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CENOZOIC STRATIGRAPHY AND STRUCTURE OF THE SOCORRO PEAK VOLCANIC CENTER, CENTRAL NEW MEXICO

VOLUME I: STRATIGRAPHY

by

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A study of an area where late Cenozoic volcanism, sedimentation, and structure related to the Rio Grande rift have been overprinted on an Oligocene resurgent cauldron.

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ABSTRACT

In central New Mexico, the Rio Grande rift has broken a northeast-trending chain of Oligocene cauldrons, and a surrounding volcanic plateau, into a series of north-trending, tilted fault-block ranges and alluvial basins. The cauldrons lie along the ancient crustal flaw of the Morenci lineament, which has been reactivated within the rift as a deep seated zone of lateral shearing. This transverse shear zone is a diffuse domain boundary at the surface, where it separates fields of tilted fault blocks that are rotated and step-faulted in opposing directions. In cross section, the closely spaced, normal fault blocks look similar to a train of fallen dominoes.

The Socorro Peak volcanic center lies within the rift, at the east end of the cauldron complex. Oligocene volcanic strata exposed here represent the remnants of the north-half of the resurgent Socorro cauldron, tentatively correlated with eruption of the Lemitar Tuff about 28 m.y. (million years) ago. Nearly 900 m of cauldron-facies Lemitar (?) Tuff, which contains caldera-collapse breccias and bedded lag-fall breccias, is exposed on the resurgent block. The Lemitar Tuff outflow sheet covered domino-style fault blocks of the early rift that were largely buried by wedge-shaped prisms of basaltic-andesite lavas. The moat of the Socorro caldera was filled to overflowing by local eruptions of lithic-rich rhyolitic ash-flow tuffs, intermediate lavas, and rhyolite domes; these moat deposits are collectively named the Luis Lopez Formation.

During Miocene time, the northern part of the Socorro cauldron was unconformably buried by the Popotosa Formation, which consists of heterolithic mudflow deposits, fanglomerates, playa mudstones, and minor interbedded basalt flows.

From 12 m.y. to 7 m.y. ago, numerous rhyodacite to high-silica rhyolite domes and tuffs of the Socorro Peak Rhyolite were erupted onto the playa floor of the Popotosa basin as it continued to fill. These silicic domes define a north-northwest-trending intrusive belt, which is widest where it crosses the buried ring fracture zone of the Socorro cauldron at Socorro Peak. Between 7 to 4 m.y. ago, continued rift faulting, combined with epeirogenic uplift, disrupted the floor of the Popotosa basin and began to form the modern ranges and basins of the Socorro Peak area. About 4 m.y. ago, basaltic lavas were erupted from vents southwest of Socorro Peak. These lavas flowed eastward across pedimented fault blocks and onto channel sands of the ancestral Rio Grande. East of Socorro Peak, the facies of the Sierra Ladrones Formation consist of these fluvial sands and intertonguing piedmont gravels shed from the modern highlands.

Eruptive events in the Socorro Peak volcanic center have been dated at 28.6, 11.9-10.3, 10.5-9.0, 7.4, and 4.0 m.y. The primary control of this recurrent magma intrusion and related hydrothermal activity has been the "leaky" vertical fabric of the Morenci lineament. In light of this history, it is not surprising that geophysically defined magma bodies, which provide a heat source for the present geothermal anomaly, are again rising under the Socorro Peak area.

INTRODUCTION

Statement of the Problem

The Socorro Peak volcanic center is located in central New Mexico (fig. 1) where the Rio Grande rift cuts across the northeast flank of the middle Tertiary Datil-Mogollon volcanic field (Chapin, 1971a; Chapin and Seager, 1975). Within this broad zone of intersection, the original volcanic plateau and associated caldera complex have been broken by rifting into a series of north-trending fault-block basins and ranges. The Socorro Mountains, northern Chupadera Mountains and southern tip of the Lemitar Mountains, which together form one of these intra-rift fault-block ranges, form the main part of a large geothermal anomaly outlined by geophysical techniques and constitute the study area of this dissertation.

Interpretations of microearthquake seismograms over a period of years by Professor A. R. Sanford and his graduate-student associates at New Mexico Tech have been used to identify and to map present-day magma bodies intruding the crust near Socorro Peak (Sanford and Long, 1965; Sanford and others, 1973, 1977a, 1977b; Caravella, 1976; Rinehart,

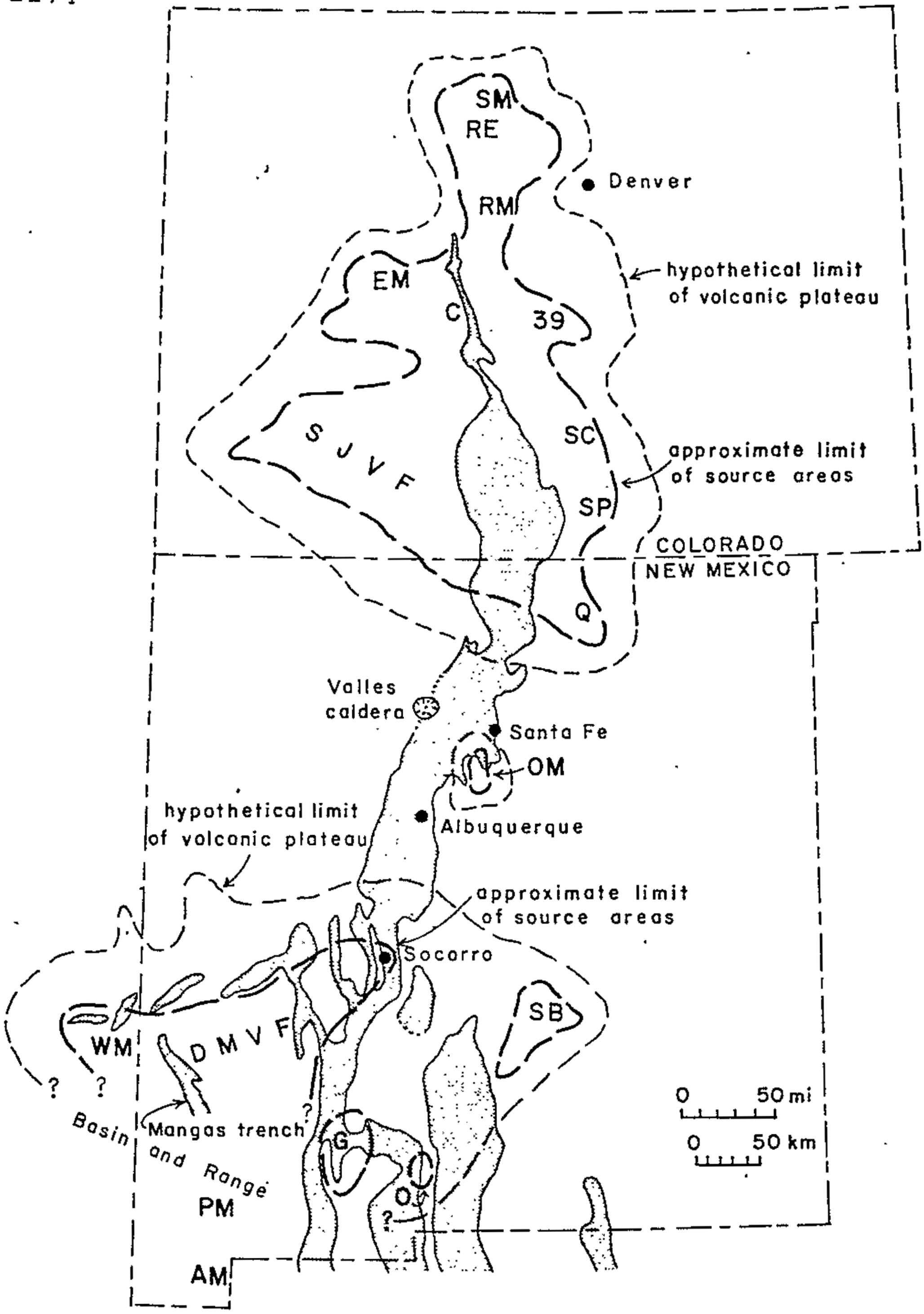

Figure 1. Map of New Mexico and Colorado showing locations of middle Tertiary volcanic fields cut by basins (shaded areas) of the Rio Grande rift (modified after Chapin and others, 1978; Elston, 1978; and Steven, 1975). Major volcanic fields are the Datil-Mogollon (DMVF) and the San Juan (SJVF). From north to south other volcanic centers are: Specimen Mountain (SM), Rabbit Ears (RE), Red Mountain (RM), West Elk Mountains (EM), Collegiate Range (C), Thirtynine Mile (39), Silver Cliff-Rosita Hills (SC), Spanish Peaks (SP), Questa (Q), Ortiz Mountains (OM), Sierra Blanca (SB), White Mountains (WM), Goodsight-Cedar Hills (G), Organ Mountains (O), Pyramid Mountains (PM), and the Animas Mountains (AM).

1976; Shuleski, 1976; Fischer, 1977; Sanford, 1977a; Shuleski and others, 1977). Five shallow (4-5 km) dike-like magma bodies clustered south of Socorro Peak and an underlying sill-like magma body at 20 km depth have been mapped using microearthquake reflections.

Supporting evidence for modern magma bodies in the Socorro area comes from heat-flow measurements as high as 11.7 H.F.U. (Reiter and others, 1975; Reiter and Smith, 1977; Sanford, 1977b) and releveling data indicating modern uplift as great as 0.5 cm/yr over the deep magma body (Reilinger and Oliver, 1976). More tangible evidence of geothermal activity is found in the thermal springs (90°F) which issue from the mountain front 4 km southeast of Socorro Peak. In 1976 the U.S. Geological Survey designated a block of 362 km^2, centered approximately at Socorro Peak, as the Socorro Peak Known Geothermal Resources Area (KGRA).

The objective of this dissertation is to provide a detailed understanding and description of the Cenozoic stratigraphic, structural and volcano-tectonic setting of the Socorro Peak volcanic center. In turn this may be applied as a geologic base for geothermal exploration and development in the Socorro Peak area. The salient aspects of the exploration framework of the volcanic center--namely stratigraphic and structural controls of potential geothermal reservoirs and where they may occur at a favorable depth with respect to the shallow dike-like magma bodies -- have been summarized by Chapin, Chamberlin and others (1978). A more comprehensive evaluation of the geothermal potential of the Socorro Peak area would be beyond the scope of this dissertation and is not attempted here. The exploration framework paper of Chapin and others was based largely on a synthesis of geologic data from this dissertation, geologic data from Osburn (1978) and geophysical data from several sources. The generalized geologic map of the Socorro volcanic center from Chapin and others (1978) is included here as plate 3. The generalized map covers a greater area than the detailed geologic map of this study (pl. 1). Comparison of these two plates should aid the reader in understanding the more fragmented outcrop pattern of the detailed geologic map.

Detailed descriptions of several new stratigraphic units are presented in this dissertation along with descriptions and redefinitions of some previously established units. New data on facies relationships, paleocurrent directions, pebble compositions, thickness variations and unconformities in the Popotosa Formation have been collected. These new data indicate a more complex evolution to the early-rift (Miocene) Popotosa basin, than previously described by Bruning (1973) or by Chapin and Seager (1975).

Particular emphasis has been given during this investigation on collecting data concerning the structural control and timing of past episodes of volcanism. Field relationships and K-Ar dating in the Socorro Peak area indicate two periods of silicic volcanism (29-28 m.y., 12-7 m.y.) and three periods of basaltic eruptions; the youngest of which occurred 4 m.y. ago. Two major structural features, which are related to magma intrusion and volcanism, are now recognized to overlap in the Socorro Peak area. These structures are: 1) the Socorro cauldron (new name), a largely buried, resurgent, ash-flow tuff caldera, which formed about (or prior to) 28 m.y. ago with Socorro Peak now on its northeast margin; and 2) a transverse shear zone related to the east-northeast-trending Morenci lineament. The transverse shear zone separates fields of tilted blocks undergoing extensional faulting and rotation in opposite directions (Chapin, Chamberlin, and others, 1978). The structure of the Socorro area is further characterized by a complex pattern of both low-angle and high-angle longitudinal normal faults of the Rio Grande rift that repeat strongly tilted volcanic strata. The relationships of these longitudinal faults indicate the onset of domino-style rifting between 32 m.y. to 28 m.y. ago (Chamberlin, 1976, 1978).

Most of the structure section of this dissertation is devoted to a discussion of the characteristics and the development of each of these structural features and in turn their control of past volcanism and present magmatism.

A large portion of the Lemitar Mountains was mapped as part of this dissertation investigation. However, this area has not been included with the geologic map of this report (pl. 1) and will not be discussed in detail.

Location and Geography

The study area of this dissertation (pl. 1 and fig. 3) is a roughly rectangular tract about 16 km long from north to south and about 10 km wide from east to west. The area includes the northernmost Chupadera Mountains, all of the Socorro Mountains, and the southern tip of the Lemitar Mountains. Socorro Peak, high-point of the Socorro Mountains, is near the center of the rectangle and approximately 6 km west of the town of Socorro in central New Mexico. The ranges are only part of a long en echelon chain of tilted fault blocks, which form a continuous looking uplift along the west flank of the Rio Grande Valley from the Ladron Mountains to the Coyote Hills, a distance of 70 km. The "uplift" is broken into a series of mountain ranges by several superposed tributaries of the Rio Grande (fig. 2), which cut across these fault blocks in the vicinity of transverse structures.

Throughout the dissertation reference is made to geologic features and their location with respect to geographic landmarks in the Socorro-Magdalena region (fig. 2) or to landmarks within the geologic-map area (fig. 3). These figures are both in the text and in the rear pocket.

Several minor unnamed topographic features in the geologic-map area (fig. 3, pl. 1) have been given informal "names of convenience" to simplify the discussion of the complex geology of the area. Like most place names which have been formalized by common usage, the informal "names of convenience" used here are descriptive of some aspect of the topographic feature, either as viewed in the field or on the topographic map. For example, these informal names are derived from shapes and colors -- Split Peak, Tripod Peak, Big Cliff, Black Hills, Black Mesa; shapes and elevation benchmarks shown on topographic maps -- 6001 Mesa, 6633 Peak; shapes and man-made artifacts -- Radar Peak, Stonewall Dome, Shrine Valley; and shapes and configurations of associated talus deposits -- Pinocchio Peak, Tepee Town Mountain; and in some cases for more subtle reasons. Anyone working in the Socorro-Lemitar Mountains during the summer months will find that biting-gnats (Spanish--jejenes) are a

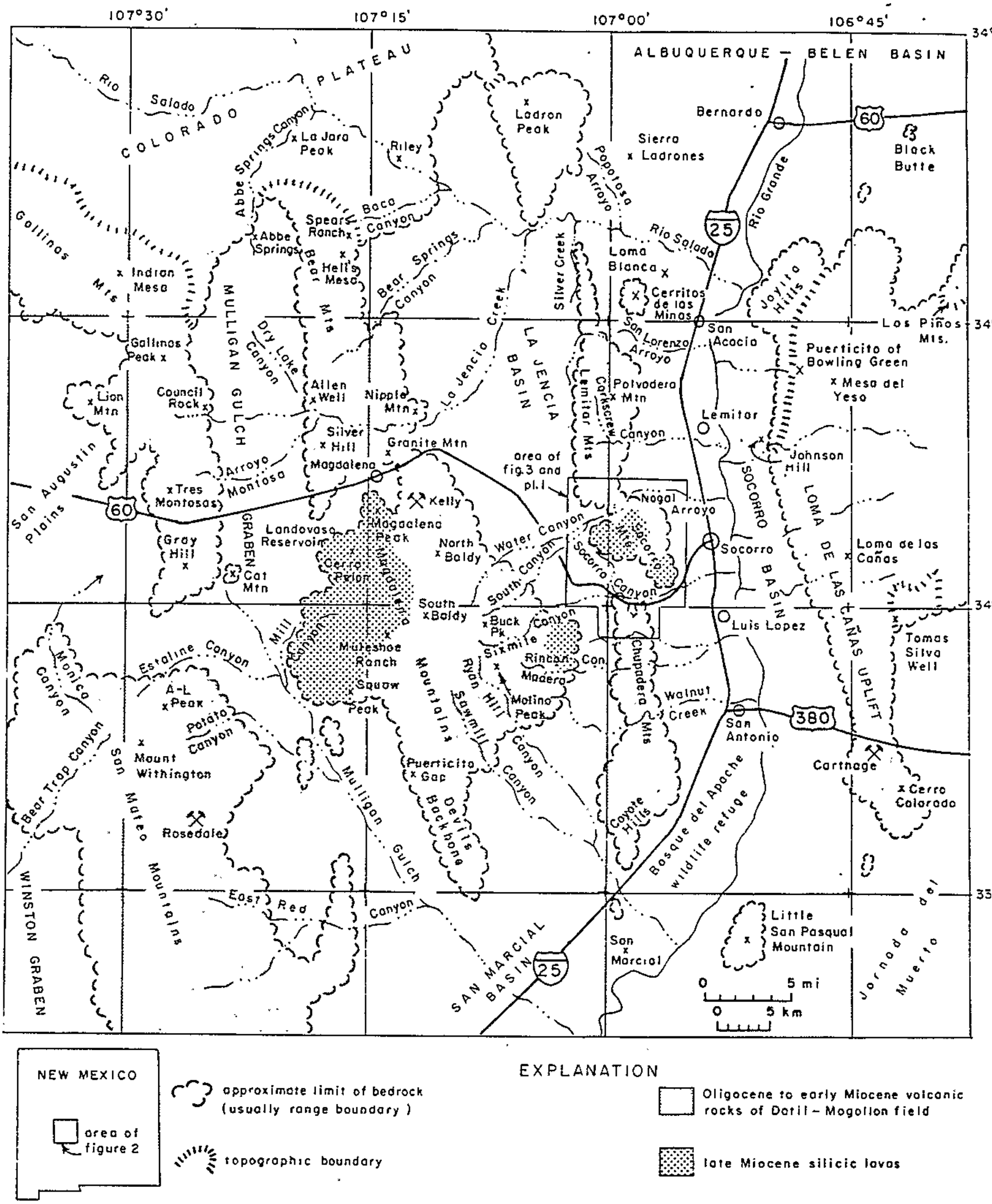

Figure 2. Index map of the Socorro-Magdalena area, central New Mexico, showing the location of geographic features and the distribution of middle Tertiary volcanic rocks of the Datil-Mogollon field. Geography from Socorro and Tularosa 1x2 degree sheets. Geology from Dane and Bachman (1965), and from Machette (1978, unpub. geologic map of the Socorro 1x2 degree sheet).

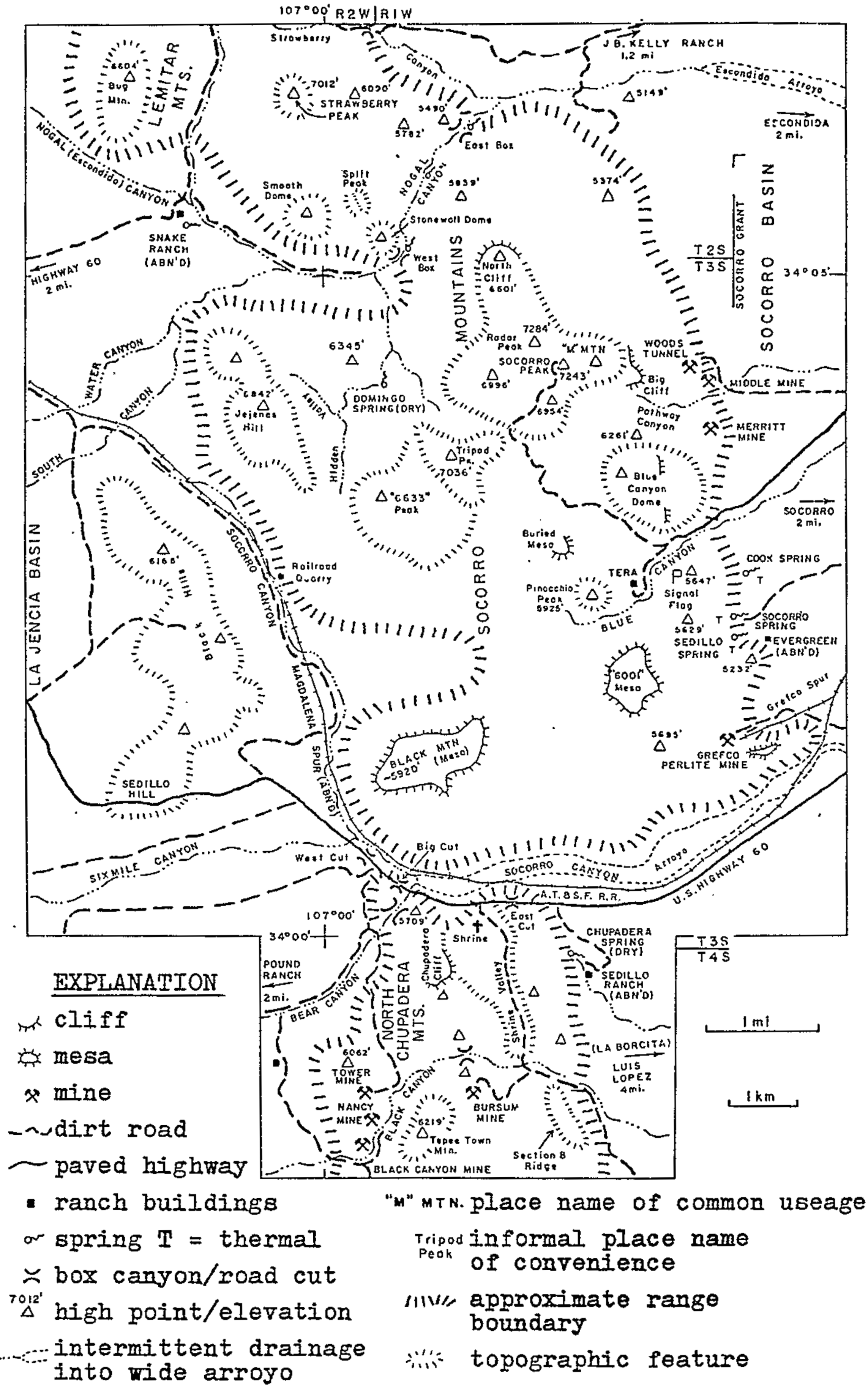

Figure 3. Geographic index map of the Socorro Peak volcanic center (pl. 1).

major nuisance factor. The hills west of Socorro Peak seem to commonly have the greatest concentration of these gnats and therefore, one has been informally named Jejenes Hill.

Access to most of the dissertation area is good via numerous ranch roads, mine roads and roads built by the New Mexico Institute of Mining and Technology for research purposes (fig. 3). Locally, however, these dirt roads require a four-wheel-drive vehicle for transit. U.S. Highway 60 traverses the southern portion of the area and provides several revealing road cuts in the volcanic deposits that filled the Socorro cauldron.

Previous Investigations

Numerous geologic investigations have been conducted in the Socorro region since before the turn of the century. Only a few of the more significant studies, particularly those concerned with the Tertiary volcanic rocks, will be mentioned here in order to provide a sketch of the historical development of geologic knowledge for the region. Many additional reports, not included here, are referenced in the main body of the thesis when pertinent to the topic of discussion.

Geologic study of southwestern New Mexico and southeastern Arizona began with a reconnaissance traverse through the area in 1873 by G. K. Gilbert (1875) as part of the Wheeler Survey. Gilbert considered the area to be "one of the great lava tracts of the world". Most reports of this early period emphasized the geology around mining districts; locally, the Magdalena (Kelly) and Socorro Peak districts attracted the most attention. Silliman (1882) was first to describe the silver chloride -- heavy spar (barite) veins of the Socorro Peak district. Silliman noted that little was known about the area because the "... danger from hostile bands of wandering Apaches, whose murderous raids have slain many adventurous miners..." had only "recently" been abated.

C.L. Herrick (1899) provided the first relatively detailed account of the volcanic geology of the Socorro and Lemitar Mountains. Herrick's geologic sketch map of the Socorro Mountains (pl. IX) and his panoramic view of the eastern face of Socorro Peak (pl. II, Fig. 3) are, with a few exceptions, remarkably accurate. However, his interpretation of the volcanic history relied heavily on the misleading assumption that the present topography is a direct result of past volcanic eruptions.

In his 1904 book on <u>New Mexico Mines and Minerals</u>, F. A. Jones briefly described the volcanic rock types and economic geology of the Socorro Peak silver district (p. 109-

111). Jones also made an astute closing remark concerning the volcanic origins of Socorro Peak,

> "Considering the recent earthquake shocks in this immediate section in connection with the thermal springs at the base of the mountain, these point to the fact that life is not yet extinct in that shattered and rock riven member."

C. H. Gordon (p. 213-285) in Lindgren, Graton, and Gordon (1910) -- Ore Deposits of New Mexico, accurately described the general volcanic sequence in the Socorro region (p. 238) as: early andesite and latite, rhyolite, followed by relatively recent basalt. The author believes it is appropriate to applaud the perceptive and profound observations of C. H. Gordon. It was a startling realization to this writer that Gordon (p. 218) and later E. H. Wells (1918, p. 71) both recognized the association of block faulting and tilting of the Tertiary volcanic strata, and also noticed the change from westerly dipping strata north of Socorro to easterly dips south of Socorro. Neither Gordon nor Wells commented on the significance of their observation. However, Gordon clearly considered it to be related to the "great structural trough" of the Rio Grande Valley, since they are both discussed in the same paragraph.

Winchester (1920, p. 10) was first to publish a measured section (considered by many to be a type section) of the volcanic rocks of the Datil-Mogollon field. Although the section was measured at the north end of the Bear Mountains, Winchester formally named the volcanic and sedimentary rocks the "Datil Formation" because as he stated: they are "...the mountain forming series of the Datil Mountains...". Wilpolt (Wilpolt and others, 1946) later renamed the basal 684.5 feet (694.5 feet according to Wilpolt) of Winchester's Datil Formation as the Baca Formation, because it consists entirely of arkosic non-volcanic detritus. Wilpolt considered these arkosic conglomerates and sandstones in the Bear Mountains to be equivalent to Eocene arkosic strata in the Joyita Hills and the Carthage area, since in all three areas they overlie Cretaceous rocks and underlie a thick section of volcanic rocks.

Loughlin and Koschmann (1942) published the results of a detailed study of the geology and ore deposits of the Magdalena mining district. They subdivided the Tertiary volcanic rocks of the district into ten informal lithologic units that they considered to be different stratigraphic intervals of andesite, latite, latite tuff and rhyolite. Loughlin and Koschmann realized that at least one of their units, the "banded rhyolite" (equivalent to the flow-banded member of the A-L Peak Tuff), is regionally extensive. However, only a thin volcaniclastic sandstone (op.cit., p. 22) exposed near Granite Mountain was considered as a possible correlative of Winchester's Datil Formation.

Beginning in the middle 1950's, the New Mexico Bureau of Mines and Mineral Resources undertook a reconnaissance mapping project of the Datil-Mogollon volcanic field as a cooperative project with the United States Geological Survey in order to provide a base for an updated geologic map of the state of New Mexico. W. H. Tonking (1957) mapped in detail the Puertecito 15' topographic quandrangle, which includes the type area of the "Datil Formation" in the Bear Mountains, at the southeastern margin of the Colorado Plateau (fig. 2). Tonking expanded and subdivided the "Datil Formation" into three members consisting, from oldest to youngest, of the: 1) Spears Member -- latitic tuffs and conglomerates, 2) Hells Mesa Member -- welded rhyolite tuffs, and 3) La Jara Peak Member -- basalt and basaltic-andesite lavas. Willard (1959) and Weber (1963) abandoned Tonking's nomenclature and subdivision of the volcanic pile because they found areas, toward the center of the field, where andesite to latite lavas and sediments were stratigraphically above or interbedded in the rhyolitic tuff section. Some of these "anomalous" areas are now recognized as cauldrons. Willard (1959, p. 96, fig. 5) and other investigators associated with the reconnaissance mapping project therefore interpreted and mapped the "Datil Formation" as intertonguing time transgressive facies with only local lateral continuity. Dane and Bachman (1965) followed the same facies concept for the "Datil Formation" on their compilation of the revised geologic map of New Mexico. As indicated on this map, rhyolitic lavas and domes capping Socorro Peak were recognized as overlying Santa Fe Group sedimentary rocks (Weber, 1963, p. 140) and therefore included as a member of the Santa Fe. Basaltic andesites of the La Jara Peak member, and "early andesites" in the Magdalena Mountains were excluded from the "Datil Formation" on the basis of apparent unconformable relationships (Willard, 1959, p. 94 and p. 98). These exclusions are now known to be inappropriate (see fig. 7 and fig. 8).

During the 1960's, a major revolution occurred in the field of volcanology and volcanic stratigraphy. In the late 50's to early 60's radiometric dating of igneous rocks using the K-Ar method became an economically feasible and generally reliable correlation tool for volcanic rocks. Radiometric ages of volcanic and intrusive rocks pertinent to this study have been published by: Weber and Bassett, 1963; Burke and others, 1963; Weber, 1971; Chapin, 1971b; Willard, 1971; Smith and others, 1976; and Bachman and Mehnert, 1978. The most important element of this revolution was the widespread publication and acceptance of the concept of regionally extensive sheets of ash-flow tuff (Smith, 1960a; and Ross and Smith, 1961) and their genetic relation to cauldron subsidence (Smith and Bailey, 1968).

Much of the present understanding of the Cenozoic volcanic stratigraphy and structure of the Socorro-Magdalena area has been developed over the last decade through synthesis of numerous detailed mapping projects (mostly 1:24000) as part of graduate thesis investigations directed by C. E. Chapin and supported by the New Mexico Bureau of Mines and Mineral Resources. Collectively these past and ongoing geologic studies, which include this dissertation, have come to be known as the "Magdalena Project". A few of these thesis investigations, particularly those of University of New Mexico students (Deal and Spradlin) were not funded as part of the Magdalena Project; however, they were completed as a cooperative effort with C. E. Chapin and the Magdalena Project. As suggested above, the key to this understanding of the volcanic history of the Socorro-Magdalena area is a stratigraphic framework of regionally extensive, petrologically and/or texturally distinctive, ash-flow tuff sheets.

The first thesis of the Magdalena Project by D.M. Brown (1972) described the detailed stratigraphic relationships and petrographic characteristics of six major ash-flow cooling units that he mapped in the southern Bear Mountains north of Magdalena. Brown subdivided and named most of these ash-flow units as members of Tonking's "Hells Mesa member", which had been elevated to formation status by Chapin (1971a). Most of Brown's nomenclature for the ash-flow stratigraphy has been modified or revised in order to give the individual major ash-flow sheets formation status; however, his original internal subdivision of the ash-flow sequence is with one exception essentially unchanged. The only significant change has been that his "tuff of La Jencia Creek" is now recognized as equivalent to -- and in the same stratigraphic position -- as his "tuff of Allen Well" (here renamed the "Lemitar Tuff").

Deal (1973) mapped about 400 square miles of the northern half of the San Mateo Mountains and studied the primary laminar-flow structures in the A-L Peak Tuff, which is probably the most extensive and distinctive of all the ash-flow sheets in the Socorro region. Deal's mapping indicated a great thickness (600m) of the crystal-poor rhyolite ash-flow tuff (A-L Peak Tuff) and an overlying sequence, 1350 m thick, of moderately crystal-rich to crystal-rich rhyolite ash-flow tuff (Potato Canyon Tuff). Deal interpreted this area of anomalously thick tuffs in the northern San Mateo Mountains, centered about Mount Withington, to be a large resurgent cauldron that was the source of both of these ash-flow tuff sheets. Part of Deal's "upper Potato Canyon" is now considered to be correlative with the Lemitar Tuff, which is interpreted here to have been erupted from the Socorro cauldron.

The original stratigraphic framework of ash-flow sheets described by Brown (1972) has been corroborated and expanded through several additional thesis mapping projects. These investigators and the general areas they mapped (see fig. 2) are as follows: Simon (1973), Silver Hill area; Woodward (1973), Lemitar Mountains; Krewedl (1974), north-central Magdalena Mountains; Chamberlin (1974), Council Rock district; Blakestad (1976), Magdalena (Kelly) mining district; Wilkinson (1976), Tres Montosas-Cat Mountain area; Spradlin (1976), Joyita Hills; Osburn (1978), northeastern Magdalena Mountains; Massingill (1978), Riley-Puertecito area; and Petty (1979), southeastern Magdalena Mountains.

Several previous geologic investigations and geologic maps of varying detail overlap with the immediate area of this dissertation. S. G. Lasky (1932, p. 121) published the first geologic map in the area of interest. Lasky's detailed map covers approximately 3 km^2 in the immediate vicinity of the silver mines of the Socorro Peak district. As part of a study of the Luis Lopez manganese district,

A. T. Miesch (1956) mapped about 65 km^2 of the northern Chupadera Mountains extending from U.S. Highway 60 south to Nogal Canyon, west of San Antonio. Reinterpretation of Miesch's map and descriptions of the volcanic rock units have been especially useful in understanding the southern portion of the Socorro cauldron, which has not yet been remapped using the present stratigraphic framework. M. E. Willard had largely completed a detailed geologic map (1:6000) and a manuscript on the manganese mineralization in the Luis Lopez district at the time of his death in 1972. Both are on file at the New Mexico Bureau of Mines and Mineral Resources in Socorro. DeBrine and others (1963, fig. 2) published an incomplete reconnaissance geologic map of the Socorro Valley area. In the same guidebook to the Socorro region, C. T. Smith (1963) published a reconnaissance map of about 13 km^2 centered about Socorro Peak and included detailed measured sections of the Pennsylvanian rocks exposed on the eastern escarpment of the peak. Two detailed geologic maps have been prepared by New Mexico Institute of Mining and Technology students (Lowell, 1967; and Burton, 1971) as "independent studies" under the supervision of C. E. Chapin. Lowell mapped about 25 km^2 extending from north of U.S. 60 to Blue Canyon and west to Black Mesa. Burton mapped about 10 km^2 from Blue Canyon north to

Socorro Peak and compiled a stratigraphic correlation chart to show the relationship of his map units to those of Lasky, Smith, and Lowell.

All of these investigations in the Socorro-Chupadera Mountains area were conducted prior to Brown's establishment of the regional ash-flow stratigraphy and some predated the general recognition of ash-flow tuffs and calderas. It is to these investigators' credit that the rock types they describe and the distribution of these units as shown on the geologic maps are mostly accurate; however, many of their interpretations must now be revised in light of a regional ash-flow tuff stratigraphic framework.

Woodward (1973) prepared a semidetailed geologic map of the northern Lemitar Mountains from Red Mountain to Corkscrew Canyon (approximately 55 km^2). Woodward interpreted major east-trending offsets of Tertiary volcanic strata as paleovalley walls because the projection of these discontinuities did not cut underlying Paleozoic strata. The author's initial mapping in the area south of Corkscrew Canyon indicated major structural complexities but no significant erosional unconformities within the Oligocene volcanic sequence. Therefore, I extended my mapping into Woodward's area to resolve this inconsistency.

Osburn's (1978) detailed mapping of the Water Canyon to Pound Ranch area in the eastern Magdalena Mountains (70 km^2) has provided pertinent data on the nature of the western boundary of the Socorro cauldron. Also, unconformable onlap of Socorro Peak rhyolite lavas (late Miocene) over late Oligocene ash-flow tuffs, mapped by Osburn in the Pound Ranch area, place an important constraint on the geometry of the early-rift Popotosa basin.

Most of the evolution of the Rio Grande rift is recorded in the intermontane basin-fill deposits of the Santa Fe Group. Pioneering studies of these dominantly fanglomerate-playa and fluvial deposits were made by Kirk Bryan (1938) and his students. Recent summaries of age relationships, facies relationships, stratigraphy, and different approaches to subdivision of the rift-basin fill can be found in Hawley and others (1969), and in Galusha and Blick (1971).

In the Socorro area, three investigations have been completed in which the Santa Fe Group was studied in detail and divided into formations. Denny (1940) mapped and described strongly tilted, "post-volcanic" (post-Oligocene) volcanic-rich conglomerates, sandstones, and mudstones in the area south of Ladron Peak along the Rio Salado. Denny named these deposits the Popotosa Formation. Bruning (1973) expanded upon Denny's work and correlated outcrops of volcanic-rich fanglomerates and intertonguing playa mudstones in the Socorro-Lemitar Mountains, northern Magdalena Mountains and northwestern Bear Mountains with the Popotosa Formation. Bruning recognized some difficulty in defining the top of the Popotosa Formation. Machette (1978) has mapped and described Pliocene fluvial sands of the ancestral Rio Grande and intertonguing fanglomerates which lie in angular unconformity on Popotosa strata near San Acacia. Machette has named these beds of the uppermost Santa Fe Group, the Sierra Ladrones Formation.

Published papers, which summarize much of the data from the Magdalena Project are: Chapin and others, 1974, 1975, and 1978; and Chapin and Seager, 1975. Three publications of the New Mexico Geological Society: Guidebook 14, 1963; Special Publication No. 5, 1976; and Special Publication No. 7, 1978, contain a wealth of geologic data concerning the Socorro area and Cenozoic volcanism of the Datil-Mogollon field.

Present Investigation

This dissertation is founded on detailed geologic mapping (1:12000) of approximately 63 mi^2 (163 km^2) of the Lemitar, Socorro and northern Chupadera Mountains (fig. 3, pl. 1). Support data has been collected in the form of measured sections (mostly in the Lemitar Mountains; R. M. Chamberlin, 1978, unpub. map), pebble counts, paleocurrent measurements, paleoflow measurements in volcanic rocks, and numerous thickness estimates of stratigraphic units from structure sections. About 150 specimens selected from measured sections and representative outcrops were thin sectioned for petrographic analysis. About 30 bulk-rock samples were collected for chemical analysis and/or K-Ar age dating.

The geologic map is based on a total of 20 months field work during the summers of 1975, 1976, 1977, and the fall through spring of 1975-76. Geologic contacts observed in the field were plotted directly onto black-and-white aerial photographs of the U.S. Geological Survey, Series GS-VMA (flown 3-7-56, scale 1:40000) and series GS-VCUA (flown 8-13-71, scale 1:29000). Contacts and faults were transferred from aerial photos and compiled on a 1:12000 topographic base map using an opaque projector, correcting for parallax as necessary. The outcrop widths shown on the geologic map of narrow geologic features such as: dikes, colluvial breccias at the base of the Popotosa Formation, and exposures of mudstones of the upper Popotosa Formation in gullies around Socorro Peak, are commonly exaggerated by 2 or 3 times their actual width.

A mylar topographic base map (1:12000) was made by joining and enlarging portions of U.S. Geological Survey topographic maps of the Socorro and Molino Peak 7.5' quadrangles, and the Magdalena and San Antonio 15' quadrangles. The spatial relationships and respective contour intervals of these topographic quadrangle maps are shown at the bottom of plate 1. Outcrop data such as strike and dip measurements were plotted directly on field copies of the topographic base map. The final geologic map (pl. 1) does not include an area of about 30 mi^2 (78 km^2) mapped in the Lemitar Mountains (see index map, pl. 1). Throughout the dissertation where reference is made to this unpublished map of the Lemitar Mountains, it will be called the "Lemitar Map" (see reference list for availability).

Seven complete sections and one partial section of Oligocene volcanic formations exposed in the Lemitar Mountains (Lemitar map) were measured using the brunton and tape method (Kottlowski, 1965, p. 112-118). Only measured sections of the A-L Peak Tuff and the Lemitar Tuff, will be included here since they are important stratigraphic units in the area of the dissertation geologic map (pl. 1). Thickness values determined from these measured sections are considered accurate to ±10 percent. Thicknesses of formations determined by sketch structure sections made from the geologic map were estimated to the nearest 25 feet (10 m) and are considered to be accurate to ±20 percent for units thicker than 100 feet (30 m) and to ±50 percent for thinner units.

The orientations of paleoflow indicators and paleocurrent indicators were determined from outcrops using a brunton compass. Where strata have been tilted more than 15-20°, the original orientation of the observed linear element was found by using a Wulff-type stereonet. To rotate the vectorial element back to its original position, the enclosing stratum was assumed to have been rotated about a horizontal axis parallel to its strike. The original attitude of faults of various ages have also been estimated using a stereonet by applying similar assumptions. The orientation of tension fractures roughly perpendicular to flow direction, ramp structures (flow thrusts) dipping sourceward, and flow banding in the more silicic lavas were all measured and plotted where observed to aid in the interpretation of the location of volcanic vents. The azimuth of pumice lineations formed by primary laminar flow in the flow-banded member of the A-L Peak Tuff and in a thin ash-flow tuff unit in the Luis Lopez Formation were noted at several outcrops as a check on their source areas.

Where observed, the attitude of imbricated pebbles, festoon trough axes, cross bedding, parting lineations and log axes have been measured to estimate paleocurrent directions (Potter and Pettijohn, 1977). These paleocurrent vectors and azimuths, which have been plotted on the geologic map (pl. 1), are considered to be within ±30° of the true paleocurrent direction. Pebble compostions and their relative abundance in Santa Fe conglomerates have been mostly determined by visual estimates, and in a few instances by means of counting clasts of discernable lithology within a square meter on selected exposures.

Petrographic interpretations are based largely on routine fabric description, mineral identification and visual estimates of modal compositions of approximately 150 thin sections of mostly volcanic rocks collected in the study area. Modal analyses of 5 thin sections were made using a Swift automatic point counter with a grid spacing of 1/3 mm by 2/3 mm. Statistically accurate analyses require a minimum of 1000 total points counted for phenocryst-rich rocks and 2000 total points for phenocryst-poor rocks. Modes listed only to the nearest percent concentration are from visual estimates rather than point-counted. About half of the thin sections were stained with sodium cobaltinitrite using the method of Chayes (1952) to aid in the identification of potassium feldspar. These stained sections also proved to be valuable in the study of widespread potassium metasomatism recently recognized in the Socorro-Magdalena area (Chapin and others, 1978).

Most stated compositions of phenocryst minerals are from routine flat-stage thin-section examination using a Zeiss binocular petrographic microscope. Estimates of the composition of plagioclase feldspar, alkali feldspar, and clinopyroxene were made from optical data determined on a four-axis universal stage with standard extinction techniques (Emmons, 1943). Plagioclase compositions (±3 percent anorthite, op cit, p. 133) were determined from maximum extinction angles (X') perpendicular to 010, using the high-temperature curves of the Rittman zone method (Troger, 1959, 0.111). Ambiguities in extinction angles for compositions between An_0 - An_{30} were resolved by determining optic axial angles (2V) (Emmons, 1943, p. 23-41) and checking against 2V versus percent anorthite curves for volcanic plagioclase (Slemmons, 1962, pl.11).

The alkali feldspars, sanidine -- anorthoclase -- orthoclase, were differentiated on the basis of 2V measurements and a common physical trait. Although not entirely diagnostic, many investigators have noted that in unaltered rocks sanidine and anorthoclase are characteristically clear while orthoclase is usually clouded.

Approximate compositions of clinopyroxene phenocrysts were determined from a combination of 2V and extinction angle (Z∧C) measurements (Heinrich, 1965, p. 218). Nomenclature of Ca-Mg-Fe clinopyroxenes used here is after Poldevaart and Hess (1951, p. 472).

Several new whole-rock major-element chemical analyses reported here have been made by D. L. White using X-ray fluorescence techniques. These analyses are tabulated in Appendix 1 of the New Mexico Bureau of Mines and Mineral Resources open file report number 88, available on request at Socorro. These preliminary analyses do not include the quantity of H_20-, H_20+ or CO_2 in these rocks[1] and should therefore not be compared directly with other chemical analyses recalculated to 100 percent on a volatile free basis. Other chemical analyses of volcanic rocks from the Socorro-Magdalena area reported in several prior investigations (Lindgren and others, 1910; Loughlin and Koschmann, 1943; Miesch, 1952; Tonking, 1957; Weber, 1963a; Deal, 1973; and Spradlin, 1976) are included in the text where pertinent.

New K-Ar radiometric dates reported here for volcanic

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[1]MgO values reported are of uncertain acuracy and the largest analytical error is probably with respect to SiO_2 content.

rocks in the Socorro area were determined by Geochron Laboratories Division of Krueger Enterprises, Inc. K-Ar ages of the Grefco perlite and 6001 Mesa rhyodacite dome were determined by A. Leroy Odom at Florida State University. To relate radiometric ages to Cenozoic time units, the Cenozoic time scale as currently accepted by the U.S. Geologic Survey (Bishop and others, 1978, p. 150) has been used here. The entire sequence of major geologic time divisions for the Precambrian and Phanerozoic Eons is shown in figure 4.

The terminology and nomenclature used here is generally in agreement with the American Geological Institute -- Glossary of Geology (Gary and others, 1972). A few terms used here may not be found in this comprehensive glossary. The usage of terms such as "caldera" (topographic volcanic depression) and "cauldron" (volcanic subsidence structure) and features in cauldron environments follows that of Smith and Bailey (1968), and Steven and Lipman (1976). The classification and terminology of faults and other structures is from Dennis (1967). The nomenclature of ash-flow tuffs used in this dissertation, with some modifications, is generally that of R. L. Smith (Smith, 1960a, 1960b; Ross and Smith, 1961). Since Smith's classification scheme provides little indication of total phenocryst volume in the ash-flow tuffs, the following modifiers are used here: very crystal-

Era	Period	Epoch	Boundary age (in m.y.)
Cenozoic	Quaternary	Holocene	
		Pleistocene	1.8
	Tertiary	Pliocene	5.0
		Miocene	26.0 [1]
		Oligocene	37.5
		Eocene	53.5
		Paleocene	65
Mesozoic	Cretaceous		136
	Jurassic		190-195
	Triassic		225
Paleozoic	Permian		280
	Pennsylvanian		320
	Mississippian		345
	Devonian		395
	Silurian		430-440
	Ordovician		500
	Cambrian		570
Precambrian	Precambrian Z		800
	Precambrian Y		1600
	Precambrian X		2500
	Precambrian W		

Figure 4. Chart of major geologic time divisions currently in use by the U.S. Geological Survey (Bishop, Eckel and others, 1978).

[1] The Oligocene-Miocene boundary at 26.0 m.y., from Harland and others (1964), is used here to provide consistency with prior studies in the Socorro region.

poor (0-3 percent phenocrysts), crystal-poor (3-10 percent), moderately crystal-rich (10-25 percent), and crystal-rich (> 25 percent). The lithic content and pumice content of ash-flow tuffs are important parameters to describe but are generally not representative when observed in single hand specimens or thin sections. Also, the apparent abundance of pumice is partly a function of the degree of welding. Loosely defined modifiers for lithic content are: lithic-poor (<2 percent) and lithic-rich (>5 percent). Some extremely lithic-rich tuffs grade into "tuff breccias" with less than 50 percent matrix. Most crystal-rich tuffs are pumice-poor containing less than 2 percent pumice in most exposures. Pumice-rich tuffs commonly consist of 5 to 25 percent pumice lapilli.

The classification and nomenclature of all rock types described here, except for the volcanic rocks, is after Travis (1955). Lipman (1975, p. 5) presents a simple although "somewhat arbitrary" classification system for volcanic rocks of the southeastern San Juan volcanic field with divisions based on percent SiO_2:

Rock name	Percent SiO_2
basalt	<52
andesite	52-60
rhyodacite	60-65
quartz latite	65-70
rhyolite	>70

As Lipman points out, a major advantage of his scheme is that "for San Juan rocks", the divisions coincide approximately with phenocryst variations and combinations visible in the field (fig. 5). Preliminary comparison of chemical analyses and associated modal compositions of Cenozoic volcanic rocks in the Socorro Peak area indicate that Lipman's classification scheme can also be applied here. Within the above classification system based on percent SiO_2, average basaltic-andesites contain 52-56 percent SiO_2, (Chapin and Seager, 1975, p. 308, Fig. B) and high-silica rhyolites contain more than about 76 percent SiO_2 (Byers and others, 1976).

One important introductory note remains to be made. In contrast to the San Juan field, much of the Datil-Mogollon field remains essentially unmapped in terms of a regional ash-flow and cauldron framework. As all geologists are aware, the recognition and interpretation of structure relies heavily on an accurate understanding of the local and regional stratigraphy. Thus, much data on the nature of overprinting of the Rio Grande rift on the Datil-Mogollon volcanic structures in the Socorro-Magdalena area, has only become available recently and is limited in areal extent. The details of regional stratigraphy, facies relationships, and paleobasin topographic development of the Santa Fe Group

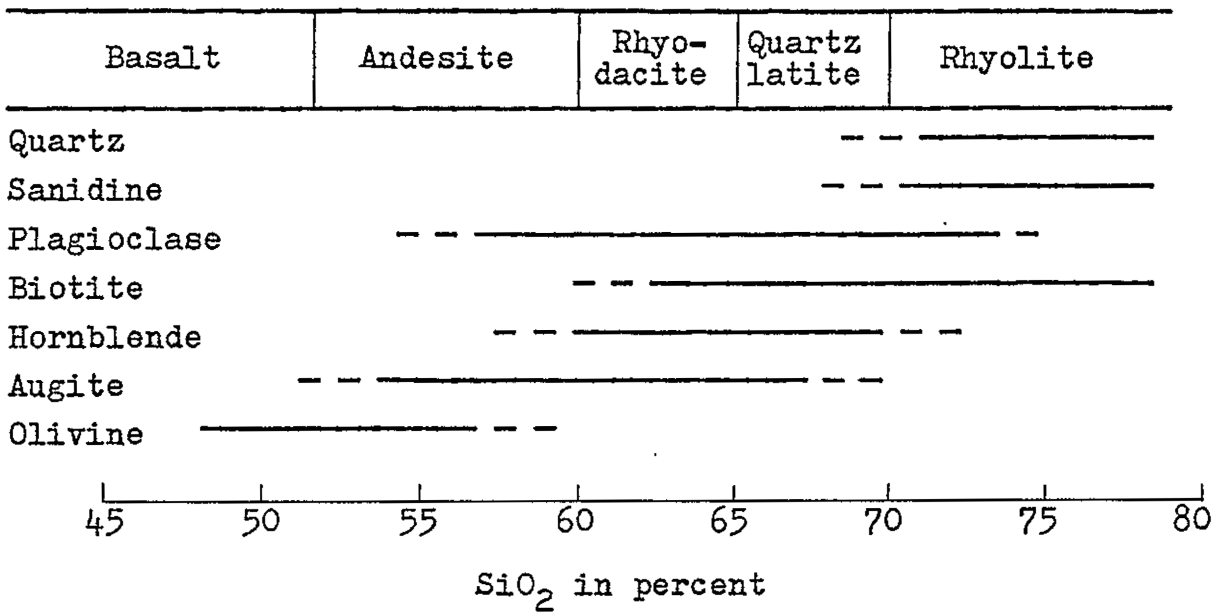

Figure 5. Approximate relationship between phenocrystic minerals and SiO_2 content of volcanic rocks of the southeastern San Juan Mountains (Lipman, 1975). Volcanic rock names used in this report follow this field classification scheme.

basin-fill deposits is even more poorly understood and complicated by the general absence of marker horizons. Consequently, many of the conclusions presented here, which may have regional implications, rely heavily on the interpretation of a limited data base and should be regarded as preliminary.

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PREVOLCANIC ROCKS

Rocks that lie below the Datil-Mogollon volcanic pile range in age from Precambrian to late Eocene. The oldest rocks exposed in the Socorro area consist of Precambrian low-grade metasedimentary and metavolcanic rocks, mostly phyllite and quartzite, intruded by small weakly metamorphosed gabbroic stocks and large unmetamorphosed granite to quartz monzonite plutons (Condie and Budding, 1979). The entire metamorphic-granitic complex is cut by numerous pre-Mississippian (Precambrian?) diabase dikes. Isochrons constructed from Rb-Sr isotopic data on Precambrian plutonic rocks in central New Mexico indicate a late Proterozoic age, 1.5-1.2 b.y. (White, 1978).

In central and northern New Mexico the Precambrian basement complex is unconformably overlain by widespread Mississippian, Pennsylvanian and Permian limestones, shales and sandstones; dominantly marine shelf deposits with some continental red beds in the lower Permian (Kottlowski, 1963). Thin Mississippian strata of the Caloso Formation and the Kelly Limestone, which are present in the southern Lemitar Mountains and northern Magdalena Moutains, are absent on Socorro Peak and east of the Rio Grande (Gordon, 1907a; Armstrong, 1958, 1963).

In the Socorro area, the distribution of late Paleozoic, Mesozoic and early Tertiary (Eocene) sedimentary strata is largely controlled by a major Laramide uplift centered to the southwest of Socorro. For example; in the Lemitar Mountains, the northern Magdalena Mountains, and near Tres Montosas (fig. 2), basal volcanic conglomerates of the Spears Formation rest unconformably on late Paleozoic limestones and sandstones (Wilkinson, 1976; Blakestad, 1976; Krewedl, 1974; Chamberlin, Lemitar Map). In contrast, along the southern margin of the Colorado Plateau (Bear and Gallinas mountains) and east of the Rio Grande (Joyita Hills, Tomas Silva Well, and Cerro Colorado) the Spears Formation lies conformably on late Eocene arkosic sediments of the Baca Formation. The Baca Formation, in turn, overlies a thick section of Cretaceous sandstones and shales, Triassic red beds, and late Paleozoic strata (Tonking, 1957; Spradlin, 1976; Wilpolt and Wanek, 1951).

Thus, the base of the Oligocene volcanic sequence evidently rests on a post-Laramide late Eocene surface of low relief and regional extent. (Epis and Chapin, 1975; Chapin and others, 1978). The change from unconformable to conformable relationships provides the general outline of the

Laramide uplift and the adjacent early Tertiary Baca basin (Snyder, 1970, 1971).

The stratigraphic sequence and general lithology of prevolcanic rocks in the Socorro volcanic center are summarized in the composite stratigraphic column for the area (fig. 6). Within the study area prevolcanic outcrops are restricted to a structurally high block on the east face of Socorro Peak that marks the northeast margin of the Socorro cauldron.

Precambrian Rocks

A small pentagon-shaped outcrop of fine grained metasedimentary rocks is exposed on the steep eastern escarpment of Socorro Peak about 225 m west of the Woods Tunnel portal (pl. 1). The outcrop consists dominantly of dark greenish-gray argillite that is pervasively and intensely jointed in comparison to surrounding sandstones of the Sandia Formation. Where well exposed in a small ravine that bisects the pentagon, the argillite contains thin interbeds of fine-grained quartzite and siltite. The relict bedding dips about 60 degrees to the south. Chlorite and quartz are the dominant minerals visible in hand specimen.

A small dike-like body of fine-grained mafic monzonite(?) is clearly exposed cutting the argillite in the ravine about

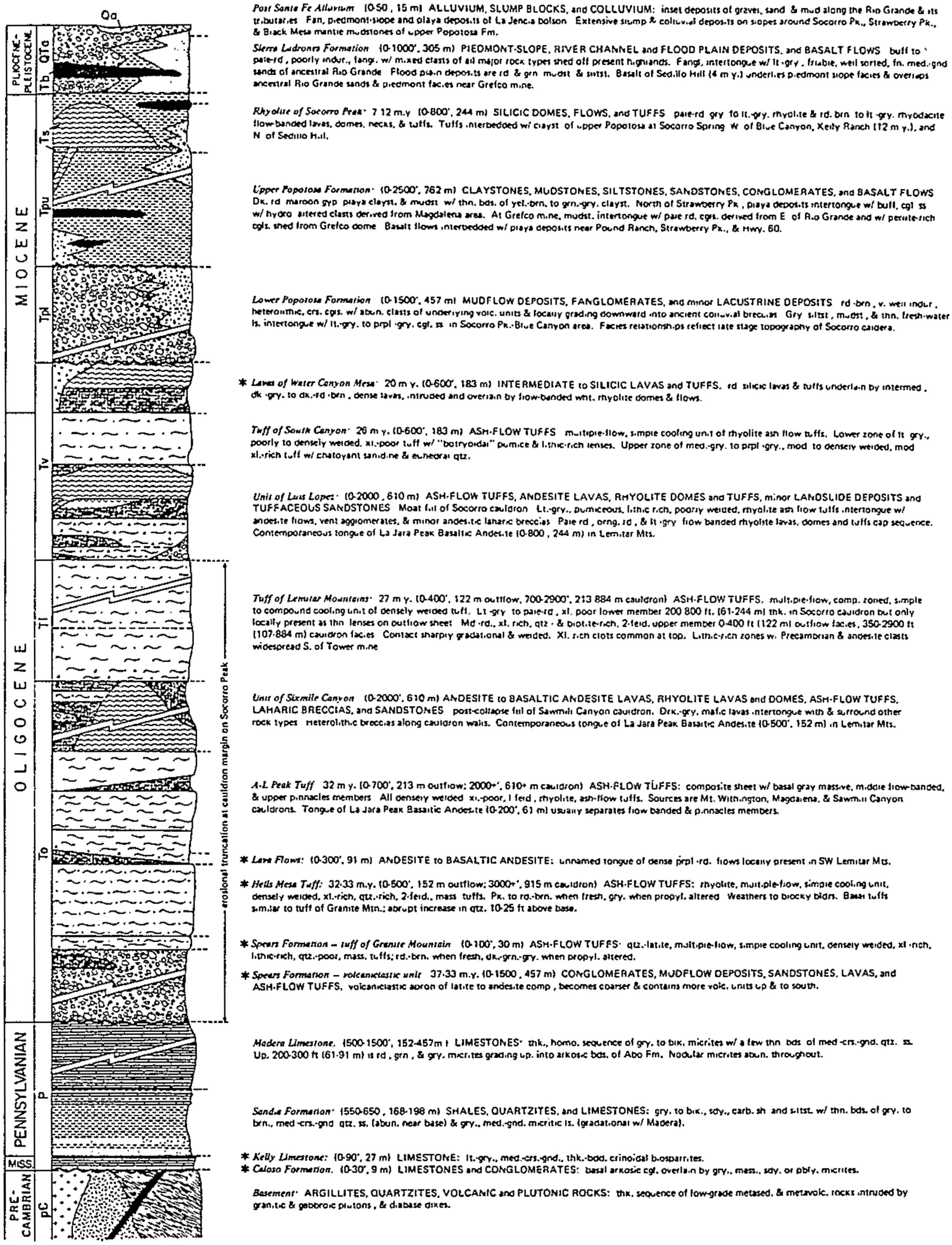


Figure 6. Composite stratigraphic column of the Socorro Peak volcanic center (from Chapin and others, 1978). Asterisks are next to units not exposed in study area but interpreted as locally present in subsurface. Thicknesses not to scale. Map symbols are keyed to Plate 3.

30 m from the west edge of the outcrop. The body is too small to show on the geologic map. This fine-grained dike is most likely a small apophysis of the medium-grained plutonic rock intersected in Woods Tunnel, about 60 m below and 45 m south of the argillite exposure (C. T. Smith, 1963, p. 186).

The northeast and southeast sides of the argillite outcrop are formed by a north-trending, low-angle normal fault dipping approximately 45 degrees to the East. Woods Tunnel intersects the subsurface projection of this fault about 100 m west of its portal and then continues another 300 m into the mountain. Inspection of numerous pieces of dump material indicates most of the tunnel was driven through a dark greenish-gray, mottled-with-pink, fine- to medium-grained, equigranular granodiorite to monzonite. Some fragments show a very crude foliation imparted by lenticular masses of greenish-black chlorite. One specimen was found exhibiting a gradational contact between a medium-grained monzonite and a fine-grained monzonite that is similar to the dike-like body exposed on the surface. One sample of medium-grained monzonite sampled from the dump was thin sectioned. The rock has a well developed mortar texture. It consists of strongly sheared pink orthoclase and lenses of chlorite with minor disseminated and broken crystals of plagioclase, quartz, epidote, and magnetite. Traces of clinopyroxene and hornblende were observed along with a few veinlets of quartz, sericite and pyrite.

Previous investigators (Lasky, 1932; Smith, 1963; Burton, 1971) and this author all consider the argillite to be Precambrian in age. An angular unconformity, overlain by Pennsylvanian sandstones, along the west face of the pentagon and the metamorphic character of these rocks are the basis of this conclusion. Lasky and Smith disagree, however, with respect to the age of the Woods Tunnel intrusive. Lasky (1932, p. 122-123, fig. 19) interpreted the monzonite in Woods Tunnel to be of Tertiary age. Smith described these same rocks as granodiorite gneiss of Precambrian age. A. J. Budding (oral commun., 1978) has observed amphibolite masses within the intrusive body in Wood's Tunnel. These relations suggest the Wood's Tunnel intrusive is most likely Precambrian in age.

Pennsylvanian Rocks

Well-stratified limestones, shales and sandstones of recognized Pennsylvanian age (Herrick, 1899, p. 75; Needham, 1937, p. 12; Kottlowski, 1960, p. 56) are exposed in a roughly rectangular area, 1.5 by 0.75 km, on the lower northeast face of Socorro Peak. These rocks are subdivided here (pl. 1) into the dominantly terrigenous strata of the Sandia Formation and the conformably overlying Madera Limestone. This nomenclature is in agreement with that currently used by the U.S. Geological Survey (Dane and Bachman, 1965).

Gordon (1907b) named these formations and defined them as part of the Magdalena Group from a type section in the Kelly district of the Magdalena Mountains. For details on the problems of nomenclature, age, stratigraphy, facies relationships, and environments of deposition of the Pennsylvanian system in central and southern New Mexico; the writer recommmends the comprehensive works of Kottlowski (1960) and Siemers (1978).

Many geologists (Needham, 1937, p. 14; Kottlowski, 1960, p. 136; etc.) including Gordon himself (1907b, p. 811) have pointed out the difficulty in recognizing the Sandia and Madera formations in southern and eastern New Mexico. Siemers (1978, p. 9) indicates this recognition problem results from rapid facies and thickness changes. This problem, however, is not encountered for the Socorro Peak area, which is only 24 km from the type section. A reference section of the Pennsylvanian strata measured on Socorro Peak (Appendix A, Section 1) indicates a lithologic sequence very similar to the type section (Gordon, 1907b, p. 807) in the Magdalena Mountains. Recent work by Siemers (oral commun.,

1978) has shown Gordon's type section to be faulted and incomplete. The total of 1102 feet (335.9 m) of Pennsylvanian strata measured at Socorro Peak must be regarded as a minimum thickness, since the section has a fault contact at the base and an erosional unconformity at the top.

As suggested by Siemers (1978, p. 147) the gradational Sandia-Madera contact is placed at the horizon below which terrigenous clastic rocks predominate and above which micritic limestones predominate. This contact typically follows a break in slope below cliff- and hogback-forming limestones of the lower Madera, and above slope-forming shales of the upper Sandia (Wilpolt and Wanek, 1951).

Sandia Formation

West-dipping beds of the Sandia Formation crop out to the north of Pathway Canyon in a wide north-trending belt near the foot of Socorro Peak (pl. 1, fig. 3). The general lithology of the Sandia Formation in this area is summarized by the measured section surveyed at the south end of the outcrop belt (Appendix A, Section 1). The formation is informally divided into a lower ledge-forming unit of dark-brown siliceous quartz sandstones (quartzites), 91.1 m thick; and an upper slope-forming unit of black carbonaceous shale, 62.5 m thick. The uppermost 25 m of the shale unit typically contains 3 or 4 ledge-forming, fossiliferous dolomitic limestone beds about 1 to 3 m thick that weather to a yellowish-brown color.

The measured thickness of the lower quartzite unit is a minimum thickness since the lower contact is a fault. About 250 m north of the line of measured section the base of the quartzite unit is exposed at an angular unconformity cut on Precambrian argillite. Above the angular unconformity, the quartzite forms a cliff representing approximately 30 m of sandstones. This apparent change in thickness is probably not real but rather the result of a truncation of the quartzite by a gently south-dipping fault which also forms the southwest edge of the Precambrian outcrop (pl. 1).

The lower quartzite unit is not exposed to the north of the Precambrian outcrop, except for a small triangular fault block juxtaposed against the argillite on its northwest side. The west side of this triangular block is marked by a long sill-like body of white rhyolite (Tirl, pl. 1) that cuts across the Sandia beds at a small angle. The intrusive contact of the rhyolite is well exposed at its south end where it pinches out against a quartzite bed near the middle of the Sandia Formation. The quartzite has been bleached white, silicified(?) and contains minor amounts of argillized feldspar and small flecks of pyrite(?) oxidized to limonite. Previous investigators (Lasky, 1932; Smith, 1963; Burton, 1971) mapped this bleached quartzite bed of the Sandia as a southern continuation of the porphyritic rhyolite. Sandia quartzites are strongly brecciated, silicified and mineralized with barite and galena along the fault zone which cuts the Precambrian argillite on its east side.

The conformable contact at the top of the Sandia Formation with the overlying Madera Limestone is well exposed and expressed by a topographic break at an elevation of 5350 feet on the northeast end of the Socorro Peak escarpment (NW/4, Sect. 4, T3S, R1W). Below this gradational contact, the black shale slopes are broken by several ledges of limestone about 15 to 30 m apart. This sequence of shales and interbedded limestones in the upper Sandia Formation is repeated twice by low-angle normal faults just west of the alluvial toe of the mountain front (pl. 1).

A small fault block of slope-forming shales (mudstones?) overlain by partially silicified limestones is located about 300 m northwest of the Woods Tunnel portal (pl. 1). The correlation of these poorly exposed slope-forming beds with the upper Sandia shales is uncertain. Alternatively, the entire outcrop could be upper Madera Limestone, including an interval of slope forming interbedded mudstones and limestones, such as units 7 and 9 of Section 1, Appendix A.

Needham (1937, p. 12, 14-15) and Siemers (1978, p. 147) consider the Sandia Formation to be entirely Atokan, (upper lower Pennsylvanian) in age. Siemers (op. cit.) has demonstrated the approximate coincidence of the Sandia-Madera contact with the Atokan-Desmoinesian faunal break in the Socorro area.

Madera Limestone

The Madera Limestone forms well-stratified light-gray ridges and cliffs on the middle slope of Socorro Peak's eastern face (pl. 1). These bold outcrops are predominantly gray, massive, thick-bedded, moderately fossiliferous, micritic limestones (Appendix A, Section 1). Zones of black chert nodules are common.

Below Big Cliff (fig. 3, pl. 1) three cliffy intervals of limestone, each 30 to 40 m thick, are separated by two slope-forming units of rhythmically (?) interbedded, gray, calcareous mudstone and micrite with some sandy limestones. The lower mudstone-limestone unit (unit 7, section 1, Appendix A) occurs at essentially the same horizon as 10 m of argillaceous to feldspathic sandstones described by Smith (1963, p. 189, fig. 4, units 9, 10 and 11) in a measured section 750 m to the north. The Madera-Sandia contact in Smith's Section II (op. cit) is at the base of his unit 26, about 43 m below the feldspathic sandstones. This abrupt lateral change in lithology presumably represents a rapid facies change similar to that described by Kottlowski (1960, p. 57).

At the upper end of Pathway Canyon, on its north side, several limestone beds at the top of the Madera have been replaced by cryptocrystalline silica. These silicified limestones are recognized by a hard flinty character, limonite staining (from oxidized pyrite?) and a lack of effervescence in acid. They do, however, retain the distinctive thick and uneven bedding of the original limestones; also fossil forms are sometimes visible on weathered surfaces. The silicified zone at the top of the Madera (unit 10, Section 1, Appendix A) gradually thins and ends 850 m north of Pathway Canyon.

On Socorro Peak, the Madera Formation is unconformably overlain by the lower lithic-rich tuff member (Tlt_1, pl. 1) of the Luis Lopez Formation, which filled the Socorro caldera. This middle Oligocene erosion surface at the caldera wall locally overlaps a more widespread late-Eocene erosion surface at the base of the Oligocene Datil volcanic pile (Chapin and others, 1975). Kottlowski (1960, p. 57) suggests that as much as 250 m of Virgilian, Missourian and Desmoinesian (middle and upper Pennsylvanian) limestones and arkosic limestones of the upper Madera were removed from the Socorro Peak section by Tertiary erosion.

Outcrops of silicified limestones in the Woods Tunnel area are almost certainly the downfaulted equivalent of the silicified upper Madera limestones in Pathway Canyon, south of Big Cliff. This downfaulted block shows a noteable similarity in stratigraphic and structural relationships to that of the Pathway Canyon area (compare rhyolite dike patterns and dip changes of Madera beds as shown on pl. 1). In contrast to their topographically higher equivalent, these silicified limestones are locally brecciated and cemented by a second silicification event. Lasky (1932, pl. 4) and Burton (1971) had previously mapped these silicified limestones at Woods Tunnel as Sandia Formation, probably interpreting them as quartzites.

MIDDLE TERTIARY ROCKS OF THE DATIL-MOGOLLON FIELD

The Datil-Mogollon volcanic field represents the largest erosional remnant of a continuous Oligocene to early Miocene volcanic plateau that once covered most of southwestern New Mexico and extended into southeastern Arizona (fig. 1). On a continental scale, the Datil-Mogollon field is a small fraction of a great middle Tertiary volcanic belt that follows the western flank of North America from southern Mexico to Oregon (Lipman and others, 1972, fig. 2b; Elston, 1976, fig. 1).

The Socorro-Magdalena area comprises about five percent of the Datil-Mogollon field on its northeastern flank. Detailed mapping and K-Ar dating of the Magdalena Project have provided a volcanic history of the Socorro region that is remarkably similar to that of the well documented San Juan volcanic field in Colorado (Lipman and others, 1970; Lipman, 1975). Both areas appear to follow the same general magmatic evolution, as summarized by Lipman (1975, p. 6), beginning with "...intermediate-composition lavas and breccias followed closely in time by more silicic ash-flow tuffs and ending with a compositionally bimodal association of basalt and rhyolite." However, some important differences between these contemporaneous volcanic fields are becoming conspicuous. Each of the three magmatic stages described above for the San Juan field begins about 2 or 3 m.y. earlier in the Datil field. Also, the ash-flow sheets of the Socorro area are considerably more silicic; most are high-silica rhyolites with abundant phenocrystic quartz, which is rare in most of the San Juan tuffs. In addition, in the Socorro area, basaltic-andesite lavas appear early in the volcanic sequence, are more voluminous, and are mostly interbedded with the major ash-flow sheets, rather than capping them, as occurs in the San Juan field. Except for the timing of initial volcanism, these differences may be interpreted as the result of regional crustal extension beginning about 32 m.y. ago in central New Mexico during the peak of middle Tertiary volcanism (Chapin, 1978). Signs of regional extension in the San Juan region appear at about 26 m.y. ago during the waning stage of the middle Tertiary magmatic event (Lipman and Mehnert, 1975).

General Stratigraphic Sequence

In the Socorro area, the Datil-Mogollon volcanic pile is commonly represented by 0.9 to 1.5 km of calc-aklaline, mafic to silicic flows and volcaniclastic sedimentary rocks.

The regional sequential and lateral relationships of the Datil volcanic strata as established through the Magdalena Project, and their correlation to the subdivisions of Tonking (1957), are shown in figure 7.

The Datil volcanics are divisable into three major petrologic-stratigraphic units comprised by: 1) early intermediate composition, andesite to quartz latite (?) lavas and volcaniclastic sedimentary rocks of the Spears Formation (37-32 m.y.); 2) major silicic, rhyodacite to high-silica rhyolite, ash-flow tuffs of several formations (32-26 m.y.); and 3) widespread basaltic-andesite lavas of the La Jara Peak Basaltic Andesite (30-24 m.y.). This sequence and subdivision of the volcanic strata is essentially that determined by Tonking (1957); however, his three-tier layer-cake geometry is now recognized to change laterally into an intertonguing pile of volcanic strata. It is significant, however, that the intertonguing strata maintain the same general sequence.

A fourth type of petrologic-stratigraphic unit now recognized is of lesser volume and restricted to calderas and caldera margins. These caldera-fill units contain lavas, agglomerates, local ash-flow tuffs and volcaniclastic sedimentary rocks that range from andesite to rhyolite in composition. Some of the calderas were evidently filled to

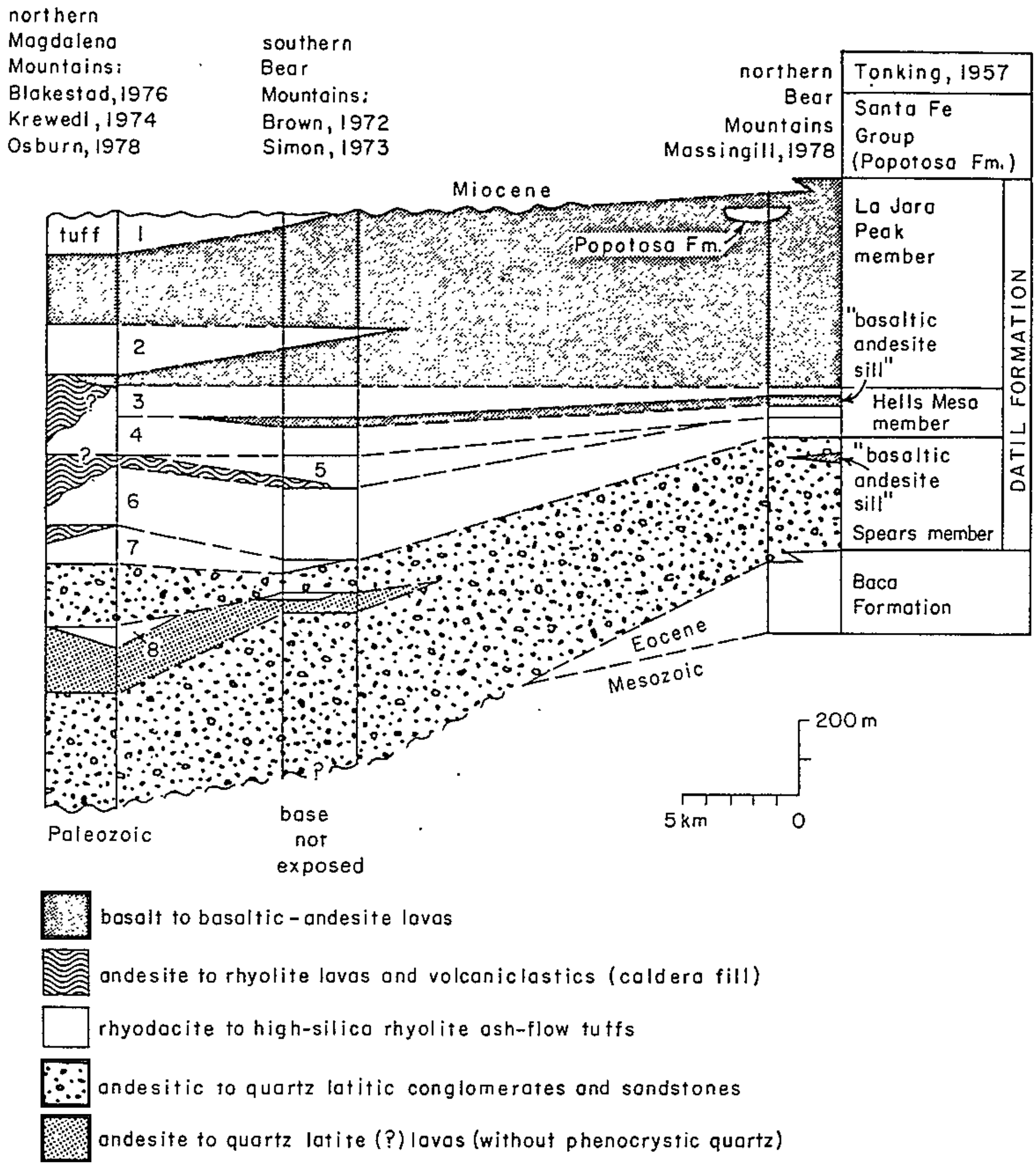


Figure 7. North-south stratigraphic cross section, applicable in general to the Socorro-Magdalena area, illustrating the sequence and lateral relationships of major petrologic-stratigraphic units within the Oligocene Datil volcanics. Arbitrary datum is top of the A-L Peak Tuff. Thicknesses of units estimated from geologic maps of authors cited; tuff thicknesses in northern Bear Mountains from C. E. Chapin (unpub. recon. section at Hells Mesa). See text for discussion. Ash flow tuffs are: 1) tuff of South Canyon, 2) Lemitar Tuff, 3) A-L Peak Tuff -- pinnacles(?) member, 4) A-L Peak Tuff -- flow banded member, 5) A-L Peak Tuff --gray massive member, 6) Hells Mesa Tuff, 7) tuff of Granite Mountain, and 8) tuff of Nipple Mountain.

overflowing so that intermediate lavas are interbedded with the major ash-flow sheets on the caldera margins. Conversely, some regional lavas of the La Jara Peak Basaltic Andesite and some younger ash-flow sheets were ponded in the older calderas either prior to, during, or in lieu of, their filling by locally erupted flows.

Relative volumes of the three major petrologic units, as suggested by figure 7 are probably not subequal. At most localities in the Socorro region, early intermediate rocks are at least twice as thick as the overlying silicic ash-flows. Younger basaltic-andesite lavas are usually equal or slightly greater in thickness compared to the underlying ash-flows; however, the mafic lavas probably cover less than a quarter as much area, in the Datil-Mogollon field in comparison to the ash-flows. Elston and others (1976, fig. 10) show that these basaltic-andesites are restricted to a wide belt along the southern margin of the Colorado Plateau and do not occur across the entire volcanic field as do the ash-flow sheets. Figure 7 is also misleading with respect to Oligocene structures. Abrupt thickness changes, which have been observed across cauldron ring fractures and across regional faults, are not shown in this figure.

Stratigraphic Nomenclature and Correlation

The history of the development of stratigraphic nomenclature, subdivision, and correlation of volcanic strata in the northeastern Datil-Mogollon field is a long and complex story that reflects the nature of the volcanic pile itself. Important events in the nomenclatural evolution and subdivision of the volcanic stratigraphy have been discussed in the "Previous Work" section of the Introduction. A chronological summary of the subdivision of the volcanic pile and associated changes in nomenclature up through this report is shown in figure 8. Table 1 summarizes the general Tertiary volcanic stratigraphy of the Socorro Peak area. This terminology follows the general format for volcanic strata used by the U.S. Geological Survey as exemplified by the geologic map of the Valles caldera in the Jemez Mountains of New Mexico (Smith and others, 1970; see also Christiansen and others, 1977).

Petrologically and texturally distinctive sheets of ash-flow tuff form the key elements of the volcanic stratigraphy in the Socorro region. The distinctive nature of these individual ash-flow sheets comes from their genetic relationship to a dynamic system of eruption and deposition in which the final make up of the ash-flow cooling unit is determined by several variables. These variables probably

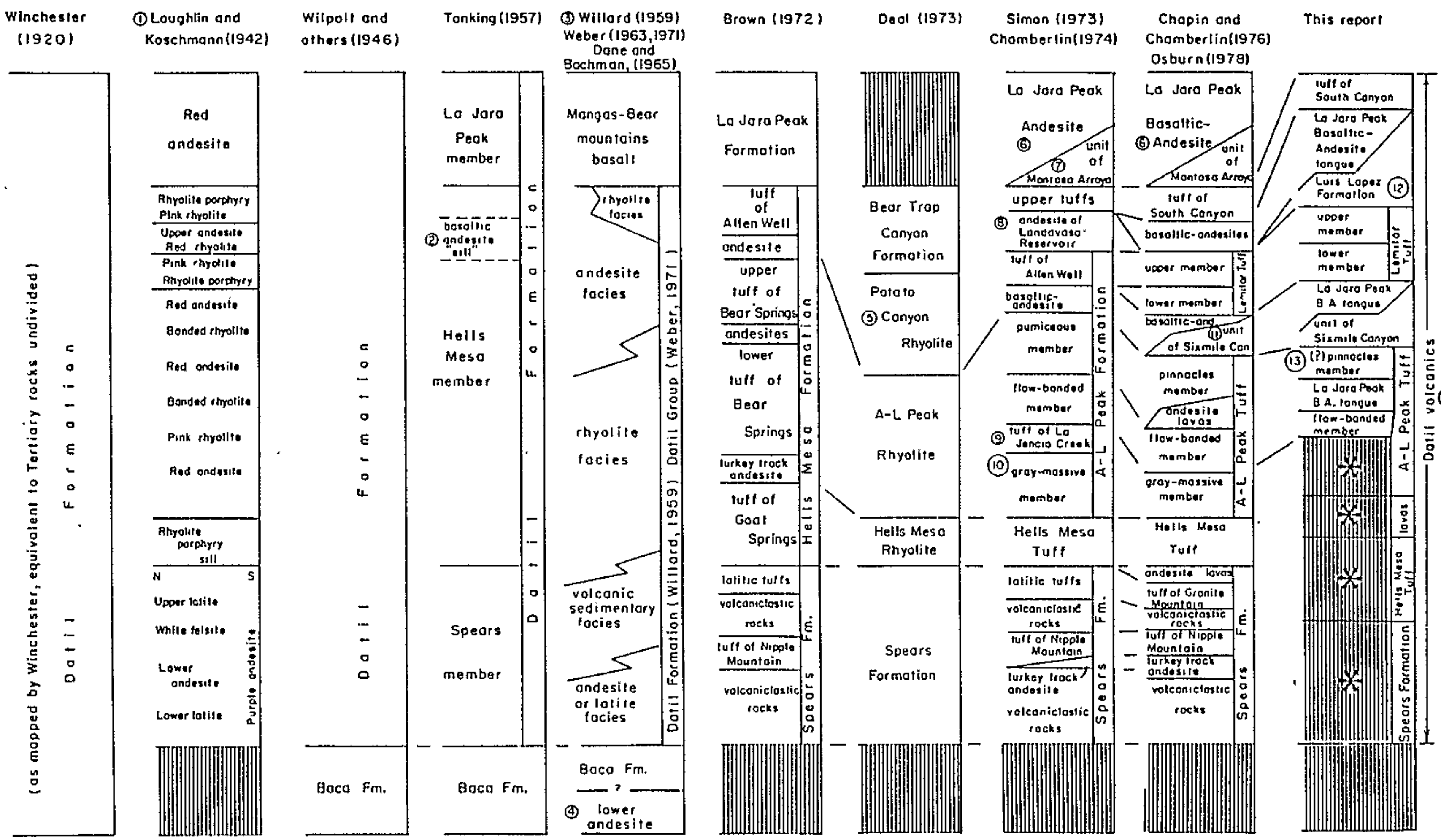


Figure 8. Correlation diagram summarizing the nomenclatural evolution of the Datil volcanics in the Socorro region (revised after Osburn, 1978).

See footnotes on opposing page.

Table 1 - Generalized Cenozoic volcanic stratigraphy of the Socorro Peak volcanic center.

Santa Fe Group

LATE BASALTS AND RHYOLITES:

Sierra Ladrones Formation:
- basalt of Sedillo Hill (4.0 m.y.)

Popotosa Formation
- basalt of Bear Canyon
- Socorro Peak Rhyolite (7.4-11.9 m.y.)[1]
- basalt of Kelly Ranch

Datil Volcanics

MAJOR ASH-FLOW TUFFS:
- tuff of South Canyon (26.2 m.y.)
- Lemitar Tuff (27.9 m.y.)[2]
- A-L Peak tuff (base, 31.8 m.y.)
  - pinnacles (?) member (upper tuff of Bear Springs, Brown, 1972)
  - flow-banded member
- *Hells Mesa Tuff (32.2 m.y.)[2]
- *Spears Formation
  - *tuff of Granite Mountain member

LAVAS AND ASSOCIATED ROCKS PENECONTEMPORANEOUS WITH ASH-FLOW TUFFS:

1) La Jara Peak Basaltic Andesite (~30-23.8 m.y.)[1]: widespread and thick pile of basaltic-andesite lavas erupted from vents mostly north of $34^\circ15'$; forms three tongues, respectively below the pinnacles (?) member of the A-L Peak Tuff, Lemitar Tuff, and tuff of South Canyon

2) Caldera filling andesite to rhyolite lavas and associated volcaniclastic rocks intertonguing with major ash-flow sheets:
   - Luis Lopez Formation (28.6 m.y.)
   - unit of Sixmile Canyon (sandstones of Tower Mine drill hole)

EARLY INTERMEDIATE LAVAS AND VOLCANICLASTIC SEDIMENTARY ROCKS:

*Spears Formation (34.5-37.1 m.y.)[1] mudflow deposits, conglomerates and sandstones derived from andesite to quartz latite(?) (without phenocrystic quartz) lavas, with minor interbedded lavas; intertongues with lower part of major ash-flow tuff sequence.

1 age range of multiple flow units

2 average of analytically synchronous dates that have overlapping error bars

* units not exposed in study area, but present in the subsurface of Strawberry Peak area (Lemitar Map)

include the rate and intensity of eruption, temperature, magma composition -- including phenocryst and volatile content, viscosity, and volume of the erupted material. R. L. Smith (1960b, p. 153) summarized this important charcteristic of ash-flow tuffs in his paper on zonal variations with this statement:

> "Any system that includes several variables where one variable can change the entire appearance of a rock body, or any part of it, is a very flexible system, and must be treated as such. The chance that two of more welded ash flows will show no differences as a whole is extremely slight. On the other hand, the close similarities that commonly exist between some welded ash flows may be confusing and it may be impossible to distinguish between their equivalent zonal facies. Differentiation may then be dependent upon detailed petrographic studies of phenocrystic minerals coupled with careful field study."

Regional correlation of ash-flow sheets between the fault-block ranges of the Socorro-Magdalena area is based on the original detailed descriptions and petrographic data of most of the major ash-flow sheets presented in Brown's thesis (1972) on the southern Bear Mountains. At least seven ash-flow sheets of sufficient thickness and extent to require cauldron collapse appear to have been erupted from a complex of overlapping and nested calderas in the Magdalena, Chupadera and San Mateo Mountains (Chapin and others, 1978) (fig. 9). The general petrologic characteristics of ash-flow sheets that are present in the study area

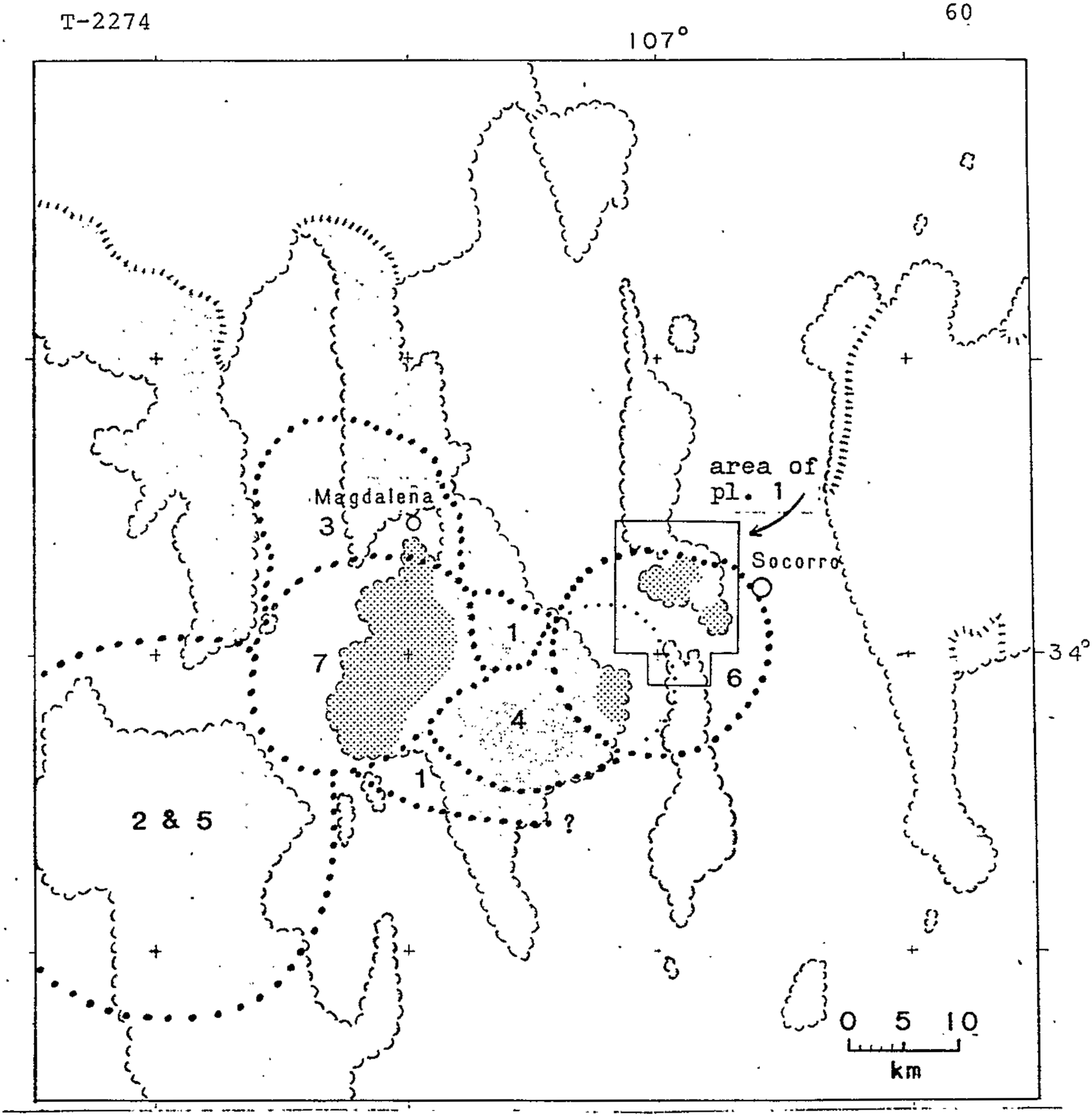


Figure 9. Complex of seven overlapping and nested Oligocene cauldrons in the Socorro-Magdalena area. From oldest to youngest the cauldrons and their associated ash-flow tuff sheets are 1) North Baldy -- Hells Mesa, 2) Mount Withington -- A-L Peak Tuff, gray-massive member, 3) Magdalena -- A-L Peak Tuff, flow-banded member, 4) Sawmill Canyon -- A-L Peak Tuff, (?) pinnacles member, 5) Mount Withington -- Potato Canyon Tuff, 6) Socorro -- Lemitar Tuff(?), 7) Hop Canyon -- tuff of South Canyon. From Chapin and others (1978). Sawmill Canyon cauldron modified after Petty (1979); smaller dots indicate projected eastern margin buried by Lemitar Tuff and younger basin fill.

are summarized in Table 2. Correlation problems related to nearly identical zones in different ash-flow sheets and also related to hydrothermal alteration are discussed in the following descriptions of the stratigraphic units.

In contrast to the distinctive ash-flow tuffs, mafic to silicic lavas that are interbedded in the ash-flow sequence at different horizons are often indistinguishable except for their positions relative to the tuffs. The three widespread tongues of La Jara Peak Basaltic Andesite are petrographically indistinguishable. Porphyritic andesite lavas of caldera-fill members in the Luis Lopez Formation are mineralogically and texturally identical to the early intermediate lavas of the Spears Formation. Specimens of 20 m.y.-old quartz latitic lavas of Water Canyon Mesa that occur at the base of the Popotosa Formation (Osburn, 1978, p. 60) are inseparable lithologically from the 10- to 12-m.y.-old quartz latites on Socorro Peak that are interbedded in the upper Popotosa Formation.

A-L Peak Tuff

A continuous section of welded crystal-poor rhyolite ash-flow tuffs about 609 m thick and exposed on A-L Peak in the northern San Mateo Mountains has been briefly described and named the A-L Peak Rhyolite by Deal and Rhodes (1976,

Table 2 - Petrologic summary of ash-flow sheets exposed in the area of the Socorro Peak volcanic center.

Unit	Number of analyses	percent SiO_2 [1]	percent phenocrysts [2]	quartz-feldspar ratio (mode)	plagioclase-sanidine ratio (mode)	pumice[5] content	lithic[5] content	radiometric age[6] (m.y.)	magnetic polarity[7]
tuff of South Canyon	1	78	3-21	0.4-1.5(1.1)	0-.1 (.06)	A	P-A	26.2±1.0	R
Lemitar Tuff:									
upper member	6	64-71	27-45	.01-.3 (.2)	.2-1.1 (.6)	P-A	R-A	27.9 (avg.)	N
lower member	3	74-77	8-15	.1-.7 (.2)	0-.2 (.05)	P-M	R-M, P		N
A-L Peak Tuff:									
pinnacles (?) member (upper tuff of Bear Springs)	1	75	0-7	[3] .1-.3 (.2)	~.02[4]	R-A, M	R, M (top)	31.8±1.7 (base)	R
flow-banded member	3	73-76	6-11	.03-.2 (.1)	~.05[4]	A	R-P		R

1) data from Miesch, 1956; Deal, 1973; Spradlin, 1976; and D. L. White, unpublished (see p. 25)
2) modal data from Brown, 1972; Simon, 1973; Osburn, 1978, and this report
3) excluding very crystal-poor basal zone, Q/F = 0 - 1.0
4) based on estimate of plagioclase in perthite-like feldspars (Brown, 1972)
5) visual estimates of volume percent: A = abundant (>5 percent), M = moderate (2-5 percent), P = poor (2-0.2 percent), R = rare (<0.2 percent); the most common occurrence is underlined.
6) data from Burke and others, 1963; E. I. Smith and others, 1976; and C. E. Chapin, unpublished

7) preliminary unpublished data from observations in Joyita Hills by G. R. Osburn, 1978; correlation with units and data of Strangway and others (1976) are uncertain

p. 54). This report adopts the subsequent terminology of "A-L Peak Tuff" later applied to this unit (Chapin and Deal, 1976; Chapin and others, 1978).

Prior to the work of Deal and Rhodes, Brown and Chapin had measured and described in detail, including modal analyses, equivalents of the A-L Peak Tuff exposed in the Southern Bear Mountains (Brown, 1972, p. 31-45). Brown observed two cooling units that he informally named the "lower -- and upper -- tuff of Bear Springs". They are locally separated by a thin tongue of basaltic lavas.

On the basis of a stratigraphic section (C. E. Chapin, 1975, unpub.) reconnoitered on the northeast flank of A-L Peak, Chapin and Deal (1976) subdivided the A-L Peak Tuff into three texturally or morphologically distinctive units using the descriptive terminology of Brown (1972). Informally named and given member status, the units consist of (1) a lower gray-massive member (Ibid., p 37) that is partially to densely welded, pumice-poor, and ranges from pinkish gray to gray in color; (2) a medial, primary welded, "flow-banded member" (Ibid., p 38) containing abundant extremely flattened and lineated pumice that gives the tuff the appearance of a flow-banded lava; and (3) an upper "pinnacles member" (Ibid., p. 42) that is densely welded, craggy, cliff-forming, and locally containing abundant lineated pumice but not flow-banded in appearance. Chapin and Deal (1976) correlated the gray-massive and flow-banded members with the "lower tuff of Bear Springs" and the pinnacles member with the "upper tuff of Bear Springs" as used by Brown. This has lead to the interpretation of the A-L Peak Tuff as a composite sheet (R. L. Smith, 1960b, p. 158). A reappraisal of this correlation has cast some doubt, presently irresovable (C. E. Chapin, 1979, oral commun.), on the equivalence of the "pinnacles member" on A-L Peak to the "upper tuff of Bear Springs". For this reason, the undoubtable equivalent of the "upper tuff of Bear Springs" in this study area is referred to as the "pinnacles (?) member" of the A-L Peak Tuff.

Detailed mapping of the Magdalena Project has demonstrated the gray-massive, flow-banded, and pinnacles(?) members of the A-L Peak Tuff to be widespread and largely coextensive ash-flow sheets that typically form two mappable cooling units. In the outflow-facies environment of the Gallinas - Bear - Lemitar mountains and Joyita Hills, the pinnacles(?) member occurs as simple-cooling unit approximately 75 to 100 m thick that has a distinctive very crystal poor (0-3 percent) basal zone. It is normally separated from the lower cooling unit by 10 to 50 m of basaltic andesite lavas. The lower cooling unit, generally 150 to 200 m thick, on the outflow sheet is largely a compound cooling unit consisting of the flow-banded member welded, across a partial cooling break, to the top of the gray-massive member. Poorly welded, lithic-rich tuffs are common at the top and base of the lower cooling unit.

In all of the investigations of the Magdalena Project the contact between the gray-massive and flow-banded members has not been mapped even though they are interpreted to be separate ash flow sheets. For this reason it is not commonly recognized that the flow-banded ash-flow sheet has a basal and a top zone that are not flow banded or lineated but which grade into the flow-banded core of the sheet (see Appendix A, section 2, unit 4, for description of the flow-banded member in the Lemitar Mountains). This progressive variation from a random fabric to a lineated fabric is interpreted as a zonal variation in ash flows that grade from "hot" secondary welded tuff to a "hotter" primary welded tuff.

The origin and characteristics of primary welded tuffs are discussed in detail by Chapin and Lowell (1979). An abnormally high emplacement temperature is evidently the critical factor in the development of primary welded ash-flow tuffs. Their strongly lineated fabric is characteristic because the high temperature allows agglutination of the hot glass fragments contemporaneously with their deposition (in a laminar-flow boundary layer) and before forward motion ceases. The flow-banded member of the A-L Peak Tuff is best distinguished from rhyolitic lava by its sheet-like geometry and its gradational boundaries with "normal" (secondary) welded tuff, that commonly retains its original vitroclastic texture. Disseminated lithic fragments and sparse broken phenocrysts are small scale indications of the pyroclastic origin of the flow-banded member.

All three members of the A-L Peak Tuff contain about 5 to 10 percent phenocrysts with medium-grained sanidine being about five times more abundant than small, but omnipresent, quartz (Brown, 1972, p. 34, 44). Chemical analyses of five bulk-rock samples of the A-L Peak Tuff range from 73.2 to 76.1 percent SiO_2 (Deal, 1973; D. L. White, unpub. data, see p. 28). When corrected for water content, most of the analyses are very close to 76 percent SiO_2. The uniform chemistry of these high-silica rhyolites parallels their esentially constant mineralogy.

A separate of sphene from a local occurrence of reddish brown vitrophyre at the base of the section on A-L Peak has been dated by the fission track method at 31.8 ± 1.7 m.y. (E. I. Smith and others, 1976). This age is in good agreement with bracketing ages of 32.1 and 32.4 m.y. (Burke and others 1963) on the underlying Hells Mesa Tuff and a 27.0 and 28.8 m.y. dates (Appendix B) on the overlying Lemitar Tuff. A K-Ar age of 27.4 m.y. ± 1.2 m.y. (Appendix B) on sanidine from a sample of the flow-banded member in the southeast Lemitar Mountains is almost certainly too young. This sample was taken from an area of pervasive potassium metasomatism (Chapin and others, 1978).

Deal and Rhodes (1976) propose the A-L Peak type section to be a caldera-facies accumulation erupted from the Mount Withington cauldron (fig. 9). Ongoing mapping of the Magdalena Project has shown that the Magdalena and Sawmill Canyon cauldrons may also be interpreted as sources for members of the A-L Peak Tuff (Chapin and others, 1978).

A small triangular area of A-L Peak Tuff is exposed at the far northwestern corner of the study area (pl. 1), just northeast of Bug Mountain. These complexly faulted outcrops occur at the south end of a north-trending belt of A-L Peak Tuff that generally forms the backbone of the western range of the Lemitar Mountains. Detailed descriptions of stratigraphic units that make up the A-L Peak Tuff in the central Lemitar Mountains are presented in a composite measured section (section 2) in Appendix A.

In the Bug Mountain area only the pinnacles(?) member (Talp, pl. 1), a thin tongue of interbedded basaltic andesites (Tala, pl. 1) and the flow-banded member (Talf, pl. 1) are exposed. Tuffs found below the flow-banded member in the central Lemitar Mountains are not exposed here. About 1 km north of the map boundary, the flow-banded member rests directly on massive pyroxene andesite and hornblende rhyodacite (?) lavas, which in turn overlie the Hells Mesa Tuff. These intermediate lavas pinch out abruptly to the north. Their stratigraphic horizon is replaced further north by the gray massive member, the mottled tuff, and lower lithic-rich tuff of the A-L Peak Tuff (Appendix A). This relationship suggests that the intermediate lavas formed a paleotopographic high, about 70 m higher in elevation than the area to the north, prior to deposition of the lower cooling unit of the A-L Peak Tuff.

The reader is referred to Appendix A, section 2, for detailed descriptions of the pinnacles(?) member and the flow-banded member; these generally apply to the Bug Mountain area. The only significant difference is that in the Bug Mountain area the steeply dipping, densely welded, ash-flows do not form the typical hogbacks and cliffs found to the north.

Petrographic study of samples from the measured section of A-L Peak Tuff in the Lemitar Mountains has shown some subtle differences in comparison to equivalent rocks in the southern Bear Mountains and other areas. Estimates of total phenocryst content and variation in abundance are in good agreement with modal data of Brown (1972, p. 34 and p. 44). In most samples subhedral sanidine forms about 5 to 10 percent of the rock and small, rounded quartz about one percent.

Outside of the Lemitar Mountains, the A-L Peak Tuffs commonly contain minor to trace amounts of plagioclase some of which occurs in a perthite-like relationship with sanidine (Brown, 1972, p. 33). No plagioclase was observed in any of the A-L Peak samples from the Lemitar Mountains section. This apparent absence of plagioclase is now recognized as a manifestation of pervasive potassium metasomatism that occurs throughout the study area. Potassic clays and secondary potassium feldspar have ubiquitously replaced plagioclase in all the Oligocene tuffs in the study area. The clays are typically washed out of the thin sections during their preparation. A minor volume of "washed-out" holes in these sections may be easily overlooked, thus leading to the incorrect conclusion that the rock did not originally contain plagioclase.

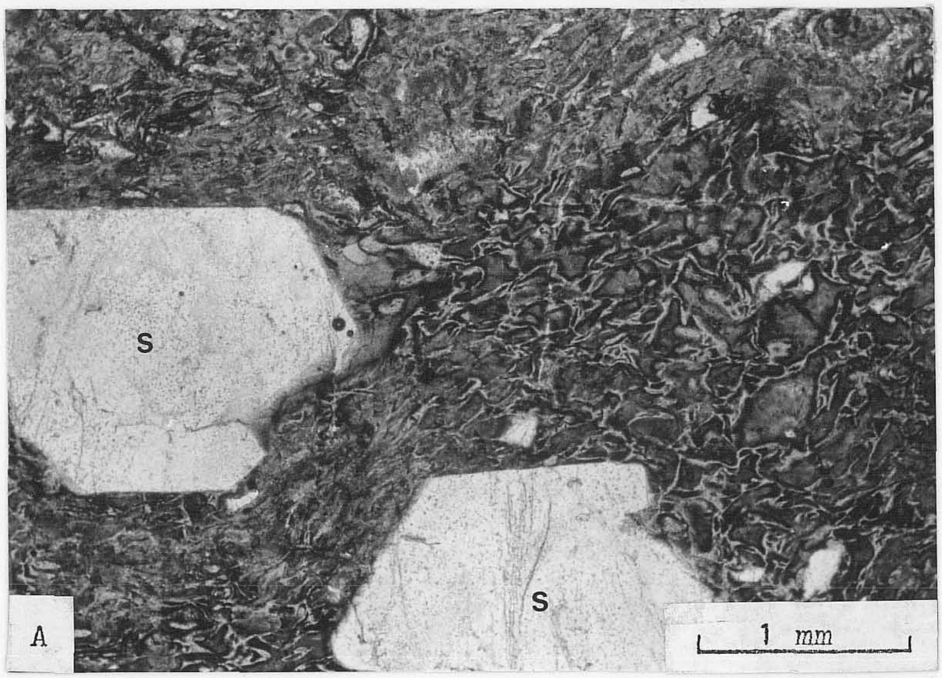
Honeycomb-like, skeletal sanidine phenocrysts are commonly observed in hand specimens of the A-L Peak tuffs. These unusual looking crystals are interpreted to be perthite-like phenocrysts of sanidine in which the plagioclase has been altered to clay and then washed out during weathering.

About one third of the samples from the measured section contain trace amounts of biotite, which in most instances is completely altered to disseminated iron oxides. The reason for this alteration is not clear, since potassium metasomatism apparently does not visibly affect primary potassic minerals (Chapin and others, 1978, p. 124).

Vitroclastic textures are conspicuous in poorly to densely welded samples of the lower lithic tuff, upper lithic tuff and the pinnacles(?) tuff (fig. 10). In the primary-welded zone of the flow-banded member all shard structures have been completely erased by flowage to produce textures indistinguishable from those in rhyolite lavas (fig. 10). Vapor-phase crystallization has largely destroyed glass shard textures in the gray-massive member.

Small-scale chevron folds, broad open folds, and variably oriented, large, similar folds occur locally in the flow-banded member. The origin of these folds is uncertain. The broad open folds are probably primary flow structures since they occur approximately perpendicular to flow lineation. The other types seem to occur adjacent to "early rift" low-angle normal faults, which may have formed while

Figure 10. Photomicrographs in plane polarized light of A-L Peak Tuff from the Lemitar Mountains. A, Specimen from densely welded contact between lower lithic tuff and the mottled tuff (Appendix A, section 2, units 1 and 2). The crystal-poor tuff contains about 5 percent subhedral sanidine(S) and traces of angular quartz, note well-preserved shard texture and conspicuous variation in the degree of deformation of shards due to differential compaction. Section approximately perpendicular to foliation plane. B, Primary-welded, flow-banded member showing typical laminar flow structure of pumice bands wrapped around a large euhedral sanidine(S) and a semiopaque lithic fragment. Vitroclastic origin is not perceptible. Flow laminae formed by stretched pumice tend to be lenticular in comparison to rhyolite lavas but this is not entirely diagnostic.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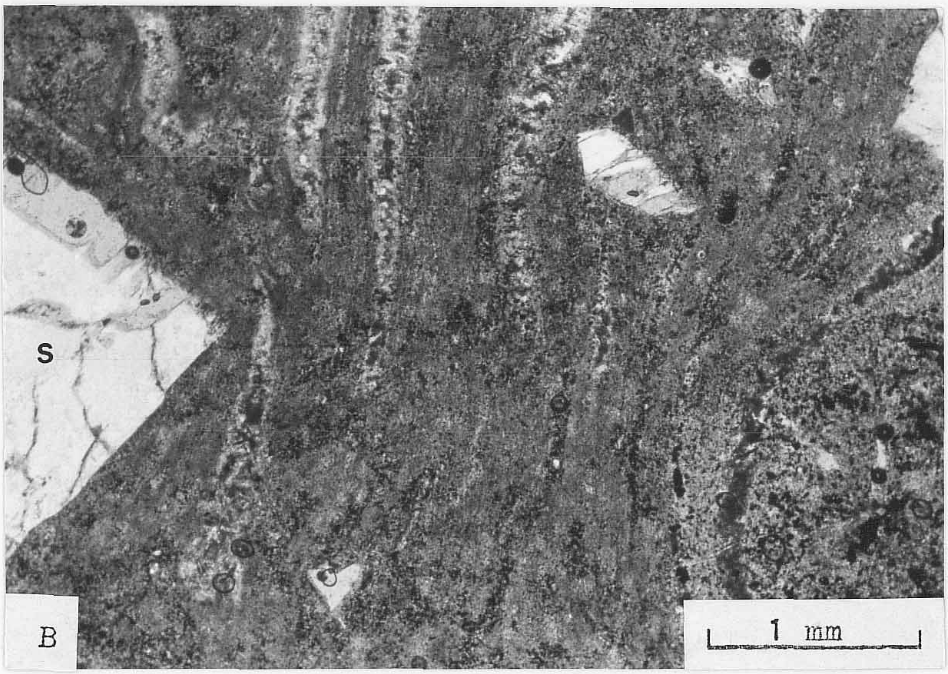
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the tuff was still hot and plastic. More data is needed on these flow structures to verify their origin.

La Jara Peak Basaltic Andesite

A thick series (350-750 m) of basaltic-andesite lava flows that forms the youngest major stratigraphic unit of the Datil volcanics in the northern Bear Mountains was named the "La Jara Peak Member" of the Datil Formation by Tonking (1957, p. 30). Chapin (1971b) reported a radiometric date (23.8±1.2 m.y.) for these mafic lavas and elevated them to formation status giving them the name "La Jara Peak Andesite". In a later paper (Chapin and Seager, 1975), this formation was referred to as the La Jara Peak Basaltic Andesite to better reflect the results of numerous chemical analyses. Seven chemical analyses (Tonking, 1957; D. L. White, unpub.; see p. 28) available for La Jara Peak lavas range from 48 to 57 percent SiO_2. The average value of approximately 53 percent SiO_2 is probably representative of the bulk composition of these mafic lavas.

The original definition of the La Jara Peak lavas by Tonking as the uppermost stratum of the Datil volcanics must now be revised. Regional mapping has demonstrated that the thick continuous piles of basaltic andesite, generally to the north of 34° 15' N (from San Acacia to Hells Mesa) intertongue to the south with A-L Peak Tuff, Lemitar Tuff, and the tuff of South Canyon (fig. 7). Spatial relationships to these ash-flows can be used to define three widespread stratigraphic intervals of basaltic lavas that are otherwise indistinguishable where the tuffs are absent. They are informally referred to in this report as the "lower, middle, and upper tongues of La Jara Peak Basaltic Andesite". These tongues are equivalent to the map units (pl. 1): Tala, Tba_1, and Tba_2 respectively.

Excluded arbitrarily from the La Jara Peak Basaltic Andesite, as defined here, are minor volumes of basaltic lavas that occur near the top of the Spears Formation in the Joyita Hills (Spradlin, 1976) and in the northern Bear Mountains (Tonking, 1957, p. 23, vesicular sill). Relatively thin strata of basalt and basaltic andesite lavas interbedded in the clastic basin-fill of the Santa Fe Group in the Socorro Peak area are likewise not included with the La Jara Peak Formation.

As illustrated in figure 7, the La Jara Peak tongues thicken northward toward their source areas and also become progressively thicker through time. In the Lemitar Mountains, the maximum thickness of the three tongues increases upwards from approximately 33 m, to 170 m, to 335 m (Lemitar Map). The wide distribution of the lower tongue of the La

Jara Peak from the Gallinas Mountains to the Joyita Hills (about 80 km), in conjunction with its average thickness of approximately 30 m, suggests that these lavas were erupted from numerous vents scattered widely over a relatively even surface.

The nature of the basal contact of the La Jara Peak has been interpreted both as conformable (Tonking, 1957) and as an angular unconformity (Willard, 1959; Chapin, 1971b). Recognition of the interfingering of the basaltic lavas with the younger major ash-flow sheets has provided the solution to this controversy. The intertonguing relationship itself necessitates a conformable association derived from penecontempraneous eruptions. However, when mapped in detail, the interbedded Lemitar Tuff forms a key structural datum that demonstrates the existence of <u>internal</u> angular unconformities within the basaltic-andesite sequence. Abrupt changes in thickness, as much as 300 m, of the middle and upper tongues of the La Jara Peak plus the interbedded Lemitar Tuff occur across north-trending low-angle normal faults (early rift faults) in the central Lemitar Mountains (Lemitar Map). This relationship requires these faults to have been developed contemporaneously with the eruption of the volcanic strata (Chamberlin, 1976, 1978). Consequently the La Jara Peak Basaltic Andesite and interbedded tuffs are clearly definable as synrift volcanics; a conclusion reached earlier by Chapin and Seager (1975, p. 307) through a different line of evidence.

North and west of this study area, the youngest La Jara Peak flows locally intertongue with heterolithic volcanic-rich conglomerates of the basal Santa Fe Group assigned to the Popotosa Formation. Within this study area, the La Jara Peak lavas are either conformably overlain by the tuff of South Canyon or unconformably overlain by conglomeratic mudflow deposits of the basal Popotosa Formation.

The age of the La Jara Peak Basaltic Andesite in the Socorro area is bracketed by the 31.8-m.y.-old A-L Peak Tuff (Smith and others, 1976) and the 26.2-m.y.-old tuff of South Canyon (Appendix B). Several K-Ar ages of La Jara Peak whole-rock samples are generally, but not entirely, in agreement with these brackets. The 23.8 m.y. K-Ar age published by Chapin (1971b) suggests a continuation of La Jara Peak basaltic-andesite volcanism after eruption of the tuff of South Canyon. These younger flows would presumably be minor in volume since the tuff of South Canyon normally caps the thick basaltic-andesite sequence in the Lemitar and Magdalena mountains. A K-Ar age of 26.3 m.y. has been reported by Machette (1978) for a basaltic-andesite flow occurring at the top of a thick lava series immediately below the Popotosa Formation at Cerritos de Las Minas. Whole rock K-Ar dates of 30.2 m.y. and 26.6 m.y. (C.E. Chapin, unpub. dates) for La Jara Peak samples from the Bear Mountains fit well within the ash-flow tuff age brackets, but are apparently not in agreement with local stratigraphic relationships.

Available data indicate that the La Jara Peak Basaltic Andesites were largely erupted from dike swarms along what is now the southern margin of the Colorado Plateau. Tonking (1957) and Massingill (1978) have mapped a major swarm of north-trending basaltic dikes intruding Mesozoic strata near the settlement of Riley, northeast of the Bear Mountains. Diabasic textures in many of the larger dikes (Massingill, 1978, p. 131) attest to a present erosion level that is significantly below the original tops of these fissures. At higher stratigraphic levels in the volcanic pile these dikes are not as conspicuous; Massingill (op. cit. pl. 1) did find one dike that cut the A-L Peak Tuff as should be appropriate for a La Jara Peak feeder dike. Two K-Ar ages on these dikes of 24.3 and 24.8 m.y. (C. E. Chapin, unpub. data) indicate their emplacement contemporaneous with the waning stage of the La Jara Peak eruptions.

Some relatively obscure vent areas for the middle or upper tongues of La Jara Peak are present in the Lemitar

Mountains. About 2 km southwest of Polvadera Mountain (Lemitar Map), a 200-m-long exposure of basaltic agglomerate and breccia about 15 m thick strikes anomalously in an east-west direction. A large, rounded hill of poorly exposed basaltic-andesite (SE/4, NE/4, NW/4, Sect. 1, T 2S, R2W), 300 m to the north of this agglomerate, is its most likely source.

Northeast of Bug Mountain, and within this study area, a poorly exposed basaltic-andesite dike and a larger plug-like body are apparently intrusive into the flow-banded member of the A-L Peak Tuff. The larger body is well exposed along the north side of a small ravine (SW/4, NE/4, NW/4, Sec. 26, T2S, R2W). It is variable in lithology and grades from a black to medium-gray, dense, aphanitic, olivine basalt at the borders to a saccaroidal, very-fine-grained diabase toward the center. Sparse, small olivine phenocrysts are present in both rock types and range from fresh to completely altered to reddish-brown iddingsite and hematite. The aphanitic plug (?) rocks match the lithology of the nearly contiguous dike and adjacent flows that overly the A-L Peak Tuff. Exposures in the area are not good enough to eliminate the alternative interpretation that the larger body is a down-faulted block of the overlying flows (Tba_1, pl. 1).

Abrupt thickness changes in the middle and upper La Jara Peak tongues occur in the Bug Mountain area (pl. 1) in association with an east-northeast-striking fault zone and hinge-line structure that is peripheral to the north margin of the Socorro cauldron. Here the middle tongue pinches out abruptly toward the south across this zone and the upper tongue wedges out toward the north, where the hinge structure is truncated by an unconformity at the base of the Popotosa Formation.

La Jara Peak lavas in the study area are generally poorly exposed as rubble covered slopes or low-rounded hills mantled by colluvium. At each locality of exposure -- Bug Mountain, northeast of Strawberry Peak, and the Tower Mine area -- the lithology of these tongues is essentially the same. Fragments of lava on these slopes are dominantly medium-gray to grayish-red-purple aphanitic rocks containing a few percent of small, reddish brown, altered ferromagnesian phenocrysts. Vesicular or scoriaceous fragments are also common; often they contain amygdaloidal calcite and tend to be reddish brown in color. Phenocrystic plagioclase is rare in these lavas.

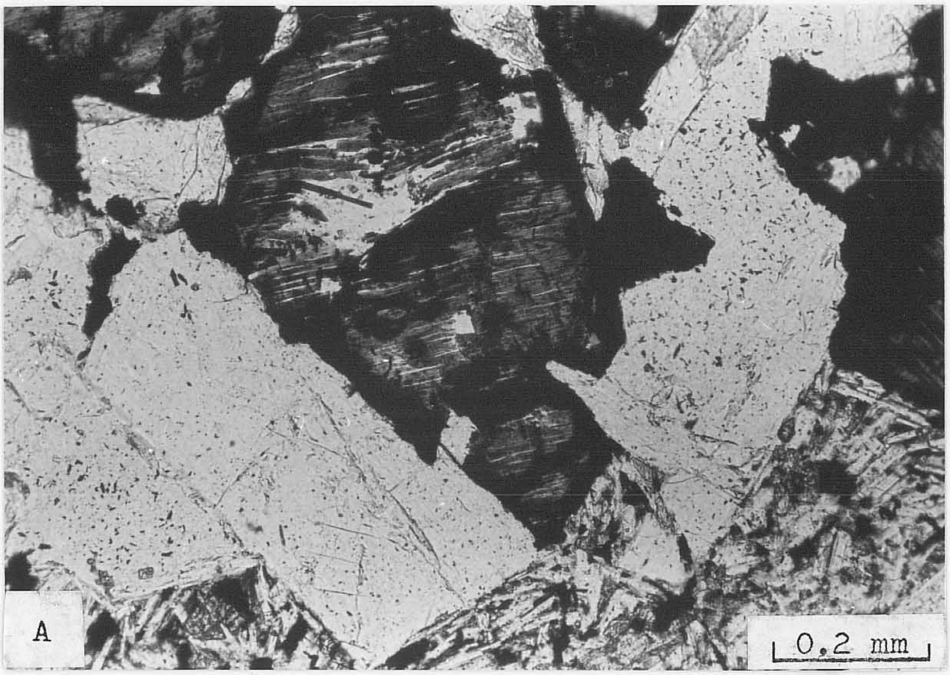
The middle and upper tongues are unusually well exposed about 2 km northeast of Strawberry Peak (SE/4 Sect. 24, T2S, R2W). Measured sections of both tongues were surveyed here and samples were collected for petrographic study (Lemitar Map, measured sections LM-5 and LM-7). After being corrected for repetition by a low-angle normal fault, the middle tongue has been estimated to be 170 m thick. The upper tongue is 179 m thick here. The middle tongue consists of as many as 40 thin flows ranging from 2 to 10 m in thickness, most are on the order of 6 m thick. Flows near the top of the upper tongue are thicker (15 m) and separated by 2-5-m-thick intervals of crudely bedded, calcite-cemented cinders and ash. The upper tongue contains as many as 35 individual flows.

In both tongues, the massive cores of flows are generally separated by 0.1 to 1.5 m of reddish-brown, vesicular, autobrecciated zones. The cores of flows are dense, dark-gray to grayish-red, aphanitic rocks often exhibiting platy or blocky jointing. Calcite, which is ubiquitous in many outcrops, fills vesicles and irregular fractures.

Petrographic study of 5 samples from the measured sections, and 2 samples collected nearby for chemical analysis, show that most of the lavas are aphanitic basalts or basaltic andesites. They typically contain 5 to 15 percent small (0.2-2 mm) phenocrysts of olivine in various degrees of alteration to a reddish-brown to yellowish-brown, weakly pleochroic, optically homogenous, phase with the properties of iddingsite as described by Brown and Stephen (1959). In many specimens, the ubiquitious small ferromagnesian phenocrysts are completely altered to iddingsite, which may be further altered such that it is commonly rimmed with opaque, reddish-brown hematite (fig. 11 A). This relationship is the apparent cause of the common interpretation of these phenocrysts as hematized pyroxene (orthopyroxene?) (Brown, 1972, p. 53; Chapin and Seager, 1975, p. 308; Osburn, 1978, p. 47). Interpretation of these altered phenocrysts as orthopyroxene probably comes from the distinct cleavage and parallel extinction of the iddingsite. Iddingsite may be distinguished from orthopyroxenes by its deep red color and a low $2V_x$ of 30 to 40 degrees. Several thin sections show relict cores of magnesian olivine with finger-like lamellae of iddingsite extending out into solid rims of iddingsite. In texturally similar crystals, such light colored cores are also commonly altered to vermicular masses of antigorite.

Clinopyroxene, typically fresh in comparison to the olivine, occurs as an abundant groundmass mineral, and often as sparse phenocrysts that are greenish gray in thin section. Universal stage measurements indicate it is mostly augite with a $2V_z$ of 58 to 62 degrees and $Z \wedge C$ of 45 to 47

Figure 11. Photomicrographs of La Jara Peak Basaltic Andesite from the Lemitar Mountains. A, Glomeroporphyritic basaltic andesite (medium power, plane polarized light) with small phenocrysts of plagioclase, An_{48}-An_{56} (light gray), and euhedral olivine (?) completely altered to iddingsite (medium gray with lamellar or micaceous character). Iddingsite is rimmed by opaque hematite (black) formed by secondary alteration. Top of middle tongue (Tba_1, Lemitar Map) 2 km northeast of Strawberry Peak. B, Olivine basalt, 49.7 percent SiO_2, with typical sparse small phenocrysts of olivine partially altered to iddingsite and hematite and a few unaltered augite (arrows) phenocrysts. Pilotaxitic groundmass of labradorite (An_{60}), granular augite, and magnetite. A small patch of calcite (C) and deuteric(?) biotite (B) fill a vug in the matrix. Crossed nicols. Near base of upper tongue (Tba_2, Lemitar Map) same locality as 11A.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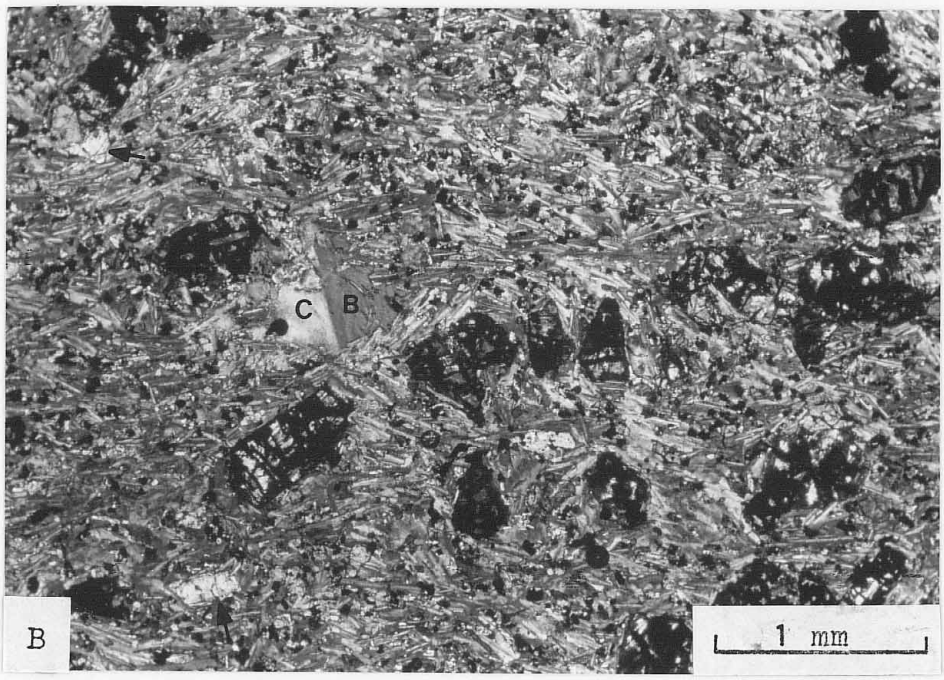

C
B
B
1 mm

degrees. No orthopyroxene was found in any of these thin sections.

The groundmass of all the rocks consists dominantly (50-75 percent) of felted to pilotaxitic plagioclase microlites (An_{45-65}, usually labradorite) with lesser volumes of finely granular augite and very-fine magnetite. Some samples contain groundmass olivine altered to iddingsite. Where clinopyroxene is more abundant, some textures approach intergranular. Xenocrysts of quartz and clouded carbonate jacketed by fibrous masses of augite occur sparsely in two samples.

One lava flow, about 20 m below the top of the middle tongue at the measured section, is atypical in its lithology because of the presence of moderately abundant phenocrystic plagioclase. The plagioclase phenocrysts are normally zoned from sodic labradorite to calcic andesine (An_{55} to An_{45}) and the groundmass plagioclase is andesine. Clinopyroxene phenocrysts are also more abundant than usual and olivine is relatively sparse. The rock has not been chemically analysed but is petrographically similar to andesites ($\sim$ 56 percent SiO_2) of the Spears Formation.

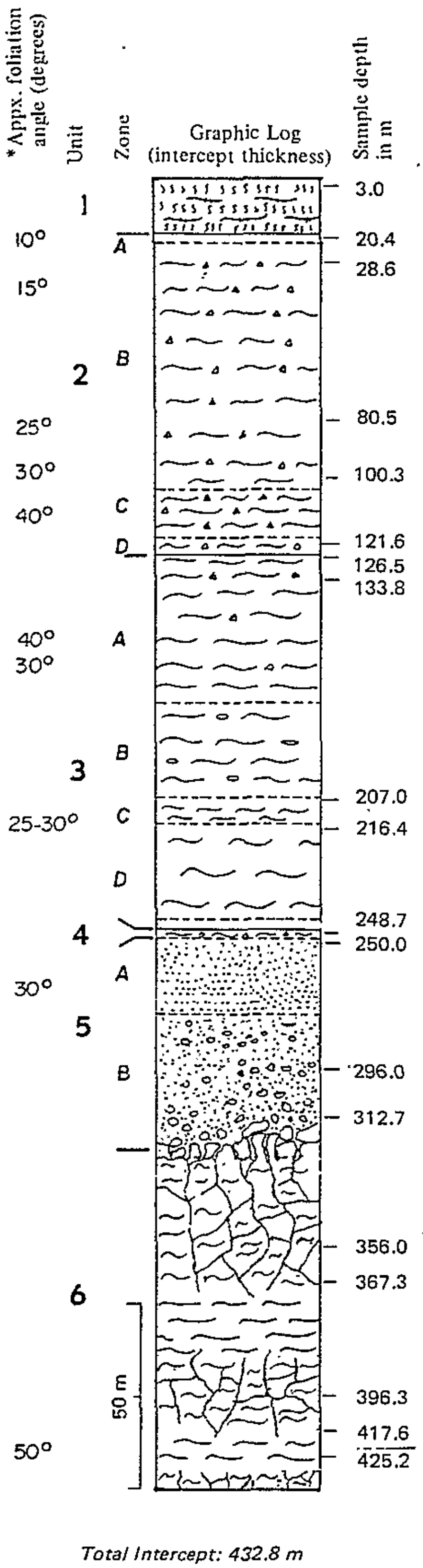
Two samples of La Jara Peak Basaltic Andesite from the Lemitar Mountains yielded analyses of 49.6 and 51.7 percent SiO_2 with 1.7 to 1.9 percent K_2O, respectively (D. L. White, unpub.; see p. 28). Both contain abundant olivine and a few small patches of calcite in association with clear to moderate brown, pleochroic biotite ($2V_x$ = 28 degrees) and traces of greenish chlorite (see fig. 11B). Similar spotty occurrences of calcite and biotite have been found in the middle(?) Miocene basalt of Kelly Ranch and the basalt of Sedillo Hill (4 m.y.). In one case, the patchy coarse-grained biotite clearly has an antipathetic relation to finely distributed groundmass magnetite, thus demonstrating that is is not a primary phenocrystic mineral. Since it is found in basalts ranging from 28 m.y. to 4 m.y. old, this non-phenocrystic biotite is most likely a low(?) temperature deuteric alteration phenomena. This unusual occurrence of deuteric(?) biotite has not been reported in previous studies of the Magdalena Project.

A rhyolitic, reddish-brown, medium-grained sandstone, about 5 m thick occurs at the base of the upper La Jara Peak tongue east of Bug Mountain and has not been differentiated from the La Jara Peak. Osburn (1978) has observed a similar thin sandstone at the same position, overlying the Lemitar Tuff, in the South Canyon area. The mineralogy, texture and stratigraphic position of this sandstone indicate that it was derived from the Lemitar Tuff. A lithologically similar sandstone as much as 25 m thick fills channels cut in the top of the Luis Lopez Formation in the northern Chupadera Mountains. The rhyolitic sandstone is here overlain by the tuff of South Canyon without intervening basaltic andesite. For this reason, the sandstone in the Chupadera Mountains is arbitrarily included in the Luis Lopez Formation (Tlsr, pl. 1).

Sandstones of the Tower Mine Drill Hole (Unit of Sixmile Canyon)

In the early 1950's (?), a diamond-drill exploration hole was driven 433 m at a 60-degree angle under the open cut of the Tower mine manganese vein to test the down-dip projection of this nearly vertical vein. The core from this drill hole is held in storage by the New Mexico Bureau of Mines and Mineral Resources at Socorro under the name "Mountain Copper hole #1". It will be referred to in this report as the "Tower mine drill hole". A log of this core was made by the author and G. R. Osburn in November of 1976 and is summarized here in figure 12.

The location of the collar and surface projection of the Tower mine drill hole are shown on the geologic map (pl. 1) near its southwestern corner. The general stratigraphic and structural interpretation of the drill hole is



Appx. true thickness in m	Unit Descriptions
19.2	*LA JARA PEAK BASALTIC ANDESITE* (upper tongue, Tba_2): red and gray aphanitic basaltic-andesite lavas, vesicular zones, sparse small FeMg phenocrysts altered to hematite.
98.5	*LEMITAR TUFF;* CRYSTAL-RICH ASH-FLOW TUFFS (upper member, 25-45 percent phenocrysts) A. Upper gray, crystal-rich, mod. welded zone (3.7). B. Red, densely welded zone with large qtz. phen. (80.8). C. Red, densely welded, pumiceous (vapor phase minerals in pumice) (16.8). D. Red and gray, mottled, densely welded, pumiceous transition zone (4.6).
NOTE:	— 0.6 m breccia zone with MnO_2 veinlets at 36.6 m; — sheared upper Lemitar Tuff cemented by red jasperoid at 97.5 m depth.
111.9	*LEMITAR TUFF;* CRYSTAL-POOR ASH-FLOW TUFFS (lower member, 5-15 percent phenocrysts) A. Mottled red and gray, pumiceous, mod. crystal content (48.8). B. Mottled red and gray, massive, lithophysal (?) zone (32.0). C. Pur.-gry. pumice in a reddish matrix, mod. welded, mottled small-pumice zone (4.0). D. Lt.-gry., crystal-poor, mod. to poorly welded zone, with dark-gray botryoidal pumice (36.6). E. Chalky, white, basal unwelded zone (2.1).
NOTE:	— Lemitar Tuff is potassium metasomatized. Plagioclase (abundant in upper member) is altered to chalky clay or milky white potassic feldspar. Traces of diamond shaped sphene altered to yellowish leucoxene (?). Sanidine and biotite are fresh.
65.5	*UNIT OF SIXMILE CANYON:* ASH-FLOW TUFF (?): Lithic-rich, qtz.-rich, crystal-rich tuff that may be a boulder of Hells Mesa or may be equivalent to the tuff interval in the unit of Sixmile Canyon (0.6). TUFFACEOUS SANDSTONES AND HETEROLITHIC CONGLOMERATES: Possibly correlative with parts of the unit of Sixmile Canyon. A. Massive, red, med.-grained, tuffaceous sandstone which grades downward into red and gray laminated and cross-laminated sandstones with minor conglomeratic beds (25.6). B. Red, conglomeratic sandstones and conglomerates with clasts as much as 15 cm across. Grades downward into brecciated beds of the underlying tuffs (48.2).
NOTE:	— Sandstones are immature and of rhyolitic to andesitic composition. — Drusy cavity with barite at 254.2 m (Willard, unpub. report). Slickensides at 296.0 m.
93.0	*A-L PEAK TUFF(?);* CRYSTAL-POOR ASH-FLOW TUFFS: Gray and massive to pur.-gray and pumiceous, crystal-poor ash-flow tuffs. Rocks are extensively brecciated (fragments a few mm to few cm in length) to an extent not seen in the overlying Lemitar Tuff. Feldspars, altered to clays (5%); small qtz. phen. (1%), andesite lithics (1%). Since vitroclastic textures are not apparent, these rocks could be massive rhyolite lavas.

* maximum angle between foliation and a plane normal to core axis.

Figure 12. Log of the Tower mine drill hole (NW/4, NW/4 Sec. 7, T4S, R1W). Intercept lengths of units, in meters, given in parentheses. See text for discussion. Revised after Osburn (1978, p. 143).

shown projected to the cross section H-H' of plate 2. This drill hole provides important subsurface control on the stratigraphy and structure within the Socorro cauldron.

Approximately 65 m of volcaniclastic sedimentary rocks were intersected below the unwelded base of the Lemitar Tuff. A densely welded crystal-rich tuff (0.6 m intercept), which occurs at the base of the Lemitar Tuff, is of uncertain correlation. It is most likely a block of older tuff blown out of the vent area during the initial eruption of the Lemitar Tuff or possibly a very thin ash flow equivalent to a crystal-rich tuff found at the top of the unit of Six-mile Canyon near Buck Peak (Osburn, 1978). The sedimentary interval consists of pale-reddish-brown rhyolitic to purplish-gray andesitic sandstones (22 m) that gradually coarsen downward into reddish-brown, muddy, heterolithic volcanic conglomerates about 43 m thick. Clasts as much as 15 cm across in the lower conglomerate are of three distinct lithologies that are similar in appearance to 1) crystal-poor A-L Peak Tuff, 2) porphyritic andesite lavas of the Spears Formation, and 3) quartz-rich, crystal-rich, Hells Mesa Tuff. The basal contact of the conglomerate unit (unit 5B, fig. 12) is gradational and typical of a local erosional unconformity. Angular clasts of crystal-poor rhyolite progressively increase at the expense of the muddy matrix and finally grade into a breccia of gray, massive, crystal-poor rhyolite (A-L Peak Tuff?).

This sedimentary interval and thin cap (?) of moderately crystal-rich tuff between the Lemitar Tuff and an A-L Peak Tuff (correlation tentative, see below) is at the same stratigraphic horizon and partly equivalent lithologically to the "unit of Sixmile Canyon" as defined by Osburn (1978). The middle tongue of La Jara Peak Basaltic Andesite (Tba_1, pl. 1) is contemporaneous with the unit of Sixmile Canyon. Osburn describes the unit of Sixmile Canyon as the heterogeneous volcanic and sedimentary fill of the Sawmill Canyon caldera (fig. 9), which is centered about 15 km to the west of the Tower mine. Mudflow breccias with Hells Mesa Tuff clasts occur near the base of the unit of Sixmile Canyon at the head of South Canyon; and 5 to 15 m of parallel laminated rhyolitic sandstone occur near its top in Sixmile Canyon (Osburn, 1978, pl. 1). The major volume of this 300- to 750-m-thick caldera-fill unit in the northeastern and central Magdalena Mountains consists of basaltic andesite (La Jara Peak type) to porphyritic andesite (Spears type) lavas intertonguing with pale-red to pinkish-gray moderately porphyritic rhyolite lavas and domes (Osburn, 1978, p. 19-31; D. Petty, 1979). If the correlation from the Sixmile Canyon area to the Tower mine hole is correct, then the absence of the thick lava sequence and overall thinning of the unit may indicate that the Tower mine area lies near the buried eastern topographic wall of the Sawmill Canyon caldera. This interpretation would also help explain the anomalously intense brecciation and hydrothermal alteration of the A-L Peak Tuff (?) that occurs as the lowest stratigraphic unit in the Tower mine hole. The brecciation and alteration might then be associated with the eastern segment of the ring-fracture zone of the Sawmill Canyon cauldron block. Alternatively the brecciated rhyolites could represent landslide blocks derived from an over-steepened caldera wall.

The correlation of the brecciated and altered crystal-poor rhyolite that forms the bottom 93 m of the Tower mine hole with a member of the A-L Peak Tuff is somewhat tenuous because of a lack of definite vitroclastic textures. A sample from a depth of 417.6 m was sectioned for petrographic study to check on this problem. The slide contains about 6 percent subhedral feldspar phenocrysts (sanidine ?) about 0.5 to 2.5 mm in length and completely replaced by calcite and sericite. Small rounded and embayed quartz phenocrysts make up another ½ to 1 percent of the rock. The groundmass consists of anhedral blobs (averaging 0.25 mm across) of microcrystaline quartz, feldspar, and amorphous (?) clays. This texture is similar to that found in hornfelsed outcrops of the tuff of Nipple Mountain exposed at the eastern periphery of the Tres Montosas stock, west of Magdalena (Chamberlin, 1974, p. 20). Some relatively non-brecciated intervals of core from this unit exhibit what appear to be rare lenses of light-gray pumice, but glass shards are not visible in any of the samples. The most likely correlation of this unit is with the gray-massive member of the A-L Peak Tuff. The absence of the normally overlying flow-banded and pinnacles (?) members could be explained by their erosion from the postulated eastern wall of the Sawmill Canyon caldera (fig. 9). The only reasonable alternative correlation is with the rhyolite lava member of the unit of Sixmile Canyon which seems unlikely because the core in question contains no visible flow-banding or spherulites typical of these lavas. The brecciated zones in the bottom 93.0 m of core probably do not represent autobrecciated lavas. The angular rhyolite fragments are uniformly fine grained, approximately 2 mm to 2 cm across; and apparently are cemented by minor amounts of silica.

The Tower mine hole did not intersect a major vein as it was apparently intended to do. A slickensided section of sandstone at 296.0 m depth is the most likely potential projection of the Tower mine vein. The absence of a vein may be due to the inability of the relatively plastic sediments to hold an open space along the minor fault. The brittle, densely welded Lemitar Tuff in comparison makes an ideal host for epithermal veins.

Lemitar Tuff

The Lemitar Tuff is named here for numerous hogback forming exposures that occur throughout the central and western Lemitar Mountains (Lemitar Map), approximately 10 km west of the town of Lemitar (fig. 2). The most prominent of these hogbacks forms the crest of the western Lemitar Mountains (SW/4, Sect. 11, T2S, R2W, elev. 6800 feet) about 1.5 km north of the head of Canoncito del Puerticito del Lemitar.

A reference measured section of the Lemitar Tuff, surveyed in the east-central Lemitar Mountains, is presented in Appendix A, section 3. The measured section is summarized graphically in Figure 13. Modal data from this measured section are reported in Table 3 and chemical data in Table 4.

The Lemitar Tuff is a distinctive, compositionally zoned, ash-flow sheet that occurs as a moderately widespread simple-cooling unit in the Socorro-Magdalena area. Three compositional subunits are recognized here consisting of: 1) a basal, crystal-poor to moderately crystal-rich, silicic

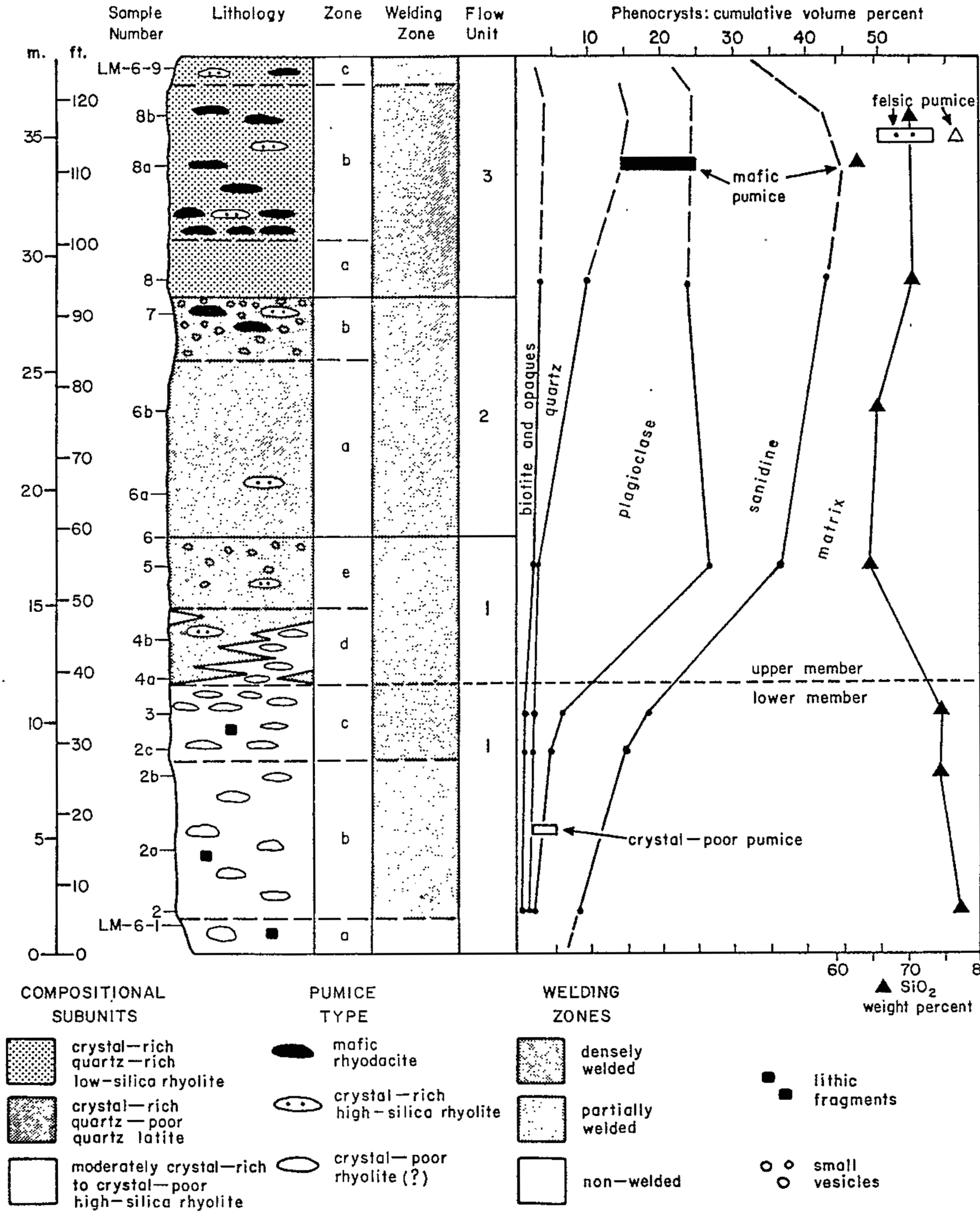


Figure 13. Graphic measured section of the Lemitar Tuff in the east-central Lemitar Mountains showing zonal variations in composition, welding and other parameters. Modal data is from Table 3 and silica data from Table 4. See Appendix A, section 3, for detailed description of individual zones.

Table 3 - Modal data in volume percent for the Lemitar Tuff reference measured section, east-central Lemitar Mountains (SE¼ Sec. 12, T2S, R2W)

Height above base in m	Sample No.	Total Pheno-crysts	Sani-dine	Plagio-clase[2]	Quartz	Biotite	Opaques	Points Counted	Pumice[5]	Lithics[5]
28.9	LM-6-8[1]	43.0	15.1	18.0	6.3	2.8	0.8	2792	tr.	0
16.4	LM-6-5[1]	36.6	10.2	23.5[3]	0.3	1.9	0.7	2395	½-1	0
upper member										
lower member										
10.4	LM-6-3[1]	18.1	11.8	3.8	1.5	1.0	0.03	2740	5-8	tr.
8.8	LM-6-2c	15.2	10.5	2.6	1.0	0.5	0.6	2579	3-5	0
1.8	LM-6-2	8.4	6.2	0.4	1.4	0.2	0.2	6941[4]	2-4	<½

[1] analysis by G. R. Osburn

[2] Plagioclase entirely altered to potassic clay and feldspar, occurs as washed-out holes and skeletal remnants

[3] may include very small vesicles (0.2-0.5 mm) that comprise 1-3% of this rock

[4] 1/3 x 1/3 mm grid used versus standard grid of 1/3 x 2/3 mm

[5] estimated from hand specimen and outcrop

Table 4 - Chemical data in weight percent for the Lemitar Tuff reference measured section, east-central Lemitar Mountains (SE¼ Sect. 12, T2S, R2W). Analyses by D. L. White using x-ray fluorescence methods at the New Mexico Bureau of Mines and Mineral Resources. (See p. 28)

	Height above base in m	Sample No.	SiO_2	Al_2O_3	Fe_2O_3 (total)	MgO	CaO	Na_2O	K_2O[1]	TiO_2	TOTAL
low-silica rhyolite	36.0	LM-6-8b	70.13	15.75	2.71	0.78	0.56	1.97	9.46	0.57	101.93
	33.8	LM-6-8a (mafic pumice)	63.16	18.28	3.88	0.70	1.06	1.57	11.57	0.70	100.92
	28.9	LM-6-8	70.88	16.15	2.26	1.16	0.72	1.54	8.67	0.50	101.88
quartz latite	23.5	LM-6-6b	65.85	15.47	2.78	0.71	1.73	1.79	9.55	0.57	98.45
	16.4	LM-6-5	64.47	17.89	3.11	0.38	0.81	2.75	10.63	0.61	100.65
	upper member										
	lower member										
high-silica rhyolite	10.4	LM-6-3	74.73	14.86	2.42	0.63	0.58	2.63	5.80	0.35	102.00
	7.6	LM-6-2b	74.02	14.54	1.01	0.71	0.57	2.22	8.12	0.29	101.48
	1.8	LM-6-2	77.13	13.34	2.02	0.08	0.73	1.98	7.94	0.25	103.47
		PR-1-77b[2] (felsic pumice)	76.51	12.75	1.88	<0.01	0.53	3.63	5.97	0.32	100.60

1 abnormally high K_2O contents reflect potassium metasomatism; plagioclase is entirely altered to potassic clay and feldspar in all of these samples

2 felsic crystal-rich pumice from uppermost quartz-rich zone of the Lemitar Tuff in the eastern Magdalena Mountains (Osburn, 1978, p. 145) petrologically similar felsic pumice occur in association with the mafic pumice (LM-6-8a) at the measured section.

rhyolite, 2) a medial, crystal-rich, quartz-poor, quartz latite, and 3) an upper, crystal-rich, quartz-rich, rhyolite. For mapping purposes, the middle and upper subunits are arbitrarily grouped together as the crystal-rich "upper member" (Tlu, pl. 1) of the Lemitar Tuff and the lower subunit is mapped separately as the crystal-poor "lower-member" (Tll, pl. 1) of the Lemitar Tuff. In the Lemitar Mountains, the approximate location of the contact between members is easily mapped, since the upper member forms a distinctive dark reddish caprock that is welded to the mostly light-gray lower member. A more exact definition of this gradational contact, for correlation purposes, is described in the section titled "outflow facies". The sheet-like geometry of the upper member contrasts with the locally lenticular and generally less extensive lower member because of greater paleotopographic control on the distribution of the lower member.

In prior investigations of the Magdalena Project, the equivalent of the upper member of the Lemitar Tuff has been mapped as the "tuff of Allen Well" (Blakestad, 1976; Brown, 1971; Simon, 1973; and Spradlin, 1976) and the "tuff of La Jencia Creek" (Brown, 1971; Simon, 1973); these stratigraphic terms have now been abandoned by members of the project. The lower member of the Lemitar Tuff where observed by some earlier investigators was erroneously correlated with various members of the A-L Peak Tuff: "gray-massive member" as used by Simon (1973, p. 18) and the "pumiceous member" (pinnacles(?) member) as used by Spradlin (1976, p. 41).

In the type area, the Lemitar Tuff is normally interbedded between the middle and upper tongues of the La Jara Peak Basaltic Andesite. This "sandwich" of basaltic lavas, with Lemitar Tuff as the filling, rests on the pinnacles(?) member of the A-L Peak Tuff and is locally overlain by the tuff of South Canyon. North of the eroded edge of the South Canyon ash-flow sheet, the "sandwich" is unconformably overlain by the lower Popotosa Formation.

Exposures in the northern Magdalena Mountains, southern Bear Mountains and Joyita Hills, which are now reassigned to the outflow facies Lemitar tuff, have stratigraphic relationships essentially identical to those described above for the type area. In the central and eastern Magdalena Mountains the Lemitar Tuff overlies complexly intertonguing andesite to rhyolite lavas and volcaniclasitic rocks that filled the North Baldy and Sawmill Canyon calderas (fig. 9) (Osburn, 1978; Petty, 1979; and Bowring, in prep.).

The preferred K-Ar radiometric age of the Lemitar Tuff is 27.9 m.y. Biotite separates taken from samples of unaltered crystal-rich tuff collected at Monica Canyon in the northern San Mateo Mountains, and at Rosa del Castillo Arroyo in the Joyita Hills, have been dated respectively at 27.0 ± 1.1 m.y. and 28.8 ± 0.7 m.y. (Appendix B). The preferred age is the numerical average of these dates that have overlapping error bars. A biotite separate from a potassium metasomatized sample of the Lemitar Tuff collected in the type area has been dated at 26.3 ± 1.0 m.y. (Appendix B); this date may be slightly young.

Available data on the thickness and distribution of the Lemitar Tuff, summarized in Figure 14, indicate that the thickest sections of the ash-flow sheet, reasonably interpreted as cauldron-facies tuff, occur in the eastern Magdalena Mountains and the adjacent Chupadera Mountains. These areas lie within the Socorro cauldron, which has been proposed as the source of the Lemitar Tuff (Chapin and others, 1978). Some aspects of this interpretation are problematical, namely: 1) apparent subsidence of 150 m along the western margin is relatively minor, 2) apparent thickness changes of cauldron-facies tuffs in the northern Chupadera Mountains from the Tower Mine area (210 m) to the Black Canyon area (greater than 880 m) are unusually abrupt and large, and 3) the correlation of the 880-m-thick

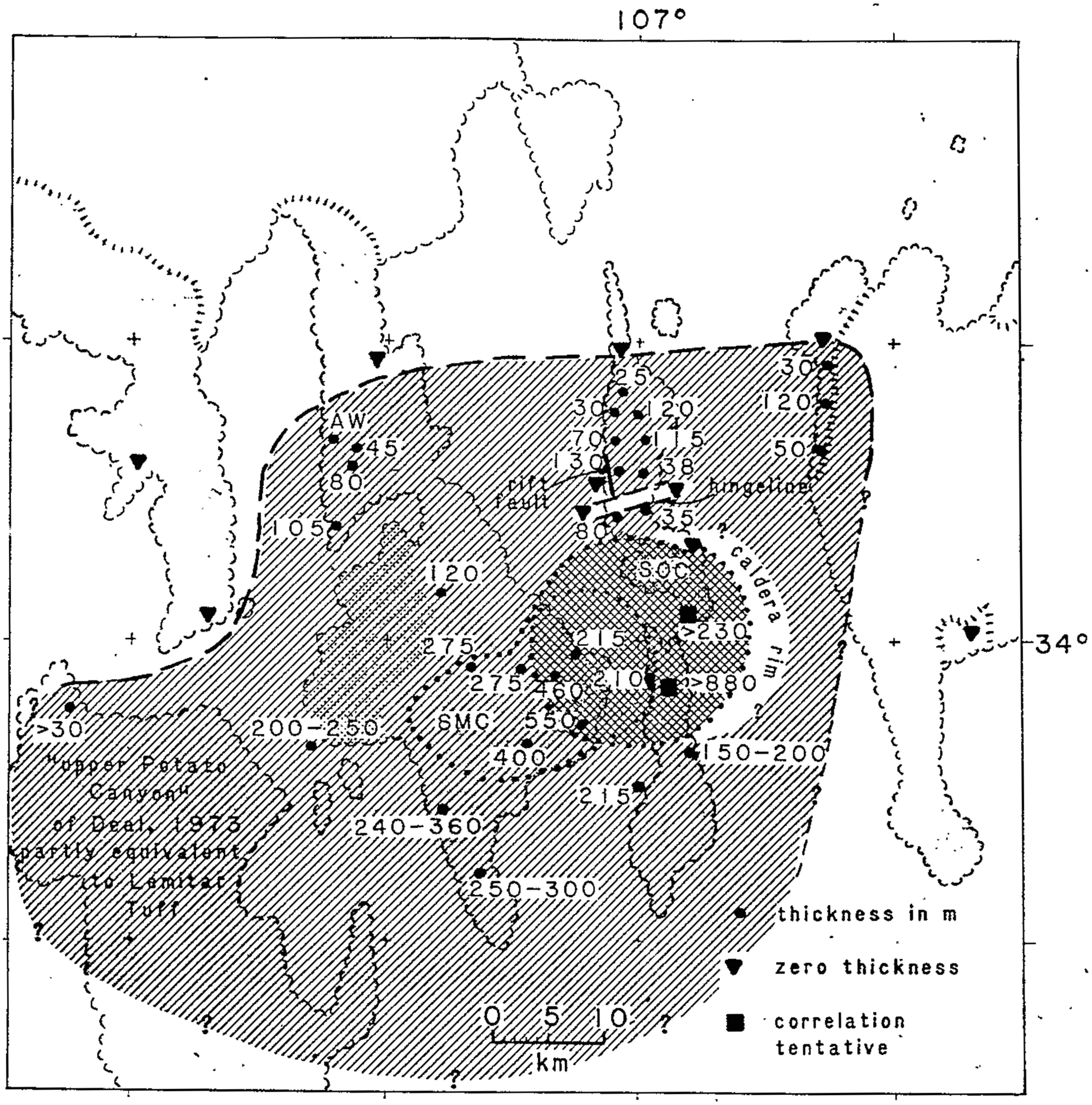


Figure 14. Outcrop control points, approximate thicknesses, and the inferred original extent of the Lemitar Tuff outflow sheet (lined). Cross hatched area within the Socorro cauldron (SOC) is interpreted as cauldron-facies tuff. Smaller dots represent the projected eastern margin of the Sawmill Canyon cauldron (SMC), which is buried by the Lemitar Tuff. Control point data are from Blakestad, 1976; Bowring, in prep.; Brown, 1972; Osburn, 1978, 1979, oral commun.; Petty, 1979; Simon, 1973; Spradlin, 1976; Wilkinson, 1976; this report and the "Lemitar Map".

cauldron-facies tuff southeast of Black Canyon with the Lemitar Tuff must be considered tenative and, alternatively, it could be cauldron-facies Hells Mesa Tuff (G.R. Osburn, oral commun., 1979). The first two problems are reasonably explained by interpreting the Socorro cauldron as: 1) a trap-door cauldron (Lipman and Steven, 1976, p. 31) hinged along its western margin (Osburn, 1978; Petty, 1979); and 2) as a cauldron formed concurrently with regional extensional stress such that preexisting, north-trending, early-rift faults (Chamberlin, 1976, 1978) allowed differential subsidence of the cauldron block during eruption of the Lemitar ash flows. Cauldron-facies tuffs southeast of Black Canyon, that are crystal-rich, quartz-rich and locally lithic-rich may be interpreted as the equivalent of the upper quartz-rich rhyolite subunit of the Lemitar outflow facies (Fig. 13). This anomously large thickness of the upper rhyolite subunit may have been caused by late-stage subsidence of the eastern half of the cauldron block along a north-trending early-rift fault. Apparently the greatest amount of subsidence, which typically occurs concurrent with all major ash-flow eruptions (Steven and Lipman, 1976, p. 31), takes place in the final stage of the eruptive cycle after the underlying magma chamber has been nearly emptied (Lipman, oral commun., 1978). The Socorro cauldron may be similar in this respect to the Long Valley caldera of California. In this well-documented caldera, the cauldron-facies tuff and younger fill deposits thicken abruptly, by as much as a kilometer, where a major range-bounding normal fault projects across the cauldron block (Bailey, 1976, p. 731).

Outflow Facies

As much as 80 m of outflow-facies Lemitar Tuff forms a low hogback in the northwestern corner of the study area near Bug Mountain. Both the lower and upper members are exposed here. The lower member, 45 m thick at the southern end of the outcrop, wedges out about 500 m to the north against a paleotopographic scarp related to a hingeline fault zone that is peripheral to the Socorro cauldron. The upper member here is relatively uniform in thickness at about 35 m. It is truncated by an erosional unconformity at the base of the Popotosa formation where it is sharply upturned along the hingline structure.

The reader is referred to Appendix A, section 3, for detailed descriptions of zonal variations that are generally applicable to the Lemitar Tuff outflow sheet throughout the Lemitar Mountains, including the Bug Mountain area. The following descriptions of the lower and upper members of the Lemitar Tuff on the outflow sheet are from petrographic study of samples from the measured section and general observations in the area.

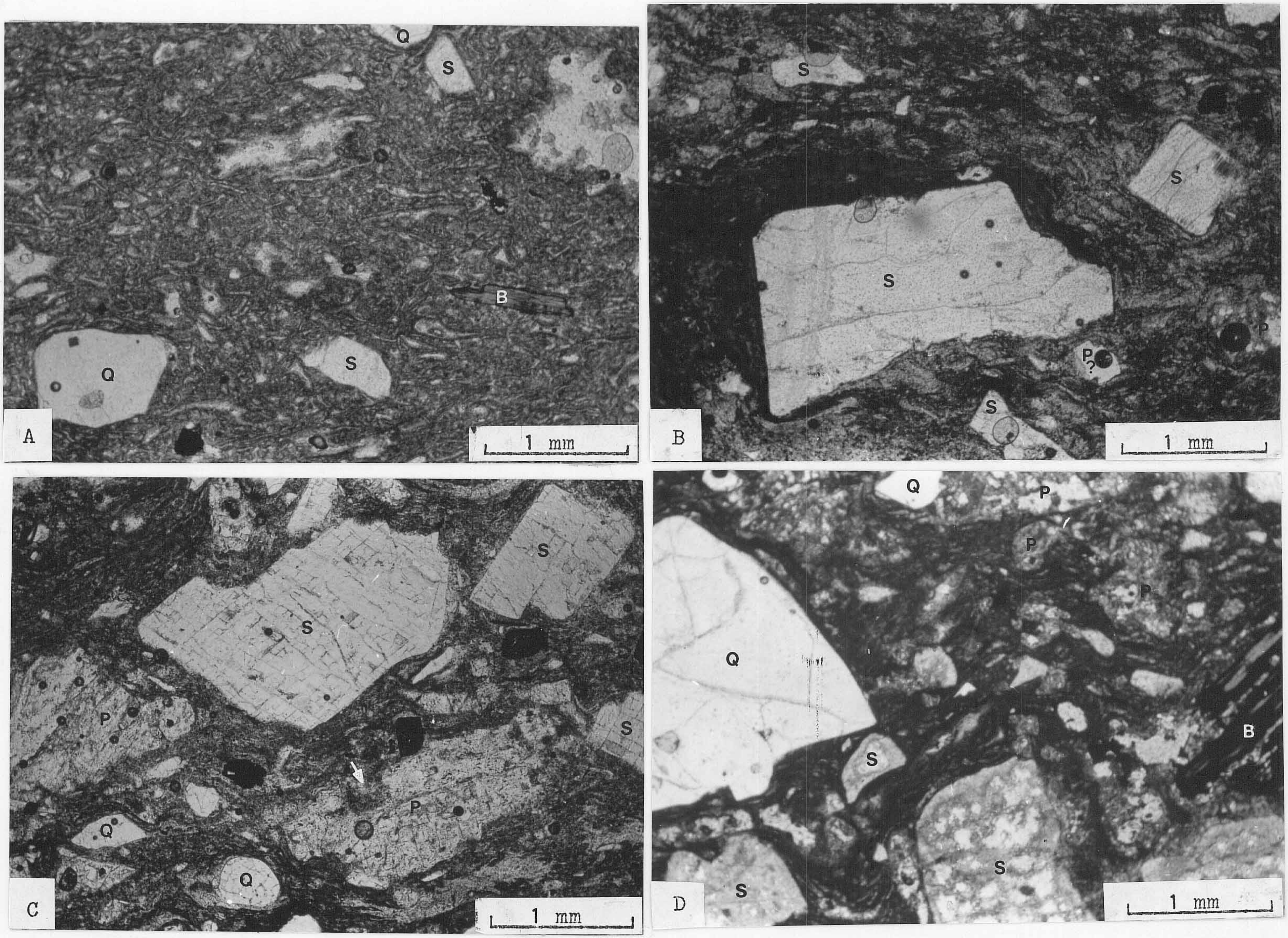
Lower member. Zonal variations and petrographic data for the lower member of the Lemitar Tuff have been described for a measured section in the eastern Magdalena Mountains (Osburn, 1978, p. 36) and for the "Crouch drill hole" of the Silver Hill area west of Magdalena (Simon, 1973, p. 21-22; his "gray-massive member"). These data are generally in good agreement with observations at the measured section in the type area of the Lemitar Mountains.

At the reference section, the lower member consists of 11.6 m of crystal-poor to moderately crystal-rich, high silica rhyolite. Total phenocryst content increases upwards, from 8 percent near the base, to about 18 percent at the top. Silica content decreases from 77 percent to 74 percent over the same interval.

It is important to note that the entire section of Lemitar Tuff at the measured section, and generally throughout the study area, is potassium metasomatized. Recognition of this hydrothermal alteration raises some question as to how representative the chemical analyses may be with respect to the SiO_2 content. Clearly considerable amounts of K_2O have been added and Na_2O removed from these rocks (Table 4; Chapin and others, 1978). Except for the one sample nearest to the base of the ash-flow sheet (LM-6-2), virtually every thin section shows some evidence of secondary silica, mostly as vapor-phase filling of pumice and void spaces. Some small discontinuous silica veinlets occur subparallel to the foliation, in an en echelon pattern, in samples LM-6-4b and LM-6-5. These veinlets cut across fresh sanidine phenocrysts and are continuous across "holes" formed where plagioclase phenocrysts have been altered to clay minerals, and then washed from the thin section during its preparation. This relationship suggests that the silica veinlets predate the potassium metasomatism. In spite of these observations, the silica content of these samples are considered to be reasonably representative of the original rocks because the observed silica variations do faithfully reflect the mineralogical variations in the different compositional zones. The tuff sample with the highest SiO_2 content (LM-6-2) has about 5 percent open pore space and contains no apparent secondary quartz. In contrast the sample with the lowest SiO_2 content (LM-6-5) has small silica veinlets that cut across the densely welded matrix and phenocrysts. These relations support the interpretation that the SiO_2 variations apparent in Table 4 and Figure 13 actually reflect primary chemical zonation of the ash-flow sheet and are not controlled by hydrothermal alteration.

Phenocryst mineralogy is essentially the same throughout the lower member; in approximate decreasing order of abundance, sanidine, quartz, biotite, plagioclase, magnetite, ilmenite(?) and traces of sphene are typically present. Plagioclase phenocrysts are consistently altered to potassic clay and feldspar, whereas sphene is altered to leucoxene. The selective alteration of only calcic minerals appears to be distinctive of potassium metasomatism. Chalky- white altered plagioclase phenocrysts are highly visible in hand specimen but only sparsely present throughout the lower member. Quartz/sanidine ratios are similar to the A-L Peak tuffs (Table 2). Traces of small coppery biotite flakes are very common in handspecimens of the lower Lemitar Tuff, and readily distinguish them from the A-L Peak tuffs, which usually lack biotite. Sanidine is slightly argillized in some samples from the lower member. Vitroclastic textures are well preserved in the lower member and are particularly visible in sections stained with sodium cobaltinitrite (fig. 15A). Vapor-phase quartz first appears in the pumice of sample LM-6-2b and increases in abundance upward toward the contact with the upper member. The lower member is generally pumiceous containing about 3 to 10 percent of crystal-poor pumice. The crystal-poor pumice are assumed to be similar in chemical composition to the matrix

Figure 15. Photomicrographs in plane polarized light of the Lemitar Tuff from the reference measured section and a nearby area in the Lemitar Mountains. The series illustrates variation in phenocryst content and matrix appearance for the three compositional subunits (fig. 13). All thin sections contain phenocrysts of sanidine (S), altered plagioclase (P), quartz (Q), biotite (B or black), and opaques (black) and except for "C" have been stained for potassium with sodium cobaltinitrite. Not all phenocrystic minerals are visible in each photomicrograph. Small, high-relief bubbles are artifacts. A, Sample LM-6-2, partially welded, crystal-poor, high-silica rhyolite. Note shard outlines are potassium-rich and enhanced by staining; unstained cores of shards probably represent crypotocrystalline silica. B, Sample LM-6-3, densely welded, moderately crystal-rich, silicic rhyolite. Note the dark, fluidal matrix adhered to the large sanidine phenocryst similar to matrix of the quartz latite in Figure 15C. Also note the lighter colored rhyolitic matrix containing extremely flattened but distinctly visible shards. C, Sample 76-1-10 (NW/4, Sec. 2, T2S, R2W), approximately equivalent to LM-6-5, densely welded, crystal-rich (35 percent), quartz-poor (1 percent) quartz latite (69 percent SiO_2). Note rounded and embayed "plagioclase" phenocrysts (arrow) replaced by clouded, inclusion-rich, potassium feldspar (compare with fig. 20B and 20C). D, Sample LM-6-8, densely welded, crystal-rich, quartz-rich, rhyolite. Dark, densely welded matrix appears homogeneous in contrast to pumiceous mixture found higher in this unit. Light colored areas in plagioclase are washed out holes once filled with clay minerals.


A
Q
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1 mm
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of the tuff which has an equivalent crystal content. Based on the degree of compaction of pumice and outcrop character, the lower member is divisable into: 1) a non-welded basal zone that is only rarely exposed, 2) a light-gray partially welded zone, and 3) a medium-gray to pale-red densely welded zone. The red interval of the densely welded zone is distinctly streaked with abundant pinkish-orange pumice. Pumice that has a botryoidal appearance, visible as light-gray colored bubbles in a darker gray framework, is only poorly developed in a thin interval near the base of the partially welded zone. Similar botryoidal pumice is found in the basal A-L Peak Tuff and the base of the tuff of South Canyon.

Lithic fragments consisting of: porcelaneous gray crystal-poor rhyolite, possibly a welded tuff, and reddish-brown, andesitic porphyry occur only in the lower member. They are quite sparse, forming less than a fraction of a percent of the outcrop; this contrasts significantly with the same interval in the eastern Magdalena Mountains, which commonly contains 5-10 percent of similar composition lithics (Osburn, 1978). This relationship is in general accordance with the observation that cauldron-facies tuffs typically have a higher lithic content (Lipman, 1978, oral commun.).

The contact between the lower and upper members of the Lemitar Tuff has generally been described as a gradational densely welded contact. It is associated with a transition zone ranging from about 1 m thick in the Silver Hill area west of Magdalena (Simon, 1973, p. 19) to about 15 m thick in the cauldron-facies tuff at South-Canyon in the eastern Magdalena Mountains (Osburn, 1978, p. 41). Based on the modal data of Simon, Osburn, and this report, the contact between the lower and upper members of the Lemitar Tuff should be placed at the transition zone across which the total phenocryst content increases abruptly from 10-15 percent to 25-30 percent. Associated with the upward increase in phenocryst content are: 1) a sharp decrease in pumice content, 2) a sharp drop in the quartz/feldspar ratio, 3) an increase in the plagioclase/sanidine ratio, and 4) a decrease in the silica content (see tables 2, 3, 4 and figure 13).

At the reference section in the Lemitar Mountains, the contact between members is more gradational than typically observed elsewhere in the Lemitar Mountains and at the above mentioned localities. It consists of a three-meter-thick zone of large intertonguing streaks, up to 10 m in length, formed by a physical mixture of lighter colored, pumiceous rhyolite (typical lithology of the top of the lower member) and darker colored, crystal-rich, quartz-poor, pumice-poor, quartz latite (typical of the lower portion of the upper member). A small scale illustration of the physical mixture of quartz latite and rhyolite magmas is shown in Figure 15B. At the reference section the contact has been arbitrarily placed at the first megascopic appearance of the typical dark-red quartz latite tuff. Figure 15B shows that the appearance of quartz latite blebs on a sub-mesoscopic scale actually occurs a few meters below the megascopic contact (sample LM-6-3).

Upper member. At the reference section, the upper Lemitar Tuff forms a caprock 26.8 m thick. It consists of dark-reddish, mostly densely welded, crystal-rich, ash-flow tuff. The upper member is primarily divisable into two compositional subunits -- a lower, quartz-poor, quartz latite averaging about 65 percent SiO_2; and an upper, quartz-rich, low-silica rhyolite with a bulk composition of 70 to 71 percent SiO_2 (fig. 13).

The compositional subunits are formed by three ash-flow depositional units that are defined by sharp welded contacts above two reddened (oxidized) and finely vesicular zones. Vesicles in these zones are very (0.1 to 0.5 mm) and may form about 1 to 5 percent of the rock. Some vesicles are partially to completely filled with polycrystalline vapor-phase quartz that in some instances exhibit multiple crystal-growth surfaces. These vesicles may be incipient lithophysae formed by partial degassing prior to burial by the overlying densely welded flow unit.

Within this framework of compositional and flow units, the upper Lemitar is further divisable into seven zones defined by interrelated changes in color, outcrop character, pumice content and degree of welding. These zone boundaries are gradational except where coincident with flow-unit boundaries.

Variations in welding observed in outcrop and checked in thin section indicate that the Lemitar Tuff at the reference section is essentially a simple cooling unit. The vesicular nature of zone b, in flow-unit 2 (fig. 13), makes it less resistant to weathering. Thus, in the outcrop, it looks like a partially welded zone, but in thin section it is clearly as densely welded as the overlying and underlying zones. The partially welded zone at the top of the Lemitar Tuff is commonly absent, apparently because of slight erosion of the Lemitar sheet prior to eruption of the overlying La Jara Peak lavas.

Routine petrographic study of all samples shown in Figure 13, indicate that modal analyses of the quartz latite (LM-6-5) and the upper rhyolite (LM-6-8) reported in Table

3 are reasonably representative and may be used to characterize the subunits. An exception may be with regard to the anomalously high plagioclase content of sample LM-6-5. Analyses reported by Simon (1973, p. 21) and Osburn (1978, p. 36) for approximately equivalent rocks indicate subequal volumes of sanidine and plagioclase, but with sanidine generally more abundant. In samples from the reference section, plagioclase is entirely altered to potassic clays and feldspar, which has been washed out and plucked from the thin sections during preparation. All such "holes" in sample LM-6-5 were counted as plagioclase, some of these may have been sanidine phenocrysts that were plucked out, and some may be the small vesicles that are locally present in this zone.

The quartz latite compositional unit is characterized by abundant and subequal amounts of plagioclase and sanidine along with minor amounts of quartz, biotite and opaques. Sanidine is mostly fresh and euhedral and ranges from 0.5 to 1.5 mm in length. Some sanidines show partial alteration to clay (kaolinite?) along fingerlike lamellae parallel to cleavage traces. Outlines of altered plagioclase phenocrysts are usually rounded suggesting marked magmatic resorbtion at the time of eruption. In thin section minor amounts of moderately birefringent clay (identified by x-ray diffraction as illite, J.A. Cima Osburn, 1978, oral commun.,) is commonly visible in some cores of altered plagioclase where not completely washed out during preparation of the section. Outlines of the altered plagioclase usually consist of irregular remnants of potassium feldspar, which stains strongly with sodium cobaltinitrite.

In one thin section (fig. 15c, sample no. 76-1-10), the replacement of plagioclase by potassium feldspar is unusually complete. This section was chosen for a universal stage examination in order to compare the optic axial angles (2V) of the primary alkali feldspar phenocrysts (sanidine) with those of the potassium feldspar formed by replacement of plagioclase during potassium metasomatism. Fifteen primary sanidine phenocrysts, distinguished by their clear homogeneous extinction and euhedral character, yielded values of $2V_x$ (rounded to nearest degree) of: 20, 20, 24, 28, 30, 31, 32, 34, 34, 36, 37, 38, 40, 42, and 48 degrees. In comparison, thirteen replacement-type potassium feldspars; characterized by clouded inhomogeneous extinction, numerous inclusions, and diffuse interference figures, yielded values of $2V_x$ of: 31, 32, 32, 36, 36, 37, 38, 38, 39, 42, and 42 degrees. Thus it appears that the metasomatic potassium feldspar has optic axial angles which fit either the sanidine-anorthoclase series ($2V_x$ = 18 to 54 degrees) or the orthoclase-orthoclase microperthite series ($2V_x$ = 33 to 103 degrees), but not the microcline-microcline perthite series ($2V_x$ = 66 to 103 degrees) as reported by Deer, Howie and Zussman (1966, p. 285). Observations in the hot springs area of Steamboat Springs, Nevada (Schoen and White, 1965) indicate that the hydrothermal potassium feldspar associated with the springs is monoclinic orthoclase; this may also be true of metasomatic feldspars in the Lemitar Mountains and the Socorro Peak area.

Phenocrystic quartz is typically sparse in the quartz latite subunit, ranging from about 1to 4 percent by volume. Sparse white, crystal-rich (rhyolite?) pumice that are generally disseminated throughout the quartz-latite unit appear to contain relatively more quartz than the host rock. In the quartz latite, biotite is generally not as oxidized as in the lower crystal-poor rhyolite or the upper crystal-rich rhyolite. Sphene, a common trace mineral in the rhyolite subunits, is rare in the quartz latite, except in association with the crystal-rich pumice.

The matrix of the quartz latite tuff is a densely welded dark-reddish-brown glass(?). It is distinctly fluidal and homogeneous in appearance in comparison with the equally densely welded lower rhyolitic tuff, which exhibits numerous shard outlines (see fig. 15B). The red homogeneous matrix is not clearly isotropic under crossed nicols and may be cryptocrystalline. The fluidal appearance of the quartz latite matrix suggests that it may be a primary welded tuff. S.A. Bowring (in prep.) has observed lineated gas pockets in the quartz latite horizon of the Lemitar Tuff in the western Magdalena Mountains, which supports the primary welding interpretation. The possibility of a widespread flow lineation in the Lemitar Tuff should be a subject for further study.

A single, anomalously fresh, sample (LM-6A-2) of the quartz latite subunit was collected in the southeastern Lemitar Mountains. In this rock, plagioclase and sanidine are subequal in volume with the former slightly more abundant. The plagioclase is strongly resorbed and often embayed by the glassy matrix. Universal stage measurements, indicate the plagioclase is calcic oligoclase to sodic andesine (An_{25} to An_{41}, seven measurements, Ritmann zone method). Traces of clinopyroxene are present in this sample, some of which exhibit a magmatic reaction relationship with biotite. One such crystal was identified with the universal stage as ferroaugite ($2V_z$ = 58 degrees; $Z \wedge C$ = 45 degrees). The apparent high iron content is supported by the reaction relationship. Traces of very small euhedral zircon and apatite are also present in this sample.

This sample also has an anomalous occurrence. It was taken from a densely welded lens about 0.5 m thick in a non-welded zone of crystal-rich Lemitar Tuff about 10 m thick that occurs just north of a south-facing paleoscarp, similar to the one at Bug Mountain (Lemitar Map; SE ¼, Sec. 24, T2S, R2W). A complete and hydrothermally altered section of Lemitar Tuff occurs on the south side of the paleoscarp.

The low-silica rhyolite subunit forms approximately the uppermost one-quarter of the ash-flow sheet at the reference section. The base of this subunit is a sharp, welded, flow-unit boundary. Compositional changes upwards across this boundary are: 1) an increase in total phenocryst content from about 35 percent to 45-50 percent, 2) an increase in quartz phenocrysts from 1-2 percent to 5-15 percent and 3) a local decrease in pumice content. The pumice-poor, quartz-rich base of this subunit (fig. 13; unit 3, zone a) is petrographically indistinguishable from the Hells Mesa Tuff. The occurrence of abundant mafic rhyodacite pumice in the top two-thirds of the quartz-rich, crystal-rich Lemitar Tuff may be a possible means for distinction between what is potentially cauldron-facies Lemitar Tuff or Hells Mesa Tuff in the northern Chupadera Mountains.

Most of the upper rhyolite subunit appears to be a physical mixture of rhyodacitic and rhyolitic magmas. This is indicated by the presence of abundant rhyodacite pumice (63 percent SiO_2, Table 4, LM-6-8a) and sparse high-silica rhyolite pumice (76 percent SiO_2, Table 4, PR-1-77b) and a bulk composition of low-silica rhyolite (70-71 percent SiO_2; Table 4, LM-6-8 and LM-6-8b). Osburn (1978, p. 43) has observed the same pumice association in the eastern Magdalena Mountains to vary laterally such that one pumice type locally becomes dominant over the other. Lipman and others (1976) have observed mixtures of pumice, different in composition from that of the host, in a compositionally zoned ash-flow sheet in southern Nevada.

The rhyodacite pumice is grayish-red, moderately crystal-rich and occurs in swarms forming as much as 5 to 10 percent of the outcrop. They contain 15-25 percent phenocrysts, dominantly altered plagioclase with minor biotite and traces of quartz and sanidine. The matrix of the mafic rhyodacite pumice is a fluidal-looking, reddish glass (cryptocrystalline?) which is locally scoriaceous. At the reference section in the Lemitar Mountains, the largest mafic pumice are 15 cm long. In the Rincon Madera Canyon of the eastern Magdalena Mountains, the writer has observed these mafic pumice to be as much a 1 meter in length, suggesting that they are closer to their source in this area. Similar but less abundant mafic pumice also occur near the top of the quartz-latite subunit (fig. 13).

The high-silica rhyolite pumice is light-gray to white, very crystal-rich (roughly 50 to 60 percent phenocrysts), and occurs as elliptical vuggy pods of relatively large euhedral crystals from which the matrix material preferentially weathers out. Total phenocryst contents are difficult to estimate because of their weathering character, their small size -- generally less than 3 cm in length, and their sparse occurence -- forming about 1 percent of the outcrop (unit 3, zone b and c, fig. 13). Sanidine forms about 50 percent of the phenocrysts followed by large (1-4 mm) quartz phenocrysts (25-30 percent) and altered plagioclase (10-15 percent). Coppery biotite and traces of sphene may also be present. The matrix, largely replaced by vapor-phase quartz, appears to consist of spherulitic quartz and feldspar. Similar white pumice, with relatively abundant quartz, occur as small, 1 cm, wisps in the quartz latite. They form significantly less than one percent of the quartz latite outcrops.

The bulk composition of the low-silica rhyolite subunit is modally and chemically between that of the associated pumice types (fig. 13). Phenocrysts generally form 40 to 45 percent of the tuff where it is densely welded. Sanidine is dominant, followed by plagioclase, quartz and biotite.

The phenocryst mineralogy of sample LM-6-8 (fig. 15D) is consistent with a physical mixture interpretation, but the dark hematitic matrix appears to be relatively homogeneous in comparison to figure 15B. Essentially complete mixing of phenocrysts and magmatic liquids may have occurred in the basal interval of the upper rhyolite. An apparent decrease in total phenocryst content near the top of the Lemitar Tuff (fig. 13, Simon, 1973, p. 21) may be related to a decrease in the degree of welding or to winnowing of phenocrysts from the upper portions of the ash flow.

Cauldron Facies

As part of a study of the Luis Lopez manganese district, Miesch (1956) published a geologic map which shows the core of northern Chupadera Mountains -- about 25 km^2, from Black Canyon on the north to Walnut Creek (Nogal Canyon) on the south -- to consist of a "massive rhyolite". Miesch (p. 11) described the "massive rhyolite" as "highly porphyritic with phenocrysts of sanidine, quartz, biotite ... and mostly altered plagioclase." Miesch considered the "massive rhyolite" to be a domal intrusive in the Black Canyon area and a lava flow further to the south. However, he was clearly uneasy with this interpretation as indicated by the statement (Ibid, p. 13):

> "The fragmental nature of the phenocrysts in the massive rhyolite and the absence of flow structures suggest the possibility of a pyroclastic origin. The groundmass, however, is not at all fragmental ..."

The "massive rhyolite" of Miesch is here reinterpreted to be a thick pile of densely-welded cauldron-facies ash-flow tuff, the eruption of which resulted in the collapse of the Socorro cauldron. This cauldron-facies tuff is tentatively correlated with the crystal-rich and quartz-rich zone of the upper Lemitar Tuff, as defined on the outflow sheet.

Two categories of cauldron-facies tuff are recognized in the study area on the basis of stratigraphic relationships, compositional zonation, outcrop characteristics and geographic location. Exposures at the Tower mine block form one category. This roughly triangular block, about 1.5 km on a side, is bound on the east side by a north-northeast-trending fault that extends from the Black Canyon mine northward almost to Chupadera Cliff (pl. 1). The north side is formed by a west-northwest trending fault, which juxtaposes the tuff of South Canyon against upper Lemitar Tuff; and by an unconformity along this trend where Santa Fe Group sediments and lavas lap southward onto the block. A fault-line escarpment bounding the west side of the Chupadera Range forms the remaining side of the triangle.

Exposures on the Tower mine block mapped as upper Lemitar Tuff (pl. 1) are not of questionable correlation. Zonal variations observed in the Tower mine drill hole (fig. 11) are essentially identical to those on the outflow sheet (fig. 13) and confirm this correlation. The overlying stratigraphic sequence of the La Jara Peak Basaltic Andesite (Tba_2, pl. 1) and tuff of South Canyon (Tsc, pl. 1) also mimics the outflow-facies relationships.

Most of the exposures on the Tower mine block appear to be equivalent to the uppermost quartz-rich zone of the outflow sheet. Abundant mafic pumice occurs in the top 20 m of the Tower hole and in outcrops near the Black Canyon mine. Outcrops of the crystal-rich, quartz-poor subunit occur in the area between the Nancy mine and the central shaft of the Tower mine open cut. Abundant waste blocks of lower crystal-poor Lemitar Tuff, distinctive in their content of sparse small biotite, are present on the dump of the 75-m-deep shaft at the Nancy mine.

Some features of the upper Lemitar Tuff at the Tower mine area are anomalous. Core of the upper Lemitar in the drill hole shows some minor shearing and cementation by red jasperoid. In contrast, surface exposures of the Lemitar Tuff along the western and northern boundaries of the triangle are strongly silicified. Pumice and compaction foliation are only vaguely visible in these silicified outcrops. Where attitudes have been obtained, they are mutually inconsistent, suggesting the outcrops may actually represent large jumbled blocks. Attitude data of this type are marked on the map with a query. Large xenoliths of upper Lemitar have been locally observed in the overlying basaltic andesite lavas at the big exploration pit 300 m northeast of Tower mine and in the basal zone of the tuff of South Canyon outcrop due west of the Tower mine. A similar occurrence of jumbled blocks occurs in tuff of South Canyon exposures in the same areas. These jumbled blocks are unconformably overlain by non-brecciated and uniformly dipping lower Popotosa strata (Tpsd, pl. 1) and younger lavas (Tsd, Tpkb, pl. 1). These relationships suggest that the Tower mine block is locally mantled by ancient landslide deposits of late Oligocene age. These landslides are probably related to the eastern margin of the Sawmill Canyon cauldron (fig. 9 and 14) or an early-rift fault zone utilized by the Socorro cauldron (pl. 2, section G-G' and H-H'). The sequence of outcrops of Lemitar Tuff through Popotosa Formation in the immediate vicinity of the Tower Mine drill hole (pl. 1), apparently do not consist of jumbled blocks.

The second category of cauldron-facies ash-flow tuff is mostly exposed in the area southeast of Black Canyon.

Also included in this category are: 1) a small exposure at the Black Canyon Box (fig. 3) on the north side of the canyon, 2) an isolated fault block in the area 1 km south of Sedillo Ranch (abandoned), and 3) an isolated block about 1 km west of the Grefco perlite mine. The correlation of these outcrops with the upper crystal-rich and quartz-rich compositional subunit of the Lemitar Tuff must be considered as tentative. Alternatively they may be correlative with the Hells Mesa Tuff (G.R. Osburn, 1979, oral commun.). Local lithologic variations, stratigraphic relationships, and radiometric age dates, which are available in the study area, are permissive of either correlation. This correlation ambiguity will not be resolved until the southern remnant of the Socorro cauldron block, exposed in the central and southern Chupadera Mountains, can be mapped in detail.

Tuffs of the second category, hereafter referred to as "cauldron-facies Lemitar(?) Tuff" or simply "Lemitar(?) Tuff", have stratigraphic relationships significantly different from the "bona fide" Lemitar Tuff at the Tower Mine. The Lemitar (?) Tuff has the following general characteristics in the study area: 1) an apparent (regionally correlative) base to the section is not exposed; 2) it is everywhere overlain by cauldron fill members of the Luis Lopez Formation, in turn locally overlain by the tuff of South

Canyon, and 3) distinctive lithic-rich zones occur near the base and top, which are not present at the Tower Mine. These differences may be explained by considerably greater and contemporaneous subsidence of the eastern half of the Socorro cauldron in comparison to the western half.

The greatest thickness and variation in lithology of the cauldron-facies Lemitar (?) Tuff are found in the area south of Black Canyon. Structure sections here (pl. 2, section I-I') indicate a minimum thickness of about 880 m of the upper Lemitar (?) with the base of the section not exposed. In this area, the Lemitar (?) Tuff dips steeply to the east and forms low, rounded hills mantled by numerous small (unmapped) patches of angular colluvium. The Lemitar (?) Tuff is uniformly densely welded, crystal-rich and quartz-rich, except locally near the top where the total phenocryst content drops to about 25 percent, although quartz remains relatively abundant.

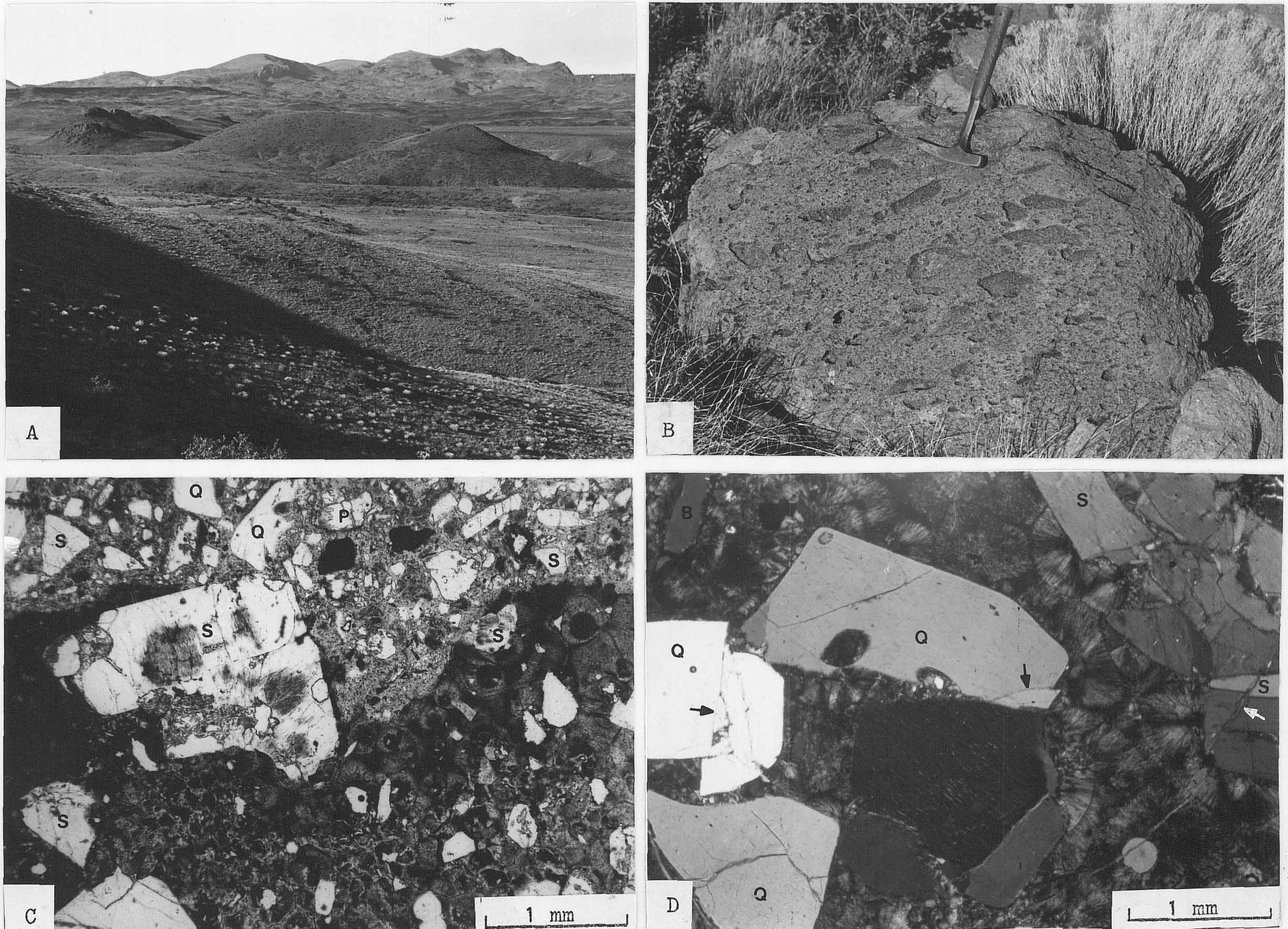
Three zones in the cauldron-facies Lemitar(?) Tuff can locally be defined on the basis of lithic content. The stratigraphically lowest zone is a distinctive heterolithic breccia containing about 5 to 20 percent xenoliths of Precambrian crystalline rocks, red siltstones and porphyritic andesites (Tlx, pl. 1). The xenolithic zone is approximately 335 m thick, and is only exposed in the area south of the Nancy mine. Outcrops of this lithic-rich zone are notably redder, because of a pale red matrix, in comparison to the typical grayish-red color of the overlying lithic-free Lemitar(?) Tuff (Tlu, pl. 1). Precambrian granitic xenoliths and red siltstone clasts, possibly derived from Permian rocks or early Tertiary sedimentary rocks, are the most abundant types of lithic fragments. A single thin section from this area shows that the granitic xenoliths contain microcline, slightly altered plagioclase, and muscovite; thus supporting the Precambrian age interpretation. This rock also contains about 30-40 percent phenocrysts(?) of partially argillized, subequal, sanidine and plagioclase, with 5-7 percent quartz and traces of biotite. A significant but unknown percentage of these phenocrysts (?) are xenocrysts derived from abrasion of the lithic fragments.

In the area south of the Nancy mine, xenoliths in the Lemitar(?) Tuff average about 4 cm in length and may be as much as 15 cm in length. Miesch (1956, p. 11) reports that Precambrian(?) granite and schist blocks as large as 1.5 m across, occur in xenolith-rich zones in the "massive rhyolite" (Lemitar(?) Tuff) about 6 km due south of the Nancy mine. Assuming these zones, 6 km apart, are approximately correlative, then the size relationship suggests that the source of these breccias may have been the southern wall of the Socorro cauldron. This inferred cauldron wall is probably now expressed by a large exposure of Precambrian rocks in the Coyote Hills of the southern Chupadera Mountains (Dane and Bachmann, 1965). The top 15 m of the "Tlx" unit is a tuff breccia containing only andesitic lithic fragments. It is clearly stratified in comparison to the underlying heterolithic tuffs. The xenolithic-rich zone most likely represents caldera-collapse breccias, mesobreccias in the terminology of Lipman (1976). Presumably the mesobreccias were derived from rock slides off the collapsing southern wall of the Socorro cauldron, which was oversteepened because of rapid subsidence of the cauldron block contemporaneous with the ash-flow eruptions.

The remainder of the cauldron facies Lemitar(?) Tuff (Tlu, pl. 1), in the area south of Black Canyon, is as much as 550 m thick and consists mostly of typical quartz-rich, crystal-rich tuff with sparse light-colored crystal-rich pumice. Mafic rhyodacitic pumice, common in this zone on the outflow sheet (fig. 13), are evidently absent here.

The uppermost 120 m of the cauldron-facies Lemitar(?) Tuff, exposed on Section 8 Ridge, consists of distinctly bedded and lithic-rich, densely welded tuffs (fig. 16A). These stratified ash-flow tuffs are apparently not typical of ignimbrite deposits in general (Ross and Smith, 1961),

Figure 16. Photographs and photomicrographs of a bedded coignimbrite lag-fall deposit in the uppermost cauldron-facies Lemitar(?) Tuff at Section 8 Ridge, northern Chupadera Mountains. A, Steeply east-dipping "strata" of welded tuff formed by resistant clast-poor intervals form slopes in foreground; Socorro Mountains on the horizon. B, Boulder of coarse clast-rich tuff illustrating the subparallel alignment and angular character of clasts. Hammer head 18 cm long. C, Photomicrograph in plane polarized light showing details of a boundary between a spherulitic clast and the host welded tuff. The boundary cuts across the spherulitic structure (arrow) but the boundary is not sharp probably because of slight resorbtion during welding. Note abundance of broken fragments in host tuff including angular bits of clast matrix (black spots at top center, spherulitic texture not visible due to high contrast in negative). Contains phenocrysts of sanidine (S), plagioclase (P), quartz (Q), and biotite (not visible). D. Photomicrograph with crossed nicols of spherulitic clast showing microfaults (arrows) and brecciated quartz phenocryst indicating occurrence of a strong shear stress while the matrix was relatively plastic. Contains phenocrysts of sanidine (S), quartz (Q), biotite (B), and plagioclase (not visible). Crossed nicols.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C
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or known to have any equivalent on the outflow-facies Lemitar Tuff in particular.

This unusual deposit is interpreted to be an autolithic, co-ignimbrite, lag-fall deposit similar to one in Central America recently described by Wright and Walker (1977). Wright and Walker (Ibid, p. 732) state that: "Criteria for recognizing lag-fall deposits are their coarse grain size, richness in lithics or dense juvenile material, stratification, and lateral correlation with ignimbrite." Evidently co-ignimbrite lag fall deposits are transitional between ash-flow and air-fall tuffs and posess characteristics of both types of deposits.

The bedded clast-rich tuffs at the top of the cauldron-facies Lemitar(?) Tuff most likely represent the effect of periodic blocking of a nearby ignimbrite vent by its own, temporarily volatile-poor, magma. Each blockage was presumably followed by explosive disruption of the partially congealed magma plug. This explosive clearing of the vent then began a new cycle of ash-flow eruption. Each bed, generally 3 to 10 m thick, in the series of at least a dozen beds, is considered to represent a single nuee ardente. The resistant ledges which define these beds (fig. 16A) are believed to be the relatively ash-rich winnowed fraction deposited during the waining stage of each eruptive cycle.

This process, may be somewhat analogous to a "sputtering stew pot" that spits out steam and water droplets at the same time. It would most likely be characteristic of the later stages of cauldron subsidence, when volatile pressures in the magma chamber are nearly at equilibrium with weight of the cauldron block (stew-pot lid) allowing the vent to be closed periodically.

Megascopic observations which support the above interpretation are: 1) the reddish brown clasts weather preferentially from the densely welded grayish-red, host tuff, thus the clast-rich horizons tend to weather to slopes and the thin clast-free intervals form resistant ledges (fig. 16A); 2) the coarsest clast-rich intervals, well exposed on the western crest of the ridge (pl. 1), show crude graded bedding with the largest clasts, 15-75 cm long, concentrated near the base of three separate beds; 3) clast geometries range from elliptical and well-rounded (particularly larger clasts) to subequant and angular (fig. 16B); 4) the elliptical clasts are generally aligned parallel to stratification; 5) length to width ratios of clasts generally do not exceed 5 to 1, with 2 to 1 most common, 6) the clast-rich beds are apparently lenticular and grade laterally into typical upper Lemitar(?) Tuff.

The same stratigraphic horizon at the top of the cauldron-facies Lemitar(?) Tuff is locally repeated through faulting. It occurs at the Black Canyon Box and also occurs on the east side of the fault block about one km north of Section 8 Ridge. At the first locality the bedded interval is only 50 m thick and clasts are less than 3 cm long; at the second locality no clasts or bedding were observed.

Microscopic observations that support the autolithic lag-fall deposit interpretation are: 1) clasts have the same mineralogy -- sanidine, plagioclase, quartz, and biotite -- as the host tuff; 2) clasts are crystal-rich (30-40 percent) with a non-pyroclastic texture of relatively large, euhedral and mostly unbroken phenocrysts (bipyramidal quartz common) in a spherulitic red matrix (fig. 16C and 16D); 3) the host tuff matrix is densely welded and contains abundant broken crystal fragments and microscopic fragments of broken spherulites (fig. 16C); 4) the clasts have irregular boundaries that cut across spherulites (fig. 15C); and 5) some clasts contain microfaulted and shattered phenocrysts (fig. 16D) indicating that the clast material was semicongealed and plastic at the time of its explosive disruption. The low degree of flattening of the clasts, and an apparent density sufficiently high to transmit shear stress in the clast, indicates that these clasts were not vesiculated pumice at the time of deposition, but rather the equivalent of a viscuous spherulitic rhyolite lava.

Other exposures of the cauldron facies Lemitar(?) Tuff west of the Grefco Perlite mine and south of Sedillo Ranch are mostly "normal" upper Lemitar(?) Tuff. At the latter locality the top 20 m of the section is unusually low in total phenocryst content (20-30 percent) and contains relatively small phenocrysts. This ash-rich, densely welded zone, like those at Section 8 Ridge, may represent the winnowed fraction deposited on the waining of an ash-flow eruption column. West of the Grefco mine, the Lemitar(?) Tuff forms unusually craggy outcrops and locally exhibits well-developed sheet jointing subparallel to the foliation. Joint sets are commonly filled with red jasperoid here and the outcrop is locally brecciated. Foliation attitudes are consistent with warping of the "strata" by adjacent faults. There is no indication of a "jumbled blocks" relationship here, as at the Tower mine area.

Luis Lopez Formation

Following the major period of subsidence of the Socorro cauldron, the depression was backfilled predominantly by alternating eruptions of andesitic lavas and lithic-rich rhyolitic ash-flow tuffs and then locally capped by silicic rhyolite domes and associated tuffs. Minor volumes of volcaniclastic sedimentary rocks including landslide deposits, andesitic laharic breccias, and rhyolitic breccias and sandstones, occur respectively near the base, middle and top of the fill sequence. This complexly intertonguing and heterogenous fill of the Socorro cauldron is here collectively named the Luis Lopez Formation for a large area of exposures in the Luis Lopez manganese district (Miesch, 1956) at the northern end of the Chupadera Mountains.

The thickest continuously exposed stratigraphic section of the Luis Lopez Formation occurs in an east-dipping block overlying cauldron-facies Lemitar(?) Tuff at the Black Canyon Box. A structure section here (pl. 2, H-H') indicates the formation — with some members not present — is approximately 530 m thick. The total thickness of a composite section of all 15 members is 1470 m; however this is not representative of the maximum thickness of the formation because several of the members are highly lenticular. The greatest thickness of the Luis Lopez Formation should occur about midway between the cauldron margin at Socorro Peak and the resurgent block south of Black Canyon. A reasonable thickness estimate in the Blue Canyon area is considered to be about 800 m (pl. 2, C-C').

The east face of Socorro Peak is essentially a cross section view of the complex topographic, and structural, northern margin of the Socorro cauldron (see fig. 38).

Here the Luis Lopez Formation laps unconformably onto the Pennsylvanian Madera Limestone and is locally overlain by the tuff of South Canyon. This relationship indicates the presence of a major intravolcanic unconformity. As much as 900 m of regional Oligocene volcanic strata, which are found in the interval between the Madera Limestone and tuff of South Canyon in the Lemitar Mountains, are absent on Socorro Peak (fig. 6). The missing strata include the Spears Formation, Hells Mesa Tuff, A-L Peak Tuff, La Jara Peak Basaltic Andesite, and Lemitar Tuff.

The Luis Lopez Formation thins abruptly to the north across the Pathway Canyon fault (pl. 2, section A-A'), which is interpreted as a ring-fracture fault of the Socorro cauldron. At least 220 m of strata wedge out, or are faulted out, across this structure. North of the Pathway Canyon fault and east of "M" Mountain, the exposed section of Luis Lopez Formation is 274 m thick. Toward the inferred major ring fracture under the Blue Canyon Dome, the Luis Lopez Formation increases in thickness to probably as much as 800 m.

The landslide deposits (Tlm, pl. 1) at Pathway Canyon mark the location of an early topographic wall of the

Socorro cauldron that was shortly thereafter buried by younger volcanic members of the Luis Lopez Formation. The volcanic members of the Luis Lopez Formation must wedge out against a younger -- northward retreated -- topographic wall, somewhere in the interval between Socorro Peak and Strawberry Peak, since the formation is not present in the southern Lemitar Mountains (Lemitar Map). This inferred younger topographic wall is not exposed. It is unconformably buried by the Miocene Popotosa Formation; as are other elements of the Socorro cauldron where locally exposed on Socorro Peak.

Bedded rhyolite tuffs (Tlrbt, pl. 1) and a thin remnant of an overlying rhyolite flow (Tlrb), which occur locally at the top of the Luis Lopez Formation to the south of Big Cliff, are abrubtly truncated at the Pathway Canyon fault. These units were truncated on the upthrown north block by an erosional unconformity at the base of the overlying Popotosa Formation in late Oligocene or early Miocene time. These relationships are attributed to late-stage subsidence on the cauldron ring fracture near the end of the cauldron filling volcanism.

The Luis Lopez Formation is locally overlain by the tuff of South Canyon (26.2 m.y., Appendix B), in the area northeast of the Tower mine, and on the northern end of the eastern face of Socorro Peak. At most localities, between the above areas, the formation is overlain disconformably and in angular unconformity by conglomeratic facies of the lower Popotosa Formation.

A biotite separate from a thin interval of welded rhyolite tuff (Tlrst, pl. 1), collected near the hilltop just south of the East Roadcut along Highway Sixty, has a K-Ar age of 28.6 ± 1.1 m.y. (Appendix B). This age is slightly old with regard to that of the Lemitar Tuff (average age 27.9 m.y., Appendix B), tentatively interpreted as underlying the Luis Lopez Formation. However, the ages are within limits of analytical error. Since the cauldron-facies Lemitar (?) Tuff, could be Hells Mesa Tuff (average age 32.2 m.y.), the slight age discrepancy may also be non-existent. The 28.6 m.y. date is from a very fresh rock containing unaltered plagioclase; it is considered to be representative of the age of volcanic rocks that filled in the Socorro cauldron after its collapse.

A separate of a dark-colored mica, originally believed to be phenocrystic biotite, from a coarsely porphyritic rhyodacite(?) dike (Tiap, pl. 2) has been dated at 22.8 ± 0.9 m.y. (Appendix B). This sample was taken from an old roadcut of Highway Sixty just north of the Big Roadcut. In thin section, this mica does not look like typical phenocrystic biotite; instead it is inhomogeneous with respect to color and is poikolitic with abundant opaque inclusions. The mica also has ragged outlines and a seriate occurrence from phenocryst size to microscopic wisps in the matrix. The mica is apparently a mixture of pale-brown muscovite and reddish-brown biotite. These mica "phenocrysts" may represent replaced ferromagnesian phenocrysts such as pyroxene or hornblende or alternatively they might represent recrystallized primary biotite phenocrysts. Replacement of plagioclase phenocrysts in this rock by potassium feldspar indicates the rock has been potassium metasomatized. The radiometric age of this mica probably approximates a potassium metasomatic hydrothermal event -- other potassium metasomatic and hydrothermal events are also recognized in the Socorro Peak area. A petrographically similar lava flow (Tlap, pl. 1) locally underlies the 28.6-m.y.-old rhyolite tuff (Tlrst, pl. 1) at the East Road Cut; this stratigraphic relationship supports the interpretation of the younger mica date as representing a post-emplacement hydrothermal event.

Reconnaissance traverses in the vicinity of Walnut Creek -- Nogal Canyon of southern Chupadera Mountains -- by the author and G.R. Osburn, have shown that a similar sequence of intermediate composition lavas, rhyolitic tuffs and domes (respectively map units: Tas, Tts, and Trs of

Miesch, 1956, pl. 1) is probably correlative to the Luis Lopez Formation. This heterogenous sequence appears to overlie the "massive rhyolite" of Miesch (1956, pl. 1) considered to be equivalent to the cauldron-facies Lemitar(?) Tuff of this study. However, an enigmatic situation presently exists in this area. The probable equivalent of the Luis Lopez Formation appears to be overlain by distinctly zoned lower and upper Lemitar Tuff at the eastern mouth of Nogal Canyon (narrow band of "Trs" as mapped by Miesch along east side of range). These apparent relationships will have to be mapped in detail before the correlation problem is solved.

The Luis Lopez Formation is evidently restricted to the eastern half of the Socorro cauldron. Geologic mapping of the western half of the Socorro cauldron, exposed in the eastern Magdalena Mountains (Osburn, 1978; Petty 1979), has shown reasonable correlatives of the Luis Lopez Formation to be entirely absent or lenticular and minor in volume. Petty (1979, p. 75-77) has observed rhyolitic tuffs and lavas in a paleochannel cut in the Lemitar Tuff. These units are overlain by the tuff of South Canyon, which places them in the same stratigraphic position tentatively assigned to the Luis Lopez Formation. If this assigned position is correct, then approximately 120 to 180 m of upper tongue La

Jara Peak Basaltic Andesite lavas exposed along the western margin of the Socorro cauldron in the South Canyon area (Osburn, 1978, p. 46) should be contemporaneous with the Luis Lopez Formation.

Exposures of the Luis Lopez Formation in the dissertation area have been previously mapped as parts of investigations by Lasky (1932, p. 118-133), Miesch (1956), Smith (1963), Lowell (1967), and Burton (1971). The reader who compares the interpretations of these previous investigators, with those of this report, will find some significant differences with respect to equivalents of the Luis Lopez Formation. The most significant differences involve the distinction of densely welded tuffs from silicic lavas, and the usefulness of these ash-flow tuffs as stratigraphic markers. Other differences in interpretation are related to the distinction of massive andesite flows from intrusive andesites, and the distinction of volcaniclastic sedimentary rocks from agglomerates and various types of volcanic breccias.

Sequence and Correlation of Members

Strata of the Luis Lopez Formation have been subdivided into 15 petrogenetic map units, each of which has been given member status (pl. 1). The general mapping approach has been to separate strata of distinctly different genesis -- lavas flows, ash-flow tuffs, bedded tuffs, and volcaniclastic sedimentary rocks -- and to combine (by color scheme, pl. 1) petrologically similar rocks, which form different stratigraphic horizons. Two thick intervals of lithic-rich, rhyolitic, ash-flow tuffs (Tlt_1 and Tlt_2, pl. 1) are the key units that permit a lithologic correlation between the geographically isolated exposures of the Luis Lopez Formation at Socorro Peak and the northern Chupadera Mountains. The general stratigraphic sequence and lateral relationships of the members of the Luis Lopez Formation, and their relationship to structural elements of the Socorro cauldron, are summarized in Figure 17.

The established stratigraphic sequence and available chemical analyses (Table 5) suggest that the caldera-filling eruptions alternated between intermediate and rhyolitic magmas. Local sequences of intermediate lavas appear to become more differentiated with time, as indicated by increases in size and abundance of plagioclase phenocrysts in successive flows. No lavas or tuffs that appear transitional in texture or composition between the intermediate rocks and the rhyolites have been observed. Thus, although the high-silica rhyolite lavas generally cap the caldera-fill sequence, there is little physical evidence to indicate the

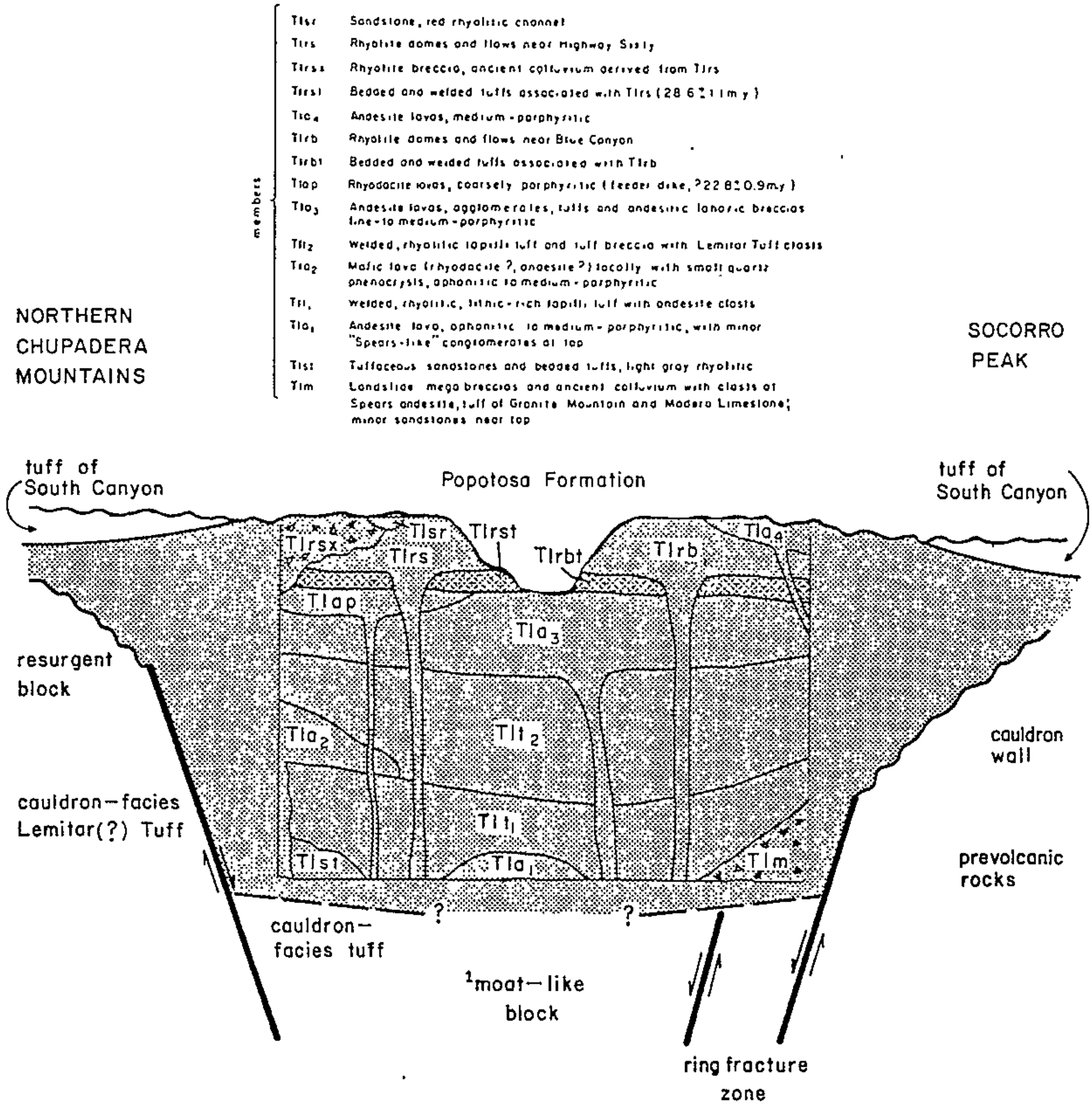

Figure 17. Schematic stratigraphic cross section of the Luis Lopez Formation (shaded); illustrating the general stratigraphic relationships of members within the formation and their general setting with regard to structural elements of the Socorro cauldron. Section is modified from the explanation of Plate 1 and is not drawn to scale. Effects of post-cauldron normal faulting related to the Rio Grande rift are not shown.

[1] An annular "moat" geometry has not been established for the Socorro cauldron.

Table 5. Analyses of lavas and a tuff in the Luis Lopez Formation. Chemical analyses by x-ray fluorescence, D.L. White, New Mexico Bureau of Mines and Mineral Resources (see p. 28).

Sample No. Member (pl. 1)	77-7-7 Tla_3	77-5-6[1] Tlrb	77-5-2 Tlrst
Major oxides (weight percent)			
SiO_2	54.79	77.81	77.80
Al_2O_3	15.32	13.26	12.45
Fe_2O_3 (total)	5.69	1.14	0.92
MgO	6.28	<0.01	<0.01
CaO	4.57	0.07	0.46
Na_2O	4.17	1.03	2.99
K_2O	3.45	8.27	6.39
TiO_2	0.92	0.31	0.31
Total	95.19	101.90	101.33
phenocrysts (volume percent)			
sanidine	-	15	1
plagioclase	20	-	∿ 1.5
quartz	-	5	tr[2]
biotite	-	-	∿ 0.5
clinopyroxene	3	-	-
opaques	2[3]	tr[2]	tr[2]
groundmass	75	80	97

[1] potassium metasomatized

[2] tr = trace

[3] mostly resorbed pyroxene(?), see description

Table 5. (continued)

Sample No.	Sample Descriptions
77-7-7	Dark-gray, slightly propylitized and potassium metasomatized, pyroxene andesite. Contains abundant clear, fresh, tabular (2-4 mm) plagioclase phenocrysts (An_{40} - An_{45}); and some sparse dark-greenish, granular-looking, stubby pyroxene phenocrysts (augite?), partly replaced by calcite, in a groundmass of felted plagioclase microlites and disseminated magnetite. Phenocrysts reported as opaques are mostly stubby prisms (some are eight-sided) of an early ferromagnesian mineral, possibly orthopyroxene. The possible orthopyroxene shows magmatic reaction to form untwinned feldspar(?) and magnetite. Also included with opaques are traces of small needle-like prisms altered to magnetite(?), which may have been hornblende. Traces of metasomatic mica are associated with hematite. Blue Canyon area (SE/4, NW/4, Sec. 16, T3S, R1W).
77-5-6	Grayish-red, potassium-metasomatized, high-silica rhyolite. Contains abundant pearly-white, subhedral, sanidine phenocrysts (1-3 mm) and small euhedral to rounded quartz (0.5 - 1 mm), in a matrix of spherulites partly replaced by polycrystalline, microgranular quartz. Sanidines are partly altered; disseminated hematite(?) occurs in exsolution lamellae, and illite(?) or sericite occur in small patches. North-northeast of "M" Mountain (SW/4, NW/4, Sec. 4, T3S, R1W).
77-5-2	Pale-red, densely welded, very crystal-poor, high-silica rhyolite ash-flow tuff. Contains sparse and unaltered phenocrysts (1-2 mm) of plagioclase (An_{31}), sanidine, biotite, and a trace of small (< 0.5 mm) quartz, in a fluidal, primary-welded, devitrified matrix. Some small relict patches in matrix exhibit densely welded glass-shard textures. Highway Sixty, south of East Roadcut (NW/4, SW/4, Sec. 32, T3S, R1W).

rhyolites are a differentiation product of the intermediate magmas. Caldera-fill lavas are not petrologically similar to the cauldron-facies tuff, suggesting that this magma type was completely expended during the main ash-flow eruptions and cauldron collapse. The fifteen members of the Luis Lopez Formation may, for the purpose of description and discussion, be grouped into the following five catagories based on similar lithologies and age relationships.

Early landslide megabreccia (Tlm) and tuffaceous sandstones (Tlst).

Several small exposures consisting dominantly of dark-purplish-gray to bluish-gray heterolithic boulder conglomerate and breccia, occur sporadically along Pathway Canyon on the east face of Socorro Peak. They are mapped here as the locally oldest member of the Luis Lopez Formation (Tlm, pl. 1). They are interpreted as ancient landslide and colluvial deposits shed southward into the Socorro caldera from its northern topographic wall. The northern side of the topographic wall is now marked by exposures of Madera Limestone. The interpretation of these bouldery deposits is complicated by the lack of a complete section with a depositional top and bottom and by partial obscuring of their spatial relationships to the cauldron wall by younger faulting.

The southward-sloping caldera wall has been down-dropped to the east and offset northward by north-trending rift faults (Woods Tunnel fault zone; pl. 2, A-A', B-B').

The largest exposure of ancient landslide deposits occurs at the box area about halfway up Pathway Canyon. Here the member consists of as much as 107 m of crudely bedded, moderately indurated, conglomerate and breccia with a sandy mud matrix. Rounded to angular cobbles and boulders of dark-gray and dark-purplish-gray andesite porphyry, similar to those in the volcaniclastic facies of the Spears Formation (fig. 1, Table 1), are dominant (fig. 18A). Other clasts consist of gray micritic limestone (Madera Limestone?), light-brownish gray silicified limestone, and a distinctive reddish-brown to purplish-gray crystal-rich welded tuff. These welded-tuff clasts are mostly lithic-rich and contain only traces of quartz phenocrysts; these are distinctive traits of the tuff of Granite Mountain member of the Spears Formation. The presence of these welded-tuff clasts distinguishes these landslide deposits from volcaniclastic conglomerates of the Spears Formation.

The most unique feature of this exposure is a large block of sheared and silicified Madera(?) Limestone, which appears to be completely enveloped by the bouldery deposits (fig. 18B). The block, about 10-15 m thick and at least

Figure 18. Ancient landslide deposits and colluvium (Tlm) of the Luis Lopez Formation exposed at Pathway Canyon Box on the eastern face of Socorro Peak. A, Indurated bouldery conglomerate containing clasts of silicified limestone (s), micritic limestone (m), andesite-latite lavas of Spears Formation type (a), and tuff of Granite Mountain (t) in a grannule to mud matrix. Tuff of Granite Mountain clasts distinguish this deposit from Spears Formation conglomerates. Hammer head 18 cm long. B, Looking west northwest at sheared and silicified ancient landslide block (lb) of Pennsylvanian limestone about 15 m thick, enveloped in crudely bedded (lower left) bouldery colluvium (Tlm). Well-stratified Madera Limestone (Pm) forming caldera-wall interpreted as source of landslide deposits. Down-faulted blocks of rhyolite lava and tuff (Tlrb), and Popotosa Formation siltstones and mudstones (Tpsl); and surficial colluvium (Qca) confuse and mask relationship of ancient landslide deposits to caldera wall.


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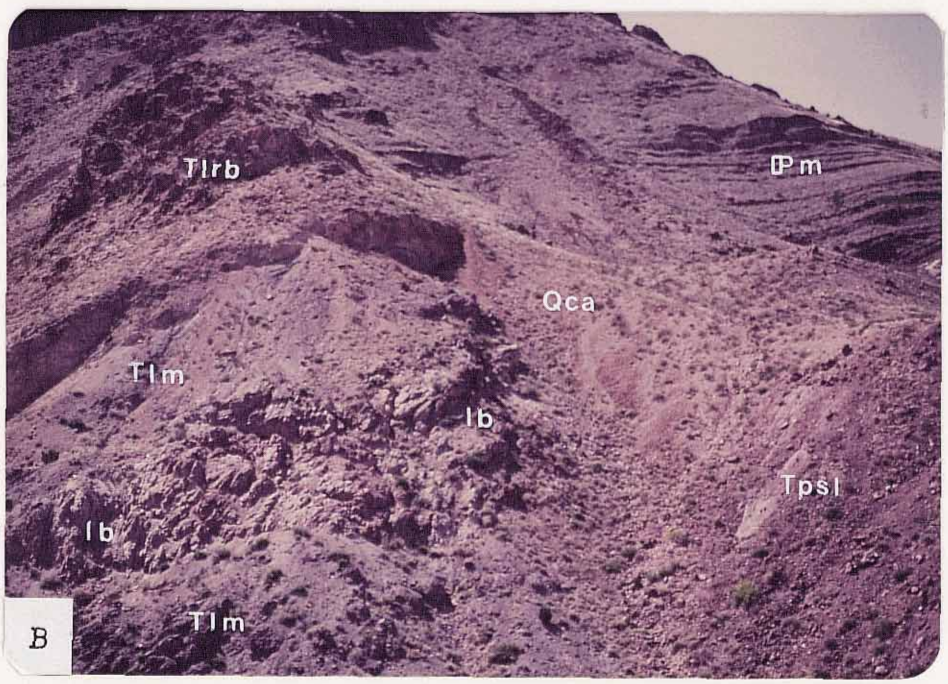
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B

110 m long, is considered to be an ancient landslide block. At its south end, the block is truncated by a northwest-trending mylonitic fault gouge exposed near the canyon floor. On the southern wall of the canyon, about 15 m above the canyon floor, is exposed a 3-m-thick interval of angular limestone blocks. This interval of limestone blocks is clearly interbedded in purplish conglomerates and sandstones, somewhat finer, but otherwise equivalent to those on the north side of the canyon. This bed of limestone blocks probably represents part of the colluvial apron that should have surrounded a large landslide block such as the one described above.

Light-gray and light-purplish-gray sandstones and siltstones, as much as 10 m thick, generally occur at the stratigraphic top of the landslide deposits, where they are overlain by poorly welded lithic-rich tuff (Tlt_1, pl. 1). Sandstones of this type are exposed along the southern margin and western margin of the Pathway Canyon Box exposure. Two similar exposures of sandstones that occur west of the box area are completely isolated by fault contacts and Quaternary colluvium. Some sandstones in the lower Popotosa Formation (Tpsw facies, pl. 1), which also crop out in the Pathway Canyon area, look very much like those in the Luis Lopez Formation. They may usually be distinguished by local stratigraphic relationships and clast lithologies; younger Popotosa strata contain abundant clasts of gray, crystal-poor rhyolite welded tuff (A-L Peak Tuff?).

The base of the landslide deposit is not exposed at the Pathway Canyon Box but to the northwest there are two small exposures of an andesitic conglomerate that appear to rest on the Madera Limestone. Further north these conglomerates are absent and lithic-rich tuffs (Tlt_1, pl. 1), which rest on the conglomerates at the box, lie directly on the Madera Limestone. These relationships indicate that the landslide deposits wedge out to the north onto the topographic wall of the cauldron. The wedge geometry of the landslide deposits is confirmed by a shallow thermal-gradient drill hole collared about 60 m north of the Middle mine portal (pl. 1). This hole intersected 15.5 m of andesitic conglomerate with clasts of silicified limestone (Tlm) below 15.0 m of lithic-rich tuff (Tlt_1) (R.W. Foster, unpub. report, circa 1970, Sanford Hole No. 6). About 160 m north of the mine portal, the lithic-tuff rests directly on silicified Madera Limestone.

Approximately 35 m of rhyolitic bedded tuffs and tuffaceous sandstones (Tlst, pl. 1) overly cauldron-facies Lemitar(?) Tuff on the northeast flank of the Chupadera Mountains about one km south of Sedillo Ranch. Light-gray to pinkish-gray, medium- to coarse-grained, parallel-bedded sandstones form the majority of the outcrop. The moderately wellindurated sandstones are rhyolitic and contain abundant grains of sanidine, quartz and some plagioclase(?) altered to friable white clay. Some intervals within the sandstones are slightly pumiceous and display graded bedding; they are thought to be air-fall tuffs. About 10 m of reddish-brown mudflow breccia with clasts of "Spears type" porphyritic andesite or latite is exposed near the middle of the unit at the south end of the outcrop. This andesitic mudflow is similar to andesitic conglomerates found at the top of the lower andesite member (Tla_1, pl. 1), where it is exposed at Black Canyon Box, and to the landslide conglomerates (Tlm, pl. 1) on Socorro Peak. This relationship suggests that all three of these members (Tlst, Tla_1, Tlm) which occur near the base of the Luis Lopez Formation, may be penecontemporaneous. The tuffaceous sandstones are overlain by andesite-rich lithic tuffs (Tlt_1). This contact is not well exposed, but appears to cut disconformably down-section into the sandstones toward the north.

<u>Lithic-rich ash-flow tuffs (Tlt_1 and Tlt_2)</u>

Approximately one third of the volume of the Luis Lopez Formation is formed by two thick intervals of mostly poorly welded, pumiceous, lithic-rich, rhyolitic ash-flow tuff. The lower lithic tuff (Tlt_1) contains abundant xenoliths of gray and purplish-gray porphyritic intermediate lavas (Spears type) and generally lacks, or only rarely contains, rhyolitic fragments. The upper lithic tuff (Tlt_2) is characterized by abundant xenoliths of reddish-brown, moderately crystal-rich or crystal-rich rhyolite with a phenocryst mineralogy -- sanidine, altered plagioclase, quartz, and biotite -- equivalent to the uppermost zones of the cauldron-facies Lemitar(?) Tuff (fig. 19A). Vitroclastic textures expressed by densely welded shards and small pieces of broken phenocrysts are apparent in these rhyolitic fragments when viewed in thin section (fig. 19B). A specimen of the upper lithic tuff from the Big Cliff area contains spherulitic blebs in some of the tuff clasts that are essentially identical to those in the cauldron facies lag-fall deposits.

Both lithic-tuff members are otherwise similar in lithology. Abundant small (2 mm to 3 cm) aphyric pumice lapilli usually altered to white clay (montmorillonite?; Miesch, 1956, p. 4) are a distinctive trait of these tuffs. Traces of small phenocrysts of sanidine and quartz are present in some of the pumice indicating that they are rhyolitic in composition. The tuff matrix often contains a greater

Figure 19. Upper lithic-rich tuff member (Tlt_2) of the Luis Lopez Formation. A, Outcrop of partially welded tuff showing typical weathered-out pumice lapilli and angular clasts of densely welded cauldron-facies Lemitar(?) Tuff in a white tuff matrix. From below Big Cliff on east face of Socorro Peak. Knife is 9 cm long. B, Photomicrograph showing partially welded tuff matrix containing microxenoliths (m) and xenocrysts (x) with adhered welded tuff matrix (black) and larger xenoliths of densely welded, crystal-rich, cauldron-facies Lemitar(?) Tuff. Note broken spherulite at bottom center. Phenocrysts and xenocrysts(?) of quartz (white), sanidine (gray and white, partially altered), altered plagioclase (grainy gray), and biotite (black). Same location as A. Crossed nicols.

proportion of quartz and sanidine crystals than is present in the pumice; this is most noticeable in the upper lithic tuff where the densely welded tuff clasts are abundant. The lithic tuffs appear to be crystal-poor tuffs containing 5-10 percent crystals of sanidine and quartz; however, most of these crystals may be phenoclasts derived from comminution of lithic fragments.

The majority of lithic-tuff outcrops are poorly welded and range from light brownish gray to light purplish gray and grayish-orange pink in color. Pumice lapilli weather out preferentially leaving a pock-marked and hackled surface; this is well displayed at the Big Roadcut along Highway Sixty. In areas adjacent to andesite dikes (Tia and Tiap, pl. 1) in the northern Chupadera Mountains and rhyolite dikes (Tirb) on Socorro Peak, the lithic tuffs are typically silicified to varying degrees. Here they take on a grayish-red to dark-reddish-brown color and the silicified pumice does not weather out preferentially.

The lower lithic tuff (Tlt_1) is generally a slope-forming unit. At Black Canyon Box it is about 45 m thick and sandwiched between two aphanitic andesite flows (Tla_1 and Tla_2). At this exposure it is poorly welded throughout. North of the West Roadcut at Highway 60, the same tuff interval contains a 1-2 m-thick tuffaceous sandstone lens near the middle of the unit. The lower lithic tuff at the Middle mine on Socorro Peak consists of two distinct zones, a basal poorly-welded lithic-rich zone about 50 m thick; and an upper moderate to densely welded zone about 30 m thick that is relatively lithic-poor and contains 5-10 percent flattened pumice disks altered to a soft, light-gray clay. These zonal relations in the lower lithic tuff indicate that it is a multiple-flow compound cooling unit.

The upper lithic tuff (Tlt_2) is partially to densely welded and normally forms resistant outcrops such as at Chupadera Cliff and the hogbacks northeast of Black Canyon Box. On the east face of Socorro Peak, the base of the upper lithic-rich tuff generally follows a break in slope where it overlies the slope-forming lower lithic tuff. At Socorro Peak, the upper lithic tuff is approximately 70 m thick; here the lithic fragments derived from the cauldron-facies Lemitar(?) Tuff generally form 1 to 10 percent of the rock and do not exceed about 5 cm in diameter. A minor amount of small andesitic lithics are also present here.

Northeast of Black Canyon, the upper lithic tuff is 275 m thick. Here, zones near the middle of the member consist of a very lithic-rich tuff or tuff-breccia, in which the clasts of cauldron-facies tuff form more than 50 percent of the outcrop and are as much as 60 cm in diameter. Subrounded boulders of crystal-rich cauldron-facies Lemitar(?) Tuff also occur sparsely in the upper lithic tuff near the Highway Sixty bridge over Bear Canyon. Just north of here, in the old road cuts of Highway Sixty, a local zone of the upper lithic tuff contains sparse clasts of cauldron-facies mesobreccia (Tlx) distinguished by granitic clasts within the tuff clasts. Since the "Tlx" member of the cauldron facies tuff should have been deeply buried at this time, these rare clasts probably represent material torn from the walls of the ash-flow vent.

Variations in clast content and degree of welding in the upper lithic-rich tuff imply that it is a multiple-flow compound cooling unit. Like the lower lithic tuff, where it is densely welded, it tends to contain less lithic fragments. Densely welded, lithic-poor, outcrops of this type are present near the top of the member along the eastern side of the Shrine Valley. Along the southwest side of the Shrine Valley, the uppermost 120 m of the upper lithic-rich tuff has been intensely silicified and replaced by dark red jasperoid (stippled area, pl. 1). Similar silicified zones are also present at the same stratigraphic horizon on the east side of the valley. In the Pathway Canyon Box area, the pumice in both lithic tuff members has a bluish-green color because of propylitic alteration.

<u>Intermediate lavas and associated rocks</u> (Tla_1, Tla_2, Tla_3, Tlap, Tla_4)

Four stratigraphic horizons of dark-colored lava flows, characterized by hydrothermally altered (potassium metasomatized) plagioclase as the dominant phenocryst or groundmass phase, are recognized in the Luis Lopéz Formation. These intermediate lavas span the entire time of cauldron filling, locally representing the oldest member (Tla_1) to the youngest member (Tla_4). A distinctive coarsely porphyritic flow (Tlap, pl. 1) that occurs at the top of the third stratigraphic horizon of mafic lavas (Tla_3) in the vicinity of Highway Sixty has been arbitrarily mapped as a separate member. In other areas, texturally different flows in one horizon have been mapped together as a unit.

The primary mineralogy and composition of the intermediate lavas in the Luis Lopez Formation are masked by pervasive potassium metasomatism, which appears to affect <u>all</u> of the Oligocene volcanic rocks in the Socorro area. Outcrops of the intermediate lavas that contain fresh plagioclase are so rare that, where observed by Lasky (1932, p. 128) and Burton, (1971, p. 24), they were mapped separately from their equivalent altered lavas as younger "diabase dikes".

In most outcrops of the intermediate lavas, plagioclase phenocrysts are typically altered to granular-looking mixtures of chalky white potassic clay and milky-white potassium feldspar. A few percent of altered ferromagnesian phenocrysts replaced by skeletal masses of magnetite and hematite are common in all the flows. Stubby prismatic crystals with four and eight sided sections suggestive of pyroxene are common in the older aphanitic to slightly porphyritic lavas (Tla_1, Tla_2, lower Tla_3) suggesting that they are mostly pyroxene andesites. Long, six-sided prisms and needle-like altered ferromagnesian phenocrysts, probably hornblende, are common in the more porphyritic lavas (upper Tla_3, Tlap, Tla_4) suggesting that they are hornblende rhyodacites. Biotite is fairly common but appears to occur sporadically within individual flows. Some biotite "phenocrysts" may be a hydrothermal phase formed during potassium metasomatism as previously suggested. Reddish-brown, euhedral, homogenous biotite occurs in the upper flow of the "Tla_3" member on Socorro Peak. This is probably primary phenocrystic biotite and the rock is tentatively classified as a rhyodacite or mafic quartz latite. Samples of the coarsely porphyritic lava (Tlap) and related dikes from the Highway Sixty area contain coarse altered plagioclase (replaced by potassic clay and feldspar), altered hornblende(?) prisms, metasomatic biotite replacing pyroxene, and sparse small, unaltered sanidine phenocrysts. This rock is tentatively classed as a rhyodacite or mafic quartz latite.

Three small, relatively fresh outcrops of the intermediate lavas are known. They occur west of the Middle mine (Tla_3), surrounded by colluvium at the west end of the exposure in Blue Canyon (Tla_3), and on the east side of the vent just north of Socorro Canyon (Tla_2). Sample number 77-7-7 (Table 5) from the Blue Canyon area appears to be a slightly altered pyroxene andesite containing approximately 55 percent SiO_2. This rock may be representative of the composition of about half of the intermediate lavas in the Luis Lopez Formation prior to being hydrothermally altered.

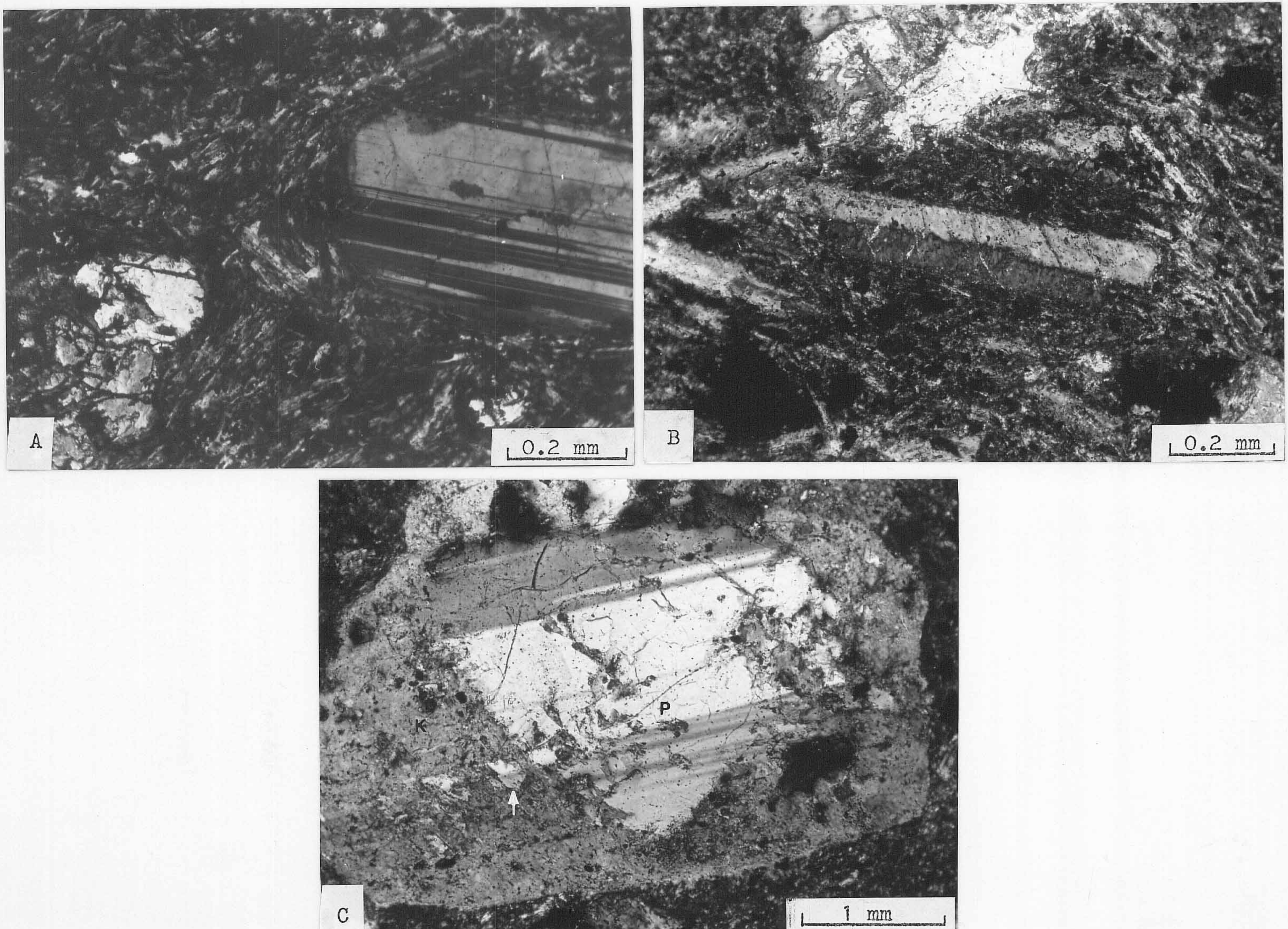
Most of the exposures of the intermediate rocks in the Luis Lopez Formation appear to be formed by massive lava flows, on the order of 30-90 m in thickness. Flow boundaries and geometries are best defined by contacts with adjacent strata. In the Black Canyon Box area, three flow units (Tla_1, Tla_2 and Tla_3) are recognized through sequential relationships with the lithic-rich tuffs (Tlt_1 and Tlt_2). Below Big Cliff, on the upper east face of Socorro Peak, the strata-like geometry of the "Tla_3" flows is also defined by the underlying lithic tuffs. Laharic breccias that contain only andesite clasts locally mark the base of the flow unit here. Directly east of the "M" two flows are apparent. The lower flow is approximately 50 m thick and contains 5 to 10 percent chalky clay pseudomorphs of plagioclase phenocrysts that are 1-2 mm long; the upper cliff-forming flow is about 70 m thick and contains 25 to 30 percent altered plagioclase phenocrysts 1 to 4 mm in length and some reddish-brown biotite.

Two stratigraphic horizons of massive, aphanitic to slightly porphyritic lava (Tla_1 and Tla_2) crop out in the Black Canyon Box area. They are distinguished primarily by their stratigraphic relationships: underlying the lower lithic tuff (Tlt_1) and the upper lithic tuff (Tlt_2), respectively. Similar aphanitic lavas with equivalent stratigraphic relations to the lithic tuffs also crop out along western Socorro Canyon and to the north of Section 8 Ridge. No lavas are recognized at these stratigraphic positions on Socorro Peak. At the Black Canyon Box, the lower flow (Tla_1), about 30 m thick, is capped by several meters of Spears-like conglomerate with porphyritic andesite clasts. The flow rock is dark-gray to purplish-gray and generally appears aphanitic, although microphenocrysts may locally be visible with a hand lens. Sparse, small, greenish-colored ovoids after altered ferromagnesian phenocrysts are fairly common. The upper aphanitic flow unit (Tla_2) is dark-reddish-brown, and has no apparent internal boundaries even though it is 120 m thick. Very sparse and small quartz crystals, believed to be xenocrysts, occur sporadically throughout this flow.

Examination of thin sections from both aphanitic flow units (Tla_1, Tla_2) at Black Canyon clearly indicates that they are intensely potassium metasomatized. Texturally these altered mafic lavas are very similar to relatively unaltered, slightly porphyritic and microporphyritic, pyroxene andesite in the "Tla_2" member north of Socorro Canyon (fig. 20A). The fresh rock at Socorro Canyon contains about 5 percent microphenocrysts (0.2-1.0 mm) of calcic andesine (An_{40-49}) that exhibits fine polysynthetic albite twins and some multiple carlsbad twins. Sparse, small microphenocrysts of clinopyroxene (augite?) are present along with a few relatively large (1-2 mm) phenocrysts of an early ferromagnesian phenocryst, possibly orthopyroxene or olivine, altered to hematite. The remainder of the fresh lava consists of a pilotaxitic-textured groundmass of plagioclase microlites with some small grains of magnetite and clinopyroxene in the interstices of the feldspar laths.

Thin sections of the aphanitic lavas (Tla_1, Tla_2) from the Black Canyon Box are texturally equivalent to the

Figure 20. Photomicrographs illustrating fresh and potassium metasomatized microporphyritic andesite lavas in the Luis Lopez Formation; and a partially metasomatized plagioclase phenocryst. A, Anomalously fresh pyroxene andesite (Tla_2) from east side of vent north of Socorro Canyon. Pilotaxitic groundmass of plagioclase microlites (gray), granular magnetite (black) and clinopyroxene contains microphenocrysts of combined carlsbad-albite-twinned plagioclase (An_{45}) and augite (?) (white to light gray). Crossed nicols. B, Potassium-metasomatized microporphyritic andesite (Tla_1) from north of Black Canyon Box. In comparison to "A" note the similarity in pilotaxitic texture of groundmass microlites and granular opaques. Lath-shaped microphenocryst of plagioclase in center is replaced by potassium feldspar. Note the highly sutured carlsbad twin plane, presumably deformed during recrystallization, the absence of albite twinning, and patchy extinction of the crystal. These traits are characteristic of metasomatic potassium feldspar. Large opaques are probably pseudomorphs after clinopyroxene and large white patches are calcite associated with scaly chlorite formed after an early ferromagnesian phenocryst. Crossed nicols. C, Plagioclase phenocryst (P), partially replaced by potassium feldspar (K). Note the small islands of unaltered plagioclase (arrow) within the metasomatic potassium feldspar, and the destruction of albite twin planes by the metasomatism process. This photomicrograph serves to illustrate the inferred nature of the transition between the fresh and completely metasomatized rocks shown above. Specimen from welded ash-flow tuff unit in the central Lemitar Mountains. Crossed nicols.


A
0.2 mm
B
0.2 mm
K
P
C
1 mm

pyroxene andesite lava at Socorro Canyon. However, the plagioclase microphenocrysts and microlites do not exhibit albite twins and are stained by sodium cobaltinitrite indicating they are now potassium feldspar. Carlsbad twins of the original plagioclase crystals apparently are not destroyed by the potassium metasomatism but some twin planes are clearly deformed by the recrystallization process (fig. 20B). In the lower aphanitic lava the larger ferromagnesian phenocrysts are replaced by greenish, scaly chlorite and calcite and microphenocrysts of clinopyroxene have been replaced by iron oxides.

A thin section of the upper aphanitic flow unit (Tla_2) from near the top of the flow at the Black Canyon box has similar textures and alteration assemblages as described above for the lower flow unit. The only significant difference is the presence of a few small, rounded quartz crystals which are outlined by narrow rims of devitrified glass. This apparent disequilibrium relationship and their sporadic occurrence suggests that the quartz crystals are xenocrysts. Miesch (1956, p. 4) classified the upper aphanitic lava at Black Canyon (Tla_2) as a "dark rhyolite", apparently because of the association of potassium feldspar and quartz.

Sparsely porphyritic lavas of the "Tla_2" member at Socorro Canyon were apparently erupted from a small andesitic volcano on the north side of the canyon. The dissected vent complex consists of remnants of a small neck and an adjacent cinder cone. Here the younger lithic tuff (Tlt_2) is plastered onto the eastern flank of the andesitic volcano along a steep east-dipping contact.

The "Tla_3" horizon, which normally overlies the upper lithic tuff (Tlt_2), is the most widespread, voluminous, and variable of all the intermediate lava members. The bulk of the "Tla_3" exposures consists of dark-gray to dark-purplish-gray, massive flows of andesitic to rhyodacitic lava that typically become more porphyritic upwards in local sections. Where propylitized, such as along the Wood's Tunnel fault zone (pl. 2, C-C') at Blue Canyon, these lavas are light-gray to greenish-gray. Subparallel alignment of plagioclase phenocrysts is a common expression of their flow structure. Contacts between individual flows, marked by reddish-brown autobrecciated zones, are only rarely exposed. Blocky and platy colluvial debris mantle much of the exposures, particularly in areas where the terrain consists of low, rounded hills.

The "Tla_3" member locally includes reddish-brown, monolithic, andesitic laharic breccias, such as those that are well exposed just north of Black Canyon near Section 8 Ridge. Several small areas of indurated, bedded andesitic cinders, interpreted as agglomerates, are also common in areas within 500 m of local vents.

The vents for these intermediate lavas are commonly expressed by craggy outcrops with a flaggy fracture (flow foliation) that forms steeply dipping concentric patterns around the vent. Vents expressed in this manner are found: northeast of the "M" on the upper face of Socorro Peak, at Blue Canyon, and east of the Shrine Valley. At the southwest end of the Bear Canyon Box, the canyon wall exposes the profile of a volcanic vent. Here the vent wall is formed by agglomerates and lithic tuff and the near vertical flow structure in the vent can be seen to flatten and fan out upwards into the adjacent flow (fig. 21). Northeast of Black Canyon Box, a long andesite dike (Tia) probably feeds the adjacent flow (Tla_3) adjacent to the eastern end of the dike. The relationships of the dike and flow are largely masked by colluvium.

An exposure of medium-gray to reddish-brown coarsely porphyritic lava (Tlap) in the area south of the East Road-cut appears to represent a single flow about 85 m thick. This flow is distinguished by 5-15 percent of elliptical to subhedral, chalky-white altered plagioclase phenocrysts about 5 to 15 mm in length. The altered feldspar weathers from the outcrop giving it a pock marked appearance. Some

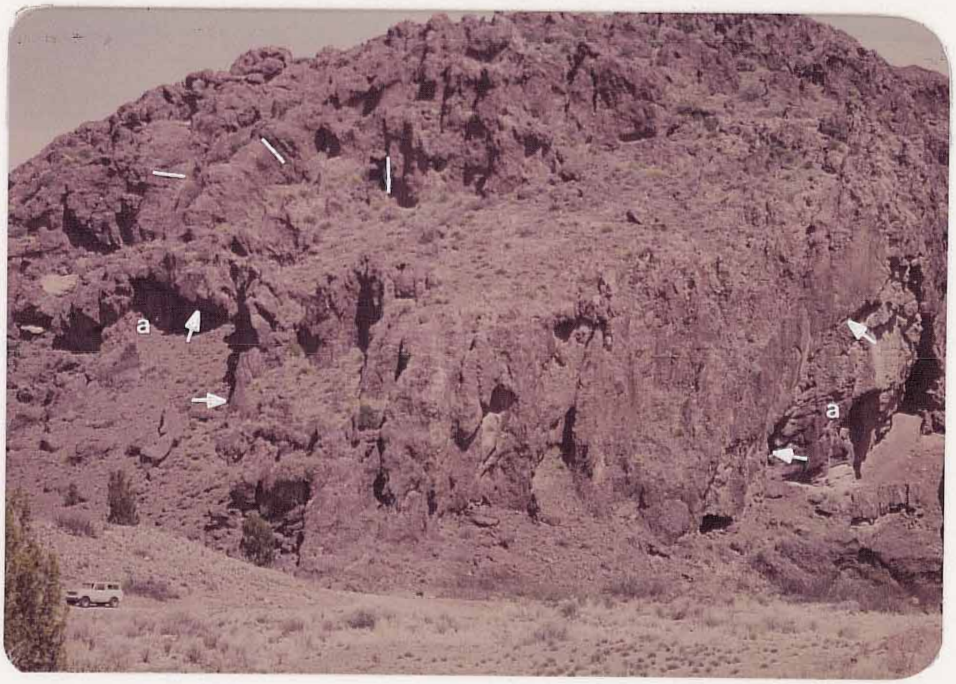


Figure 21. Southwestern wall of Bear Canyon Box showing dissected vent for an andesitic lava flow (Tla_3) in the Luis Lopez Formation. The walls of the vent (arrows) are formed by crudely bedded andesitic agglomerate (a) probably representing remnants of a cinder cone. The traces of subtle flow foliations (lines) are locally expressed by fractures along the weathered surface. Flow foliations fan out rapidly from nearly vertical above the vent to nearly horizontal outside of the vent. At the left, the base of the flow forms an "S" shaped flow fold where it breaks over the lip of the cinder cone. At lower right, light-colored rhyolitic tuffs ($?Tlt_1$) and an interbedded, ledge-forming andesitic laharic breccia and sandstone, underlie the agglomerate. Note vehicle on left for scale.

large phenocrysts are partially to completely replaced by pinkish, vuggy, potassium feldspar. About 2-3 percent of the rock is formed by black, lath-shaped, skeletal crystals of magnetite pseudomorphic after pyroxene or possibly hornblende. Less than one percent of the rock consists of crystals of colorless, fresh, feldspar identified as sanidine in thin section. A few percent of euhedral dark colored mica are common in handspecimen. In thin section, these are mixed crystals of colorless muscovite and reddish-brown biotite with abundant magnetite inclusions. They are ragged to euhedral in outline and some occur as very small wisps disseminated through the matrix. The pinkish potassium feldspar and micas are interpreted as hydrothermal minerals formed by potassium metasomatism. The coarse grain size of the "plagioclase" phenocrysts, and the presence of minor sanidine phenocrysts and possibly some hornblende, suggest the rock was originally a rhyodacite or mafic quartz latite.

On the north side of the East Roadcut, small faults, which repeat the upper contact of the flow, have given it an intrusive appearance with regard to the overlying bedded tuff (Tlrst). At the west side of the roadcut a thin conglomeratic sandstone containing clasts of porphyritic andesite is present at the base of the tuff, verifying this as a depositional contact. Coarsely porphyritic dikes (Tiap) at the Big Roadcut and a large plug of similar texture at the north end of Bear Canyon are the most likely source of the rhyodacitic "Tlap" flow.

The uppermost intermediate lava (Tla_4) occurs as small erosional remnants resting on rhyolite lava and tuffs on the north flank of the Blue Canyon dome (fig. 3). It is dark gray and moderately porphyritic. A small exposure of an altered andesite porphyry dike in the Pathway Canyon fault zone just below Big Cliff may have been the source of this flow. Appropriately, this dike appears to cut a north trending rhyolite dike that is petrologically equivalent to the rhyolite dome underlying the "Tla_4" flow.

The intermediate lavas of the Luis Lopez Formation are readily interpreted as a consanguinous rock series. They are believed to represent periodic tapping of a magma body of initial andesitic composition as it progressively crystallized and differentiated to a rhyodacitic or quartz latitic magma. The relationship of the intermediate magma body to an apparently contemporaneous rhyolitic magma body, in the same area under the cauldron block, is an intriguing and unresolved problem.

<u>Rhyolite lavas (Tlrb, Tlrs) and tuffs (Tlrbt, Tlrst).</u>

Eruption of pyroclastic followed shortly thereafter by rhyolite lavas and domes, are considered to be characteristic of the final stage of moat-filling volcanism in resurgent cauldrons (Smith and Bailey; 1968, p. 641). The Socorro cauldron also followed this pattern.

Rhyolite domes and flows that are typically underlain by petrologically equivalent, or similar, air-fall and ash-flow tuffs, generally cap the Luis Lopez Formation (fig. 17). Two geographically limited petrologic suites of rhyolite lavas and tuffs are recognized on the basis of phenocryst assemblages. One suite of moderately porphyritic, sanidine- and quartz-bearing rhyolite, seems to be associated with the cauldron margin. It will be informally referred to as the rhyolite of Blue Canyon (Tlrb, Tlrbt; pl. 1) after its most prominent exposure, which forms the Blue Canyon dome (fig. 3). The second type of rhyolite lavas and tuffs consists of a sparsely porphyritic rhyolite containing sanidine, plagioclase, and biotite with only traces of quartz. These crystal-poor rhyolites appear to be restricted to the vicinity of the resurgent block in the Chupadera Mountains, and will be informally referred to as the rhyolite of Highway Sixty (Tlrs, Tlrst; pl. 1) for exposures near the East Roadcut.

The original distribution of these rhyolite suites in the Luis Lopez Formation cannot be determined because the cauldron-fill deposits are largely buried under younger strata of the Santa Fe Group.

In the Valles caldera of northern New Mexico, three petrologic types of rhyolite domes are well exposed within the caldera. Two of these rhyolite suites, the Valle Grande Member and the Redondo Creek Member (Smith and others, 1970), have spatial relationships and phenocryst variations that are similar to those between the rhyolite of Blue Canyon and the rhyolite of Highway Sixty respectively.

<u>Rhyolite of Blue Canyon (Tlrb, Tlrbt)</u>. The rhyolite of Blue Canyon is divided into two units consisting of rhyolite lavas (Tlrb) and cogenetic tuffs (Tlrbt). The tuffs consistently crop out below the lavas where both units are exposed. The lavas generally weather to moderate-reddish-brown cliffs and resistant craggy outcrops where deeply dissected by the present erosion surface. Flow foliations, defined by differential weathering of the matrix, are readily visible on the weathered outcrops and therefore fresh handspecimens tend to look massive. Flow folding is not common except near the tops of the flows. Most outcrops are finely banded and the foliation surfaces are planar to broadly curviplanar. Near the vents, foliation planes tend to become more widely spaced and the outcrops more massive

in appearance. Roughly concentric patterns of steeply dipping flow foliations indicate the presence of a local vent under the Blue Canyon dome and the largely faultbounded outcrop north of "M" Mountain.

Lavas of the rhyolite of Blue Canyon typically contain 15 to 25 percent phenocrysts of sanidine and quartz. Near the base and outer edge of the flows, the phenocryst content may be as low as 5 to 10 percent, and near the center and top of the flows the phenocryst content is as much as 30 percent. Sanidine phenocrysts are medium grained (1-3 mm), and typically form blocky euhedra. Usually the sanidines are partially altered and have a pearly-white to pinkish-white color. Incipient alteration along traces perpendicular to their long axes, gives the altered sanidine crystals a lamellar pseudo-perthic appearance. Sanidine is normally about three to five times as abundant as quartz. Quartz phenocrysts are generally small (0.5-1 mm) and range from euhedral to rounded. Small lath-like wisps of disseminated magnetite or hematite, suggestive of biotite pseudomorphs, have been locally observed in the lavas. An atypical outcrop at the north end of the long cliff-forming exposure south of Blue Canyon does contain very sparse biotite; however, the sanidines are also more argillized than normal suggesting that the biotite may be hydrothermal.

The rhyolite lavas are consistently lithoidal and usually colored pale-red or light-red on an unweathered surface. Moderate-reddish-orange, grayish-red and rarely light-gray colors have been locally observed. With a hand lens, the matrix often has a fine silty or granular appearance. Small, pale-red spherulites and milky-white stringers of silica are also common in the matrix.

A sample from the rhyolite dome north of "M" Mountain (77-5-6, Table 5) is considered to be reasonably representative of the lavas in the rhyolite of Blue Canyon. In a thin section of this sample, the sanidine appears mostly euhedral, tabular and only rarely exhibits carlsbad twins. All the sanidines show various degrees of alteration of two distinct types. The dominant alteration occurs in a lamellar fashion along crystallographic planes and appears to consist of submicroscopic to microcrystalline reddish material. The lamellar alteration may represent hematite exsolution or alteration to a low-birefringent clay mineral. Patches of a moderately birefringent clay mineral, probably illite or possibly sericite, occur with the lamellar alteration but are less abundant.

Quartz phenocrysts are mostly euhedral with hexagonal forms common. A small fraction of the quartz phenocrysts show evidence of resorption in the form of rounding or embayments. The matrix consists of a patchy mixture of spherulites and polycrystalline quartz that appears to replace the spherulites. Lobate to slightly dendritic masses of finely disseminated hematite, 1-2 mm across, cover about 5 percent of the slide. Small subequant crystals of magnetite, or possibly ilmenite, occur in clusters in the matrix and are unrelated to the hematite stains.

A single chemical analysis and the observed phenocryst mineralogy are consistent with the classification of this rock, from the dome north of "M" Mountain, as a high-silica rhyolite (78 percent SiO_2). High K_2O content (8.3 percent) and low CaO (0.1 percent) suggest that the rock has been potassium metasomatized. Most of the additional potassium would probably be contained in the matrix. The sanidine phenocrysts, which were clearly sites of potassium concentration before the alteration, are only moderately altered and therefore do not seem to be significant sites of potassium addition. The general absence of plagioclase in the rhyolite of Blue Canyon suggests that the initial lime content of these lavas may have been negligible.

The tuffs of the rhyolite of Blue Canyon (Tlrbt) consist largely of numerous bedded air-fall tuffs with intertonguing, thin intervals of partially welded to densely welded ash-flow tuffs. Welded tuff intervals typically form resistant cliffs and ledges, while the intimately associated bedded tuffs form recessive slopes. The tuffs generally form light-gray to white outcrops except where hydrothermally altered. Bluish-green propylitized tuff outcrops north of Blue Canyon have provided the canyon with its name. Some outcrops southwest of the Middle mine in Pathway Canyon are reddened because of secondary silicification.

The most revealing and thickest exposure of the tuffs of the rhyolite of Blue Canyon is south of Big Cliff at the top of Pathway Canyon. Here, the bedded tuff section is about 98 m thick and contains two welded tuffs. A lithic-rich, pumiceous, partially welded tuff, about 20 m thick, forms a cliff at the base of the section at the southern end of the exposure. Below the cliff, the indurated base of the tuff is well exposed in a small downfaulted block. Here the welded tuff rests in sharp contact on an ancient colluvial(?) breccia developed on the top of the underlying massive porphyritic rhyodacitic flow (Tla_3). Blocks of dark-purplish-gray rhyodacitic porphyry, as much as 15 cm across, are concentrated along the base of the tuff and smaller fragments of intermediate lavas are abundant throughout the tuff.

Previous investigators (Lasky, 1932; Smith, 1963) correlated this lithic-rich welded tuff (Tlrbt) with the similar lithic-rich welded tuff (Tlt_2) exposed at about the sare elevation on the north side of the Pathway Canyon fault. In this case, and most other instances, these similar stratigraphic horizons may be distinguished by differences in phenocryst mineralogy and types of lithic fragments present. The younger bedded and welded tuffs (Tlrbt) generally contain 5 to 15 percent phenocrysts of sanidine and quartz. Pinkish or pearly colors of the sanidine is a common trait to both lavas and tuffs of the rhyolite of Blue Canyon; but is not diagnostic when absent, since it is controlled by secondary alteration. Pumice lapilli in the "Tlrbt" tuffs are usually crystal-poor and only ocassionally are they aphyric. This is in contrast to the older lithic tuffs (Tlt_1 and Tlt_2), which very rarely contain phenocrysts in the pumice.

Approximately 80 m of slope-forming, light-gray, bedded air-fall tuffs overlie the lower cliff-forming welded tuff at the exposure south of Big Cliff. This slope is interrupted by a 5 m-thick ledge of welded tuff about 15 m below the top of the tuff section where it is overlain by a cogenetic rhyolite lava (Tlrb). The bedded tuffs are light greenish gray (altered), pumiceous, and contain abundant andesitic lithics and a few rhyolite lithics like the "Tlrb" lavas. Graded bedding is common and depositional units are as much as 2 m thick.

Southwest of the Middle mine, and at the top of the east-trending ridge south of the Blue Canyon dome, a lavender-colored, densely welded, flow-banded, crystal-poor tuff is exposed that, except for color, is essentially identical in appearance to the flow-banded A-L Peak Tuff. Pumice lineations at both of these exposures are aligned to the northwest (pl. 1). An established regional pattern of pumice lineations for the A-L Peak Tuff (C.R. Osburn, 1979, oral commun.) would require northeast-trending lineations in the Socorro Peak area. This flow-banded tuff unit in the rhyolite of Blue Canyon was most likely erupted from a local vent, possibly the vent now buried under the Blue Canyon dome. Bedded tuffs below this flow-banded tuff at the southern locality have been noted for their coarse spherulitic texture (Lasky, 1932; Lowell, 1967).

The northerly dip of these bedded tuffs along this ridge must be largely primary -- tectonic dips in this area are either westerly or easterly. These north dips are interpreted as reflecting a north-facing paleoslope on an underlying andesitic volcano (Tla_3) that was centered several hundred meters to the southwest. The southwestern two-thirds of this volcano has been removed from view (pl. 1) by downfaulting of the western side and by erosion and burial of the south flank in late Miocene time. Thin beds of entirely andesitic detritus (Tpsw), presumably shed from this volcano, are found interbedded in heterolithic early Miocene conglomerates (Tpsd) to the southwest of the exposed remnant. These conglomerates buried the constructional volcanic topography of the Blue Canyon rhyolite dome and presumably that part of the andesite volcano not previously overlapped by the dome.

Four local vents -- identified with varying degrees of confidence -- are recognized as the local sources of the discontinuous exposures of the rhyolite of Blue Canyon (pl. 1). The discontinuous nature of this stratigraphic unit is primarily a function of the extreme lenticularity of the originally steep-sided rhyolite domes. Penecontemporaneous movement on cauldron-related faults and later offset by rift faults have added to the original discontinuities.

The largest and best exposed of the rhyolite domes is the Blue Canyon dome. A steeply dipping and roughly concentric flow-foliation pattern indicates the presence of a vent under the dome a little south of the center of the outcrops (pl. 1). Northwest aligned stretch marks, perpendicular to flow direction, on the northeast flank of the dome (pl. 1), and a northwest oriented pumice lineation in the cogenetic flow-banded tuff on the southeast flank of the dome, are roughly consistent with radial flow away from the proposed vent. Structure section C-C' (pl. 2) indicates the rhyolite flow may be as much as 300 m thick above the vent. A minimum of 70 m of primary topographic relief is exposed along the west side of the Blue Canyon dome where it is onlapped by an equivalent thickness of lower Popotosa strata. At the time of deposition of these conglomerates and sandstones, the west flank of the dome sloped about 27 degrees to the west.

The Blue Canyon dome probably marks the position of the major ring fracture of the Socorro cauldron. The apparent north-south alignment of vents for the rhyolite of Blue Canyon (pl. 1) is an artifact of tilted fault-block structure, and burial under thick basin-fill deposits.

The most revealing exposure of a volcanic vent in the study area is found just south of Big Cliff. Here the tuff blanket most likely related to the Blue Canyon dome has been uplifted against the structural margin of the Socorro cauldron formed by the Pathway Canyon fault. Just south of the Pathway Canyon fault, a rhyolite dike (Tirb) cuts bedded tuffs (Tlrbt) in a fault that parallels the Pathway Canyon structure. Upslope, this dike intersects the top of the bedded tuffs and fans out into an eroded remnant of a rhyolite flow. The flow is about 15 m thick above the vent.

The flow is truncated about 100 m to the south of the vent where an ancient erosion surface cuts across the edge of the flow and down into the underlying tuffs. This ancient surface is marked by a colluvial breccia (breccia symbol, pl. 1), about 12 m thick consisting of blocks derived from the lava (Tlrb). The breccia is arbitrarily mapped with the rhyolite of Blue Canyon as an undesignated map unit, but more properly belongs with the Popotosa Formation. A small, lenticular outcrop of equivalent monolithic breccia, consisting entirely of clasts of the rhyolite of Blue Canyon occurs below Big Cliff on the north side of the Pathway Canyon fault (mapped as Tlrb breccia, pl. 1).

Rhyolite dikes (Tirb) that are petrologically equivalent to lavas of the rhyolite of Blue Canyon, are exposed at several locations in the Pathway Canyon area. Only the feeder dike described above is clearly cogenetic with the lavas, the rest are assumed to be cogenetic on the basis of lithologic correlation. The "Tlrb" dikes are usually flow banded and reddish colored like the "Tlrb" lavas. The matrix of dikes of deeper erosion levels -- north-trending dike below Big Cliff and dikes near Wood's Tunnel -- generally exhibit a very fine saccaroidal texture. Phenocrystic sanidine, usually colorless or white, and quartz are characteristic of the dikes. The dikes commonly form resistant outcrops and the intrusive contacts are usually covered by small unmapped patches of colluvium. Dikes in the down-faulted blocks, such as the one west of the Middle mine are most likely cut off from their original source by younger rift faults (pl. 2, B-B').

Another rhyolite dome of the rhyolite of Blue Canyon crops out to the northeast of "M" Mountain. It is bound by two faults and a steep westerly dipping depositional contact that is overlain by the tuff of South Canyon. These steep contacts falsely give the exposure the appearance of a large volcanic neck. Crudely concentric flow foliations suggest the presence of a vent under the center of the dome. A small outcrop of gray tuffs under the lava at the eastern end of a deep ravine that bisects the dome attests to the fact that the present erosion surface is above the Oligocene ground surface onto which the tuffs and lava dome were erupted.

Three north-trending outcrops of the rhyolite of Blue Canyon occur west of Socorro Spring. These appear to represent the exhumed crest (or crests) of another domal flow (or flows) buried by conglomerates of the lower Popotosa Formation. The long cliff and ridge that forms the largest of these exposures has steeply dipping flow structure at its north end. The north end of this cliff may be on the periphery of another local rhyolite vent that is buried under the down-faulted blocks of Popotosa Formation northeast of this ridge.

Rhyolite of Highway Sixty (Tlrs, Tlrst). Rhyolite lavas (Tlrs, pl. 1) and tuffs (Tlrst, pl. 1) informally designated here as the "rhyolite of Highway Sixty" crop out in the northern Chupadera Mountains at the East Road Cut and west of Chupadera Cliff. The rhyolite tuffs of Highway Sixty consist of interbedded air-fall tuffs and thin, local, ash-flow tuffs. They are generally distinguished from other tuffs in the Luis Lopez Formation by conspicuous, small, black or coppery biotite phenocrysts that normally form about 0.5 percent of the rock. Small glassy sanidine phenocrysts (0.5-2 mm) and plagioclase phenocrysts (generally altered to white clay) occur subequally and make up about 3 to 5 percent of these tuffs. Small, rounded quartz phenocrysts are rarely present. Altered plagioclase is visible on the weathered surface as stubby lath-shaped euhedral vugs, but is nearly invisible on the freshly broken surface of the white tuffs.

Approximately 37 m of white to light-gray bedded tuffs are exposed in the south side of the East Road Cut. Here, the tuff unit is stratigraphically bound by the porphyritic rhyodacite (Tlap) and the overlying flow-banded rhyolite lava (Tlrs). The road cut provides an excellent exposure in these slope-forming tuffs that are largely mantled by a pale-brown tuffaceous soil to the north and south of the cut. The bedded tuffs at the road-cut are moderately indurated, pumice-poor, and lithic-poor. They contain less than one percent of small, red, andesitic, lithic fragments and a few red, finely flow-banded, rhyolitic lithics similar to the overlying lavas (Tlrb). Two relatively resistant intervals about 1 to 2 m thick near the top of the tuff sequence are moderately welded ash-flow tuffs with a few percent of medium-light-gray, sandy-looking pumice. The tuffs are parallel bedded and vary from thick-bedded massive nonwelded(?) ash-flow tuffs(?) to thin-bedded air-fall tuffs, some of which exhibit graded bedding.

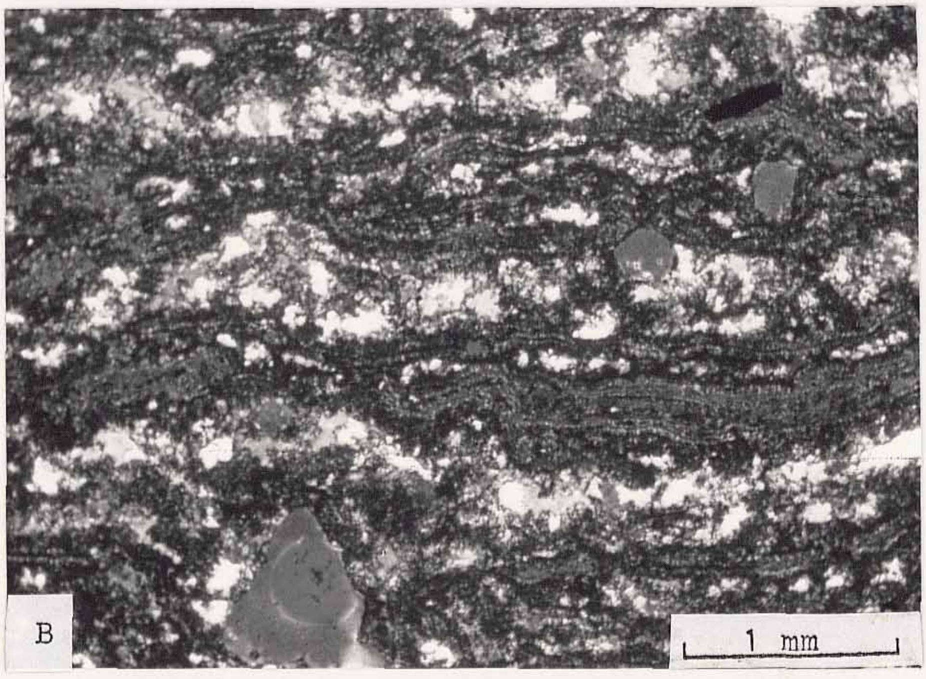
The tuff exposures (Tlrst) west of Chupadera Cliff are largely formed by a light-gray, crystal-poor, lithic-rich, partially welded ash-flow tuff. It contains abundant lithic fragments of the underlying lava flows (Tla_3 and Tlap) and 5-10 percent, small, medium-gray, pumice with the vesicular structure well preserved. This lithic-rich tuff is about 20 m thick and appears to be preserved in an erosional remnant of a north-trending paleovalley (pl. 2, G-G').

The crystal-poor rhyolite lavas of the Highway Sixty Member are typically light gray to pinkish gray and are characterized by a finely laminated flow-banding (fig. 22). Small scale contorted flow-banding and flow folds are commonly visible both in handspecimen and outcrops. The flow-banding is defined by gray cherty silica along the shear planes. The lavas crop out as low, rounded hills and often weather to a colluvial soil of platy fragments. These platy, finely flow-banded rock fragments found on modern surfaces have similar distinctive analogs of early Miocene age. Small remnants of the well-indurated lower Popotosa conglomerates (Tpsd) occur on top of the flow-banded lavas to the north and south of the East Road cut. These conglomerates grade down into an ancient colluvial breccia consisting entirely of platy lava fragments in a red siliceous matrix. Red silica in clastic dikes and veinlets are readily observed in the light gray rhyolite lava and tuff outcrops where they have been exhumed from under the siliceous red facies (Tpsd) of the lower Popotosa Formation.

A small outcrop of crystal-poor flow-banded rhyolite is exposed in a low hill southeast of the East Road Cut. Concentric and steeply dipping foliations indicate that this was a vent or crest of a small exogenous dome. This vent

Figure 22. Rhyolite of Highway Sixty Member of the Luis Lopez Formation. A, Typical finely laminated flow-banded rhyolite lava. The flow layering is locally flow folded but generally is steeply dipping. Location is vent southeast of East Roadcut in northeastern Chupadera Mountains. Knife is 9 cm long. B, Photomicrograph of convolute flow banding in rhyolite lava. Main flow bands are defined by microgranular quartz (white) that form weather-resistant light colored ridges in above photograph. Compare continuity of flow laminae with Figure 10B. Sparse phenocrysts of biotite (black) and feldspar (light gray) are present. From hilltop south of East Roadcut. Partly crossed nicols.


A


B
1 mm

is considered to be the source of the tuffs and lava flow exposed at the East Cut, even though it is now physically isolated from them by later faulting and erosion. The large flow-banded rhyolite intrusive (Tirl) that forms Tepee Town Mountain is lithologically similar to the rhyolite of Highway Sixty and may be another possible source, particularly for the exposures west of Chupadera Cliff.

Chemical analysis and petrographic study of an unusually fresh sample (Table 5, no. 77-5-2) of densely welded tuff from the hillside south of the East Road Cut both indicate that the rhyolite of Highway Sixty is a high-silica rhyolite. The mineralogy and texture of this rock are also described in Table 5. In thin section, some of the plagioclase may be mistaken for sanidine because the plagioclase (An_{31}; $2V_x$ = 75 degrees, $X' \wedge 010$ = 16 degrees) exhibits very fine albite twinning and is not noticeably zoned. Therefore, when viewed at high angles to the twin planes, the plagioclase appears to be a homogeneous feldspar readily assumed to be sanidine. Hand specimens of hydrothermally altered lavas and tuffs from the same area, which contain both altered plagioclase and fresh sanidine, confirm the greater abundance of plagioclase.

Rhyolitic breccias (Tlrsx) and rhyolitic sandstones (Tlsr)

Two distinctive units of volcaniclastic sedimentary rocks (Tlrsx and Tlsr) locally rest on the uppermost lava flows (Tlrs and Tla_3) and tuffs (Tlrst) of the caldera-filling Luis Lopez Formation. The inclusion of these sedimentary rocks with the Luis Lopez Formation is somewhat arbitrary, since the breccias and sandstones apparently record a period of normal faulting, erosion and sedimentation that post-dates the caldera-filling volcanism.

The rhyolitic breccia unit (Tlrsx) is interpreted to be dominantly an ancient colluvial deposit derived from erosion of an Oligocene fault scarp that developed along an east-northeast-trending normal fault (pl. 2, west end of section C-C'). The colluvial deposit is derived mostly from the rhyolite of Highway Sixty (28.6 m.y.) and is overlain by the tuff of South Canyon (26.2 m.y.) thereby dating the fault as late Oligocene in age.

The main area of exposure of the rhyolite breccia and associated fault lies halfway between the Tower mine and Chupadera Cliff (pl. 1). The rhyolite breccia is about 60 m thick on the southern, downthrown, side of the fault. The base of the breccia unit appears to lap unconformably across the fault plane -- locally exposed at lower elevations on the hillsides -- northward onto an eroded scarp.

The scarp cuts downward (pl. 1) across depositional contacts between older rhyolite lavas (Tlrs), tuffs (Tlrst) and andesite flows (Tla_3). Most of the rhyolite breccia is preserved in a small, west-trending, graben-like block. The southern bounding fault of this downthrown block truncates the breccia and younger overlying units and must have been active after the breccia was deposited.

Outcrops of the rhyolite breccia are massive and unstratified. Angular fragments of light-gray and pinkish-gray flow-banded rhyolite lava -- mineralogically and texturally identical to the lavas of Highway Sixty (Tlrs) -- are usually the only clasts present. The monolithic fragments range from blocks as much as a meter across down to a centimeter or less and comprise from 50 to 90 percent of the outcrop. The remaining fraction consists of a grayish-red to reddish-brown muddy(?) sandstone matrix that is much like the rhyolitic channel sandstone (Tlsr), which locally overlies the breccia unit. The breccia is cemented with silica and moderately well indurated. Small resistant outcrops are common on the west-facing hillside that forms a large part of the main exposure. Where mantled by slope wash derived from the breccia, the loose rhyolitic blocks commonly exhibit indurated sandy matrix adhered to the lava fragments. A small segment of the trace of the east-northeast-trending fault genetically related to the breccia deposit is moderately well exposed on this west facing hillside. Outcrops immediately south of this fault trace are unusual in that they contain a few fragments of andesitic lava. The mafic lava fragments are texturally similar to the flow (Tla_3) that forms the north wall of the fault; these andesitic fragments are thought to be transported only a short distance from where the upthrown block is truncated by the ancient erosion surface.

Stratigraphic relations at the top of the breccia unit change abruptly over a short distance. The main exposure preserved in the west-trending graben-like block is overlain by about 24 m of red rhyolitic sandstone and in turn by the tuff of South Canyon. The breccia to sandstone contact is not well exposed. The breccia is probably massive and unstratified up to the contact. Northwest of the graben-like block, the rhyolitic breccia appears to have been preserved on the downthrown, west side, of a north-trending normal fault. Here the upper 5 m of the breccia is crudely stratified. A dark-red, well-indurated, massive mudflow about 2 m thick and containing angular gray rhyolitic pumice fragments occurs at the top of the breccia unit. The top of this pumiceous mudflow is well exposed where it is overlain by typical well-indurated heterolithic mudflow conglomerates (Tpsd) of the lower Popotosa Formation. The reason(s) for these abrupt changes in stratigraphic relationships are not readily apparent but most likely involve complex paleotopographic relationships controlled by contemporaneous faulting, erosion, and deposition.

A small outcrop of rhyolite breccia (Tlrsx) occurs below the red rhyolitic sandstone (Tlsr) in exposures about 1 km south of the Big Road cut. The breccia, sandstone and overlying tuff of South Canyon appear to fill an ancient channel of northeasterly trend. The approximate trend of the channel wall is provided by the wedging out of the unit s onto it. The channel was later tilted to the northwest and may have been unconformably overlain by the lower Popotosa Formation, although these contact relations are not certain here. The small breccia outcrop now exposed was near the channel axis.

The rhyolitic sandstone (Tlrs), as previously indicated, occurs in paleochannels generally overlying the breccia unit and underlying the tuff of South Canyon. The sandstone may be as much as 24 m thick and is typically reddish brown, well indurated, and thin bedded to laminated. Parallel stratification is dominant and cross bedding is not apparent. However, the general transport direction of the paleochannels may be inferred from approximate limits on the configuration of the channel walls (from map relations, pl. 1) and from composition of the sandstone. The channel wall at the northern exposure trends generally northeast and at the southern exposure (in the graben like block) to the northwest. The sandstone is medium to fine grained and consists of abundant grains of quartz and "glassy" sanidine and sparse grains of plagioclase(?) altered to chalky white clay, and traces of biotite. This mineralogy is only consistent with derivation of the sandstone from a crystal-rich ash-flow tuff, such as the cauldron facies Lemitar(?) Tuff. The cauldron facies tuff can reasonably be assumed to have been exposed on the resurgent block, south of Black Canyon, at the time of deposition of the sandstone. This time of deposition would be following resurgence and filling of the adjacent depression in the cauldron. Thus, it appears reasonable that the rhyolitic sandstone filled a channel or channels that flowed northerly away from the resurgent block of the Socorro cauldron.

Related intrusive rocks (Tia, Tiap, Tirb, Tirl)

Numerous intrusive bodies of andesitic to rhyolitic composition (Tia, Tiap, Tirb, Tirl) that are shown on the geologic map (pl. 1) have been correlated -- at significantly different levels of confidence -- to the eruptive rocks of the Luis Lopez Formation. The correlations are based primarily on petrographic similarity of the intrusive, to the flows and tuffs of the Luis Lopez Formation. Most of these intrusions occur as dikes that cut slightly older to penecontemporaneous strata of the Luis Lopez Formation. Thus, a close spatial alliance is apparent on the geologic map between the intrusive bodies and their inferred equivalent volcanic flows. These dikes do not anywhere cut strata younger than those of the Luis Lopez Formation and many are demonstrably buried unconformably by the lower Popotosa Formation (fig. 23). South of Big Cliff on the east face of Socorro Peak, the present erosion surface cuts across a rhyolite dike (Tirb) and upwards across the flow (Tlrb) fed by the dike. This continuous exposure from intrusive to extrusive rhyolite clearly documents their equivalency. Northeast of Black Canyon a similar, though less well exposed, feeder-dike relationship is apparent between a moderately porphyritic andesitic dike (Tia) and a equivalent textured andesitic flow (Tla_3).

Dark-gray to dark-purplish-gray dikes mapped as "Tia" are commonly aphanitic to moderately porphyritic rocks containing small phenocrysts of hydrothermally altered plagioclase and altered ferromagnesian minerals. Prior to hydrothermal alteration, they were probably andesitic in composi-

Figure 23. Andesitic dike (Tia) of the Luis Lopez Formation, cutting upper lithic-rich tuff (Tlt_2) of the Luis Lopez Formation and unconformably overlain (arrow) by a bouldery mudflow deposit (Tpsd) of the lower Popotosa Formation. Dike is approximately 10 m thick. At old road cut of Highway Sixty in western Socorro Canyon.

tion and composed dominantly of plagioclase and pyroxene. The dark-colored dikes have textures and compositions that indicate they may have been feeder dikes for any of the intermediate flows (Tla_1, Tla_2, Tla_3, Tla_4), except for the coarsely porphyritic lavas (Tlap).

Most of the andesitic dikes occur between Black Canyon and Socorro Canyon on the large block of Luis Lopez Formation that forms the northwest end of the Chupadera Mountains. The andesitic dikes are generally 5-20 m in width. They typically stand only slightly higher than the adjacent wall rocks and usually the intrusive contacts are covered by small patches of loose andesitic debris. The 600-m-long dike (Tia) northeast of Black Canyon is highly visible as a dark colored band where it cuts across the lighter-colored lithic tuff (Tlt_2). To the west of the tuff outcrop, this moderately porphyritic dike extends as an erosion resistant wall about halfway up the dip slope of aphanitic lava (Tla_2) before it disappears under colluvium derived from the aphanitic lava. Along the same trend of this long dike, about 250 m further west, is exposed a composite dike consisting of dark-gray aphanitic andesite(?) cut at a slight angle by a younger reddish-brown rhyolite porphyry. It appears from these field relations that the same northeast-trending fracture may have been used as a conduit for three different eruptive events in this area.

Several dark-colored dikes and one irregular shaped dike-like body of intermediate composition are exposed in the new and old road cuts of Highway Sixty between Bear Canyon and Sixmile Canyon. The old highway cuts are along the south side of Socorro Canyon. Sharp, and locally chilled, phenocryst-poor contacts of the dikes are well exposed where the dikes cut the upper lithic-rich tuff (Tlt_2). Foliations in the lithic tuff are locally dragged upwards subparallel to the dike walls, indicating that the dikes were forcefully injected. Varying degrees of silicification of the tuffs occurs adjacent to the dikes.

Two coarsely porphyritic rhyodacitic(?) dikes (Tiap) are exposed toward the east end of the road cuts. The longer east-northeast-trending rhyodacitic(?) dike exhibits steeply plunging slickensides on fractures near the southern wall of the dike where it is exposed in the Big Roadcut. Some movement has clearly occurred along the dike trend since its injection, but it is believed to be minor. No significant offset is visible of the contact that unconformably truncates this dike at the west end of the Big Roadcut. This dike appears to end against a poorly indurated andesitic breccia with a sandy mud matrix. The breccia, interpreted as an ancient colluvium, is in turn unconformably overlain by heterolithic mudflow deposits of the lower

Popotosa Formation (Tpsd). The long porphyry dike probably cuts the moderately porphyritic flow (Tla_3) overlying the tuff (Tlt_2), although the relationships are masked by colluvium. Slope wash has also made it uncertain whether or not the short north-northeast trending rhyodacitic(?) dike actually joins with the longer dike. The rhyodacitic dikes contain abundant phenocrysts of altered plagioclase, as much as 1 cm in length, black iron oxide casts of elongate laths of an altered ferromagnesian -- possibly hornblende, locally some small fresh sanidine crystals, and a phenocrystic-looking euhedral black mica. As previously discussed, radiometric dating and thin-section examination of the mica indicate that it is a secondary hydrothermal phase. Some plagioclase phenocrysts are replaced by pink, spongy, potassium feldspar, but most are altered to a white clay.

Several volcanic vents for intermediate lavas in the Luis Lopez Formation that are shown on the geologic map (pl. 1) locally exhibit steeply dipping flow structure indicative of intrusive relations. However, the exact boundaries of the intrusive versus extrusive portions of these outcrops are often either uncertain, or difficult to portray on the geologic map. Therefore, the areas immediately surrounding the vent symbols may be regarded as intrusions even though the boundaries of the intrusions are not shown.

Rhyolitic intrusions (Tirb, Tirl) considered to be correlative with rhyolite lavas and tuffs of the Luis Lopez Formation are spatially associated with west-northwest-trending cauldron-margin structure on the east face of Socorro Peak; rhyolitic intrusions also occur along the major east-northeast trending fault zone that generally bounds the resurgent block of cauldron-facies Lemitar(?) Tuff exposed to the south of Black Canyon.

Rhyolite dikes (Tirb) petrographically similar to the rhyolite lavas of Blue Canyon (Tlrb) have been briefly discussed with the description of these lavas. Dikes of this designation are mostly grayish-red to pale-red, moderately phenocryst-rich rocks, and are petrographically similar to the Blue Canyon lavas in their content of pink sanidine and small quartz phenocrysts. These rhyolite dikes are generally erosion-resistant and locally form "walls" cutting across slopes of the less resistant host rocks. They range from 10 to 40 m in width and commonly exhibit flow foliations parallel to their local outcrop trends. Small patches of colluvium typically cover the intrusive contacts of the rhyolite dikes at the edges of their outcrops. None of these dikes, including the very prominent feeder dike south of Big Cliff, were recognized as such in previous studies (Lasky, 1932, Smith 1963, Burton, 1971).

Dikes correlated with the rhyolite of Blue Canyon occupy faults that commonly show significant stratigraphic offset. The dikes in down-to-the-south, west-northwest-trending faults are readily interpreted as ring-fracture dikes of the Socorro cauldron. The prominent feeder dike also exhibits some brecciation along its north margin indicating minor movement after the dike was emplaced. Rhyolite dikes of the Blue Canyon lithology also intrude steeply dipping, down-to-the-west, north-trending faults, thereby indicating that this structural trend was active during the cauldron evolution.

A sill-like body of white rhyolite porphyry (Tirl) that intrudes the Pennsylvanian rocks north of the cauldron margin on Socorro Peak is here tentatively correlated with the Luis Lopez Formation rhyolites, primarily on lithologic grounds. The white color of this rhyolite is most likely caused by intense argillic alteration and bleaching related to oxidation of associated pyrite. The white rhyolite is typically flow-banded and ranges from aphyric at the contacts to moderately porphyritic in the core.

In thin section, a sample of the core rock contains 10 percent phenocrysts of stubby, prismatic, argillized sanidine(?) (1-3 mm) mostly as washed out holes but containing some bits of gray, isotropic clay and 3 percent subhedral and embayed quartz phenocrysts (0.5-1.5 mm). The groundmass consists of a microgranular mixture of silica and feldspar(?). The silica occurs in irregular patches that are disturbingly similar in appearance to non-welded glass shards; however, on close examination the apparent shard boundaries are much too ragged to be broken glass fragments. Darker-colored flow laminae are defined by concentrations of finely disseminated opaques (iron oxide ?).

Very fine ($<$ 0.5 mm) specks of yellowish-brown iron oxide are common in the white rhyolite, particularly near the margins and in the adjacent wall rocks. The specks often have cubic outlines when viewed with a hand lens, and are considered to be oxidized pyrite. Oxidation of pyrite probably accounts for the bleaching and intense argillization of this rhyolite, which is otherwise petrographically similar to the rhyolite of Blue Canyon.

It does not seem likely that the white rhyolite intrusive is equivalent to the silicic magmas of the late Miocene Socorro Peak Rhyolite. Several vents and dikes of the Socorro Peak Rhyolite occur nearby at "M" Mountain and on the north flank of the Blue Canyon dome. However, these intrusives contain only phenocrystic plagioclase, hornblende and biotite. Vents and intrusions of rhyolitic quartz-porphyry lavas of the Socorro Peak Rhyolite are not closer than 3 km to the exposures of the white rhyolite intrusion.

The white rhyolite intrusive is sill-like, rather than a true sill or dike. It occupies a minor fault that cuts at a low-angle across shales of the upper Sandia Formation. The intrusion terminates on its south end where the structure intersects the base of the Madera Limestone. The intrusive contacts are well exposed at this southern terminus.

The white rhyolite body dips about 30 to 45 degrees to the west, approximately conformable with the Pennsylvanian strata. It typically forms resistant ledges and low cliffs surrounded by locally derived colluvium, particularly where it intrudes the Sandia Formation. The intrusive body pinches and swells along strike and has a maximum thickness of about 30 m at its most prominent outcrop northwest of Wood's Tunnel. Towards the north end of the face of Socorro Peak, the sill-like body is repeated several times by east-dipping low-angle normal faults. The low-angle faults are most likely rotated early-rift faults (Chamberlin, 1976, 1978) of late Oligocene or early Miocene age; this structural relationship favors the interpretation of the white rhyolite as being approximately equivalent in age to the Oligocene Luis Lopez Formation as opposed to the most likely alternative, which would be a late Miocene age, equivalent to the Socorro Peak Rhyolite.

A series of east-northeast trending rhyolitic dikes (Tirl) and a large plug-to-dike-shaped rhyolitic intrusion (Tirl), all of which cut cauldron-facies Lemitar(?) Tuff in the vicinity of Black Canyon, are here tentatively correlated with the rhyolite of Highway Sixty Member of the Luis Lopez Formation. In general, both the rhyolite of Highway Sixty and the proposed equivalent rhyolitic intrusions contain phenocrystic sanidine, plagioclase, biotite, and quartz. In comparison to the lavas and tuffs, the rhyolitic intrusions locally contain more phenocrysts, ranging from 2 percent to 10 percent, and locally exhibit phenocryst ratios significantly different from the lavas. Many outcrops of the rhyolitic intrusions appear to contain more alkali feldspar than plagioclase. However, the alkali feldspar to plagioclase ratio is uncertain in most outcrops because of hydrothermal alteration, which often appears to effect both feldspars. Fresh plagioclase has not been observed in any of these rhyolitic intrusions, but plagioclase altered to chalky-white clay is clearly present in some otherwise fresh specimens that contain "glassy", alkali feldspar. Quartz phenocrysts locally form a few percent of the intrusive rocks and are particularly conspicuous in the core of the plug-like intrusion.

Assuming that the rhyolite of Highway Sixty and the rhyolitic intrusions near Black Canyon are cogenetic, then the differences in crystallinity and phenocryst proportions could be explained as variations in magmas erupted from different levels of a compositionally zoned magma chamber. The typically phenocryst-poor early erupted magmas may be represented by the flows; and later magmas (more phenocryst "rich") may have been preserved in the shallow intrusive environment but not in the eroded extrusive equivalents.

Other relationships of the rhyolitic intrusions exposed in the vicinity of Black Canyon also support their correlation with the Luis Lopez Formation. These rhyolitic dikes consistently occur in east-northeast trending faults with significant down-to-the-north displacements (pl. 1). The largest dike and a contiguous plug-like intrusion north of Tepee Town Mountain occur along the major fault that forms much of the north boundary of the resurgent block. The eastern end of the large dike clearly cuts intermediate lavas (Tla_2) of the Luis Lopez Formation. At the western end, the large dike consists of a silicified, very-crystal-poor, flow-banded rock with a few sparse altered feldspar phenocrysts. This silicified outcrop is considered to represent a near surface level of emplacement of the original rhyolitic dike. It has apparently been preserved on a block dropped down-to-the-west by a major north-trending fault. This silicified dike exposure appears to be overlain unconformably by the lower Popotosa Formation, but the contact is not well exposed. A few small clasts of white silicified rhyolite(?) occur in the Popotosa mudflow deposits. They are believed to have been locally derived, along with the clasts of cauldron-facies Lemitar(?) Tuff that are dominant here. Thus, the field relationships of these rhyolitic intrusions indicate they are younger than part of the Luis Lopez Formation, but older than the lower Popotosa Formation, which is essentially the stratigraphic position of the rhyolite of Highway Sixty.

Most of the rhyolitic dikes in the Black Canyon vicinity are light-gray to light-purplish-gray and characterized by 1 to 3 percent phenocrysts of altered feldspar that form euhedral prismatic clasts (1-3 mm long) on the weathered surface. With a hand lens, the matrix typically has a very fine granular texture, which was probably induced by recrystallization during silicification or potassium metasomatism, or both.

If the predominant altered feldspar is plagioclase, then its occurrence must be somewhat irregular, because some dike rocks locally contain both fresh alkali feldspar and an altered feldspar -- presumably plagioclase -- while other dikes locally contain only one feldspar identifiable as fresh alkali feldspar. The rhyolitic intrusions nearest to the Black Canyon Box also commonly contain sparse black biotite. In thin section, the biotite is homogeneous and pleochroic from greenish brown to light brown; this implies that it is a true phenocryst, rather than a secondary potassium-metasomatic mineral.

The rhyolitic dikes generally crop out as discontinuous low walls less than a meter high. The 3-km-long dike trend from Tepee Town Mountain to Section 8 Ridge is often broken by small patches of angular colluvium derived from the cauldron-facies tuff. Intrusive contacts here, and elsewhere, are commonly covered by colluvium. Most of the dike outcrops are 5 to 20 m wide.

The large plug-like body east of the Nancy mine, which is contiguous on its east and west side with unusually wide dikes, 60-150 m, generally contains more phenocrysts than the adjacent dikes. Chilled margins of the large intrusion contain about 3 to 5 percent phenocrysts consisting of sanidine, with minor altered feldspar and locally some biotite. Quartz phenocrysts increase in size and abundance toward the core of the plug. Here, quartz forms about 2 percent of the rhyolite and is accompanied by 5 to 8 percent sanidine and altered feldspar.

In thin section, a sample from the southwest side of the plug-like intrusion was found to contain approximately 8 percent phenocrysts consisting of about 6 percent unaltered alkali feldspar, 1.5 percent quartz, and 0.5 percent altered feldspar replaced by low birefringent clay. The groundmass consists of an allotriomorphic microgranular mixture of potassium feldspar, quartz and disseminated hematite. About 95 percent of the surface area of the alkali feldspar crystals exhibits what appears to be irregular, slightly wedge-shaped, polysynthetic twins. The $2V_x$ of both the twinned and untwinned areas, estimated on the flat stage, is about 25 to 35 degrees; thus the twinned feldspar is clearly not plagioclase. The close association of homogeneous and twinned areas, which occur in optically continuous crystals of seemingly constant index of refraction, suggests that the twinned phase is anorthoclase and the untwinned phase is sanidine. Small, irregular patches of highly birefringent sericite(?) replace about 5 to 10 percent of the alkali feldspar and matrix in the rock. Traces of zircon and sphene, partially altered to lexcoxene, are also present in this sample from the rhyolitic plug.

The geometry of the plug-like intrusion is not clear cut; it appears to have conflicting aspects of both a steeply plunging neck (on the northwest side) and a gently west-dipping, north-trending dike (along its eastern margin). Flow foliations in the large intrusion commonly exhibit swirling and folding, thus the foliations do not necessarily reflect the true attitude of the body. An intrusive contact is locally visible in a ravine wall along the eastern side of the body at an elevation of about 5640 feet. Here the contact has an apparent west dip of approximately 30 degrees. This gently west dipping contact is appropriate if the plug-like intrusion is considered to be a thick north-trending dike of Oligocene age that has been strongly tilted to the east about 40 degrees, a value commensurate with adjacent east-tilted strata of the Luis Lopez Formation. Thus the 600-m-wide outcrop of the plug-like intrusion, measured in an east-west direction, may be twice the true thickness of the intrusion if it actually dips gently to the west as a dike-like body. The situation would be similar to that shown in cross section I-I (pl. 2), except the westerly dip of the intrusion could be as low as 30 degrees.

The rhyolitic portion of the composite dike north of Black Canyon is anomalous in its freshness, color, and phenocryst content. This dense, grayish-red, dike rock contains about 10 percent medium-grained phenocrysts. In decreasing order of abundances the phenocrystic minerals consist of "glassy" euhedral sanidine, subhedral quartz, black biotite, and chalky white clay pseudomorphs apparently after plagioclase. The clay pseudomorphs are commonly enveloped by fresh sanidine.

The most tenuous correlation of an intrusive body with the Luis Lopez Formation is a poorly exposed, arc-shaped, outcrop of grayish-red-purple, hydrothermally altered, andesite(?) porphyry (?Tia), which occurs within the eastern side of the large plug-like rhyolitic intrusion at Black Canyon. The andesitic intrusive weathers readily and forms a terrace-like break in slope in the surrounding rhyolite outcrops. The occurrence of the andesitic body within a rhyolitic intrusive body seems to require that it is either a younger intrusion or a large xenolith. The latter does not seem likely, since no small xenoliths of any type were observed in the rhyolite outcrops.

The dark-colored intrusive generally exhibits trachytic alignment of phenocrysts, which form about 5 to 20 percent of the rock. Chalky, altered plagioclase laths 1 to 5 m long, are abundant. Sparse pseudomorphs of reddish-brown hematite apparently occur after a stubby prismatic phenocryst, probably pyroxene. The similarity in lithology and hydrothermal alteration between this intrusion and intermediate lavas of the Luis Lopez Formation seems sufficient to assume that they are correlative. However, if the correlation of the surrounding rhyolite (Tirl) is correct, then the andesitic intrusion would have to be correlated with the youngest intermediate lava (Tla_4), which is only observed in the Socorro Peak area.

Tuff of South Canyon

The tuff of South Canyon (Tsc, pl. 1) is the youngest major ash-flow sheet in the Socorro region; where present it marks the top of the Datil volcanic pile (Table 1; fig. 8, last column). Named by Osburn (1978, p. 49) for a type section at the mouth of South Canyon, this ash-flow sheet forms a simple cooling unit of high-silica rhyolite, which is distinctively zoned, both mineralogically and texturally. Although it is normally mapped as a single unit, the tuff of South Canyon is generally divisable into a lower crystal-poor zone that is partially to densely welded, and an upper moderately crystal-rich zone that is densely welded.

A single sample of moderately crystal-rich tuff of South Canyon, from the Joyita Hills, has been analysed as a high-silica rhyolite, containing about 77.8 percent SiO_2 (D.L. White, unpub., see p. 25). A biotite separate from this same very fresh sample yielded a K-Ar age of 26.2±1.0

m.y. (Appendix B).

Correlatives of the tuff of South Canyon mapped in previous investigations of the Magdalena Project have been referred to as the "upper tuffs" (Simon, 1973; Blakestad, 1976) and as the "upper Potato Canyon Rhyolite" (Spradlin, 1976). These stratigraphic terms have been abandoned by present members of the project. The tuff of South Canyon, as defined by Osburn (1978) is not correlative with the "Potato Canyon Rhyolite" of Deal (1973). Since as mapped by Deal, the Potato Canyon Rhyolite apparently also includes the Lemitar Tuff and welded tuffs between the Lemitar Tuff and A-L Peak Tuff, which are generally not recognized outside of the San Mateo Mountains. However, Deal's informal and unmapped subdivision of the Potato Canyon Rhyolite referred to as the "upper moonstone tuff unit" could be equivalent to the tuff of South Canyon. As described by Wilkinson (1976), the "tuff of Gray Hill" and "Potato Canyon Tuff" are very similar to the lower and upper zones of the tuff of South Canyon, respectively.

Past and ongoing mapping of the Magdalena Project has shown that the tuff of South Canyon crops out discontinuously in the following areas: Joyita Hills, Lemitar Mountains, Socorro Mountains, northern and southern Chupadera Mountains, throughout the Magdalena Mountains, and near

Landavaso Reservoir (G.R. Osburn, 1979, oral commun.). Potential equivalents of the tuff of South Canyon are also present in the northern San Mateo Mountains and in the Gallinas Mountains near Gray Hill and at Lion Mountain (op. cit.).

Chapin and others (1978) have proposed a down-faulted area along the western flank of the Magdalena Mountains, designated the Hop Canyon cauldron (fig. 9), as the source of the South Canyon ash-flow sheet. This area is largely covered by Miocene silicic lavas and basin-fill alluvium (fig. 2); and a thick intracaldera equivalent of the South Canyon has not yet been found, although much of the area remains unmapped. Additional data on the distribution and thickness of the tuff of South Canyon will be needed to verify the proposed source.

In the area of this study, the tuff of South Canyon crops out as widely distributed remnants that are generally discontinuous along strike, either because of erosional truncation or deposition in paleochannels, or both. General outcrop localities occur near Bug Mountain and Strawberry Canyon in the southern Lemitar Mountains, northeast of Socorro Peak, and between the Tower mine and Bear Canyon Box in the northern Chupadera Mountains. At most localities the maximum preserved thiskness is about 60 m. The greatest preserved thickness, approximately 85 m, occurs in the paleovalley northeast of the Tower mine, where the tuff rests on a sandstone in the Luis Lopez Formation.

The petrography of the tuff of South Canyon, in the area of this report, have not been studied in detail. For detailed petrographic data and descriptions the reader should refer to Osburn (1978, p. 50-58). Some general petrologic indices of the tuff of South Canyon from Osburn's data are summarized in Table 2. Field observations and handspecimens collected in this study area are mostly in agreement with the observations of Osburn. Some differences in zones of welding, which are apparent between the type section and the Socorro Peak area, may be explained by eastward thinning of the cooling unit. Where pertinent, some thin section observations of samples collected in the central Lemitar Mountains (Lemitar Map, LM-8) are mentioned in the following descriptions.

Phenocryst mineralogy throughout the tuff of South Canyon is relatively constant and may be described independently of zonal variations. Subequal volumes of quartz and sanidine are the dominant phenocrysts in all zones of the tuff. Biotite and sometimes plagioclase are locally present in amounts of less than one percent. The high quartz-sanidine ratio, approximately 1.0, is unique to the tuff of

South Canyon. Where the uppermost densely welded zone has been preserved from erosion, sanidine may be more abundant than quartz (Osburn, 1978, fig. 14).

Quartz crystals are generally euhedral and range from 1 to 3 mm in maximum dimension. In handspecimens from this study area, four sided rhombic-looking crystals are common along with a few hexagonal dipyramids. Sanidine phenocrysts are mostly subhedral and glassy but locally may exhibit a bluish "moonstone" chatoyancy, particularly in the upper moderately crystal-rich zone. Northeast of Socorro Peak, altered outcrops of the tuff of South Canyon contain skeletal whitish sanidine, which weather to honeycomb-like crystals as previously described in the A-L Peak Tuff. Very fine flakes of coppery biotite occur sporadically in the lower crystal-poor zone and are fairly common in the upper zone. Sparse phenocrysts of plagioclase altered to white clay, highly visible in rocks containing fresh sanidine, have been occasionally observed near the top of the thickest preserved sections.

For the purpose of description, and following the approach of Osburn (1978), the tuff of South Canyon in the Socorro Peak area may be divided compositionally into a lower crystal-poor zone, generally with 2-5 percent phenocrysts (locally slightly more), and an upper moderately crystal-rich zone, with 10 to 15 percent phenocrysts. Uppermost zones with 20 percent phenocrysts (Osburn, 1978, fig. 14) are not commonly preserved in the Socorro Peak area. Texturally, the tuff of South Canyon may be divided into a basal nonwelded zone, a lower partially welded zone, and an upper densely welded zone. An upper partially welded zone has not been observed, apparently because of significant erosion of the South Canyon ash-flow sheet in late Oligocene to early Miocene time prior to burial by Santa Fe Group strata.

Essentially all outcrops of the tuff of South Canyon in this study area appear to consist of only two zones. This is because the lower nonwelded zone is rarely exposed, and also because the gradational contact from partial to dense welding is roughly coincident stratigraphically (within 3 to 5 m), with the fairly abrupt increase in phenocryst content.

In the South Canyon type section (Osburn, 1978, fig. 14), the base of the densely welded zone is about 25 m below the base of the moderately crystal-rich zone, and between them is a pale-red lithophysal zone approximately 20 m thick. This lithophysal zone, which Osburn describes as occurring "sporadically", has not been observed in this study area. The above described differences in zonation between the type section and this study area may be explained if the original South Canyon cooling unit was thinner in the Socorro Peak-Lemitar Mountains area. The type section may be unusually complete with a light-colored upper partially welded(?) zone preserved at the top (Osburn, 1979, oral commun.). In comparison to the type section, which includes 62 m of lower crystal-poor tuff and 127 m of moderately crystal-rich tuff, the thickest preserved section in this study area (in the paleochannel northeast of the Tower mine) consists of about 40 m of crystal-poor tuff and 45 m of moderately crystal-rich tuff.

The basal non-welded zone is fairly well exposed in the wall of Strawberry Canyon, in the northeast corner of the study area. Here, it is 1-2 m thick and consists of a friable, clayey, grayish-pink tuff, with traces of phenocrysts of quartz and sanidine. Also present are sparse, small, reddish-brown, basaltic andesite(?) lithics. A few meters of bedded, lithic-rich, crystal-poor, tuffs have been observed at the same position in the eastern Magdalena Mountains (Osburn, 1978; Petty, 1979).

The lower partially welded zone, generally 20 to 40 m thick, typically forms slopes locally broken by ledges of angular joint blocks. It may grade into the cliff- and ledge-forming densely welded zone, although a break in slope is common between these zones. In the central Lemitar Mountains, ledge-forming intervals near the poorly welded base appear to be unusually well indurated because of cementation by vapor-phase quartz.

The lower partially welded zone consists of light-gray to light-brownish-gray, pumiceous, crystal-poor rhyolite generally with 2 to 5 percent phenocrysts of quartz and sanidine, and 5 to 10 percent crystal-poor, medium-gray, botryoidal pumice. Botryoidal pumice appears to be distinctive of zones of incipient to poor welding in pumiceous tuffs. Botryoidal pumice is better developed and more prominent in the tuff of South Canyon because this zone is thicker than equivalent basal zones in the Lemitar Tuff and A-L Peak Tuff.

Lithic fragments at most outcrops consist of small pieces of reddish-brown to dark-gray basaltic andesite, and less commonly grayish-red, crystal-poor, rhyolite that appears to be either densely welded or flow-banded. The latter type of lithic fragment is most likely derived from the A-L Peak Tuff. Lithic fragments are mostly small (< 1 cm) and moderate in abundance (1-3 percent). Lithic-rich lenses, similar to those described by Osburn (1978, fig. 18), have been observed in the Lemitar Mountains. Lithic fragments of quartz-rich Lemitar Tuff have been observed in the lower tuff of South Canyon only near the Tower mine suggesting that they were picked up locally from the walls of the paleochannel, here filled by the South Canyon ashflows.

In the northern Chupadera Mountains, the upper densely welded zone, generally 20 to 45 m thick, forms blocky, jointed ledges, small cuestas and hill tops. Exposures in the Tower mine area generally exhibit a break in slope at the base of the densely welded zone which occurs about 3 m stratigraphically below the gradational contact between the lower crystal-poor zone and the upper moderately crystal-rich zone. The latter contact is shown on the geologic map (pl. 1) as an intraformational contact. In the northern part of the study area, where the tuff of South Canyon is overlain by cliff and hogbackforming lower Popotosa Formation (Tpsd, Tplrl, pl. 1), it has no apparent topographic expression; however, it does form resistant exposures.

Rock samples from the upper densely welded zone are mostly medium light gray to light brownish gray and streaked with 5 to 10 percent of light-gray, crystal-poor to moderately crystal-rich pumice. Pumice as much as 4 cm long are common and occasionally the pumice are rimmed with pale-red spherulites. In the northern Cupadera Mountains, thin zones of pale-red densely welded tuff have been observed locally that are streaked with medium-gray, cherty-looking irregular blebs replacing the pumice. These reddish zones may be incipient lithophysal zones. The densely welded zone is consistently moderately crystal-rich; quartz and sanidine are dominant phenocrysts and minor coppery biotite occurs sporadically. Lithic fragments tend to be small and sparse as in the lower tuff of South Canyon.

Stratigraphic relationships of the tuff of South Canyon in the Socorro Peak area are variable and reflect structures that both predate and postdate the ash-flow sheet. In the southern Lemitar Mountains, the tuff of South Canyon rests conformably on the relatively flat top of the upper tongue of La Jara Peak Basaltic Andesite; this is the typical regional stratigraphic relationship of the South Canyon outflow sheet. At Bug Mountain, the tuff of South Canyon is truncated by a south-facing erosion surface at the base of the Popotosa Formation. This surface is locally related to a long-term subsidence toward the south along a peripheral hingeline structure of the Socorro cauldron. The longevity of this cauldron-initiated structure may be caused by regional extensional stress related to the early Rio Grande rift.

Northeast of Socorro Peak, the South Canyon ash-flow sheet appears to have lapped onto a rhyolite dome of the Luis Lopez Formation on a north-facing slope at the cauldron rim. The westernmost of these two outcrops is anomalously brecciated. It has been broken throughout into fist-sized and larger angular fragments that do not appear to be rotated and only occasionally appear to be cemented by red silica. The overlying Popotosa strata are not brecciated and the underlying rhyolite lava shows relatively minor and irregular brecciation. The origin of this brecciation is probably not related to the underlying rhyolite dome and is most likely tectonically induced similar to larger-scale disrupted blocks of the tuff of South Canyon north of the Tower mine.

Stratigraphic relationships of the tuff of South Canyon in the northern Chupadera Mountains help define a north-trending fault zone, downthrown to the east and then overlapped by the South Canyon ash flows. On the west side of this north-trending fault zone, on the Tower mine block (see p. 100), the tuff of South Canyon fills a paleochannel(s) cut in the upper Lemitar Tuff, which had previously been partially filled by upper La Jara Peak Basaltic Andesite. East of the fault zone, and now at approximately the same elevation, the tuff of South Canyon fills a paleochannel(s) partly filled by channel sands of the Luis Lopez Formation which, in turn, overlie rhyolite breccias and lavas and andesitic lavas of the Luis Lopez Formation. As much as

500 m of down-to-the-east subsidence may have occurred along this fault zone prior to eruption of the tuff of South Canyon (pl. 2, G-G', H-H').

Other anomalous aspects of the tuff of South Canyon in the northern Chupadera Mountains indicate local tectonism shortly after its eruption. At the South Canyon exposure, southeast of Bear Canyon Box, the upper moderately crystal-rich South Canyon is in sharp contact, along a nearly horizontal and silicified fault, with an overlying block of quartz-poor upper Lemitar Tuff. The compaction foliation in the upper block is nearly vertical. This block is interpreted as an ancient landslide block of probable late Oligocene or early Miocene age. The antiquity of the slide block is required by the lack of any apparent source terrane and the silicification of the contact. The most likely source of this block would have been a northward continuation of the Tower mine block, uplifted to the west of this outcrop along the same north-trending fault zone previously described. Later down-to-the-west movement along the zone (pl. 2, H-H'), related to rifting, has allowed the probable source to be buried under younger Santa Fe Group strata.

Northwest of the Tower mine, strike and dip measurements of compaction foliations in the tuff of South Canyon are erratic (pl. 1, marked with ?) and appear to define large, jumbled landslide blocks. Many of these blocks are cut by veinlets of red jasperoid, which locally seems to act as matrix cementing smaller brecciated fragments of tuff of South Canyon. East of the Tower mine, a relatively unbroken block of South Canyon is cut by numerous irregular red jasperoid veinlets, which perceptibly increase in abundance toward the unconformable contact with the lower Popotosa mudflow deposits (Tpsd). These mudflow deposits are extremely well-indurated by red silica cement; this same red silica cement is interpreted to have infiltrated downwards into clastic dikes and fractures in the underlying, brittle upper tuff of South Canyon. Popotosa strata in the Tower mine area are faulted, but not jumbled like the tuff of South Canyon. Thus, the jumbled South Canyon exposures apparently define local faulting and uplift, probably along the same north-trending fault zone previously described, prior to deposition of the lower Popotosa mudflows in late Oligocene or early Miocene time.

This phenomenon of red jasperoid cemented breccias, veinlets and clastic dikes in the exhumed erosion surface originally formed on top of the tuff of South Canyon has been widely observed. It also occurs in the Lemitar Mountains (Bruning, 1973, fig. 16; Lemitar Map) and in the eastern Magdalena Mountains (Osburn, 1978, p. 50). In many instances, it indicates that the South Canyon outcrops were buried by the Popotosa mudflows, even where the mudflow deposits have been completely removed by later erosion. For example, northwest of the Tower mine, the South Canyon is overlain by a quartz latitic lava of the late Miocene Socorro Peak Rhyolite but contains abundant fracture fillings of red jasperoid. In conjunction with stratigraphic relationships on adjacent fault blocks (pl. 1), the jasperoid occurrence helps define a period of local faulting, differential uplift, and erosion between early Miocene and late Miocene time.

SANTA FE GROUP

The term "Santa Fe Group" as used by many geologists (Baldwin, 1963; Hawley and others, 1969; and Machette, 1978) is generally considered to include essentially all (?) sedimentary and minor interbedded volcanic rocks related to the Rio Grande rift, and to range in age from Miocene to middle Pleistocene. The sedimentary rocks of the Santa Fe are often described as thick intermontane basin fill, deposited in graben-like basins and derived from horst-like ranges formed by regional extension.

In the Socorro area, numerous angular unconformities within the Santa Fe Group and within underlying late Oligocene volcanic rocks (Chamberlin, 1976, 1978) have recorded a structural style of rifting similar to that seen in the Afar depression of east Africa (Morton and Black, 1975). This alternative to the horst and graben style of rifting is characterized by progressive rotation of closely spaced fault blocks through time. It is recognized by progressively decreasing dips of contemporaneous strata with decreasing age, and by many small, wedge-shaped prisms of strata bound by rotated normal faults and local angular unconformities. Relationships of this type may be commonly observed on the geologic map (pl. 1) and cross sections (pl. 2) of the Socorro Peak area. Some horst-and-graben-style structure is also evident in the Socorro area; both of these structural styles are compatible forms of crustal extension and may overlap in time and space, since they are probably controlled by variations in the local thermal regime (Chamberlin, 1978).

As used in this report, the Santa Fe Group follows the stratigraphic limits defined by Baldwin (1963) and described by Hawley and others (1969). However, the lower time limit of "generally Miocene" (op. cit., p. 52) arbitrarily excludes several hundred meters of late Oligocene basaltic-andesite lavas and high-silica rhyolite ash flow tuffs that were erupted contemporaneously with the onset of rifting in the Socorro area (Chapin, 1978). Thus, as interpreted here, the Santa Fe Group includes <u>most</u>, rather than "all", rocks related to the Rio Grande rift.

In central New Mexico, the Santa Fe Group has been redefined (Machette, 1978) to consist of the Miocene Popotosa Formation of Denny (1940) and the early Pliocene to middle Pleistocene Sierra Ladrones Formation, newly named by Machette. The Popotosa Formation has been interpreted as a fanglomerate and playa deposit that filled a closed basin (bolson) related to the early development of the Rio

Grande rift (Denny, 1940; Bruning, 1973; Chapin and Seager, 1976; Machette, 1978). The Sierra Ladrones has been described as a thick accumulation of fluvial sands of the ancestral Rio Grande (Machette, 1978; Bachman and Mehnert, 1978), which is largely overlain by, and intertongues with, piedmont gravels derived from modern ranges bordering the present river valley.

In the Socorro Peak volcanic center, the Santa Fe Group has been divided into 4 major mappable units consisting of: 1) the lower Popotosa Formation, 2) the upper Popotosa Formation, 3) the Socorro Peak Rhyolite (locally interbedded in the upper Popotosa Formation), and 4) the Sierra Ladrones Formation. At least two stratigraphic horizons of basaltic flows are interbedded in the upper Popotosa Formation and a third basalt occurs in the Sierra Ladrones Formation. Each of the three sedimentary "formations" is believed to reflect a significant change in the geometry of the local basin(s) of deposition.

The Popotosa and Sierra Ladrones Formations have been subdivided into numerous intertonguing lithofacies, some of which are locally bound by intraformational unconformities. General lithofacies groups recognized are: 1) conglomerates and conglomeratic sandstones, mostly interpreted to be of piedmont-slope and alluvial-fan origin; 2) sands and sandstones of fluvial or distal-alluvial-fan origin; and 3) claystones to siltstones of basin-floor (playa) or overbank origin, and also some minor lacustrine deposits. The coarse-grained strata are also divisable into locally unique facies that are characterized by: 1) paleocurrent directions, 2) clast lithologies, 3) color, 4) sedimentary structures (mainly bed-forms), and 5) degree of induration. Fine-grained facies are primarily differentiated by their associated equivalent coarse-grained facies, and also by color variations and associated lithologies, such as gypsum.

The base of the Santa Fe Group in the volcanic center is placed at the contact above which volcanic-rich sedimentary rocks, mostly conglomerates, are dominant and below which various types of volcanic rocks are dominant. In most instances, this contact is a local angular unconformity; below it the underlying volcanic strata -- previously broken into numerous tilted fault blocks -- dip about 10-30 degrees more steeply than the overlying conglomerates (Tpsd, Tplrl, pl. 1). Similar angular relationships occur at the base of the Santa Fe Group in the Lemitar Mountains (Lemitar Map), in the type area of the Popotosa Formation (Denny, 1940; fig. 2), and in the Magdalena Mountains (Osburn, 1978; Petty, 1979).

The top of the Santa Fe Group is defined as the "surface of the youngest basin fill pre-dating initiation of the Rio Grande valley entrenchment" (Hawley and others, 1969), which is generally accepted as occurring in middle Pleistocene time, about 0.5 m.y. ago (Kottlowski, 1958). Accurate placement of this contact typically requires detailed geomorphic and pedologic studies (Hawley and Kottlowski, 1969; Machette, 1978), which have not been attempted in this investigation. For this study, the top of the Santa Fe Group (top of Sierra Ladrones Formation) is placed at the base of thin piedmont-slope gravels which underlie broad, partly dissected, geomorphic surfaces of apparent Quaternary age. The oldest widespread surface, which generally projects about 37 to 43 m above adjacent arroyo floors, may be of youngest Santa Fe age (J. W. Hawley, 1979, oral commun.).

Strata of the Santa Fe Group are best exposed, and appear to be thickest, along the northern boundary of the study area (pl. 1). Structure sections in this area (Lemitar Map) indicate a total composite thickness for the Santa Fe of about 1.7 km. This estimate does not include the highly lenticular Socorro Peak Rhyolite. Gravity and seismic-refraction data for the basin east of Socorro Peak suggest that "low-density" basin-fill units may be as much as 1.3 km thick (A. R. Sanford, 1979, oral commun.). The well-indurated lower Popotosa Formation, as much as 0.5 km thick, presumably would not qualify as low-density strata. If the upper Popotosa playa muds and moderately indurated fanglomerates are of low density, then the 1.7-km exposed thickness of the Santa Fe basin fill would be in good agreement with the geophysical estimates.

Popotosa Formation

The Popotosa Formation, named by Denny (1940) for exposures northwest of San Acacia at Popotosa Arroyo (fig. 2), is considered the basal unit of the Santa Fe Group in the Socorro area (Bruning, 1973; Chapin and Seager, 1975; Machette, 1978). In the type area, the Popotosa consists dominantly of volcanic-rich conglomerates, conglomeratic sandstones, and mudstones to siltstones containing some gypsum beds (Denny, 1940; Bruning, 1973, Machette, 1978). These lithologies have been generally interpreted as intertonguing fanglomerate and playa deposits that filled a broad, closed basin formed by crustal extension during early development of the Rio Grande rift (Bruning, 1973; Chapin and Seager, 1975).

Deposition of the Popotosa apparently spanned nearly all of Miocene time. Near the type area, the Popotosa rests on basaltic-andesite lavas, dated as 26.3 m.y. old, and tentatively correlated with the La Jara Peak Basaltic Andesite (Machette, 1978). At San Acacia, a 4.5-m.y.-old basalt flow (Bachman and Mehnert, 1978) is interbedded in the Sierra Ladrones Formation, considered to unconformably overlie the Popotosa Formation (Machette, 1978).

In the Socorro volcanic center, basal Popotosa strata rest unconformably on the 26.2-m.y.-old tuff of South Canyon or on older late Oligocene volcanic units. North of Sedillo Hill, playa muds (Tpkp) of the upper Popotosa -- generally masked by landslide blocks of basalt (Qlb) -- are overlain by the 4.0-m.y.-old (Bachman and Mehnert, 1978) basalt of Sedillo Hill, which occurs at, or near, the base of the Sierra Ladrones Formation. Thus, the radiometric ages demonstrate that the Popotosa sections at the type area and at Socorro Peak are essentially equivalent in age.

The Popotosa Formation is here informally divided into a lower member of early (?) Miocene age and an upper member of middle (?) to late Miocene age. The author believes that future detailed mapping will allow the Popotosa Formation to be split into at least two regional genetic units that reflect major changes in basin geometry related to local evolution of rift structure.

Lower Member (Tps- , Tpl-)

The geographic subdivision of the lower Popotosa Formation as shown on the geologic map, into a "member of the Socorro cauldron" and a "member of the Lemitar Mountains" is no longer considered warranted, since the major map units (Tpsd, Tplrl) of both areas are clearly equivalent in all respects. The lower member of the Popotosa Formation in the Socorro area consists of five separate mappable units (Tpsd = Tplrl, Tpsw, Tpsl, Tplru, Tplb) and one undifferentiated unit (Tplr). These five units display both intertonguing facies relationships and also local unconformable relations. They will be labeled and referred to here by their general lithologic and stratigraphic relationships, as opposed to their interpretative descriptions on the geologic map, some of which have been slightly revised. Figure 24, a paleogeologic map of the study area in early Miocene time, summarizes the relationships of the basal facies (Tpsd, Tplrl, Tpsw) of the lower Popotosa Formation, which are described in the following sections. The basal contact of the Popotosa Formation is a widespread erosion surface of early (?) Miocene age that displays a wide range in stratigraphic relationships and is therefore described separately.

EARLY MIOCENE (circa 25 m.y.)

early lower Popotosa time

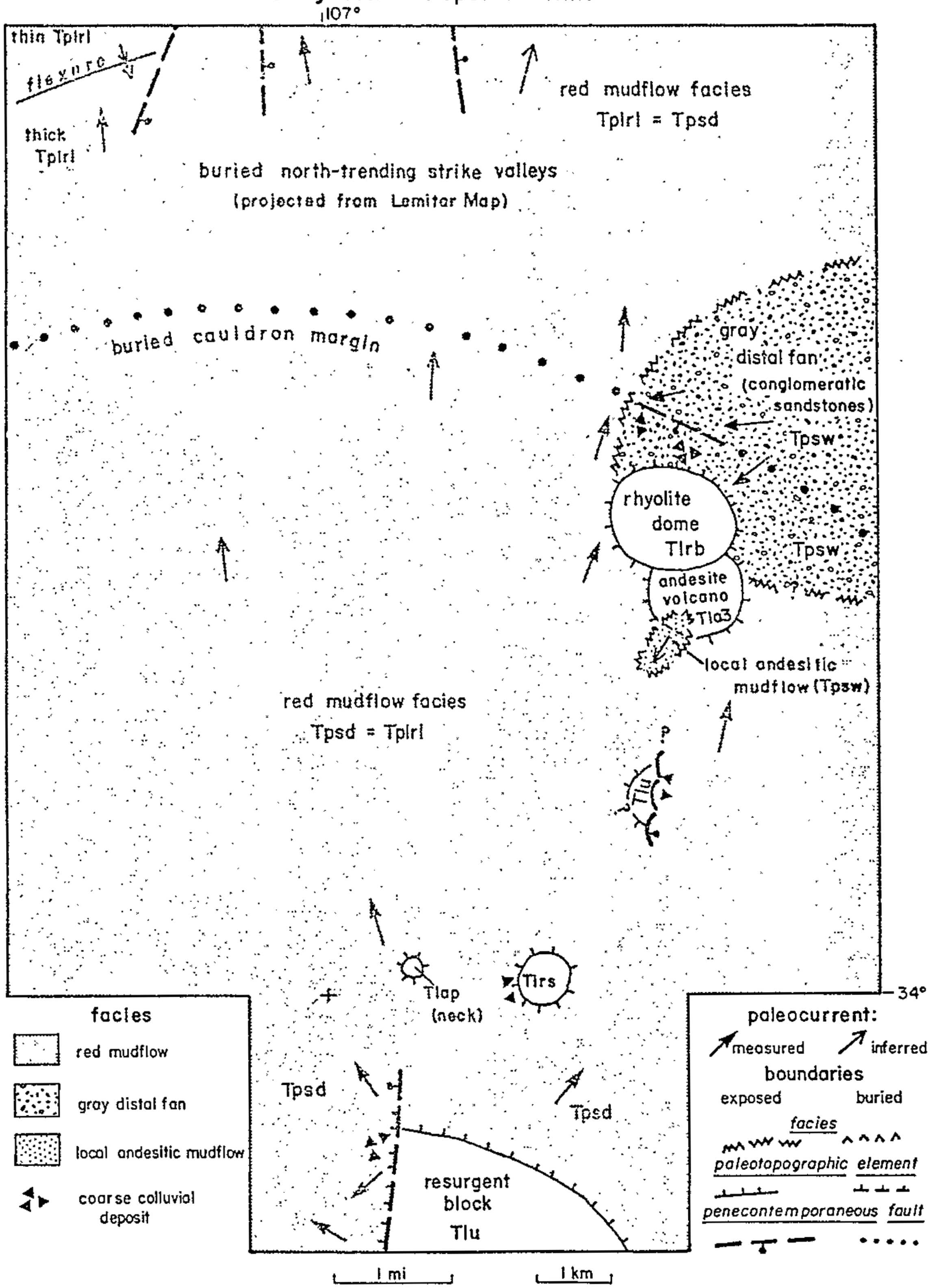


Figure 24. Paleogeologic map of the Socorro Peak volcanic center in early Miocene time summarizing the depositional relationships of the basal lower Popotosa Formation. Younger facies of the lower Popotosa Formation (Tpsl, Tplru, Tplb) are not shown. See Plate 1 for explanation of map symbols.

Basal contact relations. The basal contact of the Popotosa Formation follows a complex topographic surface of moderate to locally steep relief. The early (?) Miocene conglomerates, which bury this surface (Tpsd, Tplrl, Tpsw, pl. 1) were derived mostly from local topographic elements related to: 1) structures of the Socorro cauldron, 2) local fault-block topography of the early Rio Grande rift (north-trending paleovalleys), and 3) local constructional volcanic topography. The dominant northerly paleocurrent directions of most of these conglomerates is thought to reflect a broad north-facing paleoslope of late Oligocene age away from the resurgent highlands of the Socorro caldera and other calderas to the southwest (fig. 9).

The widespread red mudflow and conglomerate facies (Tpsd and Tplrl) buries structures related to the Socorro cauldron margin at Bug Mountain and at Pathway Canyon on Socorro Peak. In the northern Chupadera Mountains, the red mudflow deposits lap unconformably across the moat-fill deposits of the Luis Lopez Formation and onto cauldron-facies tuffs that mark the resurgent block of the cauldron.

In the northern two-thirds of the study area, the base of the Popotosa is an angular unconformity, below which the Oligocene volcanic strata are generally tilted westward about 10 to 30 degrees more than the overlying Popotosa.

The same angular relationship exists, but in the opposite direction (eastward), along the southern margin of the study area. These relations reflect local tilting of north-trending block faults prior to deposition of the lower Popotosa strata (Chamberlin, 1976, 1978). This tilting in opposite directions is the signature of the transverse shear zone (Chapin and others, 1978). Basal Popotosa strata onlapping originally west-facing slopes of 15 to 30 degrees are well exposed at the base of the cliff, northeast of "M" Mountain, and also south of the East Roadcut.

Between Chupadera Cliff and the Big Roadcut at Highway Sixty, the basal Popotosa Formation is exposed on a slightly north-tilted, exhumed erosion surface of early Miocene age. The lobate remnants of red mudflow deposits and conglomerates (Tpsd) here fill a shallow north-trending erosional valley cut into the moat deposits. This local block, which is not tilted east or west, and younger blocks like it (Black Mesa, "6001" Mesa) are located along the "null line" of the transverse shear zone (Chapin and others, 1978).

Locally, the basal Popotosa consists of monolithic or bilithic boulder breccias, which are interpreted here as ancient colluvial deposits adjacent to eroded scarps of various types. They are indicated on the geologic map with a breccia symbol (pl. 1) or mentioned here where outcrops are too small to show as such on the map. East of the Tower mine is a boulder breccia of welded tuff blocks derived from cauldron-facies Lemitar (?) Tuff and the tuff of South Canyon. This coarse breccia is probably related to a north-trending early rift fault (fig. 24), which cuts the resurgent block of the cauldron and is downthrown to the west. Similar fault-scarp breccias are found west of the Grefco mine (along the east side of the isolated block of cauldron-facies Lemitar Tuff), west of the Merritt mine, and south of Big Cliff (mapped with "Tlrb" of Luis Lopez Formation). The latter two occurrences are apparently related to cauldron-ring fracture faults, locally reactivated by rift stresses. This phenomenon of extensional stress reactivating cauldron structures and perpetuating their original sense of movement has been described in the Long Valley caldera (Bailey, 1976) and the Timber Mountain caldera (Christiansen and others, 1965). The breccia west of the Grefco mine is associated with a buried low-angle normal fault, which partially bounds the east side of the welded tuff outcrop (pl. 2, F-F'). The lithology of breccia blocks at all these localities indicates their derivation from the immediately underlying or adjacent Oligocene volcanic units.

The basal Popotosa also buries local primary volcanic topography such as the Blue Canyon rhyolite dome (Tlrb).

Along the west side of this dome, three facies of the lower Popotosa (Tpsd, Tpsw, Tpsl) wedge out onto more than 100 m of paleotopographic relief. The original slope on the west flank of the dome was about 25 degrees. The total buried relief on this dome may be much greater (pl. 2, C-C'). Colluvial breccias are also found as eroded remnants on the rhyolite lavas of Highway Sixty (Tlrs) south of the east Roadcut.

Lower red mudflow and conglomerate facies (Tpsd = Tplrl). The most distinctive, widespread, and thickest facies of the lower Popotosa Formation consists of extremely well-indurated, dark reddish-brown, heterolithic, mudflow deposits and conglomerates (Tpsd = Tplrl, pl. 1). The lower red mudflow deposits and conglomerates form the bold cliffs (Big Cliff) high on the east face of Socorro Peak. Less prominent outcrops generally weathered to dark-brown or reddish-brown low cliffs, ledges and ridges, occur throughout the study area, such as at Bug Mountain, north of Strawberry Peak, Blue Canyon, Socorro Spring, and west of Black Canyon. Talus blocks and colluvium derived from the resistant mudflow deposits are common on surrounding slopes. Soils on these outcrops contain colluvial blocks from which the matrix does not weather away from the clasts. The matrix is as hard, or harder, than the clasts and the blocks typically break across the clasts.

The red mudflow and conglomerate outcrops are generally medium- to thick-bedded and display crude subparallel stratification. Cut-and-fill structures associated with sandy bedload gravels or lag gravels are fairly common. Most of the beds are massive mudflows with the smaller volcanic clasts "floating" in a red sandy and muddy matrix (fig. 25). Larger cobbles and boulder clasts locally appear to touch each other and if elongated may appear to be some what imbricated, but at an unusually steep angle to the bedding of 45-60 degrees (see Bruning, 1973, fig. 17). Well-imbricated lag gravels generally confirm that the imbricated cobbles in the mudflows are reasonably representative of flow directions. Northerly paleocurrent directions are dominant in this facies (pl. 1).

Volcanic clasts in the lower red mudflow and conglomerate facies appear to be entirely derived from erosion of the underlying late Oligocene volcanic strata. Lateral variations in the abundance of these clasts, particularly near the base of this facies, indicates that they were derived largely from nearby topographic elements. Examples of this relationship are found: 1) west of Black Canyon where upper Lemitar Tuff clasts are dominant, 2) in Blue

Figure 25. Lower red mudflow and conglomerate facies (Tpsd, pl. 1) of the lower Popotosa Formation. The coarse clast-rich stratum near the middle of the photograph mostly exhibits characteristics of a mudflow deposit. It has a relatively sharp base (arrow) where it rests on sandy water-laid, channel-fill deposits exhibiting cut-and-fill structures (bottom of photograph). The cobbles in the clast-rich stratum are for the most part clearly floating or standing on end in the sandy-mud matrix. However, some clasts seem to be imbricated (although inconsistently within the bed) and the relatively fine-grained center of the stratum appears to be somewhat sorted, granule-sized material similar to the water-laid channel fill. Thus the clast-rich bed may be transitional between a turbid mudflow deposit and a water-laid deposit. Clasts in this outcrop are mostly derived from rhyolite and intermediate lavas in the Luis Lopez Formation and some are from the upper Lemitar Tuff. Pebble imbrications in adjacent outcrops indicate paleocurrents were generally from right to left. Hammer head is 18 cm long. Outcrop is located in arroyo floor south of Sedillo Spring.

Canyon where clasts of Luis Lopez Formation intermediate lavas and rhyolite lavas are dominant, and 3) north of Strawberry Peak where tuff of South Canyon clasts are dominant. The red mudflow deposits and conglomerates commonly become less coarse and more heterolithic upwards in individual sections. The general northward flow direction is supported by the presence of distinctive clasts of finely flow-banded, white-colored, rhyolite of Highway Sixty (Tlrs) in exposures at the mouth of Strawberry Canyon (fig. 3).

On the east face of Socorro Peak at Big Cliff the lower red mudflow and conglomerate facies (Tpsd) intertongues with and grades rapidly southward into the light-gray conglomeratic sandstone facies (Tpsw), which forms less resistant outcrops and exhibits nearly opposing southwesterly paleo current directions. Interbedding of these two conglomeratic facies (Tpsd and Tpsw) is best displayed in a small ravine on the south central end of the Blue Canyon rhyolite dome. Here, the Popotosa beds are downfaulted against the dome by the Woods Tunnel fault zone (pl. 2, C-C'). Similar interbedded lithologies occur at Big Cliff and west of the Middle mine.

The lower red mudflow and conglomerate facies is older than all the other facies of the lower Popotosa Formation (Tpsl, Tplru, Tplb) and is overlain by these units in the areas west of the Blue Canyon dome and in the southern Lemitar Mountains. In the southern half of the study area (south of Blue Canyon), the younger facies of the lower Popotosa are absent. Here the well-indurated red mudflow deposits and conglomerates (Tpsd) are overlain disconformably and with angular unconformity by poorly indurated muddy conglomerates (Tpkc) and playa mudstones (Tpkp) of the upper Popotosa Formation, or locally by silicic lavas of the Socorro Peak Rhyolite.

The lower red mudflow and conglomerate facies (Tplrl = Tpsd) generally appears to thin from north to south. It is approximately 235 m thick on Bug Mountain, 122 m thick at Big Cliff and, locally absent north of the Tower mine, where Socorro Peak Rhyolite lavas (Tsd) rest directly on the tuff of South Canyon. However, it is also apparent that abrupt east-west variations in thickness of this facies in the Tower mine area are controlled by erosion of local fault-block topography in middle to late Miocene time.

The divergence in dip between the lower Popotosa mudflow deposits and conglomerates and the underlying volcanic strata, most apparent in the vicinity of Socorro Peak, (pl. 2, A-A'), requires that this facies thicken rapidly to the west, as a wedge shaped prism, until it meets a buried fault scarp (or scarps) that produced rotation of the older volcanic strata. The location, number, and position of these buried scarps is problematical. A series of hypothetical buried scarps are shown in the cross section A-A' (pl. 2) on the basis of gravity data. North-trending strike valleys floored by west-facing unconformities and east-facing low-angle normal faults and buried by the lower red mudflow and conglomerate facies are locally exposed in the southern Lemitar Mountains, just north of this study area (Lemitar Map, Chamberlin, 1978). Such buried paleovalleys, not more than 300 m deep, probably could not be detected from gravity data.

In the volcanic center, the lower Popotosa mudflow deposits and conglomerates are loosely bracketed between the immediately underlying 26.2-m.y.-old tuff of South Canyon and the 11.8-m.y.-old quartz latite dome (Tsd) of Strawberry Peak (Appendix B). The latter is separated from the top of the lower red mudflow and conglomerate facies by two local unconformities and about 670 m of younger strata of both the lower and upper Popotosa (Lemitar Map cross sections). These relationships suggest an early (?) Miocene age for the lower red mudflow and conglomerate facies. Exposures of well-indurated, red mudflow deposits and conglomerates, generally with northerly transport directions and similar stratigraphic position have been mapped outside of this study area, mostly in the Magdalena and Lemitar Mountains (Bruning, 1973; Woodward, 1973; Simon, 1973; Osburn, 1978; Petty, 1979; Bowring, in prep.).

Northwest of Magdalena, near Montosa Arroyo (fig. 2), red well-indurated conglomerates are interbedded with rhyodacitic lavas dated at 25.2 m.y. (Simon, 1973); at Water Canyon Mesa similar red conglomerates overlie a quartz latite lava dated at 20.0 m.y. (Osburn, 1978). A tentative correlation is suggested here between these red conglomerates and the lower red mudflow and conglomerate facies (Tpsd = Tplrl) of this report. A tentative age assignment of early Miocene (25-20 m.y.) for the lower Popotosa Formation is in agreement with observations in this study area.

Probably the most distinctive traits of the lower red mudflow and conglomerate facies are its dark reddish-brown color and extreme degree of induration by cryptocrystalline silica cement (Bruning, 1973). Red-colored terrigenous sediments, commonly referred to as "red beds", are generally thought to develop by one of two processes: 1) deposition of muddy sediments, in which the clays are intimately associated with previously formed ferric hydroxide (yellow-brown limonite) and then later diagenetically dehydrated and "aged" to form reddish-brown hematite (Krynine, 1949; Van Houten, 1968); or 2) post-depositional (in situ) diagenetic alteration of sand-sized detrital grains of ferrosilicates, or black Fe-Ti oxides, to limonite by oxidizing intrastratal solutions. The second process is continued by dispersal of the limonite in intrastratal solutions and later dehydration to form hematite (Walker, 1967). Both of these processes seem to be well-supported by detailed studies and there seems to be no good reason to discount either process. They essentially follow the same geochemical path and the only difference is whether the limonite is formed before or after deposition of the sediment.

The anomalous dark-reddish-brown color and high degree of induration of the lower Popotosa mudflow deposits and conglomerates has been suggested by Bruning (1973) to have been formed in situ by ground water alteration. Chapin and others (1978) suggest that the Popotosa rocks may have acquired their unusual red coloration and high degree of induration during potassium metasomatism related to an ancient geothermal system of late Miocene age. Field observations of the red mudflow deposits and conglomerates by the author suggest that the red color and silica cement are an intrinsic (syngenetic) characteristic of this stratigraphic unit. The author suggests that an alternative origin for the red coloration and extreme induration might invoke a scenario where erosion of the late Oligocene rocks occurred contemporaneously with widespread hot-spring activity in early Miocene time. Hot springs should be an ideal environment to produce muddy sediment containing both abundant ferric-hydroxide and silica in solution (White, 1955). Several investigators have noted the accelerated oxidation process of ferrosilicate minerals at higher temperatures in tropical climates (Schmalz, 1968; Van Houten, 1972), which should also apply to hot springs. The main observations which support this hot spring environment of deposition are 1) the homogeneous color of the muddy matrix, 2) the limit in space and time of the dark-red color and silica cement to this one particular genetic stratigraphic unit (see following descriptions of intertonguing and overlying facies), and 3) the occurrence of widespread red jasperoid veinlets and red clastic dikes in the underlying light colored (unstained) volcanic units (see description of tuff of South Canyon), but not in younger strata. Outcrops of the red mudflow and conglomerate facies contain abundant potassium-metasomatized volcanic clasts exhibiting plagioclase phenocrysts replaced by white clay and secondary feldspar, and frameworks of ferromagnesian crystals leached to black iron oxide and silica(?). Preliminary field observations suggest a paucity of detrital plagioclase grains altered "in situ" to white clay, even where the associated clasts contain relatively abundant altered plagioclase. This also supports a pre-depositional alteration of the volcanic clasts as implied in the hot springs model.

Gray conglomeratic sandstone facies (Tpsw). From Blue Canyon northward along the eastern flank of Socorro Peak are numerous exposures of light-gray conglomeratic sandstones (Tpsw) characterized by southwesterly paleocurrent directions and abundant clasts of dark colored intermediate lavas and light-gray A-L Peak-like tuffs. The latter clast type must have been derived from highlands outside the immediate area. The gray-sandstone facies is interpreted to represent distal-alluvial-fan deposits shed from the early eastern margin of the Popotosa basin, probably east of the present Rio Grande. Since the Luis Lopez Formation locally filled the Socorro cauldron to overflowing, it does not seem as likely that the A-L Peak-rich detritus was derived from a topographic wall of the caldera, as suggested on the map (pl. 1).

These distal-fan deposits appear to have originally entered the Socorro Peak area in the "shadow" of the late Oligocene Blue Canyon rhyolite dome (Tlrb) and an andesitic volcano (Tla_3) on the dome's south flank. These volcanic topographic elements locally blocked the northward progradation of the red-mudflow and conglomerate facies (Tpsd).

South of Big Cliff, about 116 m of the conglomeratic sandstones are exposed in a continuous section. Here they rest on colluvial breccias developed on the rhyolite of Blue Canyon (Tlrb) and are overlain by purplish-gray siltstones and mudstones (Tpsl). Toward Big Cliff, the continuous sandstone section becomes interbedded with a northward-thickening wedge of cliff-forming red mudflow deposits and conglomerates (Tpsd).

Outcrops of the gray-conglomeratic-sandstone facies are moderately indurated and generally form ledges and moderately steep slopes. The gray facies is mostly medium to thick bedded and well stratified. Cut and fill structures and graded bedding are common. Colors of the beds are variable and controlled by the dominant clasts present. Well imbricated, tabular, gray tuff clasts are most abundant in the light-gray beds, which grade to purplish gray with increasing andesitic detritus (Spears or Luis Lopez type). Bluish-gray beds, present in several outcrops, are colored by abundant clasts of andesitic lavas that are stained with bluish-green celadonite. Minor abundances of clasts similar to the tuff of South Canyon and upper Lemitar Tuff are also present in outcrops near the mouth of

Pathway Canyon. Black scoriaceous basalt clasts are unusually abundant here.

West of the Blue Canyon dome and below the "M", the gray- sandstone facies appears to be disconformably overlain by playa muds (Tpkp) of the upper Popotosa Formation. At the northeastern corner of Socorro Peak, the gray conglomeratic sandstones (Tpsw) rest in local angular unconformity on the red mudflow and conglomerate facies (Tpsd) and also contain a thin basalt flow. Neither of these relationships has been observed in any other "Tpsw" section.

South of the TERA headquarters in Blue Canyon, two outcrops of poorly indurated, dark-purplish-gray, andesitic mudflow deposits are locally exposed in bulldozer cuts. These are believed to represent local sheet-wash deposits shed from the south flank of an andesitic volcano (Tla_3), the remnants of which are now exposed in the area entering Blue Canyon. They are arbitrarily mapped as the "Tpsw" facies.

<u>Gray mudstone and siltstone facies (Tpsl).</u> Pale purplish-gray to grayish-red, thin-bedded to massive, silty mudstones and siltstones have been mapped as a separate local facies (Tpsl), even though they are typically interbedded with the

upper portions of the gray-conglomeratic-sandstone facies (Tpsw).

These poorly to moderately indurated, slope-forming, fine-grained beds are best exposed in an outcrop area surrounded by Quaternary landslides and colluvium west of the Blue Canyon dome (fig. 24). Approximately 20 m of badland-forming silty mudstones are exposed at the local base of this section. They underlie gently west-dipping ledges of light-gray sandstone (Tpsw), which are in turn overlain by another 10 m of gray mudstone at the east side of the outcrop. Similar, or more intimately interbedded, outcrops of gray mudstone (Tpsl) and sandstone (Tpsw) are common all along the east flank of Socorro Peak (pl. 1).

At the locality west of the Blue Canyon dome, Burton (1971, p. 15) found some thin, yellowish-white, clay beds to contain "fossil imprints of leaves, pieces of reeds, and fragments of bark". At the time of this report, these flora have not been identified and the whereabouts of these specimens is uncertain. Attempts to collect additional fossils at this locality have not been fruitful.

Thin beds of wavy-laminated, muddy limestone and a black petroliferous, calcareous, siltstone have been observed in the downfaulted outcrops along the Woods Tunnel fault zone (pl. 2, C-C') where it cuts the Blue Canyon dome.

Unusually well-developed rhythmic bedding may be observed in the "Tpsl" facies north of Socorro Peak.

The close association of the gray-mudstone facies with the distal-fan sandstone facies (Tpsw) and its lithologic characteristics suggests that the silty mudstones are basin-floor sediments deposited on an alluvial flat or locally in a small lake(s?). The gray mudstone-siltstone facies is not known to intertongue with the lower red mudflow and conglomerate facies (Tpsd). All the outcrops of interbedded gray mudstone (Tpsl) and sandstone (Tpsw) in the Socorro Peak area either overlie the red mudflow and conglomerate facies (Tpsd) or locally rest directly on Oligocene volcanic units (Tlrb, Tla_4) on the northeastern flank of the Blue Canyon dome. It appears that by "late" lower Popotosa time the uppermost gray facies (Tpsw and Tpsl) completely buried the Blue Canyon dome and then prograded westward onto the red-mudflow facies (Tpsd) after deposition of that unit had ceased.

<u>Upper red conglomerate facies (Tplru).</u> Along the northern boundary of the Socorro Peak volcanic center (pl. 1) the cliff-forming red mudflow and conglomerate facies (Tplrl = Tpsd) is locally conformably to disconformably overlain by as much as 100 m (west of Bug Mountain) of moderately indu-

rated, pale-red to moderate-red or moderate-reddish-orange, slope-forming conglomerates (Tplru). The upper red conglomerates (Tplru) are considered to be alluvial-fan deposits derived from local uplift and erosion of the lower red mudflow and conglomerate facies somewhere in the southern part of the study area. Presumably this tectonostratigraphic unit was deposited in early to middle (?) Miocene time shortly after deposition of the lower red mudflows and conglomerates before they became well indurated.

Outcrops of the upper red conglomerate facies generally weather to low rounded hills and slopes that are mantled by a lag gravel of subrounded volcanic cobbles. In contrast to soils developed on the lower red mudflow and conglomerate facies (Tplrl), the clasts on the surface of the upper red conglomerate do not exhibit adhered matrix material. Where well exposed in gulches, this unit is crudely bedded, displays some cut and fill structures, and has a red sandy to pebbly matrix. Northeast of Strawberry Peak, recent erosion of the upper red conglomerates has formed pedestals below large boulders that are more resistant than the matrix of the underlying conglomerate.

Most of the andesitic clasts and numerous rhyolitic tuff clasts in the upper red conglomerate have very dark red rinds that are about 3 "values" darker than the enclosing moderate-red matrix. This observation supports their interpretation as being reworked clasts from the older red mudflows and conglomerates. Near the mouth of Strawberry Canyon, andesitic clasts in the lower red mudflows and conglomerates (Tplrl) have been observed with dark red rinds that are harder and more indurated than the cores of the clasts, which weather out preferentially.

Clast lithologies in the upper red conglomerate are essentially the same as the lower red facies. Pebbles and cobbles similar to the tuff of South Canyon, upper Lemitar Tuff, and various types of andesitic to rhyolitic lavas in the Luis Lopez Formation are common. Rounded clasts of the white, finely flow-banded, rhyolite of Highway Sixty (Tlrs), which have weathered to thin, fissle plates, are fairly common in the soils on this facies (Tplru).

Sparse observations of pebble imbrications in the upper red conglomerate (Lemitar Map) indicate northerly transport directions. At the exposures due north of Strawberry Peak, a moderately well-indurated, pale-red sandstone, about 3 m thick, appears to form the conformable base of the upper red conglomerate. Northeast of Strawberry Peak, the upper conglomerate fills a north-trending (fault controlled ?) channel cut in the top of the moderately south-dipping lower red mudflow and conglomerate facies. About 2.4 km northeast of Strawberry Peak (Lemitar Map), the same lithologic contact is an angular unconformity. West of Bug Mountain, the basal contact of the upper red conglomerate is not well exposed, and is therefore placed at a break in slope and color change similar to those observed elsewhere.

Buff-conglomerate facies (Tplb). As much as 60 m of buff to pale-orangish-pink, poorly exposed conglomerates appear to disconformably overlie the lower and upper red conglomeratic facies near the north boundary of this study area (pl. 1). Where locally well exposed in a ravine just north of Strawberry Peak, this unit displays the crudely even stratification, poor sorting, and coarse grain size -- small boulders to pebbles -- common to most fanglomerate deposits (Bull, 1968). Imbricated cobbles at this exposure indicate a northerly paleocurrent direction similar to the other conglomerate facies of the lower Popotosa Formation.

Most exposures of the buff-conglomerate facies consist of light-colored rounded hills mantled by loose, subrounded cobbles and pebbles of gray and purplish-gray, A-L Peak-like, welded tuffs. Less common volcanic clasts are similar in appearance to La Jara Peak Basaltic Andesite lavas and the tuff of South Canyon. A relatively rare clast lithology observed in this facies consists of grayish-red to purplish-gray flow-banded, hornblende, biotite, plagioclase rhyodacite or quartz latite lavas. These hornblende-bearing clasts look very similar to the quartz latitic lavas of Water Canyon Mesa dated at 20 m.y. (Osburn, 1978) or the quartz latitic lavas of Strawberry Peak -- a domal flow of the Socorro Peak Rhyolite -- dated at 11.8 m.y. Since the buff fanglomerates appear to be overlain and separated from the Strawberry Peak dome by as much as 500 m of the upper Popotosa Formation (Lemitar Map cross sections), these hornblende-bearing clasts in the buff fanglomerate are thought to have come from the Water Canyon Mesa area. A general source area for the buff fanglomerates to the southwest, in the north-central Magdalena Mountains, would help explain the abundance of A-L Peak-like clasts in this unit. The meager paleocurrent control is in general agreement with a source to the south. Potential sources of A-L Peak Tuff detritus to the southeast of this area are not recognized at present.

West of Strawberry Peak, an undifferentiated unit "Tplr" which is probably mostly the buff fanglomerate facies (Tplb), but which may also include the upper red conglomerate (Tplru), has been mapped. These pale-reddish to buff-colored hills appear to be conformably overlain by buff-colored, distal-fan deposits (Tpkf) of the upper

Popotosa Formation. The poorly exposed contact is mappable because of a distinctive change in clast lithology described in the following section.

Upper Member

Facies and units of the upper Popotosa Formation are collectively labeled on the geologic map (pl. 1) as the "member of Kelly Ranch", named for excellent exposures west of the J.B. Kelly Ranch. However, since the regional correlation of this map unit has not been established, it will be referred to here simply as the "upper member."

Figure 26, a paleogeologic map of the Socorro Peak area in late Miocene time, summarizes the spatial and depositional relationships of the upper Popotosa sedimentary facies, as described in the following sections. This figure also illustrates the relationship of the sedimentary units to the contemporaneous lavas and tuffs of the Socorro Peak Rhyolite.

The upper member of the Popotosa Formation consists largely of moderately to poorly indurated, fine-grained detritus derived from highlands to the west and east, well outside the Socorro Peak area. Playa claystones (Tpkp), which underlie about half the study area, are reflected in the wide apron of landslide deposits (Ql) that surround the upper Popotosa Formation and Socorro Peak Rhyolite time

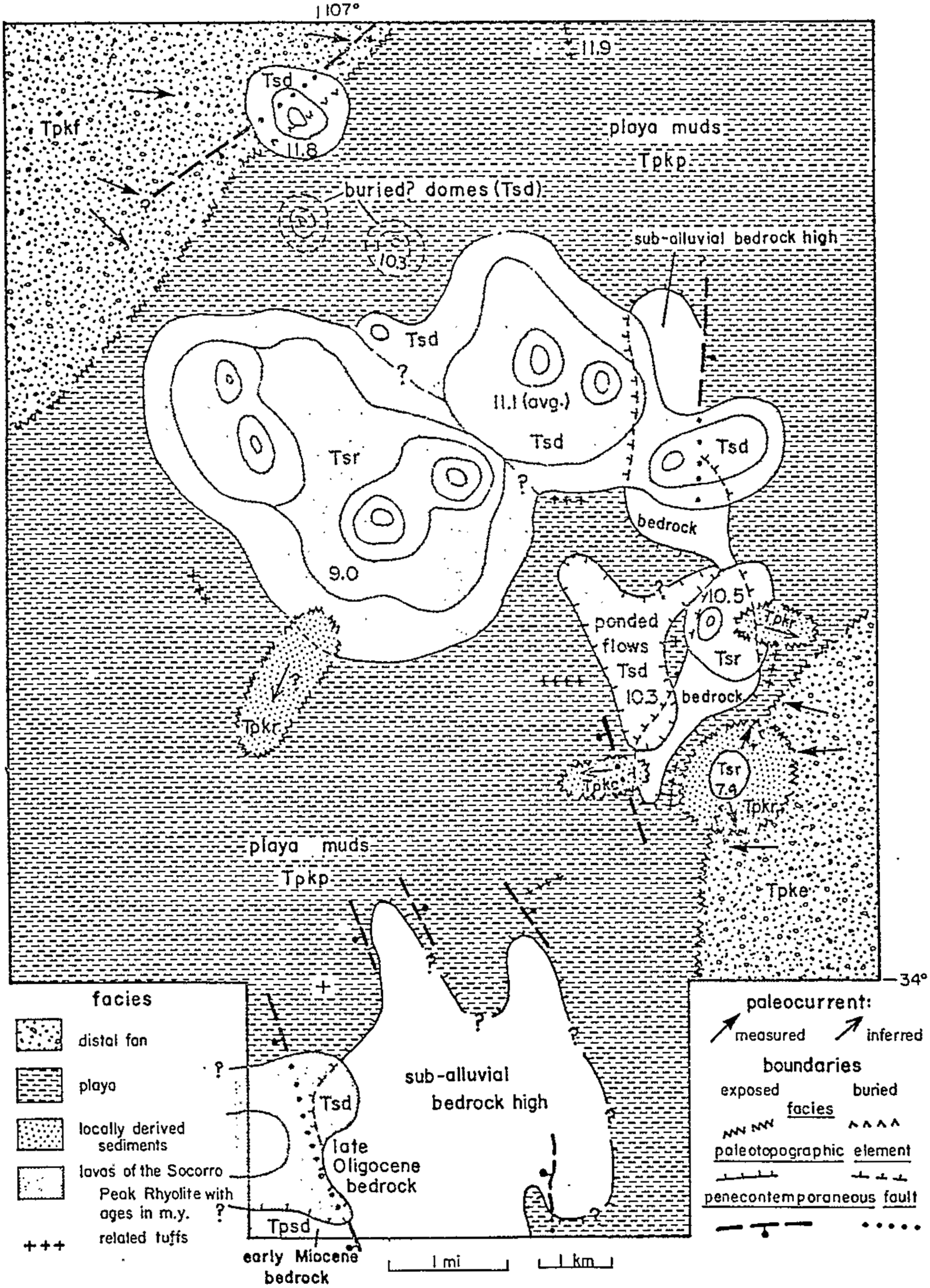


Figure 26. Paleogeologic map of the Socorro Peak volcanic center in late Miocene time summarizing depositional relationships of the upper Popotosa Formation and the contemporaneous (interbedded) Socorro Peak Phyolite. See Plate 1 for explanation of map symbols.

Socorro Mountains. Distal alluvial-fan deposits intertongue with the playa deposits in the northwest (Tpkf) and southeast (Tpke) corners of the study area. Younger strata of the upper Popotosa Formation are interbedded with silicic lavas and tuffs of the Socorro Peak Rhyolite. Rhyolitic detritus (Tpkr) eroded from these flows locally intertongues with the playa and fanglomerate deposits. Bedded tuffs of the Socorro Peak Rhyolite (Tsdt, Tsrt) and two stratigraphic horizons of basaltic flows (Tpkb) are widely intercalated in the upper Popotosa strata. These volcanic marker beds provide important control on thickness variations that would not otherwise be detectable.

The upper Popotosa is as much as 850 m thick to the northeast of Strawberry Peak (Lemitar Map cross sections). The formation generally thins toward the south and southeast portions of the study area where the playa muds appear to have lapped unconformably onto structurally high, but topographically low, fault blocks of older bedrock. These bedrock blocks are formed by variably eroded lower Popotosa and late Oligocene volcanic strata (fig. 26). These suballuvial bedrock blocks did not shed significant amounts of coarse sediment, however they do appear to mark the southern margin of the Popotosa basin.

The basal contact of the upper Popotosa Formation is an unconformity that reflects the presence of this basin margin. Stratigraphic relations described hereafter generally indicate the basal upper Popotosa strata become younger toward the south, and also that they lap onto progressively older (more deeply eroded) bedrock strata. This general north to south thinning, is complicated by longitudinal (north-south-trending) fault blocks that displace the moderately tilted (15 to 30 degrees) upper Popotosa strata. Because of their relatively low resistance to erosion the upper Popotosa strata are generally cut by broad pediment and landslide surfaces of Quaternary age. Similar pedimentation of Popotosa strata during previous geologic epochs is also recognized. Stratigraphic relations at the base of the Socorro Peak lavas (Tsd, pl. 1) in the Tower mine area demonstrate that these north-trending fault blocks were locally beveled by an erosion surface of late Miocene age in upper Popotosa time. Similarly the basalt of Sedillo Hill (Tbsh) buried beveled fault blocks of upper Popotosa strata in middle Pliocene time.

The basal contact of the upper Popotosa Formation generally appears to be a paraconformity or disconformity, or less commonly an angular unconformity. The best exposure of this contact -- typically it is poorly exposed -- is found west of the Blue Canyon dome. A photograph of this locality is shown here as figure 27. As described in the figure caption, the contact appears to be a disconformity.

The upper member of the Popotosa Formation is here divided (pl. 1) into five sedimentary facies and two horizons of interbedded basalt flows. For the purpose of description they are presented here in general ascending stratigraphic order; sedimentary units will be referred to in terms of their dominant lithology and color. The Socorro Peak Rhyolite, which is interbedded with the younger strata of the upper Popotosa Formation, is described separately.

Buff-conglomeratic-sandstone facies (Tpkf). In the northwest portion of the study area, mostly north of a line between Snake Ranch and Strawberry Canyon, the basal strata of the upper Popotosa Formation consist of pale brownish-yellow (buff) colored conglomeratic sandstones (Tpkf, pl. 1). This facies is believed to represent distal alluvial-fan and braided-channel deposits shed from the Magdalena area about 15 km to the west.

The buff-sandstone facies is characterized by easterly to southeasterly paleocurrent directions (pl. 1) and moderately abundant (5-20 percent of total clasts) yellowish-brown hydrothermally altered clasts of A-L Peak-like tuff

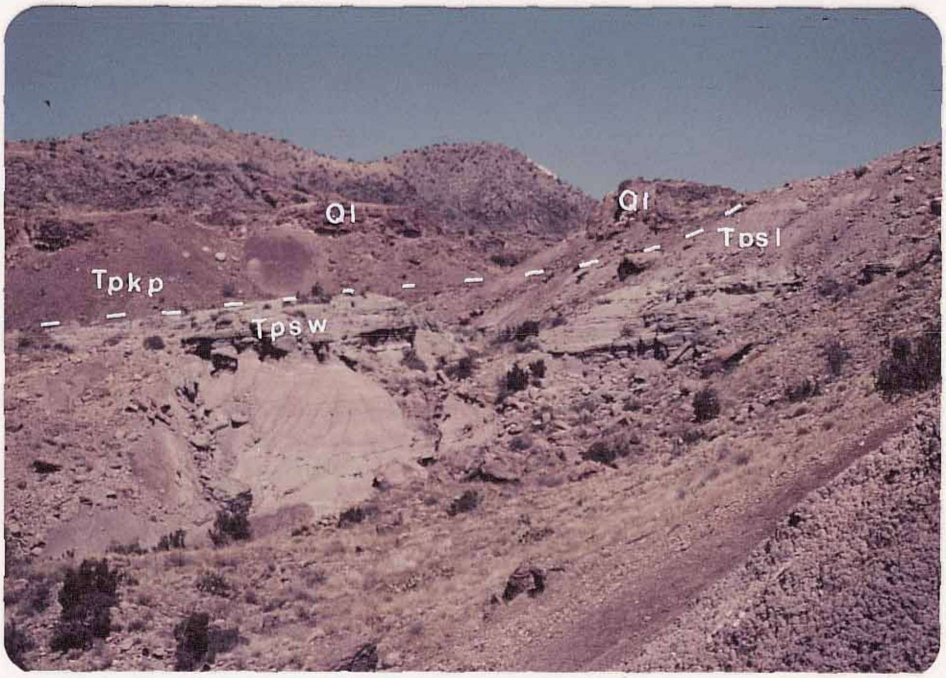


Figure 27. Basal contact (dashed line) of the upper member of the Popotosa Formation. The maroon playa claystones of the upper Popotosa Formation appear to disconformably overlie gray sandstones (Tpsw) and silty mudstones (Tpsl) of the lower Popotosa Formation. Quaternary landslide blocks (Ql) are derived from Socorro Peak Rhyolite (rhyodacite) lavas that form Socorro Peak (radio towers) and "M" Mountain on the skyline at the right. Playa claystones are downfaulted against the lower Popotosa strata in the left foreground and immediate right foreground. Looking north toward Socorro Peak from the area west of the Blue Canyon dome.

Along the south side of Strawberry Canyon, the buff sandstones are conformably overlain by reddish-brown mudstone and pinkish siltstones of the playa facies (Tpkp). This contact is transitional over a stratigraphic interval of 10-15 m. In this interval, poorly indurated buff sandstones, pinkish-orange siltstones and reddish-brown mudstones are intimately interbedded. The buff sandstone (Tpkf) is believed to be replaced toward the east (through intertonguing) by a roughly equal thickness of the red mudstone-claystone facies (Tpkp). A basaltic flow (Tpkb), which is interbedded in the mudstones (Tpkp), on the south side of Strawberry Canyon, provides a time horizon through which it can be shown that the underlying mudstones are thickening eastward at the expense of the underlying sandstones (pl. 1). About 2 km north of the study area (SE/4, 18, T2S, R1W), stratigraphic relations similar to those south of Strawberry Canyon demonstrate that approximately 300 m of the buff conglomeratic sandstones (Tpkf) must grade laterally, within 0.8 km along strike, into an equal thickness of reddish-brown mudstones (Tpkp) (Lemitar Map).

The buff-conglomeratic-sandstone facies is best exposed in the north wall of Strawberry Canyon. The continuous exposure here consists of approximately 180 m of fine- to coarse-grained channel sandstones, which commonly exhibit scour structures, some trough cross bedding and some pebbly channel bottoms. The outcrops are mostly medium-bedded, planar-bedded, and commonly show graded bedding. A structure section from Strawberry Canyon northeastward to where the base of the unit is exposed (Lemitar Map) indicates that the buff conglomeratic sandstone may be as much as 396 m thick here.

Red claystone and mudstone facies (Tpkp). Approximately half of the study area (pl. 1) consists of wide slopes mantled by landslide blocks (Q1) and smaller colluvial debris (Qca). Within these slope areas numerous small outcrops of mostly dark-reddish-brown to maroon claystone (Tpkp) have been exposed where the surficial deposits have been cut by narrow steep-walled ravines. Throughout its length, the geologic map (pl. 1) is speckled with a myriad of small claystone and mudstone outcrops. Within the landslide areas, no attempt was made to locate all of these small outcrops. Only those outcrops observed on limited field traverses through these areas have been shown on the map.

The generalized geologic map of the Socorro volcanic center (pl. 3) shows the broad areas underlain by upper Popotosa strata as if the mantle of surficial debris were not present. Except for small areas around Strawberry Peak and the Grefco mine, essentially all of these upper Popotosa Formation "outcrops" represent areas underlain by the red claystone and mudstone facies (Tpkp).

The red claystone and mudstone facies is believed to represent for the most part a thick body of playa deposits and some minor lacustrine deposits. The northwest and southeast boundaries of this Miocene playa are marked where the mudstones intertongue with distal alluvial fan deposits (Tpkf, Tpke, respectively). The general south margin of the playa occurs where the red mudstones thin abruptly onto fault blocks of older bedrock in the northern Chupadera Mountains.

The best exposures of the red playa facies (Tpkp) are found near the north boundary of the study area northeast of Strawberry Peak. Here, the poorly indurated fine-grained strata locally form low, rounded, badland type hills with negligible vegetation. Approximately 260 m of playa-facies strata, truncated by faults at the top and base of the section, are exposed just north of the map boundary (Lemitar Map).

Most outcrops of the playa facies consist of massive-appearing, reddish-brown or maroon claystone. Internal lamination or stratification is rarely observed; if present it is commonly masked by a crumbly "cauliflower" surface developed by alternating wetting and dessication events. The red claystones locally appear to grade into massive silty mudstones, however the latter lithology is less common. Grayish-yellow to pale-olive-gray claystone horizons, about 1-2 m thick, locally define the bedding in the massive red outcrops. Calcareous nodules are common with the greenish beds. Two 15-cm-thick, wavy-laminated, algal limestone beds are associated with greenish claystones in a small exposure southwest of "6001" Mesa. In this rare occurrence, the greenish strata probably represent deposits of shallow short-lived lakes. Three light-gray beds of coarsely crystalline, granular gypsum, as much as 3 m thick, occur in the well-exposed section just north of the map area. Gypsum also commonly occurs in the playa facies as fibrous satin spar veinlets, both subparallel to and cross-cutting the bedding. Many of the gypsum veinlets occur in small(?) faults and the satin-spar fibers show deformation, which may be used to infer the sense of movement on the fault assuming normal drag (simple shear). Bedded gypsum has not been observed south of Socorro Peak and satin spar veinlets generally disappear south of Highway Sixty.

Thin beds of pinkish-orange, friable, siltstones and fine-grained sandstones are common near the base of the playa facies or where it grades laterally to sandstones of the distal-fan facies. Beveled ripple marks have been observed in some of these siltstones. Reddish-brown, non-gypsiferous, silty mudstones are commonly associated with the siltstones.

Stratigraphic relations and thickness variations of the playa facies are complex, variable and generally obscured by Quaternary colluvium. Topographic relief on the landslide deposits surrounding the Socorro Mountains suggests that those slopes are underlain by at least 215 m of claystones and mudstones. A shallow thermal gradient bore hole at the northwest end of the Shrine Valley has shown that an additional 145 m of claystones fill this fault-block valley at the base of the landslide areas. Thus, the original depositional thickness of the playa claystones may have been as great as 360 m in the central part of the study area.

Northeast of Strawberry Peak, the playa facies (Tpkp) rests on the buff sandstone facies (Tpkf). Also, here the basalt of Kelly Ranch is interbedded in the lower part of the claystone section. The landslide masked claystone section capped by Strawberry Peak may be about 180 m thick.

In comparison, about 1 km north of Socorro Peak the playa facies is absent below the basalt of Kelly Ranch (Tpkb), which rests directly on lower Popotosa strata (Tpsl). Both the basalt and overlying claystones appear to wedge out rapidly toward the southeast. On the south side of "M" Mountain, the base of the silicic dome (Tsd) rests almost directly on lower Popotosa strata. This wedge of upper Popotosa claystones thickens westward (pl. 2, A-A') away from the late Miocene local bedrock highland (fig. 26) that now forms the east flank of Socorro Peak. The playa claystones also thicken eastward (away from this suballuvial bedrock high) as evidenced by their presence under the flows of Socorro Peak Rhyolite (Tsd) at the Merritt mine (Lasky, 1932; Smith, 1963). Similar stratigraphic relations indicate that this late Miocene bedrock high continued south to the Blue Canyon area and nearly to Socorro Canyon.

In the Northern Chupadera Mountains, the playa facies laps onto fault blocks from which the lower Popotosa mudflow conglomerate (Tpsd) was locally removed by erosion in middle(?) Miocene time. Near the Tower mine, no playa facies strata older than the Socorro Peak Rhyolite is present and locally these flows lie directly on the late Oligocene tuff of South Canyon (Tsc) or on lower Popotosa mudflows (Tpsd). Similar stratigraphic relations occur 5 km to the west in the eastern Magdalena Mountains southwest of Pound Ranch (Osburn, 1978). Together these areas are thought to mark the southern limit of thick playa facies mudstones and therefore indicate the southern structural margin of the late Miocene Popotosa basin. Small and thin exposures of red mudstones have been observed several kilometers to the south of the Tower mine (Osburn, 1979, oral commun.), indicating that the east-trending margin was complex (broken by north-trending fault blocks).

Playa mud sedimentation continued for some time after eruption of the Socorro Peak Rhyolite lavas. Small remnants of red mudstones and red muddy colluvial soils occur on top of the Socorro Peak Rhyolite lavas in the area west of Socorro Peak. Similar small patches of red muds rest on Oligocene volcanic rocks in the northern Chupadera Mountains. Much of the rolling-hill landscape west of Socorro Peak and in the northern Chupadera Mountains probably represents topography recently (geologically) exhumed from under a thin blanket of clay and mud. Several red mudstone sections that have locally been preserved above Socorro Peak Rhyolite lavas and tuffs may be observed on the geologic map. They range from 15 to 90 m in thickness (pl. 2, D-D', E-E', F-F', H-H'). A few soft, white pumice fragments, petrographically similar to Socorro Peak Rhyolite lavas, are common in these younger mudstone outcrops, especially south of Blue Canyon.

Basalt of Kelly Ranch (Tpkb north of Socorro Peak). A series of thin basaltic flows, which are interbedded in the lower part of the upper Popotosa Formation, and exposed in a large area west of the J.B. Kelly Ranch are here informally referred to as the basalt of Kelly Ranch. In the study area (pl. 1), these greenish-black, ledge-and cuesta forming basaltic flows crop out along Strawberry Canyon, at the east box of Nogal Canyon, about 1 km east of the confluence of the above canyons, and about 1 km north of Socorro Peak. Additional exposures of the basalt of Kelly Ranch occur along the southeast side of the Lemitar Mountains as far north as Corkscrew Canyon (fig. 2).

Outcrop relations in the Strawberry Canyon area suggest that, locally, the basalt flows occur at three different stratigraphic positions in the sedimentary strata. The oldest flow forms a small lenticular outcrop in the floor of the canyon where it appears to fill a channel in the buff sandstone facies (Tpkf). The north wall of the channel is fairly well exposed; also, here the overlying channel sandstones have unusually abundant basaltic clasts. The youngest flow is exposed on the south side of the canyon near it's east end. The youngest flow, about 12 m thick, appears to be locally preserved on the downthrown (east) side of two north-trending fault blocks; further west it is absent. The youngest flow is interbedded in the playa claystone facies (Tpkp) and is separated from the middle stratigraphic horizon of basalt by about 50 m of claystones.

The middle stratigraphic horizon of the basalt of Kelly Ranch is the most widespread and may be as much as 35 m thick. It is typically comprised by 2, to as many as 4, thin flows stacked one upon the other. The individual flows have autobrecciated, scoriaceous tops and bottoms and range from 5 to 15 m in thickness. Friable, buff-colored aeolian(?) sand locally fills crevices and forms partings between flows. Greenish-gray, baked mud chips are fairly common in these basalt flows, particularly near the base of the sequence. Black calcite and milky quartz are common as vesicle fillings.

The outcrops of the basalt of Kelly Ranch outside the Strawberry Canyon area are tentatively correlated with the middle-flow horizon. The north-trending outcrop north of Nogal Canyon consists of a single flow that apparently wedges out progressively to the north onto a low topographic swell built on mudstones. Immediately north of the map boundary, this flow reappears along the same trend. From here, it progressively thickens northward into a stack of as many as 4 flows and continues along strike for at least 2.5 km. The exposure north of Socorro Peak consists of a double stack of relatively vessicular and oxidized flows, but otherwise it is petrographically and stratigraphically similar to the basalt of Kelly Ranch at Strawberry Canyon.

All of the flows assigned to the basalt of Kelly Ranch appear to occur stratigraphically below the rhyodacitic to quartz latitic flows (Tsd) and tuffs (Tsdt) of the Socorro Peak Rhyolite. The Strawberry Peak dome (Tsd) and a related tuff (Tsdt, small outcrop north of Nogal Canyon) have been dated at 11.8 m.y and 11.9 m.y. respectively (Appendix B). A fresh sample (No. 76-1-9, table 6) of the basalt of Kelly Ranch, taken from the outcrop north of Nogal Canyon, has been dated by H. Mehnert (U.S. Geological Survey) at 9.1±0.5 m.y. (Machette, 1978). This date is believed to be too young, since north of this locality the same trend of basalt flows is evidently overlain by the 11.9-m.y.-old "Tsdt" tuff. The basaltic sample contains traces of secondary carbonate and antigorite after olivine and also some quartz xenocrysts. Near its base, this flow contains some mud chips, but these are not apparent in the date sample. Unrecognized contamination with potassic clays could have affected this whole-rock age date. This sample contains about 53.6 percent SiO_2 (Table 6) and may be classified as a xenocrystic basaltic andesite. Uncontaminated equivalents are likely to be true basalts, although additional analyses are needed to verify this.

In thin section and handspecimen, the basalt of Kelly Ranch is normally grayish-black to medium gray, or grayish red where oxidized. Sparse, clear plagioclase laths and traces of yellowish-green olivine are the only phenocrystic minerals. In most specimens, the olivines are altered to reddish-brown rims of iddingsite and cores of yellowish antigorite. Gabbroic xenoliths and a few unusually large plagioclase "phenocrysts"(?) have been observed in a few specimens. In dense samples, the matrix typically has a fine-grained diabasic (intersertal) texture. The description of sample 76-1-9 (Table 6) generally applies to most samples of the basalt of Kelly Ranch, except for the degree of alteration of the olivines.

A source area for the basalt of Kelly Ranch has not been identified. Basalt flows, which occur between the tuff of South Canyon and equivalents of the Socorro Peak Rhyolite, (Pound Ranch lavas) have been mapped in the northeastern Magdalena Mountains (Osburn, 1978). These basalt flows occur at roughly the same stratigraphic position as the basalt of Kelly Ranch. However, they are not interbed-

Table 6. Analyses of basaltic lavas interbedded in the Santa Fe Group in the Socorro Peak area. Chemical analyses by x-ray fluorescence; D. L. White, New Mexico Bureau of Mines and Mineral Resources (see p. 28).

Sample No.	76-1-9	77-5-3	76-6-11
Formation	upper Popotosa	upper Popotosa	Sierra Ladrones
Unit	basalt of Kelly Ranch	basalt of Bear Canyon	basalt of Sedillo Hill
	Major Oxides (weight percent)		
SiO_2	53.61	47.43	50.15
Al_2O_3	15.79	16.30	12.21
Fe_2O_3 (total)	8.75	10.65	10.79
MgO	8.51	7.63	8.41
CaO	7.40	7.50	8.49
Na_2O	3.41	5.16	3.06
K_2O	2.16	1.24	1.70
TiO_2	0.35	1.82	1.28
Total	99.98	97.73	96.09

Sample Descriptions

76-1-9 Grayish-black, dense, slightly porphyritic, xenocrystic, basaltic andesite. Contains 5 percent small (1-2 mm) plagioclase (An_{55-65}) laths, some with narrow reverse zoned rims (An_{70}), and traces of subhedral olivine (0.5-1 mm) slightly altered to antigorite in a pilotaxitic groundmass of plagioclase microlites, olivine, clinopyroxene and Fe-Ti oxides. Contains a few rounded and embayed quartz (2-3 mm) xenocrysts and trace of carbonate. Just north of Nogal Canyon, east of confluence with Strawberry Canyon. (SE/4, NW/4, Sec. 29, T2S, R1W)

77-5-3 Dark-gray, dense, aphanitic, ophitic basalt. See fig. 28 for description and location.

Table 6. (continued)

76-6-11 Dark-gray, dense, slightly porphyritic, olivine basalt. Contains 3 percent small (0.5-1 mm) greenish-yellow-brown olivine phenocrysts and small bluish-black subhedral clinopyroxene (augite?) microphenocrysts in a fine-grained granular groundmass of plagioclase, clinopyroxene, and Fe-Ti oxides. Contains one muddy carbonate xenolith with clinopyroxene reaction corona and a trace of olivine altered to chlorite, calcite and a pale-brown mica (muscovite?). From landslide block on southwest side of Black Mountain (Mesa) (SE/4, SE/4, Sec. 25, T3S, R2W).

ded in sedimentary strata of the Popotosa Formation, apparently because they occur south of the basin margin.

<u>Muddy conglomerate facies (Tpkc)</u>. In the area from Blue Canyon to Socorro Canyon, the basal unit of the upper Popotosa Formation consists of 0-25 m of poorly indurated, reddish-colored, muddy to sandy conglomerates (Tpkc). This coarse, unsorted debris is believed to represent local sheet-wash deposits that mixed with playa muds as they onlapped a local bedrock highland in middle to late Miocene time (fig. 26). Southwest of "6001", Mesa the muddy conglomerate contains only cobbles of aphanitic andesite lava apparently derived from an adjacent bedrock outcrop (Tla_1?). However, at other outcrops, most of the coarse material in this unit appears to be reworked from the lower mudflow conglomerates (Tpsd). Where the matrix is sandy, these outcrops are similar in appearance to the upper red conglomerate facies (Tplru). Nevertheless, the muddy conglomerates (Tpkc) are considered to be younger than the upper red-conglomerate facies (Tplru); the latter underlies the playa facies (Tpkp) which appears to be coeval with the muddy conglomerates.

Pale-red conglomeratic sandstone facies (Tpke). This facies, which is informally referred to on the geologic map (pl. 1) as the fanglomerate near Evergreen Ranch, is exposed discontinuously in the southeastern portion of the Socorro Mountains between Blue Canyon and Socorro Canyon. The pale-red conglomeratic sandstones are considered to represent distal alluvial-fan deposits derived from the eastern border of the Popotosa basin in late Miocene time. This interpretation is supported by westerly paleocurrent directions (pl. 1) and clast lithologies indicative of a source area east of the present Rio Grande. Trace amounts of well-indurated, red siltstone and fine-grained sandstone pebbles have been observed in nearly every area that this volcanic-rich facies crops out. The tabular, thinly laminated siltstone clasts are very similar to siltstones that form outcrops of the Permian Abo Formation (Wilpolt and Wanek, 1952) now exposed on the Loma de las Canas uplift east of Socorro (fig. 2).

Stratigraphic relations of the pale-red conglomeratic sandstones are variable and indicate that deposition of this facies occurred both before and after local eruptions of the Socorro Peak Rhyolite. At the east foot of Pinnochio Peak, the poorly indurated pale-red conglomeratic sandstones (Tpke) are interbedded in red playa muds (Tpkp), which evidently underlie this rhyodacitic flow (Tsd) that is petrologically equivalent to the 10.3-m.y.-old rhyodacite of "6001" Mesa (Appendix B). Northeast of the Grefco mine, the moderately well-indurated, pale-red conglomeratic sandstones form low cliffs and east-dipping cuestas where they are interbedded with vitrophyric sedimentary rocks (Tpkr) shed from the 7.4-m.y.-old Grefco rhyolite dome (Appendix B). Near Socorro Spring, the reddish conglomeratic sandstones are interbedded in red mudstones (Tpkp) which overlie a thick tuff blanket (Tsrt) associated with the Grefco dome. These are the youngest strata of the Santa Fe Group which can be shown to predate inception of the ancestral Rio Grande (Bachman and Mehnert, 1978). Therefore, they are assigned to the upper Popotosa Formation.

The older strata of this reddish-conglomeratic sandstone facies, which are also exposed southwest of the Grefco mine near Socorro Canyon, are generally 12-15 m thick. Younger strata at Socorro Spring are about 30 m thick and, northeast of the Grefco mine, this facies may be as much as 75 m thick.

The best exposures of the "Tpke" facies are man-made cuts at the earthen tank northeast of Socorro Spring and in the railroad cut at the Grefco perlite mill. At Socorro Spring, this facies consists largely of pale-red and moderate-reddish-orange, medium-to fine-grained sandstones that are poor to moderately indurated. About 15 to 25 percent of the outcrop is formed by lenticular pebbly conglomeratic sandstones generally 7-15 cm thick. These appear to mark scoured channel bottoms. The better exposed strata appear to be crudely planar bedded. Clast lithologies here are similar to those observed in a pebble count at the railroad cut outcrop next to the Grefco mill.

Detritus from a wide stratigraphic range of regional Oligocene volcanic units seems to be present near the Grefco mill. In decreasing order of abundance, the majority of clasts are similar to the A-L Peak Tuff, intermediate lavas of the Spears Formation (or Luis Lopez Formation), and La Jara Peak Basaltic Andesite lavas. Less abundant (5-10 percent each) clast types are similar to the tuff of South Canyon, Hells Mesa Tuff, and upper Lemitar Tuff. Some clasts counted as Hells Mesa could be derived from the uppermost quartz-rich Lemitar Tuff and some clasts counted as La Jara Peak lavas could be from basaltic lavas in the Santa Fe Group. About 2 percent of the smaller pebbles in the pebble count area are reddish granitic rocks or dark biotite schists. These may have been recycled from the Permian Bursum Formation or, less likely, derived from relatively small Precambrian granite outcrops now exposed east of the

Rio Grande. Traces of red siltstone clasts observed at Socorro Spring and several other outcrops were not present in the square meter of the pebble count area. Punky, white pumice(?) or rhyolitic lava fragments, similar to the Socorro Peak Rhyolite lavas, form a few percent of the clasts at Socorro Spring, but are absent at the Grefco mill.

Rhyolitic conglomeratic sandstone facies (Tpkr). The Grefco "perlite" dome (Tsr) is flanked by a conical wedge of vitrophyric conglomeratic sandstones mapped here as a unique, locally derived facies (Tpkr) of the upper Popotosa Formation. The southeastern flank of the Grefco dome (Tsr) and its adjacent glassy alluvial wedge have apparently been downfaulted out of view (pl. 2, F-F').

Where the outer edge of the vitric sedimentary wedge is exposed, the light-gray ledge-forming outcrops of the glassy conglomeratic sandstones rest on a thick blanket of slope-forming, white, bedded tuffs (Tsrt). These tuffs represent initial pyroclastic eruptions related to the Grefco dome. In previous descriptions of this area (Weber, 1963; Lowell, 1967), the equivalents of these two units (Tpkr and Tsrt) were both described as pyroclastics.

At the south side of the Grefco dome, the base of the vitric sedimentary facies is well-exposed where it apparently lapped off the steep front of the rhyolitic lava. Here, the flow-banded lava grades up into a blocky vitrophyre breccia which, in turn, grades to a crudely bedded coarse-grained conglomeratic sandstone composed almost entirely of angular to subangular, vitric rhyolite detritus. Small rounded clasts of mafic scoria may also be rarely observed in some adjacent outcrops.

Northeast of the Grefco dome, the distinctive detrital components of the light-gray vitric lithofacies (Tpkr) and the pale-red distal-fan lithofacies (Tpke) are intimately intermixed and interbedded. These strata are separated on the basis of the components that are dominant, which is reflected by the overall color of the individual beds.

Outcrops of the vitrophyric conglomeratic sandstones are moderately to well indurated (siliceous?) and generally form light to medium-gray ledges, although yellowish-brown limonite stained beds are also common, particularly near the base of the unit. Crude parallel bedding is dominant; however, some planar(?) crossbedding and cut-and-fill structures are also common. The clean, sandy matrix of these rocks is almost entirely composed of angular to subangular, and medium- to very-coarse-grained fragments of rhyolite vitrophyre. Vitrophyre clasts are generally subangular and subequant and therefore are only rarely imbricated. Angular fragments of grayish-red lithoidal rhyolite similar in phenocryst mineralogy to the vitric rhyolites occur as minor clasts. Subrounded basaltic pebbles and some crystal-rich tuff fragments (Lemitar Tuff?) have been rarely observed in the vitrophyric conglomeratic sandstones.

West of the Grefco mine, exposures of the vitric sandstones appear to thin rapidly away from the dome. The unit is at least 75 m thick on the northwest side of the dome and thins to less than 30 m in the section south of "6001" Mesa.

Locally derived rhyolitic sandstones and conglomerates (Tpkr) similar in age and origin to the vitric sandstones at the Grefco mine have been observed at three other localities in the study area. These rhyolitic sedimentary rocks crop out 1) west of Cook Spring, 2) north of the Tower mine near Bear Canyon, and 3) near the vent area for the basalt of Sedillo Hill (Tbsh). At the latter locality, the poorly exposed rhyolitic conglomerates are surrounded by basaltic colluvium derived from the cliff-forming flows (Tbsh) at the top of the slope. These rhyolitic beds are probably intercalated in red playa muds (Tpkp). They contain clasts petrologically similar to the rhyolite lavas of the Tripod Peak dome and the "6633" dome to the northeast, which is therefore interpreted as the most likely source area. At the first two localities mentioned above, the immature rhyolitic sedimentary rocks are clearly derived from the immediately underlying lava flows.

Basalt of Bear Canyon (Tpkb, south of Highway Sixty). A thin, petrographically distinctive basalt flow, which is interbedded in upper Popotosa playa claystones (Tpkp) in the northern Chupadera Mountains, is here termed the basalt of Bear Canyon for numerous low cuesta-like exposures along this drainage. This 5-10 m-thick basalt flow also forms black to brownish-black ledges, rounded hills and cuestas in the Shrine Valley. One relatively isolated outcrop occurs nearly a kilometer south of Chupadera Spring. Except for one small outcrop in Socorro Canyon, the basalt of Bear Canyon has not been observed north of Highway Sixty.

Spatial relations of outcrops of the basalt of Bear Canyon to numerous isolated patches of red claystones (Tpkp) suggest that the flow is interbedded in the playa deposits; however, contact relations are rarely well exposed. The Bear Canyon flow is fairly well exposed in a road cut on the south side of Highway Sixty. Here, it is overlain by about 20 m of gently west-dipping green and red claystones and mudstones. In a thermal-gradient drill hole west of this road cut, this basalt flow formed a 6 m intercept, approximately in the middle of 140 m of playa claystones (C. E. Chapin, unpub. drill log). The hole bottomed against "hard rock" presumed to be the lower Popotosa mudflows (Tpsd), although representative cuttings were not obtained to confirm this.

North of the Tower mine, and in the valley to the west, the basalt of Bear Canyon appears to overlie playa deposits, which rest on quartz latitic lavas (Tsd) assigned to the Socorro Peak Rhyolite. Neither the basalt flow nor the quartz latitic lavas have been dated. However, the latter is similar in petrology and in stratigraphic relations to silicic lavas mapped in the adjacent Pound Ranch area by Osburn (1978, p. 144), which have a K-Ar age of 10.5±0.4 m.y. The basalt of Bear Canyon is therefore tentatively assigned a late(?) Miocene age, and is considered to be younger than the basalt of Kelly Ranch and older than the basalt of Sedillo Hill. The basalt of Bear Canyon and the Grefco "perlite" dome occur at about the same stratigraphic level, but do not overlap, so their mutual stratigraphic relations are uncertain. The source area of the basalt of Bear Canyon is unknown. Small isolated outcrops of basalt similar to the Bear Canyon occur in the alluvial valley west of the Chupadera Mountains as far south as Nogal Canyon (Walnut Creek, fig. 2) (Osburn, 1979, oral commun.).

The basalt of Bear Canyon is a distinctive grayish-black, dense, finely saccharoidal (diabase-like), basalt that is characterized by a well developed ophitic and minor hyaloophitic texture. The interlocking ophitic texture is most likely why this rock has a tough elastic character, which is readily apparent when attempting to break it with a hammer.

The reader is referred to the caption and photomicrograph of figure 28 for a representive description of the petrography of this uniform basalt flow. In hand specimen, the glassy pockets in the rock appear as milky white irregular blebs. These glassy pockets, which may range from 5 to 15 percent of the rock's volume, commonly give freshly broken surfaces a white, speckled appearance. Locally, this basalt flow may contain traces of small, clear plagioclase phenocrysts. In a few areas, a vessicular zone has been observed near the top of the flow. Chemically (table 6) this flow appears to be an alkali basalt with about 47.4 percent SiO_2. However, this sample contains a few small vesicles filled with a white acicular zeolite(?), which could be analcite. This would help explain the unusually high soda content of the rock, 5.2 percent.

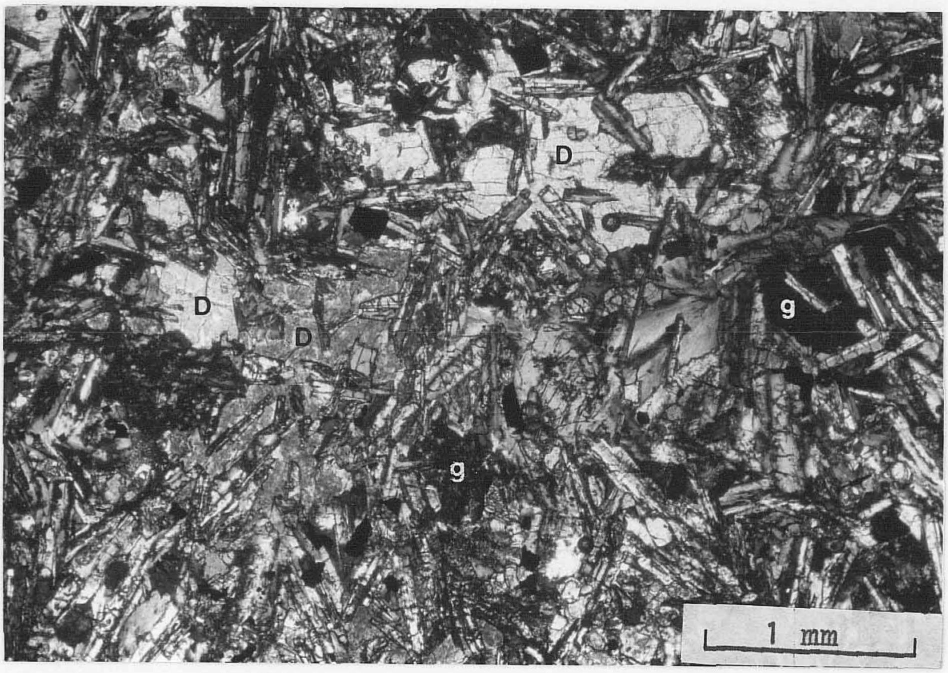


Figure 28. Photomicrograph of the basalt of Bear Canyon. This dense ophitic and hyaloophitic basalt lava consists of abundant labradorite laths (An_{64-67}) floating in a framework of subhedral diopside (D) or augite (?) crystals. Also present are sparse amounts of olivine (small rounded crystals), euhedral Fe-Ti oxide microphenocrysts (black), and small patches of partly devitrified glass (g). Crossed nicols. At water cut through low cuesta-forming basalt flow in the Shrine Valley (SW corner, NW/4, NW/4, Sec. 5 T4S, R1W). Sample No. 77-5-3.

Socorro Peak Rhyolite

In a preliminary report on the geology of the Socorro Mountains, Smith (1963) referred to aphanitic porphyritic rocks of peculiar composition that form the main mass of Socorro Peak and the adjacent high points of the Socorro Mountains as the "Socorro Peak calcitrachyte". Field observations of these porphyritic volcanic rocks has now established that they are mostly thick silicic lava flows. Petrographic study and preliminary chemical analyses of these lavas (Table 7) has indicated that they form a comagmatic calc-alkaline suite with an alkali-lime index of approximately 58. Most of the rocks analysed are vitrophyres and may reasonably be expected to contain a few percent of water. The Grefco rhyolite vitrophyre (industrial perlite) is known to contain more than 3 percent water (Weber, 1963a). A rough estimate of the SiO_2 content of these lava flows as anhydrous rocks (Table 7) indicates a wide range in composition, ranging from rhyodacite (62 percent SiO_2) to high-silica rhyolite (77 percent SiO_2). Based on petrographic similarity with the analysed rocks, field observations (pl. 1) indicate that the rhyolitic lavas are somewhat greater in volume than the rhyodacite or quartz latite lavas. Therefore, this heterogeneous, but dominantly rhyolitic, complex of lavas and associated pyroclastic

Table 7. Analyses of lava flows of the Socorro Peak Rhyolite. Chemical analyses by x-ray fluorescence, D. L. White, New Mexico Bureau of Mines and Mineral Resources (see p. 28). Trace = tr. See Appendix B for radiometric age analytical data.

Sample No.	77-5-1	76-4-1b	76-6-1	76-6-3	76-6-2a	76-6-2b	77-5-4
Map Unit (pl. 1)	Tsd	Tsd	Tsd	Tsr	Tsr	Tsr	Tsr
Related vent or domal flow	"6001" Mesa	Radar Peak (Socorro Peak)	Strawberry Peak	"6633" Peak	Signal Flag	Signal Flag	Grefco Mine
K-Ar age in m.y. (Appendix B)	10.3	11.5 & 10.7	11.8	9.0	10.5		7.4
Major oxides (weight percent)							
SiO_2	60.16 (62.3)[1]	68.28 (69.2)[1]	68.50 (71.6)[1]	69.79 (70.6)[1]	75.06 (75.5)[1]	73.33 (76.5)[1]	73.86 (77.4)[1]
Al_2O_3	15.03	15.84	14.38	14.83	13.90	12.72	13.17
Fe_2O_3 (Total)	3.95	3.48	3.17	2.97	2.15	1.14	0.20
MgO	6.66	1.91	1.14	1.12	0.34	0.63	<0.01
CaO	4.58	2.93	2.66	2.01	2.00	1.90	0.57
Na_2O	3.70	2.20	1.98	3.39	2.11	2.97	2.86
K_2O	1.79	3.57	3.55	4.40	3.61	3.11	4.86
TiO_2	0.73	0.47	0.14	0.38	0.15	0.07	0.06
Total	96.60	98.68	95.63	98.89	99.32	95.87	95.59
Phenocrysts (volume percent)							
plagioclase	8	6	7	8	5	4	>½
(percent anorthite)	(65-40)	(60-40)	(50-30)	(50-30)	—	(32-28)	(26-24)
sanidine	-	-	-	3	10	7	<½
quartz	-	-	-	1	4	3	tr.
hornblende	3	1	1	tr.	-	tr.	-
biotite	tr.	1	2	3	1	<½	tr.
opaques	>½	<½	tr.	tr.	tr.	tr.	-
groundmass	88+	91+	90	84+	80	85+	99

1. Estimated silica content of anhydrous rock based on an assumption that the total of the major oxides subtracted from 100 percent is roughly equivalent to the water content of the rock.

Table 7. (continued)

Sample No.	Sample Descriptions
77-5-1	Medium-light-gray, waxy, dense, trachytic, hornblende rhyodacite. See fig. 30A for additional description. Southeast side of "6001" Mesa near exposed base of flow. (SE/4, SW/4, Sec. 21, T3S, R1W)
76-4-1b	Pale-red, lithoidal, frothy, biotite-hornblende quartz latite. Contains sparse phenocrysts of strongly resorbed (spongy texture) and zoned plagioclase (1-5 mm) (fig. 30B), minor biotite (0.5-2 mm) and hornblende (0.5-2.5 mm) in a partly glassy to cryptocrystalline groundmass with numerous ragged (pumice like) vugs. Small hornblende needles mostly oxidized to opaque Fe oxides; some larger hornblendes show reaction relationship to form biotite. North side of Radar Peak (center SW/4, Sec. 5, T3S, R1W)
76-6-1	Medium-gray, waxy, partly perlitic, dense, trachytic, biotite-hornblende, quartz latite or rhyolite (?). Contains sparse, subhedral phenocrysts of resorbed (spongy texture) and zoned plagioclase, (0.5-3 mm) and minor biotite (0.2-2 mm) and hornblende (0.2-2 mm) in a glassy groundmass dotted with sparse, small spherulites. Feldspar microlites moderately abundant in groundmass. Crest of big slide-block due east of Strawberry Peak. (NE/4, SE/4, Sec. 25, T2S, R2W)
76-6-3	Light-gray, dense, phenocryst-rich, quartz latite to rhyolite vitrophyre. Contains moderately abundant, strongly resorbed (rounded) plagioclase (0.5-1.5 mm) and sanidine (1-1.5 mm) phenocrysts with minor biotite (0.2-1.0 mm) and sparse subhedral quartz (0.3-0.8 mm) in a slightly perlitic, glassy groundmass. Contains traces of hornblende, sphene and zircon. Large slide block (Qlr, pl. 1) at old railroad quarry in Socorro Canyon (NW/4, SW/4, Sec. 13, T3S, R2W). Block apparently derived from "6633" Peak domal flow.

Table 7. (continued)

Sample No.	Sample Descriptions
76-6-2a	Light greenish-gray with streaks of light brownish-gray, dense, flow-banded, phenocryst-rich rhyolite. Contains abundant phenocrysts of stubby sanidine (2-4 mm), plagioclase (1-3 mm), rounded and embayed quartz (1-3 mm) and minor biotite (0.5-2.5 mm) in an incipiently devitrified groundmass with some relict perlitic cracks and some crypto-crystalline streaks (brownish streaks). Contains traces of sphene and zircon. Roadcut at base of Signal Flag tower used by TERA in Blue Canyon (NW corner, SW/4, SE/4, Sec. 16, T3S, R1W)
76-6-2b	Yellowish-gray, dense, moderately phenocryst-rich, rhyolite pitchstone. Phenocryst mineralogy and texture like No. 76-6-2a. Groundmass is perlitic glass. Contains trace of hornblende. Same location as No. 76-6-2a.
77-5-4	Light-gray, dense, finely flow-banded, very phenocryst-poor, high-silica rhyolite vitrophyre. Contains a few small phenocrysts of plagioclase and sanidine with a trace of quartz and biotite in a striated, slightly frothy, glassy groundmass. From "ore" stockpile at Grefco mill. Mined from east center of Grefco "perlite" dome (NE/4, NE/4, Sec. 28, T3S, R1W).

rocks is here renamed the Socorro Peak Rhyolite.

Silicic lavas similar in petrology and age to the Socorro Peak Rhyolite (fig. 2) also occur along the western flank of the Magdalena Mountains (Weber and Basett, 1963; Chapin and Seager, 1975, p. 315; Allen, 1979) and in the northeastern Magdalena Mountains ("Pound Ranch lavas" of Osburn, 1978, p. 78; Petty, 1979). The Socorro Peak Rhyolite and the "Pound Ranch lavas" of Osburn (1978) are here considered to be stratigraphically equivalent, even though they probably did not form a laterally continuous pile. These late Miocene lavas have been faulted and locally buried by younger Santa Fe sedimentary strata. Nevertheless, their present outcrops appear to define local accumulations of viscous lavas around different eruptive centers. Small erosional remnants of a silicic flow from the Pound Ranch center locally crop out in the southwest corner of the study area near the Tower mine (Tsd, pl. 1). For the sake of convenience, these isolated exposures are included with the Socorro Peak Rhyolite.

The Socorro Peak Rhyolite represents a heterogeneous complex of domal lava flows and minor pyroclastic rocks erupted onto the floor of the Popotosa basin, beginning about 12 m.y. ago and ending about 7 m.y. ago (fig. 26). Filling of the late-stage Popotosa basin with playa muds and alluvial-fan deposits occurred prior, during, and after this period of silicic volcanism. Hence, the lenticular flows of the Socorro Peak Rhyolite and associated pyroclastic deposits are locally interbedded at different stratigraphic horizons within the upper Popotosa Formation. The rounded hills of the western Socorro Mountains essentially represent constructional volcanic topography of several silicic domes and flows, which most likely were exhumed from under a cover of playa muds in late Pliocene or Pleistocene time.

The bases of flows of the Socorro Peak Rhyolite are rarely exposed. Where the flows occur on relatively uplifted blocks, their eroded edges are marked by cliffs by flanking talus cones (Qca). The talus deposits grade downslope into numerous landslide blocks (Ql) underlain by incompetent playa muds (Tpkp) of the upper Popotosa Formation. Some large, incipient landslide blocks, only partly detached from their parent flows, have been mapped as bedrock blocks in the area south of Nogal Canyon and near "6633" Peak. East of Strawberry Peak, one unusually large slideblock and several small blocks of lava are arbitrarily separated (pl. 1) from the surrounding landslide blocks in order to show their geometric relationship or internal structure.

Stratigraphic relationships along the east flank of the Socorro Mountain block indicate that, in late Miocene time, the Socorro Peak lavas locally flowed unconformably across older or contemporaneously tilted fault blocks (pl. 2, section E-E'). These tilted blocks did not disrupt the basin floor. Instead they apparently formed sub-alluvial bedrock benches (fig. 26), much like the many shallowly buried fault blocks that occur surrounded by alluvium in the modern basins of the Rio Grande rift (Woodward and others, 1978). Thus, the Socorro Peak Rhyolite lavas may either rest conformably on thick sections of Popotosa playa muds (concealed by landslides) or unconformably on lower Popotosa conglomerates or Oligocene volcanic rocks. Continued rifting since late Miocene time has locally uplifted, downdropped, and tilted the Socorro Peak flows in various directions. These domal flows and their underlying vent structures are now exposed over a wide range of elevations and erosion levels.

The Socorro Peak Rhyolite was erupted from more than twenty vents, many of which are tightly grouped into small complexes. Overall, these vent structures appear to be concentrated along the buried northern margin (ring-fracture zone) and moat of the Socorro cauldron. The vents form a north-northwest-trending intrusive belt, which extends for more than 11 km from the Grefco mine to Strawberry Peak. The belt is widest where it crosses the projected ring-fracture zone. Linear chains of three or more volcanic vents are common on the geologic map (pl. 1). Most of these trend north-northwest, paralleling the strike of major rift faults; however, others also trend east-northeast and west-northwest. The latter trends respectively parallel the transverse shear zone (Morenci lineament) and the local trend of the ring-fracture zone of the Socorro cauldron.

Vents for the Socorro Peak Rhyolite that are shown on the geologic map (pl. 1) have been recognized by the following commonly accepted criteria (Williams, 1932): 1) roughly concentric flow foliations that form steeply dipping funnel or bulbous patterns around the vent; 2) an association with local topographic highs representing either original volcanic dome topography, eroded remnants of domes held in place (from landsliding) by intrusive underpinnings, or resistant, eroded volcanic necks; 3) roughly concentric patterns of intra-folial gash fractures (stretch marks, pl. 1) related to radial flow away from the point source vents; and 4) hydrothermal alteration or minor vein structures within or close to the vent. Most of the vents shown on the geologic map meet two or more of these criteria. Undoubtedly, some vents have been overlooked where exposures

PETROLOGIC UNITS OF THE SOCORRO PEAK RHYOLITE

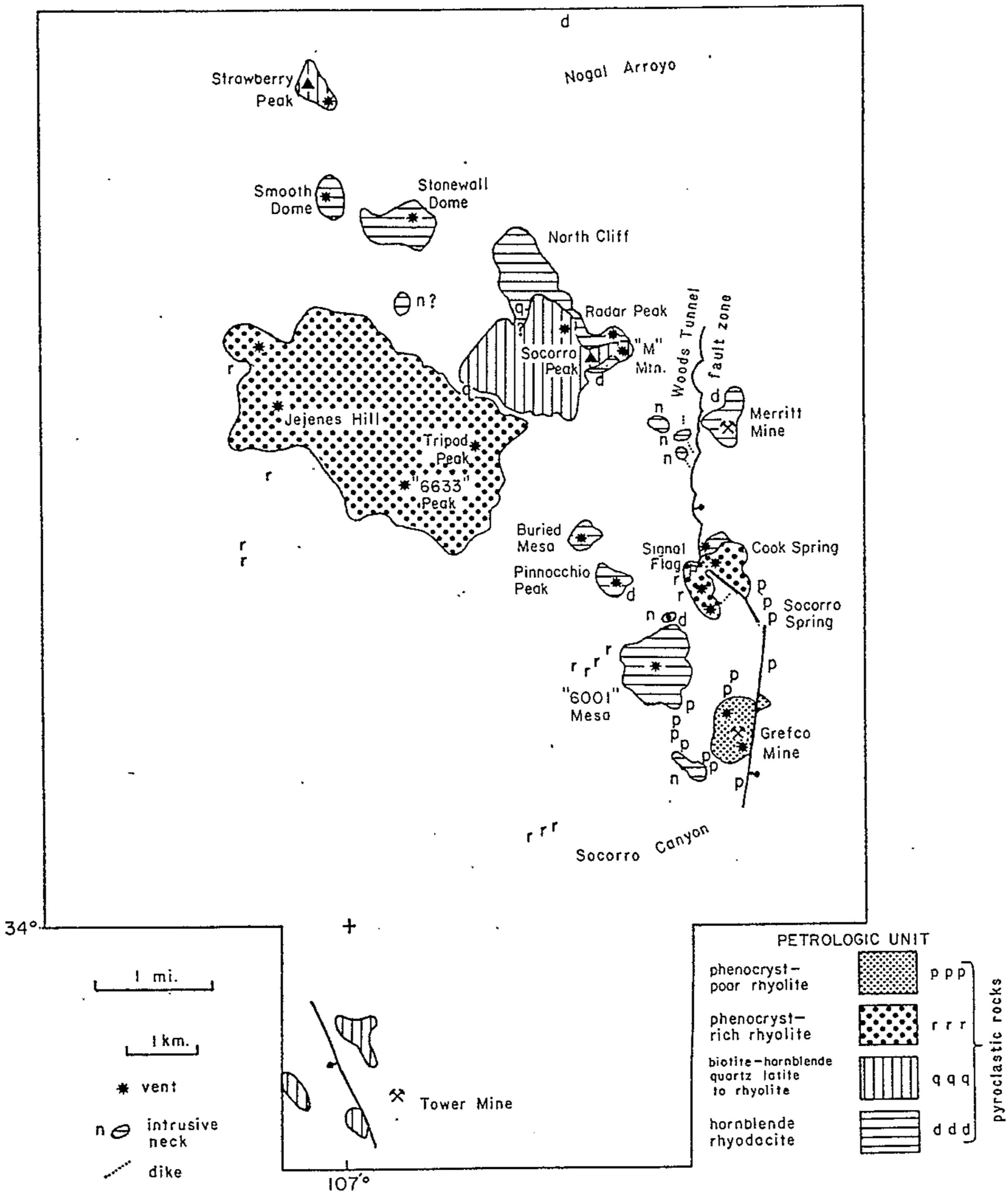


Figure 29. Distribution of petrologic units of the Socorro Peak Rhyolite subdivided into their respective lavas, pyroclastic rocks, and related intrusive rocks. Based on additional differentiation of map units: Tsd, Tsdt, Tsr, and Tsrt shown on Plate 1.

are poor and some of the less well-documented vents may be fictitious; presumably the number in either category is small.

Many vent structures expose both intrusive and extrusive relationships in a continuous outcrop. Only those structures which have been completely isolated by erosion from their eruptive counterparts are shown on the geologic map as intrusive bodies. These intrusives are commonly associated with variably hornfelsed Popotosa sedimentary rocks around their periphery.

Lavas and related pyroclastic rocks of the Socorro Peak Rhyolite are shown on the geologic map (pl. 1) as undifferentiated rhyodacites and quartz latites (Tsd, Tsdt) and undifferentiated rhyolites (Tsr, Tsrt). The units of Plate 1 have been differentiated into the four informal petrologic units shown on figure 29 in order to facilitate their description. In ascending stratigraphic sequence, the units consist of: 1) a hornblende rhyodacite unit, 2) a biotite-hornblende quartz latite-rhyolite unit, 3) a phenocryst-rich rhyolite unit, and 4) a phenocryst-poor rhyolite unit. Representative samples of these units are described in the same order in Table 7. This stratigraphic sequence is based primarily on sparse observations of superposition relationships which are generally illustrated on figure 29. The quartz latite-rhyolite unit is not overlapped by strata that can be related to the other units of the Socorro Peak Rhyolite. Therefore, the phenocryst-rich and the phenocryst-poor rhyolite units are considered to be younger than the quartz latite-rhyolite unit solely on the basis of their respective radiometric age dates (Appendix B). Radiometric age data of the rhyodacite unit and the quartz latite-rhyolite unit (Appendix B) are not entirely consistent with the observed stratigraphic sequence. However, most of these dates have overlapping error bars and the data are permissive of the preferred interpretation, namely that the rhyodacite and quartz latite-rhyolite units are penecontemporaneous. The following descriptions are based on figure 29, which may be cross referenced with plate 1.

Hornblende rhyodacite unit

The hornblende rhyodacite unit of the Socorro Peak rhyolite consists predominantly of medium-gray and reddish-brown lavas; it also includes minor volumes of pumiceous tuffs and agglomerates, and some intrusive necks. These rhyodacitic rocks are characterized by a bimodal population of sparse, medium-grained and moderately abundant, fine-grained phenocrysts of plagioclase and hornblende. The rhyodacite unit forms a north-northwest-trending belt of discontinuous outcrops mostly between "6001" Mesa and the Stonewall dome in Nogal Canyon.

The oldest rocks of the rhyodacite unit consist mostly of reddish-brown or light-gray, bedded lapilli tuffs, blocky agglomerates and thin, dark-red laharic breccias which locally underlie the associated rhyodacitic flows. Outcrops of this type (Tsdt, pl. 1) occur on the south side of Socorro Peak, northwest of the Merritt mine, on the south side of Pinnocchio Peak, and on the northeast flank of "6001" Mesa.

The exposure south of Socorro Peak is the most revealing; here, about 15 m of reddish-brown, crudely sorted pumice and ash-fall beds dip northward under the autobrecciated base of the overlying rhyodacitic lava flow (Tsd, pl. 1). Some of these agglomerate beds contain vesiculated blocks approaching a meter in length. Many of the beds fine upwards and some are capped by thin mudstone partings. A dusky red, well-indurated, laharic breccia less than a meter thick separates the bedded air-fall deposits from the 10 m-thick autobrecciated base of the overlying lava flow. The mudflow contains some light-colored pumice and a few, dense, dark-colored lava fragments. The early rhyodacitic pyroclastic rocks appear to overly a thick section of playa muds as indicated by the large landslide area to the south of Socorro Peak (pl. 1).

Intensely silicified rhyodacitic(?) lapilli tuffs and a thin laharic breccia appear to have once underlain hydrothermally altered rhyodacitic lavas in an area northwest of the Merritt mine. This area may have been associated with hot springs at the edge of the rhyodacite flow in late Miocene time. The tuff beds here are almost entirely replaced by cryptocrystalline and irregularly banded white silica. Some incompletely silicified pumice fragments weather to angular vugs in this otherwise extremely resistant horizon which caps a long saddle on the south side of Pathway Canyon. Thin, red and green claystone horizons (Tpkp), which are intercalated with the tuff beds here, are also silicified and unusually well indurated. In comparison to the Socorro Peak area, the interbedded tuffs and claystones here rest directly on gray siltstones (Tpsw) of the lower Popotosa Formation. At the east end of the saddle, the silicified tuffs appear to project under a dark-red laharic breccia (Tsdt), which continues eastward as the basal unit of an east dipping remnant of a rhyodacitic flow (Tsd). In addition to chalky white rhyodacitic pumice fragments like those observed south of Socorro Peak, this laharic breccia contains sparse rhyolitic and andesitic fragments similar to lavas in the Luis Lopez Formation.

North of Nogal Arroyo, at the northern limit of the map area (pl. 1, fig. 29), a thin horizon of light-gray, rhyodacitic lapilli tuff (Tsdt) is exposed. This tuff forms a key marker horizon in a thick section of upper Popotosa claystones (Tpkp). Structure sections in this area (Lemitar Map) indicate that this marker tuff bed occurs about 600 to 750 m above the base of the upper Popotosa Formation. In comparison, a possible equivalent rhyodacitic tuff at Pathway Canyon locally forms the basal unit of the upper Popotosa Formation in that area. Locally, this marker tuff and the surrounding Popotosa claystones are unconformably overlain by mudstones and siltstones of the lower Sierra Ladrones Formation (Tslo). The marker tuff and the same unconformable relationship are repeatedly exposed in the next two gulches north of the map boundary (Lemitar Map). Additional exposures of the marker tuff, as much as 12 m thick, occur west of the J.B. Kelly Ranch about 1 km to the north of the outcrop shown on Plate 1. Here, the lapilli-rich tuff exhibits some cross-bedding and some well-sorted beds, which may be water laid. However, it also contains a few dense blocks of hornblende-bearing rhyodacitic lava, indicating that it was originally an air-fall deposit.

Light gray rhyodacitic pumice lapilli, which characterize this tuff, consistently contain sparse amounts of black hornblende needles. A sample of the hornblende-bearing pumice lapilli from this marker tuff was collected by the author and Kim Manley of the U.S. Geological Survey. Ms. Manley subsequently dated a zircon separate from this pumice using the fission track method; this analysis has yielded an age of 11.9±0.8 m.y., making it the oldest dated rock assigned to the Socorro Peak Rhyolite. I am especially grateful to Kim Manley for providing the age control on this key tuff bed. The source of the rhyodacitic marker tuff is uncertain; vents at "M" Mountain or in the Nogal Canyon area are the most likely candidates.

The rhyodacite unit consists dominantly of medium-gray to light-gray and grayish red to reddish brown lava flows; the reddish colors being generally associated with oxidized tops of the flows or relatively minor hydrothermally altered areas. Most of the rhyodacitic lava outcrops represent gently dipping or horizontal erosional remnants of flows, which are contiguous with intrusive rocks at their source vents. The rhyodacitic lavas commonly form brownish-black weathered cliffs and steep talus mantled slopes at their eroded edges. Columnar jointing, prominently displayed in the cliffs on the southwest side of "6001" Mesa and on the north side of "M" Mountain is fairly common to the rhyodacitic lavas. The cliffs, which generally range from 60 to 150 m in height, provide a rough measure of the thickness of the flows, even though their base is rarely exposed and their tops are fairly flat, eroded surfaces. Flow layering is best developed near the base of the cliffs, where it tends to dip toward the cliff face at moderate angles. This relationship, which is particularly evident in the North Cliff area, suggests that these viscous flows cut "U" shaped channels in the underlying soft sediments much like a glacier does. Christiansen and Lipman (1966, p. 679) have described a rhyolite lava flow in southern Nevada that exhibits similar channel scouring capabilities with respect to soft underlying tuffs.

The geometry of many of the rhyodacitic flows is complex owing to a combination of emplacement on the soft sediments or on irregular bedrock topography. Continued rift faulting since late Miocene time has also locally tilted these flows in various directions and to different degrees. Most of the rhyodacitic lava flows originally had flat tops. Buried Mesa and "6001" Mesa are examples of remnants of these flat-topped flows that have not been tilted. In the Socorro Peak area, the flat top of a rhyodacitic flow (Tsd, pl. 1) is well preserved where it is overlain by a younger lava flow (Tsd_2, pl. 1), now assigned to the quartz latite-rhyolite unit on Figure 29. From a distance, either south or north of Socorro Peak, the planar contact between these flows is highly visible as a bench at the top of the darker colored rhyodacitic cliffs. This contact has an apparent dip of about 5 to 7 degrees to the west.

In contrast, two remnants of flows of the rhyodacite unit, which occur on the down side of the Wood Tunnel fault at the foot of the Socorro Mountain block are tilted to the east (fig. 29). When viewed from the road entering Blue Canyon, the flat top of a cliff-forming rhyodacitic flow is visible in the south wall of the canyon. It appears to be tilted about 20 degrees to the east. Out of sight from the canyon floor, the southern edge of this inclined flow is partly overlapped by a flow of the phenocryst-rich rhyolite unit (fig. 29). The east-tilted rhyodacitic flows at the mountain front are reflections of an anticlinal drag fold, which parallels the range bounding fault zone.

Hydrothermally altered rhyodacitic lavas, which host the silver-bearing barite-quartz veins of the Merritt and Torrance mines, were previously interpreted by Smith (1963, p. 195) as "rootless" landslide blocks. In comparison to the relatively intact outcrop just north of the Merritt mine, the rhyodacitic lavas at the mine area are notably broken and jumbled, as described by Smith. However, the lavas at the Merritt mine exhibit the most common stratigraphic relationship of the rhyodacite unit, namely resting on playa claystones of the upper Popotosa Formation. Only the brecciation here is unusual. Therefore, the outcrops in the mines area are believed to have been largely downdropped to their present position by the action of the Woods Tunnel fault zone. The unusually disrupted block at the Merritt mine may have formed by relatively recent gravity sliding over a minor distance. Presently, the downslope side of this broken outcrop of lavas seems to be essentially unsupported.

Remnants of rhyodacitic lava flows erupted from vents in the Blue Canyon area may have originally ponded in a small basin to form a single mass of lavas with a continuous flat top (fig. 26). This interpretation is supported by the following observations: 1) The relict flat tops of "6001" Mesa and Buried Mesa and the broad crest of Pinnocchio Peak are all approximately at the same elevation (±10 m), and 2) the horizontal "6001" Mesa flow exhibits considerable present day relief (paleotopographic relief) along it's base, which shows that it filled a small asymmetric basin or paleovalley (pl. 2, E-E'). In view of the significant thickness of the remnant flows, the close proximity of the vents, and apparent relative fluidity of the rhyodacitic flows, it seems likely that the individual flows once coalesced into a single flat-topped body of lava.

On the south side of "6001" Mesa, the rhyodacite lavas unconformably overlap a tilted fault block of cauldron-facies Lemitar (?) Tuff and lower Popotosa conglomerates, thereby demonstrating one occurrence of local rift faulting prior to late Miocene time. In the northeastern Magdalena Mountains, Osburn (1978) has observed similar tilted fault blocks unconformably buried by the late Miocene Pound Ranch lavas, which are equivalent to the Socorro Peak Rhyolite.

The Stonewall dome and Smooth dome in the Nogal Canyon area appear to have been only recently exhumed from under a cover of upper Popotosa playa claystones (fig. 26). The Stonewall dome is only partly exhumed and apparently extends outward from its map boundary under a cover of landslide deposits and associated playa claystones. A shallow thermal-gradient borehole on the southwest flank of this outcrop, which was collared in alluvium, intersected 60 m of this rhyodacitic lava and an approximately equal thickness of underlying Popotosa claystones. Near the northwest side of Stonewall dome, numerous well-indurated clastic dikes of red siltstone and mudstone cut the lava outcrops, thus supporting the interpretation of a recently exhumed lava.

The rhyodacitic lava flows of the Nogal Canyon area apparently had smooth and rounded upper surfaces, rather than the flat tops previously described for equivalent flows. The reason for this different occurrence is unknown. The dark-gray flow at Stonewall dome is probably the least silicic lava of the Socorro Peak Rhyolite. This rock has not been chemically analysed; however, it contains the most calcic plagioclase phenocrysts (An_{70}) observed in all of the late Miocene lavas, which may be taken as a general indication of a relatively low silica content (see Table 7).

The hornblende rhyodacite unit of the Socorro Peak Rhyolite is associated with a dozen vent structures (fig. 29), which are exposed at variable levels of erosion and recognized with variable degrees of confidence. Intrusive necks and dikes of rhyodacitic composition, shown on Figure 29 and labeled "Tid" on Plate 1, are generally regarded as vents for the rhyodacitic lavas. These hornblende-and plagioclase-bearing intrusive rocks normally stand as resistant outcrops above the adjacent lower and upper Popotosa Formation sedimentary rocks. Narrow zones of unusually well-indurated, hornfelsed or hydrothermally altered, Popotosa sedimentary rocks are common at these intrusive contacts. Flow foliations in the intrusive rocks typically parallel the steep intrusive contacts, as well illustrated by the largest intrusive neck west of the Merritt mine (pl. 1).

The rhyodacitic intrusive necks and petrologically similar flows in the Socorro Peak silver mining district (Lasky, 1932) are largely hydrothermally altered. The intrusive rocks appear to be the most intensely altered. They are also associated with barite veins similar to the Merritt-Torrance vein. The intrusives are mostly bleached(?) to a pale purplish-gray color, commonly exhibit quartz veinlets and may appear to contain phenocrystic quartz. In thin section, some of these apparent quartz phenocrysts are clearly polycrystalline and conspicuously fill euhedral clasts of evacuated hornblende phenocrysts. Medium-grained plagioclase phenocrysts, exhibiting relict spongy resorbtion textures, typical of the rhyodacite unit (fig. 30A), are partly to completely replaced by potassic clays and feldspar. This observation, verified with thin sections stained for potassium, is the only well established occurrence of potassium metasomatism in rocks younger than Oligocene within the area of the Socorro Peak volcanic center.

Lavas in the mining district are not as intensely altered as their equivalent intrusive rocks. The lavas are mostly reddish brown and characterized by abundant hornblende needles replaced by soft, black iron oxides and predominantly fresh, or slightly argillized, plagioclase phenocrysts. However, a few outcrops in the mines area do contain "pseudophenocrysts" of quartz and metasomatic feldspar. The hydrothermal alteration of these originally rhyodacitic rocks in the Socorro Peak mining district explains why they were previously classified as "banded trachyte" (Lasky, 1932; Smith, 1963; Burton, 1971) and why they were not correlated with their fresh equivalent lavas exposed near Socorro Peak.

The best exposure of a vent related to the rhyodacite unit is found on the north side of "M" Mountain. This vent is apparently continuous into the flow that extends northward to North Cliff. Although the inferred sedimentary wall rocks are not exposed at this vent, the northern limit of the outcrop seems to be clearly intrusive, since it exhibits a well-developed flow layering that parallels and dips steeply into the mountain side. Columnar jointing is very well developed here. The columns are nearly vertical in the interior (upper) portion of the vent but fan outward rapidly as they approach the steeply oriented perimeter of the outcrop. This columnar joint pattern is quite similar to the common textbook example of an intrusive neck -- the

Devils Postpile in northern California.

Other vents for lavas of the rhyodacite unit have been inferred from steeply dipping slabby fractures that represent flow foliations. Sparse foliation data, which suggest the presence of a vent under Buried Mesa, are supported by indirect evidence of an intrusive root to this outcrop. Such a root would seem to be a necessity, since Buried Mesa is acting like a bulwark, retarding the passage of rhyolitic landslide blocks (Qlr, pl. 1) downslope to the southeast.

Specimens of the rhyodacite unit of the Socorro Peak Rhyolite generally contain about 5 to 15 percent phenocrysts. However, at first glance these rocks appear to be only sparsely porphyritic, since typically only 2 to 4 percent of the rock consists of conspicuous medium-grained phenocrysts about 1 to 4 mm in length. Most of the phenocryst population is fine-grained (about 0.7 to 0.2 mm) and relatively inconspicuous when observed in handspecimen.

Plagioclase generally constitutes about two-thirds of the total phenocrysts and hornblende forms most of the remainder. Fine-grained (0.4 to 0.1 mm) black Fe-Ti oxides may form 1 to 2 percent of these relatively dark-colored rocks. Biotite occurs only in trace amounts.

In thin section, the bimodal distribution of plagioclase phenocrysts is readily apparent because of a relative paucity of plagioclase crystals between about 1.5 and 0.5 mm (fig. 30A). Blocky, medium-grained, plagioclase phenocrysts, which represent an early stage of intratelluric crystallization, are characterized by spongy resorbed cores of labradorite (An_{70} to An_{60}) and narrow, strongly zoned rims of labradorite to andesine (An_{60} to An_{40}). These resorbed plagioclase phenocrysts commonly have a yellowish or pinkish granulated appearance in hand specimen because of abundant inclusions of glassy groundmass. The fine-grained late-stage plagioclase phenocrysts typically are not resorbed and exhibit the same strong normal zoning seen in the rim material of the early-stage phenocrysts.

Hornblende phenocrysts also show a bimodal occurrence. Early hornblende phenocrysts are dominantly euhedral six-sided prisms about 1 to 3 mm long. Some of the largest hornblende phenocrysts are poikolitic and contain abundant Fe-Ti oxide inclusions. Pale yellowish-green to light-brown pleochroism and measured optic axial angles ($2V_x$) of 62 and 65 degrees indicates that the early hornblende phenocrysts are of the common variety found in intermediate members of the calc-alkaline series of igneous rocks (Deer, Howie, and Zussman, 1966, p. 167). Late-stage hornblende phenocrysts occur mostly as slender, needle-like crystals and less commonly as subequant prisms. Most thin sections of the rhyodacite unit contain traces of biotite, which appears to have formed by a magmatic reaction at the expense of hornblende.

The rhyodacites are dominantly lithoidal and in thin section most of them exhibit a devitrified murky cryptocrystalline groundmass, which commonly contains sparse trachytically aligned feldspar and hornblende microlites along with disseminated opaques. In devitrified specimens, hornblende needles are usually replaced by opaque Fe-oxides; this is probably a manifestation of deuteric alteration.

A single analysis of the plagioclase and hornblende-bearing lava from "6001" Mesa (Table 7, no. 77-5-1) indicates that it is a rhyodacite. Based on a inferred water content of about 3 percent, the estimated SiO_2 content of this sample as an anhydrous rock is about 62 percent. Gordon (1910, p. 240) reported a partial analysis of a sample from an unknown location on Pinnocchio Peak (Pyramid Peak), which yielded an SiO_2 content of 65.6 percent. Compositions of plagioclase phenocrysts in the Stonewall dome lava, as previously stated, suggest this flow is slightly lower in SiO_2 content than the "6001" Mesa flow. Therefore, most of the rocks of the Socorro Peak Rhyolite that are characterized by plagioclase and hornblende phenocrysts are believed to be rhyodacites containing about 60 to 65 percent SiO_2.

The oldest rock of the rhyodacite unit is the 11.9 m.y-. old marker tuff previously described in this section. Hornblende separates from the rhyodacitic lavas at Stonewall dome and "6001" Mesa have yielded K-Ar ages of 10.3±1.5 m.y. and 6.0±0.6 m.y., respectively. A subsequent K-Ar analysis performed at a different laboratory, gave a 10.3 m.y. age for a whole-rock split of the "6001" Mesa sample. Stratigraphic relationships of the rhyodacite unit demonstrate that it is older than both the phenocryst-rich and phenocryst-poor rhyolite units of the Socorro Peak Rhyolite, which have yielded K-Ar ages of 10.5 to 9.0 m.y., and 7.4 m.y., respectively. The above relationships strongly suggest that the 6.0 m.y. hornblende date for "6001" Mesa is anomalously young and therefore probably erroneous. Rocks now assigned to the quartz latite-rhyolite unit have been dated at 11.8 m.y. and 11.5 m.y. (Appendix B), and at 10.7 m.y. (Burke and others, 1963). Field and petrographic observations of the rhyodacite unit and the quartz latite-rhyolite unit suggest that they are penecontemporaneous derivatives from the same magma body and that the more silicic lavas are only slightly younger. Therefore, both the rhyodacite unit and the quartz latite-rhyolite unit are considered to be about 11.1 m.y. old, which is the numerical average of their six radiometric dates ranging from 11.9 to

10.3 m.y. The anomalously young 6.0 m.y. date is excluded from this calculation.

Biotite-hornblende, quartz latite-rhyolite unit

The biotite-hornblende quartz latite-rhyolite unit of the Socorro Peak Rhyolite consists dominantly of pale-red, light-gray and grayish-red moderately porphyritic lavas, and minor bedded pyroclastic rocks. They are generally characterized by phenocrysts of plagioclase, biotite and hornblende. Exposures assigned to the quartz-latite-rhyolite unit occur at three widespread localities in the study area: at Strawberry Peak, at and west of Socorro Peak, and in the vicinity of the Tower mine. These isolated outcrops generally represent erosional remnants of individual domal flows erupted from local vents.

Strawberry Peak is a partially preserved erosional remnant of a viscous lava dome; it consists of an intrusive neck on the south side and a flow remnant on the north side (fig. 29, pl. 1). The original dome was apparently erupted athwart a structurally controlled facies boundary in the underlying upper Popotosa sedimentary units (fig. 26). Thus, the northwest side of the domal flow is now largely preserved where it rests on conglomeratic sandstones (Tpkf). However, the original south and east sides of the dome have broken away from the vent area to form large landslide blocks because here the lava structure originally rested on incompetent playa muds (Tpkp). The largest of these landslide blocks, which forms a prominent ridge east of Strawberry Peak (pl. 1), exhibits foliation patterns and variations in lithology very similar to the parent outcrop of Strawberry Peak.

Observations on the north flank of Strawberry Peak suggest that this lava dome was erupted in at least two stages consisting of an early rhyodacitic magma pulse and a later quartz latitic to rhyolitic magma pulse. The base of the flow is not exposed here; however, the lowest 30-50 m of the outcrop consists of medium-gray, columnar-jointed lava with 5-10 percent phenocrysts of plagioclase and hornblende, suggesting it is of rhyodacitic composition. Where locally well exposed near the top of a cliff, the columnar jointed lava appears to grade into grayish-red, blocky-jointed lava containing the typical phenocryst assemblage of the quartz latite-rhyolite unit: plagioclase, biotite and hornblende. The blocky red lava is about 140 m thick and forms the bulk of Strawberry Peak. The boundary between the two lava types appears to be gradational over a few meters and is not a typical contact between separate lava flows. No autobrecciated zone or pyroclastic interval, which typically occur at flow boundaries, has been observed in association with this contact. A similar transitional contact is also present on the north side of the big landslide block east of Strawberry Peak.

In the vicinity of Socorro Peak, a biotite-bearing quartz latite-rhyolite lava flow (or flows?) (Tsd_2, pl.1) locally rests on a hornblende-bearing rhyodacitic lava flow (Tsd_1, pl. 1). The quartz latite-rhyolite lavas were erupted from a composite vent at "M" Mountain and a nearby vent at Radar Peak. In contrast to the Strawberry Peak area, the rhyodacitic and quartz latitic lavas here clearly represent separate flows. The base of the younger lava flow (Tsd_2) is conspicuously defined by discontinuous exposures of autobrecciated lava (pl. 1), bedded tuffs (Tsdt), and small patches of red, silty mudstone (Tpkp). These mudstones are locally baked to a pottery-like material. The basal zone of the younger biotitic lava is light-gray in color and contrasts with a reddish colored zone at the top of the underlying flow, which most likely represents an oxidized or weathered zone formed prior to eruption of the younger flow.

West of Socorro Peak and Radar Peak, the biotite-bearing quartz latitic lava appears to have flowed across the edge of the older rhyodacitic flow and onto playa mudstones (Tpkp). Although the areas of this apparent off-lap are now poorly exposed, the outcrop relationships at the eroded edges of these flows suggests that the base of the quartz latitic flow drops in elevation (toward the west) by as much as 90 m where it crossed the edge of the older flow. This observation could be explained by scouring of the soft mudstones by the viscous quartz latitic lava, by subsidence of the heavy lava into poorly consolidated mudstones, or by younger faulting along this trend. The latter alternative, as shown on the geologic map (pl. 1), seems somewhat fortuitous and is now less favored.

Discontinuous exposures of grayish-red lavas in the Tower mine area, which contain phenocrysts of plagioclase, biotite and hornblende, have been for the sake of convenience assigned here to the quartz latite-rhyolite unit of the Socorro Peak Rhyolite. These faulted remnants represent the thin edge of a lava flow (fig. 26), locally not more than about 45 m thick. They are stratigraphically and petrologically similar to the upper Pound Ranch lavas of Osburn (1978, p. 80) and therefore considered correlative. The probable source vent for these lavas, as described by Osburn (1978, p. 81), is located about 4 km west of the Tower mine. In a manner similar to the Pound Ranch area, the biotitic lavas (Tsd, pl. 1) in the Tower mine area rest variably on the Oligocene tuff of South Canyon (Tsc) and on early Miocene red mudflow and conglomerate deposits (Tpsd, pl. 1) of the lower Popotosa Formation. These stratigraphic relationships locally define block-faulting and erosion prior to eruption of the quartz latite-rhyolite lavas in late Miocene time (fig. 26). Later movement of the same fault blocks has displaced this flow since its emplacement. The absence of upper Popotosa claystones (Tpkp) under these lavas in the Pound Ranch and Tower mine area is also significant. This relationship is interpreted to mark the southern margin of the upper Popotosa basin. The presence of a thin interval of playa muds (Tpkp) and the interbedded basalt of Bear Canyon (Tpkb), which overlie the quartz latite-rhyolite lavas west of the Tower mine, indicates that the basin margin locally shifted laterally with time and had small embayments related to local north-trending block faults (fig. 26).

Lavas of the quartz latite-rhyolite unit of the Socorro Peak Rhyolite generally weather to grayish-red or light brownish gray resistant outcrops; cliffs and ledgy outcrops are common at the eroded margins of the thicker flows. Flow layering is generally well developed and defined by lighter-colored, slightly vesiculated bands along the shear surfaces. Pumice-like ragged vesicles are fairly typical of the upper portions of these flows. Intrafolial gash fractures, which strike approximately perpendicular to the local flow direction, are locally well developed in the Strawberry Peak flow. In exposures at right angles to the flow layering, the gash fractures are visible in multiple layers and occur as en echelon lenticular openings between foliation surfaces. Minor manganese and hematite staining and some calcite veins are locally associated with vent structures at "M" Mountain and Strawberry Peak.

Specimens of the quartz latite-rhyolite unit generally contain 5 to 20 percent phenocrysts of plagioclase, biotite and hornblende. Plagioclase normally forms about two-thirds of the total phenocrysts, biotite and hornblende form the remainder. Either biotite or hornblende may be the dominant mafic phenocryst; usually they are subequal in volume and ratios greater than 2:1 in either direction are not common. Locally mafic phenocrysts may be quite sparse, particularly in the middle or upper portion of these flows, thus making them difficult to distinguish from the similar plagioclase-rich rhyodacites.

Petrographically, the biotite-hornblende quartz latite-rhyolite unit (Table 7, no. 76-4-1b, 76-6-1) is very similar to the hornblende rhyodacite unit. In thin section, specimens of the quartz latite-rhyolite unit show the same bimodal phenocryst population of sparse medium-grained and moderately abundant fine-grained crystals observed in the rhyodacite unit. The same pattern of spongy resorbed cores of early plagioclase phenocrysts associated with strongly zoned rims and strongly zoned late-stage plagioclase phenocrysts is also present in the quartz latite-rhyolite unit. However, in the relatively biotite-rich rocks, the plagioclases are generally less calcic; cores of early phenocrysts range from An_{60} to An_{50} and outer zones of late phenocrysts approach An_{30}. A small percentage of early plagioclase phenocrysts exhibit patchy reverse zoning from andesine to labradorite (fig. 30B), which is also observed in the rhyodacite unit. Some specimens of the quartz latite-rhyolite unit contain sparse, blocky plagioclase phenocrysts that are coarse-grained and as much as 8 mm long. These textural similarities of plagioclase phenocrysts and variations in composition between lavas of the quartz latite-rhyolite unit and the rhyodacite unit strongly suggests that these units represent differentiates formed in the same magma system This would explain the observed close association of these units in space and time.

Biotite generally occurs as thin-euhedral plates (1-3 mm) and is mostly pleochroic from pale yellowish brown to reddish brown and less commonly pleochroic from pale yellow-

Figure 30. Photomicrographs illustrating spongy resorbtion textures and zoning of plagioclase phenocrysts common in both the rhyodacite unit and the quartz latite unit of the Socorro Peak Rhyolite (Tsd, pl. 1). A, Hornblende rhyodacite from "6001" Mesa (sample no. 77-5-1, Table 7) containing moderately abundant phenocrysts of plagioclase (light-gray), hornblende (dark-gray), and minor small Fe-Ti oxides (black) in a cryptocrystalline to glassy groundmass (medium-gray). Several features common to both the rhyodacite unit and quartz latite unit are exhibited here: 1) a bimodal population of large (1-3 mm) early-stage phenocrysts and small (0.2-0.5 mm) late-stage phenocrysts, 2) spongy strongly resorbed calcic cores (An_{65}) of early-stage plagioclase phenocrysts, and 3) unresorbed rims on early plagioclase phenocrysts and small late plagioclase phenocrysts that both exhibit strong normal zoning (An_{60-40}). Most labradoritic cores of early plagioclase phenocrysts are unzoned or show slight normal zonation. Plane polarized light. B, Reverse-zoned, early stage, plagioclase phenocryst in quartz latitic lava near Radar Peak (approximately equivalent to sample no. 76-4-1b, table 7). Numbers indicate optically determined percent anorthite in various zones of the crystal (Rittman zone method, p. 27). Note the broad step-like reverse zonation of the core, spongy resorbed calcic zone, and narrow normally zoned rim. In the rhyodacite to quartz latite lavas, about 10 to 25 percent of the early-stage plagioclase phenocrysts exhibit similar reverse zoning. Crossed nicols.

ish green to dark brown. Some biotites appear to have formed by magmatic reaction at the expense of hornblende. Hornblendes are present as both euhedral, medium-grained prisms (2-4 mm) and smaller, needle-like cyrstals. Most hornblende is pleochroic from pale yellowish green to light brown. In one thin section from Socorro Peak, both biotite and hornblende are similar in color (yellowish brown to dark reddish brown) and pleochroism. In this case, they may be distinguished by cleavage and crystal form. Both biotite and hornblende are often partially altered to opaque Fe-oxides; needle-like hornblendes seem particularly susceptible to this alteration, which is probably deuteric. Fine, equant grains of Fe-Ti oxides usually form less than one-half percent of these rocks.

Fine-grained phenocrysts of plagioclase, biotite and hornblende consistently grade into groundmass microphenocrysts of the same minerals. Microphenocrysts and phenocrysts are usually flow aligned within a cloudy cryptocrystalline or partially spherulitic devitrified groundmass. Microphenocrysts may form 5 to 30 percent of the groundmass.

Two analyses of the plagioclase-biotite-hornblende lavas from the study area have nearly equivalent uncorrected SiO_2 contents of about 68 percent suggesting that they are quartz latites (Table 7, no. 76-4-1b, 76-6-1). However, when the estimated water content is used to recalculate these samples as anhydrous rocks the results span the boundary from quartz latite to low-silica rhyolite. Osburn (1978, p. 81) reported a partial, uncorrected analysis of the petrographically equivalent upper Pound Ranch lava which yielded a SiO_2 content of 72.3 percent, clearly placing it in the rhyolite field. Therefore, most rocks mapped as the quartz latite-rhyolite unit (fig. 29) are believed to contain about 68 to 72 percent SiO_2. However, should additional analyses be obtained from the Strawberry Peak flow, a greater range in silica content may be expected.

Burke and others (1963, p. 229) reported a whole rock K-Ar date for their "trachy-andesite of Socorro Peak" of 10.7±1.5 m.y. This sample was taken from Radar Peak and is here reassigned to the quartz-latite rhyolite unit of the Socorro Peak Rhyolite. A subsequent sample from the same location has been dated by C. T. Smith (unpub. data reported by C. Burton, 1971) at 11.5 m.y. Biotite from the big slide block east of Strawberry Peak has yielded a K-Ar age of 11.8±0.5 m.y. The preferred radiometric age of both the quartz latite-rhyolite unit and the rhyodacite unit is 11.1 m.y., which is the average of six dates from these two units. The basis of this calculation has been described in the previous section.

Phenocryst-rich rhyolite unit

The phenocryst-rich rhyolite unit of the Socorro Peak Rhyolite is predominantly comprised of light gray, light brownish-gray and pinkish-gray lavas, which are characterized by abundant medium-grained phenocrysts of sanidine and plagioclase, and lesser amounts of quartz and biotite. Sparse phenocrysts of hornblende also occur sporadically in these lavas. Bedded lapilli tuffs and one isolated intrusive neck, with similar phenocryst assemblages, make up the minor remainder of this unit. The phenocryst-rich rhyolite appears to be the most voluminous of the late Miocene lavas and comprises nearly half the volume of the Socorro Peak Rhyolite.

Exposures of the phenocryst-rich rhyolite unit occur in two principal areas (fig. 29), each of which is associated with a local vent complex. The largest outcrop area occurs in the western Socorro Mountains. This broad area of lava outcrops is formed by contiguous lava flows and domes erupted from four widely spaced vents. The second principal outcrop area occurs at the Signal Flag vent complex just west of Cook Spring.

Stratigraphic relationships of the phenocryst-rich rhyolite unit are variable. In the western Socorro Mountains, these phenocryst-rich lavas apparently rest on a thick interval of playa claystones (Tpkp) of the lower Popotosa Formation. Drill holes in the northern reaches of Socorro Canyon bottomed in playa deposits at a depth of 150 m (C.E. Chapin, 1978, unpub. drill logs). West of Cook Spring, lavas and tuffs of the phenocryst-rich rhyolite variably overlie eroded fault blocks of the lower Popotosa conglomerates (Tpsd) and upper Popotosa claystones (Tpkp) (pl. 2, D-D'). Just northwest of Cook Spring, the phenocryst-rich rhyolite lavas overlap the edge of an east-tilted flow of the rhyodacite unit (Tsd). South of Cook Spring, a small bulldozer cut clearly exposes the top of the same phenocryst-rich lava flow; here it is overlain by playa mudstones (Tpkp) and, in turn, by bedded tuffs of the phenocryst-poor rhyolite unit (Tsrt). This key area around Cook Spring provides most of the stratigraphic control on the petrologic units of Socorro Peak Rhyolite.

The rounded peaks of the western Socorro Mountains -- Tripod Peak, "6633" Peak, Jejenes Hill, and an unnamed peak to the north of Jejenes Hill -- all exhibit foliation patterns consistent with their interpretation as domal accumulations of pasty lavas roughly centered over pipe-like vent structures. The surrounding rolling topography at lower elevations, such as in the area between Hidden Valley and Domingo Spring, generally exhibits irregular foliation patterns. The low, rolling hills and valleys are considered to represent several individual lava flows, which have coalsced into a single mass. The geometry of these individual flows is uncertain; small outcrops of autobrecciated lava mapped north and northeast of Jejenes Hill (pl. 1) may locally mark the sutured boundaries of at least two separate flows.

The eroded edges of these lava flows are generally marked by steep slopes and cliffs, which grade into talus cones (Qca) and landslide blocks (Qlr) further downslope. For practical purposes, the bottom of these cliffs may be considered as essentially equivalent to the base of these flows. From observation of the general westerly decrease in elevation of the approximate base of these flows, an overall apparent dip of 3 to 4 degrees to the west-northwest has been calculated. Since the phenocryst-rich rhyolite flows here are gently dipping, the topography on the top of these lavas may then be used as direct indication of variations in their thickness. Local elevation variations indicate that the domal structures (rounded peaks) are underlain by about 120-180 m of lava, whereas the outer eroded edges of the lava flows range from 45-60 m in thickness. The significance of a small hill (elevation 6345) northwest of Domingo Spring is uncertain. This mound of lavas, about 100 m thick, could be a small endogenous dome.

In the western Socorro Mountains, the phenocryst rich rhyolite lavas generally weather to a medium-light gray or a light-brownish-gray color. Ledgy outcrops are common above the surrounding cliffs. Large areas mantled by a colluvial soil (pl. 1) of subangular lava blocks in a reddish-brown, muddy-sand matrix may represent areas incompletely stripped of their inferred original cover of playa muds.

Flow layering is most prominently developed in the outer reaches and lower portions of the flows, which also tend to be less porphyritic. Near the vent areas, the lavas become the most phenocryst rich (30-35 percent) and tend to have a dense, massive appearance. Here, the lavas often weather to rounded, granitic-looking outcrops. Nearly all of the phenocryst-rich rhyolites are lithoidal; spherulitic devitrification is common. However, one large area of gray and black vitrophyric breccia has been observed southwest of Jejenes Hill (pl. 1); this outcrop is believed to represent a basal flow breccia.

Lenticular and irregular-shaped xenoliths of red hornfelsed mudstone are common in the phenocryst-rich rhyolite lavas near Domingo Spring. Most of the inclusions are a few centimeters long but a few are more than one meter long. The mudstone xenoliths are typically concordant with flow foliation. Some well-indurated red muds similar in appearance to the xenoliths occur in conjugate fractures and may be clastic dikes.

Outcrops of the phenocryst-rich rhyolite at the Signal Flag vent complex west of Cook Spring have a variable topographic expression, which may be related to differential uplift and downwarping along the range bounding fault zone. Immediately west of Cook Spring, the phenocryst-rich rhyolite outcrops clearly have the geometry of an east-tilted lava flow; here they are overlain by similarly dipping conglomeratic sandstones (Tpkr) and mudstones (Tpkp) of the upper Popotosa Formation. Further upslope to the west, the same phenocryst-rich rhyolites form three steep-sided knobs representing eroded plug domes. The lava outcrops immediately surrounding these knobs appear to be remnants of nearly horizontal flows. The bases of the two knobs south of the Signal Flag are evidently eroded to a level that is at, or very near to, the late Miocene surface onto which these lavas were extruded. This is indicated by the flow layering, which locally fans outwards to subhorizontal attitudes at the base of the knobs.

Flow foliations in this area are primarily defined by varying degrees of vesiculation of the lavas along shear planes. Some vesicular layers are quite wide and may represent pods of magma with a slightly greater orignal volatile content. This relationship is well exposed in road cuts leading to the Signal Flag area. Here, dense flow bands and vesicular (pumiceous) flow bands, a meter or more in thickness, form nearly vertical alternating layers that are respectively light gray and white in color. The pumiceous bands locally contain small lenses of dense material and the dense bands may contain thin vesiculated lenses. On weathered surfaces the punky vesicular flow layers form recesses, which often tend to give the outcrops a ribbed appearance. This local phenomena of differential weathering of flow layers in the phenocryst-rich rhyolite has combined (in a chance circumstance) with the relatively recent development of a large landslide scarp to produce an excellent exposure of the large flow fold shown here in figure 31.

Two small exposures of white air-fall tuffs (Tsrt), about 5-10 m thick, are locally exposed at the base of the phenocryst-rich lava flows in the area southwest of the Signal Flag. These bedded tuffs contain abundant white pumice lapilli, which are characterized by sparse but conspicuous black biotite and glassy quartz phenocrysts in addition to less obvious feldspar phenocrysts. Similar biotite and quartz-bearing tuff beds have been observed at several isolated outcrops where they are interbedded in the

Figure 31. Large recumbent flow fold in the phenocryst-rich rhyolite unit of the Socorro Peak Rhyolite. The flow structure is largely evident through differential weathering of resistant (dense) flow layers and non-resistant (vesiculated) flow layers. This large flow fold may be part of an endogenous dome (vent structure) centered to the left of the photograph. A vent structure is recognized to the left of this photograph; however, its relationship to the above outcrop is obscured by minor(?) post-lava faulting and talus deposits. This cliff face is below and to the east of the Signal Flag, both of which are visible when entering Blue Canyon. The visible portion of the cliff is about 45 m high; note bulldozer in foreground for relative scale.

upper Popotosa playa deposits. The locations of these tuff beds, tentatively assigned to the phenocryst-rich rhyolite unit, are shown on figure 29.

At the northeast corner of "6001" Mesa is a small, but prominent, isolated knob of phenocryst-rich rhyolite. This knob is clearly exposed at a deeper erosion level than the plug domes to the east of here. It exhibits a sharp intrusive contact with the rhyolite of Blue Canyon and is therefore shown on the geologic map as an intrusive neck (Tir, pl. 1). Both this small intrusive neck and the plug dome due west of Socorro Spring are hydrothermally altered. Hematite has stained these intrusives an anomalous reddish-orange color and plagioclase phenocrysts in them are replaced by chalky clay. The altered plagioclase phenocrysts weather out to give the outcrops a vuggy or pseudovesicular appearance; this is particularly noticeable at the smaller intrusive neck. Even where altered, the late Miocene phenocryst-rich rhyolites are readily distinguished from nearby outcrops of the Oligocene rhyolite of Blue Canyon by the presence of black biotite phenocrysts; also quartz phenocrysts are larger and more abundant in the younger rhyolite.

Handspecimens of lavas from the phenocryst-rich rhyolite normally contain about 15-35 percent phenocrysts.

Both sanidine and plagioclase generally occur as fresh, clear tabular crystals, making them quite difficult to differentiate in handspecimen. The feldspars typically make up more than two-thirds of the total phenocrysts; the remainder consists of subhedral to rounded quartz and black biotite flakes. Quartz tends to be larger and more abundant at the Signal Flag complex, where it forms as much as 5 percent of the rocks. Specimens from the outer edges or bases of these flows (Table 7, no. 76-6-3) generally contain smaller phenocrysts and less total phenocrysts than the bulk of these flows. Pumice lapilli in related tuffs may appear to be only moderately porphyritic because of their relatively low density.

In thin section, homogeneous sanidine phenocrysts are clearly more abundant or subequal in volume to finely albite twinned and zoned plagioclase phenocrysts. Sanidine usually occurs as stubby, euhedral, single crystals and carlsbad twins are relatively rare. Tabular and subequant plagioclase phenocrysts commonly exhibit oscilliatory zoning, which is superimposed on crystals that overall are broadly and normally zoned. The total range of plagioclase composition in one specimen may vary from calcic to sodic andesine (An_{50}-An_{30}); however, individual crystals exhibit smaller variations usually of less than 10 percent anorthite.

Weakly zoned sodic andesine is typical of the most crystal-rich samples. Spongy resorbtion textures are rare or absent in plagioclases of the phenocryst-rich rhyolite unit.

Rounded and embayed quartz phenocrysts are common and generally comprise from 2 to 5 percent of the phenocryst-rich rhyolites. Both reddish-brown and greenish, pleochroic biotite occur, but the former is more common. Biotite is slightly less abundant than quartz in most samples and forms about 1-3 percent of the rock. Hornblende, sphene, zircon, and Fe-Ti oxides are commonly observed in trace amounts as microphenocrysts in a cryptocrystalline or spherulitically devitrified groundmass. Perlitic structure, indicative of a significant water content, is commonly well developed in the vitrophyric samples.

Analyses of rocks assigned to the phenocryst-rich rhyolite unit (Table 7, no. 76-6-3, 76-6-2a, and 76-6-2b) have indicated a range from low-silica to high-silica rhyolite (70 to 76 percent SiO_2). Sample no. 76-6-2a, which contains about 75 percent SiO_2, is probably the most representative of the bulk of the phenocryst-rich lavas.

A sample of the phenocryst-rich rhyolite from bulldozer cuts at the Signal Flag has yielded a biotite K-Ar age of 10.5±0.4 m.y. (Appendix B). Similar phenocryst-rich lavas in the western Socorro Mountains may be somewhat younger.

A biotite separate from a sample collected at the large vitrophyric slideblock (Qlr, pl. 1) of the old railroad quarry in Socorro Canyon (fig. 3) has yielded a K-Ar age of 9.0±0.4 m.y. (Appendix B). This slide block was most likely derived from the outer margin of the "6633" Peak domal flow (fig. 29). The 10.5 to 9.0 m.y. age range for the phenocryst-rich rhyolite is in general accordance with the assigned stratigraphic position of this unit, which is above the rhyodacite unit and the quartz latite-rhyolite unit. The latter units have given seemingly reliable ages ranging from 11.9 to 10.3 m.y. Some of the phenocryst-rich rhyolite lavas may have been erupted contemporaneously with the less silicic flows; no stratigraphic evidence is available to refute this possibility.

Phenocryst-poor rhyolite unit

Rocks of the phenocryst-poor rhyolite unit of the Socorro Peak Rhyolite are characterized by less than two percent phenocrysts consisting of plagioclase, sanidine, quartz and biotite. As defined here, this unit is synonymous with the 7.4-m.y.-old (Appendix B) rhyolite vitrophyre dome at the Grefco perlite mine and a surrounding cogenetic pyroclastic horizon (fig. 29). The salient features of this small body (about 0.05 km^3) of high-silica rhyolite lava

(Table 7, no. 77-5-4) were well-described by Weber (1963b). Some new interpretations of the stratigraphic relationships of this economically important lava dome, and the new K-Ar date mentioned above, may be considered as an update on Weber's fundamental report. The Grefco perlite dome, which is the youngest dated rhyolite in central New Mexico, provides an approximate upper limit on playa (bolson) sedimentation of the Popotosa Formation and a rough lower limit on the initiation of fluvial (ancestral Rio Grande) sedimentation of the Sierra Ladrones Formation.

The general stratigraphic and structural relationships of the Grefco perlite dome (Tsr) and a related air-fall tuff horizon (Tsrt) are illustrated in cross section F-F' of Plate 2. The area around the Grefco mine provides the most conspicuous example of intertonguing of the Socorro Peak Rhyolite and the upper Popotosa Formation because this lava dome is only partially exhumed from under a cover of younger Popotosa sedimentary strata.

The contact relationships between the lava dome (Tsr) and its cogenetic tuff bed (Tsrt) are largely concealed by younger sedimentary facies of the Popotosa Formation (Tpkp, Tpke, Tpkr), which bury the steep flanks of the dome. However, at a small exposure on the south side of the dome, the brecciated base of the glassy flow appears to have banked onto a north-facing, moderately dipping slope at the top of the massive white tuff. The same lava-tuff contact on the north side of the dome has been largely obscured by debris from mining operations.

The tuff horizon associated with the Grefco dome forms numerous discontinuous outcrops between Socorro Canyon and Socorro Spring. The general distribution of these outcrops is shown on figure 29. This horizon apparently thins away from the vent area: about 1 km west of the Grefco mine it is about 60 m thick, about 1.5 km north of the mine at Socorro Spring, it is less than 25 m thick. It generally crops out as creamy white ledges and slopes consisting of moderately well-indurated lapilli tuff. Pumice lapilli and the tuffaceous matrix are partially altered to white clay in most outcrops. Angular fragments of light-gray rhyolite vitrophyre, petrologically equivalent to the phenocryst-poor Grefco dome lavas, are common and distinguish this tuff from other tuff units (fig. 29) assigned to the Socorro Peak Rhyolite. Other types of lithic fragments include a pale-red lithoidal equivalent of the Grefco lava, non-indurated red claystones, and sparse accidental fragments of older volcanic rocks. Some clasts of red claystone, as much as 1.5 m long, have been observed in outcrops within one kilometer to the north and west of the Grefco mine (fig. 32).

The size of these soft plastic mud fragments, clearly representing air-fall debris, attests to the proximity of the source vent.

Lithic fragments similar to the hornblende rhyodacite unit of the Socorro Peak Rhyolite have rarely been observed in the long outcrop of bedded tuff southwest of the Grefco mine. This observation indicates that the large body of intrusive rhyodacite (Tid) in this area may be older than this tuff. The contact between these units in the arroyo floor (pl. 1) is not well exposed; however, the tuff does not appear to be hornfelsed or unusually altered here. If these observations are correct, then there seems to be a local erosional unconformity between the phenocryst-poor rhyolite unit and the hornblende rhyodacite unit of the Socorro Peak Rhyolite. South of Sedillo Spring, small exposures of the tuff related to the Grefco dome, which are surrounded by alluvium, may have lapped unconformably onto an eroded block of lower Popotosa conglomerates (Tpsd). In marked contrast, the tuff exposures west of the Grefco mine and at Socorro Spring are conformably interbedded within the upper Popotosa strata.

The tuff related to the Grefco dome appears to grade laterally from a non-sorted, coarse tuff breccia at outcrops nearest to the mine, to typical well-bedded and sorted air-fall tuffs further away from the mine. As mapped at Socorro Spring, the upper half of this unit consists of parallel bedded and cross bedded tuffaceous sandstones that are well-indurated with silica cement. This upper portion of the tuff horizon may be equivalent to the vitrophyric sandstone facies (Tpkr) of the upper Popotosa Formation, which consistently overlies the air-fall deposits around the Grefco mine. In comparison to the sandstones near the mine, which are gray, granular, and glassy, the white sandstone at Socorro Spring contains mostly shard-like sand grains partially altered to white clay.

The general features of the Grefco perlite body, as previously described by Weber (1963b), and those shown on the geologic map (pl. 1) both lead to the same conclusion, namely that it is a silicic lava dome. However, some aspects of the geometry of the dome are not entirely clear, particularly along the west side where most of the contact relationships are masked by colluvium. The north and south margins of the dome are overlapped by vitric conglomeratic sandstones (Tpkr); these sandstones are described in an earlier section of this report as a local facies of the upper Popotosa Formation derived from erosion of the glassy lava dome. Most of the eastern boundary of the dome is formed by high-angle normal faults which juxtapose the lava against younger sedimentary rocks of the upper Popotosa and Sierra Ladrones Formations. The spatial relationships of the lava outcrops along the poorly exposed western flank of the dome (pl. 1) indicate that this is a steeply dipping contact. Weber (1963b, p. 144) previously described the perlite dome as "occupying the position of a horst block" and interpreted the western contact as a down-to-the-west fault. Even though exposures are poor here, the map relationships of the surrounding Popotosa strata (pl. 1) do not indicate any significant displacement along this western contact. The vitric sandstone (Tpkr) that flanks the dome on its north and south end also occurs at the same or higher elevations west of the dome. Therefore, this western contact is here interpreted as the original steep flank of the dome, somewhat modified by erosion and later buried by upper Popotosa sedimentary units.

The geometry of the base of the Grefco dome is its least certain aspect. As an alternative to the flat-bottomed flow shown in cross section (pl. 2, F-F'), the base of the dome may be more saucer-shaped, and the domal flow may be largely restricted to the inward-facing slopes of an underlying tuff cone. The single observation of an inward dipping base on the flow at its south end supports the latter interpretation.

The main body of the dome is roughly rectangular in plan view, measuring about 850 m from north to south and about 550 m from east to west. Topographic relief across the outcrop is approximately 115 m; this is considered to be minimum thickness for the domal flow. Including inferred underlying intrusions, the present volume of the Grefco dome is estimated to be about 0.05 km^3.

The rhyolite vitrophyre lava of the Grefco dome is mostly light gray on fresh surfaces and weathers to resistant blocky outcrops that are medium gray or light brownish gray. Flow layering is conspicuous throughout the dome and generally appears to be formed by the development of lineated and slightly pumiceous bands along shear planes. Concentric patterns of steeply dipping foliations suggest the presence of vent structures under the northwest and south east corners of the lava dome; however, the latter structure is less constrained and is considered less certain.

As pointed out by Weber (1963, p. 144), most of this rhyolite vitrophyre lava does not display an onion-skin perlitic structure, rather it normally appears to be a slightly pumiceous glass containing less than 2 percent phenocrysts. Thus the term perlite is used in an industrial sense, because this water-rich glass can be processed in furnaces to form artificial pumice. Two small outcrops of pale-red lithoidal rhyolite lava, associated with pumiceous layers altered to white clay, were observed in a ravine on the northeast side of the dome and at the southwestern corner of the dome. The position of these atypical outcrops and an association with subhorizontal flow layering suggests that they occur very near to the base of the flow.

The Grefco perlite dome appears to consist of a homogeneous, very phenocryst-poor lava throughout the body. A random sample taken from the ore stockpile (Table 7, no. 77-5-4) is considered to be representative of the bulk of the lava dome. This sample contains about 1 percent medium- to fine-grained phenocrysts, most of which are clear and inconspicuous in the glassy groundmass. In thin section, tabular and embayed crystals of slightly zoned oligoclase (An_{25}) and sanidine appear to be about subequal in volume and form the dominant phenocrysts. Small, rounded quartz crystals are about five times less abundant than the feldspars. Fine flakes of reddish-brown biotite are present in trace amounts in this specimen; however, biotite appears to occur sporadically in outcrops at the open pit area.

Chemical analyses reported by Weber (1963b, p. 144) and in this report (Table 7, no. 77-5-4) indicate that the Grefco perlite dome is a high-silica rhyolite containing about 76 to 77 percent SiO_2 as an anhydrous rock. The earlier analysis yielded a water content of about 3.6 percent. The sample collected from the ore stockpile at the Grefco mill has given a K-Ar whole rock date of 7.4 m.y. This age is in good agreement with the stratigraphically older phenocryst-rich lavas at Cook Spring dated at 10.5 m.y., and the overlying basalt of Sedillo Hill dated at 4.0 m.y. (Bachman and Mehnert, 1978).

Sierra Ladrones Formation

Based on mapping of the San Acacia quadrangle, about 20 km north of Socorro (fig. 2)., Machette (1978) has designated the upper subdivision of the Santa Fe Group in central New Mexico as the Sierra Ladrones Formation. The formation is defined as follows (ibid.):

> "The Sierra Ladrones Formation of early Pliocene to middle Pleistocene age is here named for the Sierra Ladrones (low foothills of the Ladron Mountains) and consists of alluvial fan, piedmont slope, alluvial flat, floodplain, and axial stream deposits and locally derived basalt flows."

Only the "axial stream deposits" (Machette, 1978, "QTsa"), synonomous with "deposits of the ancestral Rio Grande", clearly distinguish the valley-fill sequence of the Sierra Ladrones Formation from the bolson deposits of the underlying Popotosa Formation. Sands of the ancestral Rio Grande crop out almost continuously along the course of the present Rio Grande from as far north as Santa Fe to south of El Paso (Chapin and Seager, 1975; Hawley, 1975; Bachman and Mehnert, 1978). In northern New Mexico, these ancient river sands are included in the upper buff member of the Santa Fe Formation of Bryan and McCann (1937). In southern New Mexico and Texas, these fluvial sands are included in the Camp Rice Formation of Strain (1966) and of Seager and others (1971).

Prior studies in the Socorro region indicate that the axial stream deposits of the Sierra Ladrones Formation form a continuous stratum in the subsurface of the Rio Grande Valley between the type area at San Acacia and Socorro. As mapped by DeBrine and others (1963, fig. 2, "Tsfg"), "axial river sands and gravels" of the "ancestral Rio Grande" form a wide belt of locally discontinuous outcrops paralleling the east side of the Rio Grande from San Acacia to 5 km south of Socorro. Although individual outcrops are discontinuous, their belt-like distribution coupled with the fluvial nature of the deposit implies continuity in the subsurface. A mastadon jaw and horsetooth found in these light-gray, arkosic fluvial sands in Arroyo de la Parida (near Johnson Hill, fig. 2) have been assigned a late Pliocene (Blancan) age (Needham, 1936).

As described by Machette (1978), the remainder of the Sierra Ladrones Formation consists predominantly of piedmont-slope deposits generally derived from areas of the present-day ranges. In the type area, two units of piedmont-slope deposits are recognized to intertongue with and lie below or above, the axial stream sands.

In the Socorro Peak volcanic center, correlatives of the Sierra Ladrones Formation crop out in large patches on the northeast, southeast and southwest flanks of the Socorro Mountains; respectively; these localities are: Nogal (Escondida) Arroyo, Socorro Canyon (arroyo), and Sedillo Hill. Smaller isolated outcrops of piedmont-slope gravels assigned to the Sierra Ladrones Formation also occur at the east and west margins of the northern Chupadera Mountains, near Black Canyon and along the eastern margin of the mapped area, in the Socorro Basin (fig. 3).

In the Socorro Peak area, the Sierra Ladrones Formation generally rests in angular unconformity on strata of the upper Popotosa Formation or locally on older strata. The top of the formation is placed at the base of thin veneers of gravelly alluvium (Qoa), which rest on pediment surfaces cut across poorly indurated strata of both the upper Popotosa and the Sierra Ladrones formations. These pediment veneers are considered to be of Quaternary age because of their geomorphic expression.

Age limits of the Sierra Ladrones Formation are not as well defined in the Socorro Peak area as they are in the type area. The early Pliocene to middle Pleistocene age assigned to the formation in the type area (Machette, 1978) is thought to generally apply here also. The Sierra Ladrones exposures near Socorro Canyon are evidently younger than the 7.4-m.y.-old Grefco dome and older than the basalt of Sedillo Hill dated at 4.0 m.y. by Bachman and Mehnert (1978). Here, the basalt of Sedillo Hill locally underlies the eroded top of the Sierra Ladrones Formation; however, at its vent area near Sedillo Hill, these basalt flows occur at the base of the Sierra Ladrones Formation. Thus the base of the formation is clearly not everywhere the same age.

The distribution and thickness of the Sierra Ladrones Formation is primarily controlled by varying amounts of erosion and preservation across tilted fault blocks. The thickest preserved sections occur in the Socorro and La Jencia fault block basins (fig. 3), which have been superimposed across the older bolson of the Popotosa basin. The thickest exposed section, about 350 m thick, crops out in the Nogal Arroyo area. The thickness of the Sierra Ladrones Formation in the Socorro basin east of Socorro Peak is unknown; it may be significantly greater than 350 m.

Within the Socorro Peak area, the Sierra Ladrones Formation generally consists of an upward coarsening sequence of mud, silt, sand and gravels or their weakly indurated equivalents. This upward coarsening is indicative of a progressive increase in local topographic relief since late Miocene or early Pliocene time (about 7-4 m.y. ago). The older (middle or early Pliocene) fluvial deposits of the Sierra Ladrones Formation do not indicate the presence of significant relief where the modern Lemitar, Socorro and Chupadera mountains now exist (see fig. 33). However, the younger piedmont-slope gravels of the Sierra Ladrones Formation can be directly related to the modern topography.

Both the inception of the ancestral Rio Grande and the progressive topographic development of the modern mountain ranges at Socorro can be explained by a combination of rifting and contemporaneous epeirogenic uplift of the southern Rocky Mountains and adjacent regions since about 7 m.y. ago (Chapin, 1978). The writer believes that epeirogenic uplift is the primary cause for the progressive topographic development of the Socorro Peak area, since local rift faulting in early to late Miocene time was <u>not</u> associated with the development of significant topographic relief (cf. fig. 26, fig. 33, pl. 1).

lower Sierra Ladrones Formation and basalt of Sedillo Hill time

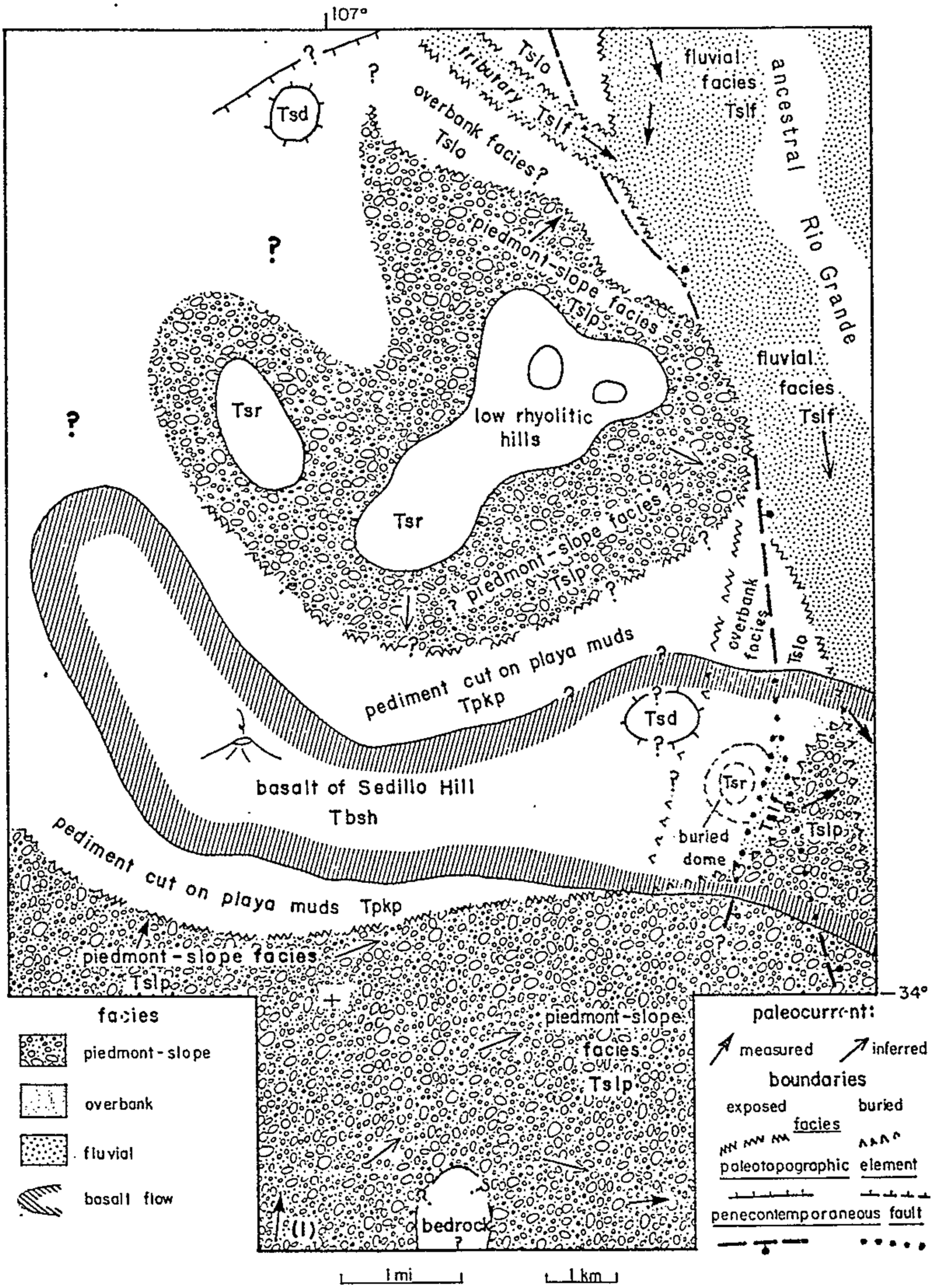


Figure 33. Paleogeologic map of the Socorro Peak volcanic center in middle Pliocene time summarizing depositional relationships of the lower Sierra Ladrones Formation and the interbedded basalt of Sedillo Hill. See Plate 1 for explanation of map symbols.

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[1] Paleocurrent direction from G.P. Osburn (1979, oral commun.)

The Sierra Ladrones Formation has been divided here into three sedimentary facies: 1) an overbank mud and silt facies of the ancestral Rio Grande (Tslo), 2) a fluvial sand facies of the ancestral Rio Grande (Tslf), and 3) a piedmont-slope facies composed mainly of sandy gravels derived from adjacent mountain ranges. The basalt of Sedillo Hill (new name), erupted about 4 m.y. ago from vents near Sedillo Hill, exhibits relationships indicating that it is interbedded with the piedmont-slope facies. The general relationships of these sedimentary facies immediately after the eruption of the basalt of Sedillo Hill are summarized on figure 33. Since that time, an eastward prograding wedge of piedmont-slope gravels has been shed from the developing Socorro-Lemitar Mountains causing the river to retreat to the east and largely burying the older fluvial sands in the Socorro basin. A similar wedge of alluvial gravels derived from the eastern Magdalena Mountains also later prograded across the basalt of Sedillo Hill as far east as Black Mesa. The following descriptions of units in the Sierra Ladrones Formation are generally arranged from oldest to youngest, with the recognition that the ages of the units overlap.

Overbank Facies (Tslo)

Reddish-brown to light-brown, weakly indurated beds of mud, silt, and sand, which locally grade upward into distinctive channel sands (Tslf) of the ancestral Rio Grande, are here interpreted as overbank or floodplain deposits (Tslo) of the fluvial system assigned to the Sierra Ladrones Formation. As mapped here, the overbank facies may be equivalent to the uppermost fine-grained part of the "piedmont-slope and alluvial-flat deposits" of Machette (1978, "Tsp"). The relationship between the "alluvial-flat" and "axial-stream deposits" in the type area is uncertain because the inferred contact between them is not exposed (ibid). The overbank facies is believed to interfinger laterally with the fluvial facies (Tslf), although this relationship has not been locally verified because of poor exposures. DeBrine and others (1963) have reported such interfingering at the east side of the valley near Socorro.

North of Nogal Arroyo approximately 60 m of these fine-grained strata appear to unconformably overlie claystones (Tpkp) of the upper Popotosa Formation and underlie a gradational contact with the fluvial facies (Tslf). The angular unconformity at the base of the overbank facies is only readily apparent near the north boundary of the map area. Here the lighter colored non-gypsiferous silty muds of the overbank facies truncate a rhyodacitic marker tuff (Tsrt) of the Socorro Peak Rhyolite and its enveloping gypsiferous maroon claystones (Tpkp).

The distinction between fine-grained overbank deposits of the Sierra Ladrones Formation and playa deposits of the upper Popotosa Formation is not everywhere certain, particularly where the distinctive fluvial-facies sands (Tslf) are not also present. As much as 6 m of thin- to medium-bedded silty muds and fine-grained, friable, arkosic-looking sandstones have been observed on the "under side" of landslide blocks of the basalt of Sedillo Hill (Tbsh) on the north and south flanks of Black Mesa. Because of their similarity to arkosic sands of the fluvial facies (Tslf) and their apparent original stratigraphic position, these fine-grained beds are tentatively assigned to the overbank facies. Should this correlation be verified, then the limit of the overbank facies shown on the southeast part of figure 33, can be moved westward under Black Mesa.

Several small outcrops of reddish-brown, non-gypsiferous mudstone and siltstone are exposed in the older road cuts on the south side of Highway Sixty and in the upper slopes of the flat-bottomed canyon further south. The correlation of these isolated fine-grained outcrops with the Sierra Ladrones overbank deposits is considered to be the most tenuous; alternatively these small exposures could represent a transition between the "Tpke" and "Tpkp" facies of the upper Popotosa Formation. In the future, petrographic studies of sands in the Popotosa and Sierra Ladrones formations could resolve some of these correlation problems.

Fluvial Facies (Tslf)

As used here the "fluvial facies" (Tslf) is largely equivalent to the "axial-stream deposits" of the Sierra Ladrones Formation of Machette (1978, "QTsa"). However, here the fluvial facies also includes some deposits of a tributary to the trunk stream of the ancestral Rio Grande (fig. 33). The writer is indebted to Mike Machette for pointing out the similarities of the sand bodies north of Nogal Arroyo to the ancestral Rio Grande sands near San Acacia via field excursions to both areas.

The tributary subfacies, which is undifferentiated on the geologic map, crops out mostly west of the common line between sections 28 and 29 in the Nogal Arroyo area. The distinction between trunk stream and tributary stream subfacies is based on minor differences in pebble lithology and paleocurrent directions in otherwise similar sandstones.

The fluvial facies, in general, consists predominantly of light-gray sand and friable sandstones with lesser volumes of pebbly conglomeratic sandstones, and rarely some muddy abandoned-channel fills. The sands are fine to coarse grained, moderately well sorted, medium to very thick bedded, and commonly cross bedded (fig. 34). Fine grained sands tend to be parallel laminated and locally exhibit parting lineations; whereas coarse grained sands and conglomeratic beds commonly display large-scale trough cross-bedding. The coarse beds are often moderately well indurated with calcite cement; therefore festoon-shaped sandstone outcrops surrounded by slopes of loose or friable sands are fairly common. Some yellowish-brown limonite staining and black manganese locally occur in association with silicified logs in the trunk-stream deposits. Sands of the fluvial facies commonly have an arkosic appearance imparted by pink or reddish granitic(?) feldspars.

A wide variety of subrounded volcanic pebbles, many of which are probably recycled from moderately indurated Popotosa fanglomerates (Tpkf, Tplb, etc.), are commonly the dominant clasts in the conglomeratic sands of fluvial facies. However, in the trunk-stream-facies conglomeratic beds; more than 10 percent of the clasts characteristically consist of smooth well-rounded pebbles of quartzite,

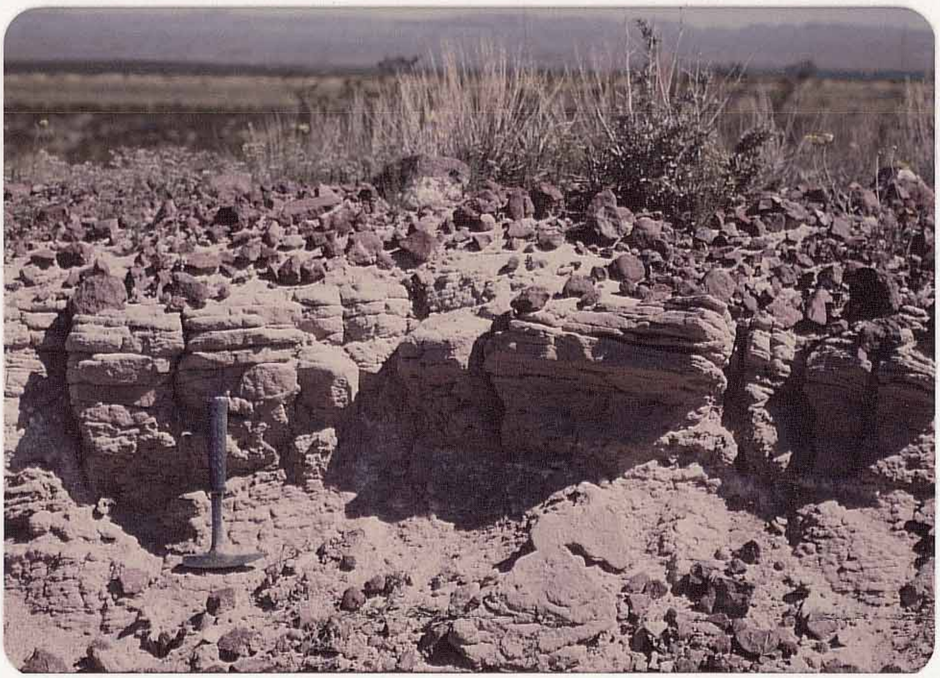

Figure 34. Fluvial facies sandstones (Tslf) of the Sierra Ladrones Formation exhibiting trough cross bedding common in this unit. Paleocurrent direction is generally from left to right. The surficial lag of rounded volcanic cobbles is derived from local deflation of a sandy gravel pediment veneer (Qoa) underlying the grassy surface behing the outcrop. Hammer is 38 cm long. Location is about 0.8 km north of Nogal Arroyo (SW corner, SE/4, SW/4, Sec. 21, T2S, R1W).

granite, and chert. These well rounded pebbles, most likely from distant source areas to the north, are absent in the tributary sands. Red and greenish non-indurated clay balls are also common clasts in the fluvial facies; they probably indicate lateral scouring of overbank deposits by sinuous or meandering channels.

Volcanic clasts in the tributary facies seem to be primarily of local origin. Sparse amounts of yellowish-brown altered A-L Peak Tuff clasts are apparently recycled pebbles from the upper Popotosa fanglomerates (Tpkf). Rare clasts of red volcanic conglomerate or breccia, apparently derived from the extremely indurated lower Popotosa conglomerates (Tplrl, Tpsd), have also been observed in the tributary facies. These rare clasts are interpreted as an early signal of the complete disruption of the Popotosa basin by a rising ancestral Socorro-Lemitar Mountains.

Numerous observations of paleocurrent directions in the fluvial facies (including "Lemitar Map") indicate a general southerly transport for the trunk-stream subfacies and a dominant southeasterly transport direction for the tributary subfacies. Variations in paleocurrent direction within each of the subfacies suggest these sands were deposited in braided or sinusoidal channels.

The sands of the ancestral Rio Grande form a major aquifer nearly 10 km wide and of uncertain thickness under the valley at Socorro. The western limit of these fluvial sands is essentially structurally controlled by the range bounding fault zone at the foot of Socorro Peak. About 300 m due east of the Middle mine, and east of the locally inferred range-bounding fault, a thermal gradient hole intersected 82 m of arkosic fluvial sands below 70 m of piedmont-slope gravels (pl. 2, B-B'). Topographic relief on outcrops of the fluvial facies at Nogal Arroyo and Socorro Canyon indicates minimum thicknesses of about 120 and 90 m, respectively. Assuming a 10 degree southerly dip in the Nogal Arroyo area there may be as much as 210 m of fluvial strata south of the arroyo.

Relationships at the basalt-capped ridge north of Socorro Canyon demonstrate that the range bounding fault zone has acted as a growth fault during Sierra Ladrones deposition. For this reason, the writer would not be surprised if the fluvial facies were considerably thicker on the east side of this fault zone, possibly as much as 305 m.

The eastern limit of the fluvial facies, which occurs outside of the mapped area, is most likely a depositional pinch out. East of Socorro these ancient river sands have been described as intertonguing with alluvial-fan deposits that are rich in Paleozoic detritus (DeBrine and others, 1963) shed from the Loma de las Canas uplift (fig. 2).

Piedmont-slope facies (Tslp)

In the Socorro Peak volcanic center, the piedmont-slope facies (Tslp) of the Sierra Ladrones Formation consists of wedge-shaped alluvial fan and coalescent fan (bajada) deposits. These sandy gravels were shed from the rising Socorro-Lemitar Mountains and from the eastern Magdalena Mountains, mostly during late Pliocene and early(?) Pleistocene time. The relatively low block of the northern Chupadera Mountains, largely buried by the gravel apron from the Magdalena range (fig. 33), was not a significant sediment source during Sierra Ladrones time.

Northeast of Socorro Peak, in the Socorro Basin, the piedmont-slope facies conformably overlies the fluvial facies (Tslf). Intertonguing relationships of these facies, generally not well exposed in the study area, have been observed by J.W. Hawley and the writer in the road cuts along I-25 just north of Nogal Arroyo (fig. 2). The second cut northeast of the Escondida exit is the most revealing. As illustrated at the east end of section A-A' (pl. 2), the alluvial-fan gravels in the Socorro Basin are thought to have prograded eastward as Socorro Peak grew, thus burying the ancestral Rio Grande sands and forcing the channel eastward.

Piedmont-slope deposits, which could be appropriately assigned to the Sierra Ladrones Formation, have not been observed below the fluvial facies in the Socorro Peak area. Thus Machette's "Tsp" unit (1978) has no apparent correlative in this map area.

In a similar manner, the alluvial fan deposits from the northeastern Magdalena Mountains originally prograded northeastward across the basalt of Sedillo Hill (fig. 33) and onto the western slopes of the Socorro Mountains. Isolated outcrops of well-indurated gravels, cemented by pedogenic carbonate (caliche), which occur south of Jejenes Hill and which partly cap Black Mesa, contain volcanic clasts (Lemitar Tuff, etc.) that must have come from the Magdalena Mountains.

The piedmont facies is mostly poorly indurated and therefore typically forms rounded hills and slopes mantled with cobbles and pebbles, mostly from volcanic rocks. White caliche rinds are common on these cobbles at Sedillo Hill; however, this aspect is a more distinctive trait for the younger surficial gravel deposits (Qoa).

Where locally exposed in arroyo cuts, this unit consists of pale-red, light-brown, or pinkish-orange sandy gravels and loamy sands. Individual outcrops are medium to thick bedded and generally exhibit crude subparallel stratification. Cut and fill structures are common.

Essentially all major stratigraphic units, from Precambrian to late Miocene age that are exposed in the ranges of the map area, are represented in the clasts of the piedmont-slope facies. Lateral variations in clast composition also matches those seen in the bedrock of the adjacent ranges. Volcanic clasts are dominant in all areas. Northeast of Socorro Peak, clasts from the Socorro Peak Rhyolite and basal Popotosa conglomerates are abundant. North of Nogal Arroyo, sparse Precambrian metamorphic clasts and Paleozoic limestone clasts appear indicating input from the Lemitar Mountains. South of Socorro Peak and west of the range front, the piedmont-slope deposits are characterized by abundant upper Lemitar Tuff clasts, which commonly contain mafic rhyodacite pumice (fig. 13) thereby indicating a source in the Sixmile Canyon area.

The best exposure of the piedmont-slope facies occurs near the range front on the north side of Socorro Canyon arroyo. Here the piedmont-slope facies and the fluvial facies are both capped by remnants of the basalt of Sedillo

Hill, which form an east-trending ridge. As indicated on figure 33 these facies were juxtaposed by a fault zone prior to eruption of the basalt of Sedillo Hill about 4 m.y. ago. Continued movement on the fault zone has also offset the basalt flow. The piedmont gravels in this fault zone are anomalously well indurated with carbonate cement and, locally, they are cut by west-northwest trending banded calcite veins exposed in the arroyo floor. These features suggest some minor hydrothermal activity has occurred here since middle Pliocene time.

The thickness of the piedmont-slope facies may be generally estimated from local topographic relief, since these beds are only slightly tilted by faulting. The thickest exposed section, at Sedillo Hill, is about 110 m thick. Most other exposures are at least 60 m thick. In both the Socorro basin and the La Jencia Basin (Sedillo Hill area), these gravels appear to thicken toward the range fronts and their bounding fault zones.

Basalt of Sedillo Hill (Tbsh)

Olivine-bearing basalt flows, erupted 4 m.y. ago (Bachman and Mehnert, 1978) from vents north of Sedillo Hill, are here informally named the basalt of Sedillo Hill for prominent outcrops around the vent area. Erosional remnants of the basalt of Sedillo Hill form a dogleg-shaped band of topographically high cuestas and mesas with the vent area at the crook. From the vent area, these black cliff-forming outcrops trend north towards Snake Ranch Flat and east toward Black Mesa and the Grefco mine area.

A stack of three flows (9-15 m thick) in the vent area is as much as 33 m in total thickness. The number of flows in individual outcrops and their overall thickness decreases away from the vent area. Thin remnants of a single flow northeast of the Grefco mine are less than 5 m thick. However; south of here, west-tilted blocks of lava at Socorro Canyon are anomalously thick with respect to surrounding mesa forming outcrops. Here the flow may have ponded in a sag structure related to the range bounding fault zone. The basalt horizon on Black Mesa thickens to the south; two flows are present here in comparison to a single flow on the north side.

The apparent abrupt thinning of the basalt flows south of the vent area toward Sedillo Hill is probably the result the lavas having wedged out onto a north-facing piedmont slope (fig. 33); local erosion of the upper surface of the flows during their burial by piedmont-slope gravels (Tslp) may also have decreased their thickness. These gravels can be seen to fill a 15-m-deep channel cut in the basalt flows when viewed looking northwest from the base of Sedillo Hill (see pl. 1 south of southern vent).

Northwest of the vent area, reddish-brown bedded tuffs and agglomerates overlie the basalt flows and have been mapped as a separate unit (Tbsht, pl. 1). The tuff beds consist of small lapilli of yellowish-brown (hydrated?) basaltic pumice, and granules of black lava, bound by a splotchy matrix (?) of reddish-brown mud. The cementing muds were probably originally lithic fragments of playa deposits (Tpkp).

The basalt of Sedillo Hill is considered to be interbedded in the piedmont-slope facies of the Sierra Ladrones Formation, even though it can not be seen as such at a single location. North of Sedillo Hill, these flows rest unconformably on playa deposits of the Popotosa Formation (Tpkp) and are disconformably overlain by the piedmont-slope facies. Southeast of the Grefco mine, the basalt of Sedillo Hill lies conformably on the piedmont-slope deposits and unconformably(?) on older fluvial-facies deposits (fig. 33).

Two vent structures for the basalt of Sedillo Hill are exposed in steep walled drainages cut in the thickest portions of these flows north of Sedillo Hill. The northern vent, exposed at the mouth of a small canyon, is visible in the south wall as an east-northeast striking vertical dike defined by vertically oriented flattened vesicles. Flaggy fractures are aligned subparallel to the vesicle flow structure in the dike. Above the feeder dike, these flaggy fractures fan out and merge with the overlying subhorizontal flow. Steeply dipping curviplanar fracture foliations define a pipe-like structure at the southern vent.

Most of the Sedillò Hill lavas are similar to the sample from the Black Mesa area (Table 6, no. 76-6-11). These dark-gray, fine-grained aphanitic rocks contain sparse, small phenocrysts of yellowish-brown olivine and blue irridescent pyroxene in a fresh, grainy looking groundmass. In thin section, subhedral phenocrysts of olivine and microphenocrysts of clinopyroxene (augite) and plagioclase (An_{65}) appear to grade with a seriate texture into an intergranular groundmass of plagioclase, olivine, clinopyroxene and opaque Fe-Ti oxides.

An uncommon variety of the basalt of Sedillo Hill is found in the flow near the north end of the Black Hills (fig. 3). This medium-gray lava contains sparse, tabular phenocrysts of plagioclase, which are mostly 1-2 mm long and rarely over 1 cm long. Phenocrystic olivine is either rare or absent in this flow. Traces of xenocrystic(?) quartz have also been observed here. This uncommon variant of the basalt of Sedillo Hill may be a basaltic andesite with about 54 percent SiO_2, since it is similar to an analysed specimen from the basalt of Kelly Ranch (Table 6, no. 76-1-9). An analysis of the common olivine-bearing variety of Sedillo Hill lavas (Table 6, no. 76-6-11) indicates that the bulk of this unit is an alkalic basalt containing about 50 percent SiO_2. Bachman and Mehnert (1978, no. 25) have reported a K-Ar whole-rock age of 4.0±0.3 m.y. for a sample from the Black Mountain area (Black Mesa).

QUATERNARY SURFICIAL DEPOSITS

The Quaternary history of south central New Mexico, with emphasis on the Rio Grande valley system, has been described in a summary paper by Hawley (1975). This paper provides a regional framework applicable to the Socorro Peak area most of the concepts of Quaternary evolution expressed herein draw heavily on this summary paper.

On a regional scale, landscape development during the Quaternary period has been controlled by continued epeirogenic uplift and more localized effects of rifting (Hawley, 1975; Chapin, 1978). Individual cycles of erosion and deposition are respectively associated with the waxing and waning periods of Quaternary glaciations (Hawley, 1975). These episodes of drainage entrenchment and back filling are evident in the stepped landscape of broad piedmont surfaces (Qoa), terraces (Qoa) and entrenched arroyos (Qya) shown on the geologic map (pl. 1). The downcutting of these drainages, tributary to the Rio Grande, has occurred chiefly since middle Pleistocene time (0.2-0.6 m.y.) (Hawley, 1975). This lowering of base level of the Rio Grande and its tributaries was triggered by capture of the ancestral Rio Grande near El Paso and its integration with a lower drainage extending to the Gulf of Mexico (Kottlowski, 1958; Hawley, 1975).

The area of Socorro Peak and the adjacent basins is presently underlain by a deep magma body and shallow extensions of this body (Chapin and others, 1978; see pl. 3, map 4). Thus, magma intrusion may also be a locally significant factor controlling uplift and deformation in the Socorro Peak area in relatively recent geologic time. Variations in elevation of the flows of the Socorro Peak Rhyolite and the basalt of Sedillo Hill -- discussed at greater length in the structure section -- appear to define an elliptical uplift trending west-southwest from Socorro Peak that has formed since 4 m.y. ago. This elongated dome, roughly coincident with the main locus of shallow magma intrusion (pl. 3, map 4), is approximately outlined by the elliptical drainage pattern created by South Canyon -- Water Canyon-Nogal Arroyo on the north side, and by Sixmile Canyon -- Socorro Canyon on the south side. This consequent drainage pattern was generally established in early(?) Pleistocene time at the maximum aggradation of the ancestral Rio Grande (Sierra Ladrones Formation). Since then, these drainages have episodically cut down into the poorly indurated basin-fill deposits (upper Popotosa Formation and Sierra Ladrones Formation) and now locally cut across resistant blocks of volcanic bedrock. Thus, much of the present bedrock topography of the northern Chupadera Mountains and the lower flanks of the Socorro Mountains has only been recently exhumed. These areas lie below remnants of the basalt of Sedillo Hill (Tbsh) and some remnants of late Pliocene piedmont-slope deposits (Tslp) west of Black Canyon (see Osburn, 1978, pl. 1).

Approximately two thirds of the mapped area is covered by a thin mantle of surficial deposits of Quaternary age. These morphologically defined units have been mapped primarily on aerial photographs at a scale of about 1:29,000. Field checks on the placement of contacts and the lithology of the deposits have been relatively limited. Fault scarps and fault-line scarps displacing geomorphic surfaces developed on the older alluvial deposits (Qoa) are also primarily based on photo mapping.

The oldest surficial deposits (Qoa) consist of alluvial-fan and piedmont-slope deposits resting unconformably on broad pediment surfaces cut on tilted strata of the upper Popotosa Formation and the Sierra Ladrones Formation. Because of this widespread unconformity at the base of the oldest surficial units, they are here arbitrarily and collectively labeled as "post Santa Fe Group" deposits (pl. 1). The highest (oldest) pediment veneers of the older alluvium (Qoa) will probably be reassigned to the Santa Fe group when their absolute age is clearly established.

Older Alluvium (Qoa)

Areas mapped as older alluvium (Qoa) consist of broad and dissected pediment veneers, alluvial-fan deposits, and strath terraces. Four elevations of geomorphic surfaces associated with these deposits are recognized to project at: 70-75 m, 37-43 m, 18-21 m, and 6 m above the adjacent arroyo floors. Sanford and others (1972, map 2) have correlated these surfaces at Socorro with geomorphic surfaces defined by Denny (1941) in the San Acacia area. However, for this report, these geomorphic surfaces are grouped together, as a general unit. Their regional correlation by elevation alone is considered uncertain, since these surfaces are displaced and warped in several places along the range-bounding fault zone of the Socorro Mountains (pl. 1).

The older alluvial deposits are usually less than 6 m thick and most likely are never more than about 20 m in thickness. Gravel and sandy gravel, compositionally similar to the piedmont-slope facies (Tslp) of the Sierra Ladrones Formation, are the dominant lithologies of the older alluvium. Therefore, where these two units (Qoa, Tslp) are shown in contact (pl. 1) with each other, the younger unit is arbitrarily defined by its surface expression. Older alluvial-flat (?) deposits (Qoa) in the Snake Ranch area locally include dark-brown mud and silt.

The older pediment veneers are commonly poor to moderately well indurated by carbonate cement associated with pedogenic horizons near their upper surface and with groundwater accumulations near an impervious (playa mud, Tpkp) lower surface. The highest, and smallest, remnant of a pediment veneer occurs south of Nogal Arroyo. Here the gravelly deposit has a well indurated cap of caliche about 2 m thick.

A general comparison with pedologically dated units in the San Acacia area (Machette, 1978) suggests that the older alluvium deposits may range in age from middle to late Pleistocene. The two lower surfaces associated with strath terraces are clearly of late Pleistocene, post-Santa Fe age.

Landslide Deposits (Ql)

Landslide deposits around Strawberry Peak, Socorro Peak, Black Mesa and the Black Hills are characterized by hummocky topography formed by slump blocks of lava.

These toreva and glide blocks range from 30 to 600 m in length and 6 to 180 m in thickness. The lava blocks are sliding downslope on playa claystones (Tpkp) of the upper Popotosa Formation. Based on the lithology of the slump blocks, the landslide deposits have been subdivided into three separate fields (Qlr, Qld, Qlb), which match composition of the bedrock lavas upslope (pl. 1). These fields are topographically bound by local drainages and saddles. Some slump blocks along the west side of the Socorro Mountains have been isolated from their source by modern drainages.

As mapped, the landslide areas include undifferentiated talus and colluvial aprons around large blocks. Fine-grained young alluvium, which commonly fills small, closed depressions developed on the upslope side of larger blocks, is generally not shown.

East of Strawberry Peak, individual slump blocks of quartz latitic lava (Tsd), colluvium (Qca), and young alluvium (Qya), have been locally subdivided to show their geometric relationships. In the same area south of Strawberry Canyon, crudely bedded colluvial deposits, mapped as landslide (Qld), have been downdropped against playa deposits (Tpkp) by a late Pleistocene fault. This fault also appears to have warped or displaced(?) a terrace deposit (Qoa) a little further to the north.

Telephone lines and poles across the landslide area between Blue Canyon and Socorro Peak have been stretched and broken during active periods of sliding after long rainy periods. The development of these landslide areas probably began about middle Pleistocene time when significant arroyo entrenchment produced moderately steep slopes in the adjacent playa deposits.

Colluvium and Minor Alluvium (Qca)

The "Qca" unit consists primarily of talus cones, rock-fall deposits and stabilized colluvial aprons deposited on steep slopes and formed by mass wasting of well-indurated bedrock. Also included are extensive slope-wash veneers of reworked "Qoa" deposits that mantle moderate to gentle slopes cut on playa claystones (Tpkp) or poorly indurated deposits (Tslf, Tslo) of the Sierra Landrones Formation.

Numerous small and isolated patches of colluvium are not shown on the geologic map. Larger areas of colluvium are shown as accurately as possible, particularly where stratigraphic contacts or faults are obscured.

The contact between talus cones (Qca) adjacent to cliffs of the Socorro Peak Rhyolite and the landslide areas (Ql) below the cliffs is gradational. This locally obscure boundary is generally placed at the transition from talus to large landslide blocks evident as hummocks. Colluvial deposits locally grade into older alluvium (Qoa) or younger (Qya) alluvium. Smaller areas of this nature have not been differentiated from the colluvial deposits.

The largest talus cones are probably not more than 50 m thick, measured perpendicular to their upper surface. Colluvial deposits are considered to be mostly late Pleistocene to Holocene in age.

Younger Alluvium (Qya)

Younger alluvium (Qya) is comprised by mud, silt, sand and gravel in arroyo floors, valley bottoms, and low terraces less than 3 m above the drainages. Some fine-grained sandy deposits in small, closed depressions are included with the younger alluvium unit. Arroyo-fill deposits are most likely less than 15 m thick. These non-indurated sediments are considered to be of latest Pleistocene or Holocene age.

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VOL II

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CENOZOIC STRATIGRAPHY AND STRUCTURE OF THE SOCORRO PEAK VOLCANIC CENTER, CENTRAL NEW MEXICO

VOLUME II: STRUCTURE

by

Richard M. Chamberlin

1980

A study of an area where late Cenozoic volcanism, sedimentation, and structure related to the Rio Grande rift have been overprinted on an Oligocene resurgent cauldron.

A Thesis submitted to the Faculty and the Board of Trustees of the Colorado School of Mines in partial fulfillment of the requirements for the degree of Doctor of of Philosophy (Geology).

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ABSTRACT

In central New Mexico, the Rio Grande rift has broken a northeast-trending chain of Oligocene cauldrons, and a surrounding volcanic plateau, into a series of north-trending, tilted fault-block ranges and alluvial basins. The cauldrons lie along the ancient crustal flaw of the Morenci lineament, which has been reactivated within the rift as a deep seated zone of lateral shearing. This transverse shear zone is a diffuse domain boundary at the surface, where it separates fields of tilted fault blocks that are rotated and step-faulted in opposing directions. In cross section, the closely spaced, normal fault blocks look similar to a train of fallen dominoes.

The Socorro Peak volcanic center lies within the rift, at the east end of the cauldron complex. Oligocene volcanic strata exposed here represent the remnants of the north-half of the resurgent Socorro cauldron, tentatively correlated with eruption of the Lemitar Tuff about 28 m.y. (million years) ago. Nearly 900 m of cauldron-facies Lemitar (?) Tuff, which contains caldera-collapse breccias and bedded lag-fall breccias, is exposed on the resurgent block. The Lemitar Tuff outflow sheet covered domino-style fault blocks of the early rift that were largely buried by wedge-shaped prisms of basaltic-andesite lavas. The moat of the Socorro caldera was filled to overflowing by local eruptions of lithic-rich rhyolitic ash-flow tuffs, intermediate lavas, and rhyolite domes; these moat deposits are collectively named the Luis Lopez Formation.

During Miocene time, the northern part of the Socorro cauldron was unconformably buried by the Popotosa Formation, which consists of heterolithic mudflow deposits, fanglomerates, playa mudstones, and minor interbedded basalt flows.

From 12 m.y. to 7 m.y. ago, numerous rhyodacite to high-silica rhyolite domes and tuffs of the Socorro Peak Rhyolite were erupted onto the playa floor of the Popotosa basin as it continued to fill. These silicic domes define a north-northwest-trending intrusive belt, which is widest where it crosses the buried ring fracture zone of the Socorro cauldron at Socorro Peak. Between 7 to 4 m.y. ago, continued rift faulting, combined with epeirogenic uplift, disrupted the floor of the Popotosa basin and began to form the modern ranges and basins of the Socorro Peak area. About 4 m.y. ago, basaltic lavas were erupted from vents southwest of Socorro Peak. These lavas flowed eastward across pedimented fault blocks and onto channel sands of the ancestral Rio Grande. East of Socorro Peak, the facies of the Sierra Ladrones Formation consist of these fluvial sands and intertonguing piedmont gravels shed from the modern highlands.

Eruptive events in the Socorro Peak volcanic center have been dated at 28.6, 11.9-10.3, 10.5-9.0, 7.4, and 4.0 m.y. The primary control of this recurrent magma intrusion and related hydrothermal activity has been the "leaky" vertical fabric of the Morenci lineament. In light of this history, it is not surprising that geophysically defined magma bodies, which provide a heat source for the present geothermal anomaly, are again rising under the Socorro Peak area.

STRUCTURE

Structural elements of the Socorro Peak volcanic center represent small segments of regional structures that continue far beyond the limits of the mapped area. The significance of these isolated structures has been realized primarily through integration with regional mapping of the "Magdalena Project" (Chapin and others, 1978).

Regional Structural Setting

The major structural features of the Socorro-Magdalena area and their relationship to the Socorro Peak volcanic center are illustrated by Figure 35. This generalized structure map may be cross referenced with figures 2 and 9 for names of features not labeled here.

The Socorro-Magdalena area has a complex structural grain formed by the intersection of several regional structures that have been successively overprinted, one upon another, during five or more periods of crustal deformation. The dominant regional structural strikes and their bearings (plus or minus 15 degrees) are: north-northwest (N10W), north-northeast (N20E), northeast (N60E) and west-northwest

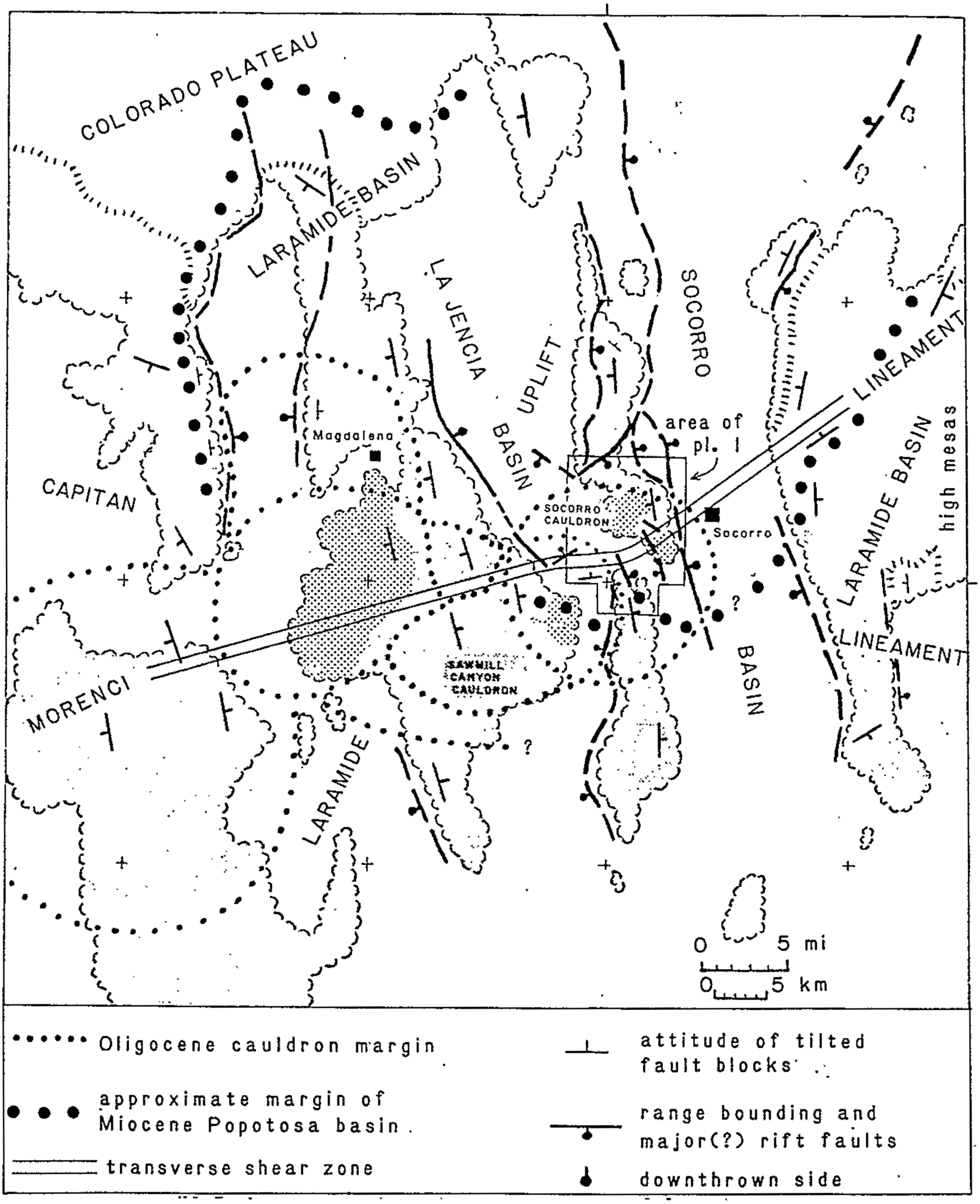


Figure 35. Major structural features of the Socorro-Magdalena area; after Chapin and Seager (1975), Chapin and others (1978), and Machette (1978, Socorro 2-degree sheet, unpub.) with modifications. Numerous closely spaced early rift faults that cut ranges are not shown (cf. pl. 3, map 1). See figures 2 and 9 for additional names of features.

(N75W). With respect to the modern structural grain the first two strikes are longitudinal and the latter two are transverse. From oldest to youngest, regional structures may be grouped into three general categories consisting of: 1) prevolcanic (pre-Oligocene) structures, 2) Oligocene volcanic structures, and 3) structures related to the Rio Grande rift.

Prevolcanic Structures

Prevolcanic structures are poorly known because of widespread cover by Oligocene volcanic rocks and younger basin-fill deposits. However, these older structures are quite significant in providing a pre-existing crustal grain that has been utilized in various ways by extensional stresses which have formed the Rio Grande rift.

The oldest structures in the Socorro region appear to be the northeast-trending (N60E) Morenci lineament and the west-northwest trending (N75W) Capitan lineament of Chapin and others (1978, fig. 1). These deeply penetrating lithospheric flaws, which intersect near Socorro (fig. 35), are characterized by their tendency to "leak" magmas. On regional maps (ibid., fig. 1), they appear as linear belts of mostly Cenozoic intrusions and volcanic vents that cut across New Mexico. The age of inception and the origin of the Morenci and Capitan lineaments have not been established. However, these crustal defects respectively parallel the Precambrian wrench-fault zones of the Colorado lineament (Warner, 1978) and the more debatable Texas lineament (Albritton and Smith, 1956; Moody and Hill, 1956; Turner, 1962), which suggests that they may be similar in age and origin. In the central Magdalena Mountains, foliations in late Precambrian rocks turn from northerly strikes to northeasterly strikes where they approach the projection of the Morenci lineament (Condie and Budding, 1979). However, Precambrian deformation along this northeast trend has not yet been established and the warping could be post-Precambrian (W. Sumner, in prep.). The Morenci and Capitan lineaments as depicted by Chapin and others (1978, fig. 1), and as on Figure 35, are considered to represent major fault zones within broader zones of lesser subparallel faults and fractures, much like the setting of the San Andreas fault in California (Moody and Hill, 1956). The vertical fabric of wrench faults most likely provides a favored path of least resistance for the passive ascent of magmas generated in the lower crust (C.E. Chapin, 1979b).

In late Mississippian to Permian time (ancestral Rocky Mountain orogeny) and later in late Cretaceous to middle

Eocene time (Laramide orogeny), regional compressional stresses formed overall north-trending anticlinal and fault-block uplifts that spanned the length of New Mexico and on into Colorado (Chapin and Seager, 1976, fig. 2). In the Socorro-Magdalena area, a late Paleozoic unconformity in the Joyita Hills (Kottlowski and Stewart, 1970) and bedded breccias in the Madera and Abo Formations near Magdalena (Chapin and others, 1975) indicate deformation of unknown scale related to the ancestral Rocky Mountain orogeny.

Late Mississippian to early Pennsylvanian structures are also apparent in strata of this age exposed on Socorro Peak and in the Lemitar Mountains. In the latter area (Lemitar Map), fault blocks tilted about 55 degrees west expose cross-sectional views of a late Paleozoic thrust fault that originally had a northwest to west-northwest strike. From the southern block to the northern overthrust block, the Kelly Limestone abruptly disappears, and the Sandia Formation thins abruptly from about 180 m to less than 50 m. The thrust fault does not cut the upper Sandia shales or the Madera Limestone.

The absence of the Kelly Limestone on Socorro Peak may be related to local normal faults of late Paleozoic age. The pentagon-shaped Precambrian exposure west of Woods

Tunnel is bounded on the northwest and southwest sides by faults, which do not appear to displace the Madera-Sandia contact along their projections to the west (see pl. 1 for fault attitudes). If the middle to late Cenozoic local rotation of strata is removed (about 40 degrees to west-southwest), then these apparent Pennsylvanian-age faults become northeast and north-northeast-striking high-angle normal faults bounding a small horst block. The Kelly Limestone is absent on this horst block where the base of the late Paleozoic section is exposed. However, it would not be surprising to this writer, if the Kelly Limestone were preserved on the adjacent downthrown blocks where the base of the late Paleozoic section is not exposed. This untested possibility could have economic significance in the hydrothermally mineralized setting of Socorro Peak.

The Socorro-Magdalena area lies on the broad crest of a major Laramide uplift (Chapin and Seager, 1975), which in late Eocene time was beveled by an erosion surface of low relief (Epis and Chapin, 1975). Both this uplift and flanking basins were buried by the Datil volcanic plateau in Oligocene time (Chapin and others, 1975). Stratigraphic relations at the base of the volcanic pile that define the uplift and adjacent basins have been previously described under the heading "Prevolcanic rocks". Because this contact is discontinuously exposed in tilted fault blocks of the rift, only the general positions of the basins and uplift can be reconstructed from isolated control points (fig. 35). Socorro Peak can be indirectly interpreted as a control point on the Laramide uplift. Here, the widespread early Oligocene unconformity at the base of the Spears Formation has been overlapped by a local intravolcanic unconformity related to the north margin of the Socorro cauldron. Evidence that the Spears Formation did once rest unconformably on the Madera Limestone at Socorro Peak is provided by boulders clearly derived locally from the Spears Formation. These distinctive boulders occur in late Oligocene landslide deposits on an inner topographic wall of the Socorro caldera (fig. 18).

The eastern edge of the Laramide uplift at Socorro is most likely marked by small exposures of Precambrian granite about 15 km due east of Socorro Peak. These granite outcrops are locally bound on their eastern sides by north-northeast striking (N20E), high-angle reverse faults upthrown on the west (Wilpolt and Wanek, 1951). Similar relationships distinguish Laramide structures in northern New Mexico (Woodward, 1976).

Oligocene Volcanic Structures

Approximately 37 m.y. ago, intermediate composition volcanism broke out near the heart of the Datil-Mogollon field. Penecontemporaneous erosion of volcanic highlands (large stratovolcanoes?) produced a thick sedimentary apron now known as the Spears Formation, which spread northward across the Socorro-Magdalena area. These early Oligocene calc-alkaline lavas, which may have formed in a subduction environment (Lipman and others, 1972), are considered to have been erupted in an essentially neutral stress field (Chapin, 1974). A few transverse oriented normal(?) faults of early Oligocene age, which control major thickness changes in the Spears Formation (Chapin and others, 1975; Lemitar Map), are of uncertain tectonic significance. These early Oligocene faults may be related to the Capitan or Morenci lineaments.

In middle Oligocene time, volcanism spread into the Socorro-Magdalena area, not long before a shift to eruptions of silicic ash-flow tuffs occurred. Between about 32 to 26 m.y. ago at least seven major ash-flow sheets were erupted from a complex of overlapping and nested cauldrons (fig. 35, fig. 9) that form a large chain extending about 80 km to the southwest from Socorro (Chapin and others, 1978). The general alignment of these cauldrons along the

Morenci lineament is an early indication of its "leaky" nature. Like similar cauldron (caldera) complexes in the San Juan Mountains (Steven and Lipman, 1976), these circular to elliptical subsidence structures are interpreted to mark roofs of individual shallow plutons in a composite batholith. The roofs collapsed as their upper portions were evacuated during voluminous ash-flow eruptions. Ponding of ash-flows within cauldrons (i.e. contemporaneous subsidence) to produce thicknesses often an order of magnitude greater than their equivalent outflow sheet is now recognized in essentially all well-documented cauldrons (Lipman, oral commun., 1978). The Socorro cauldron apparently formed during the initial stages of crustal extension; thus some early rift faults may have accomodated cauldron subsidence.

In contrast to the slightly eroded San Juan calderas, the Socorro-Magdalena cauldron complex and its surrounding ignimbrite plateau have been broken into a myriad of tilted fault blocks by crustal extension within the Rio Grande rift. Several cauldron margins of Figure 35 are generally placed at the transition from thick cauldron-facies tuffs, associated with relatively massive appearing ranges, to thin outflow sheets typically associated with multiple hogback topography. Younger rift faults have offset the Socorro cauldron margin at Socorro Peak (pl. 1). However, similar displacements of most cauldron margins have generally not been recognized (fig. 35), since they have been projected across large areas of younger basin fill.

Rio Grande Rift Structures

The Socorro Peak volcanic center and surrounding region lie within a north-trending zone of middle and late Cenozoic crustal extension known as the Rio Grande rift (Chapin, 1971a; 1979a). On the regional map (fig. $\overset{35}{25}$) only the nearly horizontal strata of the Colorado Plateau and the "high mesas" east of Socorro represent relatively unstretched crust at the margins of the rift. Since the onset of crustal extension, about 32 m.y. ago, the Socorro region has undergone a complex multistage evolution recorded within contemporaneous volcanic and sedimentary strata (Chapin and Seager, 1975; Chamberlin, 1978; Chapin, 1979a). The fault-block uplifts of the Chupadera-Socorro-Lemitar ranges and adjacent basins expose an approximately complete synrift stratigraphic section. The effects of essentially continuous normal faulting, contemporaneous with the deposition of these late Oligocene to Quaternary formations, have been emphasized throughout their descriptions.

Much, but not all, of the structural features of the rift at Socorro may be explained in terms of a hypothetical model of crustal attenuation (rifting) for the Afar depression of northeast Africa. This model proposed by Morton and Black (1975) is described and shown here in Figure 36. In terms of the Morton and Black model, Oligocene ash-flow sheets that dip 5-10 degrees west in the northern Bear Mountains (Masingill, 1978) lie near the rift margin, whereas the same tuffs that dip 50-60 degrees west in the Lemitar Mountains (Lemitar Map) occur near the rift axis.

The writer has referred to progressive westerly rotation of fault blocks in the Lemitar Mountains -- similar to the Morton and Black model -- as "domino style" normal faulting, because the rotated blocks are somewhat analogous in origin and appearance to a train of fallen dominoes (Chamberlin, 1978). Strongly rotated strata repeated by low-angle normal faults -- considered here to be a primary signature of domino-style faulting -- have been observed in many areas of the Basin and Range province, commonly in association with penecontemporaneous volcanism (Mackin, 1960; Anderson, 1971; Profett, 1977). In the Lemitar-Socorro mountains area, two episodes of domino style faulting involving relatively rapid rates of rotation (extension) occurred during, and somewhat after, periods of silicic volcanism from 32-20 m.y. ago and 12-7 m.y. ago (Chamberlin, 1978).

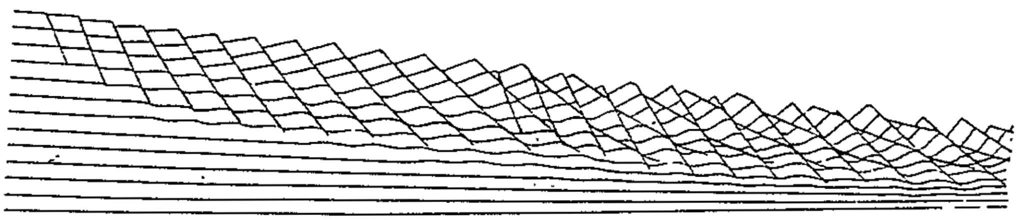


Figure 36. Morton and Black (1975, fig. 4) model of crustal attenuation (rifting) for the Afar Depression of Ethiopia. The depth of faulting is on the order of 10-30 km and the section is to true sacle (i.e. true dips are shown). Essential elements of this model are: 1) progressive necking of the ductile lower lithosphere grades upward into extension of the brittle upper lithosphere by means of closely spaced subparallel normal faults; 2) during rifting progressive rotation of bedding and fault planes occurs such that originally high-angle normal faults become less inclined with time; 3) with 20-30 degrees rotation the original set of faults is no longer of favorable orientation and a new set of faults forms to continue the process; 4) neglecting the effects of contemporaneous erosion, deposition, intrusion, and warping of strata (drag folds, etc.), the dip of bedding may be used as a measure of the amount of crustal attenuation. Horst and graben structures (not shown) are common at the rift margins (Morton and Black, 1975, fig. 3 and 8).

Stewart's (1971) discussion of the origin of Basin and Range structure has exemplified the common viewpoint that crustal extension should be mani-fested either as horsts and grabens or as tilted (domino style) normal fault blocks. In the Socorro area, these contrasting styles of normal faulting are considered to be compatible end members of crustal extension controlled by different thermal regimes. Horst and graben structures are favored by moderate heat flow and a relatively thick brittle crust. In contrast, domino-style faulting is associated with very high heat flow and a relatively thin brittle crust (Chamberlin, 1978). With spatial and temporal variations in thermal regimes, these contrasting structural styles may be superimposed on each other in an otherwise constant extensional stress field.

The rift zone at Socorro is characterized by three general types of structures: 1) longitudinal "domino-style" fault blocks and subsidiary horst-graben structures, both of which dominantly trend north-northwest (N10W) apparently perpendicular to the direction of extension; 2) discontinuous transverse faults, flexures, and twists, which commonly appear to follow pre-rift structural trends and which locally accommodate differential rotation of longitudinal blocks; and 3) Neogene sedimentary basins, which have been broken by uplifts and superimposed one on another as changes in style and locus of block faulting have occurred. These subtly interrelated structures have been overprinted upon each other during crustal extension in the Socorro area.

Crustal extension was first "felt" about 31-32 m.y. ago with the appearance of rhyolitic ash flows of the A-L Peak Tuff and interbedded mafic lavas of the La Jara Peak Basaltic Andesite (Chapin, 1978). Domino-style normal faulting was well underway in the Lemitar Mountains area by 28 m.y. ago when the Lemitar Tuff outflow sheet buried fault blocks already tilted 5-15 degrees to the west (Chamberlin, 1978). Much of the relatively minor fault block topography (100-200 m) was already filled in by contemporaneous basaltic andesite lavas of the middle La Jara Peak tongue. Therefore the distribution of the <u>synrift</u> Lemitar Tuff was relatively unaffected by ongoing crustal extension. The common interpretation of widespread Oligocene ash-flow sheets as "pre-Basin and Range" (meaning pre-rift) presumes horsts and grabens to be the only style of crustal extension.

Since the onset of crustal extension, the segment of the Morenci lineament within the rift zone has been reactivated at depth as an incipient transform fault connecting en echelon axes of extension in the rift (Chapin and others, 1978). The surface expression of this transform structure is a diffuse transverse-oriented shear zone. The transverse shear zone separates two fields of domino-style fault blocks that have been rotated to the west on the north side and to the east on the south side. The transverse shear zone subtly cuts across the southern half of the Socorro Peak volcanic center. Rifting has apparently perpetuated or enhanced the original leaky vertical fabric of the lineament as evidenced by late Miocene lavas of the Socorro Peak Rhyolite and present day shallow magma bodies related to the Socorro geothermal area (pl. 3, map 4). The Capitan lineament apparently was not of a favorable orientation to be used like the Morenci lineament. Some relatively minor transverse faults in the Socorro Peak area appear to follow the Capitan trend.

The present basins and ranges of the rift in the Socorro area are relatively young features that have been superimposed on a broad early basin of the rift (Bruning, 1973; Chapin and Seager, 1975). In early Mocene, time the Popotosa basin formed as a wide, shallow sag across the entire rift from the Gallinas Mountains and Colorado Plateau on the northwest to the "high mesas" east of Socorro. Stratigraphic relations in the Santa Fe Group in the Socorro

Peak area provide new control on the south and eastern margins of the basin (fig. 35) and on the timing and nature of disruption of the Popotosa basin by creation of the present Chupadera-Socorro-Lemitar ranges. These ranges apparently existed as suballuvial fault blocks prior to late Miocene time. They became prominent topographic features in Pliocene and Pleistocene time primarily because of epeirogenic uplift of the southern Rocky Mountains combined with continued block faulting. Development of alpine terrane in southwestern Colorado at this time provided additional runoff to form the ancestral Rio Grande (Chapin, 1979a).

Local Structure

Several regional structures form a broad zone of intersection in the Socorro Peak area that has been instrumental in creating this center of repeated magmatism. These structures are the Morenci lineament, the Rio Grande rift and the Socorro cauldron. The Morenci lineament of probable Precambrian ancestry is now reflected in the overlying Cenozoic strata as a transverse shear zone of the rift. The general aspects of these regional Cenozoic structures in the Socorro Peak area are described hereafter. Relatively limited expressions of prevolcanic structures in the Socorro Peak area have been described with the regional setting. The stages of development of Neogene rift basins in the Socorro Peak area are reviewed in the "Cenozoic Geologic History" section, which summarizes this report.

Socorro cauldron and related structures

The Socorro cauldron (fig. 35) has many aspects common to well-defined calderas (cauldrons) of the San Juan field and other volcanic fields. Common attributes of cauldrons and calderas are: 1) a very thick, densely welded, "dirty" intra-caldera tuff; 2) an intravolcanic unconformity at the caldera wall; 3) post-collapse, sedimentary and locally erupted volcanic fill (especially resurgent calderas); and 4) a circular or elliptical geometry, with diameters of about 10-40 km (see Steven and Lipman, 1976; Lipman, 1975; Byers and others, 1976; Smith and Bailey, 1968). The thick and "dirty" nature of cauldron-facies tuffs results from contemporaneous collapse and avalanching or slumping of the caldera walls to produce mesobreccias and megabreccias in the terminology of Lipman (1976). Lag-fall breccias (Wright and Walker, 1977) related to local ignimbrite vents may also be common in cauldron-facies tuffs (Lipman, 1978, oral commun.). All of the above aspects of cauldrons have been observed in fault-block remnants of the Socorro cauldron and provide the main evidence for its existence. Except for some arcuate drainages of questionable interpretation in the eastern Magdalena Mountains, the Socorro cauldron (caldera) presently has no obvious physiographic expression.

The Socorro cauldron, as shown on Figure 35, is interpreted as a complex Oligocene collapse structure tentatively correlated with eruption of the Lemitar Tuff about 28 m.y. ago (Chapin and others, 1978). The Socorro cauldron is tentatively interpreted here as a complex combination of a trapdoor and resurgent cauldron, which may have formed during two stages of eruption and collapse. It may initially have formed as a trapdoor structure hinged on the west with maximum subsidence on the east. Then possibly during late stage eruption of the uppermost quartz-rich Lemitar Tuff -- which lies above a slight cooling break on the outflow sheet -- the eastern half of the cauldron subsided differentially along a pre-existing north-trending fault zone. Only the eastern half of the cauldron resurged and developed a semicircular moat. This tenuous history is based on all presently available data, but is clouded by ambiguity due to uncertainty in correlation of the cauldron-facies tuff.

The Socorro cauldron is clearly fragmented by block faulting of the Rio Grande rift, which requires its reconstruction based on remnants exposed in the Chupadera, Socorro and eastern Magdalena mountains. The western half of the Socorro cauldron overlaps and buries the Sawmill Canyon cauldron, which is considered to be the source of the pinnacles (?) member of the A-L Peak Tuff erupted about 31-32 m.y. ago. Mapping of the western remnant of the Socorro cauldron in the Magdalena Mountains (Osburn, 1978; Petty, 1979) has indicated relatively minor subsidence of about 150 m contemporaneous with eruption of the Lemitar Tuff. For this reason, these authors have interpreted the Socorro cauldron as a trapdoor structure, hinged on the west, with maximum subsidence on the east. The cauldron margin in the eastern Magdalena Mountains is expressed by an arcuate zone of post-Lemitar rhyolite intrusions, hydrothermal alteration, and local anomalous (overturned) dips in the Lemitar Tuff. Minor rhyolite domes between the Lemitar Tuff and the tuff of South Canyon represent the only reasonable candidates for post-collapse fill of the Socorro cauldron in the Magdalena Mountains. Post 26 m.y. subsidence and eruption of rhyolite domes about 20 m.y. ago in the Water Canyon Mesa area (Osburn, 1978) may be related to reactivation of the western Socorro cauldron margin by rifting and intrusion along the developing transverse shear zone.

Structurally high blocks of Precambrian and Paleozoic rocks at Socorro Peak, west of Water Canyon, and in the Coyote Hills (see fig. 2) mark the maximum possible outer limits of the collapsed block of Socorro cauldron. The southern portion of the Socorro cauldron block, which lies in the central and southern Chupadera Mountains, has not yet been mapped as part of the "Magdalena Project". The geologic map by Miesch (1956) of this area suggests the general outline of a resurgent block (his massive rhyolite) and cauldron-fill deposits dipping southeasterly off the resurgent block. The location of the southern margin of the Socorro cauldron shown on Figure 35 is uncertain.

Structural relationships of the eastern half of the Socorro cauldron, which lies east of the buried Sawmill Canyon cauldron margin, are notably different from that of the western half. The eastern half clearly has the elements of a resurgent cauldron, which is anomalous since most trapdoor cauldrons do not resurge (Steven and Lipman, 1976, p. 31). However, some aspects of the resurgent eastern half of the Socorro cauldron are ambiguous because the correlation of the cauldron-facies tuff here is uncertain.

The eastern cauldron-facies tuff has been tentatively correlated with the uppermost quartz-rich Lemitar Tuff as defined on the outflow sheet in the Lemitar Mountains (fig. 13). Alternatively this cauldron facies Lemitar(?) Tuff could be equivalent to the petrographically similar Hells Mesa Tuff (see "Lemitar Tuff - cauldron facies" for details). For the purpose of this report, structures of the Socorro cauldron in the mapped area have been interpreted on the assumption that both the western and eastern cauldron-facies tuffs are Lemitar Tuff. Should mapping of the southern Chupadera Mountains prove the alternative to be true, sufficient data have been provided here to make the relatively minor revisions that would be necessary.

Several relationships do support the Socorro cauldron as the source of the Lemitar Tuff. The distribution and thickness of the Lemitar ash-flow sheet (fig. 14) are in good agreement with a source in the Socorro cauldron. Also, a peripheral hingeline structure in the southern Lemitar Mountains appropriately indicates pre-Lemitar tumescence and post-Lemitar subsidence. Finally, a 28.6±1.1 m.y. K-Ar age for the Luis Lopez Formation (cauldron fill), which clearly overlies cauldron-facies Lemitar(?) Tuff, is in reasonable agreement with reliable K-Ar ages of 28.8±0.7 m.y. and 27.0±1.1 m.y. for the outflow-facies Lemitar Tuff.

Faults and other structures related to the Socorro cauldron depicted on the geologic map and cross sections (pl. 1 and 2) may be distinguished in several ways. Cauldron faults commonly contain intrusions petrographically similar to (and assigned to) caldera-fill members of the Luis Lopez Formation. Also they are truncated by an early Miocene erosion surface at the base of the Popotosa Formation. Therefore, caldera structures are also consistently offset or truncated by rift faults; all faults that cut Santa Fe Group strata are rift faults. Ring fracture and peripheral faults of the cauldron have been indicated on the map and cross sections by an extra heavy line.

Eruption of the Lemitar Tuff and subsidence of the Socorro cauldron apparently occurred concurrently with early regional crustal extension (Chamberlin, 1976, 1978). Thus, some north-trending Oligocene age faults within the cauldron may be early rift faults utilized during cauldron subsidence. However, in this case they would no longer be "true" rift faults, since transmission of regional extensional stress to a floating cauldron block seems unlikely.

Structural elements of the Socorro cauldron and the buried margin of the Sawmill Canyon cauldron in the mapped area (pl. 1) are illustrated in Figure 37. These cauldron-related structures are exposed in variably tilted rift fault blocks of Oligocene rocks in three areas: 1) the northern Chupadera Mountains, 2) the eastern escarpment of Socorro Peak, and 3) in the southern Lemitar Mountains. To appreciate the original attitude and sense of motion on these

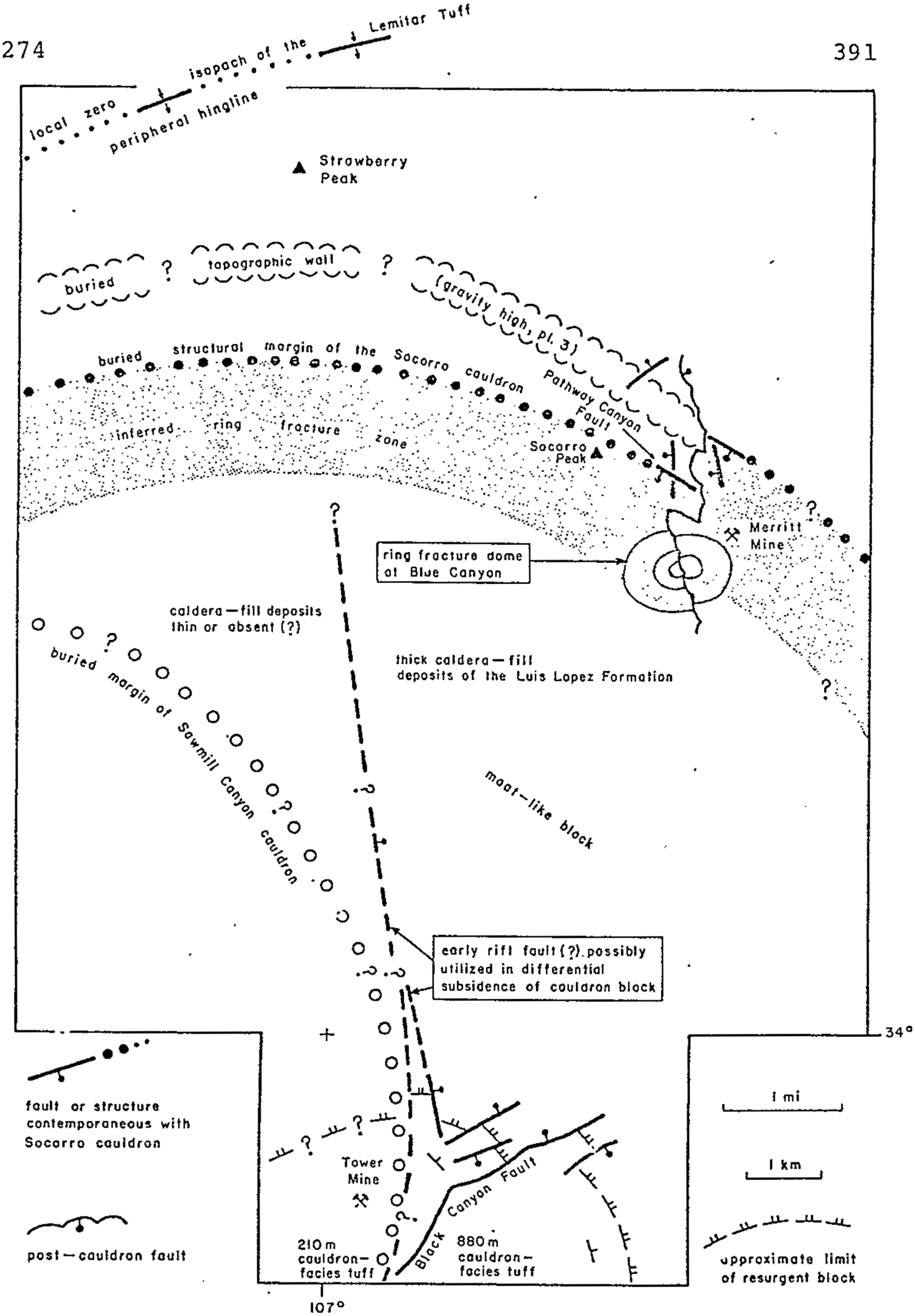


Figure 37. Structural elements of the Socorro cauldron and related structures within the Socorro Peak volcanic center.

Oligocene structures the effects of younger fault-block rotation must be subtracted.

Resurgent Block. A remnant of the resurgent block (dome?) of the Socorro cauldron is expressed by a large outcrop area of cauldron-facies Lemitar(?) Tuff mostly to the south of Black Canyon in the Chupadera Mountains. The cauldron facies interpretation is supported by the great thickness -- as much as 880 m -- of the densely welded tuff, and the presence of both cauldron-collapse mesobreccias (Tlx) and lag-fall breccias (upper Tlu) as illustrated in section I-I' of Plate 2 (see "Lemitar Tuff - cauldron facies" for details). The northeast side of the resurgent block is partly bound by several northeast-trending normal faults that have been rotated about 40 degrees to the east giving them the appearance of oblique slip faults. The northeast strike of these caldera related faults has not been significantly altered by later block rotation. Their trend suggests that they are an expression of pre-existing basement fractures of the Morenci lineament within the cauldron block. The Black Canyon fault, which is downthrown toward the adjacent moat deposits, is intruded by a large rhyolite body and exhibits unconformable relationships indicating that it was partly buried by moat deposits of the Luis Lopez Formation.

Some doming(?) (northeast tilting) of the resurgent block prior to filling of the moat-like area is suggested in cross section H-H' (near middle) where cauldron-fill strata apparently wedge out onto the resurgent area. Most of the present easterly dip here is attributed to rotation of rift fault blocks which occurred after filling of the moat and crystallization of the underlying magma body. A horst-block that extends from Chupadera Cliff to Socorro Canyon (pl. 1) has not been significantly tilted by rifting. Gentle to moderate northerly dips of Luis Lopez strata on this block most likely reflect resurgent doming.

The resurgent block of Figure 37 is bisected by a north-trending late Oligocene structure that may be interpreted in two ways. Evidence for this structure consists of a much reduced thickness of cauldron-facies tuff (210 m) intersected in the Tower mine drill hole; and the absence of cauldron-fill deposits of the Luis Lopez Formation on the Tower mine block (see pl. 2, H-H'). These relationships imply differential subsidence of the eastern segment of the cauldron block, both during eruption of the Lemitar Tuff and later during resurgence and filling of the moat-like area. Alternative explanations for this situation are: 1) the buried Sawmill Canyon cauldron block acted as a bouyant area that resisted subsidence, or 2) a preexisting north-trending rift fault zone was utilized as a locus of down-to-the-east subsidence.

Evidence for the latter possibility is found in a north-trending late Oligocene fault west of Chupadera Cliff. This fault cuts lower members of the Luis Lopez Formation but is overlapped by the upper lithic tuff member (Tlt_2) as shown in section G-G' (pl. 2). A similar down-to-the-east fault -- across which the Luis Lopez Formation apparently wedges out -- has been inferred on the geologic map less than 1 km west of the above-mentioned structure. Here the basal Popotosa Formation appears to lap across the inferred structure variably resting on Lemitar Tuff on the west side and on Luis Lopez Formation on the east side. The northward projection of this proposed early rift fault zone used by cauldron subsidence is shown in section A-A' (pl. 2) where its position is interpreted from a gravity profile (see pl. 3, map 2 for gravity map).

Evidence for the presence and influence of the eastern margin of the Sawmill Canyon cauldron comes mainly from the Tower Mine drill hole (fig. 12). This hole intersected a relatively thin interval of sandstones and conglomerates that are most likely correlative with the post-collapse fill of the Sawmill Canyon cauldron, (unit of Sixmile Canyon, Osburn, 1978). Extremely brecciated A-L Peak Tuff below these sandstones (pl. 2, H-H') may represent landslide blocks derived from the buried cauldron wall. These relationships suggest the east margin of the Sawmill Canyon cauldron lies just east of the Tower mine.

Mapping by Petty (1979) and Osburn (1979, oral commun.) has shown the southern structural margin of the Sawmill Canyon cauldron to project northeasterly toward the central Chupadera Mountains. The present expression of the eastern margin of the Sawmill Canyon cauldron, which was buried by the Lemitar Tuff, is most likely a north-trending, down-to-the-west, rift fault designated the Tower Mine fault (see fig. 39). The inferred relationships of this fault are shown in section H-H' (pl. 2) where it passes on the east side of the Tower Mine. In this interpretation, the original down-to-the-east sense of movement in late Oligocene time has been changed to a down-to-the-west sense, by rifting since Miocene time. Whatever its final interpretation may be, this north-trending structure that cuts the resurgent block of the Socorro cauldron is clearly of late Oligocene age and related to cauldron subsidence.

Collapse of the Socorro cauldron during eruption of the Lemitar Tuff combined with later resurgence formed a moat-like area between the northern Chupadera Mountains and the cauldron margin at Socorro Peak (fig. 17). This structural and topographic low filled during resurgence with as much as 800 m of locally erupted lavas and tuffs along with relatively minor landslide deposits and sediments shed from the caldera wall. These heterogeneous moat-fill deposits, collectively named the Luis Lopez Formation, were apparently restricted to a semicircular partial moat that flanks the resurgent block in the eastern half of the cauldron. Probable equivalents of the Luis Lopez Formation are present in the southern Chupadera Mountains, but appear to be absent or minor in volume where the western half of the cauldron is exposed in the eastern Magdalena Mountains.

Cauldron Margin. The northeastern structural-topographic margin of the Socorro cauldron is exposed in a complex of west-tilted fault blocks on the east face of Socorro Peak. The primary expression of the cauldron margin is a major intravolcanic unconformity where moat-fill deposits of the Luis Lopez Formation lap across the outer limit of the ring fracture zone (Pathway Canyon fault) onto a structurally high block of Pennsylvanian Madera Limestone. At the highest point on the cauldron margin (below Big Cliff) the entire Oligocene volcanic section consists of less than 200 m of moat volcanics. Outside the cauldron, about 10 km to the northwest in the Lemitar Mountains, the entire Oligocene volcanic section is more than 1000 m thick and is sandwiched between the same bounding formations as on Socorro Peak. Also, about 10 km to the southwest in the northern Chupadera Mountains, the minimum thickness of the Oligocene volcanic section is greater than 1400 m.

Before it was buried by volcanic units of the moat fill, the structurally high Madera Limestone block formed an early topographic wall (caldera rim) at the edge of the main ring fracture zone. Landslide deposits (Tlm) containing reworked boulders of the pre-cauldron Spears Formation lie against this topographic wall, thus demonstrating the <u>intra</u>volcanic nature of the unconformity. Later movement on the outer ring fracture -- Pathway Canyon Fault -- caused a greater thickness of volcanic moat-fill strata to be preserved on the downthrown south side of the fault. This may be interpreted as late-stage subsidence along the ring fracture zone during moat filling or continued uplift of the caldera rim block on the north side. Pennsylvanian strata on Socorro Peak have a northerly strike in comparison to the north-northwest strike of overlying moat-fill strata, which could be explained by northward tilting of the cauldron margin block prior to moat filling. These cauldron margin relationships are illustrated in cross sections A-A' and B-B' of Plate 2. Some of the more visible aspects of the cauldron margin are also shown in Figure 38.

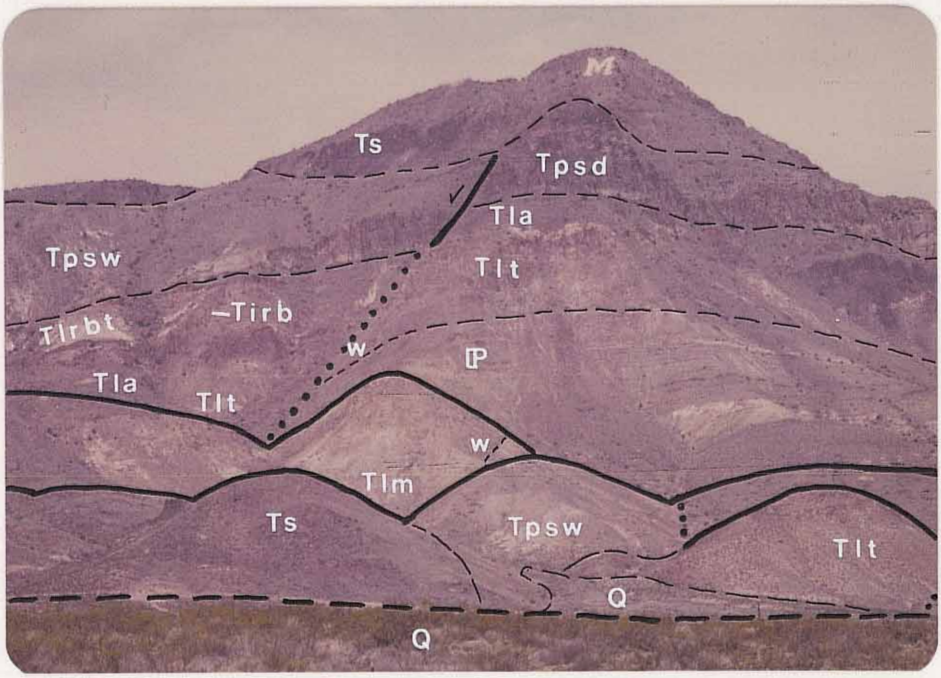


Figure 38. East face of Socorro Peak showing cross-section-like view of northeastern margin of the Socorro cauldron. West-tilted rift fault blocks (heavy lines) and range-bounding fault (dashed heavy line) have displaced the structural margin (dotted line) and a buried inner topographic wall (w) of the Socorro cauldron. Oligocene moat-fill deposits of the Luis Lopez Formation (Tl) thicken aburptly to the south where the cauldron margin truncates Pennsylvanian strata (P). Ancient landslide deposits (Tlm) shed from buried topographic wall and bedded tuffs (Tlrbt) are absent north of margin. Lithic tuffs (Tlt) and intermediate lavas (Tla) are present on both sides of cauldron margin. Rhyolite feeder dike (Tirb) marks trend of ring fractures. Early Miocene lower Popotosa conglomerates (Tpsd, Tpsw) unconformably overlie moat-fill deposits and are capped by late Miocene lavas of the Socorro Peak Rhyolite (Ts); apparent hump below "M" is due to perspective. Minor offset of Popotosa Formation along cauldron margin (Pathway Canyon fault) indicates local reactivation as a transverse rift fault. Several lesser faults and some Quaternary deposits (Q) are not delineated.

The exposures on Socorro Peak provide little more than a control point for the cauldron margin. Exposures of west-northwest trending ring-fractures, mostly intruded by Oligocene rhyolite dikes (Tlrb), are limited to about one km^2 in the Pathway Canyon area. These exposed ring fractures lie somewhere between the main ring fracture of the cauldron and the outer topographic wall of the caldera. The main ring fracture is probably marked by the Blue Canyon dome, which is the largest exposed rhyolite dome of the moat-fill sequence. The outer topographic wall of the caldera lies buried under younger rocks of the Santa Fe Group somewhere to the north of Socorro Peak.

The location and trend of the cauldron margin is supported by a gravity high that trends west-northwest from Socorro Peak (see pl. 3, map 2; fig. 37). The gravity map also indicates truncation of the cauldron margin by a major north-trending gravity gradient that represents the range-bounding fault zone and the west side of the Socorro basin. The decrease in this north-south gravity gradient going from Socorro Peak to the latitude of Blue Canyon is a reflection of the lower density moat deposits and greater depth to basement rocks within the cauldron. In the Socorro basin, the gravity map shows a broad saddle-like high in the basin floor to the northeast of Socorro Peak. This may be a reflection of the deeply buried cauldron margin offset to the north by the range-bounding Socorro fault zone. This 1-2 km (?) offset of the cauldron margin would be considerably greater than the visible offset by the Woods Tunnel fault zone (fig. 37).

The ring fracture faults on Socorro Peak are mostly poorly exposed, but in general they appear to be steeply dipping west-northwest-striking fractures consistently downthrown to the southwest. Shear fractures dipping about 78° southwest were observed in a small adit driven near the top of a talus chute that covers most of the Pathway Canyon fault trace. Upslope from here a vertical barite vein cutting lower Popotosa conglomerates is thought to represent an extension of the original Pathway Canyon fault rejuvenated by rifting. Smith (1963) reported the Pathway Canyon fault (his section 9 fault) to strike N45W and dip 60 degrees NE, which gives it the attitude of a high-angle reverse fault. However, he did not indicate where this observation was made. In any case, with as much as 40 degrees of post-Oligocene westerly rotation in the Pathway Canyon area, the original steeply dipping normal (?) faults of the ring fracture zone may reasonably be expected to now appear as near vertical faults or high-angle reverse faults (see pl. 2, A-A', C-C').

The minimum stratigraphic throw on the Pathway Canyon fault attributable to cauldron subsidence is about 300 m. The maximum displacement along the ring fracture zone southwest of Socorro Peak could be as much as 1500 m. This estimate is a conservative sum of the maximum thickness of cauldron-facies tuff and moat-fill deposits.

Some minor north-trending scissors faults also related to the Socorro cauldron crop out in the Pathway Canyon area. They are consistently downthrown to the west and invaded by Oligocene rhyolite dikes (Tirb). Below Big Cliff one of these north-trending dike and scissors fault combinations dips 85 degrees west and apparently joins with a ring fracture dike further south. This scissors fault dies out a short distance north of the cauldron margin. As interpreted in section B-B' (pl. 2), the original updip continuation of the scissors fault and dike below Big Cliff now occurs in a downfaulted block east of the Woods Tunnel fault zone as a truncated rootless dike.

Peripheral Hingeline. In the southern Lemitar Mountains near Bug Mountain, west-tilted Oligocene strata contain an east-northeast-trending fault zone and a related drag flexure (fold), which may reasonably be interpreted as a hingeline structure peripheral to the Socorro cauldron. Stratigraphic relationships across this structure indicate the following sequence of events: 1) initial uplift on the south side of the hingeline -- interpretable as pre-eruption tumescence (Smith and Bailey, 1968) -- shortly before eruption of the Lemitar Tuff, 2) relatively rapid subsidence on the south side of the hingeline after eruption of the Lemitar Tuff and development of a tight drag fold in the ash-flow sheet, 3) erosion of the Lemitar Tuff from the high north side of the hingeline, 4) continued but slower subsidence on the south side of the hingeline during deposition of late Oligocene volcanic strata (Tba_2, Tsc, pl. 1), 5) burial of the hingeline structure in early Miocene time by lower Popotosa conglomerates but with some gentle southward subsidence continuing into early Miocene time (see fig. 24). Essentially all of the stratigraphic relationships that define this kinematic development of the hingeline are apparent on the geologic map (pl. 1) when viewed looking downstructure to the west-southwest.

The main fault of the hingeline is locally intruded by a basaltic-andesite dike (Tiba) of probable late Oligocene age. The peripheral faults are not well exposed but appear to be nearly vertical. About 40-50 degrees westerly rotation of the Oligocene tuffs in this area (see Lemitar Map) has had essentially no effect on the strike and dip of the originally steep hingeline faults, which strike approximately perpendicular to the axis of rotation. However, the associated drag fold axis now clearly plunges to the southwest. A similar, but more subdued, hingeline structure is exposed in the eastern part of the southern Lemitar Mountains (Lemitar Map). Together, these control points do not make an arcuate trend paralleling the inferred ring fracture zone. This hingeline structure, evidently controlled by changing magmatic pressures, may have formed along a basement fracture of the Morenci trend rather than following the ring fracture trend. Younger transverse rift faults (Lemitar fault zone, fig. 39) also appear to be influenced by this inferred basement fracture.

Rift Structures

The Socorro Peak volcanic center is dominated by late Oligocene to late Pleistocene normal faults of the Rio Grande rift (Table 8), which have broken the area into a complex combination of tilted blocks and subsidiary horst-graben blocks (fig. 39). All faults and associated structures that deform early Miocene and younger Santa Fe Group strata are considered to be related to the rift. The recognition of early rift faults of late Oligocene age is less certain in the Socorro Peak area because of contemporaneous development of the Socorro cauldron. An angular uncon-

Table 8. Ages and evidence for middle and late Cenozoic rift faulting in the Socorro Peak volcanic center.

Age	Evidence
late Pleistocene	3m scarp cutting terrace gravels (Qoa) in Socorro Canyon at Socorro fault zone
middle Pleistocene to late Pleistocene	30m offset of broad geomorphic surface (Qoa) south of Socorro Canyon at Socorro fault zone
middle Pliocene	basalt of Sedillo Hill (Tbsh) variably overlies faulted facies of Sierra Ladrones Formation (Tslf, Tslp) north of Socorro Canyon at Socorro fault zone (pl. 2, F-F';fig. 33)
late Miocene to early Pliocene	10-15 degree angular unconformity between Sierra Ladrones and upper Popotosa formations; 11.9-m.y.-old tuff bed (Tsdt) truncated by ancestral Rio Grande overbank deposits (Tslo) north of Nogal Arroyo
late Miocene	lava flow of the Socorro Peak Rhyolite (Tsd) variably rests on lower Popotosa Formation (Tpsd) and tuff of South Canyon (Tsc) on opposite sides of West Chupadera fault in Tower mine area
early Miocene to middle Miocene	indirectly indicated by change in source areas of lower and upper Popotosa Formation (cf. fig. 24, fig. 26)
early Miocene	colluvial deposits in basal Popotosa Formation (Tpsd, Tpsw) near Tower mine and west of Merritt mine (fig. 24)
late Oligocene to early Miocene	10-30 degree angular unconformity at base of lower Popotosa Formation
late Oligocene	colluvial breccia (Tlrsx) preserved in transverse graben and overlain by tuff of South Canyon (Tsc), northeast of Tower mine (pl. 2, G-G')

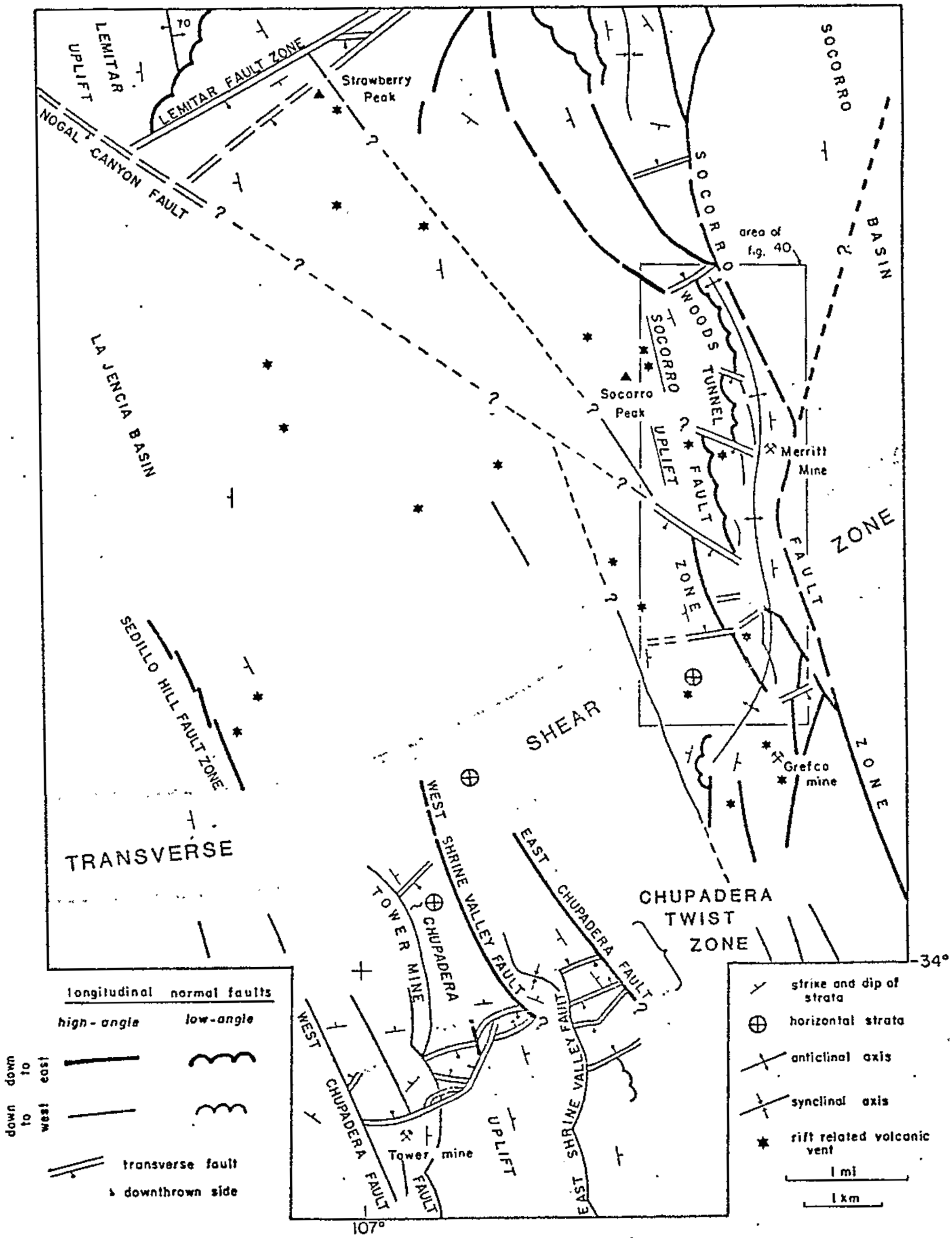


Figure 39. Structural elements related to the Rio Grande rift within the Socorro Peak volcanic center. Several minor and indefinite faults of Plate 1 are not shown.

formity below the Lemitar Tuff outflow sheet indicates that the Socorro cauldron formed after the onset of rifting; thus there need be no significant hiatus between subsidence and resurgence, and the initiation of rifting within the cauldron block. Locally rifting appears to have reactivated (perpetuated?) cauldron faults with the same sense of movement.

The oldest fault block reasonably attributed to rifting within the Socorro cauldron is a late Oligocene transverse graben (Table 8), which is part of the Chupadera twist zone (fig. 39). The late Oligocene graben exhibits a differential rotation of strata across it similar to other portions of the twist zone that locally deform the Popotosa Formation. Thus the twist zone and its elements are considered to be a long-lived structure related to rifting.

The strongest evidence of late Oligocene domino-style rifting is a widespread 10-30 degree angular unconformity at the base of the early Miocene lower Popotosa Formation. The area of this unconformity far exceeds the limits of cauldron subsidence (Lemitar Map). Observations listed in Table 8, and many others included with previous descriptions of Santa Fe Group formations, indicate that crustal extension has been a progressive (essentially continuous) process since late Oligocene time.

Like the Socorro fault zone, which has been active more than 4 million years, most rift faults are probably long-lived growth faults that may span the time of deposition of several stratigraphic units (pl. 2, F-F'). Thus, the thickness of strata may vary abruptly across rift faults; which in turn often makes estimates of the throw on rift faults uncertain. In addition, estimates of displacements on relatively young range-bounding faults may be greatly exaggerated because of their overprinting on early-rift low-angle normal faults. Where this occurs, the apparent stratigraphic throw on the younger fault is actually the cummulative effect of movement on both the older and younger faults (see right side of fig. 36).

For the purpose of their description, rift faults and related structures in the map area are divisable into two categories: 1) longitudinal structures that strike between north-northwest to north-northeast, and 2) transverse structures that strike between northeast to southeast. The development of these contemporaneous longitudinal and transverse structures also controlled the local evolution of past and present-rift basins (see summary in Cenozoic Geologic History)

Longitudinal Normal Faulting. Many aspects of longitudinal fault blocks in the map area either support, or may be explained, by the Morton and Black model of rifting (fig. 36), referred to here as domino-style normal faulting. However, the interpretation of progressive rotation of longitudinal blocks is complicated by the presence of the transverse shear zone, which separates blocks tilted in opposing directions. The transverse shear zone of Chapin and others (1978) as shown on Figure 39 is considered here to be the axis of a diffuse belt (or en echelon belt ?) of discontinuous transverse faults about 6-8 km wide. The diffuse transverse shear zone may extend as far north as Socorro Peak and south to the Chupadera twist zone. As used here, the transverse shear zone breaks the map area into three domains of longitudinal blocks, with the shear zone as the central domain. In the north domain, longitudinal faults are dominantly downthrown to the east and repeat, in a domino style, strata that are tilted to the west (pl. 2, A-A'; see also pl. 3, map 1, A-A'). Down-to-the-west faults are dominant in the south domain and form a series of east-tilted "dominoes" (pl. 2, H-H', I-I'; see also pl. 3, map 1, B-B'). In the central domain, the blocks generally become less tilted and narrow horst-graben blocks appear to form by the interfingering of the opposing domino sets present to the north and south. Thus, the north end of the Chupadera uplift (fig. 39) and the south end of the Socorro uplift (pl. 2, E-E', F-F') appear to be horst blocks; although in the latter area the horst-like block is superimposed on older west-tilted blocks. Away from the transverse shear zone, these same uplifts essentially represent the structurally high side of broad tilted blocks.

The progressive rotation of longitudinal blocks is most apparent in the north and south domains, where there is a general decrease in the dip of strata as they become younger. In these areas, Oligocene volcanic strata dip at about 40-60 degrees, in contrast to 20-30 degree dips in Miocene Popotosa Formation strata, and to 0-10 degree dips in Pliocene-Pleistocene strata of the Sierra Ladrones Formation. In the Lemitar Mountains (Lemitar Map), the 28-m.y.-old Lemitar Tuff typically dips 35-45 degrees west, whereas the 31-32 m.y. old A-L Peak Tuff dips at 50-60 degrees west. This late Oligocene angular unconformity records the onset of domino-style rifting in the Lemitar Mountains. When the Lemitar Tuff is rotated back to its original horizontal attitude, the down-to-the-east low-angle normal faults associated with the late Oligocene unconformity become early rift high-angle normal faults dipping about 70 degrees east (Chamberlin, 1978, fig. 1; or see Hawley, 1978, fig. S-42).

Low-angle normal faults, which dip 45 degrees or less where observed in the Socorro Peak center, consistently occur in moderately to strongly tilted strata of Miocene to Oligocene age, or locally in prevolcanic rocks. These low-angle faults typically form cuspate traces as they cross numerous small drainages. The cusps normally point toward the downthrown block. These low-angle normal faults, which dip east in the north domain and west in the south domain (fig. 39), may be generally interpreted as inactive, rotated early rift faults, even where definite age control is lacking, because of their stratigraphic association. One low-angle normal fault of late Oligocene-early Miocene age is inferred to be present west of the Grefco mine (pl. 2, E-E', F-F'). Here mudflow deposits of early Miocene age (Tpsd) unconformably lie on a gently east-dipping erosion surface that cuts at a high angle across compaction foliations in the underlying cauldron-facies tuff (Tlu). Since the cauldron-facies tuff and overlying andesite lavas of the Luis Lopez Formation (Tla_1) dip steeply to the west, this buried erosion surface may then be reasonably interpreted as an early rift domino fault that was buried in early Miocene time. Similar buried domino-style faults about 1 km apart are present in the southern Lemitar Mountains (pl. 3, map 1, A-A'). These observations suggest that many early-rift domino-style faults may be buried below the Popotosa Formation in the western Socorro Mountains. If domino-style faults were inferred about 1 km apart to the west of Socorro Peak then the amount of westward thickening shown in the Miocene Popotosa strata of cross section A-A' (pl. 2) could be much less and still accomodate a 15-20 degree angular unconformity between the Oligocene and Miocene strata.

In the northern domain of longitudinal rift faults, some minor rift faults are downthrown to the west making them "antithetic" to the master set of down-to-the-east domino faults. These "antithetic" faults commonly appear to have westerly dips greater than 80 degrees suggesting that they have been rotated with the domino faults. Several "overturned" antithetic faults, which now have attitudes of high-angle reverse faults, have been observed in steeply tilted Oligocene strata in the Lemitar Mountains (Lemitar Map). A similar high-angle reverse fault occurs in the Lemitar Tuff east of Bug Mountain. When the enclosing tuff is rotated back to its original horizontal position this seemingly anomalous "reverse" rift fault becomes an early rift antithetic normal fault dipping about 75 degrees to the west.

On the geologic map and cross sections, the West Shrine Valley fault and the East Chupadera fault have been projected south of the Chupadera twist zone as minor "antithetic" faults. As indicated by Figure 39, this interpretation is questionable. These projected antithetic faults (as shown in H-H' and I-I', pl. 2) may alternatively be interpreted as east-dipping stratigraphic contacts where upper Popotosa strata unconformably lap onto steeply dipping Oligocene volcanic strata. The writers reconnaissance has shown the eastern boundary of the southern Chupadera Mountains at Walnut Creek to be such an east-dipping unconformity. Thus, the Chupadera uplift has the appearance of a horst block only in the vicinity of the transverse shear zone and for the most part is essentially an east tilted fault block.

The major longitudinal structural blocks in the area are relatively young (late Miocene to Pleistocene) topographically defined features referred to on Figure 39 as the Socorro, Chupadera and Lemitar uplifts, and the Socorro and La Jencia basins. As a first order of approximation, these uplifts and basins respectively represent the structurally high and low sides of relatively wide domino blocks that are moderately tilted and bound by major high-angle normal faults or fault zones. The main uplift-bounding structures in the map area are the Socorro fault zone, the West Chupadera fault and the Lemitar fault zone. The apparent stratigraphic throw on these bounding-fault zones appears to vary greatly as they cut across older fault blocks. At the north boundary of the map area, the Lemitar fault zone has a stratigraphic throw of about 600 m. North of here, the fault zone bends to a northeast strike (pl. 3, map 1) and cuts across older west tilted blocks; this produces erratic and abrupt changes in the apparent stratigraphic throw. From south to north along 3 km of the northeast-trending segment of the Lemitar Fault zone, the apparent stratigraphic throws are: 1220 m, 1830 m, 2440 m, and 1980 m. However, there are no appropriate changes in the topographic expression of the range front to suggest that such changes are real. East of Socorro Peak, the Socorro fault zone appears to have a stratigraphic throw of several kilometers based on gravity and stratigraphic data. Much of this displacement is locally exaggerated by the effects of the Socorro cauldron margin and the Pathway Canyon fault zone, both of which predate the Socorro fault zone. The "true" Pliocene-Pleistocene displacement on the Socorro fault zone is uncertain, but it must be at least 330 m based on offset of Sierra Ladrones strata in the Nogal Arroyo area. Where locally exposed in arroyos and ravines, all faults cutting Pliocene and Pleistocene strata have been found to be high-angle normal faults dipping between 70 and 80 degrees.

Figure 39 shows a long anticlinal fold between the Woods Tunnel fault zone and the Socorro fault zone, which is interpreted as a drag flexure related to the Socorro fault zone. This drag flexure is expressed along the mountain front as a band of east-tilted (or less westerly tilted) strata. The flexure is most apparent in the Miocene strata south of Pathway Canyon. Drag flexures are most common in the relatively plastic strata of the upper Popotosa Formation, but have also been observed in normally brittle Oligocene ash-flow tuffs. The Sedillo Hill fault zone apparently cuts a thick section of relatively elastic upper Popotosa claystones, which may explain the en echelon character of this fault zone. Drag flexures often introduce a complicating factor in evaluating the amount of rotation of strata. Drag along domino-style faults tends to reduce the amount of apparent rotation.

Most longitudinal faults have been approximately located by juxtaposition of different stratigraphic units. Where actually exposed over small areas, the faults appear planar and may be expressed as breccia or gouge zones, closely spaced fractures, slickensided fractures, veins, and silicified breccia zones. Slickensides consistently have east-northeast bearings and are essentially dip slip with only minor components of oblique slip indicated. Minor components of oblique slip may be dextral or sinistral; no particular pattern of oblique-slip components has been recognized. High-angle faults usually have gently curved traces concave toward the downthrown side.

The down-dip geometry of normal faults formed by crustal extension is a controversial question. Profett's (1977) conclusion that they are gently curved surfaces concave upwards (toward the footwall block) is based on considerable subsurface control. Proffett estimated that normal faults in the Yerington, Nevada, area "flatten" at rates of 3 to 7 degrees per kilometer of depth. Similar values of decreasing dip have been obtained from first motion studies of microearthquakes in the Socorro area (A.R. Sanford, 1976, oral commun.). For the purpose of constructing cross sections, rift faults in the Socorro Peak area have been presumed to be gently curved toward the footwall block and flatten with depth. The writer has observed subhorizontal early rift faults in the Lemitar Mountains that locally appear to be significantly curved, platter or spoon shaped surfaces. Some of these may actually be composites of multiple fault planes with different dips (see Morton and Black, 1975, fig. 6, "stage 5"). Warping of one early rift fault plane by a drag flexure at the range bounding fault is also apparent in the Lemitar Mountains.

Many longitudinal faults in the map area appear to make sharp bends; sometimes to become transverse faults. Others seem to terminate on transverse rift faults, whereas others are deflected or offset by the transverse faults. The complex interaction of transverse and longitudinal faults is explained by the nature of the transverse shear zone described next.

Transverse shear zone and transverse faulting. In the Socorro Peak volcanic center, transverse rift faults (fig. 39) commonly parallel or follow reactivated elements of the Socorro cauldron, which in turn are believed to reflect pre-rift fractures of the Morenci and Capitan lineaments. North of Socorro Peak, several longitudinal rift faults make abrupt bends or gradually turn toward orientations parallel to the margin of the Socorro cauldron suggesting that they have locally utilized ring fractures and peripheral fractures of the cauldron. The Pathway Canyon ring fracture fault has been reactivated as a transverse rift fault as evidenced by a 60 m offset of the lower Popotosa Formation. Stratigraphic juxtaposition along transverse faults generally indicates that they are normal faults with short or discontinuous traces. On the east face of Socorro Peak, transverse faults parallel to the ring fracture trend, tend to deflect or offset longitudinal faults; whereas in the northern Chupadera Mountains many northeast-trending transverse faults appear to terminate on, or be offset by, longitudinal faults.

Observations of slickensides on transverse faults would at first suggest that they are mostly oblique slip faults with large components of lateral slip. However, the admittedly limited observations are also consistent with their interpretation as original dip-slip faults that have later been rotated. North of the transverse shear zone (Lemitar Map), the slickensides on transverse faults consistently plunge at small angles to the northeast and on the south side they plunge to the southwest. A transverse dike of Oligocene age (Tiap) at the Big Roadcut exhibits near vertical (dip slip) slickensides indicating some post Oligocene movement. This dike occurs in an essentially non-rotated block within the transverse shear zone.

The primary function of transverse rift faults is well illustrated by relationships along the Woods Tunnel fault zone, which are summarized in Figure 40. Although the transverse faults are not well exposed, outcrop control is sufficient to demonstrate several offsets (deflections?) of the main fault of the Woods Tunnel zone along transverse faults. Abrupt changes in dip of adjacent strata and in

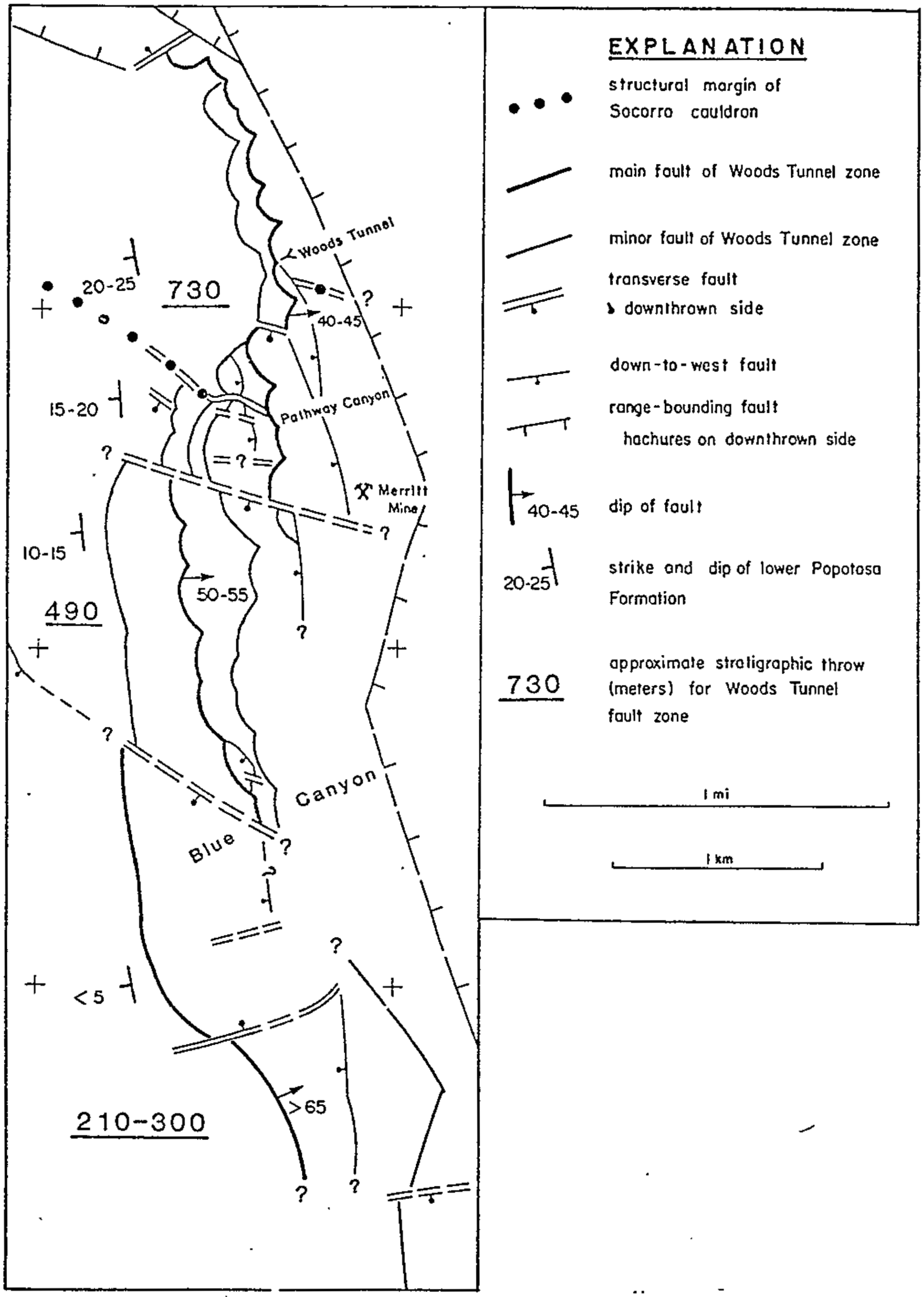


Figure 40. Interpretative map of the Woods Tunnel fault zone. Map illustrates accomodation of differential rotation of longitudual fault blocks by reactivated segments of ring fracture faults utilized as transverse rotary faults. Note southerly decrease in dip of strata is matched by approximately equivalent increase in dip of faults. Several deflections of the main Woods Tunnel fault and related minor faults occur along transverse faults.

dip of the longitudinal fault zone appear to occur between segments of the zone separated by transverse faults. Thus, the conclusion is that the transverse faults are accommodating differential rotation of the longitudinal blocks. As defined by Gary and others (1972), the transverse faults are acting as rotary or hinge faults, which permit rotation about an axis roughly perpendicular to the fault plane.

The incremental decrease in dip of strata going south from Woods Tunnel may be interpreted as the effect of a decreasing amount of crustal extension toward the axis of the transverse shear zone. Such an interpretation is in accordance with the Morton and Black (1975) model of crustal attenuation, even though they did not consider such variations with regard to transverse structures. An overall decrease in stratigraphic throw of the Woods Tunnel zone toward the south also supports this interpretation.

The accommodation of differential rotation of "domino" blocks is most apparent across northeast-striking transverse faults of the Chupadera twist zone (fig. 39). One of these transverse hinge faults at the northeast end of the zone is also a scissors fault that separates blocks tilted in opposing directions. Another element of the Chupadera twist zone is a synclinal fold in upper Popotosa strata in the Shrine Valley. This fold axis takes a transverse orientation where the East- and West- Shrine Valley faults die out, which suggests the fold may be related to a scissors like zone in the underlying fault blocks. However, away from this transverse segment, the fold may be readily interpreted as a drag flexure. The Chupadera twist zone appears to end on the southwest against the West Chupadera fault and may project to the northeast as far as the Socorro fault zone. Several longitudinal faults appear to die out or change their direction of displacement across the eastern end of Socorro Canyon, which lies along the projection of the twist zone. Both the Chupadera twist zone and the transverse hinge faults on Socorro Peak are local expressions of the transverse shear zone of Chapin and others (1978). They are considered here to mark the north and south limits of the shear zone whose axis is shown as a shaded band in figure 39.

Chapin and others (1978) recognized -- primarily from maps of the "Magdalena Project" -- that the rotation of domino style blocks in opposite directions occurs along a linear zone extending west-southwest from Socorro. We noted that this twisting motion in the surface rocks requires lateral shearing in the ductile lower lithosphere and therefore referred to this structure as the Socorro transverse shear zone (fig. 35, 39). The transverse shear zone occurs along the segment of the Morenci lineament that lies within the Rio Grande rift. From the onset of rifting, the vertical fabric of the lineament has been reactivated as a transform fault connecting en echelon axes of extension in the rift. Since the surface expression of this transform fault is <u>not</u> a major strike slip fault, there must be some decoupling (horizontal shearing) between the ductile and brittle zones of the crust. Diffuse continental transform structures, which contain elements of both extension and shearing (similar to the Socorro shear zone), have been observed in the Basin and Range province (Eaton, 1979) and in the Afar depression (Barberi and Varet, 1977). The first visual observations of a seafloor transform structure (valley) have shown it to be a complex transverse graben, which contains both longitudinal and transverse elements, flanking a narrow zone of lateral shearing in the deepest part of the valley (Choukroune and others, 1978). Thus, even in a seafloor spreading environment, transform faults are not necessarily simple strike-slip faults.

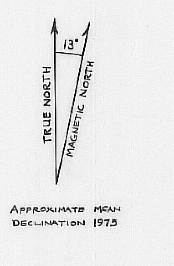
As defined by Chapin and others (1978) the surface expression of the Socorro transform structure is a "zone of jostled fault blocks 1.5 km or more in width across which the tilt of beds and sense of extension changes markedly". Where the transverse shear zone crosses the Socorro Peak volcanic center it appears to be an 8-km-wide zone of transverse hinge faults, and closely spaced horsts and grabens formed by intermeshing of domino-style faults. The most visible aspect of this diffuse zone is the change in dip directions of strata across it, as illustrated in Figure 41. Dips of strata generally decrease toward the axis of the shear zone, which is locally marked by a null line of mesas.

The transverse shear zone has apparently tended to be structurally and topographically high, probably because it is a band of less extended crust relative to areas to the north and south. This relationship is evident by shallowing of the Mulligan Gulch graben and the La Jencia basin southward against the transverse structure. The southeastern margin of the Popotosa basin in Miocene time (fig. 36) was most likely a manifestation of the transverse shear zone.

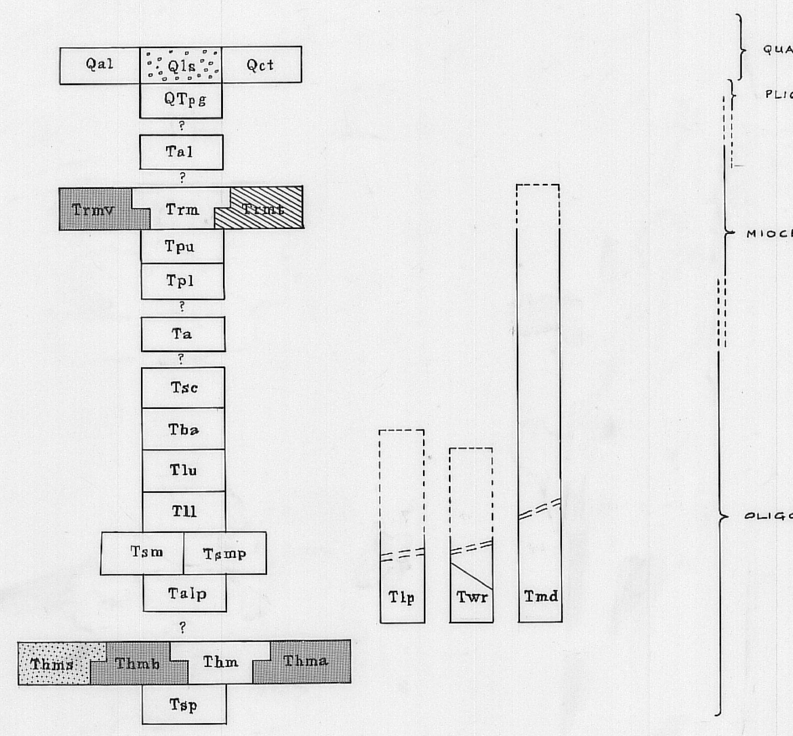
The primary control of magmatism at the Socorro Peak volcanic center has been the Precambrian(?) Morenci lineament, which is now expressed in the transverse shear zone of the rift. The Socorro cauldron represents only one of several Oligocene plutons intruded along the leaky vertical grain of the lineament (fig. 36). The late Miocene volcanic vents of the western Magdalena Mountains, the Pound Ranch area and the Socorro Peak area are distributed along the

GEOLOGIC MAP AND SECTIONS OF THE WEST FLANK OF THE MAGDALENA MOUNTAINS
SOUTH OF THE KELLY MINING DISTRICT
SOCORRO COUNTY, NEW MEXICO

by
Philip Allen
1979



CORRELATION OF MAP UNITS



DESCRIPTION OF MAP UNITS

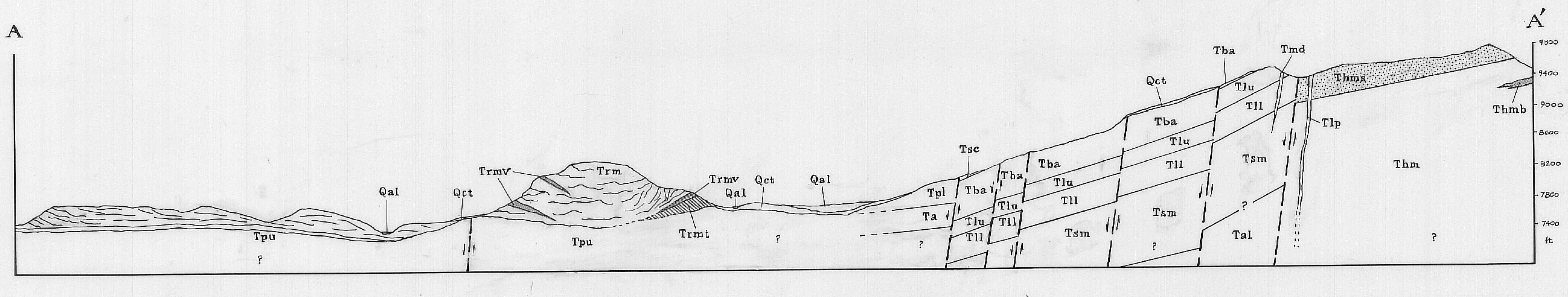
EXTRUSIVE AND SEDIMENTARY ROCKS

Qal	ALLUVIUM
Qls	LANDSLIDES
Qct	COLLUVIUM – TALUS
QTpg	PIEDMONT GRAVELS and older valley fill
Tal	ALLUVIUM
Trm	RHYOLITE OF MAGDALENA PEAK — Flow banded where foliation trends shown, undivided elsewhere. Locally includes: Trmv and Trmt
Trmv	Vitrophyre and perlite
Trmt	Tuffs
Tpu	POPOTOSA FORMATION – Upper member
Tpl	POPOTOSA FORMATION – Lower member
Ta	ANDESITE
Tsc	TUFF OF SOUTH CANYON
Tba	BASALTIC ANDESITE AND HORNBLENDE ANDESITE
Tlu	TUFF OF LEMITAR MOUNTAINS — Upper member
Tll	TUFF OF LEMITAR MOUNTAINS — Lower member
Tsm	UNIT OF SIXMILE CANYON — Locally includes: Tsmp
Tsmp	Latite porphyry
Talp	A-L PEAK TUFF – Flow-banded member
Thm	HELLS MESA TUFF — Locally includes: Thms, Thmb, and Thma
Thms	Sandstones and conglomerates
Thmb	Laharic Breccias
Thma	Andesites
Tsp	SPEARS FORMATION

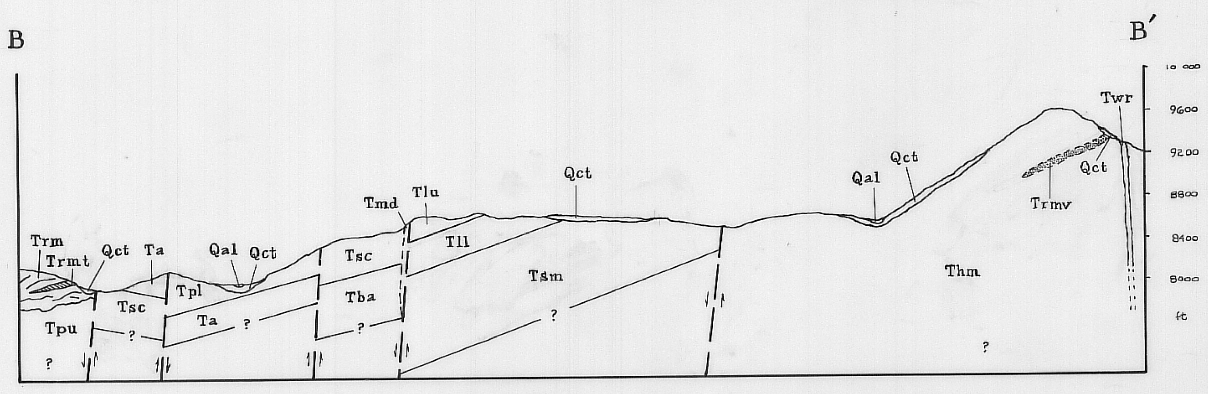
INTRUSIVE ROCKS

Tlp	LATITE PORPHYRY OF MISTLETOE GULCH
Twr	WHITE RHYOLITE DIKES AND DOME
Tmd	MAFIC DIKES

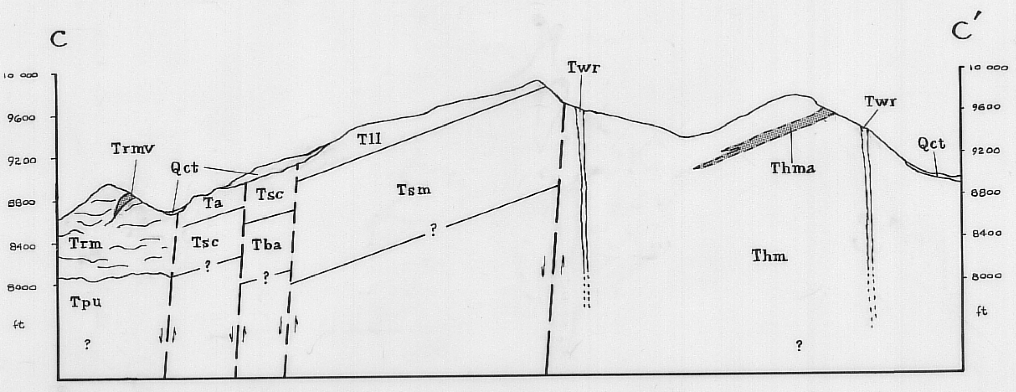
A A′



B B′



C C′



SYMBOLS

	CONTACT — Dashed where approximately located
	FAULT — Dashed where approximately, dotted where concealed, ball on downthrown block
20	STRIKE AND DIP OF BEDS
50, vertical	STRIKE AND DIP OF FLOW LAYERING
	TRANSPORT DIRECTION
	VOLCANIC VENT
	SILICIFIED ZONE
75	VEIN — Arrow shows inclination
X	PROSPECT
	MINE — Includes adits, shafts, and large open pits

transverse shear zone. Magmas are again bleeding upwards along the lineament at Socorro to provide the heat source for the geothermal area. As shown on Plate 3 map 4; the deep magma body generally ends against the shear zone and the shallow dike-like bodies are grouped in an overall northeast-trending array along it. The north trend of these dikes is consistent with brittle extension of the upper crust in a general east-west direction.

Reilinger and Oliver (1976) have demonstrated modern uplift associated with the deep magma body at Socorro. Since magmatic inflation and doming is common in active volcanic areas, the question arises: does the cluster of shallow magma bodies around Socorro Peak have any surficial structural expression? The writer offers a tentative yes. Several anomalous and apparently unrelated features when taken together suggest the rough outline of an elliptical neomagmatic uplift extending west-southwest from Socorro Peak. The axis of the possible uplift crosses the headwaters of Socorro Canyon and extends into the east side of the La Jencia basin. If it exists, the magmatic uplift appears to be largely truncated by the Socorro fault zone, although confluent drainages in the basin due east of Socorro Peak may also be part of the uplift pattern.

The possible magmatic uplift is defined by three elements: 1) variations in elevation of the base of the Socorro Peak Rhyolite lava flows, 2) elevation changes at the base of the basalt of Sedillo Hill in the area west of Socorro Canyon, and 3) drainage patterns in the La Jencia and Socorro basins. Flows of the rhyodacite and quartz latite units of the Socorro Peak Rhyolite form a north-northwest-trending belt from 6001 Mesa to Strawberry Peak (fig. 29). The elevations of these separate flows decreases abruptly to the north and south away from Socorro Peak. From Nogal Canyon to Strawberry Peak the elevation increases; this would place Nogal Canyon along a synclinal trough. In the western Socorro Mountains, the bases of flows of the phenocryst-rich rhyolite appear to slope away from vents at 6633 Peak and Tripod Peak, although the decrease in elevation to the south here is minor.

The basalt of Sedillo Hill forms a belt of north-northwest-striking cuestas to the north of Sedillo Hill. This outcrop belt anomalously increases in elevation to the north <u>away</u> from the vent area. The highest point on the basalt flow occurs west of the headwaters of Socorro Canyon. This general trend of structurally(?) high points from Socorro Peak to the headwaters of Socorro Canyon appears to extend into the La Jencia and Socorro basins as topographically high drainage divides (see Quaternary surficial deposits).

Taken together these elements suggest recent (or progressive?) uplift along this trend. Several aspects of these observations are problematical. Most importantly the axis of the hypothetical uplift lies significantly north of the locus of shallow intrusions (pl. 3). It is quite possible that the above observations represent a fortuitous combination of unrelated structures. However, the hypothesis of neomagmatic uplift may be worth testing by independent means, such as a tiltmeter survey.

CENOZOIC GEOLOGIC HISTORY

The major conclusions of this investigation may be summarized in the form of a geologic history. The Cenozoic history of the Socorro Peak Volcanic center has been greatly influenced by a preexisting regional structural grain. Two major crustal flaws -- the northeast-trending Morenci lineament and the west-northwest-trending Capitan lineament of Chapin and others (1978, fig. 1) -- intersect near Socorro. These deeply penetrating flaws, which are partly defined by their tendency to "leak" magmas, appear on regional maps as linear belts of Cenozoic intrusions and volcanic vents cutting across central New Mexico. The Morenci and Capitan lineaments respectively parallel the Precambrian wrench fault zones of the Colorado lineament (Warner, 1978) and the Texas lineament (Albritton and Smith, 1956), suggesting that they may be of similar origin.

The northerly structural grain of the Socorro area was developed (or reactivated?) during periods of regional compression in late Mississippian to Permian time (ancestral Rocky Mountains orogeny) and in late Cretaceous to middle Eocene time (Laramide orogeny) (Chapin and Seager, 1975).

Socorro Peak lies on a broad Laramide uplift about 15 km from its eastern flank, which is expressed by a north-northeast-trending high-angle reverse fault (Wilpolt and Wanek, 1951). By late Eocene time, this Laramide uplift was stripped of a thick sequence of Mesozoic strata and beveled by an erosion surface of low relief (Chapin and others, 1975; Epis and Chapin, 1975). The massive Pennsylvanian Madera Limestone apparently formed a resistant caprock on the Laramide uplift in the Socorro area. These limestones most likely now form the top of the prevolcanic basement throughout the subsurface of the volcanic center.

Cenozoic strata in the mapped area provide an essentially complete geologic record since middle Oligocene time. However, the middle and late Cenozoic sections are also broken by numerous local erosional and angular unconformities developed across contemporaneous fault blocks related to both the late(?) Oligocene Socorro cauldron and the Rio Grande rift. Regional stratigraphic data of the New Mexico Bureau of Mines and Mineral Resources "Magdalena Project" have filled in some of these gaps in the record.

In early Oligocene time, great volumes of andesite to quartz latite lavas began to erupt from central volcanoes near the heart of the Datil-Mogollon field (Chapin and others, 1975). Penecontemporaneous erosion of these volcanic highlands produced a thick volcaniclastic and alluvial apron that spread northeastward across the Socorro area during the period about 37-32 m.y. ago. These mudflow deposits, conglomerates and sandstones of intermediate volcanic detritus, are now assigned to the Spears Formation. The Spears Formation is locally absent on Socorro Peak, where the base of the Oligocene section is exposed, because of a late(?) Oligocene erosional unconformity at the north margin of the Socorro cauldron (caldera). To the north and south of the caldera margin, the Spears Formation should be present in the subsurface of the map area.

About 32-33 m.y. ago major volcanic activity shifted to voluminous eruptions of silicic ash-flow tuffs, which continued intermittently until 26 m.y. ago. During this period seven major ash-flow sheets were erupted from a great chain of overlapping and nested cauldrons, which extend about 80 km to the southwest from Socorro (Chapin and others, 1978, fig. 2). These cauldrons are the surface expressions of shallow granitic plutons intruded along the zone of weakness of the Morenci lineament.

Crustal extension, which has produced the Rio Grande rift of today, began in the Socorro region about 32 m.y. ago (Chapin, 1978). The first sign of extensional stress was the eruption of a bimodal suite of rhyolite ash flows and voluminous basalt to basaltic andesite lavas, a magmatic association commonly attributed to extensional tectonism (Christiansen and Lipman, 1972). Near the north end of the Socorro Peak volcanic center, in the outflow environment of the Lemitar Mountains, the record of this magmatic change consists of the silicic ash-flow sheets of the 31 to 32-m.y.-old A-L Peak Tuff, the 28-m.y.-old Lemitar Tuff and the 26-m.y.-old tuff of South Canyon. Here the ash-flow sheets are separated by northward thickening tongues of the La Jara Peak Basaltic Andesite, erupted mostly from north-trending fissures along what is now the southern margin of the Colorado Plateau (Tonking, 1957; Massingill, 1978).

Near the axis of the developing rift, domino-style normal faulting was well underway by 28 m.y. ago when the Lemitar Tuff ash-flow sheet buried multiple fault blocks already tilted 5-15 degrees to the west in the Lemitar Mountains area (Chamberlin, 1978). Progressive rotation of the closely spaced domino blocks produced relatively minor fault-block topography that was mostly filled in by contemporaneous basaltic-andesite lavas of the middle La Jara Peak tongue. Thus, the distribution of the Lemitar Tuff was relatively unaffected by ongoing crustal extension.

Since the onset of crustal extension, the vertical grain of the Morenci lineament has been utilized as a deep-seated zone of lateral shearing, which has accommodated extension of the crust in opposing directions. In the Socorro-Lemitar mountains area, north of this transverse shear zone, the domino blocks have been stepped down toward an axis of extension, located to the east, concurrent with their westerly rotation. South of the shear zone, in the Chupadera Mountains area, the domino blocks have been down-thrown toward a westerly axis and rotated to the east. In other words, this segment of the Morenci lineament has been reactivated at depth as an incipient transform fault connecting en echelon axes of extension in the rift (Chapin and others, 1978). Where the transverse shear zone crosses the south-central portion of the Socorro Peak volcanic center, this differential rotation of longitudinal blocks has been accommodated by a diffuse zone of transverse faults, most of which are hinge faults and some of which are scissors faults.

Subsidence of the Socorro cauldron, which may be a complex combination of a trapdoor and a resurgent cauldron, has been tentatively correlated with eruption of the compositionally zoned Lemitar Tuff about 28 m.y. ago. The following tenuous history assumes cauldron-facies tuffs exposed in the eastern (resurgent) half of the cauldron to be uppermost quartz-rich Lemitar Tuff rather than the possible alternative, which is the 32-m.y.-old Hells Mesa Tuff (see "Lemitar Tuff -- cauldron facies"). Eruption of the lower crystal-poor zone, and the middle crystal-rich quartz-poor zone, of the Lemitar tuff initially formed a trapdoor cauldron hinged on the west (Osburn, 1978; Petty, 1979). Then apparently after a short time break, eruption of the uppermost crystal-rich, quartz-rich zone of the Lemitar Tuff caused differential subsidence of the east-half of the cauldron along a pre-existing north-trending fault zone. This older fault zone may have been an early-rift fault or may reflect the eastern margin of the Sawmill Canyon cauldron.

As much as 0.9 km of densely welded upper Lemitar(?) Tuff accumulated in the eastern part of the caldera as it collapsed. A thick section of cauldron-collapse mesobreccias (Lipman, 1976) were formed near the base of the upper zone as a result of landsliding on the oversteepened southern wall of the caldera. Near the end of the Lemitar(?) ash-flow eruptions, rhythmic blocking and explosive clearing of a local ignimbrite vent produced bedded lag-fall breccias containing abundant fragments of semicongealed magma. The magma fragments were essentially nonvesiculated and compositionally equivalent to the host tuff.

Although the Socorro cauldron apparently formed during the initial stages of crustal extension, its development was primarily controlled by changing magma pressures. A peripheral hingline structure in the southern Lemitar Mountains indicates pre-Lemitar tumescence and post-Lemitar subsidence. However, subsidence and resurgence was apparently accommodated locally by prevolcanic structures.

Following the major period of subsidence of the Socorro cauldron, mostly contemporaneous with the ash-flow eruptions, a period of erosion partially filled the depression with volcaniclastic sedimentary rocks. The northern margin of the caldera -- now exposed on the east face of Socorro Peak -- shed landslide deposits and colluvium derived from the pre-cauldron Spears Formation and the underlying Madera Limestone into the caldera. Shortly thereafter the eastern core of the cauldron began to be resurgently uplifted. This formed a semicircular moat between the resurgent block (now exposed in the northern Chupadera Mountains) and the eastern rim of the caldera. During resurgence the moat-like area between Socorro Peak and the Chupadera Mountains was filled by alternating local eruptions of intermediate (andesite to rhyodacite) lavas and lithic-rich rhyolitic ash-flow tuffs that were then capped by high-silica rhyolite domes and associated tuffs. These complexly intertonguing and heterogeneous deposits, collectively assigned to the Luis Lopez Formation, completely filled in the 500-800-m-deep moat and overlapped the inner topographic wall of the caldera in the Socorro Peak area. Minor subsidence of the moat-like block, along ring-fracture faults and faults partly bounding the resurgent block, occurred contemporaneously with moat filling. The latter faults, which trend north and northeasterly, apparently reflect prevolcanic fractures (i.e. Morenci lineament) in the cauldron block.

Major ash-flow volcanism ended in the Socorro region about 26 m.y. ago with eruption of the tuff of South Canyon, which apparently flowed into the Socorro cauldron area from a source cauldron to the west. The tuff of South Canyon filled paleovalleys in the moat area and wedged out onto the cauldron margin northeast of Socorro Peak. Minor volcanism related to the waning Datil-Mogollon field continued into early Miocene time as represented in a few isolated areas outside the Socorro Peak volcanic center.

The Neogene history of the Socorro Peak area is provided by basin-fill sedimentary strata and interbedded volcanic rocks of the Santa Fe Group. The development of rift basins in the map area may be divided into three stages that are recorded in strata of: 1) the lower Popotosa Formation of early Miocene age, 2) the upper Popotosa Formation of middle to late Miocene age, and 3) the Sierra Ladrones Formation of Pliocene to middle Pleistocene age. Silicic domes and tuffs of the Socorro Peak Rhyolite, erupted during the period 12-7 m.y. ago interbedded in the upper Popotosa Formation, provide important marker horizons in the basin fill. These late Miocene silicic domes, erupted along the north margin of the Socorro cauldron, are the most visible aspect of this locus of repeated magmatism. Figures 24, 26, and 33 respectively summarize the stratigraphic and structural relationships of the Popotosa and Sierra Ladrones formations at the time of their deposition.

By early Miocene time, about 25 m.y. ago, domino-style normal faulting had produced a broad sag basin (lower Popotosa basin) that included most of the map area. The resurgent block of the Socorro cauldron and other cauldron highlands to the southwest shed mudflow deposits and conglomerates northward onto the distended floor of the basin and buried domino blocks tilted as much as 30 degrees. The less extended area along the transverse shear zone may have tended to remain high and thus added to the elevation of these volcanic highlands that formed the southern margin of the early Popotosa basin. Contemporaneous hot spring activity in the volcanic highlands may account for the extreme induration of the silica-cemented mudflow deposits and their apparent high ferric iron content now expressed by their deep red coloration. This inferred early Miocene period of extensive hydrothermal alteration may be related to widespread potassium metasomatism (Chapin and others, 1978) that has affected essentially all the Oligocene volcanic rocks in the Socorro Peak area.

In the early part of lower Popotosa time, distal fan deposits, derived from a highland to the east, entered the Socorro Peak area in the "shadow" of some small, late Oligocene volcanoes in the Blue Canyon area. In the later portion of lower Popotosa time this eastern highland -- probably now represented by the Loma de las Canas uplift (fig. 2) -- continued to be a source of sediments, which overlapped these domes and then prograded west across the older mudflow deposits. These highlands to the east of the Rio Grande were apparently a constant sediment source throughout lower and upper Popotosa time. This relatively fixed eastern margin of the Popotosa basin may have been locally controlled by reactivation of a Laramide reverse fault as a down-to-the-west normal fault.

By middle Miocene time, about 20 m.y. ago, the extreme heat-flow regime of the Datil volcanic period had largely dissipated and there was a change in the structural style of rifting (Chamberlin, 1978). At about this time the western Popotosa basin was broken by a horst block in the area that now includes the northern Magdalena Mountains and the southern Bear Mountains. Chapin and Seager (1975) referred to this area as the ancestral Magdalena Mountains. This modification to the geometry of the original lower Popotosa basin shifted the basin axis into the Socorro Peak area and gave it a general north-south trend. As much as 800 m of playa muds were deposited along the axis of the relatively deep upper Popotosa basin. The playa was bordered by toes of alluvial fans at the east and west fringes of the map area. The south margin of the playa was more structurally than topographically controlled. Here the playa muds lapped onto a bedrock high in the northern Chupadera Mountains, which was probably related to the structurally high transverse shear zone. This south margin of the basin extended to the west into the Pound Ranch area of the Magdalena Mountains (Osburn, 1978).

In late Miocene time, from 12 m.y. to 7 m.y. ago, numerous rhyodacitic to rhyolitic domes and tuffs of the Socorro Peak Rhyolite were periodically erupted onto the playa floor. At this time, the east side of the present Socorro uplift was already a discrete tilted fault block, but it had no topographic expression since it was awash under a thin cover of playa deposits. Unconformities in the upper(?) Popotosa Formation in the Lemitar Mountains suggest that it too already existed as a tilted block largely mantled by a thin cover of alluvium (Lemitar Map). Deposition of playa and fan deposits of the upper Popotosa Formation continued during eruption of the silicic domes and until sometime after the last eruption 7 m.y. ago, which formed the Grefco perlite dome. During this interval of rhyolitic volcanism a second period of domino-style normal faulting occurred in the Socorro Peak area. This event rotated strata of the upper Popotosa Formation (and older strata) as much as 15 degrees prior to deposition of the Sierra Ladrones Formation. Some hydrothermal alteration and mineralization in the Socorro Peak area is clearly related to late Miocene intrusion. Here barite-silver veins cut rhyodacitic flows of the Socorro Peak Rhyolite, and related volcanic necks have been potassium metasomatized.

Between 7 and 4 m.y. ago, epeirogenic uplift and high-angle normal faulting (horst and graben style?) exhumed the Socorro and Lemitar uplifts from under their thin alluvial cover and elevated them sufficiently to topographically disrupt the floor of the Popotosa basin. During this period the ancestral Rio Grande was formed, apparently as a result of increased runoff from newly elevated alpine areas in

Colorado and northern New Mexico (Chapin, 1978). Fluvial sands of the ancestral Rio Grande now occur at the immediate foot of Socorro Peak, suggesting that the Socorro Mountains were not much of a topographic feature in middle Pliocene time. By 4 m.y. ago, when the basalt of Sedillo Hill was erupted, the Socorro fault zone had become the structural margin of the Socorro basin. These basalt lavas flowed, from their vents at Sedillo Hill, down a broad valley cut on upper Popotosa playa deposits and finally onto the flood-plain of the ancestral Rio Grande. The 4-m.y.-old lavas buried the Socorro fault zone and overlapped piedmont slope gravels shed from the eastern Magdalena Mountains. These piedmont slopes of the middle Pliocene eastern Magdalena Mountains apparently covered most of the Chupadera Mountains. Similar piedmont gravels locally derived from the Socorro and Lemitar Mountains, were deposited across broad pediments truncating tilted strata of the upper Popotosa Formation. These ranges continued to rise during late Pliocene and Pleistocene time. Since middle Pleistocene time continuing epeirogenic uplift, faulting, and lowering of the base level of the Rio Grande by its capture at El Paso has enhanced the topography of the modern ranges. Socorro Peak probably represents the boldest topographic relief in the map area since formation of the

Socorro cauldron in late(?) Oligocene time.

As a first order of approximation, the present ranges and basins of the map area represent relatively wide and slightly tilted domino blocks of Pliocene and Pleistocene age, which are bound by high-angle normal faults. The late Cenozoic fault blocks are superimposed on numerous closely spaced and moderately to strongly tilted domino blocks of late Oligocene to late Miocene age bounded by normal faults of moderate to low dip. The change from west-tilted to east-tilted domino blocks in the south-central portion of the volcanic center marks the general location of the transverse shear zone. The present surface expression of the transverse shear zone in the mapped area is a diffuse zone of discontinuous transverse hinge faults, which have allowed differential rotation of the longitudinal domino blocks. As much as 8 km wide, the shear zone also contains numerous horst-graben blocks formed by intermeshing of the opposing domino sets. Dips of strata generally decrease toward the axis of the transverse shear zone. The change in direction of dip and a null line of mesas along the axis of the transverse shear zone are the most visible aspects of this subtle transform structure (fig. 41).

The "leaky" vertical fabric of the Morenci lineament, which is now expressed by the transverse shear zone, has been the primary control of recurrent magma intrusion, silicic volcanism and hydrothermal activity in the Socorro Peak volcanic center. The deeply penetrating lineament has facilitated the ascent of magmas in the Socorro-Magdalena region in: middle to late Oligocene time (32-26 m.y. ago), early Miocene time (20 m.y. ago), late Miocene time (13-7 m.y. ago), and middle Pliocene time (4 m.y. ago). Within the Socorro Peak volcanic center, silicic eruptive events have been dated at: 28.6, 11.9-10.3, 10.5-9.0, and 7.4 m.y. The span of late Oligocene intrusion and volcanism, which may have been as long as 4 million years, is underrepresented by the single 28.6 m.y. date. Late Miocene calc-alkaline lavas of the Socorro Peak Rhyolite are associated with undated (middle to late Miocene) precursor and antecedent eruptions of alkali olivine basalts and xenocrystic basaltic andesites assigned respectively to the basalt of Kelly Ranch and the basalt of Bear Canyon. The alkalic basalt of Sedillo Hill was erupted 4.0 m.y. ago from vents within the transverse shear zone near the southwest side of the volcanic center. However, basalts of Pliocene age (3-5 m.y. old) are widely distributed along the Rio Grande rift (Chapin, 1979), suggesting that the transverse shear zone has little control on basaltic magmas.

In light of past episodes of volcanism, it is not surprising that shallow magma bodies are presently rising along the transverse shear zone at Socorro. These shallow (4-5 km) dike-like magma bodies, geophysically defined by Sanford and others, apparently form the heat source of the Socorro geothermal area (Chapin and others, 1978).

GEOTHERMAL POTENTIAL

For a detailed discussion of the geothermal framework and geothermal potential of the Socorro Peak area, the reader is referred to Chapin, Chamberlin and others (1978). In brief, the potential for commercial geothermal reservoirs in the Socorro Peak area is good. Favorable aspects include the presence of: 1) geophysically defined shallow magma bodies at depths of 4-5 km (pl. 3, map 4), 2) high heat flow, as high as 11.7 HFU (pl. 3, map 4), 3) potential reservoir horizons in the stratigraphic column which have been downfaulted to favorable depths (near tops of magma bodies) by cauldron subsidence and by subsidence in rift basins, and 4) relatively impermeable caprocks.

Brittle stratigraphic horizons, such as densely welded tuffs (i.e. Lemitar Tuff, A-L Peak Tuff) and the extremely indurated lower Popotosa Formation (Tpsd), are likely to be good reservoir rocks because of fracture induced permeability. Most perennial springs in the Socorro Peak volcanic center issue from the lower Popotosa Formation, where groundwater flow toward the Rio Grande Valley is blocked by faults that juxtapose impermeable claystones of the upper

Popotosa Formation (Tpkp) against the lower Popotosa conglomerates (Tpsd, Tplrl). Poorly welded, moat-fill tuffs of the Luis Lopez Formation (Tlt, Tlt_2, Tlrbt, Tlrst) apparently had a high initial porosity and permeability, as indicated by numerous silicified zones adjacent to intrusions. Basaltic to intermediate lavas (La Jara Peak Basaltic Andesite, Luis Lopez Formation lavas: Tla_1, Tla_2, Tla_3, Tlap, Tla_4) tend to act as caprocks because they are less brittle and have low primary permeability. The Pennsylvanian Sandia shales, andesitic conglomerates of the Spears Formation (fig. 6), and playa claystones of the upper Popotosa Formation are probably the most widespread potential caprocks.

The depth to potential reservoir horizons is controlled by the overprinting of multiple generations of domino-style fault blocks on the Oligocene Socorro cauldron, which is in turn superimposed on a Laramide uplift. Much of the complex older structure is masked by basin fill of the Santa Fe Group. Younger structures that cut the basin fill are deceptively simple.

Hydrothermal alteration patterns in the Socorro Peak volcanic center appear to represent at least two ancient geothermal systems. Silicified Pennsylvanian limestones, at Socorro Peak (pl. 1, fig. 13), and jasperized tuffs

(Tlt_2) of the moat-fill sequence, at the Shrine valley (pl. 1, pl. 2), are most likely late Oligocene features related to the Socorro cauldron. Barite and silver veins of the Socorro Peak district, which occur in the Popotosa Formation and in flows of the Socorro Peak Rhyolite, appear to be related to a late Miocene geothermal system.

The Socorro Peak area lies within a large potassium metasomatism anomaly that extends far outside the limits of the Socorro cauldron or the intrusive belt of the Socorro Peak Rhyolite (Chapin and others, 1978). Chemical analyses of Oligocene rocks (mostly welded tuffs) from the Socorro region commonly show K_2O contents of 6 to 11 percent in rocks that normally contain 4 to 5 percent K_2O. In thin section, the potassium metasomatized rocks are characterized by plagioclase phenocrysts that have been replaced by potassium feldspar and potassic "clays" (fig. 15, fig. 20). Other calcic minerals, such as clinopyroxene and sphene, are commonly replaced by Fe-oxides and leucoxene, respectively. In contrast, potassic minerals like sanidine and biotite are normally fresh in the potassium metasomatized rocks. The experimental data of Orville (1963) have shown that in vapor dominated geothermal systems, potassium is leached from the hotter rocks and then displaces sodium from the cooler rocks.

Ongoing chemical and petrographic studies indicate that the Socorro potassium anomaly occurs within a large triangular(?) area, which is about 60 km long on each side (C.E. Chapin, 1980, oral commun.). The anomaly extends from the western Magdalena Mountains, east to the southern Chupadera Mountains, and north to near the Ladron Mountains. Two features: 1) a field of strongly tilted Oligocene volcanic strata, and 2) an area of anomalously well indurated, red lower Popotosa Formation, both appear to be largely coincident with the potassium anomaly. The potassium anomaly and unusually red sediments are believed to be geochemical signatures of a metamorphic core complex of late Oligocene to early Miocene age (C. E. Chapin, 1980, oral commun.). The field of domino style normal faulting, which involved 30 to 40 degrees rotation of blocks in late Oligocene to early Miocene time, is considered to be the structural expression of the metamorphic core complex.

Additional research on hydrothermal alteration, mineralization, and the potassium metasomatism anomaly at Socorro is warranted. This type of data should be quite useful in evaluating the modern geothermal system.

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APPENDIX

A: Measured Sections

Note: Rock colors are from Geological Society of America "Rock Color Chart" and bed thickness classification follows that of Ingram (1954). Colors are of fresh rocks except where indicated otherwise.

Section 1: Socorro Peak Pennsylvanian Rocks

Formations: Madera Limestone and Sandia Formation

Location: General reference section measured on east face of Socorro Peak in NE/4 and NW/4 Sect. 9, T3S, R1W. Measured parallel to and 250 feet (76 m) south of section line common to sections 4 and 9, T3S, R1W. Begins in lower Sandia Formation, at elevation of 5400 feet, about 175 feet (53 m) north of gully bottom where light purplish gray volcanic-rich conglomeratic sandstone of lower Popotosa Formation (Tpsw, pl. 1) is in fault contact with lower Sandia Formation sandstone (pl. 1). Section proceeds upslope due west. Average dip, 25° west. Described by R. M. Chamberlin and W. T. Siemers, May 1977.

Unit No.	Thickness and Lithology
--	Top of section at base of covered slope with red muddy soil and limestone cobbles in float, below cliff of red-brown, lithic-rich, lower-tuff member (Tlt_1, pl.1) of the Luis Lopez Formation.
10	98 feet (29.9 m); silicified limestone; ledge and cliff forming, flinty, light- to medium-gray; shows thick, uneven bedding; black chert nodules and silicified fossils typical of limestones in lower units; limonite staining locally on fractures.
9	69 feet (21.0 m); (?) interbedded mudstone and limestone; poorly exposed, slope forming, (?) medium- to thick-bedded, dark-gray micrite and light- to medium-gray limy mudstone, some sandy limestones in float.
8	141 feet (43.0 m); limestone; bold cliff former, light- to medium-gray micrite; thick to very thick, uneven bedding; moderately fossiferous (fusulinids, brachiopods, crinoids), black chert nodules abundant in top 20 feet (6.1 m).
7	115 feet. (35.1 m); (?) interbedded mudstone and limestone, slope former like unit 9.
6	6 feet (1.8 m); silicified limestone; flinty, medium-gray, weathers moderate reddish brown.

5 126 feet (38.4 m); limestone; ledge and cliff former, forms prominent ridge in NW/4, sect. 4, T3S,R1W; light-bluish-gray to medium-gray micrite, thick- to very thick-bedded with a few laminated limy mudstone partings, moderate to slightly fossiliferous (crinoids, brachiopods, fusulinids, horn corals), 8-10 cm black chert bed with white fusulinid tests occurs about 30 feet (9 m) above the base.

4 24 feet (7.3 m); white rhyolite; sill-like intrusive in fault subparallel to bedding; repeats small thickness of lower Madera; contorted and parallel flow-banded, phenocrysts of white argillized feldspar and glassy quartz, locally speckled with yellow-brown limonite after (?) pyrite.

3 43 feet (13.1 m); limestone, ledge and slope forming, light-bluish-gray to medium gray-micrite, thick-bedded, laminated to massive, with minor interbedded medium- to dark-gray limy mudstone. Base of Madera Limestone, total thickness with eroded top: 598 feet (182.3 m).

2 205 feet (62.5 m); shale with minor interbedded limestone and sandstone; slope forming, dark-gray to black carbonaceous shale; with 1-3 m interbeds of ledge forming, dark-brownish gray, moderately

to richly fossiliferous (brachiopods, fusulinids) biomicrite, and reddish-brown to dusky-brown, fine- to coarse-grained, siliceous, feldspathic and quartz sandstone. Top of Sandia Formation

1 299 feet (91.1 m), siliceous quartz sandstone with minor interbedded siltstone and mudstone; ledge and cliff forming, weathers dark-reddish-brown to black, moderate yellowish-brown to light-gray on fresh surface, fine- to coarse-grained, massive to cross laminated, siliceous quartz sandstone; with several 1-2 m beds of medium-light-gray siltstone and limy mudstone. Lowermost 75 feet (22.9 m) extensively brecciated and silicified. Base of Sandia Formation not exposed, minimum thickness: 504 feet (153.6 m). Total thickness of Pensylvanian strata: 1102 feet (335.9 m).

Section 2:

Formation: A-L Peak Tuff

Location: Composite section measured in central and west-central Lemitar Mountains consisting of 3 separate sections: LM-4A, LM-4B, and LM-4C; respectively surveyed at (A) NW/4, SE/4, NW/4 Sect. 11, T 2 S, R 2 W, from 6320 to 6600 feet elevation, (B) NE corner of SE/4, SE/4, SE/4, Sect. 12, T2 S, R 2 W, at 5600 feet elevation, and (C) NW/4, SE/4, NW/4 to NE/4, SW/4, NW/4, Sect. 11, T 2 S, R 2 W, across ridge crest from 6580 to 6560 feet elevation. Described by R.M. Chamberlin, December, 1976.

Unit No.	Zone	Thickness and lithology
		LM-4C
		pinnacles(?) member (upper tuff of Bear Springs; Brown, 1972): (Talp, Lemitar map; 160 feet, 48.8 m)
9	d	25 feet (7.6 m) covered dip slope, top of unit estimated from projection of contact visible 120 m to south where overlain by basaltic andesite (Tba_1, Lemitar Map). Where locally exposed in

Unit No.	Zone	Thickness and lithology
		NE/4, SE/4 of Sect. 24, T 2 S, R 2 W, the top 10 m of the (?) pinnacles member consists of grayish-pink, partially welded, lithic-rich, pumiceous, crystal-poor rhyolite with 5 percent glassy sanidine and minor quartz, 3 to 5 percent small (< 1 cm) brown andesite lithics; and 10 percent small (0.5 - 3 cm) grayish-yellow pumice lentils.
9	c	127 feet (38.7 m); densely welded and vapor-phase zone; cliff- and hogback-forming; dark-grayish-red-purple to light-brownish-gray at top; very crystal-poor to crystal-poor rhyolite ash-flow tuff with 3 to 8 percent medium-grained phenocrysts of dull to pearly white sanidine, minor small quartz and rare small biotite; mostly moderately pumiceous, light-gray sandy-looking pumice filled with vapor-phase quartz.

Unit No.	Zone	Thickness and lithology
		A distinctive interval of very large ellipsoidal "fat" pumice up to a meter in length and 15 cm thick occurs near middle of this zone; some sanidine phenocrysts weather to honeycomb-like vuggy remnants. A noticeable increase in crystal content occurs about 40 feet (12.2 m) above the base of zone c.
9	b	1 foot (0.3 m); spherulitic zone; grayish-red mottled with pale-red spheroids and small orangish-pink streaks (pumice?); very crystal-poor rhyolite ash-flow tuff with traces of fine-grained white sanidine; sharp base and gradational top, exposed on cliff face.
9	a	7 feet (2.1 m); basal nonwelded to partially welded zone; well indurated, cemented by silica; medium-gray, aphyric rhyolite ash-flow tuff with trace of small andesitic lithics haloed

Unit No.	Zone	Thickness and lithology
		by light-gray silica; 15 cm of laminated air-fall tuff at cliff base.
		<u>LM-4B</u>
		<u>basaltic</u> <u>andesite</u> <u>and</u> <u>sandstone</u>: (Tala, Lemitar Map; 94 feet, 28.7 m)
8		85 feet (25.9 m); basaltic andesite lava; slope and ledge forming; grayish-red-purple, mottled with 10 percent, fine- to medium-grained, reddish-brown FeMg phenocrysts (olivine?) altered to iddingsite and hematite. A poorly exposed base (16 feet, 4.9 m) and top (31 feet, 9.4 m) are probably formed by scoriaceous flow breccia; the massive and relatively resistant center (38 feet, 11.6 m) contains numerous stringers of calcite.
7		9 feet (2.7 m); rhyolitic tuffaceous sandstone; pale-red to light-brownish-gray; fine- to medium-grained, with sub-angular grains of sanidine, quartz,

Unit No.	Zone	Thickness and lithology
		tuff fragments, and minor biotite; shows parallel lamination and graded bedding; well indurated, siliceous (?).
		flow-banded and gray-massive members, and other tuffs: (Talf, Lemitar Map; 491 feet, 149.7 m)
		upper lithic-rich tuff
6	b	5 feet (1.5 m) partially welded zone; ledge forming; pale-red mottled with reddish-brown lithics and grayish-yellow pumice; lithic-rich, pumiceous, crystal-poor to moderately crystal-rich rhyolite ash-flow tuff, with 10 to 15 percent fine grained phenocrysts of quartz, subequal to white, chalky sanidine and a trace of biotite; 3 to 5 percent small (<1 cm) lithics of red-brown andesite and pale-grayish-red, crystal-poor, rhyolite (? welded tuff); 10 to 15 percent small, angular, grayish-yellow and white,

Unit No.	Zone	Thickness and lithology
		crystal-poor, pumice lapilli are altered to clay.
6	a	5 feet (1.5 m) nonwelded zone, poorly exposed slope forming nonwelded equivalent of unit 6 zone b.
		<u>gray pumiceous tuff</u>
5		1 foot (0.3 m) densely to partially welded; light-brownish-gray; pumiceous, very crystal-poor, rhyolite ash-flow tuff, with 1 percent fine-grained sanidine and traces of quartz and biotite; abundant small (1-2 cm) light-gray pumice lentils; sharp welded contact at base with top of flow-banded member.

LM-4A

		<u>flow-banded member</u> (249 feet, 75.9 m)
4	d	47 feet (14.3 m) densely to partially welded, gray, well-foliated zone; cliff-

Unit No.	Zone	Thickness and lithology
		to bench-forming; light-brownish-gray crystal-poor rhyolite ash-flow tuff with 3 to 7 percent fine- to medium-grained glassy sanidine, sparse small quartz and rarely traces of biotite; medium-light-gray pumice forms 5 to 10 percent of rock as subtle thin folia locally with faint lineation apparent in the foliation plane. At the base of section LM-4B, the top 10 m of this zone is pale pink, partially welded and contains relatively abundant (½ to 1 percent) large flakes of reddish brown biotite and sparse, small, non-lineated pumice lentils; foliation attitudes in this zone are consistent and conformable to adjacent strata; lower contact with contorted and lineated zone gradational over about 5 m interval.
4	c	152 feet (46.3 m) densely welded, flow-banded, lineated and contorted zone;

Unit

No.	Zone	Thickness and lithology
		cliff forming; grayish-red-purple, microcrystalline matrix (granophyric crystallization?) with abundant light-gray to white folia of stretched pumice replaced by vapor-phase quartz producing the appearance of a flow-banded lava, crystal-poor rhyolite ash-flow tuff with 5 to 10 percent fine- to medium-grained phenocrysts of glassy (sometimes perthite-like) sanidine, minor small quartz and traces of biotite (mostly oxidized to black hematite); gray extremely flattened and stretched pumice forms 10 to 15 percent of rock and is strongly lineated in the foliation plane; foliation attitudes are variable indicating broad open folds; small-scale chevron folds are present locally; well-developed differential compaction occurs around sparse andesite and crystal-rich rhyolite lithics, vitro-

Unit No.	Zone	Thickness and lithology
		clastic textures not observed; lower contact gradational into pumiceous zone.
4	b	50 feet (15.2 m) lower, densely welded, pumiceous zone; steep-slope former commonly mantled by talus; grayish-red-purple streaked with moderately abundant medium-light-gray to light-gray pumice; crystal-poor rhyolite ash-flow tuff; mineralogy same as flow-banded contorted zone; moderately flattened glass bubbles visible in pumice where only partially replaced by vapor phase quartz; pumice commonly shows distinctive flame-like ends; foliation attitude consistent and conformable with adjacent strata; pumice becomes lineated near top of zone; base not exposed.
4	a	6 feet (1.8 m) colluvium covered interval, possibly represents a minor fault or basal nonwelded zone; lowermost 5 cm of unit 4, zone b, has hackled or

Unit No.	Zone	Thickness and lithology
		brecciated appearance, color change occurs across interval, welded contact occurs at this horizon in Bear Mountains and Joyita Hills.
		gray-massive member: (61 feet, 18.6 m)
3	b	51 feet (15.5 m) densely to partially welded; steep-slope forming; medium-light-gray; massive (nonfoliated); crystal-poor rhyolite ash-flow tuff with 5 to 10 percent fine- to medium-grained phenocrysts of glassy perthitic looking sanidine and a minor amount of small quartz in a flinty, lithoidal matrix; top 3 m has sparse, small, white pumice (filled with vapor-phase quartz) and small andesitic lithics; pumiceous top appears to be more welded than center indicating compound cooling, base not exposed.
3	a	10 feet (3.0 m) colluvium covered interval, possibly represents a minor

Unit No.	Zone	Thickness and lithology
		fault or basal nonwelded zone; color change across interval, no break in welding recognized at this horizon in Bear Mountains or Joyita Hills.
		<u>mottled tuff (? zone of gray-massive member)</u>: (46 feet, 14.0 m)
2		46 feet (14.0 m) densely welded; steep-slope forming; grayish-red-purple micro-crystalline matrix mottled with irregular-shaped splotches of grayish-red cryptocrystalline matrix; crystal-poor rhyolite ash-flow tuff with 5 to 10 percent phenocrysts of glassy sanidine (a few show faint blue chatoyancy) and minor small quartz; total phenocryst content increases slightly toward top, 2 to 3 percent small, medium- to light-gray pumice at base grades to very sparse pumice at top; short (3-5 cm) discontinuous irregular fractures

Unit No.	Zone	Thickness and lithology
		(cutting through phenocrysts) subparallel to eutaxitic structure and filled with white very fine-grained crystalline quartz (vapor phase?) are common near the middle of the unit; sharp densely welded contact with top of lower lithic-rich tuff.
		<u>lower lithic-rich tuff</u>: (118 feet, 36.0m)
1	c	15 feet (4.6 m) densely welded zone; ledge forming, pale-reddish-brown, lithic-rich, moderately pumiceous, crystal-poor rhyolite ash-flow tuff with 5 to 10 percent fine- to medium-grained phenocrysts of glassy sanidine and minor small quartz; 2 to 3 percent small reddish-brown andesite lithics and purplish-gray aphanitic rhyolite lithics, and 1 to 2 percent small, medium-gray pumice; abrupt change in color, dissappearance of lithic fragments and slight decrease

Unit No.	Zone	Thickness and lithology
		in size and amount of phenocrysts occurs at welded contact with mottled zone indicating flow unit boundary.
1	b	71 feet (21.6 m) partially welded zone, hackled, ledge- to slope-forming, moderate-orange-pink to light-brownish-gray, lithic-rich, pumiceous, crystal-poor rhyolite ash-flow tuff with 5 to 10 percent fine- to medium-grained phenocrysts of glassy sanidine, minor small quartz and traces of biotite (near base of zone); 5 to 10 percent small and medium sized (up to 5 cm) lithic fragments of reddish-brown porphyritic to aphanitic andesite, medium-gray, spherulitic to massive, crystal-poor rhyolite and some red muddy sandstone; and 2 to 4 percent small (1-2 cm), very crystal-poor, grayish-orange-pink to light-gray pumice; pumice commonly has internal botryiodial texture near middle

Unit No.	Zone	Thickness and lithology
		of zone.
1	a	32 feet (9.6 m) nonwelded zone; slope to bench forming; poorly exposed, light-gray to pinkish-gray, clayey soil former with abundant weathered-out andesitic lithic fragments; crystal-poor rhyolite ash-flow tuff; overlies ledge forming top of densely welded Hells Mesa Tuff.

Section 3:

Formation: Lemitar Tuff (Lemitar Map, section LM-6; 126 feet, 38.4 m)

Location: Type area of Lemitar Tuff in the central Lemitar Mountains adjacent to Canoncito del Puerticito del Lemitar where a small tributary arroyo cuts across a hogback of the welded tuff (NW corner, SE/4, SE/4, Sect. 12, T2S, R2W). At this locality the Lemitar Tuff forms two hogbacks because it is repeated by a normal fault. Section was surveyed along the north wall of the water cut through the western hogback. Average strike and dip is N-5, 55°W. Described by R.M. Chamberlin, December, 1976.

General comments: At this locality, the Lemitar Tuff departs only slightly from a simple-cooling unit consisting of at least three separate ash flows (nuées ardentes) designated as flow units 1, 2, and 3, from bottom to top. Flow unit 3 is a distinctive compositional subunit of crystal-rich, quartz-rich, rhyolite. The tops of flow units 1 and 2 are marked by finely vesicular zones (scoria-like) that are abruptly terminated at a densely welded contact with the essentially non-vesicular base of the overlying flow-unit. Changes in color occur at both contacts; the top of flow unit 2 is also marked by changes in weathering character and pumice content. Other flow-unit boundaries are probable but have not been recognized. The ash-flow sheet is here subdivided into 10 zones with gradational boundaries marked by one or more interrelated changes in color, weathering character, phenocryst content, pumice content and degree of welding. Additional or alternative subdivisions are possible depending on the parameters chosen. Throughout the section potassium metasomatism has altered plagioclase phenocrysts to chalky white clay and milky white granular appearing potassic feldspar. The clays are washed from the weathered surface leaving rounded to subhedral holes in the surface; these should not be confused with the fine scoriaceous zones at the top of the flow units. Other calcic minerals, clinopyroxene and sphene, which locally occur in trace amounts, are altered by potassium metasomatism to hematite(?) and leucoxene respectively. Modal and chemical analyses of the major rock types are included in the text (tables 3 and 4 respectively); sample numbers representative of different lithologies are included with each description for cross reference to these tables.

Flow Unit	Zone	Thickness and lithology
		upper member (Tlu, Lemitar Map; 88 feet, 26.8 m)
3	c	4 feet (1.2 m); partially welded, pumiceous, crystal-rich, quartz-rich, rhyolite (like LM-6-8) ash-flow tuff; grayish-pink, weathers to grayish-orange-pink; mottled with 3 percent of medium-brownish gray, moderately crystal-rich, scoriaceous rhyodacite pumice (like LM-6-8a); sugary looking vapor phase minerals in pumice and matrix; conspicuous coppery biotite. Overlain by lavas of the upper tongue of La Jara Peak Basaltic Andesite with 1 m of red silty mudstone at their base.
3	b	21 feet (6.4 m); densely welded, pumiceous, crystal-rich, quartz-rich, rhyolite (like LM-6-8 and LM-6-8b) ash-flow tuff; light-grayish-red to pale-red, weathers pale-red,

Flow Unit	Zone	Thickness and lithology
		mottled with abundant (3-10 percent) grayish-red, scoriaceous, moderately crystal-poor, rhyodacite pumice (LM-6-8a) containing approximately 10 percent plagioclase (altered), 2 percent glassy sanidine and 1 percent coppery to black biotite. Also present are sparse (< 1 percent), white, vuggy, very crystal rich (45-60 percent), high silica rhyolite pumice (like PR-1-77b) containing coarse-grained, euhedral sanidine and quartz. A swarm of large (as much as 60 cm long by 7 cm thick) mafic pumice about 2 m thick marks base of zone b, gradational top at color change.
3	a	8 feet (2.2 m); densely welded, massive, crystal-rich, quartz rich, rhyolite (LM-6-8) ash flow tuff; light grayish red, trace of white crystal-rich

Flow Unit	Zone	Thickness and lithology
		pumice, forms basal zone of flow-unit no. 3 with sharp welded contact at base, contact highly visible because of change in color and weathering character.
2	b	9 feet (2.7 m); finely vesicular (vesicles: 1-3 percent, 0.2-0.5 mm diam.), densely welded, moderately pumiceous, crystal-rich, quartz-poor, quartz latite (like LM-6-6b and LM-6-5) ash-flow tuff; light-red to pale-red, weathers moderate-red to grayish-red, joint blocks weather to rounded forms, abundant altered plagioclase phenocrysts weather out to form subhedral vugs generally much larger than the associated vesicles contains 1-2 percent of large (10 cm) mafic crystal-poor pumice and small (< 1 cm) white crystal-rich pumice. Biotite in flow unit 2 is black to slightly coppery. Sharp welded contact at top, lower contact

Flow Unit	Zone	Thickness and lithology
		placed at gradational change in color and outcrop character.
2	a	25 feet (7.6 m); non-vesicular, densely welded, pumice-poor, crystal-rich, quartz-poor, quartz latite (LM 6-6b and LM-6-5) ash-flow tuff; light-grayish-red-purple, weathers grayish-red-purple, forms angular joint blocks, traces of mafic and felsic pumice, upper contact gradational, lower contact is marked by a subtle but sharp color change (more reddish below contact).
1	e	10 feet (3.0 m); massive, finely vesicular, densely welded, pumice-poor, crystal-rich, quartz-poor, quartz latite (LM-6-5) ash-flow tuff; moderate-grayish-red, weathers grayish red, traces of white crystal-rich pumice; locally contains small enechelon quartz stringers cutting across both altered

Flow Unit	Zone	Thickness and lithology
		and fresh phenocrysts and filling vesicles, lower contact gradational. Offset 3 m to north across minor low-angle normal fault that repeats the basal 1.5 m of zone "e".
1	d	11 feet (3.3 m); compositionally streaked zone consisting of intimately mixed bands (1-10 m long and 1-20 cm thick with flame-like terminations) formed by a mixture of densely welded, crystal-rich, plagioclase-rich, quartz-poor, quartz latite (like LM-6-5) and densely welded, moderately crystal-rich, plagioclase-poor, pumiceous, silicic, rhyolite (like LM-6-3); quartz latite is massive looking, moderate-grayish-red and weathers grayish red; rhyolite is pale-red-purple and streaked with abundant, small, medium-light-gray, cherty-looking, pumice. Darker colored quartz latite bands are dominant at top

Flow Unit	Zone	Thickness and lithology
		of zone and locally contain small granitic-looking pods of coarsely crystalline sanidine, quartz and plagioclase; gradational basal contact of "upper member" and zone "d" placed at downward disappearance of small wisps of quartz latite.
		lower member (Tll, Lemitar Map; 38 feet, 11.6 m).
1	c	11 feet (3.3 m); densely welded, pumiceous, moderately crystal-rich, rhyolite (LM-6-3 and LM-6-2 c) ash-flow tuff; top 2.4 m of this zone is pale-red-purple, weathers moderate-grayish-red, and is mottled by abundant, crystal-poor, light-gray, pumice that are outlined with moderate-pink spherulitic rims; lower 0.9 m of zone is light-brownish-gray, weathers pale brown and has abundant pumice like in top part

Flow Unit	Zone	Thickness and lithology
		of zone; small, grayish-red lithic fragments of porcellaneous crystal-poor rhyolite occur sparsely. Trace amounts of small, yellowish-brown, skeletal, crystals of leucoxene, after sphene, are conspicuous in this zone. Color change from red to gray, used locally for reference in mapping lower versus upper member, occurs within 3 m of the actual contact.
1	b	22 feet (6.7 m); partially welded, pumiceous, lithic-poor, moderately crystal-rich, to crystal-poor, rhyolite to high-silica rhyolite (LM-6-2b and LM-6-2) ash-flow tuff; light-brownish-gray, weathers brownish-gray, mottled with small, crystal-poor, very-light-gray pumice locally of botryoidal appearance in the lower 3 m of this zone, small lithic fragments of brownish-gray crystal-poor rhyolite

Flow Unit	Zone	Thickness and lithology
		(welded tuff?) and grayish-red, porphyritic andesite, and reddish brown mudstone(?) occur sparsely throughout this zone. "Sandy" vapor phase minerals replace pumice in top half of this zone and lower half of overlying zone c.
1	a	23 feet 7.0 m; talus covered slope; La Jara Peak Basaltic Andesite lavas (Tba_1 , Lemitar Map) exposed at base of slope; 5 feet (1.5 m) of non-welded, friable, clayey, grayish-pink, crystal-poor rhyolite(?) ash-flow tuff is exposed at the south end of the eastern hogback of Lemitar Tuff about 100 m southeast of the bottom of this measured section; lower 18 feet (5.5 m) of the covered slope assumed to be underlain by basaltic-andesite lavas.

APPENDIX B: Data for unpublished radiometric ages cited in text. Dates are unpublished data of C. E. Chapin, New Mexico Bureau of Mines and Mineral Resources and were performed by by Geochron Laboratories, except where noted otherwise. Dates that may be errorneous are queried; see text for additional interpretations.

Formation (unit) location/sample no.	Age (m.y.)	Method	Ave Ar^{40} (ppm)	K^{40} (ppm)	percent K
A-L PEAK TUFF (flow-banded):					
Lemitar Mts., 76-6-7	?27.4±1.2	K/Ar-sanidine	0.006509	4.027	3.301
LA JARA PEAK BASALTIC ANDESITE:					
Bear Mts., SC-PO-1	30.2±1.4	K/Ar-whole rock	0.004155	2.330	1.910
Bear Mts., SC-PO-2	26.6±1.1	K/Ar-whole rock	0.008471	5.241	4.296
LEMITAR TUFF.					
Joyita Hills, 77-12-5b	[1]28.8±0.7	K/Ar-whole rock	(7.205)[A]	(82.70)[B]	6.38
San Mateo Mts., 77-9-5	27.0±1.1	K/Ar-biotite	0.01022	6.431	5.272
Lemitar Mts., 76-1-10	?26.3±1.0	K/Ar-biotite	0.01185	7.644	6.266
LUIS LOPEZ FORMATION:					
(rhyolite of Hwy. 60)					
N. Chup. Mts., 77-5-2	28.6±1.1	K/Ar-biotite	0.01400	8.299	6.803
(rhyodacitic dike, Tiap, pl. 1)					
N. Chup. Mts., 76-6-4	?22.8±0.9	K/Ar-biotite	0.01090	8.128	6.662

1 Data from A. L. Odom, Florida State University
A Ar^{40} (rad) x 10^{-6} sec
B percent radiogenic K

Formation (unit) location/sample no.	Age (m.y.)	Method	Ave Ar^{40} (ppm)	K^{40} (ppm)	percent K
TUFF OF SOUTH CANYON:					
Joyita Hills, 76-4-2	26.2±1.0	K/Ar-biotite	0.01199	7.755	6.357
SOCORRO PEAK RHYOLITE:					
(hornblende rhyodacite)					
tuff west of J. B. Kelly Ranch, P-40	[2]11.9±0.8	fission track-zircon	--	--	--
Stonewall dome, 77-8-1	10.3±1.5	K/Ar-hornblende	0.000396	0.658	0.539
"6001" Mesa, 77-5-1	[3]10.3±	K/Ar-whole rock	(0.7558)[A]	(38.1)[B]	1.88
"6001" Mesa, 77-5-1	?6.0±0.6	K/Ar-hornblende	0.000233	0.664	0.545
(biotite-hornblende quartz latite-rhyolite)					
Strawberry Peak, 76-6-1	11.8±0.5	K/Ar-biotite	0.005605	8.103	6.642
Radar Peak	[4]11.5±1.0	K/Ar-whole rock	--	--	--
(phenocryst-rich rhyolite)					
Signal Flag, 76-6-2	10.5±0.4	K/Ar-biotite	0.005387	8.738	7.163
"6633" Peak, 76-6-3	9.0±0.4	K/Ar-biotite	0.004267	8.068	6.613
(phenocryst-poor rhyolite)					
Grefco mine, 77-5-4	[3]7.4±	K/Ar-whole rock	(1.355)[A]	(51.9)[B]	4.71

2 Data from Kim Manley, U.S.G.S. Denver, CO
3 Data from A. L. Odom, Florida State University
4 Data from C. T. Smith, New Mexico Institute of Mining and Technology
A Ar^{40} (rad) x 10^{-6} sec
B percent radiogenic K

Appendix C: Petrographic Reference Collection

The following list includes one representative sample of each major volcanic-stratigraphic unit (from oldest to youngest), or particular stratigraphic zones of interest in the Socorro Peak volcanic center. A hand specimen and thin section of each sample in this list is housed in the Petrological Reference Collection, Department of Geological Engineering, Colorado School of Mines, Golden, Colorado. A more comprehensive petrological collection for the Socorro Peak volcanic center and the Lemitar Mountains is on file at the New Mexico Bureau of Mines and Mineral Resources, Socorro, New Mexico.

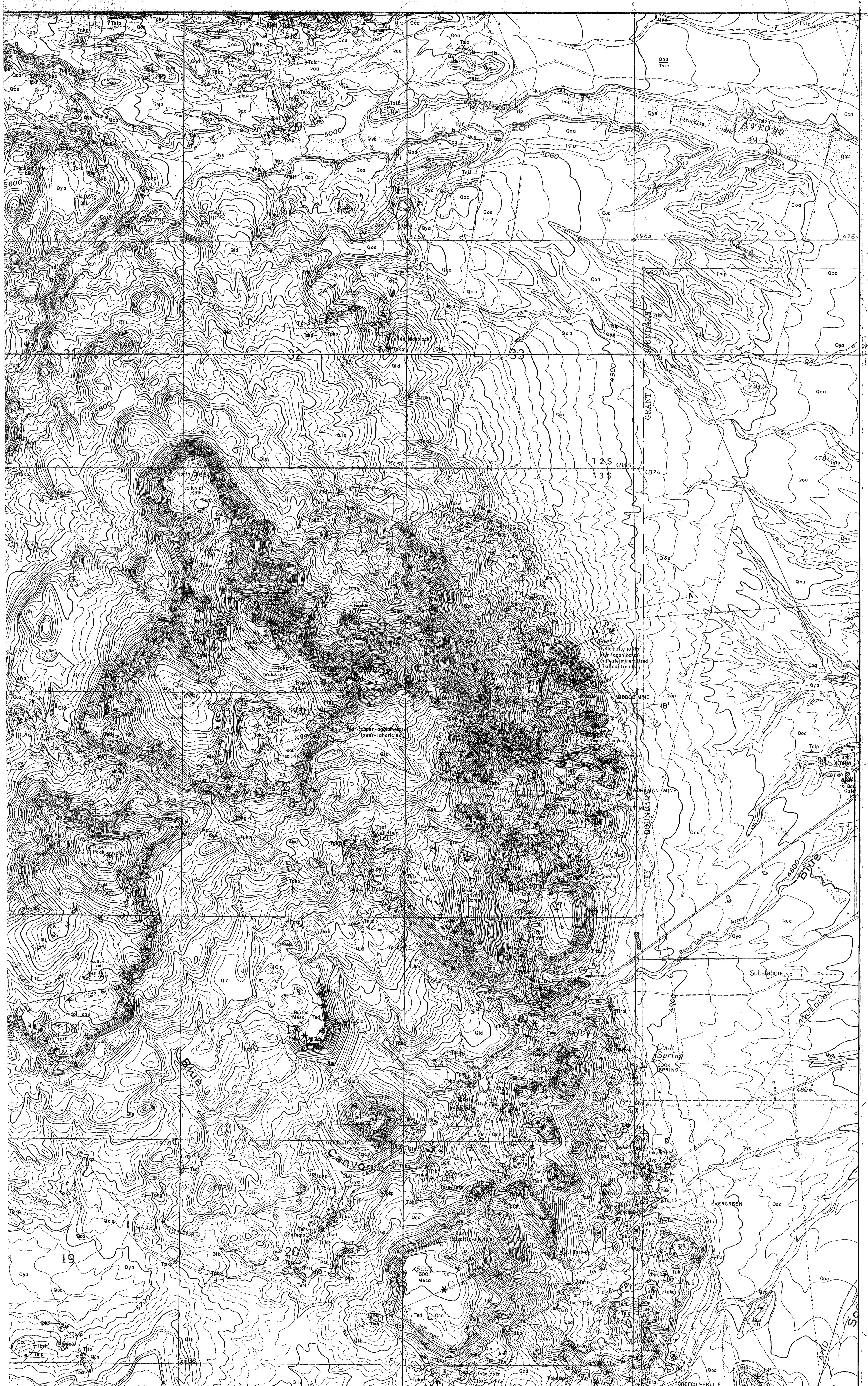
P.R.C. No.	Sample No.	Formation	Unit	Symbol (pl.1)
165-1	LM-4-4b	A-L Peak Tuff	flow-banded member	Talf
165-2	LM-4-7d	A-L Peak Tuff	pinnacles(?) member	Talp
165-3	SP-375	cauldron-facies Lemitar(?) Tuff	mesobreccia unit	Tlx
165-4	SP-376	cauldron-facies Lemitar(?) Tuff	lag-fall breccia	Tlu
165-5	LM-6-2	Lemitar Tuff	crystal-poor zone	Tll
165-6	LM-6-5	Lemitar Tuff	quartz-poor zone	Tlu
165-7	76-1-10	Lemitar Tuff	potassium metasomatized quartz-poor zone	Tlu
165-8	LM-6-8	Lemitar Tuff	quartz-rich zone	Tlu
165-9	77-2-2	La Jara Peak Basaltic Andesite	upper tongue	Tba_2
165-10	SP-303	Luis Lopez Fm.	lithic-rich tuff	Tlt_2
165-11	77-7-7	Luis Lopez Fm.	andesite lava	Tla_3
165-12	77-5-6	Luis Lopez Fm.	rhyolite of Blue Canyon	Tlrb
165-13	77-5-2	Luis Lopez Fm.	rhyolite of Highway Sixty	Tlrst
165-14	LM-8-3a	tuff of South Canyon	upper moderately crystal-rich zone	Tsc
165-15	76-1-9	upper Popotosa Fm.	basalt of Kelly Ranch	Tpkb
165-16	77-5-1	Socorro Peak Rhyolite	rhyodacite unit	Tsd
165-17	76-6-1	Socorro Peak Rhyolite	quartz latite-rhyolite unit	Tsd

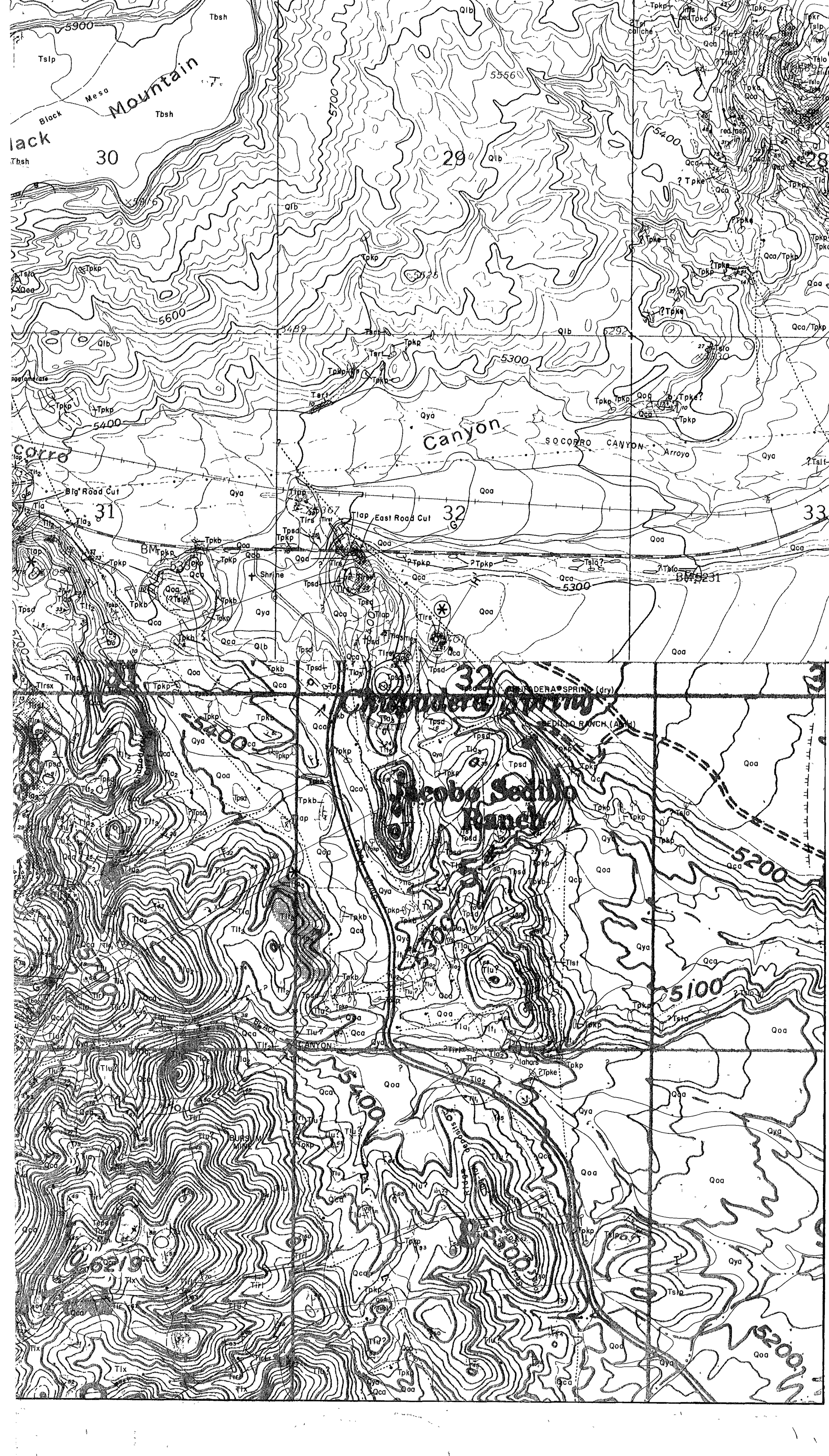
P.R.C. No.	Sample No.	Formation	Unit	Symbol (pl.1)
165-18	76-2-2b	Socorro Peak Rhyolite	crystal-rich rhyolite unit	Tsr
165-19	77-5-4	Socorro Peak Rhyolite	crystal-poor rhyolite unit	Tsr
165-20	77-5-3	upper Popotosa Fm.	basalt of Bear Canyon	Tpkb
165-21	76-6-11	Sierra Ladrones Fm.	basalt of Sedillo Hill	Tbsh

PLATE I

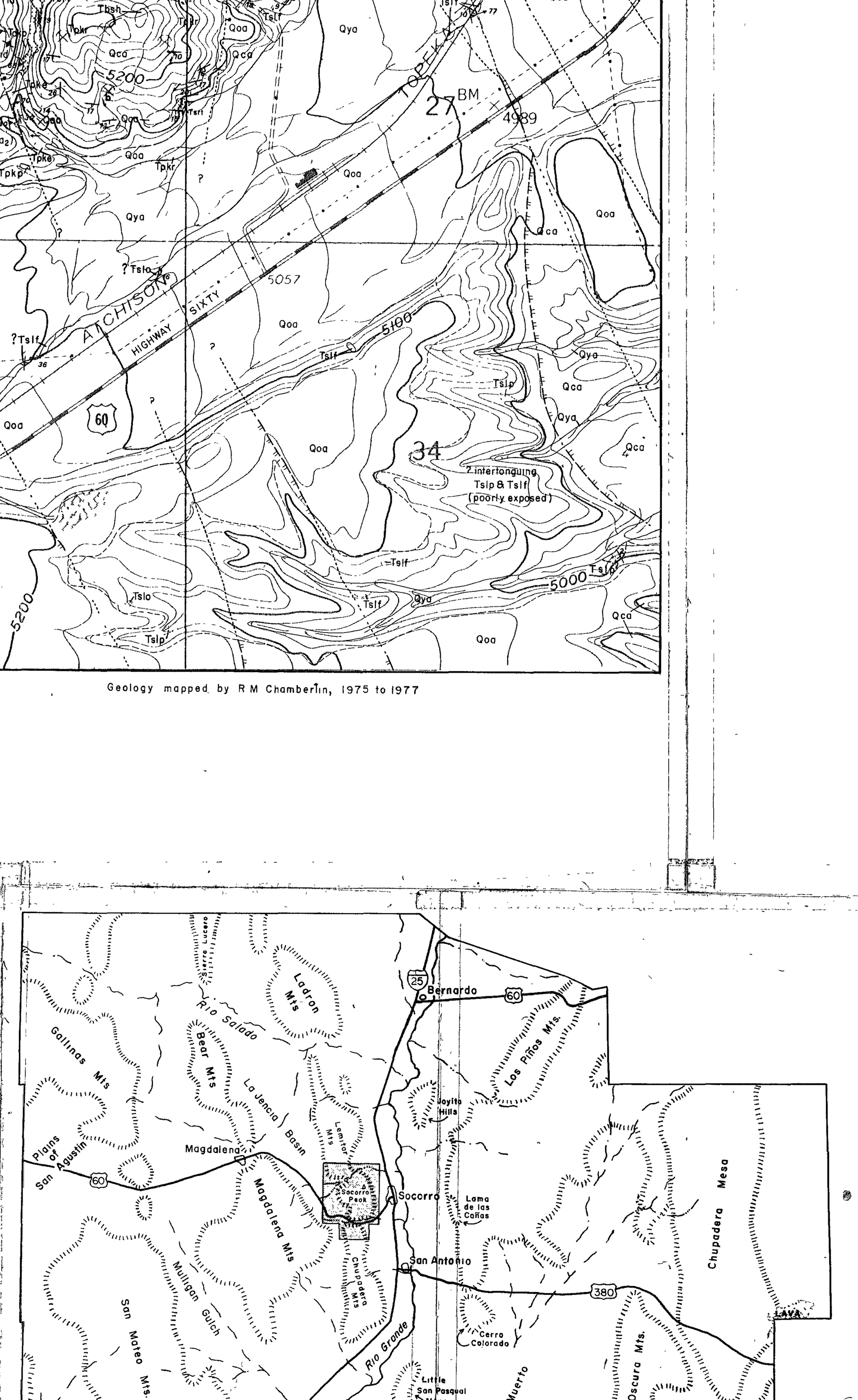
.O, AND NORTHERN CHUPADERA MOUNTAINS, COUNTY, NEW MEXICO

by
Richard M. Chamberlin, 1980





Geology mapped by R M Chamberlin, 1973 to 1977



INDEX MAP OF SOCORRO COUNTY
(APPROXIMATE AREA OF GEOLOGIC MAP SHADED)

CORRELATION OF MAP UNITS

QUATERNARY — Holocene, Pleistocene

Qca — Colluvium and minor alluvium undivided

Qya — Younger alluvium

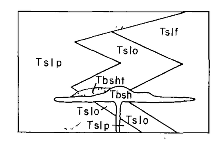
Qoa — Older alluvium

Qlb, Qlr, Qld — Landslide areas characterized by large lava blocks, hummocky topography and small closed basins. Includes undifferentiated alluvium and colluvium. Modifying suffixes b = basalt blocks, r = rhyolite blocks, d = dellenite blocks

UNCONFORMITY

TERTIARY — Pliocene

Santa Fe Group — Sierra Ladrones Formation



facies and basalt flows

Tslf	Fluvial sands of ancestral Rio Grande and tributaries
Tslo	Overbank mud-silt of ancestral floodplain
Tslp	Piedmont-slope and alluvial-fan gravels
Tbsh	Basalt flows of Sedillo Hill (4 0 ± 0 3 m y)
Tbsht	Basaltic pyroclastics consisting of agglomerates, tuffs and cinders

UNCONFORMITY

Tsr | Tsrt | Tsd | Tsdt

Socorro Peak Rhyolite
(interbedded with member of Kelly Ranch, Popotosa Formation)

Tir | Tid

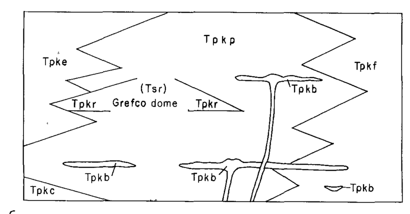
Tir — Intrusive rhyolite like Tsr
Tid — Intrusive dellenite like Tsd

Tsr	Rhyolite and quartz latite(?) domes and flows with phenocrysts of quartz, sanidine, biotite, plagioclase (7 4 to 10 5 m y)
Tsrt	Rhyolitic tuff beds associated with Tsr
Tsd	Dellenite, rhyodacite to quartz latite domes and flows with phenocrysts of plagioclase, ± hornblende, ± biotite (10.3 to 11 8 m y) Locally subdivided into older (Tsd_1) and younger (Tsd_2) flows west of Socorro Peak
Tsdt	Dellenitic tuffs and agglomerates associated with Tsd, includes pumiceous conglomerate near Kelly Ranch (11 9 m.y) and silicified lapilli tuff-breccia and laharic breccias near Socorro Peak

Miocene

Santa Fe Group — Popotosa Formation

upper Popotosa Formation (member of Kelly Ranch)

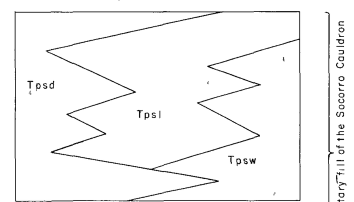


facies and basalt flows

Tpkp	Mudstone, claystone, siltstone and gypsum, playa/alluvial flat
Tpkf	Sandstone and conglomeratic sandstone, fluvial (braided channel)
Tpke	Fanglomerates near Evergreen
Tpkr	Rhyolite-vitrophyre-rich conglomerates and conglomeratic sandstones around the Grefco "perlite" mine, includes other small outcrops of rhyolitic sandstones and conglomerates locally derived from Socorro Peak Rhyolite lavas
Tpkb	Basalt flows, lower flows near Kelley Ranch (9 1 ± 0 5 m y), upper flows near Bear Canyon (vent areas unknown)
Tpkc	Basal muddy conglomerates and sandy conglomerates, locally derived from reworking Tpsd, possibly equivalent to Tplb or Tplru

UNCONFORMITY

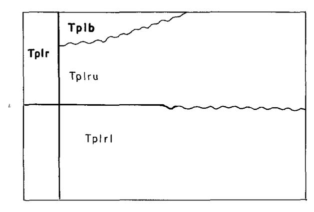
late sedimentary fill of the Socorro Cauldron — lower Popotosa Formation (member of Socorro cauldron)



facies

Tpsd	Lahars and fanglomerates shed from resurgent dome area of the Socorro cauldron
Tpsl	Lacustrine and alluvial-flat mudstone, siltstone and sandstone near moat axis
Tpsw	Fanglomerate and sandstone shed from caldera walls

lower Popotosa Formation (member of Lemitar Mountains)



facies and local units

Tplb	Buff fanglomerates near Strawberry Peak (? may locally intertongue with basal Tpk strata)
	Local unconformity
Tplru	Upper red fanglomerates, in part derived from reworking Tplrl
	Local unconformity
Tplrl	Lower red fanglomerates and lahars (equivalent to Tpsd, but outside Socorro cauldron)
Tplr	Undivided Tplr and Tplr

UNCONFORMITY

Oligocene — COLLAPSE AND RESURGENCE OF THE SOCORRO CAULDRON

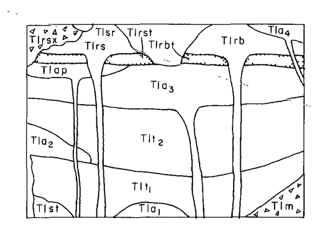
regional volcanic units which partially fill the Socorro Cauldron and locally cap the Luis Lopez Formation

Tsc — Tuff of South Canyon (26 2 ± 1 0 m y)

Tba_2 — Upper basaltic andesite lavas (upper tongue of La Jara Peak Basaltic Andesite, in part time equivalent to the Luis Lopez Formation)

Tiba ? — Basaltic-andesite dike in southern Lemitar Mountains, like Tala, Tba_1, and Tba_2 (age uncertain, cuts Talf)

Luis Lopez Formation — local volcanic and minor sedimentary moat-fill deposits of the Socorro Cauldron



Tirb | Tirl

Tirb — Intrusive rhyolite, like Tlrb
Tirl — Other types of intrusive rhyolite related to the Luis Lopez Formation

Tia — Intrusive andesite, like $Tla_{1,2,3,4}$
Tiap — Intrusive rhyodacite porphyry, like Tlap

members

Tlsr	Sandstone, red rhyolitic channel
Tlrs	Rhyolite domes and flows near Highway Sixty
Tlrsx	Rhyolite breccia, ancient colluvium derived from Tlrs
Tlrst	Bedded and welded tuffs associated with Tlrs (28 6 ± 1 1 m y)
Tla_4	Andesite lavas, medium-porphyritic
Tlrb	Rhyolite domes and flows near Blue Canyon
Tlrbt	Bedded and welded tuffs associated with Tlrb
Tlap	Rhyodacite lavas, coarsely porphyritic (feeder dike, ?22 8 ± 0.9 my)
Tla_3	Andesite lavas, agglomerates, tuffs and andesitic laharic breccias fine-to medium-porphyritic
Tlt_2	Welded, rhyolitic lapilli tuff and tuff breccia with Lemitar Tuff clasts
Tla_2	Mafic lava (rhyodacite ?, andesite ?) locally with small quartz phenocrysts, aphanitic to medium-porphyritic
Tlt_1	Welded, rhyolitic, lithic-rich lapilli tuff with andesite clasts
Tla_1	Andesite lava, aphanitic to medium-porphyritic, with minor "Spears-like" conglomerates at top
Tlst	Tuffaceous sandstones and bedded tuffs, light gray rhyolitic
Tlm	Landslide mega breccias and ancient colluvium with clasts of Spears andesite, tuff of Granite Mountain and Madera Limestone; minor sandstones near top

LOCAL UNCONFORMITY AT CAULDRON WALL

Lemitar Tuff

Tlu — Upper crystal-rich member (avg 27 9 m y ± 1.4 m y)

Tll — Lower crystal-poor member

Tlx — Thick zone of cauldron collapse ? breccias (heterolithic) with abundant Precambrian and andesite clasts in crystal-rich matrix

NOTE Tlx and Tlu ? south and east of Tower mine may be caldera-facies Hells Mesa Tuff, see text for discussion

Ttms — Conglomerates and sandstones of Tower Mine drill hole shown in cross section H-H' only (equivalent in part to "unit of Sixmile Canyon," Osburn, 1978)

Tba_1 — middle basaltic-andesite lavas (middle tongue of La Jara Peak Basaltic Andesite)

PARACONFORMITY TO DISCONFORMITY

A-L Peak Tuff

Talp — Pinnacles member

Tala — Basaltic-andesite lavas and minor rhyolitic sandstones (lower tongue of La Jara Peak Basaltic Andesite)

Talf — Lower-lithic, mottled, gray-massive, flow-banded, and upper-lithic members undivided

NOTE: The following formations are present in the subsurface of the Strawberry Peak area and probably in the subsurface of the northern Chupadera Mountains unnamed andesite lavas below the A-L Peak Tuff, Hells Mesa Tuff, tuff of Granite Mountain, and Spears Formation.

REGIONAL UNCONFORMITY

PENNSYLVANIAN

IPm — Madera Limestone

IPs — Sandia Formation

REGIONAL UNCONFORMITY

PRECAMBRIAN

p€m — Argillite with a monzonite intrusion exposed in Woods Tunnel

EXPLANATION

Contact long dashed where approximately located, short dashed where inferred, dotted where concealed

Thin ash bed or tuff interbedded in Santa Fe Group

Intraformational contact

strike and dip of stratification: horizontal; 46 Inclined; vertical; 84 overturned

strike and dip of foliation: horizontal; 60 Inclined; vertical; 69 overturned

b

Paleocurrent direction, from pebble imbrication where unlabeled, where labeled **b** = cross bedding, **a** = festoon trough axis, **ll** = parting lineation, **lo** = log axis, **i** = inferred (in Qoa only), double headed where ambiguous

S

Paleoflow direction in volcanic rocks, as labeled **s** = stretch marks, **v** = elongated vessicles, **p** = pumice lineation, **f** = flow thrust

Pyroclastic strata associated with lavas, domes and local volcanic vents

Silicified rocks, most originally limestones

Breccias and bouldery deposits interpreted as ancient colluvium

mining symbols: Shaft; Bulldozer trench or open cut; Adit or portal; Prospect; Large pit or cut; Inclined diamond-drill hole showing collar and surface projection

Faults all faults are long dashed where approximately located, short dashed where inferred, and dotted where concealed Measured dip directions and angles indicated with arrows and numbers

74 ?

High-angle normal fault, dip >45°, ball and bar on downthrown block, queried where hypothetical

30 L L

Low-angle normal fault, dip <45°, "L" on downthrown block

Ring fracture or circumfrential fault related to the Socorro cauldron

Fault scarp (solid) and fault line scarp (dashed) in alluvium, hachures on downside

66

Epithermal or telethermal vein

S

Gravity-slide fault mostly toreva style slump blocks

Trace of plunging synclinal axis

A A'

Line of structure section

Pull apart structures related to slumping

T

Perennial spring, T = thermal spring

Earthen dam

Ancient landslide block of Madera Limestone in "Tlm" southeast of Socorro Peak

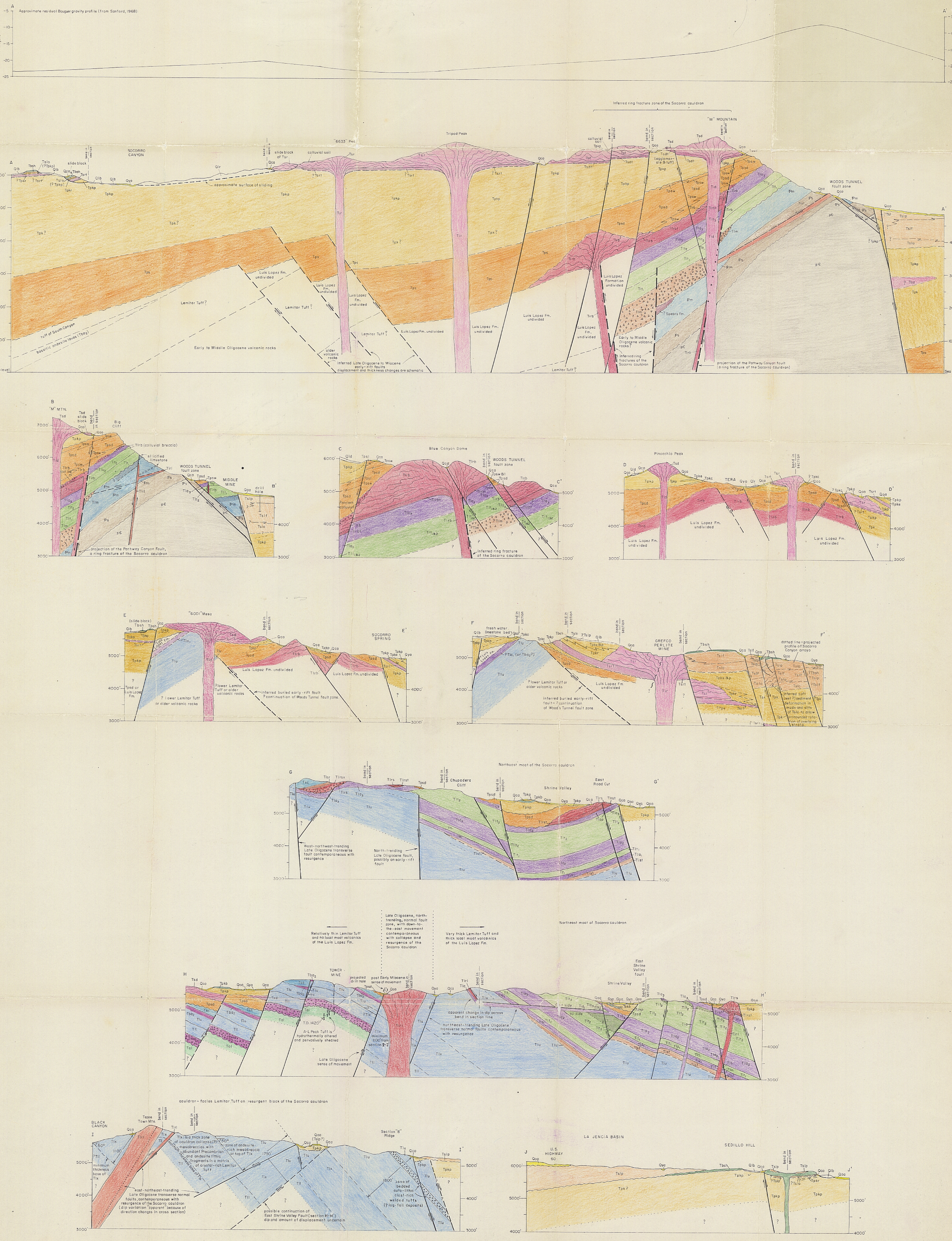
Volcanic vent

Inferred structural margin (outer limit of inferred major ring-fracture zone) of the Socorro cauldron The margin is largely buried by younger rocks, but is generally supported by gravity maps (Sanford, 1968) and aeromagnetic maps (U.S.G.S open file map NI 13-4).

118b

GEOLOGIC CROSS SECTIONS OF THE SOCORRO AND NORTHERN CHUPADERA MOUNTAINS, SOCORRO COUNTY, NEW MEXICO

by
Richard M. Chamberlin, 1980



Note: Contacts and faults dashed where uncertain; short-dashed lines indicate stratification

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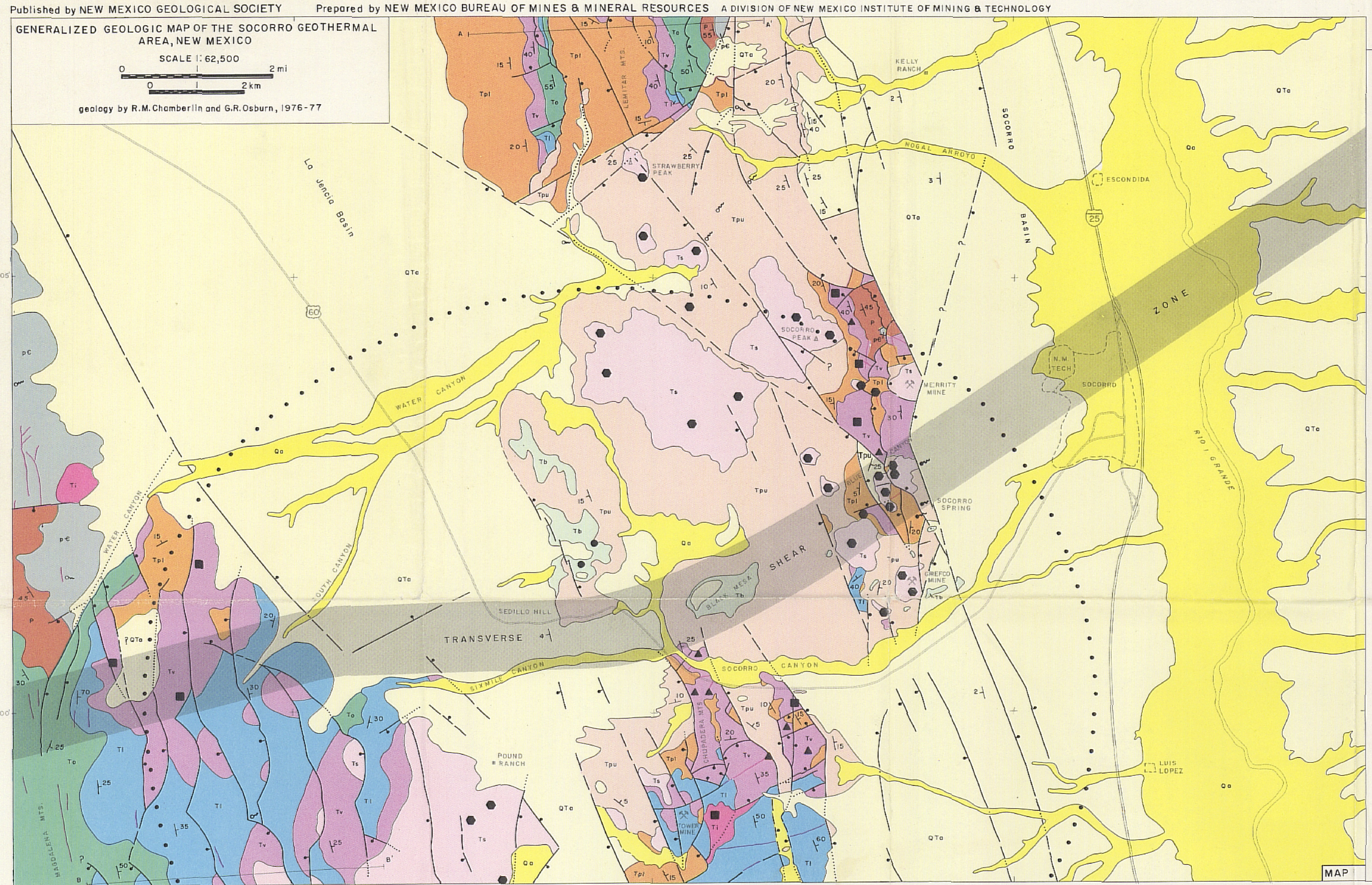
GEOLOGICAL AND GEOPHYSICAL MAPS OF THE SOCORRO GEOTHERMAL AREA, NEW MEXICO

by C.E. Chapin, R.M. Chamberlin, G.R. Osburn, A.R. Sanford and D.W. White, 1978

PLATE 1: NEW MEXICO GEOLOGICAL SOCIETY SPECIAL PUBLICATION NO. 7



EXPLANATION

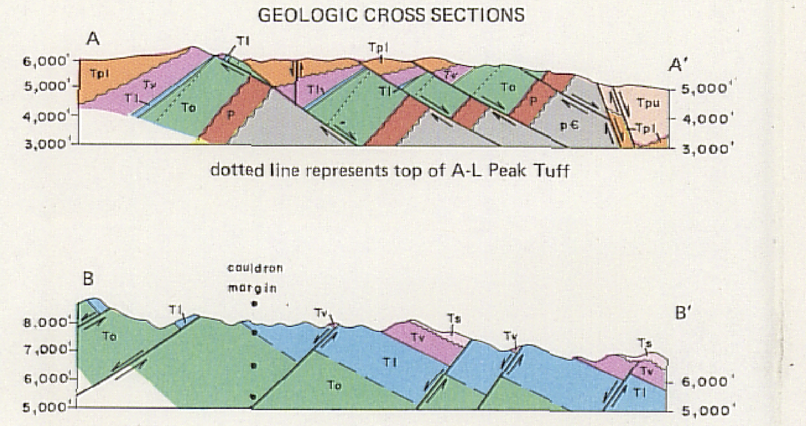
Volcanic vents:
- basalt of Sedillo Hill
- rhyolitic lavas of Socorro Peak
- rhyolite } related to the Socorro cauldron
- andesite } related to the Socorro cauldron
- inferred structural margin of the Socorro cauldron (mostly buried)
- normal fault
- downthrown side
- low-angle normal fault
- dike
- spring
- thermal spring
- 20 strike and dip of strata

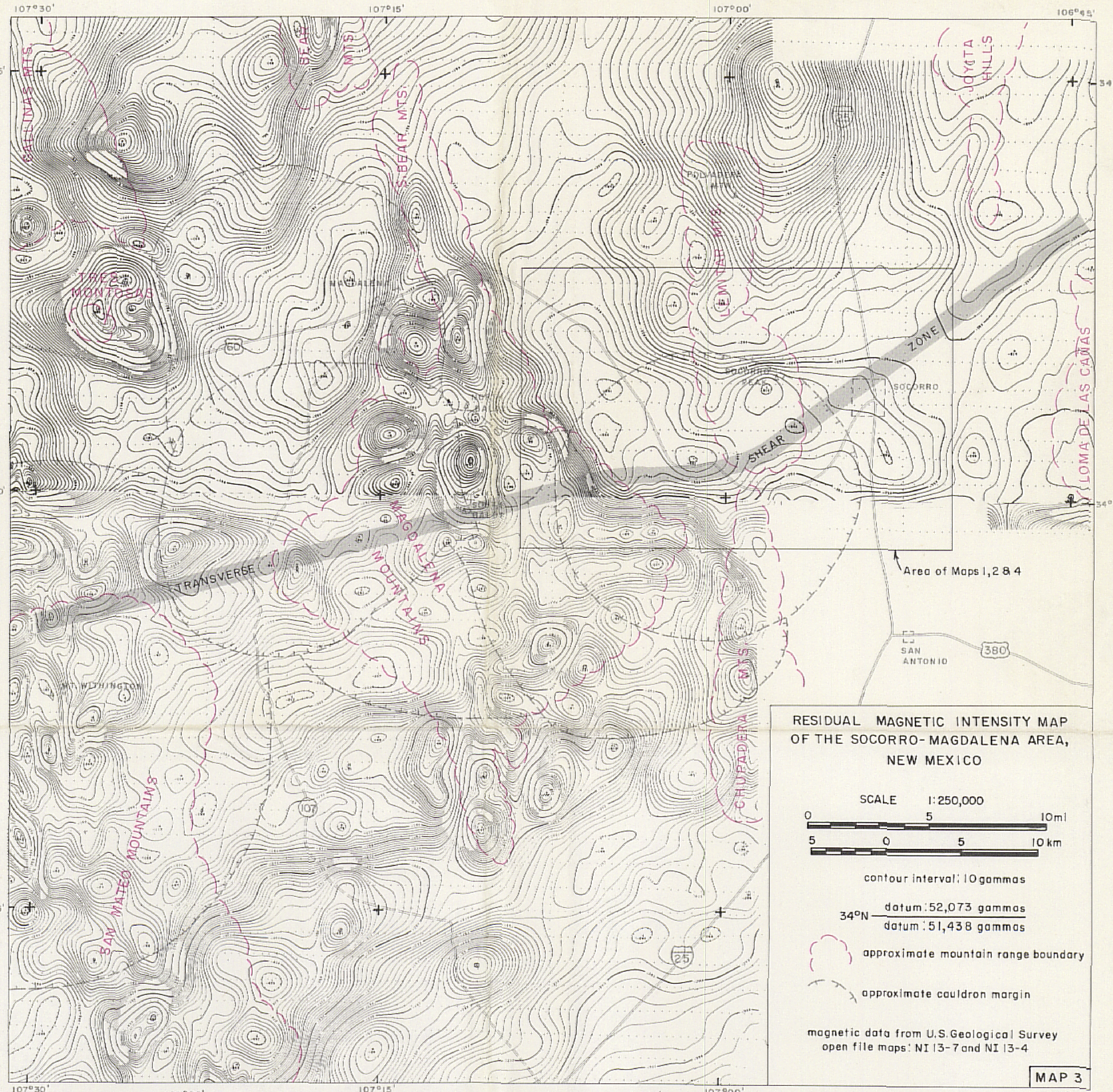
- **Qa** Upper Quaternary alluvium; gravel, sand and mud of major arroyos and the Rio Grande Valley, local alluvium and colluvium.

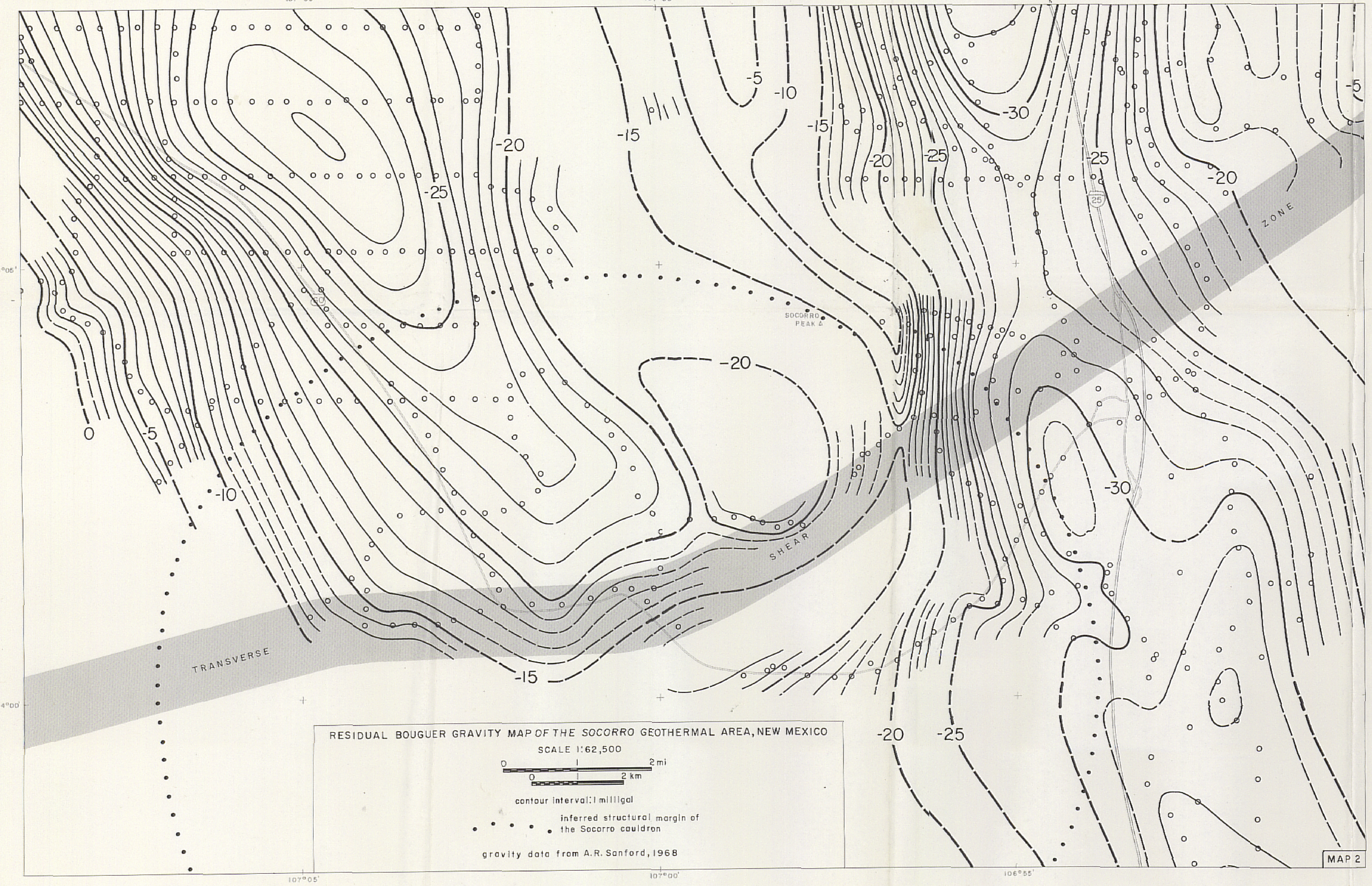
Santa Fe Group:
- **QTa** Plio-Pleistocene basin and valley fill (largely equivalent to Sierra Ladrones Fm.); poorly consolidated piedmont-slope fangls. intertonguing with friable ancestral Rio Grande sandst. and flood-plain siltst. and mudst., includes some post-Santa Fe Group terrace deposits.
- **Tb** Basalt flows of Sedillo Hill (4 m.y.); interbedded in QTa.
- **Ts** Rhyolite to rhyodacite domes, flows, necks and tuffs of Socorro Peak (12-7 m.y.); tuffs interbedded in upper Popotosa included in Tpu.
- **Tpu** Upper Popotosa Fm.; gypsiferous playa clayst. and mudst. with minor intertonguing fangls. and channel sandst., and several interbedded basalt flows (largely masked by landslides and colluvium).
- **Tpl** Lower Popotosa Fm.; well-indurated red mud-flow deposits and fangls. intertonguing with minor prpl.-gry. fangls. and lacustrine siltst. and mudst.

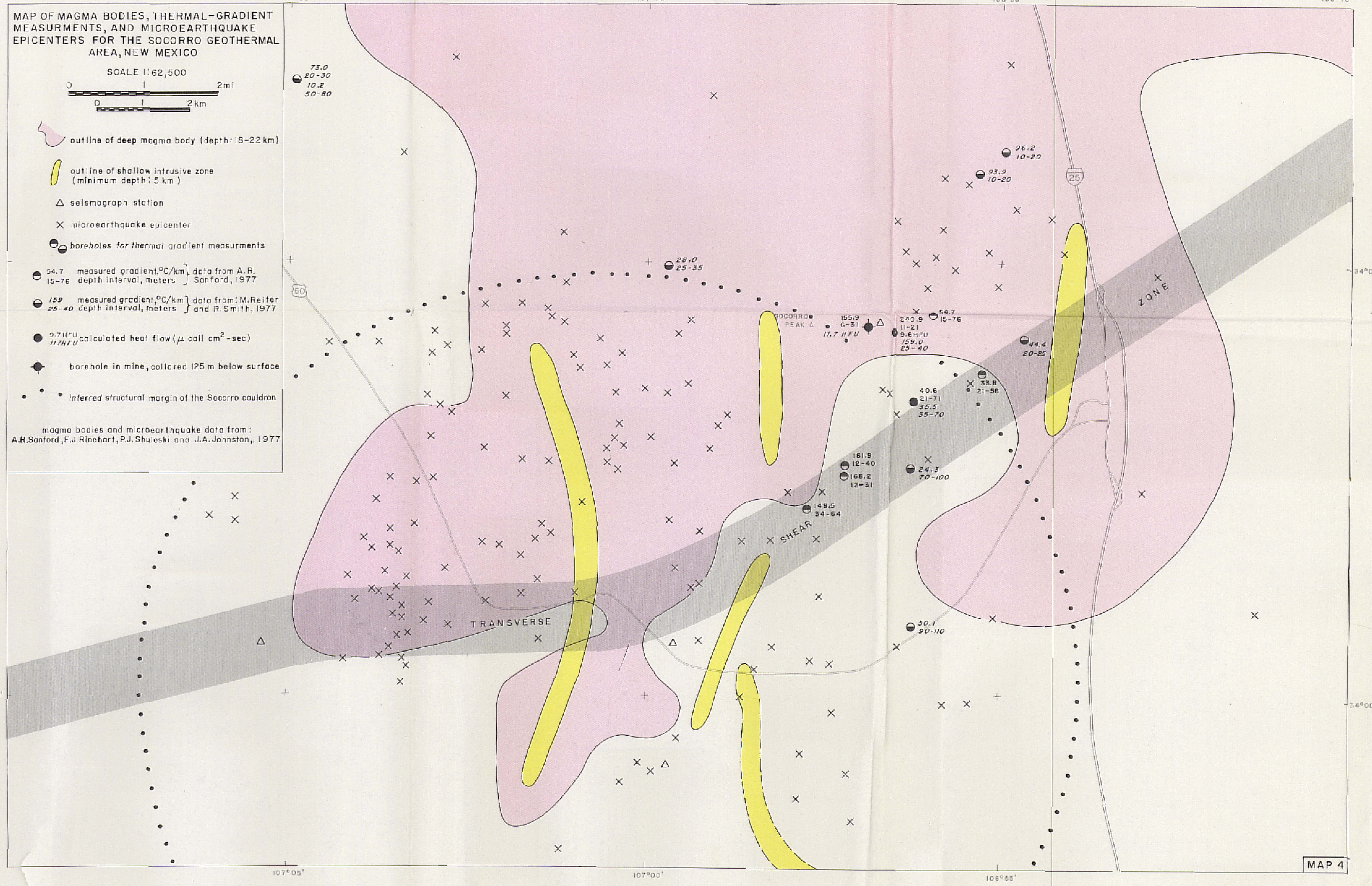
Other units:
- **Ti** Intrusive rocks; stocks and dikes of silicic to andesitic composition (Oligocene-Miocene).
- **Tv** Volcanic rocks post-dating tuff of Lemitar Mts. (26-20 m.y.); lithic-rich tuffs, andesite flows and rhyolite domes of Socorro cauldron moat, local intermediate lavas and rhyolite domes on Water Canyon Mesa, includes undiff. regional ash-flow tuffs and basaltic-andesite lavas in Lemitar and Magdalena Mts.
- **Tl** Tuff of Lemitar Mountains (27 m.y.) compositionally zoned, densely welded, rhyolite ash-flow tuff erupted from the Socorro cauldron.
- **To** Older volcanic rocks (37-32 m.y.) predating tuff of Lemitar Mts.; latitic-andesitic cgls. capped by thick sequence of densely welded, xl.-rich and xl.-poor rhyolite ash-flow tuffs and some associated moat-fill deposits.
- **P** Paleozoic limestones, shales and sandstones (Mississippian-Pennsylvanian)
- **pЄ** Precambrian granites, gabbros, diabase dikes, metavolcanic and metasedimentary rocks.

GEOLOGIC CROSS SECTIONS





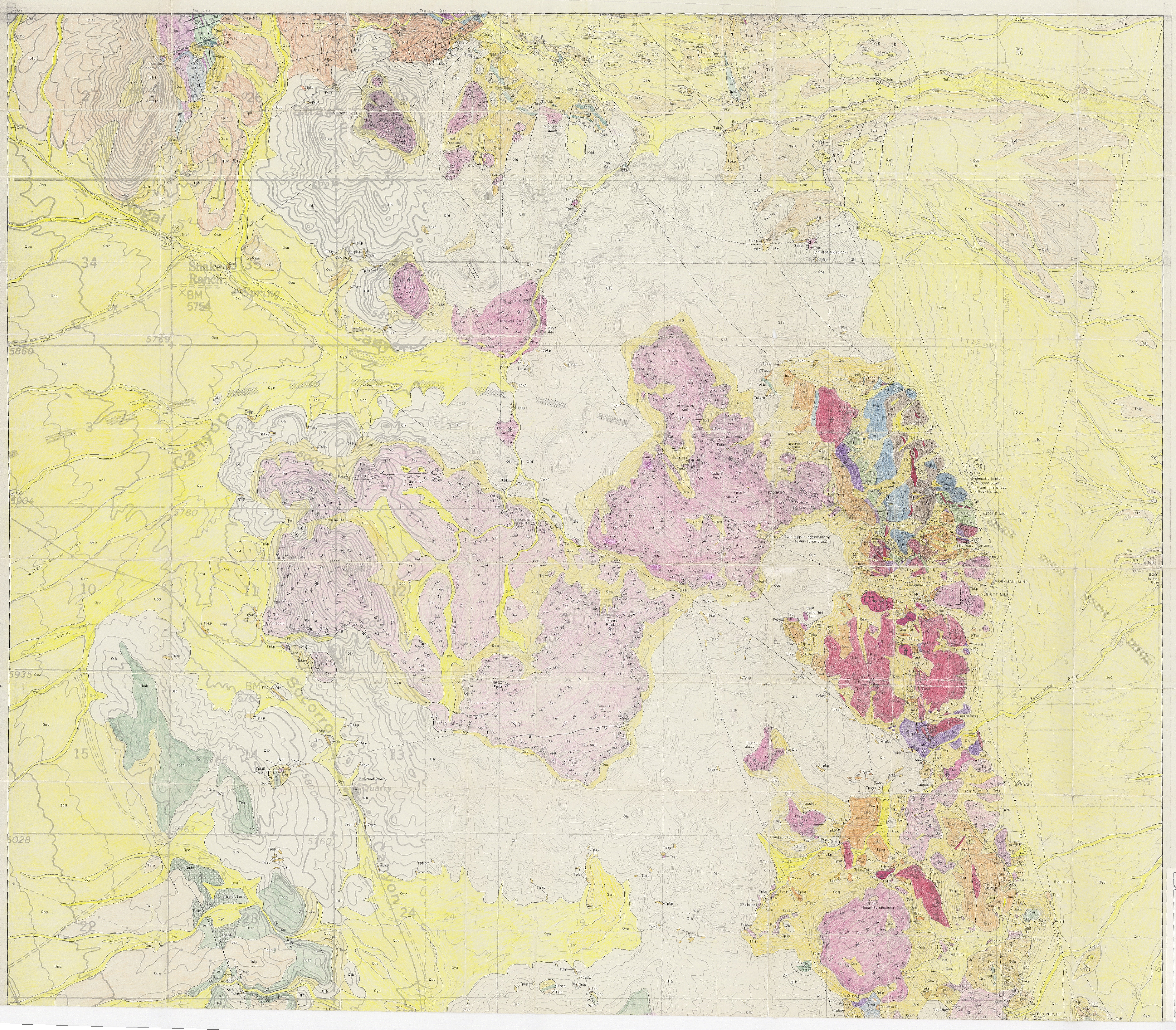




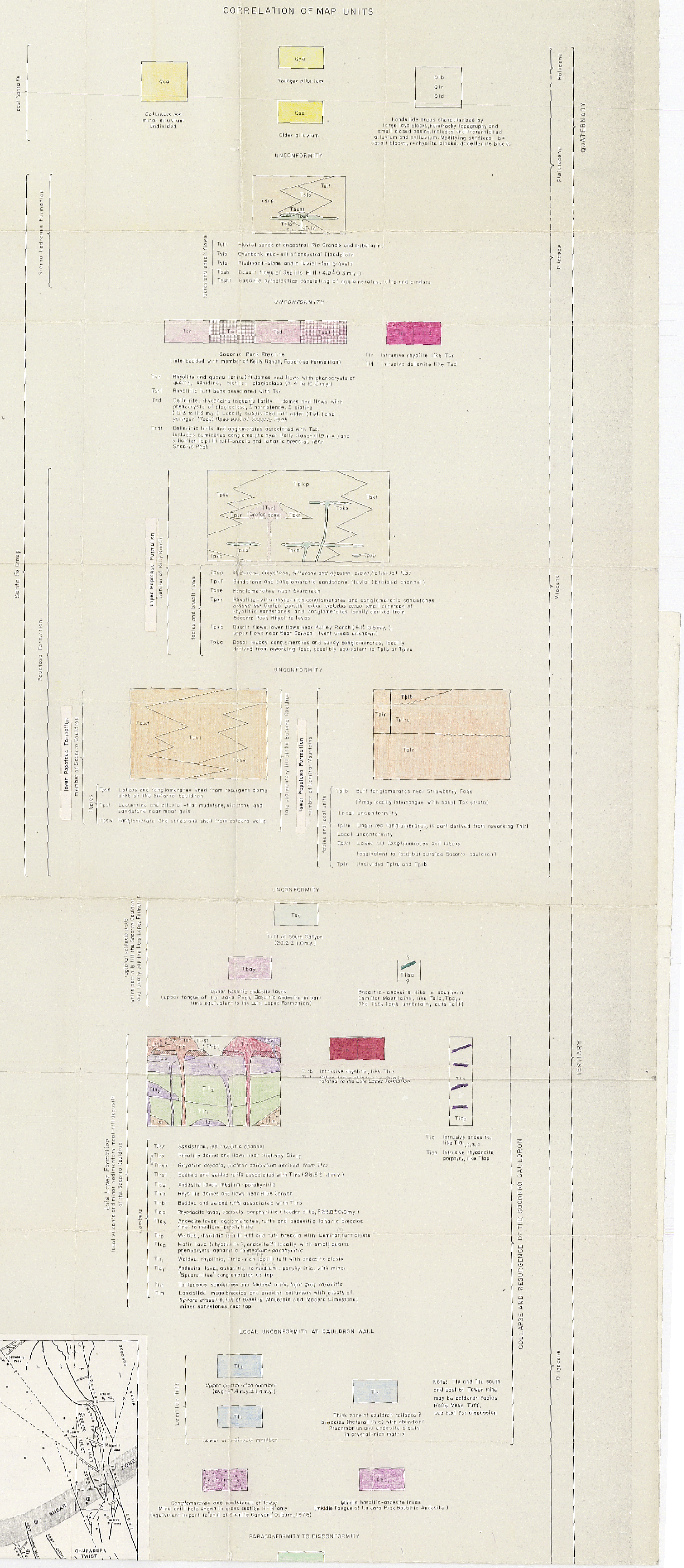
GEOLOGIC MAP OF THE SOCORRO, AND NORTHERN CHUPADERA MOUNTAINS, SOCORRO COUNTY, NEW MEXICO

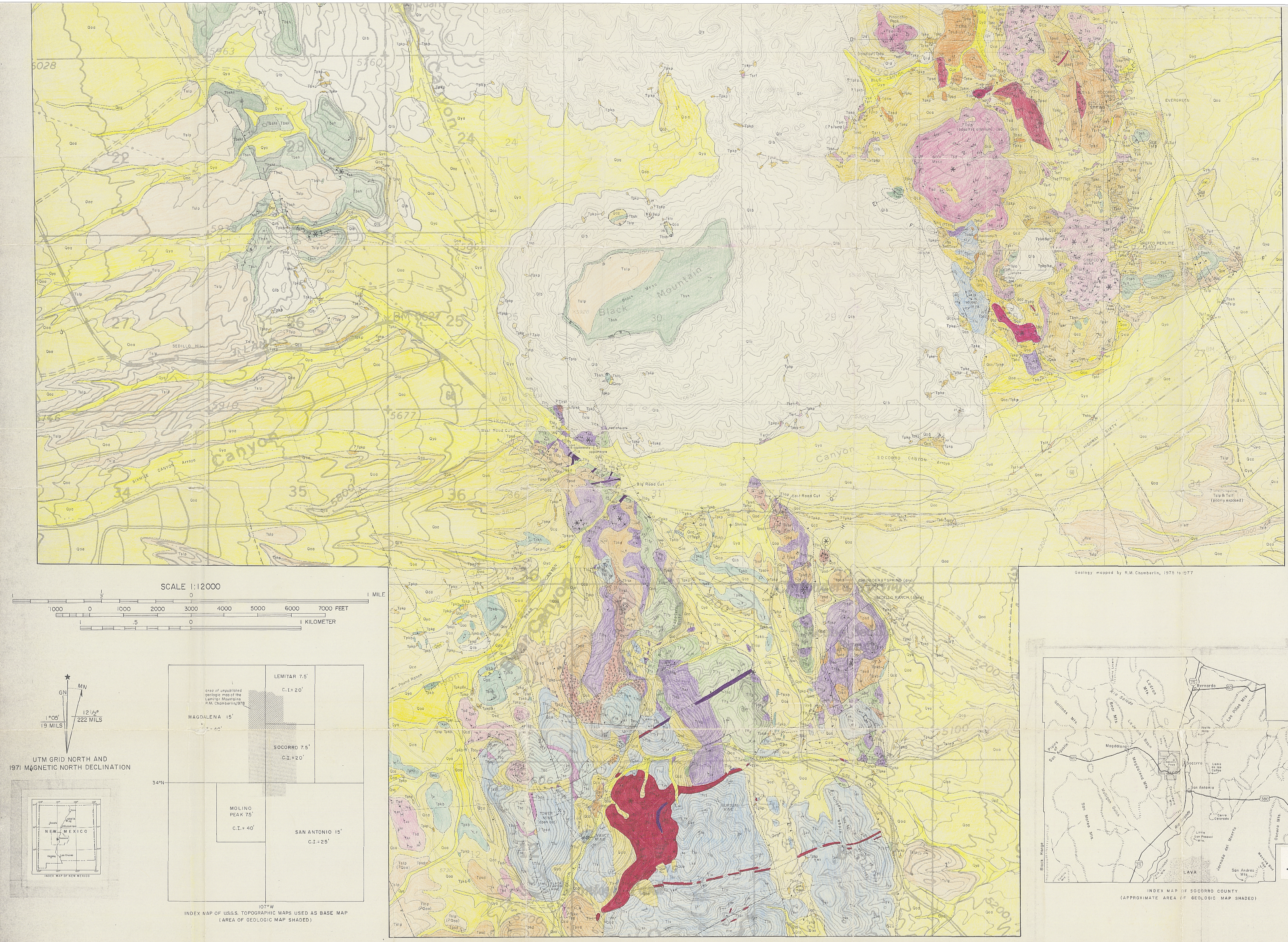
by

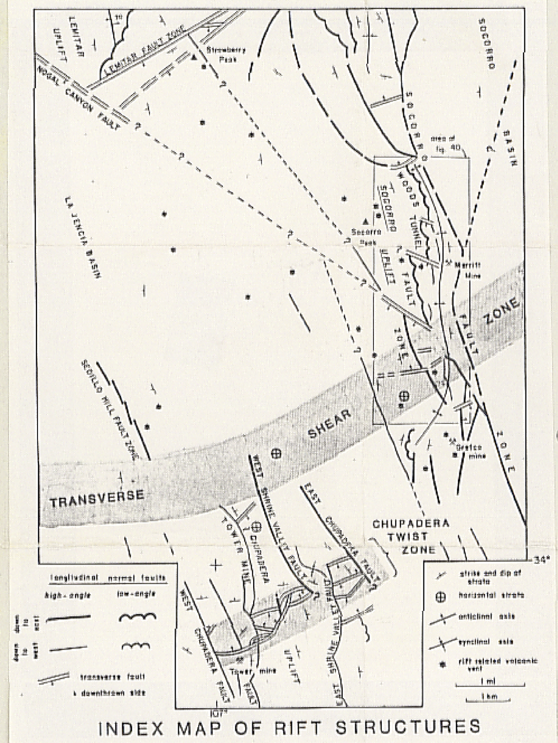
Richard M. Chamberlin, 1980



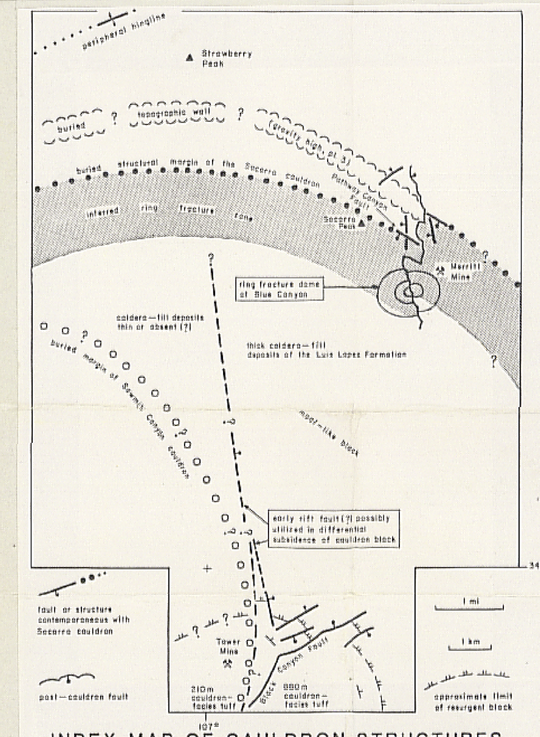
CORRELATION OF MAP UNITS



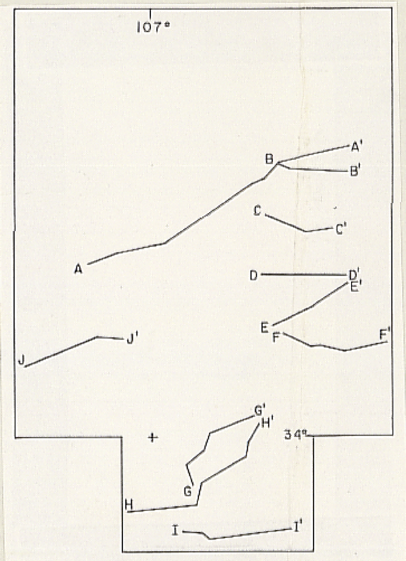




INDEX MAP OF RIFT STRUCTURES



INDEX MAP OF CAULDRON STRUCTURES



INDEX MAP OF CROSS SECTIONS

INDEX MAP OF SOCORRO COUNTY
(APPROXIMATE AREA OF GEOLOGIC MAP SHADED)

Tll_2 Welded, rhyolitic lapilli tuff and tuff breccia with Lemitar tuff clasts

Tla_2 Mafic lava (rhyodacite ?, andesite ?) locally with small quartz phenocrysts, aphanitic to slightly medium - porphyritic

Tll_1 Welded, rhyolitic, lithic-rich lapilli tuff with andesite clasts

Tla_1 Andesite lava, aphanitic to slightly medium - porphyritic, with minor "Spears-like" conglomerates at top

Tlsl Tuffaceous sandstones and bedded tuffs, light gray rhyolitic

Tlm Landslide mega breccias and ancient colluvium with clasts of Spears andesite, tuff of Granite Mountain and Madera Limestone; minor sandstones near top

COLLAPSE AND RESURGENCE O

Oligocene

LOCAL UNCONFORMITY AT CAULDRON WALL

Lemitar Tuff

Tlu — Upper crystal-rich member (avg: 27.4 m.y. ± 1.4 m.y.)

Tll — Lower crystal-poor member

Tlx — Thick zone of cauldron collapse ? breccias (heterolithic) with abundant Precambrian and andesite clasts in crystal-rich matrix

Note: Tlx and Tlu south and east of Tower mine may be caldera-facies Hells Mesa Tuff, see text for discussion

Tlrs — Conglomerates and sandstones of Tower Mine drill hole shown in cross section H-H only (equivalent in part to "unit of Sixmile Canyon," Osburn, 1978)

Tba₁ — Middle basaltic-andesite lavas (middle Tongue of La Jara Peak Basaltic Andesite)

PARACONFORMITY TO DISCONFORMITY

A-L Peak Tuff

Talp — Pinnacles member (?)

Tala — lower basaltic-andesite lavas and minor rhyolitic sandstones

Talf — Lower-lithic, mottled, gray-massive, flow-banded, and upper-lithic members undivided

NOTE: The following formations are present in the subsurface of the Strawberry Peak area and probably in the subsurface of the northern Chupadera Mountains: unnamed andesite lavas below the A-L Peak Tuff, Hells Mesa Tuff, tuff of Granite Mountain, and Spears Formation.

REGIONAL UNCONFORMITY

PENNSYLVANIAN

IPm — Madera Limestone

IPs — Sandia Formation

REGIONAL UNCONFORMITY

PRECAMBRIAN

p€m — Argillite with a monzonite intrusion exposed in Woods Tunnel

EXPLANATION

Contact long dashed where approximately located, short dashed where inferred, dotted where concealed

Thin ash bed or tuff interbedded in Santa Fe Group

Intraformational contact

strike and dip of stratification: horizontal; 46 inclined; vertical; 84 overturned

strike and dip of foliation: horizontal; 60 inclined; vertical; 69 overturned

Paleocurrent direction, from pebble imbrication where unlabeled, where labeled b = cross bedding, a = festoon trough axis, li = parting lineation, lo = log axis, i = inferred (in Qoa only), double headed where ambiguous

Paleoflow direction in volcanic rocks, as labeled s = stretch marks, v = elongated vesicles, p = pumice lineation, f = flow thrust

Pyroclastic strata associated with lavas, domes and local volcanic vents

Silicified rocks, most originally limestones

Breccias and bouldery deposits interpreted as ancient colluvium

mining symbols: Shaft; Bulldozer trench or open cut; Adit or portal; Prospect; Large pit or cut; Inclined diamond-drill hole showing collar and surface projection

Faults: all faults are long dashed where approximately located, short dashed where inferred, and dotted where concealed. Measured dip directions and angles indicated with arrows and numbers

High-angle normal fault, dip >45°, ball and bar on downthrown block, queried where hypothetical

Low-angle normal fault, dip <45°, "L" on downthrown block

Ring fracture or circumfrential fault related to the Socorro cauldron

Fault scarp (solid) and fault line scarp (dashed) in alluvium, hachures on downside

Epithermal or telethermal vein

Gravity-slide fault mostly toreva style slump blocks

Trace of plunging synclinal axis

A — A' Line of structure section

Pull apart structures related to slumping

T — Perennial spring, T = thermal spring

Earthen dam

Ancient landslide block of Madera Limestone in "Tlm" southeast of Socorro Peak

Volcanic vent

Inferred structural margin (outer limit of inferred major ring-fracture zone) of the Socorro cauldron. The margin is largely buried by younger rocks, but is generally supported by gravity maps (Sanford, 1968) and aeromagnetic maps (U.S.G.S. open file map NI 13-4).