on the downthrown western side of north to northwest trending normal faults (eg. F–F', Sheet 2) implies local block faulting and development of shallow strike valleys between 28.7 and 28.0 Ma.

In the Nogal Canyon area, late Oligocene ignimbrites are generally tilted 20 to 30° to the ENE. Older ignimbrites (Tj and Tvp) maybe slightly more tilted to the east than the younger ones (Tll, Tlu, Tsc); however, local orientations of compaction foliation are too variable to be certain of this possibility. Total differential tilt of early rift fault blocks at Nogal Canyon between 28.7 to 27.4 Ma was probably not more than 2–3 degrees, much less than the average differential tilt of 15–20 degrees observed in the Lemitar Mountains for the equivalent time period (Chamberlin, 1983; Chapin, 1989). It seems unlikely that the mid-crustal thermal regime of the southern Chupadera Mountains was significantly different from the Lemitar Mountains in late Oligocene time. The apparent difference in rate of block tilting and extension between these domains (south and north of SAZ) is therefore attributed here to relatively rapid *loading* of the upper crust north of the SAZ by as much as 500 m of basaltic andesite lavas. Calculations concerning the buoyancy forces on tilting normal fault blocks by Reiter and others (1992) show that rates of hanging wall block subsidence should be significantly accelerated by localized sedimentation on the depressed hanging wall. The apparent preferential accumulation of *dense* mafic lavas in hanging wall depressions, as in the Lemitar Mountains, is most likely an expression of significant acceleration of crustal extension by loading. The gravitational potential energy of the system must have been greatly increased by the thick pile of mafic lavas. We recognize that ignimbrite sheets and intracaldera tuffs also rapidly load the upper crust and may themselves cause accelerated crustal extension around their source calderas (eg. Ferguson, 1991; Hagstrum and Lipman, 1986), *if* the mid-crustal thermal regime is sufficiently high and tectonic boundary conditions are favorable to make "room" for extension. Timing, rate and degree of extensional strain must have varied widely in distended domains of the Socorro-Magdalena region, much like the Basin and Range province (Wernicke, 1992).

MIOCENE BASINS AND RIFTING

North trending alluvial basins in the Socorro-Magdalena region (Fig. 2) are generally interpreted as the depressed hanging wall zone of relatively wide (15–30 km) first-order tilted blocks of mostly Miocene age (eg. Chapin and Cather, 1994, fig. 7). Observations presented below suggest that local segmentation and distension by smaller second-order blocks also formed secondary strike valleys (sub-basins) that played a significant role in sediment dispersal patterns and the thickness of Miocene basin fills.

Moderate to strongly tilted volcanic-rich conglomerates that locally grade laterally and vertically into thick gypsum bearing red claystones are commonly exposed on the more recently dissected and uplifted flanks of the first-order basins near Socorro (Fig. 2). These tilted alluvial deposits are typically assigned to the Popotosa Formation (Miocene) and generally interpreted as intermontane fill of topographically closed early rift basins (Denny, 1940; Bruning, 1973) or asymmetric "half grabens" (eg. Cather and others, 1994).

K-Ar and $^{40}Ar/^{30}Ar$ ages of local volcanic units interbedded within the regional Popotosa Formation indicate that it ranges from at least 24 Ma to about 7 Ma (Osburn and Chapin, 1983;

Chapin, 1989; Cather and others, 1994; McIntosh and others, in prep.) Stratigraphic and structural relationships of dated volcanic rocks demonstrate complex temporal and spatial evolution of the Miocene Popotosa *basins* (emphasis on plural). For example, the 16 Ma basalt of Council Rock in the Abbey Springs Basin (Fig. 2) is horizontal or very slightly tilted and locally interbedded with the top of the Popotosa Formation. In contrast, the 16 Ma basaltic andesite of Silver Creek exposed on the east margin of the La Jencia Basin (Fig. 2) is tilted about 35 degrees to the west and locally lies near the base of the Popotosa Formation (Cather and others, 1994).

Volcanic-rich Popotosa conglomerates and sandstones have been commonly called "fanglomerates," and generally interpreted as piedmont-facies alluvial-slope deposits derived from stream erosion of tilted fault-block highlands within the distended Middle Tertiary volcanic terrane. Other potential sources of coarse sediment in Miocene time (see Fig. 2) were: 1) the hinge line of the downwarped Mogollon slope near the edge of the Colorado Plateau (Chamberlin and Cather, 1994), 2) The deformed "bumper like" emergent margin of the Great Plains province (Loma de las Cañas uplift, Fig. 2; Bruning, 1973), and 3) relict volcano-tectonic topography associated with the Socorro-Magdalena caldera complex.

Thick intervals of tilted and faulted gypsum-bearing red claystones and mudstones are generally interpreted as playa or intermittent lake (lacustrine) deposits of the Popotosa Formation. Presumably they accumulated preferentially in subsiding troughs, above hanging wall zones of the more active interbasin normal faults (Cather and others, 1994). Most exposures of Popotosa playa deposits are found along the uplifted western flank of the Socorro Basin from the

Ladron Mountains to Socorro Peak (Fig. 2). Discontinuous small exposures of red playa mudstones and the interbedded basalt of Broken Tank (Tbt) continue southward across the SAZ in the northern Chupadera Mountains and further south to the buried southeastern margin of the Sawmill Canyon caldera near Broken Tank Mesa (Fig. 1, Fig. 2 sheet 1, Osburn and others, 1981, pl. 1). Moderate down-to-the-northwest displacement along this northeast-striking segment of the Sawmill Canyon margin, which trends SW from Red Canyon, may have locally controlled the southern limit of Popotosa Playa deposits in middle to late Miocene time (see Osburn and others, 1981, pl.1).

Popotosa outcrops in the quadrangle are mostly preserved in hanging wall depressions adjacent to moderate and low-angle normal faults downthrown to the west (A-A' to F-F', Sheet 2). Map patterns and structure sections generally suggest rapid increases in thickness of the Miocene Popotosa beds across these down-to the west faults. Indeed, where interbedded volcanic units provide local marker horizons and time lines these abrupt thickness changes can be confirmed. The most obvious examples of Miocene growth fault relationships are found across the Chupadera fault near the Tower mine (control from Tsd) and at Broken Tank Mesa (control from Tbt; see D-D' sheet 2). Map patterns and a locally derived basalt boulder bed (Tpfb) along the low-angle Extraño fault south of Walnut Creek arroyo, also imply a Miocene growth fault relationship for this structure (see E-E' sheet 2).

Aeromagnetic flight-line profiles for the Tularosa 1x2° sheet (U.S. DOE, 1979; part of NURE program), which include the Luis Lopez quadrangle, show frequency and amplitude variations of total magnetic intensity that can be reasonably interpreted to reflect variations in thickness of Santa Fe Group basin fill under the Socorro Basin. Three east-west profiles located near the latitude of the Tower mine, the Esperanza Mine, and Table Mountain provide magnetic intensity data across the quadrangle. Inspection of *regional* aeromagnetic profile data show that exposed Tertiary volcanic rocks are associated with large amplitude and high frequency variations in total intensity. Where Tertiary volcanics are readily projected under alluvial basins, amplitude and frequency of magnetic variations become distinctly subdued (lower amplitude and broader wavelength anomalies), presumably as a general expression of basin-fill thickness (or distance between magnetic source and detector). Along these intrabasinal profiles lower amplitude magnetic highs appear to be spatially associated with the projected footwalls of the Sinuoso, Extraño and Shrine Valley faults. Thus the aeromagnetic profile data also support the inferred Sinuoso "growth fault" shown in E-E' (sheet 2). A similar association with low amplitude magnetic highs suggests that down-to-the-west Quaternary scarps in the Socorro Basin, such as the Laborcita fault and the Oeste Valle fault, are also the subtle expression of concealed Miocene growth faults (see A-A' and E-E' sheet 2).

Basal contact relationships, compositional lithofacies patterns (from volcanic clast assemblages), sparse paleocurrent observations and structural-stratigraphic relationships of interbedded volcanic units, when combined, strongly imply a distinctly different structural and depositional framework for Popotosa beds north of Red Canyon as compared to those south of Red Canyon.

In the northern area, the lower fanglomerates (Tpfm and Tpf) were locally derived from subjacent volcanic units and generally shed northerly or radially away from the Red Canyon horst (G-G' sheet 2). Usually they are less tilted than underlying volcanic strata and locally appear to fill north-trending strike valleys (Sheet 1). Equivalent, but thicker (deeper), north-trending strike-valley-fills are present in the lower Popotosa Formation of the Lemitar Mountains about 15–40 km to the north (Chamberlin, 1983). This longitudinal transport relationship suggests early to middle Miocene, paleotopography (ca 25–16 Ma) was locally controlled by the interplay of relict Oligocene volcano-tectonic highland(s) and more rapidly extending domino-type blocks to the north of the SAZ.

Red playa mudstones (Tpp) that commonly overlie lower Popotosa fanglomerates (Tpfm, Tpf) in the northern Chupadera Mountains are poorly exposed. Depositional contact relationships of these units are rarely exposed, but a few key exposures suggest the boundary is sharp to weakly gradational and may range from paraconformable to conformable. Map relationships, geothermal drill hole data (C.E. Chapin, unpub. notes), and isotopic ages of interbedded or overlapping volcanic units in adjacent quadrangles (Osburn and others, 1981; Chamberlin, 1980) show that these playa deposits are thickest (450–800 m) under the northwest flank of Socorro Peak and rapidly thin southward toward the north end of the Chupadera Mountains and the northeastern Magdalena Mountains. Late Miocene dacitic lavas (ca. 9.5 Ma; Tsd equivalent, Table 3) and tephras interbedded with thick playa deposits near Socorro Peak, also lap directly onto slightly positive bedrock highs of Oligocene rocks at the Tower Mine and Pound Ranch area (Fig. 1) to define the southern margins of the playa.

A large fraction of these late Miocene basin-floor muds (ca. 10–7 Ma) may have been delivered by growing regional stream systems (or small rivers) associated with the emerging Las Cañas uplift or the southern Colorado Plateau (Bruning, 1973; Cather and others, 1994). It also seems likely that one or more climate shifts —from normal semiarid to relatively wet and temperate—in late Miocene time (ca. 9–6 Ma; J.W. Hawley, oral commun. 1996) could have caused increased integration of the smaller middle Miocene stream systems.

In the southern Popotosa basin, south of Red Canyon, piedmont-facies conglomerates (Tpf) appear to mostly represent paleovalley fills of a westerly flowing exotic stream system that headed near the Loma de las Cañas uplift. This drainage pattern probably originated in Oligocene time as a cleft in the southeast wall of the Socorro caldera. This entrenched drainage was apparently able to maintain its overall westerly flow into the Broken Tank Basin (Miocene equivalent), in spite of ongoing easterly tilting of underlying fault blocks in the footwalls of the Extraño, Bianchi, Chupadera faults and minor unnamed intervening faults. The westerly flowing paleovalleys are most deeply entrenched in the footwall of the Bianchi fault, where buried topographic relief is as much as 300 m. Near the south margin of the quadrangle, the deepest part of this paleovalley cuts down to the top of the Luis Lopez Formation (Tzs1) where it lies adjacent to the exhumed caldera wall. Westerly trending paleovalley walls are evident (Sheet 1) at numerous stratigraphic levels depending on the local structural relief of underlying blocks (eg. F-F' sheet 2).

$^{40}Ar/^{39}Ar$ ages of basalt flows (Tbo, Tbt, Table 3) interbedded in the paleovalley-fill complex show that it is of middle to late Miocene age. Age and stratigraphic relationships of the basalt of of Broken Tank (Tbt at Broken Tank Mesa; R. M. Chamberlin, unpub. reconnaissance mapping) suggest that this paleovalley system turned northward in the Broken Tank basin and fed the southwest margin of the "Socorro Peak" playa near Broken Tank.

Paleovalleys and associated fills (including lavas) in the Bianchi footwall block are strongly tilted to the east; however contemporaneous fills near Table Mountain and Broken Tank Mesa are subhorizontal in their upper parts (late Miocene beds). This relationship suggests these mesas occupy the crest of a broad anticlinal fold associated with draping of Miocene strata over the footwall of the Chupadera fault. This large anticline would be similar in origin to smaller anticlines associated with the Oeste Valle fault (A-A' sheet 2) and the Shrine Valley fault near Walnut Creek (sheet 1).

Moderately tilted upper "fanglomerate" facies beds (Tpfu) locally exposed near "South" Blue Canyon may be associated with a poorly known transition from latest Miocene closed basins to early Pliocene thoroughgoing valley-fill deposits associated with the ancestral Rio Grande. A mudclast-rich conglomerate exposed within this facies on the north flank of "South" Blue Canyon, may represent incision and redistribution of playa muds during this early transition stage.

PLIOCENE AND PLEISTOCENE BASINS AND RIFTING

As previously described, numerous local unconformities of middle to late Miocene age in and near the Luis Lopez quadrangle generally show that modern fault block ranges and highlands (Fig. 2) were already in existence in early Pliocene time. These previously established tilt blocks were, however, apparently not as topographically prominent until development of the regional thoroughgoing ancestral Rio Grande at about 5–6 Ma. This "new" river system permitted regional erosion of older closed basin deposits and generally exhumed the flanks of the older tilt blocks. Integration of a regional river system was probably triggered in latest Miocene or early Pliocene time by a moderately short lived (ca. 1 Ma) wetter climate (J.W. Hawley, 1996, oral commun.).

The role of Pliocene and Pleistocene high-angle normal faults in outlining modern basins at Socorro is important, but may be overrated. Cumulative total displacement of the 4.0 Ma basalt of Socorro Canyon on the southeast side of Socorro Peak is approximately 208 m (Chamberlin, 1996). Thus active and recently active faults near Socorro probably do not have much more than 300 m of total displacement in Plio-Pleistocene time (last 5.3 Ma).

It has been noted (Chamberlin, 1983), that younger high-angle normal faults, of moderate displacement, locally take on the *appearance* of large range bounding faults where they truncate large displacement low-angle normal faults of Miocene age. This appears to be the case on the east flank of Socorro Peak and the northern Lemitar Mountains (Fig. 2).

The transition from Miocene closed basins to a Pliocene river network is poorly known and even more poorly exposed in the Luis Lopez quadrangle. Ongoing mapping in the Socorro quadrangle immediately to the north (Chamberlin, in progress) suggests that the earliest record of this drainage shift, consists of well-sorted crossbedded conglomerates and sandstones of early Pliocene age (post 7 Ma, pre 4 Ma) found on the uplifted south and northeast flanks of Socorro Peak. This unit is lithologically equivalent to lower Sierra Ladrones transition facies (Tsft) as mapped in the Luis Lopez quadrangle. Abundant non-volcanic (metasedimentary and plutonic) sand-sized detritus, rare recycled clasts of metasomatized lower Popotosa fanglomerate (Tpfm) and pebble imbrications suggest these well-sorted deposits represent a Pliocene *fluvial* fan eminating from the Water Canyon area of the Magdalena Mountains. This inferred fluvial fan was probably inset into and around the flanks of Socorro Peak. At Socorro Canyon, it is preserved by the overlying 4.0 Ma basalt of Socorro Canyon; northeast of Socorro Peak it is preserved by coarser, and more poorly sorted, alluvial conglomerates (Tsp) of the ancestral Water Canyon-Nogal Canyon drainage, where locally down faulted. The latter poorly sorted gravels may reflect a return to semi-arid climate prior to 4 Ma.

Similar stratigraphic relationships of gently tilted well sorted buff sandstones and sands (Tsft), overlain by poorly sorted reddish-brown conglomerates (QTsp) can be observed along "South" Blue Canyon in the quadrangle. The relationship of these Pliocene piedmont facies (fluvial and alluvial) to the ancestral Rio Grande cannot be demonstrated within the quadrangle. Field relationships in the San Acacia area (Machette, 1978) suggest that faulted and uplifted ancestral Rio Grande deposits (QTsa of Machette) as much as 270 m thick, are probably, in part, older than 4.5 Ma ("basalt of San Acacia"). Thus, regional ancestral Rio Grande deposits at the north end of the Socorro Basin may be about the same age as transitional piedmont deposits (Tsft) in the Luis Lopez quadrangle.

Exhumation of the Chupadera Mountains block must have occurred mostly in late Pliocene to early Pleistocene time. Major run off and down cutting events may have been tied to post-glacial pluvial periods, locally enhanced by the presence of small mountain glaciers on the alpine crest of the Magdalena Mountains. Early Pleistocene alluvial-fans that graded to the ancestral Rio Grande in the Luis Lopez quadrangle (Qlc/QTsp, sheet 1) were mostly derived from large canyons recently incised into the northern and eastern flanks of the Magdalena Mountains (Fig. 1, Fig.2). These high-energy intermittent streams cut through and effectively "by passed" the topographically lower Chupadera range block. Only relatively minor alluvial-fan deposits (Qpu, sheet 1) were locally derived from the Chupadera range in early to middle Pleistocene time. These smaller fans may or may not have graded to the ancestral Rio Grande.

Early Pleistocene alluvial-fan deposits intertongue with well-sorted ancestral Rio Grande sands, gravels, and floodplain mudstones near the eastern margin of the quadrangle, about 1–2 km west of I-25. New $^{40}Ar/^{39}Ar$ ages and geochemical correlation of water-laid rhyolite ash beds (Sheet 1, Table 3; Dunbar and others, 1996) that locally overlie floodplain mudstones and channel sands demonstrate that the upper Sierra Ladrones axial-river deposits (QTsf) are 1.6 to 1.2 million years old.

Stream piracy and diversion of the lower Rio Grande to the Gulf of Mexico at El Paso occurred between about 730 and 900 ka (Mack and others, 1993). Since then, the Rio Grande and its tributaries have cut down into the Sierra Ladrones beds concurrent with progressively lowering base levels of erosion. Climatically driven cycles of down cutting and back filling in middle to late Pleistocene time (eg. McGrath and Hawley, 1987) have produced a series of associated terrace and small entrenched fan deposits (Qvo1 to Qvo5) that locally intertongue with the receding (narrower) Rio Grande channel deposits (Qvof2 and Qvof3). Piedmont to axial-river facies changes of these post Santa Fe morphostratigraphic units are nearly identical to those in the Sierra Ladrones Formation. The easterly shift of these younger facies boundaries (sheet 1) reflects the entrenchment of the Rio Grande and eastward progradation of late Quaternary alluvial fans (eg. Qvo3) that unconformably overlie the ancestral Rio Grande deposits (QTsf).

Remnants of constructional graded Quaternary surfaces (Qlc and unmapped tops of Qvo1-Qvo3) are locally displaced by north-trending fault scarps. Surface displacements have been estimated visually and from "pseudo-structure contour" maps generated by projecting apparent relict topography of older surfaces. Older surfaces were "restored" in the more dissected terrains by assuming interdrainage surfaces (ridge crests) commonly approximate the maximum level of aggradation of the underlying deposits. Estimated displacements generally decrease with decreasing age of the surface, a relationship that indicates recurrent movement on the Quaternary faults. Quaternary scarps that lie to the west of the middle to late Pleistocene Socorro Canyon scarp appear to represent less active and/or older splays of the Socorro Canyon fault zone. Field relationships north of the quadrangle indicate the MCA fault was mostly active in Pliocene time

(Chamberlin, 1996).

Where the "main" Socorro Canyon fault trace transects the axial-river-piedmont facies boundary of the Sierra Ladrones Formation (near Black Canyon), there is no obvious displacement of the facies trend. However, an older buried splay, southwest of the main scarp at Red Canyon (sheet 1) does apparently displace this facies zone as much as 12 m down to the northeast. This is indicated by a northward drop in elevation of the exposed facies boundary, which is opposite to the regional trend (lower elevations to south). These limited field observations, suggest that the thickness of the lower Sierra Ladrones formation of Pliocene age (Tsft, Tsp, lower QTsp) may be significantly influenced by the MCA fault, as interpreted in sections A-A' and E-E' (sheet 2).

SOCORRO ACCOMMODATION ZONE

Published maps of the SAZ (Chapin and others, 1978; Chapin, 1989) show this ENE-trending regional tilt-block domain boundary passing under Black Mountain (Fig. 1) about 3 km north of the Luis Lopez quadrangle. Recent mapping in the Socorro quadrangle (R. M. Chamberlin, in progress) shows that the *horizontal* basalt flows capping Black Mountain (4.0 Ma basalt of Socorro Canyon) were erupted onto a gently east sloping (1–2°) piedmont surface extending from the Magdalena Mountains toward the Rio Grande. Assuming an eastward primary dip of 1–2° for these flows, then requires the Socorro Mountain block to be locally tilted as much as 2° to the west since early Pliocene time.

The locus of the SAZ is reinterpreted here (Fig. 2) to lie about 5 km south of Black Mountain at a distinct zone of ENE-striking scissors faults (fault traces marked with "S", sheet 1). Structural and stratigraphic relationships along and near this transverse boundary indicate that it has been active from Oligocene to Pliocene time. Regional ENE dip of tilt blocks on the south side of this scissors zone (SAZ) generally decrease with age, from about 30° in late Oligocene strata, to 20° in middle Miocene strata to about 5–15° in late Miocene strata. North of the scissors zone (SAZ) the blocks are tilted WNW in the area north of the Tower Mine and WSW in the footwall block of the Risco fault. The latter appears to act as a "breakaway" fault, which accommodates a longitudinal reversal in dip direction.

Prior to rifting, the nascent SAZ (Morenci lineament) apparently controlled the emplacement of Oligocene dikes associated with the Sawmill Canyon caldera (Tiaz, Tirj). The Nancy fault of Oligocene to Miocene age shows three sharp "bends" to the left (splays of mostly down-to-south displacement marked with "S") in the area north and east of the Tower Mine. The northernmost bend, about 1 km NE of Tower Mine, appears to be locally buried by a late Oligocene (ca 28.7–28.0 Ma) unconformity associated with the caldera margin. The well exposed medial bend cuts 28.0 Ma Lemitar Tuff, and the southernmost bend passes about 200 m north of the Tower Mine. Stratigraphic throw of Miocene age on the latter scissors fault is down to the south at its east end and down to the north at its west end. These Oligocene to Miocene scissors faults within the hanging wall of the Nancy fault appear to represent local block corners that shifted southward through time. Observed dip slip on the longitudinal Nancy fault, implies the sense of slip on these hanging-wall block corners is dextral oblique.

About 1.5 km NE of the Tower Mine, the Nancy fault trace splays and then turns E to ENE along a scissors fault readily interpreted as a block corner in the *footwall* of the Nancy fault zone, (NFWC, sheet 1; "C" for corner). This early to middle Miocene, east- tilted footwall block is now comprised by the older Black Canyon and Red Canyon caldera-related blocks that have been "welded" together by Oligocene intrusions. The Miocene tilt block (Nancy fault + NFWC) is now bounded by the Shrine Valley fault and the Chupadera-Nancy fault system, respectively to the east and west. Local relative uplift of this broader east-tilted block shoulder has apparently formed a faulted anticlinal warp that crests just north of this sharp bend. West of this axis, Oligocene to Miocene strata are gently tilted to the NNW (5–10°) and east of this axis, near surface beds in the hanging wall of the Risco fault are gently tilted to the ENE (sheet 1). Further to the east, the Risco hanging wall block is tilted to 15 to 25° west. A sinuous synclinal fold axis is locally defined by inward dipping cuestas of late Miocene basalt (Tbt) where the SAZ crosses the Shrine Valley. The sinistral bend in this fold axis indicates a left oblique component of torsional slip along the scissors fault in late Miocene to Pliocene time.

Minor north-striking high-angle strike-slip and oblique-slip faults of post-late Miocene age are locally present (but unmapped) along the SAZ in the northeastern sector of the Chupadera Mountains. These minor late-stage faults are defined by subhoriztonal to moderately plunging striations on near vertical jasperoid cemented fault zones. These third or forth-order structures suggest a minor north oriented compressive stress has locally occurred along the SAZ in Pliocene or Pleistocene time. Similar Pliocene to Pleistocene transverse oriented faults locally show right lateral oblique slip (SE¼ sec. 13 T4S R2W) or normal dip slip (NE¼ SW¼ sec. 5 T4S R1W).

Near surface and deeper stress fields locally associated with extensional fault block corners that are rotating in opposite directions must be exceedingly complex in space and time. However, additional work is warranted, especially to evaluate the present day stress field along the SAZ (see discussion of microearthquakes along SAZ, by Chapin, 1989). The net regional lateral shear on the SAZ is probably near zero.

ALTERATION AND MINERALIZATION

Zones of hydrothermal alteration and cogenetic mineralization, respectively represented by low-value gangue minerals and higher value ore minerals, are common aspects of ignimbrite calderas (Lipman, 1984). Flow of hot waters through fractured and porous volcanic rocks, or underlying "basement" rocks, can occur at *any* time after magma rises to shallow crustal levels. Multiple ages of hydrothermal alteration and mineralization are well documented for the Uncompahgre nested caldera complex of Oligocene age in the western San Juan Mountains, Colorado (Lipman and others, 1976). Some of the richest hydrothermal deposits in and adjacent to the Uncompahgre complex (Silverton caldera) are spatially and temporally linked with small quartz porphyry intrusions of late Miocene age that post date the caldera by 10–15 million years (Lipman and others, 1976).

Shallow magmas cool rapidly by hydrothermal convection, thus most geothermal systems should be relatively short lived (ca. 1 Ma; Lipman 1984). However, where shallow magmas are repeatedly replenished, as part of a larger longer-term magma cycle (deep-seated thermal anomaly), then the associated hydrothermal systems will remain active over longer periods of time. Also, they may become more laterally extensive in response to shifting loci of intrusions.

Available geochronologic data (Table 3 and Table 5) indicate three or four major periods of *local* silicic to intermediate volcanism and shallow intrusion that should have been capable of driving small to large hydrothermal systems in the Luis Lopez quadrangle area (c.f. G–G' sheet 2). Potential periods of hydrothermal activity since middle Oligocene time are: 32.0–31.9 Ma, 30.1–28.0 Ma, 11.3–10.3 Ma and 8.6–7.0 Ma (Table 3 and Table 5). As mentioned above, about 1 million years can be added on to these local eruptive episodes in order to estimate the potential duration of each inferred hydrothermal event.

Approximate areas of hypothetical hydrothermal activity can be estimated from the size and location of associated volcanic structures. From oldest to youngest, the inferred hydrothermal systems would have been associated with: 1) the Socorro caldera, 2) the nested Sawmill Canyon caldera, and 3) a late Miocene cluster of silicic lava domes, the locus of which generally shifted northeastward between 11 and 7 Ma. These late Miocene domes and flows generally occupy the northeast sector of the older Socorro caldera. Epithermal silver chloride and associated manganese mineralization in the Socorro Peak district (Lasky, 1932) occur within late Miocene dacite lavas and adjacent volcanic necks (Chamberlin, 1980). The southernmost limit of late Miocene silicic intrusions is indicated by the quartz-sanidine porphyry dike and plugs near Red Canyon (Tirs, sheet 1; G-G', sheet 2).

Vents for middle to late Miocene silicic domes and intermediate composition lavas are widely distributed along the SAZ (Fig. 2) from the west flank of the Magdalena Mountains to Socorro Peak (Chapin, 1989). The Miocene eruptive centers generally become younger to the north and east. Centers in the western Magdalena Mountains range from 18 to 11 Ma, whereas centers in the Pound-Ranch-Socorro Peak area range from 11 to 7 Ma (H. Newell, written commun., 1996). The apparent "core area" of late Miocene eruptive centers between Pound Ranch and Socorro Peak lies near Black Mountain about 2 km north of the Luis Lopez quadrangle. The 4.0 Ma basalt capping Black Mountain (Bachman and Mehnert, 1978, basalt of Socorro Canyon of Osburn and Chapin, 1983a) was erupted from a fissure vent on its south flank. This relationship implies the inferred late Miocene composite pluton (or plutons?) below the Pound Ranch-Socorro Peak area, had crystallized by 4.0 Ma.

Active geothermal areas provide useful analogs by which to interpret fossil geothermal systems, ie. epithermal mineral deposits. The mineralogy and composition of secondary hydrothermal minerals in active systems is primarily controlled by the chemistry of the altering fluids. Total concentration of ionic species (eg. salinity) is not a critical factor, but relative concentrations (eg. activity coefficient) are important. If the activity coefficient of potassium exceeds that of sodium, then adularia and illite would tend to be formed where reactive phases (eg. andesine and augite) are present (Browne, 1992). If sodium is more active than potassium, then albite tends to be formed.

Alkali chloride waters of near neutral pH and relatively *low* salinity are the most common type of geothermal fluid (Browne, 1992). In alkali-chloride systems the common alteration products formed by replacement of primary igneous minerals, or found as fissure fillings, include: quartz, calcite, adularia, albite, illite (or coarser sericite), smectite, chlorite, epidote, pyrite, anhydrite, wairakite (analcime), fluorite, and zeolite (laumontite and mordenite).

Intensity of hydrothermal alteration (Ia), which may be defined by the degree of replacement of a reactive primary reactive mineral, can range from nil (Ia=0.0) to 100 percent complete (Ia=1.0). Where replacement is incomplete (Ia<1.0), the alteration reactions are "frozen" in the rock sample because fluid-rock equilibrium was not achieved. In areas of widespread intense alteration, smaller areas of incomplete alteration commonly represent zones of low permeability during the period of hydrothermal activity (Browne, 1992).

Previously published descriptions of alteration and mineralization in the Chupadera Mountains have mostly emphasized the character and origin of steeply dipping manganese oxide + calcite veins that collectively represent the Luis Lopez manganese district (eg. Miesch, 1956; Farnham, 1961; Hewett, 1964; Willard, 1973; Eggleston and others, 1983; Norman and others, 1983). More recent geologic studies in the area have addressed the mineralogy, geochemistry, $^{40}Ar/^{39}Ar$ age, and thermal regime of regional K-metasomatism, which extends southeastward into the northern Chupadera Mountains (eg. Chapin and Lindley, 1986; Dunbar and others, 1994; Ennis, 1996; and Dunbar and others, 1996).

Based on reconnaissance level petrographic data for the northern Chupadera Mountains (see Appendix I) and previous publications (listed above), 4 major zones and episodes of alteration and mineralization can be recognized in late Miocene and older rocks. The following descriptions emphasize apparent age relationships of the different alteration zones and attempt to correlate them with known periods of local volcanism and intrusion as previously described. Field relationships and petrographic data indicate that two zones (or types) of alteration are probably of Oligocene age and formed in association with caldera structures. The younger alteration zones are apparently of late Miocene to earliest Pliocene age. They may represent a continuum of hydrothermal activity that changed in character (eg. CO_2-poor to CO_2-rich fluids) in response to a tentatively recognized climate shift (see discussion of Tsft under heading of Pliocene-Pleistocene basins).

Petrographic analysis of variably altered samples from the northern Chupadera range (Appendix I) provides a new perspective (at the mm scale) of relative age relationships of alteration minerals, which is in general agreement with the volcano-tectonic history of the quadrangle as previously described. Alteration zones are described below in order of apparent decreasing age. Minor zones of alteration in the southern Chupadera Mountains are loosely constrained with respect to relative age and therefore are described separately under the heading of "other alteration zones."

DEUTERIC MICROPERTHITE ZONE

Field observations show that sanidines in the lower and medial caldera-facies Hells Mesa Tuff (Thm and lower to middle Thu) are typically milky white or less commonly pearly white, as displayed on well developed cleavage facies. In contrast upper caldera facies Hells Mesa (Thf, upper Thu) is characterized by mostly clear glassy looking sanidine. Petrographic observations show that cloudy sanidines contain crystallographically aligned submicroscopic (0.02 to 0.2 mm long) rod-like and string-like inclusions distributed in wedge-shaped zones, or uniformly throughout the individual sanidine grains. In spite of abundant inclusions, the cloudy sanidines remain optically continuous and yield "fuzzy" interference figures. The abundance and size of rod like inclusions generally appears to increase at greater depth (westward) within the caldera facies tuff.

These cloudy sanidines are interpreted here as microperthites, formed by slow cooling within the thick lower levels of the intracaldera tuff. The clear microscopic rod like inclusions probably represent albite exsolved from the potassium-rich alkali feldspar at high temperatures (eg. 600–660°C; Deer, Howie, and Zussman, 1966; Kerr, 1959).

This zone of microperthitic sanidine within the caldera-facies Hells Mesa has not been mapped in detail (sheet 1). It locally appears to extend stratigraphically higher than the red sericitic alteration zone, which also appears to be mostly confined to the lower caldera-facies Hells Mesa (Thm; sericitic zone is identified with cross pattern on sheet 1). Evidence of a pre-late Oligocene age for the perthitic zone is provided by crystal poor rhyolite dikes (Tirj, sheet 1) that contain clear non-perthitic sanidine where they cut the perthitic Hells Mesa (App. I, no. 9

and 10). Also middle Miocene conglomerates and sandstones (Tpfm) locally derived from the Red Canyon horst block (G-G; sheet 2) contain a mixture of both non-perthitic and perthitic sanidines apparently representing the upper and lower caldera-facies Hells Mesa (Thf and Thm; see App. I, no. 11).

Sanidines within the stratigraphically lower part of the red-sericitic zone have a pinkish white color. In these rocks, the microperthitic albite and host sanidine may be partially altered to clay minerals and hematite.

RED SERICITIC ZONE

A large, roughly elliptical (2x5 km) zone of unusually *red* lower caldera-facies Hells Mesa Tuff centered near Red Canyon was first recognized by Willard (1973), and later mapped by Eggleston (1980). The longest axis of the "red" zone (cross pattern on sheet 1) lies near the western flank of the range where lower caldera-facies Hells Mesa is truncated by the Chupadera fault. Just north of Red Canyon the "red" zone is about 2 km wide.

As described by Eggleston and others, (1983), the "red" zone is characterized by alteration of ferromagnesian minerals and feldspars to sericite, clay minerals and hematite. Geochemical data indicate the red zone is significantly depleted in Na_2O and CaO and slightly enriched in K_2O relative to fresh Hells Mesa Tuff. Eight of ten samples from the red zone typically contain 3.0 ± 0.8 wt% Na_2O, 0.4 ± 0.1 wt% CaO and 6.0 ± 0.5 wt% K_2O (Eggleston and others, 1983, Table 1). In the same relative order, concentrations of these oxides in unaltered

Hells Mesa Tuff are typically 4.0 ± 0.3, 1.0 ± 0.2, and 4.7 ± 0.3 wt%. Eggleston and others (1983) interpreted the red zone as a relatively small epithermal zone locally *overprinted* on the southeastern sector of a much larger potassium-rich area attributed to regional K-metasomatism. Because the red alteration zone was found to be strongly depleted in manganese, Eggleston and others (1983) selected it as the most likely source area for the surrounding epithermal manganese oxide veins that comprise the Luis Lopez manganese district (Eggleston and others, 1983, fig. 5).

The concept of a hydrothermal alteration zone being *overprinted* on a preexisting alteration zone is important and warrants additional discussion. Studies in active geothermal areas have shown that where geothermal activity is centered on low-grade metamorphic rocks, the limits of modern hydrothermal alteration are difficult to detect (Browne, 1992). This is because the low-grade rocks already contain abundant albite, epidote, chlorite, illite (sericite), calcite and zeolites, formed in a low-temperature "metamorphic-hydrothermal" environment. Thus the low grade metamorphic rocks react only feebly with near neutral pH alkali chloride waters typical of geothermal systems (Browne, 1992). For the same reason, once relatively more reactive volcanic rocks have been hydrothermally altered; it seems likely that during a later hydrothermal event in the same area, the previously altered rocks will be relatively unreactive to the new system (eg. App. I, no. 7). Only where there is a significant change in the chemistry of the younger hydrothermal fluids (eg. oxidizing vs reducing) is the younger event likely to be distinctly recorded by "new" alteration products. If the later hydrothermal event produces abundant open space fillings, such as calcite and manganese oxide veins (eg. Luis Lopez vein system), then its existence would be obvious even if the host rocks were already altered.

Field observations and petrographic data, as described in later sections, indicate that manganese oxide-calcite veins of the Luis Lopez district are indeed overprinted on preexisting K-metasomatism. Recently published $^{40}Ar/^{39}Ar$ age dates of metasomatic adularia from the northernmost Chupadera Mountains (Box Canyon area, Socorro quadrangle) demonstrate that the adularia-producing system was locally active in late Miocene time from about 10.4 to 7.4 Ma (Dunbar and Miggins, 1996).

Distinctly different field relationships and petrographic data relevant to the "red" alteration zone indicate that it is probably of late Oligocene age. Structural relationships, petrographic data (app. I, no. 1) and geochemical data (previously mentioned) all strongly imply that the "red" zone is a relatively high-temperature sericitic alteration zone centered on the magmatically uplifted Red Canyon horst (G-G', sheet 2). This relatively low potassium sericitic zone (6.0 wt% K_2O) appears to be genetically linked to crystal-poor rhyolite dikes and plugs (Tirj, G-G' sheet 2) within the "red" zone that display similar sericitic alteration and also minor higher temperature "potassic alteration." The latter is expressed by traces of secondary biotite (siderophyllite?) that replace minor amounts of groundmass (quartz + feldspar + hematite) in the rhyolite dikes. This observation suggests that hydrothermal temperatures may have locally exceeded 350°C for a short period after dike intrusion (Browne, 1992). Secondary titanite, sericitic sanidine and adularia are also locally associated with the higher temperature potassic alteration (App. I, no. 9 and 10).

Other field and petrographic observations that support a pre Miocene age for the red sericitic zone are: 1) it is locally cut by an unaltered late Miocene rhyolite dike (Tirs) in the area south of Red Canyon (SW¼, NW¼, sec. 19, T4S, R1W); 2) middle Miocene fanglomerate and sandstone locally derived from the Red Canyon horst contains minor amounts of detrital sericitic plagioclase (App. I, no. 11); and 3) the overall asymmetry of the red sericitic zone (eg. widest on the west) strongly suggests that it has been uplifted, tilted to the east and truncated by the Chupadera fault zone.

Finally, if the red zone was previously K-metasomatized then it ought to contain about 8 wt% K_2O, not 6 wt% K_2O. The median K_2O content of eleven metasomatized Hells Mesa samples known to contain significant amounts of secondary adularia is 8.0 ± 0.5 wt% K_2O (Ennis, 1996, table 1, app. IV). This interpretation assumes that "overprinting" of relatively oxidizing late-Miocene K-metasomatising fluids on a relatively oxidized late Oligocene sericitic zone would have virtually no effect on the already potassic rock. In this case, secondary sericitic plagioclase is chemically unreactive, just like primary potassium-bearing sanidine and biotite (cf. App. I, no. 7).

POTASSIUM (K) METASOMATISM ZONE

Extreme enrichment of potassium, as much as 12 wt% K_2O, is commonly found in widespread crystal rich (plagioclase-rich) ignimbrites of the Socorro-Magdalena region (eg. Chapin and Lindley, 1986). The unaltered ignimbrites normally contain about 4–5 wt% K_2O.

This large K_2O anomaly is fundamentally an expression of secondary adularia (15–16 wt% K_2O) that has replaced phenocrystic and/or groundmass plagioclase (<0.4 wt.% K_2O; Deer, Howie and Zussman, 1966) in rhyolitic ignimbrites, mafic to intermediate lavas, and volcanic-rich conglomerates derived from these lavas and tuffs.

Potassium metasomatism in the Socorro region has been well studied (eg. Dunbar and others, 1994) and is the subject of ongoing research by N. W. Dunbar, C. E. Chapin and others. A considerable regional data base, pertinent to K-metasomatism near Socorro now exists. Types of data include: major and trace element geochemistry of metasomatized and fresh ignimbrites, mineralogy and abundance of metasomatic alteration products, stable isotope values ($\delta^{18}O$) of metasomatized rocks and adularia, $^{40}Ar/^{39}Ar$ ages of metasomatic adularia, and fission track data for apatite and zircon in metasomatized rocks. Most of the recent and older K-metasomatism data can be found in the M.S. Thesis of David Ennis, 1996, and in previous publications (eg. Chapin and others, 1978; D'Andrea-Dinkleman and others, 1983; Chapin and Lindley, 1986; Dunbar and others, 1994; and Dunbar and Miggins, 1996).

Interpretations of the origin of regional K-metasomatism near Socorro have evolved with the regional data base. Initially it was interpreted to be the expression of a large fossil geothermal system of early rift affinity (ie. late Oligocene to early Miocene age; Chapin and others, 1978; D'Andrea-Dinkleman and others, 1983). More recently it has been commonly interpreted as the reaction product of alkaline-saline brines that generally percolated downwards from a large evaporitic playa system (Chapin and Lindley, 1986; Dunbar and others, 1994; Ennis, 1996;

Dunbar and Miggins, 1996). This large playa system, apparently acting hydrologically like a recharge area, is presumed to have blanketed most of the Socorro-Magdalena region in middle to late Miocene time (Dunbar and others, 1994).

The Socorro K_2O anomaly is typically shown on regional maps as a large amoeboid area centered on the Magdalena range and including portions of several adjacent basins and ranges. The northern and central Chupadera Mountains lie on the southeastern margin of this regional anomaly. K-metasomatism (adularia replacing plagioclase) is commonly recognized in volcanic rocks and intercalated volcanic-rich conglomerates of middle Oligocene to late Miocene age (ca. 33–11 Ma). $^{40}Ar/^{39}Ar$ ages of metasomatic adularia from the Water Canyon and Box Canyon (1 km north of Luis Lopez qudrangle) areas range from 14.1 to 7.4 Ma (Dunbar and Miggins, 1996).

Lower Popotosa "fanglomerates" of middle to late Miocene age within the K_2O anomaly are typically very well indurated and dark red (eg. brick red). Petrographic examinations of well indurated lower Popotosa beds (eg. App. I, no. 11) reveal red jasperoidal cement. Also numerous field tests with dilute HCl indicate that carbonate is normally absent. Metasomatized lower Popotosa beds (Tpfm) and overlying playa deposits (Tpp) in the Socorro Peak-Chupadera Mountains area are obviously faulted, uplifted and eroded by later events (Chamberlin, 1980; this report). Pebbles and cobbles of red well-indurated volcanic-rich conglomerate, which are interpreted as detritus derived from metasomatized lower Popotosa, are found as rare fragments in latest Miocene(?) to early Pliocene fluvial and alluvial piedmont facies shed from the northeastern Magdalena Mountains (Tsft and Tsp, see Pliocene and Pleistocene Basins section). These observations indicate that metasomatized "bedrock" areas locally became a significant source of piedmont deposits between 7 and 4 Ma, most likely at about 6 Ma.

Key exposures of K-metasomatism boundaries in the northern Chupadera Mountains, local stratigraphic and structural relationships of metasomatized rocks, and observations of core from the Tower Mine drill hole (A-A', sheet 2; Chamberlin, 1980) suggest that the potassium-rich zone may be subdivided into an upper jasperoidal subzone (eg. Tpfm) and a lower non-silicified subzone.

UPPER JASPEROIDAL SUBZONE

A key exposure of the southern margin of regional K-metasomatism is locally mappable by variations in outcrop morphology of lower Popotosa fanglomerates southwest of the Torres Mine (Sheet 1). This alteration boundary is expressed as an abrupt south-facing topographic break between well cemented (jasperoid cemented) Popotosa conglomerates to the north (Tpfm) and poorly cemented Popotosa conglomerates of equivalent clast lithology to the south (Tpf). The boundary (hachure line, sheet 1) cuts across bedding roughly parallel to the easterly dip direction of the conglomerates. Pale orangish red conglomerates and sandstones are locally exposed in arroyo walls on the south side of the boundary. These poorly indurated beds appear to be weakly cemented with clay minerals and minor carbonate.

Ennis, 1996 (assisted by N. Dunbar and R. Chamberlin) collected ten samples of Hells

Mesa Tuff boulders and cobbles along a general north-south traverse that spans this topographic break. From north to south, 5 samples from the well indurated zone respectively contain 7.6, 7.7, 8.8, 8.4 and 6.8 wt% K_2O. Continuing southward in the poorly indurated zone, the samples respectively yielded 6.2, 5.8, 6.4, 5.7 and 5.6 wt% K_2O. The highest K_2O concentrations are from samples collected within 50 to 100 m of an extremely well indurated minor transverse fault (normal-dextral sense) that is locally filled with dark red jasper. These observations strongly suggest a silica-rich fluid, most likely carrying colloidal silica and hematite, was locally responsible for in situ K-metasomatism of the detrital Hells Mesa clasts. Based on lithic-content, most of the larger Hells Mesa clasts were apparently derived from the upper caldera-facies tuff in the Red Canyon block (Thu).

All of the Hells Mesa clasts from the well indurated K_2O-rich zone contain adularia that replaced plagioclase. Adularia and kaolinite completely replace the plagioclase except for the sample closest to the topographic break; it contains about 25% residual plagioclase (Ennis, 1996, Table 1, samples 112-116). These data indicate a gradational nature for this lateral K-metasomatism boundary. There is no field evidence that the Popotosa conglomerate originally had a significant permeability contrast along the strike of the beds in this area. If this is true, then this metasomatism boundary must have been controlled by differences in chemistry of the local interstitial fluids. This inferred geochemical fluid boundary may reflect merging or mixing zones of regional "ground water" flow or some type of flow/no-flow boundary.

The jasper and K_2O enriched zone is generally delineated by the distribution of well indurated lower Popotosa conglomerates in the northern Chupadera Mountains (Tpfm, sheet 1). Hereafter this zone enriched in K_2O and presumably silica is referred to as the jasperoidal subzone of K-metasomatism. From the Torres mine area, the southern boundary of the jasperoidal subzone trends roughly ENE, subparallel to and approximately coincident with the Black Canyon fault. North of this boundary, jasper fills virtually every well exposed fault zone or fault breccia. Exposed stratigraphic horizons likely to have had significant intergranular permeability (eg. Txx breccias, and upper nonwelded Tzt 2; stipple pattern on sheet 2) are also now well indurated by red jasperoidal silica. However, drill core at the Tower Mine shows that the subsurface equivalent of initially permeable Txx is not silicified, but is instead loosely cemented with red clay minerals that fill void spaces. This clay-rich zone is now approximately 300 m lower in elevation than surface exposures of its silicified equivalent. This significant change in alteration with depth strongly suggests that the K-metasomatism zone is internally subzoned and lacks excess silica in its lower part. This zonal interpretation is generally supported by regional data, which shows that most K-metasomatized rocks are not associated with jasper or excess SiO_2 (Dunbar and others, 1994; Ennis, 1996).

At present there is relatively little geochemical or mineralogical data available to address the origin of the upper jasperoidal subzone. The association of red well indurated lower Popotosa outcrops with K-metasomatism was recognized almost immediately (Chapin and others, 1978), but the causative jasperoidal silica cement has not been studied in detail. However, there is one local field observation that suggests the upper jasperoid zone may have been a two phase geothermal reservoir locally associated with boiling of the reservoir fluids.

Approximately 1.2 km NNE of the Tower Mine, small exposures of red jasper and brecciated white vein quartz locally occur in fractured andesite porphyry lavas (Tza3) that define the footwall of the Nancy fault. These silica-in-silica breccias are tentatively interpreted to represent hydrothermal breccias formed by explosive boiling. If this preliminary interpretation is correct, then these vein breccias would have formed within 300 m of the late Miocene land surface (Browne, 1992; Norman and others, 1983). The upper jasperoid subzone could also have been associated with cooling and oxidation of generally rising geothermal fluids. Of course boiling would also tend to cool the reservoir and raise pH favoring additional adularia precipitation (Browne, 1992).

Strong evidence of a hydrothermal origin for secondary adularia that replaced late Miocene plagioclase is found in fresh to highly altered and silicified dacite lava (Tsd) exposed several hundred meters northwest of the Tower mine. The chilled glassy base of the dacite lava contains fresh plagioclase, hornblende and biotite phenocrysts in a glassy groundmass with distributed minor Fe-Ti oxides, probably magnetite. Biotite yields an $^{40}Ar/^{39}Ar$ age of 12.23 ± 0.04 Ma (Table 3), but empirical observations and regional stratigraphic relationships suggest this rock would yield, if present, a sanidine age of about 11.0 Ma (H. Newell written commun.1996). The most likely (preferred) age of this dacite is about 11 Ma.

Upslope to the south near the middle of the dacite flow, the lava shows incipient "potassic" alteration and silicification (App. I, no. 12). Hornblende is completely replaced by a mixture of high and low birefringent clays, probably illite and smectite, plus hematite and leucoxene. Plagioclase is only slightly altered to clay minerals. Chalcedony and thin films of a dull, jet-black opaque mineral, probably Mn oxide, partially fill vugs in the lava. Adularia was not observed in this partially altered dacite.

Within a few tens of meters, at the eroded top of the flow, the dacite is highly fractured and contains abundant pink to white quartz veinlets. In thin section, plagioclase is completely replaced by adularia and clay minerals. Jasper veinlets and quartz locally cut the adularia indicating they are slightly younger (App. I, no. 13). Traces of fluorite and barite (or celestite?) are also present in the quartz veinlets. The apparent close proximity of jasper and adularia in a typical hydrothermal setting (eg. quartz veins) strongly implies that this adularia is hydrothermal in origin.

The close association of jasper and adularia also mimics that observed in well indurated lower Popotosa fanglomerates, which have been interpreted as K-metasomatized rocks of low-temperature diagenetic origin (Chapin and Lindley, 1986; Dunbar and others, 1994). As yet, the hydrothermal adularia near the Tower Mine has not been isotopically dated.

Alkaline-saline alteration of rhyolitic ash beds interbedded with semiarid lacustrine deposits of the Morrison Formation, Jurassic Lake T'oo'dichi' of Turner and Fishman (1991), has been suggested as a model for K-metasomatism near Socorro (Dunbar and others, 1994). However, the common occurrence of fresh plagioclase in basaltic lavas and in dacitic to rhyolitic ash beds interbedded in Popotosa playa deposits near Socorro Peak (Chamberlin, 1980) and in the northern Chupadera Mountains (eg. Tpbb, App. I, no. 14) does not seem to favor a Lake T'oo'dichi' model. An unknown type of greenish clay and carbonate alteration does, however, occur at some isolated exposures in the basalt of Broken Tank in the northern Chupadera Mountains. This poorly known alteration warrants further investigation.

If the K-metasomatism in the northern Chupadera Mountains is of hydrothermal origin, then the overlying playa deposits may have formed a fairly extensive impermeable caprock on a late Miocene geothermal reservoir. The jasperoidal subzone of K-metasomatism probably represents initially high permeability zones near the crest of tilted fault blocks that were locally buried by relatively impermeable playa deposits.

LOWER NON-SILICIFIED SUBZONE

Isolated occurrences of adularia-bearing and K_2O-rich Hells Mesa Tuff are known in the central and southern Chupadera Mountains (Hewett, 1964; Eggleston and others, 1983; Ennis, 1996). The best known adularia occurrence is at the MCA Mine (aka. Red Hill and State Lease open cut), where metasomatized lithic-rich caldera-facies Hells Mesa forms the wall rocks of steeply dipping manganese and calcite veins. This occurrence of adularia was first described by Hewett (1964) and later verified by Ennis (1996). Chemical analysis of the adularia bearing rock shows that it contains 7.2 wt% K_2O and 72.4 wt% SiO_2; the latter is only slightly higher than average unaltered Hells Mesa Tuff, at 72.2 wt% SiO_2 (Ennis, 1996, appendix IV, p. 149). Hematite and silica are only minor alteration phases at the MCA Mine (Norman and others, 1983). The MCA Mine area is defined here as the type occurrence of non-silicified K-metasomatism in the central Chupadera Mountains.

Another five localities in caldera facies Hells Mesa Tuff are known to contain over 7 wt% (Eggleston and others, 1983, Table 1). Two of these localities lie near the Chupadera fault, one south of the Torres Mine and the other near Nogal Canyon. Two others lie near the Shrine Valley fault and one is found within the red-sericitic zone near Red Canyon. Petrographic data also indicate non-silicified K-metasomatism occurs locally in the footwall of the Shrine Valley fault (App. 1, no. 4).

Aerial radiometric flight-line profiles (gamma ray spectrometer data; U.S. Department of Energy, 1979) that cross the Chupadera range about 1 km south of Red Canyon and 1 km south of Nogal Canyon show that the ^{40}K content of the caldera-facies Hells Mesa Tuff is 4.5 to 5.2 wt% K. Equivalent K_2O values would be about 5.4–6.2 wt% K_2O, well below that of known metasomatized rocks. Additional mapping of K-metasomatized rocks is needed in the central and southern Chupadera range. Preliminary observations suggest that metasomatism is sporadically distributed and locally associated with Miocene fault zones.

The absence of silica-enrichment in stratigraphically and structurally lower zones of potassium metasomatism can be attributed to higher hydrostatic pressures that prevented boiling and permitted a more stable temperature regime. This conclusion presumes that late Miocene potassium metasomatism in the Chupadera Mountains was formed by oxidizing alkaline-chloride geothermal waters of slightly alkaline pH (eg. pH=8.0) and unknown salinity. The primary source of thermal energy would have been late Miocene quartz porphyry intrusions centered in the Socorro Peak-Pound Ranch area, about 6–7 km north of the Luis Lopez quadrangle.

CALCITIZATION AND MANGANESE OXIDE ZONE

Epithermal vein deposits of manganese oxides and calcite that comprise the Luis Lopez manganese district (Norman and others, 1983; Eggleston and others, 1983; Hewett,1964; Farnham, 1961; and Miesch, 1956) are here referred to as the "calcitization and manganese oxide zone." The epithermal character of Luis Lopez manganese deposits is well documented by fluid inclusion and other data presented by Norman and others (1983, fig. 2). Their data indicate that the ubiquitous calcite and contemporaneous manganese oxides were mostly deposited at temperatures of 250–350° C. Most of the fluid inclusions are vapor-rich and represent boiling of the mineralizing fluids. Analyses of gases trapped in the inclusions suggest that boiling occurred primarily as CO_2 effervescence. Homogenization temperature data indicate that most mineralization near the Tower occurred at 275 to 300°C. In contrast, mineralization at the MCA Mine occurred at about 250°C. Freezing temperature data and gas analyses by Norman and others (1983) show that mineralizing fluids were of relatively low salinity, contained significant CO_2 (or bicarbonate) and notably lacked H_2S. This absence of H_2S is interpreted as a key missing link, which generally precludes the likelihood of sulfide and precious metal mineralization in the subsurface. For additional details concerning the Luis Lopez manganese district the reader is referred to the comprehensive analysis and summary of Norman and others, 1983.

Regional distribution patterns, petrographic data (app. I), and relative age relationships suggest that manganese oxide mineralization followed closely "on the heels" of local hydrothermal potassium metasomatism. The apparent close proximity of minor manganese mineralization (App. 1, no. 12) to the fringe of hydrothermal adularia formation suggest a genetic link between potassium metasomatism and mobilization of manganese oxides. In this particular example, (App. I, no. 12) completely altered hornblende (ca 0.2–0.4 wt% Mn; Deer, Howie, and Zussman, 1996) was a potential source of the tentatively identified manganese oxide. In the northern Chupadera Mountains, manganese and calcite veins consistently cut jasperoid veinlets, jasperized conglomerates (Tpfm) and other K-metasomatized rocks; thereby demonstrating that manganese mineralization post dates K-metasomatism.

Near Black Canyon, metasomatized andesites also show later replacement of adularia by calcite and epidote (App. I, no. 6). Epidote is an indicator of high temperature alteration at >260°C (Browne, 1992). Combined with the observations of Norman and others (1983), this observation suggests that carbonate replacement, particularly of Ca-bearing phases, may occur at some distance beyond the post-metasomatic manganese oxide veins. Another example of apparent calcitization, outside the margins of manganese veins, is found south of the MCA Mine (App. I. no. 2). Here the calcite replaces plagioclase in non-K-metasomatized caldera-facies Hells Mesa Tuff.

One occurrence of manganese oxide, silica, and calcite-cemented piedmont-slope conglomerate (SW¼, SE¼, sec. 9, T4S, R1W) locally suggests manganese oxide mineralization may be as young as latest Miocene or early Pliocene, or about 5 to 6 million years old. This unusual occurrence may represent the "last gasp" of hydrothermal activity of mostly late Miocene age. As with potassium metasomatic fluids, the most likely heat source for manganese depositing hydrothermal fluids was also the late Miocene quartz porphyry intrusions that underlie the Socorro Peak-Pound Ranch area. The youngest of these high-silica rhyolite lava domes is dated at 7.02 ± 0.01 Ma (Table 5). The intrusive roots of this lava dome could have supplied significant heat by itself until about 6 Ma. However, a slightly younger andesite lava and vent complex at Sedillo Hill could have added some heat to the deeper magma bodies thus allowing hydrothermal activity until perhaps 5.5 or 5 Ma.

If manganese mineralizing and K-metasomatizing fluids had the same heat source, then some external event may have changed the chemistry of the hydrothermal fluids. Field observations, mostly near Socorro Peak (Chamberlin, unpublished mapping), strongly suggest the formation of a large fluvial fan system emanating from the Water Canyon area sometime between 7 and about 4 million years ago. This fluvial fan is tentatively interpreted to represent a relatively short lived climate change, from semi-arid to temperate, roughly near the Miocene-Pliocene epoch boundary. Increased run off and deeper meteoric flow associated with this hypothetical climate shift could have significantly increased the CO_2 or bicarbonate content of the hydrothermal fluids. Obviously this hypothesis warrants additional study, but it does appear to explain the available data, particularly high-precision $^{40}Ar/^{39}Ar$ age data combined with field observations.

OTHER ALTERATION ZONES

Silicification and clayey alteration adjacent to faults in Oligocene rocks has been noted in the southern Chupadera Mountains. Just south of the Chupadera Spring ring fracture fault, the caldera-wall facies of the Hells Mesa Tuff (Thw) is locally bleached (mottled white and red) and contains white altered plagioclase and sanidine. Also, the silicified footwall of the Nogal Canyon fault locally forms steep eroded "flatirons" on the north wall of the Canyon near its eastern end. A linear zone of clayey altered plagioclase, about 90 m wide, is well exposed in the footwall of the Nogal Canyon fault where the country rock consists of Oligocene rhyolite lava (Tzbr1). Finally, the Esperanza pseudo-reverse fault is locally silicified and contains some black manganiferous calcite near the center of the quartz-jasper vein. The association of these altered zones with structures most likely formed during collapse and resurgence of the Socorro and Sawmill Canyon calderas strongly suggests these alteration zones are of Oligocene age. Considering their location a late Miocene age of alteration seems less likely. Minor manganese mineralization in the Esperanza fault might be cogenetic with the red-sericitic zone of late Oligocene age, as previously described.

ECONOMIC GEOLOGY

The most notably absent hydrothermal mineral in the Chupadera Mountains is pyrite, or its weathering product goethite (aka. limonite). Norman and others (1983) recognized a low probability for sulfide and associated precious metal mineralization in the Chupadera Mountains. The absence of H_2S and organic components in fluid inclusions of the Luis Lopez district was the basis of their conclusion.

In spite of Norman's observations, the development of jasperoidal silicified zones in the northern Chupadera Mountains is impressive and most of these zones lack obvious prospecting. One of the most impressive silicified outcrops is the jasperoidal breccia pipe developed in intrusive rhyolite (Tirj) just south of the Torres Mine (B-B' sheet 2). Several samples of this breccia pipe and other jasperized rocks from the northern Chupadera Mountains were fire-assayed for gold content, all the results were negative (<0.005 oz/ton Au). However, Willard 1973, reported one positive gold value of about 0.02 oz/ton from a sample taken at the Optimo portal about 600 m north of the Tower Mine. One minor occurrence of goethite, possibly after pyrite, has also been noted near the Torres dike zone (NW¼, NE¼, sec. 18, T4S, R1W). Cubic looking pseudomorphs of goethite found here are associated with a narrow north-trending quartz-calcite vein that cuts caldera facies Hells Mesa Tuff (Thm). Both of these areas may warrant additional sampling and evaluation for potential gold mineralization.

Approximately 50 tons of manganese oxides have been produced from the Luis Lopez district (estimate based on data of Farnham, 1961). Most of this production occurred under wartime conditions or during the aftermath when manganese was a highly priced strategic mineral. Considering modern economic conditions, none of the remaining narrow manganese veins are likely to yield economically recoverable ore. However, mineralization is known to continue deeper than the level of production at the Black Canyon Mine (Farnham, 1961; Willard, 1973).

Small areas underlain by well-sorted sand and gravel of the ancestral Rio Grande (QTsf) which also lack a significant overburden (eg. Qvo), are locally present near the mouth of Red Canyon (SE¼, sec. 11, T4S, R1W) and adjacent to I-25 near the southeast corner of the quadrangle. Both of these areas could provide moderate quantities of easily removed sand and gravel.

HYDROGEOLOGIC FRAMEWORK

Mostly unconsolidated, well-sorted, sand and gravel deposits of the ancestral Rio Grande constitute a major regional aquifer that underlies the eastern portion of the quadrangle (QTsf, sheet 1, A-A' and E-E' sheet 2). Most of the domestic water for the villages, of Luis Lopez and San Antonio, just east of the quadrangle, is drawn from shallow wells near the west margin of this regional aquifer (SE¼, sec.12, T4S, R1W; and NW¼, sec. 31, T4S, R1E; see sheet 1). Recent production of potable water from these and other nearby consumer association wells is approximately 83.9 acre feet per year (Ramsey, 1994, table 3). The thickness and storage capacity of this highly transmissive aquifer near Luis Lopez is poorly known. A resistivity log from the well west of San Antonio (E-E' sheet 2) indicates the highly permeable portion of the aquifer is about 72 m thick. However, the thickness of ancestral Rio Grande deposits may be quite variable because of contemporaneous faulting. More work is needed to define its true capacity.

Two relatively deep water wells (~ 170 m) have been drilled near the old MCA mill. The older well (1955, NW¼, sec. 23, T4S, R1W) is reported to have produced 250 gpm (Roybal, 1991), presumably as an industrial water supply for milling of manganese ore. In 1985 the BLM drilled a well near the mill site (BLM-MCA well, sheet 1) that is reported to produce as much as 800 gpm. The location and depth of these wells suggests that they do not draw from typical ancestral Rio Grande deposits (QTsf). Possible aquifer beds at this site include poorly sorted piedmont gravels (QTsp) or underlying well sorted fluvial piedmont facies sands and gravels (Tsft). The fluvial piedmont facies locally represents a transition between Miocene closed-basin-fill deposits and Pliocene through going river deposits. Considering the apparently high sustainable yield of these wells, the latter choice of aquifer beds (Tsft) seems more likely.

Thick playa deposits (Tpp) locally exposed long the northeast flank of the Chupadera Mountains (A-A; sheet 1), are generally recognized to represent a regional aquitard (Anderholm, 1987). These low permeability beds probably form a local barrier to easterly groundwater flow under the northern Chupadera range, which is primarily driven by areally significant recharge in the alpine Magdalena Mountains further west (Roybal, 1991). Small perennial man-made seeps (from Tpp, sheet 1) in the northeast sector of the range and a shallow domestic well for the Olney Ranch at Nogal Canyon, generally indicate the water table is at shallow depths (approximately. 10 m) in topographically low areas within the Chupadera range. Most of this shallow easterly groundwater flow range probably occurs along abundant fractures in the well consolidated volcanic rocks. Larger fracture zones within the range might yield significant quantities of fresh water, but there is no demand for such at present.

In summary, two regional groundwater flow systems are present in the Luis Lopez quadrangle. The southerly flowing system, which occurs in deposits of the ancestral Rio Grande, is mostly fed by alpine run off in southwestern Colorado and some smaller tributaries along the west flank of the Rio Grande.The easterly flowing system passes through fractured volcanic rocks and locally porous basin fill deposits adjacent to the Rio Grande (Roybal, 1991). Easterly flow is fed by recharge in the high Magdalena Mountains. The latter system has a shorter flow path and generally represents higher quality water. In comparison, shallow groundwaters along the east flank of the Rio Grande valley at San Antonio are relatively saline (Anderholm, 1987). Deciphering the actual potential of groundwater aquifers near Socorro will most likely require a significant financial investment in exploratory wells, which may be guided by the geologic framework provided in this report.

ACKNOWLEDGEMENTS

Financial support of Chamberlin's dissertation mapping was provided by a geothermal research grant to C. E. Chapin from the New Mexico Energy Resource Board through the Energy Institute at New Mexico State University. General operating funds of the New Mexico Bureau of Mines and Mineral Resources provided for most of the field expenses and laboratory expenses.

This report contains many small branches of a large "tree of geologic knowledge" first planted in the early 1970's by C. E. Chapin, now Director of the New Mexico Bureau of Mines and Mineral Resources. This "tree" is rooted in a solid foundation of regional geologic mapping organized and supervised by Dr. Chapin, which was originally called the "Magdalena Project." Both authors learned the basics of field geology in volcanic terranes from Dr. Chapin, a fact we appreciate very much.

The senior author owes special thanks to W. C. McIntosh and his colleagues at the New Mexico Geochronological Research Laboratory on the New Mexico Tech campus. The geologic understanding represented in this report would not have been possible without the high precision $^{40}Ar/^{39}Ar$ ages provided by the NMGRL. Also the senior author much appreciates the unwavering encouragement of F. E. Kottlowski, Director Emeritus of the NMBMMR. Finally R. M. C. wishes to say a very special "thank you" to his wife, Louise, and sons, Robert and Chris, for their loving support, which often goes beyond the call of duty.

Word processing for this report was done by Terry Gonzales. Final figures were computer drafted by Rebecca Titus. Both individuals are part of the excellent support staff at New Mexico Bureau of Mines and Mineral Resources.

REFERENCES

Anderholm, S. K., 1987, Hydrogeology of the Socoro and La Jencia Basins, Socorro County, New Mexico: U.S. Geological Survey, Water Resources Investigations Report 84-4342, 62 p.

Angelier, J. and Colletta, B., 1983, Tension fractures and extensional tectonics: Nature, v. 301, p.

49–51.

Bachman, G. O., and Mehnert, H. H., 1978, New K-Ar dates and late Pliocene to Holocene geomorphic history of the Rio Grande region, New Mexico: Geological Society of America Bulletin, v. 89, p. 283–292.

Beck, W. C., 1993, Structural evolution of the Joyita Hills, Socorro County, New Mexico: Ph.D. thesis, New Mexico Institute of Mining and Technology, 187 p.

Brown, D. M., 1972, Geology of the southern Bear Mountains, Socorro County, New Mexico: M.S. Thesis, New Mexico Institute of Mining and Technology, 110 p. New Mexico Bureau of Mines and Mineral Resources, Open-file Report 42, 110 p.

Browne, P., 1992, Hydrothermal processes and their mineralogical signatures: unpublished course notes, Geol. 572-3, New Mexico Institute of Mining and Technology, Socorro, N. M., 68 p.

Bruning, J. E., 1973, Origin of the Popotosa Formation, north-central Socorro County New Mexico: Ph.D. dissertation, New Mexico Institute of Mining and Technology, 131 p.; New Mexico Bureau of Mines and Mineral Resources, Open-file Report 42, 110 p.

Cather, S. M., 1996, Geologic maps of upper Cenozoic deposits of the Loma de las Cañas and Mesa del Yeso 7.5-minute quadrangles, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Open-file Report 417.

Cather, S. M., Chamberlin, R. M., Chapin, C. E., and McIntosh, W. C., 1994, Stratigraphic consequences of episodic extension in the Lemitar Mountains, central Rio Grande rift, *in*

Keller, G. R., and Cather, S. M. (eds.), Basins of the Rio Grande rift: Structure, stratigraphy, and tectonic setting: Boulder, Colorado: Geological Society of America, Special Paper 291, p. 157–170.

Cather, S. M., and McIntosh, W. C., 1990, Volcanogenic flood deposits near San Antonio, New Mexico: Depositional processes and implications: Sediments 1990; 13th International Sedimentological Congress, Abstracts, p. 80.

Chamberlin, R. M., 1978, Structural development of the Lemitar Mountains, an intrarift tilted fault block uplift, central New Mexico (abs.) *in* Proceedings, International Symposium on the Rio Grande Rift, Santa Fe, New Mexico, October, 1978; Los Alamos Scientific Laboratory, Conference Proceedings, LA-7487C, p. 22–24.

Chamberlin, R. M., 1980, Cenozoic stratigraphy and structure of the Socorro Peak volcanic center, central New Mexico: Ph.D. dissertation, Colorado School of Mines, 495 p.; New Mexico Bureau of Mines and Mineral Resources, Open-file Report 118, 532 p.

Chamberlin, R. M., 1983, Cenozoic domino-style crustal extension in the Lemitar Mountains, New Mexico: A summary; *in* Chapin, C. E., and Callender, J. F. (eds.), Socorro Region II: New Mexico Geological Society, Guidebook 34, p. 111–118.

Chamberlin, R. M., and Osburn, G. R., 1984, Character and evolution of extensional domains in the Socorro area of the Rio Grande rift, central New Mexico (abs.): Geological Society of America, Abstracts with Programs, v. 16, p. 467.

Chamberlin, R. M., Cather, S. M., McIntosh, W. C., Anderson, O. J., and Ratté, J. C., 1994, First-day road log from Socorro to Magdalena, Datil, Western Crosby Mountains, Sawtooth Mountains, Pie Town, Quemado and Quemado Lake; *in* Chamberlin, R. M., Kues, B. S., Cather, S. M., Barker, J. M. and McIntosh, W. C. (eds.), Guidebook of Mogollon Slope, West-Central New Mexico and East Central Arizona: New Mexico Geological Society, Guidebook 45, pp. 1–45.

Chamberlin, R. M., 1997, Pliocene and Pleistocene displacement history of the Socorro Canyon fault, central Rio Grande rift, New Mexico (abs.): New Mexico Geology, v. 18, no. 2, p. 45.

Chapin, C. E., 1989, Volcanism along the Socorro accomodation zone, Rio Grande rift, New Mexico; *in* Chapin, C. E., and Zidek, J. (eds.), Field excursions to volcanic terranes in the western United States, Volume 1, Southern Rocky Mountain region: New Mexico Bureau of Mines and Mineral Resources, Memoir 46, p. 46–57.

Chapin, C. E., and Cather, S. M., 1994, Tectonic setting of the axial basins of the northern and central Rio Grande rift; *in* Keller, G. R., and Cather, S. M. (eds.), Basins of the Rio Grande rift: Structure stratigraphy and tectonic setting: Geological Society of America, Special Paper 291, pp. 5–25.

Chapin, C. E., Chamberlin, R. M., Osburn, G. R., Sanford, A. R., and White, D. L., 1978, Exploration framework of the Socorro geothermal area, New Mexico; *in* Chapin, C. E., and Elston, W. E. (eds.), Field guide to selected cauldrons and mining districts of the Datil-Mogollon volcanic field, New Mexico: New Mexico Geological Society, Special

Publication 7, pp. 115–129, plate in pocket.

Chapin, C. E., and Lindley, J. I., 1986, Potassium metasomatism of igneous and sedimentary rocks in detachment terranes and other sedimentary basins: Economic implications: Arizona Geological Society Digest, v. 16, pp. 118–126.

Condie, K. C., 1982, Plate tectonics model for Proterozoic continental accretion in the southwestern United States: Geology, v. 10, pp. 37–42.

D'Andrea-Dinkleman, J. F., Lindley, J. I., Chain, C. E., and Osburn, G. R., 1983, The Socorro K_2O anomaly: a fossil geothermal system in the Rio Grande rift: New Mexico Geological Society Guidebook 34, p. 76–77.

Deer, W. A., Howie, R. A., and Zussman, J., 1966, An introduction to the rock-forming minerals: Wiley and Sons, New York, N. Y., 528 p.

Denny, C. S., 1940, Tertiary geology of the San Acacia area, New Mexico: Journal of Geology, v. 48, pp. 73–106.

Dunbar, N. W., and Miggins, D., 1996, Chronology and thermal history of potassium metasomatism in the Socorro, New Mexico area: Evidence from $^{40}Ar/^{39}Ar$ dating and fission track analysis (abs.): New Mexico Geology, v. 18, no. 2, pp. 50–51.

Dunbar, N. W., Chapin, C. E., Ennis, D. J., and Campbell, A. R., 1994, Trace element and mineralogical alteration associated with moderate and advanced degrees of K-Metasomatism in a rift basin of Socorro, New Mexico; *in* Chamberlin, R. M., Kues, B. S., Cather, S. M., Barker, J. M. and McIntosh, W. C. (eds.), Mogollon Slope, West-Central

New Mexico and East Central Arizona: New Mexico Geological Society, Guidebook 45, pp. 225–231.

Dunbar, N. W., McIntosh, W. C., Cather, S. M., Chamberlin, R. M., Harrison, B., and Kyle, P. R., 1996, Distal tephras from the Jemez volcanic center as time-stratigraphic markers in ancestral Rio Grande sediments from the Socorro area; *in* Goff, F., Kues, B. S., Rogers, M. A., McFadden, L. D. and Gardner, J. N. (eds.), The Jemez Mountains Region: New Mexico Geological Society, Guidebook 47, pp. 69–70.

Eggleston, T. L., 1982, Geology of the central Chupadera Mountains, Socorro County, New Mexico: M.S. thesis, New Mexico Institute of Mining and Technology, 161 p.; New Mexico Bureau of Mines and Mineral Resources, Open-file Report 141, 162 p.

Eggleston, T. L., Norman, D. I., Chapin, C. E., and Savin, S., 1983, Geology, alteration, and genesis of the Luis Lopez Manganese district, New Mexico; *in* Chapin, C. E., and Callender, J. F. (eds.), Socorro Region II: New Mexico Geological Society, Guidebook 34, p. 241–246.

Ennis, D. J., 1996, The effects of K-metasomatism on the mineralogy and geochemistry of silicic ignimbrites near Socorro, New Mexico: M.S. thesis, New Mexico Institute of Mining and Technology, Socorro, New Mexico, 160 p.

Farnham, L. L., 1961, Manganese deposits of New Mexico: U.S. Bureau of Mines, Information Circular 8030, 176 p.

Faulds, J. E., Geissman, J. W., and Mawer, C. K., 1990, Structural development of a major extensional accommodation zone in the Basin and Range province, northwestern Arizona and southern Nevada: Implications for kinematic models for continental extension, *in* Wernicke, B. (ed.), Basin and Range extensional tectonics near the latitude of Las Vegas, Nevada: Geological Society of America, Memoir 176, pp. 37–76.

Ferguson, C. A., 1991, Stratigraphic and structural studies in the Mt. Withington caldera, Grassy Lookout quadrangle, Socorro County, New Mexico: New Mexico Geology, v. 13, pp. 50–54.

Garmezy, L., 1990, Strike-slip deformation along the southeastern margin of the Colorado Plateau: Geological Society of America, Abstracts with Programs, v. 22, p. A-276.

Glazner, A. F., 1988, Stratigraphy, structure, and potassic alteration of Miocene volcanic rocks in the Sleeping Beauty area, central Mojave Desert, California: Geological Society of America Bulletin, v. 100, pp. 424–435.

Hagstrum, J. T. and Lipman, P. W., 1986, Paleomagnetism of the structurally deformed Latir volcanic field, northern New Mexico: Relations to formation of the Questa caldera and development of the Rio Grande rift: Journal of Geophysical Research, v. 91, pp. 7383–7402.

Hewett, D. F., 1964, Veins of hypogene manganese oxide minerals in the southwestern United States: Economic Geology, v. 59, no. 8, p. 1429–1472.

Jones, C. H., Unruh, J. R., and Sander, L. J., 1996, The role of gravitational potential energy in active deformation in the southwestern United States: Nature, v. 381, pp. 37–41.

Kent, S. C., 1982, Geologic maps of Precambrian rocks in the Magdalena and Chupadera Mountains, Socorro County, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Open-file Report 170, 3 sheets.

Kerr, P. F., 1959, Optical Mineralogy, McGraw Hill, New York, 442 p.

Krewedl, D. A., 1974, Geology of the central Magdalena Mountains, Socorro County, New Mexico: Ph.D. dissertation, University of Arizona, 128 p.; New Mexico Bureau of Mines and Mineral Resources, Open-file Report 44, 142 p.

Lasky, S. G., 1932, The ore deposits of Socorro County: New Mexico Bureau of Mines and Mineral Resources, Bulletin 8, 139 p.

Lipman, P. W., Fisher, F. S., Mehnert, H. H., Naeser, C. W., Luedke, R. G., and Steven, T. A., 1976, Multiple ages of middle-Tertiary mineralization and alteration in the western San Juan Mountains, Colorado: Economic Geology, v. 71, pp. 371–588.

Lister, G. S., Etheridge, M. A., and Symonds, P. A., 1986, Detachment faulting and the evolution of passive continental margins: Geology, v. 14, pp. 246–250.

Machette, M. N., 1978, Geologic map of the San Acacia quadrangle, Socorro County, New Mexico: U.S. Geological Survey, Geologic Quadrangle Map, GQ 1415, scale 1:24,000.

Mack, G. H., Salyards, S. L., and James, W. C., 1993, Magnetostratigraphy of the Plio-Pleistocene Camp Rice and Palomas Formations in the Rio Grande rift of Southern New Mexico: American Journal of Science, v. 293, pp. 19–77.

McGrath, D. B., and Hawley, J. W., 1987, Geomorphic evolution and soil-geomorphic relationships in the Socorro area, central New Mexico; *in* McLemore, V. T., and Bowie, M. R. (eds.), Guidebook to the Socorro area: New Mexico Bureau of Mines and Mineral Resources, pp. 55–67.

McIntosh, W. C., Kedzie, L. L., and Sutter, J. F., 1991, Paleomagnetism and $^{40}Ar/^{39}Ar$ ages of ignimbrites, Mogollon-Datil volcanic field, southwestern New Mexico: New Mexico Bureau of Mines and Mineral Resources, Bulletin 135, 79 p.

McIntosh, W. C., Chapin, C. E., Ratté, J. C., and Sutter, J. F., 1992, Time-stratigraphic framework for the Eocene-Oligocene Mogollon-Datil volcanic field, southwest New Mexico: Geological Society of America Bulletin, v. 104, pp. 851–871.

McIntosh, W. C., and Chamberlin, C. M., 1994, $^{40}Ar/^{39}Ar$ geochronology of middle to late Cenozoic ignimbrites, mafic lavas, and volcaniclastic rocks in the Quemado region, New Mexico: *in* Chamberlin, R. M., Kues, B. S., Cather, S. M., Barker, J. M. and McIntosh, W. C. (eds.), Guidebook of Mogollon Slope, West-Central New Mexico and East-Central Arizona: New Mexico Geological Society, Guidebook 45, pp. 165–185.

Miesch, A. T., 1956, Geology of the Luis Lopez manganese district, Socorro County, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Circular 38, 31 p.

Newell, H. N., 1997, $^{40}Ar/^{39}Ar$ Geochronology of Miocene silicic lavas in the Socorro-Magdalena area, New Mexico: M.S. Thesis, New Mexico Institute of Mining and Technology, 73 p.

Norman, D. I., Bazrafshan, K., and Eggleston, T. L., 1983, Mineralization of the Luis Lopez epithermal manganese deposits in light of fluid inclusion and geologic studies: New

Mexico Geological Society Guidebook 34, p. 247–251.

Osburn, G. R., 1978, Geology of the eastern Magdalena Mountains, Water Canyon to Pound Ranch, Socorro County, New Mexico: M.S. thesis, New Mexico Institute of Mining and Technology, 150 p., New Mexico Bureau of Mines andMineral Resources, Open-file Report 113, 160 p.

Osburn, G. R., and Chapin, C. E., 1983a, Ash-flow tuffs and cauldrons in the northeast Mogollon-Datil volcanic field: A Summary: New Mexico Geological Society, Guidebook 34, pp. 197–204.

Osburn, G. R., and Chapin, C. E., 1983b, Nomenclature for Cenozoic rocks of northeast Mogollon-Datil volcanic field, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Stratigraphic Chart 1.

Osburn, G. R., Laroche, T. M., and Weber, R. H., 1993, Geology of Lion Mountain and northern Arrowhead well quadrangles, Socorro County, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Geologic Map 68.

Ramsey, R. R., 1994, Socorro-Sierra Regional Water Plan: Soil and Water Conservation District, 203 Manzanares Avenue, Socorro, New Mexico, 98 p.

Ratté, J. C., 1989, Geologic map of the Bull Basin quadrangle, Catron County, New Mexico: U.S. Geological Survey, Geologic Quadrangle Map GQ 1651, scale 1:24,000.

Reiter, M., Barroll, M. W., and Cather, S. M., 1992, Rotational buoyancy tectonics and models of simple half graben formation: Journal of Geophysical Research, v. 97, pp. 8917–8926.

Relinger, R. E., Oliver, J. E., Brown, L. D., Sanford, A. R., and Balzas, E., 1980, New Measurements of crustal doming over the magma body, New Mexico: Geology, v. 8, pp. 291–293.

Roybal, F. E., 1991, Ground-water resources of Socorro County, New Mexico: U.S. Geological Survey, Water Resources Investigation Report 89-4083, 103 p.

Sanford, A. R., Mott, R. P., Jr., Shuleski, P. J., Rinehart, E. J., Caravella, F. S., Ward, R. M., and Wallace, T. C., 1977, Geophysical evidence for a magma body in the crust in the vicinity of Socorro, New Mexico; *in* Heacock (ed.), The Earth's Crust: American Geophysical Union, Geophysical Monograph 20, pp. 385–403.

Schneider, R. V., and Keller, G. R., 1994, Crustal structure of the western margin of the Rio Grande rift and Mogollon-Datil volcanic field, southwestern New Mexico and Southeastern Arizona; *in* Keller, G. R., and Cather, S. M. (eds.), Basins of the Rio Grande rift: Structure, stratigraphy and tectonic setting: Boulder, Colorado: Geological Society of America Special Paper 291, pp. 207–226.

Smith, R. L., and Bailey, R. A., 1968, Resurgent cauldrons: Geological Society of America, Memoir 116, pp. 613–662.

Spell, T. L., Harrison, T. M., and Wolff, J. A., 1990, $^{40}Ar/^{39}Ar$ dating of the Bandelier Tuff and San Diego Canyon ignimbrites, Jemez Mountains, New Mexico: temporal constraints on magmatic evolution: Journal of Volcanology and Geothermal Research, v. 43, p. 175–193.

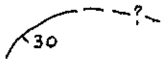
Turner, C. E., and Fishman, N. S., 1991, Jurassic Lake T'oo'dichi': A large alkaline, saline lake, Morrison Formation, eastern Colorado Plateau: GeologicalSociety of America Bulletin, v. 103, p. 533–558.

U.S. Department of Energy, 1979, Aerial radiometric and magnetic survey, Tularosa National topographic map, New Mexico: U.S. Department of Energy Open-file Report GJBX-67 (79), v. 2, data for 33 flight lines.

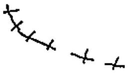
Wernicke, B., 1992, Cenozoic extensional tectonics of the U.S. Cordillera; *in* Burchfiel, B. C., Lipman, P. W., and Zoback, M. L. (eds.), The Cordilleran Orogen: conterminous U.S.: Geological Society of America, The Geology of North America, v. G-3, pp. 553–581.

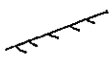
Willard, M. E., 1973, Geology of the Luis Lopez manganese district, New Mexico: New Mexico Bureau of Mines and Mineral Resources, Open-file Report 186, 81 p.

TABLE 1. Explanation of Map Symbols



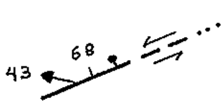
depositional contact, dashed where approximately located, queried where inferred. Direction and angle of dip shown where observed.


intravolcanic unconformity at topographic wall of Oligocene caldera (longer side of hachure toward center of caldera). Dashed where projected under alluvium.

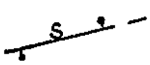

approximate projected trend of Miocene paleovalley wall; generally defined by truncation of valley floor strata or wedge out of valley fill. Hachures point toward local valley axis; acute angles point toward inferred downstream direction.

......oooooo approximate location of facies boundary (piedmont to axial river) in upper Pliocene to Pleistocene basin-fill deposits; dotted where projected under younger deposits. Placed at *western* limit of well-sorted axial river sands in the Sierra Ladrones Formation (QTsf) and *western* limit of axial river sands in post Santa Fe Group deposits (Qvof).

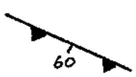

alteration boundary that cuts across bedding. Separates well indurated (jasperoidal silica cemented) and potassium metasomatized Popotosa fanglomerate facies (Tpfm) from poorly indurated, non-metasomatized Popotosa fanglomerate facies (Tpf) in area southwest of Torres Mine.



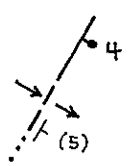
fault trace, dashed where approximately located, dotted where projected under younger deposit. Ball and bar on downthrown side. Short dash indicates direction and angle of dip. Arrow indicates bearing and plunge of striations or mullions on fault surface. Half arrows indicate apparent lateral component of slip where inferred from combination of stratigraphic offset and orientation of striations on fault surface.



trace of northeast-striking scissors fault; locally defines south margin of Socorro accomodation zone (modified after Chapin, 1989)



trace of "pseudo-reverse" fault showing direction and angle of dip. Barbs on upthrown block. Interpreted as early antithetic *normal* fault (downthrown to east) rotated to present orientation of reverse fault by dominant down-to-west rotational normal faults.



trace of Quaternary normal fault. Ball and bar on downthrown side; number indicates estimated surface displacement, at this location, in meters. Arrows indicate local downwarping along fault trend. Dip symbol in parenthesis indicates estimated local inclination of deformed land surface (sum of tilt and primary dip).

slump fault bounding small landslide block.

a local marker bed (or horizon) in upper Santa Fe Group. Includes rhyolitic water-laid ash beds (a), buried calcic soil horizons (bc), gypsiferous sandstone beds (gyp. sandst), and a manganese cemented (Mn) conglomerate bed. Lenticular beds of rhyolitic pumice and scattered pumice fragments (r) are also indicated in some localities.

c good exposure of pedogenic carbonate horizon or calcrete.

x representative exposure of late Cenozoic basin fill unit; indicates location of well sorted axial river sand or gravel where shown in QTsf.

x24 8p/2f/4p/10f selected reference exposure of basin-fill map unit showing approximate total thickness in meters, and approximate thickness of intertonguing lithofacies. Example is read as 8 m of piedmont facies, over 2 m of axial river facies, over 4 m of piedmont facies, over 10 m of axial river facies. Gravel-sand ratios (g:s) are indicated for some piedmont facies exposures. Post Santa Fe reference exposures are divided into piedmont facies (vo) and axial river facies (vof).

50 strike and dip of bedding or compaction foliation in welded tuffs.

horizontal bedding.

70 strike and dip of flow foliation in lava flow.

vertical flow foliation.

Mn 75 manganese vein or veinlets showing strike and dip.

abandoned manganese mine (name from Willard, 1973).

syncline, showing approximate trend of trough line.

orientation of flow lineation (elongated pumice) in rheomorphic La Jencia Tuff (Tj)

general direction of paleocurrents (± 30°) based on pebble imbrication.

excavation, or landfill, in late Cenozoic basin fill deposits.

c — water supply well; "C" indicates community supply well of San Antonio Mutual Domestic Water Consumers Association, "D"= minor domestic well, "S"= livestock well.

groundwater seep at man made cut and cattle tank (South Blue Canyon and Shrine Valley).

MC-1 — collar and surface projection of inclined diamond-drill hole at Tower Mine (Mountain Copper hole #1, MC-1).

b — earthen safety berm at Socorro Gun Range.

area of "red alteration" (Eggleston and others, 1983) associated with hematite replacing biotite, Fe-rich phases and some small mafic lithic fragments in lower cauldron-facies Hells Mesa Tuff.

stratigraphically and structurally controlled zones of jasperoidal silicification (hematitic chalcedony) in Oligocene volcaniclastic and volcanic rocks. Includes a silicified breccia pipe in footwall of Chupadera fault near Torres Mine, a silicified tectonic breccia of South Canyon Tuff at portal of Nancy Mine, a zone of high primary permeability in the upper pumiceous tuff (Tzt2), a basal andesite flow breccia in Tza3, and all exposures of Txx. Presumably represents same jasperoidal silicification event as that associated with potassium metasomatism of Popotosa fanglomerates in the northern Chupadera Mountains. Most mappable faults adjacent to and north of Black Canyon are cemented by jasperoidal silica.

j.b. — area of large (>10 m) jumbled blocks of South Canyon Tuff that are bound together by jasperoidal silica; exposed north of Tower Mine. May represent *very* coarse grained basal Popotosa Formation.

inferred approximate location of volcanic vent, queried where more speculative

Qlc — Las Cañas geomorphic surface of McGrath and Hawley, 1987.

8 — sample location for isotopic age analysis (numbers referred to in Table 3).

449 — sample location for paleomagnetic analysis (numbers referred to in Table 4).

TABLE 2—Description of Map Units

POST-SANTA FE GROUP UNITS

Qvy	Younger valley alluvium, piedmont facies (Holocene to uppermost Pleistocene) — Active channel, low terrace and alluvial-fan deposits of tributary arroyos. Consists of poorly sorted, nonindurated, volcanic-rich gravel, sand, silt, and clay. Associated with low graded surfaces formed during last major episode of valley entrenchment and backfilling. Thickness 0–30 m.
Qvyf	Younger valley fill, axial-fluvial facies (Holocene to uppermost Pleistocene) — Floodplain deposits of Rio Grande; consists mostly of clay to sand. Grades to well sorted gravel at basal scour surface as much as 30 m below floodplain (McGrath and Hawley, 1987). Intertongues with Qvy. Thickness 0–30 m.
Qe	Eolian deposits (Holocene) — Light gray, fine grained, well sorted sand; locally buries late Pleistocene scarp of the Socorro Canyon fault. Thickness 0–5 m.
Qca	Colluvium and alluvium, undifferentiated (Holocene to Pleistocene) — Talus and colluvium on steep to moderate slopes in the Chupadera Mountains. Also, gravelly slope wash on erosion surfaces cut in poorly consolidated upper Cenozoic deposits in Socorro Basin. Gravelly colluvium is typically shown where it masks ancestral Rio

Grande deposits (QTsf). Includes small upper Pleistocene debris flow on south wall of Blue Canyon. Widespread colluvial veneers on dissected piedmont slopes in the Socorro Basin are generally not delineated. Usually 0.3–3 m thick, locally as much as 10 m thick.

af Artificial fill and mine waste (uppermost Holocene) — Mostly represents compacted fill, as roadway subgrades along Interstate 25 and adjacent flood control levees. Includes waste-rock piles east of abandoned MCA mines and a 10 m-high tailings pile east of the abandoned MCA mill site. Thickness 0–12 m.

Qvo Older valley alluvium, piedmont facies (middle to upper Pleistocene) — Inset terrace and alluvial-fan deposits of tributary arroyos associated with five episodes of valley entrenchment and backfilling. From oldest to youngest, graded surfaces associated with these alluvial deposits project: 64–70 m (Qvo1), 43–52 m (Qvo2), 30–40 m (Qvo3), 15–21 m (Qvo4) and 6–9 m (Qvo5) above the Rio Grande floodplain. Correlation of Qvo units into western third of quadrangle is tentative. Consists of poorly to moderately sorted, bouldery to cobbly, volcanic-rich gravel, gravelly sand and muddy silts. Generally reddish orange to reddish brown, locally light brown to tan in color. Mostly nonindurated, however, uppermost beds beneath graded surfaces are variably cemented by pedogenic carbonate horizons (c) approximately 0.6 m to

0.1 m thick. Distal alluvial-fan deposits contain siliceous sand and rare quartzite pebbles recycled from the underlying Sierra Ladrones fluvial facies (QTsf). Tentative correlation with Quaternary alluvial units in the San Acacia and Las Cruces areas (based on projected height above Rio Grande) suggest these deposits range in age from about 250,000 years to 25,000 years old (Machette, 1985). Thickness 0–24 m, average thickness 6–9 m.

Qvof Older valley fill, fluvial facies and minor piedmont facies (upper Pleistocene) — Complexly intertonguing axial-river deposits (older Rio Grande) and distal alluvial-fan deposits of tributary arroyos. Locally divided into older (Qvof2) and younger (Qvof3) units based on lateral continuity with older piedmont deposits (Qvo2 and Qvo3). Consists of light gray to light yellowish brown, well sorted, fluvial sands and gravels (transport to south), interbedded with reddish orange, poorly sorted, sandy to cobbly, volcanic-rich gravels (transport to east), and red-silty mudstones. Well rounded to subrounded quartzite, granite and metamorphic rock pebbles are typical of the axial-river deposits. However, lenses of locally derived volcanic cobbles (reworked piedmont facies) are also present in axial-river deposits. Facies boundary with Qvo placed at approximate western limit (crosses on map) of axial river beds that commonly *overlie* volcanic-rich piedmont-facies gravels. As much as 24 m thick.

Qpu Piedmont deposits undifferentiated (upper to lower Pleistocene) — Coarse debris flows and poorly sorted volcanic-rich gravels, sand and muds, deposited on small alluvial fans derived from small catchment areas on the east flank of the Chupadera Mountains. Contains buried soils similar to those in QTsp; not demonstrated to intertongue with QTsf. Probably represents locally continuous aggradation, equivalent to QTsp and older Qvo deposits. Base not exposed, minimum thickness 30–60 m. Includes remnants of high-level terraces and alluvial fans, as much as 9 m thick, within the Chupadera Mountains.

SANTA FE GROUP

Intermontane basin fill of the Rio Grande rift. As redefined by Machette, 1978, includes the Pliocene and Pleistocene valley fill of the Sierra Ladrones Formation (of Machette, 1978), and Miocene bolson fill of the Popotosa Formation, named by Denny, 1940. Also includes intercalated volcanic units such as the Socorro Peak Rhyolite, rhyolitic ash-falls, and local basalt flows (Osburn and Chapin, 1983).

Qlc
QTs
Ts Sierra Ladrones Formation (lower Pleistocene to lower Pliocene) — Late-stage basin fill of the Rio Grande valley characterized by an axial-river facies (ancestral Rio Grande) and intertonguing piedmont facies (large alluvial fans) derived from

adjacent mountain ranges. Capped by Las Cañas geomorphic surface (Qlc) of McGrath and Hawley, 1987. As mapped here, younger piedmont facies (QTsp) and older piedmont facies (Tsp) lie in angular unconformity on tilted Popotosa beds at the basin margin. Also, the younger piedmont facies (QTsp) is locally inset against or disconformably overlies the older piedmont facies (Tsp). The base of the younger axial-river facies (QTsf) is not exposed here. A water well north of Walnut Creek (see sheet 2, E-E´) intersected 24.4 m of Qvo over 48.8 m of highly permeable QTsf, overlying a less permeable interval of greenish lacustrine(?) mudstones (6.1 m; E. Fry, oral commun. 1996) and fluvial sandstone (17.7 m). This lower, less permeable, interval probably represents moderately indurated, older, tributary-river deposits (Tsft). Maximum thickness of Sierra Ladrones Formation probably 200–300 m.

Qlc Las Cañas geomorphic surface of McGrath and Hawley, 1987 (lower Pleistocene). Erosional remnants of geomorphic surface representing maximum level of aggradation of late Cenozoic basin fill along the Rio Grande valley near Socorro. Defines top of Santa Fe Group and Sierra Ladrones Formation where present. Characterized by discontinuously exposed laminar calcrete as much as 2 m thick. Locally mantled with thin desert pavement or dark red argillic horizon. Projects 91 to 101 m above modern Rio Grande. Correlation with geomorphic surface in type area, east of the Rio Grande, is based on equivalent landscape position and soil development. Tenatively correlated with the lower La Mesa surface of Las Cruces area, which has recently been dated at 0.73–0.9 Ma (Mack and others, 1993).

QTsp Sierra Ladrones Formation, upper piedmont facies (lower Pleistocene to upper Pliocene) — Alluvial-fan deposits below Las Cañas surface (Qlc) of McGrath and Hawley, 1987. From north to south, large, low-gradient alluvial fans (paleoslopes of 120–140 ft/mi) emanate from Socorro Canyon, Red Canyon, Nogal Canyon (Walnut Creek) and "South Chupadera Spring Canyon." Distinctive volcanic clasts demonstrate that all but the latter were fed by large canyons in the eastern Magdalena Mountains, namely Sixmile Canyon, Molino Canyon and Caronita Canyon. Abundant metamorphic clasts and large basalt boulders in the "South Chupadera Spring" alluvial fan indicate that it was derived from a large ampitheatre between Table Mountain and Chupadera Mountain. Consists of reddish orange to reddish brown, poorly consolidated, volcanic-rich, boulder to cobble conglomerates, conglomeratic sandstones, muddy sandstones, and silty mudstones; generally fining to the east. A muddy transitional distal-fan facies (subscript "d") is locally delineated where well exposed in the area north of lower Red Canyon. The distal facies consists mostly of reddish orange silty mudstones and fine-grained sandstones, with a few pebbly conglomerates. Exposures of the distal facies average about 500 m in width and lie just west of the QTsp–QTsf facies boundary. A fine-grained, gypsum cemented sandstone within the distal facies (SE¼ Sec. 11, T4S, R1W) probably represents an overbank splay deposit of the ancestral Rio Grande. Buried calcic soil horizons (bc) are common within the piedmont facies. The gently east tilted base of the upper piedmont facies (QTsp), is locally exposed in the walls of "South Blue Canyon" where it appears to conformably overly poorly exposed fluvial fan deposits (well-sorted sands) of the lower Sierra Ladrones Formation (Tsft). Maximum thickness of upper piedmont facies, estimated from cross section (sheet 2, A A´) is 140 m. West of the MCA fault, on a structurally higher block, the uppermost piedmont facies is 9–18 m thick.

QTsf Sierra Ladrones Formation, axial-fluvial facies (lower Pleistocene to upper Pliocene) — Axial river deposits of the ancestral Rio Grande, and minor intertonguing distal piedmont-slope deposits (see reference outcrops on sheet 1 for examples of intertonguing localities). QTsf–QTsp facies boundary is placed at *western* (mountainward) limit of well sorted, light-gray to pale yellowish brown, quartz-rich river sands or pebbly gravels (crosses on map). Piedmont facies gravels that locally *cap* the fluvial facies (east of the facies boundary) are delineated as QTsp on the map. Axial river sands are mostly non indurated, however, a few lenses of friable calcite-cemented sandstone are locally present (NE¼ sec. 30, T4S, R1E). Pebbly gravel lenses typically contain well-rounded to subrounded clasts of quartzite, chert, granite, metamorphic rocks, and various volcanic rocks (a few recycled from local piedmont facies). Lenses of rhyolitic pumice (r) are also locally present. River sands often fine upwards into red or greenish gray, poorly indurated, mudstones that probably represent floodplain deposits or abandoned channel fills. Water-laid rhyolitic ash beds (a), apparently derived from eruptions in the Jemez Mountains (Dunbar and others, 1996), commonly overlie floodplain mudstones near the western limit of the axial-river facies. Ash beds are 0.1–0.6 m thick and occur at three stratigraphic levels approximately 18, 27 and 33 m below the projected top of QTsf. The stratigraphically highest ash found near Walnut Creek, was deposited at 1.22 Ma or slightly later (Table 3, no. 1); and the middle ash (Table 3, no. 3A) is geochemically equivalent to pumice dated at 1.6 Ma (Table 3, no. 2 and 3). The oldest ash is widely exposed in the area southwest of Luis Lopez. This undated ash may be equivalent to the 1.78 Ma San Diego Canyon ignimbrite of Spell and others, 1990, (N. Dunbar, pers. commun. 1996). Base of axial-river facies is not exposed in the Luis Lopez quadrangle. Minimum thickness, based on water well north of Walnut Creek, is 72.5 m. Possibly 300 m thick near down faulted axis of Socorro Basin.

Tsp Sierra Ladrones Formation, lower member, piedmont facies (lower to upper Pliocene) — Easterly transported piedmont-slope deposits (SE¼, NE¼, Sec. 8, T4S, R1W). Consists of poor to moderately indurated, light brown to pale reddish brown, conglomeratic sandstones with abundant locally derived volcanic clasts and rare dark red clasts of jasperoidal cemented (potassium metasomatized) lower Popotosa Formation (Tpfm). Lies in angular unconformity on upper Popotosa Formation (Tpf) and is disconformably overlain by upper Sierra Ladrones piedmont facies (QTsp). Locally contains an unusually well indurated conglomerate bed, which is cemented by manganese oxides and silica (SW¼ sec. 9, T4S, R1W). Lithologically similar to easterly transported piedmont-slope conglomerates that underlie 4.1 Ma basalt flow at Socorro Canyon (sec. 27, T3S, R1W; Chamberlin, 1980). An intertonguing relationship with lower Sierra Ladrones fluvial-tributary facies (Tsft) is indicated by exposures on the northeast flank of Socorro Peak (R. Chamberlin, unpublished mapping, 1996). Minimum exposed thickness is 90 m.

Tsft Sierra Ladrones Formation, lower member, fluvial-fan transition facies (lower Pliocene) — Light gray to yellowish brown, well sorted, fine to medium-grained, moderately quartz-rich sand and sandstone, with a few thin beds of volcanic-pebble conglomerate and red or pale green mudstone. Black heavy-mineral laminations are common in the well sorted, sand beds. Calcite cemented conglomerate and sandstones more common toward base. Well sorted fine-grained sands contain about 60 percent monocrystalline quartz, 20 percent feldspar (microcline and non volcanic plagioclase), and 20 percent volcanic lithic fragments. Considering their location, these relatively arkosic and volcanic-poor sandstones may represent the distal portion of a large fluvial fan emanating from the Water Canyon area of the Magdalena Mountains. Transition unit is poorly exposed in walls of "South Blue Canyon." Similar well-sorted quartz-rich sands and volcanic-rich conglomeratic sandstones locally underlie ancestral Socorro Canyon alluvial conglomerates (Tsp) and the 4.1 Ma basalt of Socorro Canyon (Chamberlin, 1980) at Socorro Canyon. The same unit is also found on the upper northeast flank of Socorro Peak, where it contains numerous well sorted crossbedded conglomerates and sandstones of *fluvial* character (Chamberlin, mapping in progress). Easterly paleocurrents and rare pebbles or small cobbles of well indurated (K-metasomatized) lower Popotosa Formation (Tpfm) in the Socorro Peak exposures indicate this fluvial unit is partly recycled from uplifted lower Popotosa beds in the Magdalena Mountains. Unit probably marks transition from closed basin to axial river deposition, possibly in response to a wetter climate. Transition unit was probably inset against Popotosa playa beds and Miocene lavas on flanks of Socorro Peak. Thickness estimated from cross section A-A' is 90–120 m; it maybe as much as 300 m thick north of Socorro Peak and it is probably absent due east of Socorro Peak.

Tp Popotosa Formation, undivided (lower to upper Miocene) — Intermontane bolson fill deposits of early Rio Grande rift half grabens. Defined by Denny, 1940; and redefined by Machette, 1978. Locally divided into a conglomeratic piedmont (fanglomerate) facies (Tpf, Tpfm), a claystone dominated playa (lacustrine) facies (Tpp), a transitional mudstone-sandstone facies (Tpt), an upper fanglomerate facies (Tpfu), and a basaltic boulder-rich bed (Tpfb). Contains interbedded ashes and lava flows of Miocene age (Osburn and Chapin, 1983; Cather and others, 1994). Undivided Popotosa is shown in cross section only.

Tpfu Popotosa Formation, upper "fanglomerate" (piedmont) facies (uppermost Miocene) — Light brown sandstone and volcanic-rich conglomeratic sandstone locally overlying thick playa facies near "South" Blue Canyon. One exposure of mudclast-rich conglomerate (NE¼, SW¼ sec. 4 T42 R1W) suggests reworking of underlying playa facies at disconformity(?). If disconformity can be demonstrated, then this unit could be reassigned to the basal Sierra Ladrones Formation (as part of Tsft). Maximum exposed thickness is 75 m.

Tpp Popotosa Formation, playa facies (lower to upper Miocene) — Mostly red or maroon claystone with minor greenish claystones and thin bedded, yellowish brown sandstones. Gypsum, as selenitic veinlets, occurs near north boundary of quadrangle. Claystones are poorly indurated and commonly masked by colluvium (Qca). Appears to conformably or paraconformably overlie piedmont facies sandstones and conglomerates (Tpfm) in northern Chupadera Mountains; grades laterally to distal piedmont deposits northeast of Socorro Peak. Thickness at "South Blue Canyon" is 460 m; northeast of Strawberry Peak (Socorro 7.5′ quadrangle) the playa facies is 850 m thick (Chamberlin, 1980).

Tpf Popotosa Formation, "fanglomerate" (piedmont) facies (lower to upper Miocene) — Heterolithic volcanic-rich conglomerates and sandstones derived from stream-erosion of early rift, tilted fault blocks. Generally described as "fanglomerates" (Bruning, 1973) although alluvial fan geometries have not been delineated. Consists of poorly to moderately well indurated, tan, buff and pale red, conglomerates, conglomeratic sandstones, sandstones and minor mudstones. Conglomeratic sandstones are dominant, and range from moderately well sorted stream deposits with well developed pebble imbrications, to poorly sorted, matrix supported debris flow deposits. Pebble imbrications indicate that westerly paleo- transport directions are dominant in the lower half of the piedmont facies at Walnut Creek,"South Chupadera Spring Canyon," and Table Mountain. Sequence adjacent to Walnut Creek fines upwards from boulder conglomerate at base to pebble conglomerate below the intercalated basalt flows (Tbt). Above the basalt flows, the sequence generally coarsens upwards. Basal contact is a slight angular unconformity or disconformity associated with as much as 100 m of erosional relief. Colluvial breccias derived from the immediately underlying volcanic unit are common at the basal contact. Maximum exposed thickness adjacent to Walnut Creek is 730 m.

Tpfb Popotosa Formation, basaltic boulder-rich bed (upper Miocene)—Poorly exposed bouldery bed in hanging wall of Extraño fault, primarily expressed as a narrow (30–60 m) north-trending belt of loose (surface mantling) basaltic blocks and subangular to subrounded boulders. In situ basaltic boulders are locally exposed in light gray rhyolite-dominated conglomerates near the north end of the belt of basaltic float (unit dips 15° east here). Basaltic boulders consist of two types: dense dark gray to black diabase and vesicular, medium gray, moderately porphyritic basalt. These lithologics are texturally and compositionally similar to zones in the basalt of Olney Ranch. Basalt boulders are as much as 1–2 m across near the south end of the belt and generally become smaller toward the north end (30–50 cm). This unique boulder bed, within the Popotosa piedmont facies (Tpf), was probably derived from erosion and mass wasting (?) of a faulted cuesta of the basalt of Olney Ranch located on the uplifted footwall of the Extraño fault in late Miocene time. Estimated maximum thickness of boulder bed is 6–12 m; unit grades northward into typical heterolithic conglomerates (Tpf).

Tpfm Popotosa Formation, silicified and potassium metasomatized "fanglomerate" facies (lower to upper Miocene) — Silicified and potassium metasomatized lateral equivalent of Tpf in northern Chupadera Mountains; characterized by dark red to brick red color and extreme induration in association with jasperoidal silica cement. Silicification boundary (hachure line) locally cuts across fanglomerate beds southwest of the Torres Mine (SE¼, sec. 13, T4S, R2W). Fanglomerates here were derived from the footwall of the low-angle Chupadera fault and contain a mixture of Hells Mesa Tuff clasts and crystal poor rhyolite clasts similar to Tirj. Chemical analyses of Hells Mesa clasts collected across this color and induration boundary demonstrates the red, well indurated exposures to the north are potassium metosomatized and poorly indurated exposures to the south are not metasomatized (Ennis, 1996). Overall, the silicification and potassium metasomatism boundary, as defined by well indurated (jasperized) Popotosa outcrops, trends ENE across the northern Chupadera Mountains, subparallel to the Black Canyon fault zone. Jasperoidal silica also cements fault breccias and zones of presumed high initial permeability in the underlying volcanic units (heavy stipple pattern) in the northern Chupadera Mountains. $^{40}Ar/^{39}Ar$ ages of metasomatic adulalaria from the Socorro Canyon (Box Canyon) area indicate the K-metasomatism system was active from about 10.4 to 7.4 Ma (Dunbar and others, 1994; Dunbar and Miggins, 1996). Silicified and metasomatized fanglomerates west of the Torres Mine are at least 300 m thick.

Tpt Popotosa Formation, transitional facies (upper Miocene) — Consists of red non-gypsiferous mudstones and volcanic-rich lithnic arenites, may be transitional to a playa facies or an axial-fluvial facies. Occurs near middle of piedmont facies at Walnut Creek; 0–60 m thick.

Tbt Basalt of Broken Tank (upper Miocene)—Moderately widespread trachybasalt flows interbedded within Popotosa Formation on southeastern; western and northern flanks of the Chupadera Mountains. Singular to multiple stacked flows range from dark gray to grayish black, slightly pophyritic to fine-grained diabasic, dense to microvesciular and amygdaloidal, trachybasalt (47–50% SiO_2). Slightly porphyritic basal zones locally contain sparse fine-grained (<1.5 mm) phenocrysts of plagioclase, clinopyroxene and olivene, the latter commonly altered to red-brown iddingsite. Amygdaloidal calcite and calcite veinlets are fairly common. Dense cores of flows typically exhibit fine grained diabasic textures, which range from ophitic to subophitic. Flows north of Walnut Creek are greenish gray, propylitically altered and locally contain minor manganese mineralization; a low precision age date from this altered zone (Table 3, no. 9) is superceded by a more precise analysis derived from outcrops along strike to the south (Table 3, no. 4). New $^{40}Ar/^{39}Ar$ ages (Table 3, no.

4–10) and petrographic studies (Chamberlin, unpub. data) indicate that the basalt of Bear Canyon (Osburn and Chapin, 1983, Chamberlin, 1980), is temporally and compositionally equivalent to the basalt of Broken Tank. The "basalt of Bear Canyon" (Osburn and Chapin, 1983) is here subsumed into the basalt of Broken Tank. Outcrop distribution, thickness variations, and paleocurrent observations in underlying Popotosa conglomerates (Tpf) are consistent with a concelaed source area near San Antonio. Flows at Walnut Creek apparently flowed westward down a piedmont slope and paleovalleys toward Table Mountain and Broken Tank. From Broken Tank these lavas flowed northward down a basin axis and then spread out as a thin flow on playa deposits near Bear Canyon. Correlation of basaltic lavas 1 km SW of MCA mine is tentative. Average of 9 $^{40}Ar/^{39}Ar$ age dates on this unit is 8.42 ±0.3 Ma (McIntosh and Chamberlin, unpublished data). Stacked flows at Walnut Creek and near Broken Tank are as much as 30 m thick; singular flow near Bear Canyon is 3–6 m thick. Unit locally wedges out against paleovalley wall about 2 km SE of Broken Tank (NW1/4 Sec. 31, T4S, R1W).

Tbo Basalt of Olney Ranch, new name (upper Miocene)—The new name "basalt of Olney Ranch" is here proposed for exposures of a porphyritic tabular basalt flow on the north flank of Table Mountain (SW1/4 Sec. 7, T5S, R1W), which lies on the southwest portion of Olney Ranch (formerly Nogal Ranch, sec. 6, T5S R1W) located in the west-central Chupadera Mountains. Correlative outcrops of this unit are found in the east-central Chupadera Mountains about 4.5 km ESE of Table Mountain. The basalt of Olney Ranch consists of medium to dark gray, slightly to moderately porphyritic trachy-basalt (49% SiO_2) with about 1–4 percent fine to medium grained (1–4.5 mm) phenocrysts of tabular plagioclase and subhedral greenish olivine. Distinctly porphyritic basal zones are 3–6 m thick and typically grade upwards into a dense core zone of fine-grained diabase with rare medium-grained phenocrysts of plagioclase. Sample from northeast Table Mountain yields an $^{40}Ar/^{39}Ar$ age of 9.64±0.06 Ma (Table 3, no. 11). Petrographically similar basalt outcrops in the east central Chupadera Mountains dip about 20° east, in sharp contrast with the nearly flat lying outcrop on Table Mountain. Similar attitude differences are observed in the younger basalt of Broken Tank (Tbt, sheet 1). Basalt of Olney Ranch is interbedded with poor to moderately indurated conglomerates and sandstones of the Popotosa Formation that locally show westerly paleocurrent directions and inset paleovalley geometries with respect to underlying Oligocene volcanic rocks. The basalt of Olney Ranch apparently flowed westward down a Miocene paleovalley from a now concealed source area between San Antonio and the Bosque del Apache. Thickness 0–10 m; locally wedges out against older rhyolite (Tzbr1) at paleovalley wall on south side of Table Mountain.

Tsd Socorro Peak Rhyolite, rhyodacite member (upper Miocene) — Light gray to pale red, flow banded, porphyritic rhyodacite lava flow containing 20–30 percent phenocrysts (1–3 mm) of plagioclase (locally replaced by adularia) hornblende, and biotite. Outcrops north and west of Tower Mine contain what appears to be hydrothermal adularia in close proximity to numerous chalcedony and minor jasperoid veinlets of irregular geometry. Unaltered dark gray vitrophyre is locally associated with the chilled base of flow approximately 700 m northwest of Tower Mine. Biotite from unaltered basal vitrophyre yields an $^{40}Ar/^{39}Ar$ age of 12.23 ± 0.04 Ma (Table 3, no. 13) which is analytically equivalent to new $^{40}Ar/^{39}Ar$ ages of biotite for the upper dacite flow of the Pound Ranch member of the Socorro peak rhyolite, collected about 4 km to the west of Tower Mine. However, the upper dacite at Pound Ranch overlies lower rhyolite that yields an average $^{40}Ar/^{39}Ar$ age from two sanidine separates of 11.31 ± 0.07 Ma (H. Newell, written communication 1996). Assumed equivalent sanidine age for Tsd is approximately 11 Ma (Table 3). Source area uncertain, possibly erupted from Pound Ranch volcanic center; round knob immediately NW of Tower mine *might* represent a vent area. Maximum exposed thickness on downthrown western side of Chupadera fault is 110 m.

MOGOLLON GROUP

Tm Mogollon Group undivided, shown in cross section only. The Mogollon Group includes all *volcanic* formations in the upper part of the Mogollon-Datil volcanic pile, which range in age from approximately 29 to 24 Ma (Cather and others, 1994). The Spears Group (not seen in Luis Lopez quadrangle) includes all *volcaniclastic sedimentary* formations throughout the Mogollon-Datil volcanic pile, except volcaniclastic sedimentary units associated with calderas (Cather and others, 1994). As delineated here, the Mogollon Group includes 5 regional ignimbrites, locally erupted volcanic units and volcaniclastic-sedimentary fill within the Socorro caldera (Luis Lopez Formation), and a thin tongue of regional basaltic andesite lavas (La Jara Peak Basaltic Andesite). Unnamed rhyolitic sandstones and conglomerates (Tvs1 and Tvs2) that occur locally between ignimbrite outflow sheets are not assigned to the Spears Group because they are more likely related to early rift block faulting than constructional volcanic highlands. Formation nomenclature is from Osburn and Chapin, 1983.

Tlp2 La Jara Peak Basaltic Andesite, upper tongue (Oligocene)—Medium to dark gray trachybasalt or basaltic trachyandesite lava flows, with sparse fine-grained phenocrysts of olivine altered to reddish brown iddingsite. Locally fills paleovalley cut in South Canyon Tuff on south side of Nogal Canyon (SW1/4, Sec. 6, T5S, R1W). Undated flow is tentatively correlated with the "basalt of Madera Canyon" as mapped by Osburn et al (1981), on the basis of similar lithology and stratigraphic position. The "basalt of Madera Canyon" is here renamed the basalt of Rincon

Madera Canyon, member of the La Jara Peak Basaltic Andesite, in order to avoid confusion with the Madera Formation of Pennsylvanian age. The basalt of Rincon Madera Canyon member has recently been dated at 26.95±0.2 Ma (McIntosh and Chamberlin, unpublished data). Thickness 0–20 m.

Tsc South Canyon Tuff (Oligocene) — Partially to densely welded, light gray to pale grayish red, phenocryst-poor to moderately phenocryst-rich, pumiceous, high-silica rhyolite ignimbrite. Medium grained (1–3 mm) phenocrysts of subequal quartz and sanidine, with traces of biotite and plagioclase, progressively increase upwards from about 5% near partly welded base to as much as 25% near densely welded top (where preserved). Generally lithic poor except near base; light gray pumice (1–5 cm) is moderately abundant (5–15%). Moderately crystal rich upper zone only locally preserved below Miocene erosion surface. Represents remnants of thin outflow sheet erupted from the Mount Withington caldera in the northern San Mateo Mountains (Ferguson, 1991). Mean $^{40}Ar/^{39}Ar$ age is 27.37 ± 0.07 Ma; magnetic polarity is reverse (McIntosh and others, 1991). Correlation here is based on lithology and relative stratigraphic position. Thickness 30–60 m.

Tlp1 La Jara Peak Basaltic Andesite, medial tongue (Oligocene) — Reddish brown to dark gray, massive to vesicular, fine-grained, basaltic andesite lava flows characterized by small (1–2 mm) moderately abundant (10–15%) Fe-Mg phenocrysts altered to hematite or iddingsite. The latter presumably represent altered olivine. The La Jara Peak Basaltic Andesite, thickens greatly to the north and west, towards a probable eruptive center near Riley, 50 km northwest of Luis Lopez. Distal margin of formation in area of Tower Mine is 0–30 m thick.

Tvs2 Volcaniclastic sedimentary unit, upper (Oligocene) — White rhyolitic sandstone, siltstone and pebble conglomerate in area south of Nogal Canyon. Occupies same stratigraphic interval as the La Jara Peak Basaltic Andesite at Tower Mine. Thickness 0–30 m.

Tlu
Tll Lemitar Tuff; upper and lower members (Oligocene) — Compositionally zoned (77–65 wt% SiO_2), ignimbrite subdivided into a partially to densely welded, light gray, phenocryst-poor (5–15%), rhyolite lower member (Tll), and a densely welded, dark red, phenocryst-rich (30–45%), quartz latite upper member (Tlu). Contains sparse to abundant, medium-grained (1–4 mm) phenocrysts of quartz, sanidine, plagioclase, and biotite with traces of augite and sphene. Lower third of upper member is relatively quartz poor (<5%) compared to upper two thirds, which is quartz rich (10–15%). Small (1–3 cm) phenocryst-poor pumice is moderately abundant (3–5%) in lower member. Sparse, phenocryst-rich pumice and small (<2

cm) grayish red "magma blobs" of dacite porphyry are typical in outflow of the upper member. Distribution and thickness trends suggest that the Lemitar Tuff was erupted from the Hardy Ridge caldera on the west flank of the Magdalena Mountains (Fig. 2; G. R. Osburn oral commun. 1999; see McIntosh and others, 1991, fig. 29). Lemitar ignimbrite locally appears to pinchout against eastern wall of the older Sawmill Canyon caldera in the Tower Mine area. Mean $^{40}Ar/^{39}Ar$ age is 28.00 ± 0.08 Ma; paleomagnetic polarity is normal (McIntosh and others, 1991). Correlation here based on distinctive zonation in drillhole at Tower Mine (MC-1), a low precision $^{40}Ar/^{39}Ar$ age of 27.38 ± 0.48 Ma from slightly altered sanidine (Table 3, no. 16), and determination of normal polarity for the lower Lemitar Tuff exposure northeast of the Tower Mine (W. C. McIntosh, unpublished data). Thickness 0–230 m.

POST COLLAPSE FILL OF SAWMILL CANYON CALDERA

Tx Sawmill Canyon Formation, undivided (Oligocene) — Shown in cross section only. Volcaniclastic sedimentary deposits mostly derived from the eastern wall of the Sawmill Canyon caldera (Osburn and Chapin, 1983); locally exposed northeast of the Tower mine.

Txs
Txx Sawmill Canyon Formation, upper and lower members (Oligocene) — Fining upward sequence is divided into lower breccia and conglomerate member

(Txx) about 40–50 m thick and an upper sandstone member (Txs) 20–25 m thick. In surface exposures the lower member is well indurated by red jasperoidal silica, which is associated with potassium metasomatism. Lower member consists of rhyolitic breccia (ancient colluvial deposit derived mostly from flow-banded Tzc) grading upward into crudely bedded debris flows with abundant locally derived clasts of Tzc, Tza3 and Thu. Upper member consists of pale red to grayish red, well sorted, fine-grained rhyolitic sandstone with minor conglomeratic beds near the base. Upper sandstone may be partly eolian. Tower Mine drill hole (MC-1) intercepted 73.8 m of Sawmill Canyon Formation below the Lemitar Tuff (Chamberlin, 1980); Txx in the drill hole appears to be potassium metasomatized and clay rich, but it is *not* cemented by jasperoidal silica. Thickness 0–70 m.

Tvs1 Volcaniclastic sedimentary unit, lower (Oligocene) — Light gray tuffaceous sandstone and pebble conglomerates probably derived from the underlying La Jencia Tuff. These alluvial sandstones may locally represent late-stage erosional enlargement of the southeastern wall of the Sawmill Canyon caldera, since they occupy the same stratigraphic position as the Sawmill Canyon Formation (Txs/Txx). Alternatively Tvs1 may be related to erosion of uplifted early rift fault blocks, since Tvs1 also occupies the stratigraphic position of the Vicks Peak Tuff (Tvp). Wedge-shaped geometries (SW¼ sec. 4, T5S, R1W) locally favor the latter interpretation.

Thickness 0–24 m.

Tvp Vicks Peak Tuff (Oligocene) — Light gray to pale red, phenocryst poor, densely welded rhyolite ignimbrite. Distinctive aspects include lithophysal zone near base and large pumice lapilli as much as 30 cm long near the top. Contains 1–5 percent phenocrysts of sanidine and sparse quartz. Thin outflow sheet here was erupted from the Nogal Canyon caldera in the southern San Mateo Mountains (Deal and Rhodes, 1976). Appears to be preferentially preserved on the downthrown (southwestern) side of northwest to north- striking normal faults near southern margin of quadrangle. Mean $^{40}Ar/^{39}Ar$ age is 28.56 ± 0.06 Ma; paleomagnetic polarity is reverse (McIntosh and others, 1991). Correlation here based on lithology and relative stratigraphic position. Thickness 0–125 m.

ROCKS ASSOCIATED WITH COLLAPSE OF SAWMILL CANYON CALDERA

Tjus Upper zone of La Jencia Tuff and minor sedimentary rocks (Oligocene) — Brick red, densely welded, phenocryst poor, rhyolite ignimbrite with abundant dark gray to black pumice fiamme. Contains sparse (1–3%) small phenocrysts of sanidine and quartz with traces of plagioclase and biotite. Grades downward into main body of the La Jencia Tuff (Tj). Brick red upper La Jencia Tuff is preserved only where locally overlain by the Vicks Peak Tuff (Tvp). As mapped here, includes a thin interval of tuffaceous sandstone and pebble conglomerate that discontinuously overlies the brick red zone. Thickness 0–30 m.

Tj La Jencia Tuff, main body (Oligocene) — Light gray, pale red and grayish red, phenocryst poor, rhyolite ignimbrite characterized by a thick medial zone of very densely welded, rheomorphic (flow banded) ignimbrite. Flow-banded core grades to normal eutaxitic ignimbrite near base and top. Contains sparse (3–5%) phenocrysts of sanidine and quartz with traces of plagioclase and biotite. Orientations of lineated (stretched) pumice are consistent with a source area to the west or northwest. Erupted from the composite Sawmill Canyon-Magdalena caldera in the north central and western Magdalena Mountains (Osburn and Chapin, 1983). Mean $^{40}Ar/^{39}Ar$ age is 28.85 ± 0.04 Ma; paleomagnetic polarity is reverse (McIntosh and others, 1991). Correlation here based on lithology, relative stratigraphic position and sanidine age of 28.75 ± 0.05 Ma (Table 3, no. 17). Thickness in area south of Nogal Canyon ranges from 45–360 m; locally absent in northern Chupadera Mountains.

POST COLLAPSE FILL OF SOCORRO CALDERA

Tz Luis Lopez Formation (Oligocene) — Heterogeneous fill of the Socorro caldera consisting of volcaniclastic sedimentary rocks, pumiceous lithic-rich ignimbrites, mafic to intermediate lavas, and rhyolitic lavas and tuffs. $^{40}Ar/^{39}Ar$ ages of volcanic units (Tzt1 and Tzbr2; Table 3, no. 24 and 18) near the bottom and top of the formation, respectively 30.1 ± 0.07 Ma and 28.61 ± 0.17 Ma, suggest that they are premonitory events related to a developing magma system which culminated in eruption of the La Jencia Tuff from the Sawmill Canyon caldera. Overall distribution and thickness, of the Luis Lopez Fm indicate that it filled a preexisting depression created by eruption of the Hells Mesa Tuff. Local unconformities in the Luis Lopez Formation suggest that the Red Canyon horst (Sheet 2, G-G') was uplifted (presumably by magma pressure) *during* eruption of volcanic units in the upper Luis Lopez Formation. The general correlation of six members within the Luis Lopez Formation, from north to south, is primarily based on lithologic similarity, relative stratigraphic position and new $^{40}Ar/^{39}Ar$ ages (Table 3, no. 18–28). Luis Lopez Formation typically lies in moderate angular unconformity (5–12°) on caldera facies Hells Mesa Tuff, except south of the ring fracture zone (at the south margin of quadrangle) where the contact is conformable. Facies relationships of early sedimentary fill see desription of (Tzs1) near Nogal Canyon imply moderate uplift of a resurgent core shortly after the main caldera forming eruption. Formation north of Black Canyon is approximately 700 m thick; south of Nogal Canyon maximum thickness is approximately 900 m.

Members of Luis Lopez Formation in southern Chupadera Mountains:

Tzs3 Upper sedimentary member (Oligocene) — Light gray, light brown or pale red rhyolitic conglomerates sandstones and minor mudstones. Conglomerates locally fill paleovalley cut in underlying rhyolite lava (Tzbr1) in area south of Nogal Canyon. Assignment to Luis Lopez Formation is arbitrary. Exposed thickness 0–30 m, possibly 90 m thick in subsurface.

Tzbr2
Tzbrt
Tzbr1 Upper rhyolite member, rhyolite of Bianchi Ranch, 3 units (Oligocene) — Includes upper rhyolite flow unit (Tzbr2), medial rhyolite tuff unit (Tzbrt) and lower rhyolite flow unit (Tzbr1). All units of the Bianchi Ranch Member have reverse paleomagnetic polarity (McIntosh and others, 1991). Composite thickness of Bianchi Ranch Member is 0–550 m.

Upper unit (Tzbr2) is a phenocryst poor rhyolite lava flow consisting of a basal dark gray vitrophyric flow breccia that grades upwards into a dense flow-banded black vitrophyre and an upper zone of reddish brown spherulitic rhyolite. Contains sparse, small phenocrysts of plagioclase, sanidine and rare biotite. Upper flow unit wedges out to south against crest of underlying lava dome (Tzbr1). Yields $^{40}Ar/^{39}Ar$ age of 28.61 ± 0.17 from bulk sanidine sample (Table 3, no. 18). Exposed thickness of upper unit is 0–60 m.

Medial tuff unit (Tzbrt) consists of light gray, phenocryst poor, pumiceous, crossbedded, base-surge deposits and thin, poorly welded, pumiceous ignimbrites compositonally similar to the overlying rhyolite flow unit (Tzbr2). Fragments of lithoidal rhyolite; vitrophyric rhyolite and andesite are locally abundant. Interpreted as a proximal pyroclastic deposit associated with a vent clearing event. Thickness is 0–20 m.

Lower flow unit (Tzbr1) is a rhyolite lava flow and dome complex that appears to be centered (thickest) near the latitude of Nogal Canyon. Dominant lithology is light brown, phenocryst-poor, flow-banded, lithoidal rhyolite with minor vitrophyric and spherulitic zones. Contains sparse phenocrysts of plagioclase, sanidine and biotite with traces of hornblende and sphene. Steep flow foliations and paleotopographic crest of lava dome south of Nogal Canyon (center, sec. 5, T5S, R1W), suggest a primary vent near here. Inferred vent area probably marks a point on the older ring fracture zone of the Socorro caldera. Correlation of a similar phenocryst-poor rhyolite locally overlying Thm near Madera Canyon (SE¼ sec. 36, T4S, R2W) with Tzbr1 is tenuous. Maximum thickness near Nogal Canyon is 550 m; unit locally wedges out to south near caldera margin (Sheet 2, G-G′).

Tzas Upper andesitic member, andesitic breccias, lavas and minor sandstones of rhyolitic composition (Oligocene) — Consists predominantly of well indurated, maroon, andesitic conglomerates and sandstones with thin intercalated andesite lava flow and flow breccias. Andesite porphyry lavas contain 15–25 percent phenocrysts of fresh to variably altered medium-grained plagioclase, pyroxene and hornblende. Andesitic conglomerates also contain sparse rhyolite tuff and lava clasts. Grades to light gray rhyolitic (tuffaceous) sandstones near south end of outcrop belt. Locally fills paleovalleys cut in Tzt, thickness 0–30 m.

Tzt Middle tuff member (Oligocene) — Light gray to pale red, poorly welded, pumiceous, lithic-rich, rhyolite ignimbrites with intercalated debris flows near the base. Debris flows contain abundant andesite clasts and rare rhyolitic clasts, including Hells Mesa Tuff fragments. Small (1–3 cm) aphyric to very crystal poor, white pumice lapilli form 10 to 20 percent of the rock. Pumice is commonly altered to white clay of zeolites; rare phenocrysts of plagioclase, sanidine and quartz indicate the pumice is rhyolitic. Lithic fragments of andesite porphyry are common through out the unit; clasts of densely welded Hells Mesa Tuff are abundant in the uppermost 30 m. Sparse crystals (3–5%) of sanidine, plagioclase, quartz and biotite found in the matrix of the upper zone are most likely xenocrysts derived from fragmentation of the accompanying Hells Mesa clasts. A thin bed of water laid ash and mudstone that contains plant impressions occurs at the base of the medial tuff member about 1 km north of Nogal Canyon. Sanidine from lower zone yields an 40Ar/39Ar age of

30.06±0.14 Ma (table 3, no. 25), which is analytically equivalent to the age of Tzt1 near Black Canyon (30.1±0.06; table 3, no. 24). A lithologically similar outflow sheet in the Socorro-Datil region is commonly mapped with the base of the La Jencia Tuff and correlated with a tuff or tuffs in the lower South Crosby Peak Formation (Osburn and others, 1993; McIntosh and Chamberlin, 1994). As much as 150 m thick, locally wedges out to south against caldera wall.

Tza Lower andesitic member (Oligocene) — Dark gray to reddish gray and bluish gray, phenocryst-poor (5–10%) andesitic lava. Contains small phenocrysts of pyroxene and plagioclase with traces of biotite and quartz. Platy fracture and weakly developed flow banding are common. As much as 180 m thick where preserved on a structural bench south of the Red Canyon fault. Tza is absent south of Nogal Canyon where equivalent stratigraphic interval (below Tzt) is occupied by basaltic lavas (Tzb).

Tzb Lower basaltic member (Oligocene) — Dark gray to black, massive to vesicular, trachybasalt lava flows characterized by sparse (3–7%) small phenocrysts of iddingsite after olivene. Groundmass separates from samples near Nogal Canyon yield a low precision $^{40}Ar/^{39}Ar$ age of 28.32±0.98 Ma (Table 3, no. 27) and a moderately precise plateau age of 29.2±0.47 Ma (Table 3, no. 28). Preferred age of these basaltic lavas is approximately 30.2 Ma, since they are immediately overlain by

Tzt dated at 30.06 Ma. As much as 120 m thick; pinches out to south near caldera wall at Chupadera Spring. Tzb is absent to north where same stratigraphic interval is occupied by andesitic lava (Tza). Lateral relationship of Tzb and Tza is concealed.

Tzs1 Lower sedimentary member (Oligocene) — Monolithic to heterolithic volcaniclastic sedimentary rocks consisting of a southern andesitic facies, a northern rhyolitic facies and a late-stage heterolithic facies. Southern facies was presumably derived from the southern wall of the Socorro caldera and northern facies was presumably derived from a weakly resurgent dome or horst of caldera-facies Hells Mesa Tuff to the north or northwest. Southern facies generally fines upward from coarse andesitic debris flows and breccias, to bed-load conglomeratic sandstones and mudstones, all of which are mostly purple to dark red in color. Northern facies, at Nogal Canyon, consists mostly of light gray to pale red rhyolitic sandstones, mudstones and minor conglomerates with rare Hells Mesa clasts. Uppermost 20 m near Nogal Canyon consists of cobble-rich conglomeratic sandstones derived from an unknown highland of spherulitic rhyolitic lavas and minor andesite lavas. Sanidine from spherulitic rhyolite clasts yield $^{40}Ar/^{39}Ar$ ages of 33.74±0.10 and 33.53±0.14 Ma (Table 3, no. 29 and 30); which supports the intepretation that they were derived from precaldera volcanic units on the south margin of the caldera. Overlies Hells Mesa Tuff near Chupadera Spring and in area about 1.5 km south of Nogal Canyon. As much as 240 m thick, wedges out to south against Precambrian rocks that locally define caldera wall just south of map area (Eggleston, 1982).

Members in northern Chupadera Mountains: Note that Oligocene volcanic rocks in northern Chupadera Mountains are commonly potassium metasomatized. In metasomatized rocks, plagioclase is replaced by white clay or adularia, pyroxenes by hematite and clays, and sphene by leucoxene. Biotite and sanidine commonly appear to be unaltered. Other types of hydrothermal alteration are also present.

Tzc
Tzct

Upper rhyolite member, rhyolite of Cook Spring (Oligocene) — Locally divided into an upper rhyolite lava flow unit (Tzc) and a lower cogenetic tuff unit (Tzct). Composite thickness of erosional remnants northeast of Tower Mine is 0–30 m. Upper flow unit (Tzc) consists of light gray to pinkish gray, finely flow banded, phenocryst-poor rhyolite lava. Contains small phenocrysts (2–5%) of sanidine and plagioclase with traces of quartz and biotite. Lower tuff unit (Tzct) consists of light gray bedded ash-fall tuffs and thin welded ignimbrites mineralogically equivalent to overlying flow unit. Welded ignimbrite in lower tuff unit near Highway 60 yields a K-Ar age of 29.4 ± 1.1 Ma from biotite (Osburn and Chapin, 1983). Drill hole, MC-1, intersected 60 m of intensely brecciated Tzc; probably representing an Oligocene slump block on the eastern wall of the Sawmill Canyon caldera.

Tzap
Tza3

Upper andesitic member (Oligocene) — Locally divided into a coarsely porphyritic upper flow unit (Tzap) and a lower medium-grained porphyritic flow unit (Tza3). Upper flow unit consists of medium gray to reddish brown andesitic to rhyodacitic lava with abundant (5–20%) phenocrysts of altered plagioclase, pyroxene and minor fresh biotite. Large subhedral to rounded plagioclase phenocrysts (5–15 mm long) are commonly replaced by pink adularia and less commonly by white clay. Northeast striking, coarsely porphyritic dikes and a circular plug near Socorro Canyon represent probable sources for the upper flow unit. Upper flow unit is 0–30 m thick. Lower flow unit (Tza3) consists of dark purplish gray, gray, and reddish brown, medium grained (1–3 mm) andesite porphyry lavas, flow breccias and finely vesicular, near vent, cinder deposits. Massive to platy and locally vesicular lavas contain moderately abundant (5–20%) phenocrysts of altered plagioclase and pyroxene. Northeast striking andesite porphyry dike about 2 km NE of Tower Mine appears to be a feeder dike for Tza3 flows. Thickness 60–180 m.

Tzt2
Tza2
Tzt1

Middle tuff and lava member (Oligocene) — Subdivided into an upper pumiceous tuff unit (Tzt2), a *locally* intercalated andesite flow unit (Tza2) and a lower pumiceous tuff unit (Tzt1). Upper tuff unit (Tzt2) consists of light brownish gray, poorly welded, pumiceous, lithic-rich rhyolitic ignimbrite compositionally similar to uppermost 30 m of Tzt in southern Chupadera Mountains. Contains sparse to abundant (5–50%) pebble to boulder size, rounded to angular, lithic fragments of densely welded Hells Mesa Tuff and andesite porphyry. Locally includes thin debris flows of same clast lithology in lower part. Hells Mesa clasts appear to be coarsest (90–120 cm) in the lower Tzt2 in the area north of Black Canyon. Upper part of Tzt2 is intensely silicified by red jasperoid in area north of Black Canyon. Sparse crystals of sanidine, quartz and biotite in matrix are probably xenocrysts. As much as 300 m thick near Black Canyon.

Andesite flow unit (Tza2) is a purplish gray to dark gray, moderately porphyritic to aphanitic, andesite lava flow. Contains sparse, small phenocrysts of plagioclase and pyroxene replaced by metasomatic adularia and Fe oxides, respectively. Thick hogback former cut by Black Canyon appears to represent a singular massive flow, about 160 m thick, which may have ponded against the Black Canyon fault zone. Locally erupted from northeast striking feeder dike (SE¼, sec. 5, T4s, R1W). Maximum thickness 160 m; absent in southern Chupadera Mountains and on Socorro Peak (Chamberlin, 1980).

Lower tuff unit (Tzt1) consists of light gray, poorly welded, pumiceous, lithic-rich, rhyolitic ignimbrite compositionally and texturally similar to lower part of Tzt in southern Chupadera Mountains. Andesite porphyry clasts are locally abundant and Hells Mesa Tuff clasts are typically rare to absent. Rare crystals of sanidine and quartz in matrix are probably primary phenocrysts. Sanidine from lower Tzt1 at Black Canyon yields a precise $^{40}Ar/^{39}Ar$ age of 30.1 ± 0.07 Ma (Table 3, no. 24) and a less precise age of 29.85±0.31 Ma (Table 3, no. 23). Medial tuff (Tzt2, Tzt1, and Tzt) was probably erupted from a now buried vent area within the eastern Sawmill Canyon caldera (Fig. 2), which is referred to here as the Black Canyon caldera. Tzt, Tzt1 and Tzt2 are here tentatively correlated with a regional ignimbrite that is generally assigned to the lower South Crosby Peak Formation (Osburn and Chapin, 1983; Osburn and others, 1993; McIntosh and Chamberlin, 1994).

Tza1 Lower andesite member (Oligocene) — Purplish gray, aphanitic to slightly porphyritic, andesite lava and minor andesitic conglomerates. Intensely altered lavas at Black Canyon contain metasomatic adularia and hematite replacing plagioclase and pyroxene respectively; also later stage calcite and epidote partially replace adularia and hematite. Vesicular or amygdaloidal zones common. Thickness 0–30 m.

Tzs Lower sedimentary member (Oligocene) — Light gray, pinkish gray, and purplish gray, medium to coarse grained rhyolitic sandstones, and thin andesitic debris flows. Rhyolitic sandstones most likely derived from underlying Hells Mesa Tuff. May intertongue with lower part of Tza1. Thickness 0–30 m.

ROCKS ASSOCIATED WITH COLLAPSE OF SOCORRO CALDERA

DATIL GROUP: Includes all volcanic strata younger than the Baca Formation and older than (and including) the Hells Mesa Tuff (Osburn and Chapin, 1903; Cather et al, 1994)

Th__ Hells Mesa Tuff (Oligocene) — Mostly densely welded, reddish, phenocryst-rich (40–50%), quartz-rich, lithic-rich, rhyolite ignimbrite. Divided into 6 intracaldera facies (4 stratigraphic intervals), primarily on the basis of size, abundance and lithology of enclosed lithic fragments. Typically contains abundant, medium grained (1–3 mm) phenocrysts of sanidine, plagioclase, quartz and minor biotite (in order of decreasing abundance). Plagioclase is commonly and widely altered to "clay" minerals in association with potassium metasomatism or "argillic" alteration. Biotite and other Fe-rich phases, including small mafic lithic fragments, are typically replaced by hematite in the red alteration zone, centered north of Red Canyon. Cleavage faces of normally clear sanidine (and non-clayey plagioclase) commonly show milky to pearly and pink pearly lusters within and around the red hematitic zone. Erupted from Socorro caldera, which extends from Chupadera Mountains westward across the Magdalena Mountains (Osburn and Chapin, 1983). Mean $^{40}Ar/^{39}Ar$ age is 32.06 ± 0.10 Ma; paleomagnetic polarity is reverse (McIntosh and others, 1991). Correlation based on lithology, stratigraphic position and three $^{40}Ar/^{39}Ar$ sanidine ages ranging from 31.84–31.95 Ma (Table 3, no. 31, 32 and 33). Base not exposed. Maximum *exposed* thickness estimated from cross sections is 2740 m. Hells Mesa Tuff is one of several older regional ignimbrites assigned to the Datil Group.

Thr Grayish red, phenocryst-rich, spherulitic rhyolite lava, texturally and mineralogically equivalent to "magma" clasts in Thf. Abundant coarse-grained phenocrysts of quartz, sanidine and plagioclase (latter altered to clay). Yields $^{40}Ar/^{39}Ar$ age of 31.88 ± 0.11 Ma from sanidine (Table 3, no. 31). Small isolated exposures 3 km SW of Red Hill mine represent paleotopographic high of lava about 60–90 m thick (Sheet 2, G-G´). This late-stage lava probably represents a resurgence in magma pressure and was presumably associated with moderate doming shortly after the main stage of ash-flow eruption and caldera collapse.

Thf Coignimbrite autoclastic, lag-fall breccias, thin ignimbrites, and pyroclastic *fall* deposits in uppermost part of caldera facies Hells Mesa Tuff. Displays a distinctly bedded appearance derived from resistant ledges of ash-rich tuff (>75% matrix) at the top of numerous individual fall deposits. Partially to densely welded. Coarse lag breccia at base of unit contains large dark red clasts of devitrified (congealed) Hells Mesa "magma" with distinctly euhedral (unabraded) phenocrysts of quartz, sanidine, plagioclase and biotite in a spherulitic matrix. Explosive origin of autoclastic lag breccia is demonstrated by fragmented spherulites in the ignimbrite matrix and microfaults (mm scale offsets) in larger quartz or feldspar phenocrysts in the spherulitic "magma" clasts. Coarsest interval of lag breccias (clasts 5–60 cm) is on ridge in sec. 8, T4S, R1W, which was apparently in a relatively proximal location compared to finer breccias to north and northwest. $^{40}Ar/^{39}Ar$ age of sanidine is 31.95± 0.11 Ma (Table 3, no. 32). Thf is absent on Red Canyon horst block. Maximum thickness is 150 m.

Thu Lithic poor (0–2%), phenocryst-rich, rhyolite ignimbrite in *upper part* of caldera-facies Hells Mesa Tuff. Contains 40–50% phenocrysts of quartz, sanidine, plagioclase, and biotite in a densely welded locally vitroclastic groundmass. Mostly grayish red to dark red, pumice poor, blocky jointed, ignimbrite. Grades downward into lithic-rich mesobreccia unit (Thm), basal contact approximately located. Maximum thickness south of Black Canyon is 460 m; truncated by Miocene erosion surface on Red Canyon horst near Red Hill mine.

Thuf Thf and Thu undifferentiated. Mapped in southeastern Chupadera Mountains near Esperanza mine. Base not exposed here, minimum thickness 240 m.

Thm Mostly lithic-rich (5–50%), reddish orange to reddish brown to grayish red, densely welded, phenocryst rich, quartz-rich, rhyolite ignimbrite *mesobreccia*. Angular to subrounded accidental lithic fragments, mostly 1–50 cm across, consist mostly of dark gray to purplish-gray Spears-type andesite porphyry (containing medium-grained plagioclase phenocrysts), with minor amounts of granite and schist of Precambrian affinity, plus limestones and sandstones of Late Paleozoic affinity. Mesobreccias in lower caldera-facies tuff are intimately interleaved with large coignimbrite landslide blocks of andesite and adjacent zones of very lithic-rich ignimbrite (Thb). Thm is interpreted as a, finer grained, mesobreccia facies laterally equivalent to proximal caldera collapse breccias (Thb) apparently derived from the southern wall of the Socorro caldera. As mapped, Thm includes a 100 m thick, lithic-poor zone in its lower part, south of the Torres mine. Maximum estimated thickness of Thm plus Thb in area south of Red Canyon is 2740 m.

Thb Caldera-collapse *megabreccia* and very lithic-rich *mesobreccias*. Dominated by large slabs of Spears-type andesite porphyry as much as 1 km long and 100 m thick. Rare "dikelets" of quartz-rich Hells Mesa ignimbrite injected into fractures in the andesite slabs locally help distinguish them from bonafide Spears Formation andesites (Eocene) that would represent caldera floor rocks. Thb includes very lithic-rich mesobreccias (>50% clasts) that contain abundant blocks of andesite, granite, schist and limestone typically 0.5 to 3 m across. These two major lithologies probably represent large toreva blocks and rock avalanche deposits, respectively. Wedges of Thb thicken to the south, indicating that they were most likely derived by slumping of the south wall of the Socorro caldera, during early stages of eruption of the Hells Mesa Tuff.

Thw Caldera *wall* facies. Poorly to moderately welded, light-gray to pale red, locally lithic-rich to lithic-poor, phenocryst-rich (20–30%), quartz rich, rhyolite ignimbrite. Locally contains abundant andesite and granitic lithic fragments. Also contains xenocrysts of microcline derived from Precambrian rocks. Lower half is bleached and "argillized" (milky sanidines) in area near Chupadera Spring. Previously assigned to lower Luis Lopez Formation (Tzt1: Eggleston, 1982; Osburn and Chapin, 1983); here reassigned to Hells Mesa Tuff on basis of a $^{40}Ar/^{39}Ar$ age from sanidine of 31.84 ± 0.13 Ma (Table 3, no. 33). Interpreted as medial to upper Hells Mesa Tuff (equivalent to Thm) filling a shallow embayment in the southern topographic wall of the Socorro caldera. Maximum thickness near Chupadera Spring is 270 m. About 1 km south of the quadrangle, Thw wedges out against Precambrian rocks that locally define the inner topographic wall of the caldera (Eggleston, 1982).

PRE-CALDERA ROCKS

Yp Porphyritic schists (middle Proterozoic) — light to medium gray, quartz-feldspar porphyritic schist and porphyry. Includes minor mafic schists and amphibolite (Kent, 1982).

Yf Feldspathic schists (middle Proterozoic) — reddish orange quartzo-feldspathic schists, locally grades into minor pelitic schists (Kent, 1982). Foliated granite, also in this basement block, yields U-Pb age from zircon of 1659 ± 3 Ma (Bowring and others, 1983).

Yu Metamorphic basement rocks undivided (middle Proterozoic) — shown in cross section only.

INTRUSIVE ROCKS

Tim Mafic dikes (Oligocene to Miocene?) — Medium gray to purplish gray, aphanitic mafic to intermediate dike about 3–10 m wide. Locally cuts upper rhyolite member of Luis Lopez Formation (Tzbr1) in area southeast of Esperanza Mine. Also includes a small, grayish red, andesite porphyry dike in large rhyolite plug (Tirj) south of Black Canyon. Aphanitic dike near Esperanza Mine is texturally dissimilar from diabasic flows in Popotosa Formation and is unlikely to be related to the Miocene flows.

Tirs Coarsely porphyritic rhyolite dikes and plug-like intrusion (Miocene) — Light pinkish gray, moderately phenocryst rich (15–20%) coarsely porphyritic rhyolite. Contains large phenocrysts (3–7 mm) of sanidine, with smaller plagioclase, quartz and biotite. Cuts caldera-facies Hells Mesa Tuff in area near Red Canyon. Larger exposures at west mouth of Red Canyon may represent top of cupola or small stock. Plagioclase is locally altered to clays adjacent to the Chupadera fault, but main part of dike contains fresh plagioclase. $^{40}Ar/^{39}Ar$ age of 11.03 ± 0.12 Ma (Table 3, no. 36) agrees well with published K-Ar age of 11.6 ± 0.6 Ma (Willard, 1971). This $^{40}Ar/^{39}Ar$ age is analytically equivalent to the lower rhyolite flow of Pound Ranch member of the Socorro Peak rhyolite of Osburn and Chapin, 1983 (H. Newell, written commun., 1996)

Tirx Moderately phenocryst rich rhyolite dike (Oligocene) — Pale red, moderately phenocryst rich (15–20%), massive rhyolite dike in northeast-striking fracture north of Black Canyon (part of composite dike adjacent to Tia). Contains medium to coarse-grained phenocrysts of sanidine, quartz, plagioclase (clayey) and biotite. $^{40}Ar/^{39}Ar$ age of 28.34 ± 0.11 Ma from sanidine (Table 3, no. 35) indicates intrusion was contemporaneous with magmatic resurgence in the Sawmill Canyon caldera, just west of this locality.

Tirj Phenocryst-poor rhyolite dikes and plugs (Oligocene) — Light gray to light purplish gray, phenocryst poor, flow banded to massive, east-northeast striking rhyolite dikes and small to large plugs in area south of Black Canyon. Width of dikes ranges from 5 to 60 m. Rhyolite plugs near Torres Mine and Black Canyon are 300 to 600 m across and locally merge with northeast trending dikes. Plug at Torres mine includes minor purplish gray dacitic phase along eastern margin. Dikes and plugs typically contain sparse (3–8%) fine grained (1–3 mm) phenocrysts of plagioclase, sanidine, quartz and biotite. Plagioclase is commonly altered to white clay and generally weathers out leaving tabular holes in rock face. Mineralogy, phenocryst size, and abundance are similar to upper rhyolite member of Luis Lopez Formation (Tzc) and the La Jencia Tuff (Tj). $^{40}Ar/^{39}Ar$ age of sanidine in dike at east range front (sec. 8, T1S, R4W) is 28.50 ± 0.11 Ma (Table 3, no. 34), which is analytically equivalent to upper Luis Lopez rhyolites and the La Jencia Tuff. Equivalent ages and lithologies suggest that these dikes were feeding Tzc flows and/or eruptions of La Jencia Tuff.

Tiaz Andesite dikes (Oligocene) — Dark gray to purplish gray, aphanitic to moderately porphyritic, northeast striking, andesite dikes (5–20 m wide) cutting Hells Mesa Tuff and lower Luis Lopez members in area north of Black Canyon. These potassium metsomatized dikes contain sparse phenocrysts of plagioclase altered to white clay and pyroxene altered to Fe oxide. Dike approximately 2 km northeast of Tower Mine appears to feed Tza3 flow. Dike 3 km east-northeast of Tower Mine appears to feed Tza2 flow.

Table 3. Summary of $^{40}Ar/^{39}Ar$ ages of volcanic rocks in the Luis Lopez 7.5 minute quadrangle. Data from W. C. McIntosh, New Mexico Geochronology Laboratory, New Mexico Tech. Except for Map No. 18, all sanidine ages represent single crystal laser fusion analyses that tend to be slightly younger (0.1-0.2 Ma at ~30 Ma; McIntosh and Chamberlin, 1994) than bulk furnace analyses published by McIntosh and others, 1991. "N" equals number of crystals analyzed; letters, eg. "C-H," indicate number of heating steps in furnace analysis used to determine plateau age (C-H=6 steps). Preferred age based on relative age relationship to other dated units and average ages of correlative dated units in addition to analytical error at approximately ±2σ. Geochemical correlationof No. 3A from Dunbar and others, 1996.

Map No.	Unit	Field No.	Lat. (N)	Long. (W)	Mineral or material dated	N: number of analyses	Age (Ma)	Error ±2σ	Comment	Preferred age (Ma)
1	Qtsf(a)	LLZ-1 (NM 1072)	33°55′47.5"	106°54′08"	sanidine + microcline?	3	1.22	0.06	3 youngest analyses approx. depositional age	1.2
2	QTsf(r)	LLZ-5	33°58′02.5"	106°54′26"	sanidine	25	1.57	0.07	max. depositional age	1.6
3	QTsf(r)	LLZ-27	33°59′12"	106°54′26"	sanidine	23	1.67	0.06	max depositional age	1.6
3A	QTsf(a)	LLZ-26	33°59′09.5	106°54′21"	geochemical correlative of No. 3				depositional age	1.6
4	Tbt	NM-1531	33°53′55"	106°56′19"	groundmass	C-H	8.36	0.05	eruption age	8.4
5	Tbt	NM-1530	33°54′11′	106°55′40"	groundmass	A-I	8.39	0.16	eruption age	8.4
6	Tbt	LLZ-47	33°53′16"	106°59′25.5"	groundmass	A-F	8.41	0.12	eruption age	8.4
7	Tbt	LLZ-33	33°59′42"	106°58′21"	groundmass	B-I	8.68	0.22	eruption age	8.4
8	Tbt	LLZ-45	33°55′20"	106°59′28.5"	groundmass	A-E	8.70	0.20	eruption age	8.4
9	Tbt	LLZ-40	33°54′49.5"	106°56′35"	groundmass	A-H	10.1	0.60	hydrothermally altered; superceded by no. 4	8.4
10	Tbt	LLZ-99-10	33°56′27.5	106°59′54.5"	groundmass		in progress		eruption age	
11	Tbo	LLZ-46	33°53′30.5"	106°59′27"	groundmass	B-H	9.64	0.12	eruption age	9.6
12	Tbo	LLZ-99-8	33°52′48.5"	106°56′42"	groundmass		in progress			
13	Tsd	HN-95-16B	33°59′07"	106°59′53"	biotite	G-J	12.23	0.04	biotite often 10-20% older than sanidine in same rock	~9.5
14	Tsc	LLZ-99-1	33°54′32"	106°57′11"	sanidine		in progress			
15	Tlu	LLZ-99-2	33°54′35.5"	106°57′13"	sanidine		in progress			
16	Tlu	KMET-93-56	33°58′49"	106°59′34.5"	sanidine	15	27.38	0.48	lg. range, K metasomatized	27.9
17	Tj	LLZ-43	33°54′25.5	106°57′29"	sanidine	10	28.75	0.05	eruption age	28.7
18	Tzbr2	NM-449	33°54′56"	106°57′23.5"	sanidine	Bulk	28.61	0.17	eruption age	28.7

Map No.	Unit	Field No.	Lat. (N)	Long. (W)	Mineral or material dated	N: number of analyses	Age (Ma)	Error ±2σ	Comment	Preferred age (Ma)
						sample				
19	Tzbr1	LLZ-37	33°56′11"	106°57′44"	sanidine (+glass?)	10	27.4	0.8	glassy, contaminated with ^{40}Ar?	28.7
20	Tzbr1	LLZ-99-5F	33°56′11"	106°57′39.5"	sanidine		in progress		felsic equivalent of no. 19	
21	Tzbr1	LLZ-99-4	33°54′27"	106°57′46.5"	sanidine		in progress			
22	Tzas	LLZ-99-6	33°56′17"	106°57′50"	groundmass		in progress			
23	Tzt1	LLZT-1	33°59′05.5"	106°58′48"	sanidine	?	29.85	0.31	K metasomatized	30.1
24	Tzt1	LLZ-34	33°59′02.5"	106°57′42"	sanidine	12	30.1	0.06	supercedes no. 23, eruption age	30.1
25	Tzt	LLZ-44	33°54′24.5"	106°58′18.5"	sanidine	7	30.06	0.14	5 Plag=28.79±0.3 sanidine xenocryst=31.77±0.08	30.1
26	Tza	LLZ-99-7	33°56′16"	106°57′55.5"	groundmass		in progress			
27	Tzb	NM-1533	33°53′38.5"	106°58′23"	groundmass	total gas	28.32	0.98	underlies Tzt, >30.1	~30.2
28	Tzb	NM-1532	33°54′03.5"	106°58′16.5"	groundmass	A-I	29.2	0.47	underlies Tzt, >30.1	~30.2
29	Tzs1	NM-1534A	33°53′49"	106°58′17"	sanidine	15	33.74	0.10	max. depositional age, precaldera rhyolite clast in moat sediment	33.7
30	Tzs1	NM-1534B	33°53′49"	106°58′17"	sanidine	13	33.53	0.14	max. depositional age	33.6
31	Thr	LLZ-41	33°55′41"	106°57′49"	sanidine	15	31.88	0.11	eruption age	31.9
32	Thf	KMET-93-58	33°59′07"	106°58′52"	sanidine	12	31.95	0.11	eruption age	31.9
33	Thw	LLZ-36	33°52′34.5"	106°56′54.5"	sanidine	14	31.84	0.13	eruption age	31.9
Silicic Intrusions										
34	Tirj	NM-1337	33°58′37.5	106°57′50"	sanidine	6	28.5	0.11	intrusion age	28.7
35	Tirx	LLZ-9	33°59′13"	106°59′00"	sanidine	12	28.35	0.08	intrustion age	28.3
36	Tirs	LLZ-35	33°56′50"	106°59′43"	sanidine	15	11.03	0.12	intrusion age	11.1

TABLE 4—Site-mean paleomagnetic data for volcanic rocks in the Luis Lopez 7.5 minute quadrangle (from McIntosh and others, 1991). Explanation: Second through fourth columns show age constraints provided by stratigraphically adjacent $^{40}Ar/^{39}Ar$-dated ignimbrites (abbreviations explained below); demag is alternating field demagnetization level for site-mean data; n/t denotes number of samples used in site mean/total samples; inc, dec, and int are site-mean inclination, declination, and intensity; k is Fisher's (1953) precision parameter; α_{95} is radius of cone of 95% confidence, type denotes demagnetization behavior: 1=TRM, 2=TRM+IRM; Letter R following type number identifies sites with poor precision (α_{95}>15°).

Ignimbrite abbreviations: LJ=La Jencia, HM=Hells Mesa.

Unit Symbol	constrained age (Ma)	bracketing ignimbrites under	over	site	lat (°N)	long (°W)	dip	strike	demag (mT)	n/t	inc	dec	intensity (A/m)	k	$0/\alpha_{95}$	type
Tzbr1	32.1–28.9	HM	LJ	446	33.907	106.965	8	0	40	4/4	-67	102	0.11	186	7	1
Tzbr1	32.1–28.9	HM	LJ	447	33.909	106.964	8	0	30	4/4	-74	127	0.19	62	12	1
Tzbr2	32.1–28.9	HM	LJ	449	33.916	106.956	0	0	40	4/4	-30	165	0.18	11	29	2 R
Tzbr1	32.1–28.9	HM	LJ	450	33.931	106.956	8	0	30	4/4	-57	108	0.04	188	7	1

TABLE 5—Representative geochronologic data for late Miocene silicic lavas and dikes in the Pound Ranch - Red Canyon - Socorro Peak area. Unpublished $^{40}Ar/^{39}Ar$ data from H. Newell (1997) and the New Mexico Geochronology Research Laboratory (W. C. McIntosh and M. Heizler). K-Ar age of Sedillo Hill lava from C. E. Chapin, unpublished data. Units listed from oldest to youngest, as based on sanidine age data and/or local stratigraphic relationships. Biotite ages are commonly, but not always, about 10% older than equivalent sanidine ages (H. Newell, written commun., 1996). Analytical error at 2 σ.

Age (Ma)	Material dated	Unit/location
11.34 ± 0.14	sanidine	lower Pound Ranch rhyolite lava dome
11.03 ± 0.24	sanidine	rhyolite porphyry dike at Red Canyon (Tirs)
12.04 ± 0.10	biotite	upper Pound Ranch dacite lava
12.23 ± 0.08	biotite	dacite lava at Tower Mine (Tsd)
11.91 ± 0.38	biotite	lower dacite lava at Socorro Peak
10.15 ± 1.70	plagioclase	upper dacite lava at Socorro Peak
10.36 ± 0.16	biotite	rhyodacite lava dome at Strawberry Peak
8.61 ± 0.04	sanidine	rhyolite lava dome at Signal Flag
7.85 ± 0.08	sanidine	rhyolite lava dome at Grefco Mine
7.49 ± 0.12	sanidine	rhyolite lava dome at Jejenes Hill
7.02 ± 0.02	sanidine	rhyolite lava dome at Tripod Peak
7.1 ± 0.4	whole rock (K-Ar)	andesite lava at Sedillo Hill

Appendix I: Petrographic data

This appendix summarizes petrographic data for selected, variably altered volcanic and volcaniclastic sedimentary rocks collected in the northern and central Chupadera Mountains. Fourteen thin sections of altered rocks have been analyzed by the senior author using a Zeiss petrographic microscope. Mineral identification in thin section is based on the following optical characteristics: color, opacity, form, cleavage, relief, birefringence, extinction, twinning, orientation, interference figure and occurrence. Diagnostic optical characteristics of most primary and "secondary" (non-magmatic) rock forming minerals can be found in Kerr, 1959, or Deer, Howie and Zussman, 1966.

Cryptocrystalline clay minerals are generally identified here as high birefringent clays (hbc.) or low to moderate birefringent clays (lbc.). The former (hbc.) is commonly interpreted as illite, since it is often associated with coarser-grained microcrystalline sericite (hydromuscovite) of relatively high birefringence. The latter (lbc.) probably represent non-potassic clays such as smectite or kaolinte. Kaolinite tends to be yellowish and semiopaque in comparison to clear smectite (Kerr, 1959).

The distinctive petrographic signature of Miocene K-metasomatism near Socorro is patchy replacement of albite twinned plagioclase by clear untwinned *optically continuous* adularia that displays low birefringence, low $2V_x$ and negative optic sign (Osburn, 1978; Chamberlin, 1980). Sanidine and biotite in K-metasomatized rocks appear to be essentially unaltered. More recent x-ray diffraction analyses of metasomatic adularia have been interpreted as mixtures of adularia plus minor quartz (Chapin and Lindley, 1983; Ennis, 1996). Potassium-rich plutonic orthoclase (Or 90–100) has a relatively low $2V_x$ of 33–40° in comparison to more sodium rich plutonic orthoclase ($2V_x$=41–103°; Deer, Howie, and Zussman, 1966, p. 308). Potassium-rich metasomatic adularia, as observed here, has an estimated $2V_x$ of 5–15°, based on a few observations of centered interference figures.

Strongly K-metasomatized crystal-rich ignimbrites commonly contain fresh looking sanidine and biotite; apparently these potassic phases were *essentially* in equilibrium with metasomatizing fluids (Ennis, 1996). Sanidine is optically similar to adularia; sanidine also exhibits low birefringence, low $2V_x$ and negative optic sign. In intensely metasomatized rocks adularia is commonly distinguished by its pseudomorphism after lath-shaped plagioclase and irregularly distributed small inclusions of hbc., sericite or other clay minerals.

Lower to medial levels of the intracaldera Hells Mesa (Thrm, lower Thu) are characterized by milky white to pinkish sanidine phenocrysts. In thin section the cloudy sanidines contain sparse to abundant small (.002–.02 mm long) rod-like inclusions preferentially aligned along crystallographic planes. The inclusions most likely represent microcrystalline albite formed by perthitic unmixing during slow cooling of the thick intracaldera tuff.

In strongly altered mafic to intermediate volcanic rocks, ferromagnesian minerals such as olivene, pyroxene, and hornblende are often replaced by mixtures of hematite + clays ± leucoxene ± chlorite. Where replacement is complete (Intensity of alteration, Ia=1.0) the original Fe Mg mineral may not always be discernable from its crystal form. Totally altered mafic minerals of uncertain primary mineralogy are simply identified here as Fe-Mg minerals. Likewise opaque fine-grained primary Fe-Ti oxides, such as magnetite and ilmenite, are not easily discriminated in reflected light. However, alteration products such as red semitranslucent hematite and white opaque leucoxene are readily discriminated with reflected light.

Petrographic analysis of secondary minerals formed by hydrothermal alteration can provide important control on millimeter scale cross cutting relationships and textural relationships that indicate relative age of secondary minerals. Easily recognized incomplete replacement relationships may reflect locally decreased permeability in the rock unit. Also, multiple ages of replacement may reveal changes in chemistry or temperature of the hydrothermal solutions in a dynamic geothermal area (eg. Browne, 1992). Petrographic identification of primary and secondary minerals is made here at varying levels of confidence; less confident identifications are presented here as choices (either/or) of optically similar mineral phases, or as "best guesses" identified with a query(?).

The following petrographic analyses are listed in order of decreasing depositional age or emplacement age of the parent rock. Analyses are not represented as complete in all respects; for instance, minor primary resistant minerals (eg. zircon) may be overlooked; also some more exotic secondary minerals (eg. chabazite and pumpellyite) may not be recognized. These limited petrographic analyses serve to characterize and discriminate age relationships of the different alteration mineral suites and to help "define" alteration zones that occur in the northern Chupadera Mountains. Additional detailed sampling, petrography, mineral identification, and chemical analyses are needed to *accurately* discriminate and map alteration zones, mineral suites, and ages of alteration in the study area.

Petrographic data are listed in the following format. Each sample is given a sequential reference number: Data categories are:

A. Field number

B. Latitude and longitude of sample location

C. Map unit symbol (sheet 1) and general location

D. General description of primary rock type

E. Primary major minerals (in some cases inferred from pseudomorphic character, or from unaltered equivalents sampled elsewhere)

F. Secondary or tertiary minerals replacing primary phases (primary in italics): full arrowheads (→) indicate complete replacement, half arrowheads (⇀) indicate partial replacement; listed in order of decreasing age, ie. *primary* → secondary → tertiary

G. Secondary minerals filling fractures, vesicles, or pore spaces (cross cutting relationships with other secondary minerals indicated by word "cuts").

H. Assigned zone, or zones

I. Comments

No. 1 (reference number)

A. SP–375

B. 33°58′16.5"N, 106°59′12.5"W

C. Thm, about 1 km SSE of Tower Mine

D. densely welded, lithic-rich, crystal-rich, quartz-rich, low-silica rhyolite ignimbrite

E. quartz, sanidine, plagioclase, biotite, Fe-Ti oxides, FeMg (hornblende? and clinopyroxene?), and minor deuteric(?) hematite; with xenocrysts of microcline and muscovite

F. *biotite* → hematite + sericite + quartz(?), *plagioclase* → hbc.(illite) + sericite + quartz (?), *sanidine* → stringlets and rods of microperthitic albite (albite → lbc.?) *FeMg* → hematite + hbc. + sericite, *Fe-Ti oxides* → hematite + leucoxene

G. no void fillings

H. "red" sericitic zone; lower perthite zone

I. cloudy sericitic plagioclase crystals show albite twins and no indication of earlier replacement by adularia (ie. "erasing" of twins; c.f. Eggleston and others, 1983); all sanidines are cloudy and look microperthitic; greater sericitic alteration of a few andesitic lithic-fragments may represent pre-Hells Mesa hydrothermal alteration; some smaller andesitic lithics have red oxidized rims.

No. 2

A. RC-KM-54

B. 33°56′47.5"N, 106°57′24" W

C. upper Thm about 0.8 km SSE of Red Hill manganese mine

D. densely welded, moderately lithic-rich, crystal rich, quartz rich, low silica, rhyolite ignimbrite

E. quartz, sanidine, plagioclase, biotite, Fe-Ti oxides, FeMg, with xenocrysts of microcline, muscovite and orthoclase(?)

F. *Plagioclase* → calcite + lbc., *sanidine* → microperthitic albite

G. no void fillings

H. weak calcitization zone, probably associated with manganese oxide veins to north; biotite is fresh; upper part of perthitic zone

I. densely welded vitroclastic texture partially preserved, plagioclase is mostly fresh, only a few crystals are partly replaced by calcite and lbc. (Ia=0.2); perthitic unmixing of sanidine is moderately developed (Ia=0.6).

No. 3

A. RC-KM-11

B. 33°59′12.5"N, 106°57′52"W

C. Thu approximately 2.8 km ENE of Tower Mine

D. densely welded, lithic-free, crystal rich, quartz rich, low-silica rhyolite ignimbrite.

E. quartz, sanidine, plagioclase, biotite, Fe-Ti oxides, Fe-Mg, (hornblende?) and minor deuteric hematite.

F. *plagioclase* → adularia + hbc. (illite?); *FeMg (hornblende?)* → sericite + leucoxene + hematite, *sanidine* → cryptoperthite, *Fe-Ti oxides* → hematite

G. no void fillings

H. "lower" K-metasomatism zone, high intensity replacement by adularia; uppermost fringe of perthitic zone

I. soft clays were mostly *washed* out of "boxes" (outlined by adularia) during thin section preparation; most sanidines show slight development of perthite (Ia=0.2) and some sanidine is totally clear, biotite is fresh. Intensity of K-metasomatism is highest possible (Ia=1.0)

No. 4

A. 78-3-5

B. 33°58′34"N, 106°57′47"N

C. basal Thf, about 2.8 km ESE of Tower Mine

D. spherulitic devitrified "magma" clast in densely welded crystal-rich, quartz rich ignimbrite of equivalent mineralogy

E. quartz, sanidine, plagioclase, biotite, Fe-Ti oxides, hornblende, clinopyroxene(?) and minor deuteric hematite in spherulites

F. *plagioclase* → adularia + hbc., *sanidine* → cryptoperthite

G. no void fillings

H. lower non-silicified K-metasomatism zone; above main perthitic zone

I. optically continuous metasomatic adularia forms boxwork (chamber walls) filled with hbc.; incipient cryptoperthitic texture in sanidine most likely formed during devitrification under high vapor pressure (i.e. represents pre-eruptive deuteric alteration), biotite is fresh; microfaults and microbrecciated mosaics of quartz and sanidine indicate explosive origin of "magma" clast.

No. 5

A. SP-377

B. 33°58′45.5”N, 106°57′54”W

C. upper Thf, about 2.6 km E of Tower Mine

D. densely welded, ash-rich, pyroclastic fall deposit and underlying crystal-rich ignimbrite, phenocryst content ranges from 15 to 45% respectively

E. quartz, sanidine, plagioclase, biotite, Fe-Ti oxides, hornblende, clinopyroxene(?) and hematite

F. *plagioclase* → hbc. + lbc. (illite + kaolinite?), trace *FeMg* → sericite + leucoxene + hematite

G. no void fillings

H. permeability controlled fringe of lower non-silicified K-metasomatism zone, no metasomatic adularia present; above perthitic zone

I. plagioclase is slightly altered to patches of clay minerals (Ia=.1), sanidine is unaltered

No. 6

A. SP-368

B. 33°59′16.5”N, 106°58′58.5”W

C. medial Tza1 about 1.1 km ENE of Tower Mine

D. fine grained, non vesicular, trachytic, andesite or basaltic andesite lava

E. plagioclase, hypersthene(?), augite(?), and Fe-Ti oxides

F. *plagioclase* → (adularia + hbc.) → calcite, *hypersthene(?)* → (sericite + leucoxene +

hematite) → calcite, *augite(?)* → (sericite + leucoxene + hematite) → chlorite + epidote + calcite + lbc.(?), *Fe-Ti oxide* → hematite + tr. leucoxene

G. no void fillings

H. "lower" K-metasomatism zone with superimposed high temperature calcitization zone; contemporaneous calcite and epidote formed at >260°C (Browne, 1992).

I. Intensity of late-stage calcite alteration is strong within original FeMg phenocrysts (Ia=.8); replacement of earlier metasomatic adularia by calcite is relatively minor (Ia=.1)

No. 7

A. SP-369

B. 33°59′17.5”N, 106°58′56.5”W

C. Medial Tzt1 about 1.2 km ENE of Tower Mine

D. poorly welded, pumiceous, lithic-rich, phenocryst-poor, rhyolite ignimbrite with 5–10% small variably altered andesitic lithic fragments

E. sparse plagioclase, quartz, sanidine, Fe-Ti, oxides and traces of biotite with xenocrysts of sericitic plagioclase and altered FeMg

F. *plagioclase* → adularia + lbc. (semectite or kaolinite?), *pumice (glass)* → sericite + quartz + lbc. or zeolite, *Fe-Ti* oxides → leucoxene + hematite, xenocrystic *Fe-Mg* → chlorite (?) + lbc. + leucoxene + hematite, xenocrystic *plagioclase* → sericite + lbc.

G. no void fillings

H. "lower" K-metasomatism zone, high intensity

I. Phenocrystic plagioclase completely replaced by boxwork adularia filled with lbc. (smectite or kaolinite?); clay commonly washed out of boxes; some andesitic lithics show all plagioclase phenocrysts replaced by adularia, other andesitic lithics contain only sericitic plagioclase and show no indication of secondary adularia. Since phenocrystic plagioclase is altered to adularia; it seems likely that sericitic andesite clasts were altered prior to deposition at 30.1 Ma (Table 3) and, adularia bearing andesite lithics were altered in situ, after 30.1 Ma.

No. 8

A. SP-370

B. 33°59′05.5"N, 106°58′45"W

C. lower Tza2

D. slightly porphyritic andesite lava (~5% phenocrysts)

E. Sparse plagioclase, larger Fe-Mg possibly hypersthene(?) smaller FeMg possibly augite(?), and fine Fe-Ti oxides

F. *plagioclase* → lbc. + hbc. (kaolinite? + illite?), *hypersthene(?)* → (tremolite? + leucoxene + hematite) → quartz + lbc.(?), *augite* → quartz + lbc.(?) *Fe-Ti oxides* → hematite + leucoxene

G. microcrystalline quartz fills narrow veinlets

H. strong silicification superimposed on earlier illitic zone, which is possibly an extension of "red" sericitic zone

I. generally within zone of strong K-metasomatism, but illitic plagioclase is not apparently not replaced by adularia, probably because it predates K-metasomatism; as a secondary potassic mineral illite is not affected by K-metasomatic fluids

No. 9

A. SP-371

B. 33°58′27″N, 106°59′9.5″W

C. wide ENE striking Tirj dike about 0.9 km SE of Tower Mine

D. moderately phenocryst-rich (6–8%), medium to fine grained (1–4 mm), high silica (?), rhyolite dike

E. moderately abundant phenocrysts of quartz and plagioclase, with minor biotite and sanidine, plus traces of FeMg (hornblende?, augite?)

F. *plagioclase* → adularia + sericite + "coarse" smectite (?), *biotite* → sericite + leucoxene + hematite, tr. *FeMg (hornblende?)* → sericite + lbc. (smectite?) + leucoxene + hematite, tr. *FeMg (augite?)* → hematite + titanite + "coarse" smectite (?) + sericite; groundmass (*quartz* + *feldspar* + *hematite*) → biotite or siderophyllite?

G. no void fillings

H. May represent high-temperature potassic alteration of late Oligocene age as suggested by presence of titanite and hydrothermal biotite(?) replacing groundmass. Although adularia replacing plagioclase is present, this is not typical late Miocene K-metasomatism.

I. Plagioclase is almost entirely replaced by adularia + sericite + smectite(?), only a trace of plagioclase remains (Ia=0.99). Alteration of FeMg phases are complete (Ia=1.0). Microcrystalline groundmass with *finely disseminated hematite* is locally "consumed" in wisps and blebs (Ia=0.01) to form clear birefringent fine-grained biotite or siderophyllite?.

No. 10

A. RC-KM-43

B. 33°57′59.5”N, 106°59′02”W

C. narrow east-striking Tirj dike about 1.7 km SSE of Tower Mine

D. phenocryst poor (2–3%), fine to medium grained (1–4 mm), high-silica (?) rhyolite dike

E. sparse plagioclase, quartz, sanidine, with trace of biotite, Fe-Ti oxides, and FeMg (hornblende?)

F. *plagioclase* → lbc. (smectite or kaolinite) + hbc. (illite) + sericite + quartz?, FeMg (augite or biotite?) → titanite + leucoxene + hematite + lbc. (smectite or halloysite?), *sanidine* → sericite, *Fe-Ti oxide* → hematite + leucoxene, *groundmass* (*quartz* + feldspar + hematite) → biotite or siderophyllite?

G. no void fillings

H. high temperature potassic alteration with hydrothermal biotite

I. Plagioclase alteration is very intense (Ia=1.0). Large euhedral biotite phenocrysts show small flamelike *secondary overgrowths* into groundmass at their margins; small wispy inclusion-rich biotites (siderophyllite?) in groundmass are probably of hydrothermal origin. One FeMg (augite or biotite?) altered to titanite and lbc. may also include some hydrothermal biotite. Sanidine is moderately replaced by sericite along cleavages and in patches (Ia=0.3); sanidine is not perthitic.

No. 11

A. RC-KMET-115 A

B. 33°57′35.5"N, 106°59′39"W

C. Tpfm, potassium metasomatized "fanglomerate"

D. moderately well sorted, fine-to-medium-grained, well indurated (jasperoidal cement), rhyolitic sandstone derived mostly from Thuf, upper Thm, and Tirj.

E. sand grains are comprised by quartz (40%), variably altered and devitrified rhyolitic rock fragments (30%) and mixed variably altered feldspars plus minor mostly altered FeMg grains (30%). Detrital feldspars include: clear unaltered sanidine plus slightly to strongly perthitic sanidine (~10%) and minor (2–4%) sericitic feldspars (plagioclase?) and/or sericitic rock fragments. About 3 percent of rock consists of sand-sized "holes" formed where soft clay-rich grains were washed out during thin section preparation. Rare microcline and andesitic rock fragments were presumably recycled from lithic-rich upper Thm. Fresh biotite is rare. All the above variably altered detrital grains presumably represent pre-middle Miocene alteration events in the source area (Red Canyon horst block; sheet 2, G-G'). The following replacement reactions and void fillings represent post-depositional (in situ) alteration associated with late Miocene K-metasomatism.

F. previously unaltered detrital *plagioclase* → adularia + lbc. (smectite or kaolinite)

G. intergranular void space (~2%) filled with red hematitic cryptocrystalline silica (jasperoidal silica)

H. "upper" K-metasomatism zone, which is typically associated with jasperoidal silica filling void spaces

I. previously unaltered plagioclase grains are completely replaced by adularia and lbc. (smectite or kaolinite; Ia=1.0) and form about ½ to 1% of rock; they are typically expressed under crossed nicols as light-gray thin shells or boxlike grains with empty cores. Low birefringent clays (lbc.) are almost always "washed out" of cores or boxworks of adularia grains.

No. 12

A. SP-374

B. 33°58′59.5”N, 106°59′47.5”W

C. medial Tsd about 500 m NW of Tower Mine

D. moderately phenocryst-rich (25%), medium-grained (1–4 mm), slightly vesicular (vuggy) dacite porphyry lava

E. spongy (resorbed) plagioclase, biotite, hornblende, and Fe-Ti oxides

F. *plagioclase* → hbc. (illite?) + clear lbc. (smectite?) + yellowish lbc. (kaolinite?), *hornblende* → hbc. (illite?) + lbc. (smectite?) + hematite + leucoxene, *biotite* → hematite

G. colliform radial chalcedony lines walls of irregular vesicles; thin films (0.1–0.5 mm) of a slightly younger dull jet black opaque mineral, most likely manganese oxides, coat the innermost walls of some vesicles

H. incipient or distal (permeability controlled) upper jasperoidal K-metasomatism subzone, presumably associated with a slight increase in bulk potassium content

I. alteration and Mn(?) mineralization must be younger than late Miocene (<ca. 11 Ma); occurs within 50 m of high intensity K-metsomatism (see No. 13). Plagioclase only slightly replaced by potassic clay and other clays (Ia=0.05). Hornblende completely altered (Ia=1). Rims of biotite flakes are partly oxidized (Ia=0.2). Fe-Ti oxide (magnetite?) does not appear to be altered (Ia=0.0)

No. 13

A. RC-KM-55

B. 33°58′58.5”N, 106°59′48.5”

C. upper Tsd about 450 m NW of Tower Mine

D. moderately phenocryst rich (20%) medium grained (1–4 mm), moderately vesicular, fractured, dacite porphyry lava

E. spongy (resorbed) plagioclase, biotite, hornblende and Fe-Ti oxides

F. *plagioclase* → lbc. (smectite?) + hbc. (illite?) + adularia, *hornblende* → hematite + quartz + clay? (washed out), *biotite* → hematite, *Fe-Ti oxides* → hematite

G. irregular vesicles (2% of rock) completely filled with early red jasperoid and later polycrystalline quartz; veinlet filled with early jasperoid and later polycrystalline quartz that becomes coarser toward the center of veinlet; part of jasperoid veinlet locally cuts adularia; small pods of high-relief purple fluorite and well cleaved barite or celestite occur as trace minerals in the quartz vein.

H. intense upper jasperoidal subzone of K-metasomatism zone with superimposed later-stage silicification

I. Most plagioclase is expressed as tabular holes where clays have been washed out. Plagioclase and hornblende are completely replaced (Ia=1.0). Close proximity of adularia and slightly younger jasperoid/quartz veinlets strongly suggests hydrothermal origin for adularia. Hydrothermal alteration must be younger than 11–12 Ma.

No. 14

A. Tb-8

B. 33°58′43.5"N, 107°00′17"W

C. medial Tpbb (basalt of Bear Canyon) about 1.1 km west of Tower Mine (Molino Peak quadrangle; Osburn and others, 1981)

D. fine grained, ophitic, olivene basalt lava

E. plagioclase, interstitial augite, small phenocrystic olivene, and Fe-Ti oxides

F. Olivene → antigorite + hematite

G. no void fillings

H. deuteric (?) alteration zone

I. Olivene partly replaced by antigorite (Ia=.3); completely fresh plagioclase forms 60% of rock. Lack of plagioclase alteration suggests that alteration of olivene is unrelated to regional K-metasomatism observed in nearby underlying rocks. This thin basalt lava is intercalated in Popotosa playa facies and lies 20–30 m stratigraphically above intensely K-metasomatized and silicified Tsd similar to No. 13.

Captions for Figures and Sheets

FIGURE 1—Index map of the Luis Lopez quadrangle and adjacent quadrangles. Small circles delineate lateral extent of ancestral Rio Grande deposits (early Pleistocene) that form a major aquifer along the Socorro Basin; east boundary from Cather, 1996.

FIGURE 2—Tectonic sketch map of the Socorro-Magdalena region. Map shows active rift faults, rift basins, tilted fault block ranges, the Socorro accommodation zone (SAZ) and the Oligocene Socorro-Magdalena caldera complex; all relative to the Luis Lopez quadrangle (LLZQ). Calderas and derivative ignimbrites are: Socorro caldera (SC), source of 31.9 Ma Hells Mesa Tuff; Black Canyon caldera (BCC), hypothetical source of 30.1 Ma lithic-rich tuff member of Luis Lopez Formation; Sawmill Canyon – Magdalena caldera (SMC), source of 28.7 Ma La Jencia Tuff; Hardy Ridge caldera (HRC), source of 28.0 Ma Lemitar Tuff; Mount Withington caldera (MWC), source of 27.4 Ma South Canyon Tuff; and Bear Trap Canyon caldera (BTC), source of 24.3 Ma Turkey Springs Tuff. Council Rock transverse fault zone (CRTZ) separates moderately extended domain on south from slightly extended domain to north. S=Socorro, M=Magdalena. Numerous domino-style early rift faults are not shown here. Modified after Chapin, 1989; and McIntosh, et al, 1991; BTC and HRC from G. R. Osburn (oral commun., 1999).

FIGURE 3—Correlation diagram for map units in the Luis Lopez 7.5 minute quadrangle.

SHEET 1— Geologic map of the Luis Lopez 7.5 ′quadrangle, Socorro County, New Mexico.

SHEET 2—Geologic cross sections of the Luis Lopez 7.5′ quadrangle, Socorro County, New Mexico.

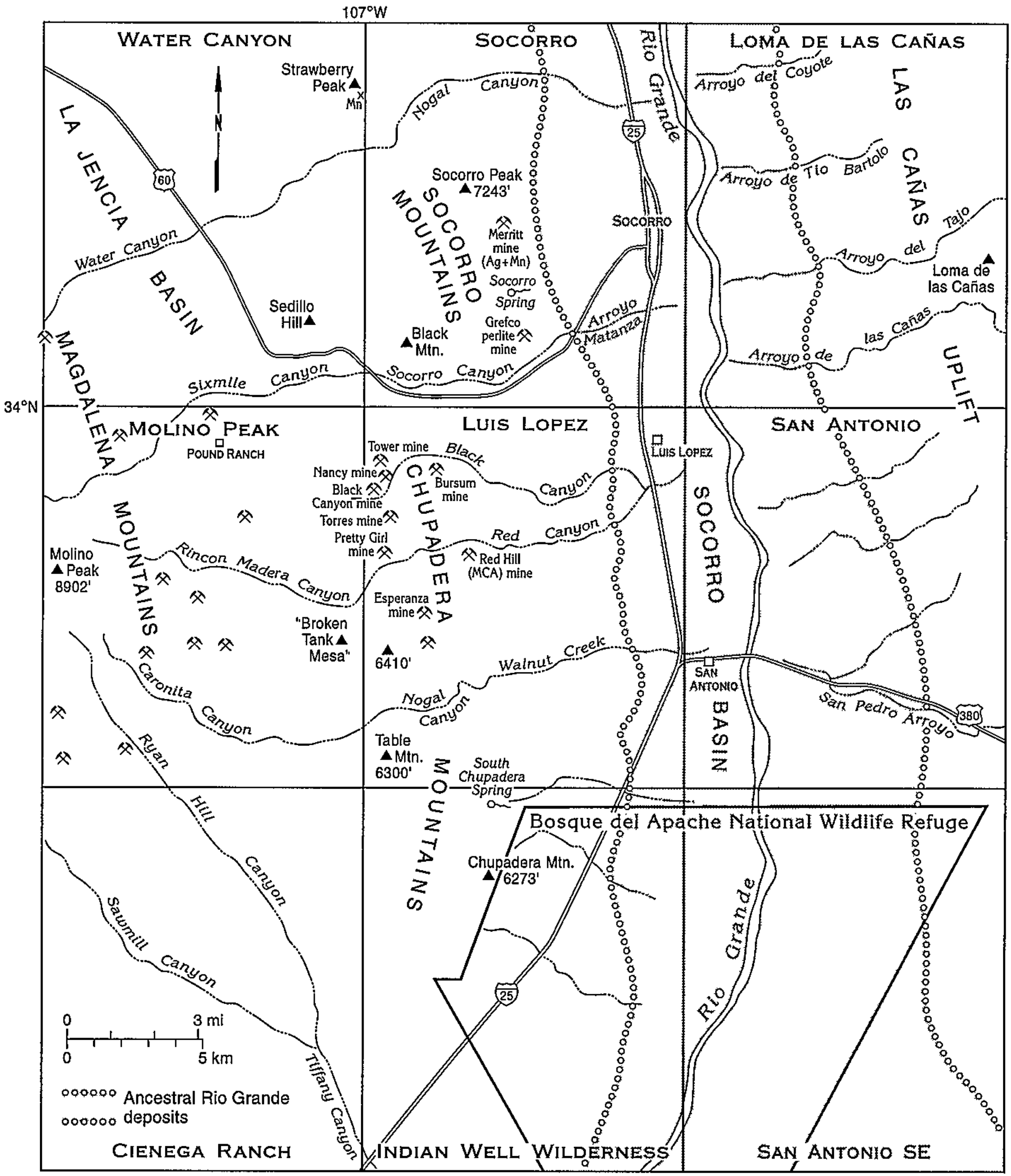


Figure 1. Index map of the Luis Lopez quadrangle and adjacent quadrangles.

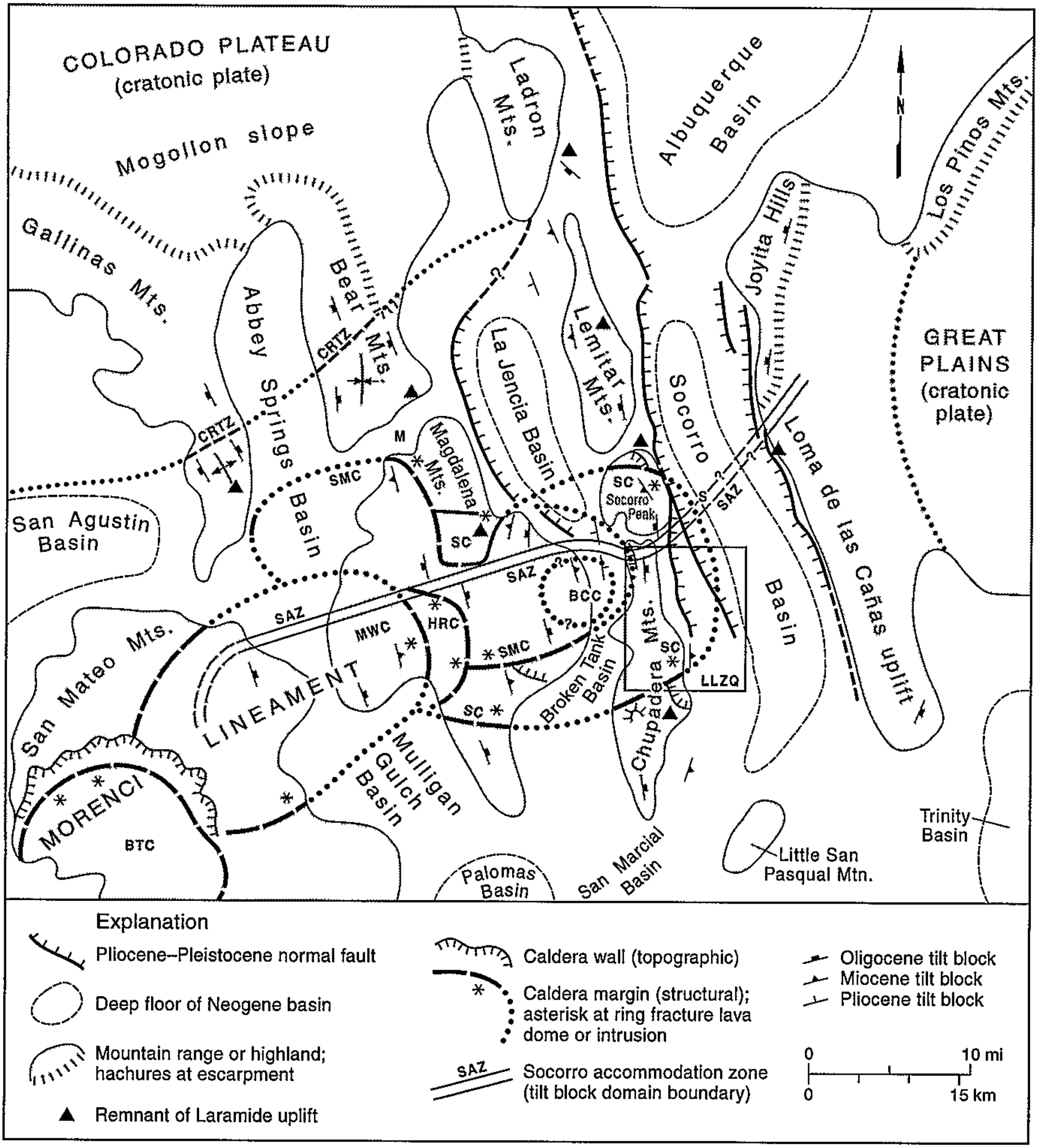


Figure 2. Tectonic sketch map of the Socorro–Magdalena region.

Correlation of Map Units

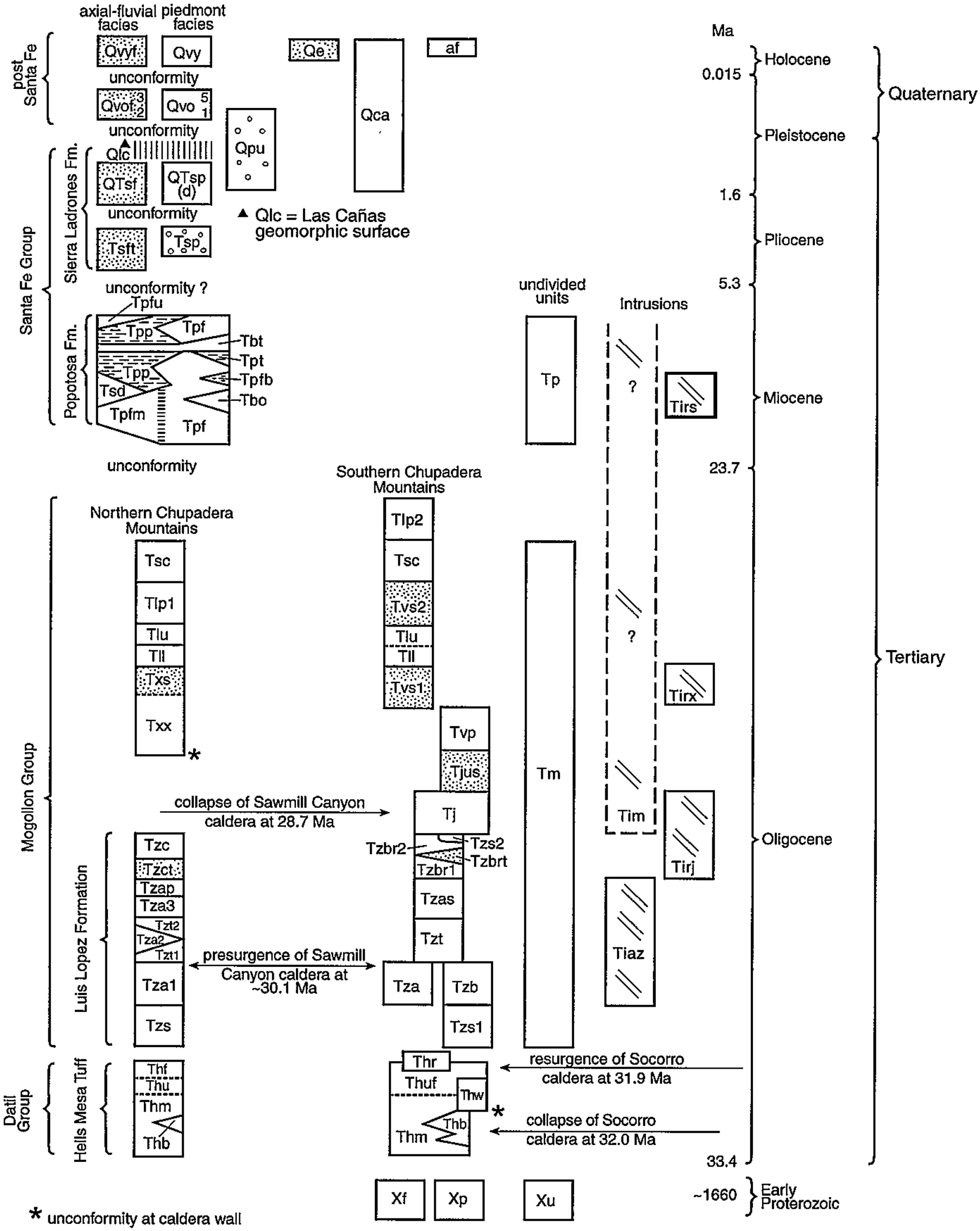


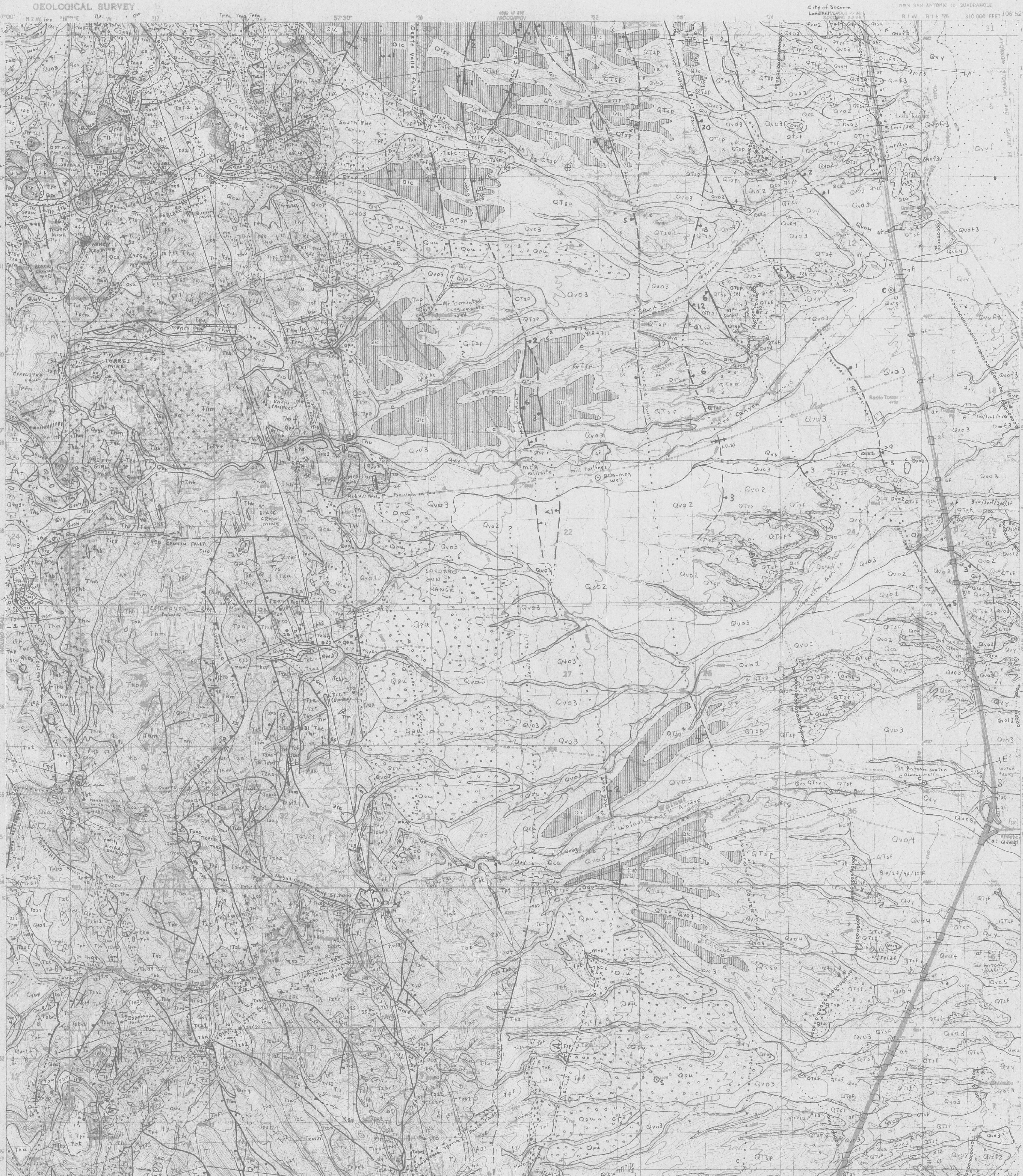
Figure 3. Correlation diagram for map units in the Luis Lopez 7.5 minute quadrangle.

New Mexico Bureau of Mines and Mineral Resources

Open-file report 421 sheet 1

UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

LUIS LOPEZ QUADRANGLE
NEW MEXICO–SOCORRO CO.
7.5 MINUTE SERIES (TOPOGRAPHIC)



Mapped, edited, and published by the Geological Survey

Control by USGS and NOS/NOAA

Topography by photogrammetric methods from aerial photographs taken 1972. Field checked 1975. Map edited 1982

Projection and 10,000-foot grid ticks: New Mexico coordinate system, central zone (transverse Mercator) 1000-meter Universal Transverse Mercator grid, zone 13 1927 North American datum

To place on the predicted North American Datum 1983 move the projection lines 4 meters south and 53 meters east as shown by dashed corner ticks

Fine red dashed lines indicate selected fence lines

SCALE 1:24 000

CONTOUR INTERVAL 20 FEET
DOTTED LINES REPRESENT 10-FOOT CONTOURS
NATIONAL GEODETIC VERTICAL DATUM OF 1929

THIS MAP COMPLIES WITH NATIONAL MAP ACCURACY STANDARDS
FOR SALE BY U.S. GEOLOGICAL SURVEY, DENVER, COLORADO 80225, OR RESTON, VIRGINIA 22092
A FOLDER DESCRIBING TOPOGRAPHIC MAPS AND SYMBOLS IS AVAILABLE ON REQUEST

ROAD CLASSIFICATION

Primary highway, hard surface — Light-duty road, hard or improved surface

Secondary highway, hard surface — Unimproved road

Interstate Route — U.S. Route — State Route

NEW MEXICO

QUADRANGLE LOCATION

LUIS LOPEZ, N. MEX.

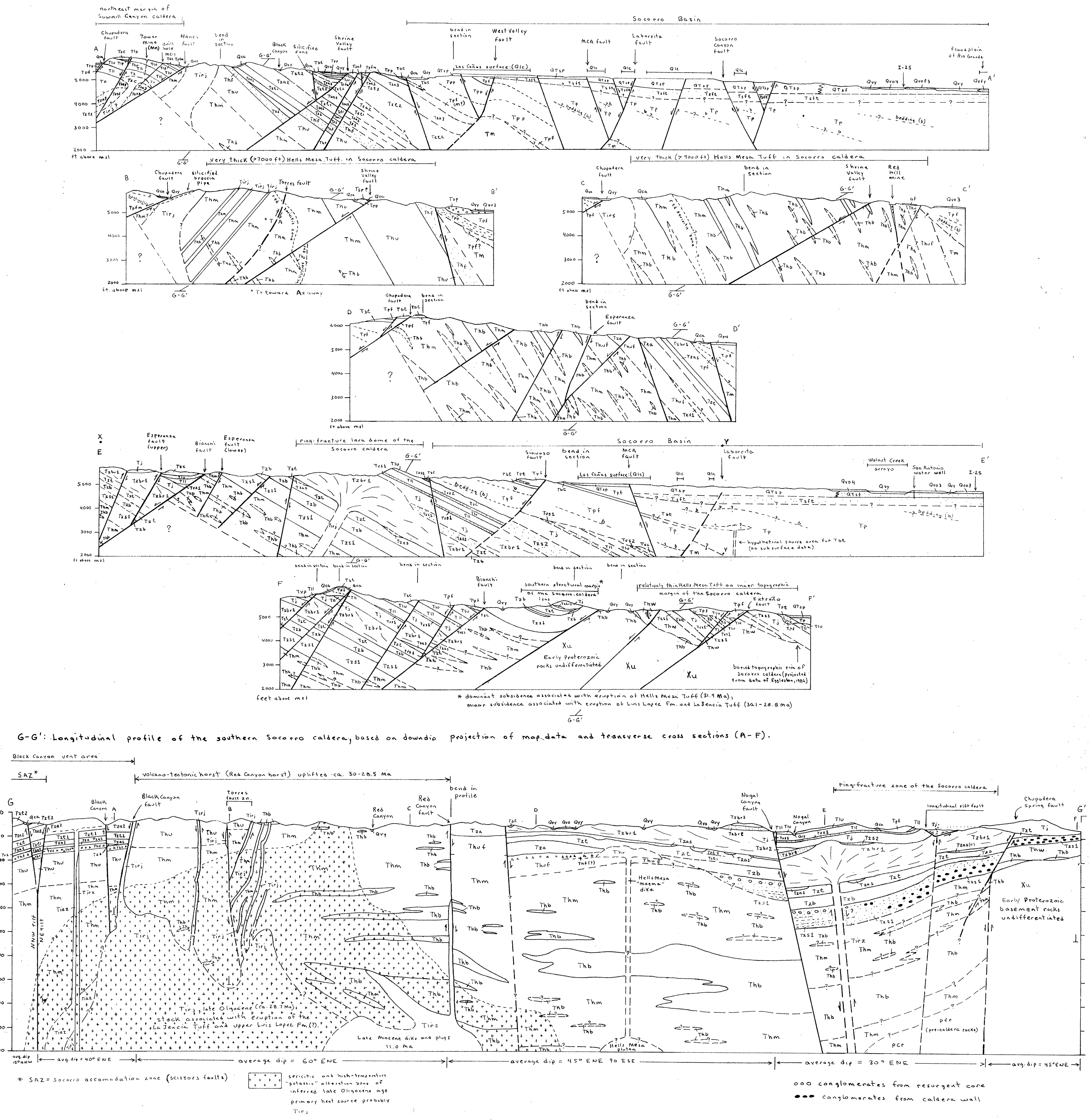
1982

MAPPING RESPONSIBILITIES

Geologic map of the Luis Lopez 7.5 minute quadrangle, Socorro County, New Mexico

by Richard M. Chamberlin and Ted L. Eggleston, 1996

revised 8/4/99



Geologic cross sections of the Luis Lopez 7.5 minute quadrangle, Socorro County, New Mexico

by Richard M. Chamberlin and Ted L. Eggleston, 1996

revised 8/4/99