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GEOLOGY OF THE CORKSCREW CANYON -- ABBE SPRING AREA, SOCORRO COUNTY, NEW MEXICO

by

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Abstract

Lithologic and paleontologic evidence suggests that two major transgressive-regressive marine cycles are preserved in the Cretaceous rocks of the study area. The units included in these cycles are the Dakota Sandstone to lower Tres Hermanos Sandstone and the upper Tres Hermanos Sandstone to lowermost Mesaverde Formation. The Eocene Baca Formation is gradationally overlain by the Oligocene Spears Formation which indicates that the study area is situated within an area of continual deposition from Baca into Spears time. Ash-flow tuffs in the study area are thin; two of them exhibit unconformable relationships with underlying units that are not observed to the south. These observations reflect the distance from source cauldrons, and possibly the influence of the Tijeras lineament.

The broad Abbe Spring anticline incorporates rocks as young as the Mesaverde Formation and was probably formed by Laramide compressional tectonics. Extensional faulting, beginning locally between 28 and 32 m.y. B.P., broke the area with numerous down-to-the-west normal faults. Many of the faults were intruded by mafic dikes. Faults with greater than 500 ft (152.4 m) of vertical displacement are paralleled on their downthrown sides by axes of narrow anticlines. These folds are most likely attributable to reverse drag effects. Late Oligocene-early Miocene block faulting caused

the Puertecito fault system which is the western border of the Mulligan Gulch graben. Activation of a transverse structural zone, the Tijeras lineament, during extensional faulting absorbed displacement on many of the extensional faults, and locally caused reversals of structural dips. The Mulligan Gulch graben is also offset to the west in two locations by this transverse structural zone.

Discontinuous coal beds occur at both the bottom and the top of the Mesaverde Formation. Coarse-grained sandstones of the Baca Formation locally contain carbonaceous material and uranium mineralization. Both of these resources have been locally developed on a small scale.

Introduction

Purpose of the Investigation

The objectives of this investigation are to determine the structural and stratigraphic relationships in the Corkscrew Canyon -- Abbe Spring area, Socorro County, New Mexico. These relationships are important for the following reasons:

1. The area is located along the poorly-defined but common margins of the Colorado Plateau, Rio Grande the rift, and the Datil-Mogollon volcanic field. Structural and stratigraphic data developed in this study will help to decipher the origin and evolution of these major tectonic and magmatic features.

2. The area is located along an outcrop belt of Cretaceous and Eocene formations that have potential for coal and uranium resources.

3. Oil tests have recently been drilled in the area. Evaluation of the oil and gas potential requires detailed stratigraphic and structural studies.

4. Mapping of this area will provide a link between studies in the adjacent Bear and Gallinas mountains.

Location

The study area is located about 15 mi (24.1 km) northwest of Magdalena, New Mexico, within the broad saddle between the north-northwest-trending Bear Mountains to the east and the northwest-trending Gallinas Mountains to the west. The study area is on the margin of the Rio Grande rift and within the boundary area between the Datil-Mogollon volcanic field and the Colorado Plateau. All but the northeastern corner of the study area is included within the Indian Spring Canyon 7.5-minute quadrangle. The northeastern corner extends into the Mesa Cencerro 7.5-minute quadrangle. Major arroyos -- Jaralosa Creek, Chavez Creek, and Abbe Spring Canyon -- drain northward into the Rio Salado which drains eastward into the Rio Grande. Fault-controlled springs -- among them Bird Spring (sec. 9, T1N, R5W), Abbe Spring (sec. 8, T1N, R5W), Montoya Spring (sec. 2, T1N, R6W), and Reid Spring (sec. 23, T1N, R6W) -- flow year-round.

Access

Main access routes into the study area are by NM 52, which leaves US 60 at the western edge of Magdalena, and by Forest Road 123, which leaves NM 52 ten miles (16.1 km) north of Magdalena. The dry bed of Jaralosa Creek and many dirt tracks provide access by four-wheel-drive vehicle to within 1.5 mi (2.4 km) of nearly every point in the study area. Almost three-quarters of the study area is privately owned.

Sections 11, 12, 14, and the northern half of 13 (T1N, R6W) are controlled by the Alamo Tribal Council.

Methods of Investigation

The surface geology of the study area was mapped on U.S. Geological Survey topographic maps of Indian Springs Canyon and Mesa Cencerro (1:24,000) during the summer and fall months of 1977. Two small, complex areas underlain by volcanic rocks were mapped on enlargments (1:12,000) of these maps, and were subsequently simplified in reduction. Black-and-white and color aerial photographs (1:31,680) from the U.S. Forest Service were utilized as guides to the location and configuration of outcrops and structures.

Twenty-eight thin sections were made from rock samples of the sandstone and volcanic units in the study area. These were examined using a Zeiss binocular petrographic microscope. Point counts of approximately 500 grains per slide were made of the sedimentary units using a Swift point counter with a grid spacing of 1 mm by 1 mm. Parenthetical references in the text refer to petrographic descriptions in Appendix II. Modal compositions of the volcanic rocks were estimated visually.

Previous Investigations

This study represents the first comprehensive geologic mapping of the Corkscrew Canyon -- Abbe Spring area utilizing recent advances in the stratigraphy of Cretaceous sedimentary rocks and Tertiary volcanic rocks. Herrick (1900) made the first geologic investigation of the area along the Rio Salado, then called Alamosa Creek, in which he described the Cretaceous rocks. Winchester (1920) produced the first geographic and geologic map which included the study area at a scale of 1:125,000. Darton's (1928) study of New Mexico's "Red Beds" and associated formations includes a geologic description and a sketch map which encompassed part of the study area. Tonking (1957) published a reconnaissance geologic map and report of the Puertecito 15-minute quadrangle. His study, which was excellent for its time, referred to the Cretaceous strata between the Dakota Sandstone and the Mesaverde Formation of this report as the La Cruz Peak Formation of the Mesaverde Group. Current workers have abandoned this terminology and recognize three tongues or members of the Mancos Shale overlain by the Gallego Sandstone (Hook, 1977, oral commun.). A field trip log through the study area based upon Tonking's report was published by Weber and Willard (1963). Snyder (1971), Johnson (1978), and Cather (in prep.) have studied the Baca Formation, including exposures in the study area. Brown (1972), Simon (1973), and Chamberlin (1974) have studied the volcanic rocks north and northwest of Magdalena. A composite stratigraphic column of volcanic rocks in the Socorro-Magdalena area has been published by Chapin and others (1978). A preliminary report on the coal, uranium, and oil and gas potential of the Riley-Puertecito area has been placed on open file by Chapin and others (1979). Figure 1 shows the location and relationship of the study area to other studies in the vicinity.

Acknowledgments

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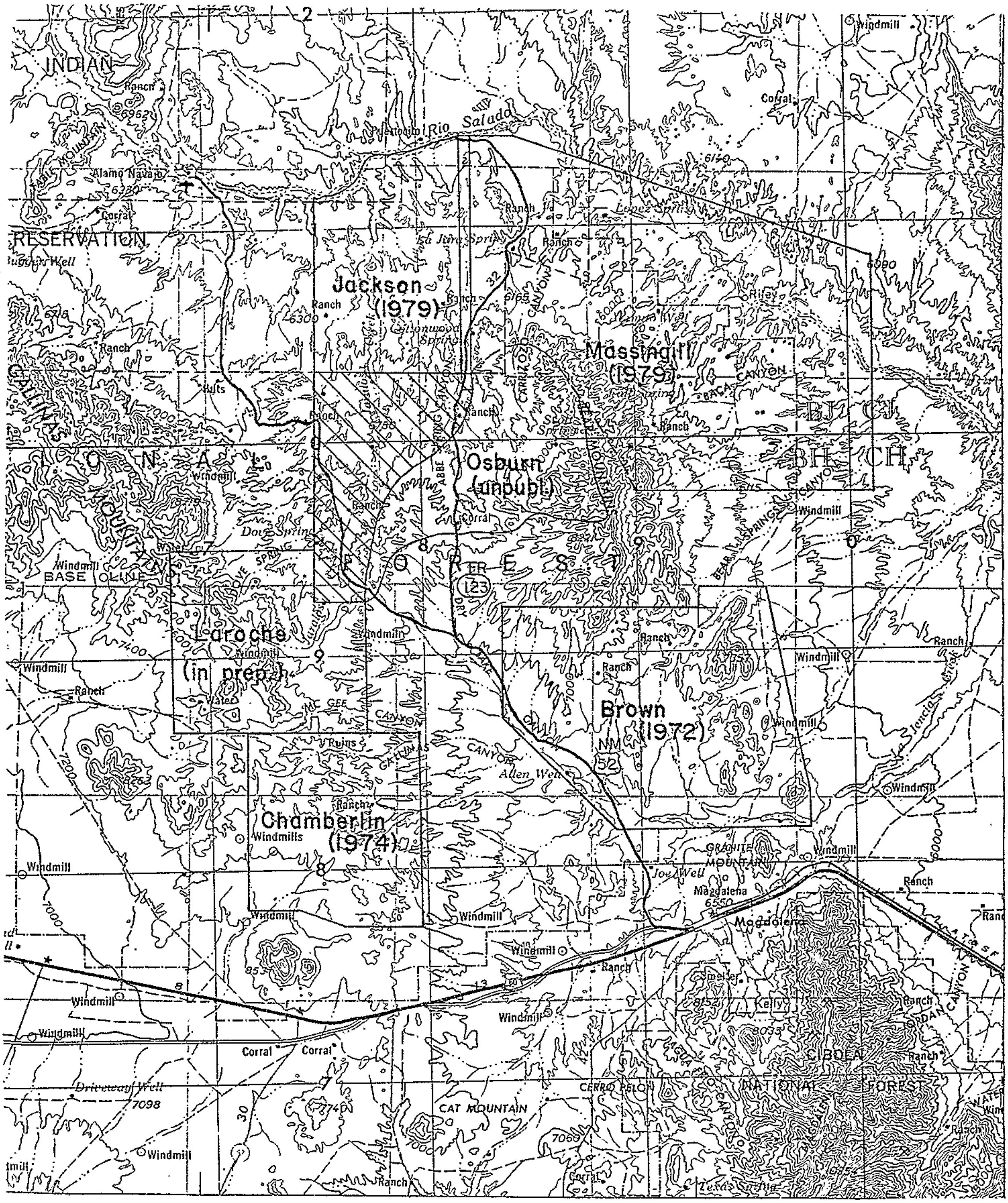


Figure 1 : RELATION OF STUDY AREA (RULED) TO ADJACENT AND NEARBY THESIS AREAS.

SCALE 1:250,000

Stratigraphy

Triassic

Chinle Formation

Winchester (1920) made reference to "Red Beds" which he believed to be of Triassic age and which which were unconformably overlain by the Dakota Sandstone. Wells (1919) referred to these rocks at exposures near Puertecito as the Puertecito Formation, but Darton (1928) believed that careful observation would reveal correlation between these rocks and Triassic rocks recognized in the Zuni uplift. Tonking (1957) first called these beds the Chinle Formation, the name given by Gregory (1917) to exposures of Triassic rocks in Chinle Valley, Arizona. The Chinle Formation exposed in the study area is probably Tonking's upper siltstone-shale unit, to which he ascribed a maximum thickness of 200 ft (61 m). Colbert and Gregory (in Reeside and others, 1957) report a middle Upper Triassic age for the Chinle Formation from land vertebrate fossils found in northern and central New Mexico.

The Chinle Formation in the study area is a slope-forming unit composed dominantly of maroon and variegated mudstones, siltstones, and shales. Petrified wood fragments were found in float at one locality. The Chinle is poorly exposed beneath hogbacks formed by the overlying resistant Dakota Sandstone in the northeastern part of the study area (sec. 5, T1N, R5W). The maximum thickness of the

Chinle exposed in the study area is about 12Ø ft (36.6 m) and occurs on the west-facing slope of hill "6687" (E 1/2, NW 1/4, sec. 5).

The upper contact of the Chinle Formation is poorly exposed due to slumping of blocks of Dakota Sandstone following erosion of the underlying Chinle siltstones. Where exposed, the top of the Chinle is scoured and eroded. The uppermost Chinle Formation varies from a very-dark-red (5R2/6) siltstone in the northernmost exposures to a dark-gray (N3) silty shale or a light-gray (N7), weathering to dark-yellowish-orange (1ØYR6/6), siltstone in the southernmost exposures. Angular unconformity with the overlying Dakota Sandstone is not obvious, although it is evident on a regional scale (Givens, 1957).

The Chinle Formation is dominantly dusky-red (5R3/4) to very-dark-red (5R2/6) calcareous siltstones and thinly laminated silty shales. These rocks are often variegated in shades of red and light gray. This variegation is especially prominent in nodular beds 1.5- to 4-in. (3.8- to 1Ø.2-cm) thick. The nodules are very calcareous and ellipsoidal, with an average long diameter of 1.5 in. (3.8 cm). They range in color from very dark red (5R2/6) through light gray (N7).

A few thin-bedded limestones occur near the top of the Chinle Formation. These limestones are commonly grayish red (1ØR4/2) and mottled by calcareous areas of pale green (5G7/2). A typical limestone (Trc-1) consists of 25% silt grains, 45% microspar, and 30% hematitic-limonitic stained clays. Porosity is essentially absent due to infilling by blocky calcite cement. Grain-size distribution is bimodal. One mode has an average apparent grain diameter of 0.05 mm and is comprised of well-sorted, subangular, elongate to intermediate, rutilated quartz with undulose extinction and slightly sericitized potassium feldspar. The other mode, which comprises 85% of the detrital components, has an average apparent grain diameter of 0.6 mm. It is comprised primarily of mudstone and hematitic mudstone fragments. The mudstone fragments range in size from 0.2 mm to 1.5 mm, and are generally subangular and elongate in shape. They are composed of 65% clay minerals (with as much as 5% replacement by chlorite); 20% clear, angular quartz grains with undulose extinction; 7% hematite; 3% felted muscovite replacing clay minerals; 1% angular, sericitized potassium feldspar; and 3% patchy calcite cement. Chapin (1979, oral commun.) asserts that the mudstone fragments represent areas of an original lithology which subsequently was largely calcified.

Three stacked, channel-shaped sandstone bodies are exposed in the top 20 ft (6.1 m) of the Chinle Formation in La Jara Canyon (SW 1/4, NW 1/4, sec. 5; fig. 2). The sandstone is moderate yellowish brown (10YR5/4), weathering to grayish orange (10YR7/4), medium grained, moderately indurated, and calcareous. Petrographically (Trc-2), the sandstone is poorly sorted and comprised of 35% framework grains, 15% patchy calcite cement, 3% porosity, and 47%

Figure 2: Channel sandstones in the Chinle Formation; La Jara Canyon (S 1/2, NW 1/4, sec. 5, T1N, R5W). Hammer handle is 15 in. (38 cm) long.

matrix. Untwinned, partially sericitized potassium feldspar comprises 40% of the framework grains; quartz comprises 18% of the framework components. Polycrystalline quartz grains are slightly more abundant than grains of monocrystalline, clear quartz with undulose extinction. Lithic fragments, which constitute at least 24% of the framework components, are soft masses comprised of 93% clays and sericite and 7% subangular corroded potassium feldspar anhedra. The matrix material of clay and sericite, and at least some of the potassium feldspar, may be derived by deformation of mudstone lithic fragments. Other framework components include subrounded, slightly elongate chert (5%); ragged muscovite laths (4%); and fresh calcic plagioclase.

The portions of the Chinle Formation exposed in the study area are nonmarine floodplain deposits (Tonking, 1957). The nonmarine origin is substantiated by observations of vertebrate remains (Tonking, 1957), petrified wood, the red color of the sediments, and the absence of marine fossils. Nodular beds may represent areas of subaerial exposure to desiccating conditions (Reineck and Singh, 1975).

Upper Cretaceous

Dakota Sandstone

The Dakota Sandstone, the name given by Meek and Hayden (1862) to the basal Cretaceous sandstone unit near Dakota, Nebraska, is now the accepted identity of the basal Cretaceous sandstone in central New Mexico. Earlier geologists in New Mexico used the name "Dakota" with a suffixed query to imply this probable correlation, a practice which recently has been discontinued. Tonking (1957) observed that the Dakota Sandstone unconformably overlies progressively older beds southward. To the north of the study area, the Dakota Sandstone unconformably overlies the Morrison Formation of Jurassic age. Southward, including the study area, it overlies progressively older beds of the the Triassic Chinle Formation.

Lee (1915) postulated that the basal Dakota Sandstone was deposited in a nearshore environment over a base-leveled surface. Long periods of wave sorting and reworking of sediments produced the characteristic well-sorted quartzose sandstone and quartzite conglomerate which make up the Dakota throughout the Rocky Mountain area. The Dakota Sandstone in northwestern New Mexico is a formation of late Early Cretaceous and early Late Cretaceous ages (Dane and Bachman, 1957). It is comprised of shale, sandstone, conglomeratic sandstone, conglomerate, and coal, representing deposits of marine, marginal marine, and continental origins. Young (1960) proposed the elevation of the Dakota to group status with two subdivisions for its occurrences on the Colorado Plateau.

The Dakota Sandstone is a resistant, well-indurated, quartzose sandstone with thin conglomeratic beds and lenses which contain pebbles of quartzite and chert. It is a cliff-forming unit and crops out as a series of gently dipping hogbacks in the northeastern corner of the study area (secs. 5 and 6, T1N, R5W). The hogbacks are formed by eight down-to-the-west and three down-to-the-east normal faults which repeat the Dakota Sandstone and the top of the underlying Chinle Formation. The locations of the fault planes are generally well-defined by breccia and slickensides in the brittle Dakota Sandstone (fig. 3). The Dakota Sandstone in the study area is estimated to be about 30 ft (9.1 m) thick.

The contact with the underlying Chinle Formation is poorly exposed due to erosion of the soft Chinle shales and siltstones, and the resultant slumping of Dakota Sandstone cliff faces. (These slump blocks are more extensive than are shown on the geologic map; many have been omitted due to limitations of scale.). Where exposed, the top of the Chinle clearly has been eroded and scoured. The base of the Dakota Sandstone is commonly marked by an irregular 0 to 7 in. (17.8 cm), yellowish-gray (5Y7/2), deeply weathered conglomeratic sandstone, containing well-rounded gray and white quartzite and chert pebbles and 2-in. (5.1 cm)-long white siltstone

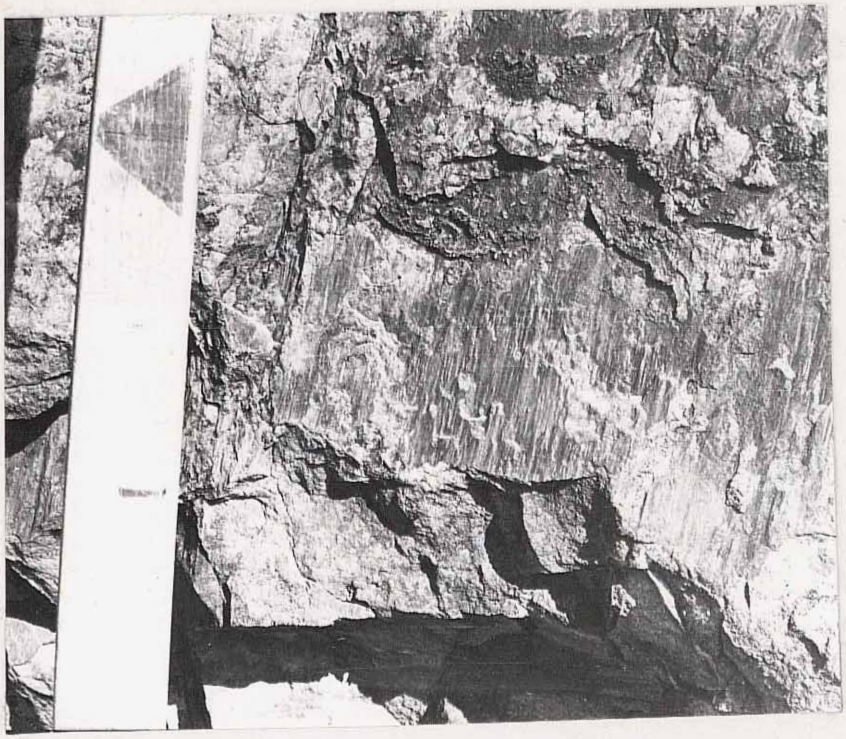


Figure 3: Slickensides along a fault cutting the Dakota Sandstone; Abbe Spring anticline (NW 1/4, SE 1/4, sec. 5, T1N, R5W). Jacob's staff graduation is 6 in. (15.2 cm).

chips weathered and eroded from the top of the Chinle. The contact appears to be conformable, although it is unconformable on a regional scale (Givens, 1957). The contact of the Dakota Sandstone with the overlying Alamito Well tongue of the Mancos Shale is exposed only on the eastern slope of the most southward-extending Dakota hogback (SE 1/4, sec. 5). The contact is poorly exposed but appears conformable and gradational.

The main body of the Dakota Sandstone is greenish gray (5GY6/1), and weathers grayish yellow (10R7/4) or moderate yellow brown (10YR5/4). The exposed face is often covered with a layer of dark-reddish-brown (10R3/4) to black (N1), shiny desert varnish. On some exposures, dusky-red (5R3/4), angular hematite plates about 30 mm across coat the exposed surface. Red, round to oblong, ferruginous sandstone nodules weather to positive relief and occur commonly. Liesegang banding is also a common surficial feature. Small dark particles (presumably hematite) on many exposures highlight thin planar laminae and 6-in. (15.2 cm)-thick tabular sets of straight, high-angle cross-laminae. These frequently change upwards in the unit to large-scale, trough-shaped cross-laminae. Wood casts as much as 1 in. (2.5 cm) thick by 12 in. (30.5 cm) long also occur (fig. 4). Grain size is medium to fine sand; the sand grains are well sorted to very well sorted throughout the unit. Sand grains in hand specimen appear rounded and frosted.

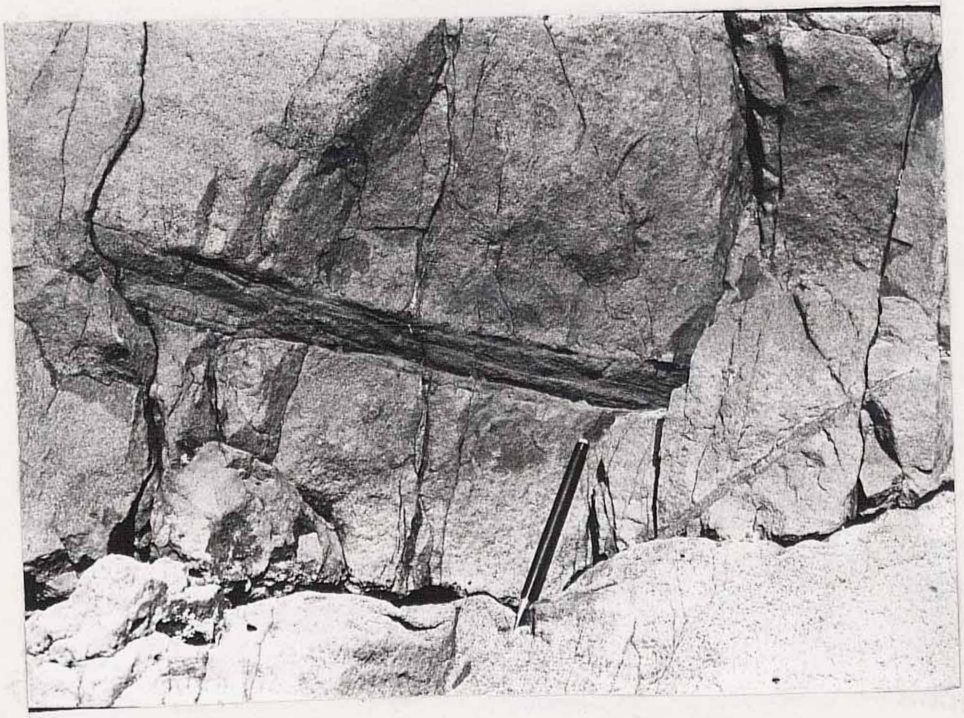


Figure 4: Wood casts in Dakota Sandstone; Abbe Spring anticline (NE 1/4, SW 1/4, sec. 5, T1N, R5W). Pencil is 5.5 in. (14 cm) long.

Petrographically, the Dakota Sandstone (Kd-1) is comprised of rounded, well-sorted, predominantly monocrystalline quartz grains with an average apparent grain diameter of Ø.3 mm. The grains are cemented by quartz syntaxial rim cement. The cement overgrowths form an interlocking mosaic of quartz crystals, producing a rock with an estimated 3% porosity. The detrital quartz grain outlines can usually be discerned under low light with the nicols uncrossed by virtue of a thin dusting of hematite, limonite, or clay around their peripheries. These grains comprise 9Ø% of the framework components of the rock. The majority of the quartz grains are slightly undulose. Common identifiable inclusions which occur in the majority of the detrital quartz grains are zircon subhedra, muscovite shreds, and randomly distributed bubbles. Well-rounded detrital zircon is also observed in trace amounts. The detrital lithic fragments, which constitute 3% of the framework components, are comprised primarily of subangular quartzite. One well-rounded lithic fragment of claystone and one rounded lithic fragment of felted muscovite with a limonitic center also occur.

The conglomeratic lenses and beds have maximum thicknesses of 4 in. (1Ø.2 cm) and average about 2 in. (5.1 cm) thick. The pebbles are subrounded and polished gray and white chert and quartzite, averaging about Ø.25 in. (Ø.64 cm) in diameter. A cast of a small crinoid columnal was observed on one white chert pebble.

Petrographically, the conglomeratic lenses and beds (Kd-2) are comprised of about 82% detrital grains, 15% quartz syntaxial rim cement, and 3% pores. Monocrystalline, rounded, moderately sorted quartz grains with undulose extinction comprise 50% of the detrital grains. About 5% of these quartz grains exhibit small nuclei which are not in optical continuity with the rest of the grain; 15% of the quartz grains exhibit tiny internal cracks. Inclusion mineralogies are similar to those observed in the quartz grains of the sandstone beds. Cloudy, partially sericitized potassium feldspar comprises another 5% of the framework components. Lithic fragments and chert, comprising 40% of the detrital components, are very poorly sorted and subrounded. The dominant lithology is chert, which is commonly cut by veins of drusy and blocky, undulose, sutured quartz crystals. Other lithologies, in order of decreasing abundance, are slightly sericitized arkosic wacke (Pettijohn and others, 1975), metaquartzite, and quartz-muscovite schist.

The top 3 to 5 ft (0.9 to 1.5 m) of the Dakota Sandstone crops out as an uneven soil-covered slope. Medium-dark-gray (N4) shale float found here suggests that there may be some thin shale beds intercalated with the sandstones. These sandstones are fine-grained and occur in 4-in. (10.2 cm)-thick planar beds.

The Dakota Sandstone is commonly assumed to have been deposited by a transgressing sea. A nearshore environment is indicated by wood casts, and the absence of marine fossils and bioturbation. The conglomeratic beds and lenses may represent small fluvial or tidal channels, or sediments introduced by storms.

The Dakota Sandstone has been recognized as a good hydrocarbon reservoir rock in the San Juan Basin since the first discovery of gas in 1921 (Reese, 1952; Burton, 1955). Uranium production from this formation is reported as minor (Smith, 1955).

Mancos Shale

The Mancos Shale was first named by Cross (1899) for a thick shale body near Mancos, Colorado. Its original definition limited it to the stratigraphic interval between the Dakota Sandstone and the Mesaverde Formation. The Mancos Shale is thus a lithologic unit that transcends time lines throughout its areal extent. The Mancos Shale thins southward from its type locality and is known to intertongue with the basal Mesaverde Formation (Hunt, 1936; Dane and others, 1957).

In the study area, the Mancos Shale consists of two shale tongues -- the Alamito Well and D-Cross tongues -- separated by the Tres Hermanos Sandstone Member. Total thickness of this formation is about 850 ft (259 m). Above the Mancos Shale in the study area is the Gallego Sandstone, presently a unit of unknown extent but thought to belong in the Gallup Sandstone, which is elsewhere part of the Mesaverde Group (1978, Hook, oral commun.).

Alamito Well tongue

The Alamito Well tongue of the Mancos Shale, defined as the shale interval between the underlying Dakota Sandstone and the overlying Tres Hermanos Sandstone Member, was named at exposures in Canon del Alamito (sec. 20, T2N, R4W) by Massingill (1979). Massingill reports a thickness of about 556 ft (169.5 m) at the type section, which includes 7 ft (2.1 m) of sills and 81 ft (24.7 m) of cover. To the north (Jackson, 1979) and northeast (Massingill, 1979) of the study area, the Alamito Well Tongue is split by the Twowells Tongue of the Dakota Sandstone. The stratigraphic intervals below and above the Twowells Sandstone are called the Whitewater Arroyo and the Rio Salado tongues of the Mancos Shale respectively. An 8-foot (2.4 m)-thick sandstone recognized on electric logs of drillhole H2 and H3 (Transocean Oil Co., Houston, TX) occurs between 195 and 210 ft (59.4 and 64.0 m) below the Tres Hermanos Sandstone and may be the equivalent of the Twowells Tongue of the Dakota Sandstone. However as this sandstone is not observed at the surface within the study area, the Whitewater Arroyo-Twowells-Rio Salado nomenclature has not been used. The Alamito Well tongue is also present in the Carthage area southeast of the study area (Massingill, 1978, oral commun.).

Outcrops of the Alamito Well tongue are limited to the northeast corner of the study area (secs. 5, 6, and 8; T1N; R5W). Unfortunately, this is the area of intersection of the Puertecito fault system, forming the western border of the Mulligan Gulch graben, and a northeast-trending zone of faults and dip reversals related to the Tijeras lineament. Complex faulting, igneous intrusion, and erosional dissection in this area make the task of a thorough stratigraphic description difficult. Observations are mainly limited to the basal 20 ft (6.1 m; SW 1/4, SE 1/4, sec. 5, north of Tijeras lineament), the top 100 ft (30.5 m; north and east slopes of Tres Hermanos hogbacks), and an intruded 60-foot (18.3 m)-thick section exposed along La Jara Creek presumably nearer to the base than to the top (sec. 5, T1N, R5W).

The base of the Alamito Well tongue is gradational with the underlying Dakota Sandstone. This contact was delineated at the first medium-light-gray (N6) shale unit of approximate 2-foot (0.61 m) thickness. Exposures in this area are extremely poor, but float suggests several thin, very-fine-grained, moderate-brown (5YR3/4), silty sandstones may occur above this basal shale. The dominant lithology, however, is dark-gray (N3) to grayish-black (N2), fissile, calcareous, silty shale.

The abrupt upper contact of the Alamito Well tongue with the overlying Tres Hermanos Member is well-exposed on the east slope of hill '6597' (sec. 6). The dominant lithology of the Alamito Well tongue again is shale, now medium light gray (N6). Mammites depressus was found within limestone concretions in these upper shales exposed along the east side of Abbe Spring Canyon. The concretions average 1.5 ft (0.46 m) in diameter, and are smooth, dark-yellowish-orange (10YR6/6)-weathering ellipsoids. Sciponoceras gracile was collected from a medium-gray (N5) limestone bed approximately 4 in. (10.2 cm) thick and about 90 ft (27.4 m) below the Tres Hermanos Member at the same location. Occurrences of Pycnodonte newberri are also reported from this limestone (Massingill, 1977, oral commun.).

The section of the Alamito Well tongue exposed along La Jara Creek is comprised of interbedded dark-gray (N3), calcareous shales and grayish-olive (10Y4/2)-weathering, very-fine-grained, massive sandstone. The shale units are between 6 in. (15.2 cm) and 3 ft (0.9 m) thick; the sandstones range from 2 in. (5.1 cm) to 12 in. (0.3 m) thick. Contacts between these lithologies are relatively sharp.

The Alamito Well tongue consists of low-energy, offshore shelf deposits. Included within this record is a change from marine transgression at the base to marine regression at the top; shoreline sands are preserved below and above this unit as the Dakota Sandstone and Tres Hermanos Member, respectively. Interbedded sandstone beds may represent storm-generated deposits (Reineck and Singh, 1975).

The *Sciponoceras gracile* zone is a widespread ammonite zone and marks the boundary between Cenomanian and Turonian time (89 m.y.; Obradovich and Cobban, 1975). *Pycnodonte newberri*, found at the same stratigraphic location as *S. gracile*, is found only in the Four Corners states (Hook and Cobban, 1977).

Tres Hermanos Sandstone Member

The Tres Hermanos Sandstone Member of the Mancos Shale was named by Herrick (1900) for exposures near Tres Hermanos Buttes (sec. 26, T3N, R7W). Tonking (1957) misinterpreted Winchester's (1920) stratigraphic section south of Puertecito, and mistakenly identified the Twowells Tongue of the Dakota Sandstone as the Tres Hermanos Member of the Mancos Shale. The Tres Hermanos Member of this report was included within Tonking's La Cruz Peak Formation. This mistake was corrected by Dane, Landis, and Cobban (1971). They also reported an early Carlile age for the Tres Hermanos Member based upon recovery of *Collignonceras woollgari*. Massingill (1979) reports a thickness of approximately 231 ft (70.4 m) to the northeast of the study area.

The Tres Hermanos Member crops out in the northeastern section of the study area, and again west of Jaralosa Creek and south of Dove Spring Canyon. Exposures are generally poor due to slumping and extensive erosion. In addition, all outcrops of the Tres Hermanos in the study area are faulted, and in the northeastern section, extensively intruded. The Tres Hermanos is estimated to be 250 ft (76.2 m) thick in the study area.

The basal contact of the Tres Hermanos Member with the Alamito Well tongue is well-exposed on the east side of hill '6597' (sec. 6, T1N, R5W). The upper contact with the overlying D-Cross Tongue is well-exposed in Jaralosa Creek west of drillhole H1 (sec. 35, T1N, R6W). Both of these contacts are marked by sharp lithologic changes. The Tres Hermanos Member is informally divided into three gradational units by the occurrence of a nonmarine interval of organic-rich sandstones, siltstones, shales, and lignite between upper and lower intervals of marine sandstone containing minor intercalated shale.

The lower marine section forms gently dipping hogbacks in the northeastern corner of the study area. This section is estimated to be about 80 ft (24.4 m) thick. Thin- to medium-bedded sandstone is dominant in this section. Sandstone colors on a freshly broken surface are yellowish gray (5Y8/1), light greenish gray (5GY8/1) or very light gray (N7). Colors on weathered surfaces are very pale orange (10YR8/2) or pale yellowish brown (10YR6/2), at times mottled by dark yellowisn orange (10YR6/6). Massingill (1979) reports <u>Collignonceras (Selwynoceras) mexicanum (Bose)</u> from the base of this section. Fossil debris is common in the sandstones near the top of the lower marine sandstone unit. High-angle, small-scale, foreset crossbeds and asymmetric ripples are common. Framework grains are medium to fine sand with some tendency to coarsen upwards; the sandstones are moderately to well-sorted. Intercalated shale partings are medium gray (N5) to dark gray (N3), generally silty, and average 2 in. (5.1 cm) thick. Petrographic analysis (Kth-1) indicates clear, angular quartz grains, with up to 60% exhibiting relict fine-grained silciceous cement, comprise the dominant framework mineralogy. Lesser amounts of subangular, sericitized potassium feldspar and highly corroded plagioclase (An66, average of 4 grains, Michel-Levy method) are minor framework components. Lithic fragments are generally subrounded balls of phyllosilicate "hash". These together with rounded chert grains constitute 17% of the detrital components. One gneissic-textured lithic fragment was also observed. Rounded glauconite pellets are present in trace amounts. Matrix and empty pores constitute about 2% of the total rock; framework grains comprise 73% and fine-grained calcite cement constitutes 25%. Intercalated

The central nonmarine section is well-exposed along the arroyo southwest of Middle Well (sec. 35, T1N, R6W). It is estimated to be approximately 100 ft (30.5 m) thick. The nonmarine interval crops out as low sandstone benches, separated by shallow valleys underlain by shale and siltstone. Moderately sorted, thin- to medium-bedded, fine-grained quartzose sandstone containing as much as 20% organic debris is intercalated with greenish-gray (5GY6/1) to greenish-black (5G2/1), organic-rich siltstone and shale. These units are generally between 1- and 10-feet (0.3 and 3.0

m)-thick. The sandstones are chiefly yellowish-gray (5Y8/1), weathering to light-yellowish-gray (5Y7/2), and commonly contain tabular sets of small-scale, low-angle tangential crossbeds. Wood casts in the sandstones are common. The organic content of the shales is locally high enough to comprise thin lignite beds.

The top 70-foot (21.3 m)-thick marine section of the Tres Hermanos Tongue is similar to the lower section. A 10-foot (3.0 m)-thick, massive, extensively bioturbated sandstone occurs approximately 10 ft (3.0 m) below the top of the Tres Hermanos. From this sandstone, which in the field is very similar in appearance to the Gallego Sandstone, was recovered Lopha bellaplicata (NE 1/4, sec. 6, T1N, R6W). Coilopoceras colleti was recovered from the silty shale beneath the bioturbated sandstone.

The lower and upper marine sections of the Tres Hermanos Tongue are deposits left by regressing and transgressing seas, respectively. The reverse grading in the sandstones of the lower marine section can be attributed to wave-swash sorting (Clifton, 1969). Crossbedding and asymmetrical ripples which occur in both marine intervals are characteristic of the upper shoreface depositional environment (Reineck and Singh, 1975). The bioturbated sandstone in the upper marine interval may represent somewhat deeper shoreface sedimentation. The central nonmarine section preserves deposits of nearshore or inland environments. This interval, characterized by wood casts, fine-grained sandstones, organic shales, and thin lignite beds, represents nearshore deposits of backshore marshes and tidal flats similar to those described for the Menefee and Lance formations (Selley, 1972). Within this interval, the sea level probably remained nearly static and the shoreline position changed from regression to transgression.

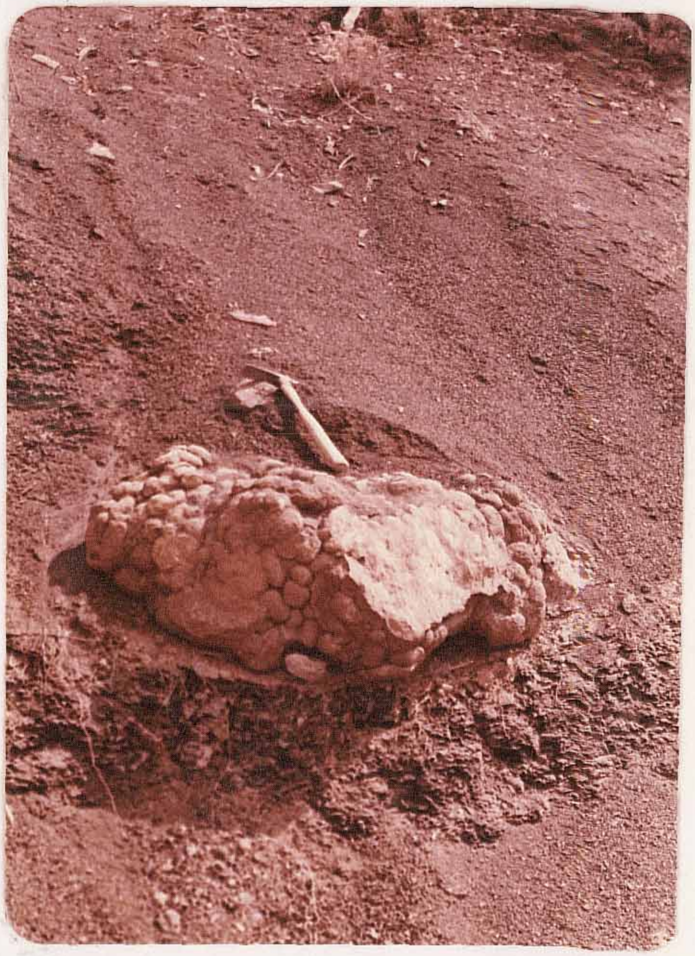
D-Cross Tongue

Dane, Wanek, and Reeside (1957) proposed to rename Pike's (1947) Pescado Tongue of the Mancos Shale at D-Cross Mountain (secs. 17 and 18, T3N, R8W) the D-Cross Tongue. They recognized this stratigraphic interval by faunal evidence to be a distinct and higher tongue than Pike's Pescado Tongue. They defined the D-Cross Tongue as the persistent marine shale of latest Carlile age underlying the Gallego Sandstone (then the upper Gallup Sandstone). Tonking (1957) had included the D-Cross Tongue within his La Cruz Peak Formation.

The D-Cross Tongue of the Mancos Shale is best exposed in a faulted and intruded 40-foot (12.2 m)-thick section on the west side of Jaralosa Creek (NW 1/4, SE 1/4, sec. 35, T1N, R6W). Maximum thickness in the study area is approximately 95 ft (29 m). The basal contact of the D-Cross Tongue with the underlying Tres Hermanos Member is sharp; a marked lithologic change occurs between the moderately consolidated sandstones of the Tres Hermanos Member and the silty, gray shales of the D-Cross Tongue. The upper contact of the D-Cross Tongue with the overlying Gallego Sandstone is gradational and is best exposed in a canyon cut into the northern edge of the extensive pediment surface (SE 1/4, SW 1/4, sec. 6, T1N, R5W). Within the top 10 ft (3.0 m) of the D-Cross Tongue, the shale grades upwards into medium-gray (N5)-weathering, fine-grained, massive sandstone. The grain size increases and color lightens upwards into the Gallego Sandstone. The contact is drawn beneath the first resistant sandstone. Shales are silty near both the base and the top of the D-Cross Tongue.

The D-Cross Tongue is generally comprised of medium-dark-gray (N4), weathering to grayish-black (N2), fissile shale. At the previously mentioned exposure on Jaralosa Creek, 1- to 5-in. (2.5- to 12.7-cm)-thick olive-black (5Y2/1)-weathering, very-fine-grained, silty sandstones are gradationally interbedded with the shales. The shales are generally concretionary, with concretions locally forming irregular discontinuous beds. Two distinct types of sandy limestone concretions occur within the D-Cross Tongue. The first of these are smooth and oblate, weather to moderate brown (5YR4/4), and often contain ammonites (fig. 5A). The second type are septarian and oblate, weather to light brown (5YR5/6) or grayish black (N2), and are rarely fossiliferous (fig. 5B). The centers of both types of concretions (usually the latter type) may be comprised of brownish-black (5YR2/1) massive calcite. Concretions range

A



B

Figure 5: Concretions in the D-Cross Shale
A. Smooth variety
B. Septarian variety
Hammer handle is 12.5 in. (31.8 cm) long.

in diameter from 6 in. (15.2 cm) to 30 in. (0.76 m). Fossils recovered from the D-Cross Tongue in the study area include *Scaphites whitfieldi*, *Prionocyclus novimexicanus*, and *Inoceramus sp.*.

A 6-in. (15.2 cm)-thick, moderate-reddish-orange (10R6/6)-weathering, silty limestone with an irregular surface is well-exposed to the east of the stock pond dam in the arroyo west of Middle Well (NW 1/4, SE 1/4, sec. 35, T1N, R6W). This bed is located approximately 8 ft (2.4 m) above the base of the D-Cross Tongue, and may represent the Juana Lopez equivalent in the study area (Hook, 1977, oral commun.).

The D-Cross Tongue of the Mancos Shale records the change from marine transgression at the base to marine regression at the top. The D-Cross Tongue accumulated in a marine shelf environment. Sandstone beds may record storm deposits (Reineck and Singh, 1975).

Gallego Sandstone

Winchester (1920) gave the name Gallego Sandstone to the resistant sandstone occurring about 900 ft (274 m) above the Dakota Sandstone at Pueblo Viejo Mesa near Gallego Creek (sec. 17, T4N, R7W). He included this sandstone unit within his Miguel Formation. Pike (1947) claimed to have traced the Gallup Sandstone in the San Juan basin into Winchester's Gallego Sandstone Member of his Miguel

Formation. Dane, Wanek, and Reeside (1957) regard the Gallego Sandstone as being a regressive sandstone member of the upper Gallup Sandstone. Molenaar (1974) suggests from his studies of measured sections that the Gallego Sandstone Member of the upper Gallup Sandstone at Puertecito is comprised of two barrier-bar sandstones separated by a transgressive marine shale. His lower barrier bar is the Gallego Sandstone of this report; the upper barrier bar and the intervening shale have been included in the Mesaverde Formation in this report. Recently, the United States Geological Survey Stratigraphic Names Committee informally suggested that the correlation of the Gallego Sandstone with the Gallup Sandstone is, at present, invalid (Hook, 1978, oral commun.). The Gallego Sandstone pinches out to the east of the study area (Massingill, 1979).

The Gallego Sandstone in the thesis area crops out as rounded cliffs or low hogbacks. Thickness of the Gallego Sandstone at a canyon cut into the northern edge of the extensive pediment surface (sec. 6, T1N, R5W) is approximately 50 ft (15.2 m). In an arroyo west of Jaralosa Creek (sec. 35, T1N, R6W), the Gallego Sandstone is approximately 40 ft (12.2 m) thick. The contact with the underlying D-Cross Tongue of the Mancos Shale is well-exposed at the former location. The contact here is gradational and is drawn at the first resistant sandstone. The upper contact with the overlying Mesaverde Formation is relatively sharp. The base of the latter is a silty, dark-gray (N3) shale at exposures at Middle Well (sec. 35, T1N, R6W) and south of Corkscrew Canyon (sec. 1, T1S, R6W). The base of the Gallego Sandstone is an olive-black (5Y2/1) thick-laminated sandstone that weathers to moderate yellowish brown (YR5/4). It is 8- to 10-ft (2.4- to 3.0-m)-thick. Color on a fresh surface becomes yellowish gray (5y8/1) upwards into the main body of the Gallego Sandstone. The basal section, which is approximately 20 ft (6.1 m) thick, is medium- to very thick-bedded. Large-scale, low-angle, tabular sets of planar crossbeds occur locally. Local erosional surfaces are marked by channel scours and very thin siltstone lenses. Bedding, however, is generally obscured by numerous vertical burrows 0.25 in. (0.64 cm) wide and as much as 6 in. (15.2 cm) long. These burrows stand out prominently on a weathered surface and are a characteristic feature of the Gallego Sandstone. As a result of the extensive burrowing, the Gallego Sandstone commonly assumes a massive appearance. Organic material usually occurs on a fresh rock surface.

Petrographically (Kg-1), these sandstones are comprised dominantly of well-sorted, monocrystalline, subangular, quartz grains with slighly undulatory extinction of fine sand size. Although most of these grains are clear, a small percentage contain linear trains of bubbles. Three grains with relict fine-grained siliceous cement also occur. Cloudy, partially sericitized potassium feldspar and trace amounts of microcline comprise 14% of the rock's framework components. Only 4% plagioclase grains were observed;

however, these were extensively replaced by sericite and calcite, and were usually recognizable only by twinning "ghosts" in the replacing minerals. Other framework constituents include ragged muscovite cleavage laths and a few rounded chert grains. The matrix of clays and organic debris comprises 28% of the rock. Interstitial clays commonly show partial replacement by sericite or chlorite. Patchy calcite cement and trace amounts of phyllosilicate cement comprise 10% of rock. Open pore space is estimated at about 2%. It should be noted, however, that induration of the Gallego Sandstone is only moderate, and effects of weathering processes are thought to permeate quite deeply into the rock. Thus, an accurate characterization of true petrographic properties is difficult.

Generally one or two 14-in. (35.6 cm)-thick fossiliferous sandstone beds are prominent in the upper 8 to 10 feet (2.4 to 3.0 m) of the Gallego Sandstone. These beds commonly weather brownish gray (5YR4/1) to dark yellowish brown (19YR4/2) and are more resistant than the rest of the Gallego Sandstone. On a fresh surface, color varies from moderate brown (5YR3/4) to medium light gray (N6) with decreasing sand content. The dominant fossil is Lopha sannionis, an oyster with thick, curving ribs that averages 1.5 in. (3.8 cm) long (fig. 6). Nearly unbroken fossils are common. These fossiliferous beds are characteristic of the Gallego Sandstone in the study area, as well as in exposures to the north (Jackson, 1979) and east (Massingill, 1979).

Figure 6: Lopha sannionis from Gallego Sandstone "brown bed"; Jaralosa Creek at Middle Well (SW 1/4, NE 1/4, sec. 35, T1N, R6W). Pencil is 5.5 in. (14 cm) long.

The overlying Mesaverde Formation contains nonfossiliferous sandstone beds similar in appearance to the Gallego (as at exposures in the northeast corner of the study area; sec. 8, T1N, R5W). Thus, the fossiliferous beds in the Gallego are important to formation identification.

Petrographically (Kg-2), the framework components of the fossiliferous beds are comprised of 15%, predominantly monocrystalline quartz with slightly undulatory extinction; 8% angular, extensively sericitized potassium feldspar; and 7% plagioclase (An49, average of 8 grains, Michel-Levy method) replaced locally by calcite. Burrowed fragments of Lopha sannionis, which comprise another 40% of the framework components, commonly exhibit preservation of internal shell structure. A few smaller fragments are replaced internally by pseudospar, which often shows "ghosts" of the original internal structure. Other framework constituents include 3% subrounded chert, 3% muscovite shreds and laths, 4% organic trash, and traces of subrounded lithic fragments of quartz-mica schist. Matrix material, comprising 15% of the total rock, is calcareous mud. About 30% of the mud occurs as relatively sand-free rounded masses, which are inferred to represent erosion of locally-derived calcareous mud. Cement material is principally small crystals of clear calcite.

Casts of Cardium sp. and Inoceramus sp. are ubiquitous and locally abundant in the middle and upper Gallego Sandstone. Prionocyclus sp. was collected from the top of the Gallego Sandstone from exposures southwest of La Jara Canyon (SE 1/4, SE 1/4, sec. 6, T1N, R6W).

The Gallego Sandstone is obviously marine as inferred from the ubiquity of marine fossils throughout the formation. The distribution of this formation and its pinchout to the northeast of the study area suggests that the Gallego Sandstone represents an elongate sand body. This fact, together with observed sedimentary structures and bioturbation, is strongly suggestive of a barrier-bar deposit, in agreement with Molenaar's (1974) interpretation. Such features are described for barrier sand-bars in Selley (1972), Reineck and Singh (1975), and Pettijohn and others (1973). Selley (1972) states that clay matrix and low porosity are characteristics of regressive barrier sands. Channel scours may be ascribed to storm washovers. Fossiliferous "brown beds" probably represent short episodes of slight transgression and slightly higher energy conditions as indicated by the presence of intraclasts. Swamp deposits, commonly comprised in part of coals, may occur to the landward side of such barrier bars (Selley, 1972; Pettijohn and others, 1973). The lowest coal-bearing units of the basal Mesaverde Formation may be the coals of younger swamps left by the regressing sea.

Mesaverde Formation

The Mesaverde Group was named by Holmes (1877) for exposures on Mesaverde in southwestern Colorado. Winchester (1920) included the rocks herein mapped as the Mesaverde

Formation within his Chamiso Formation. Pike (1947) observed that the term "Mesaverde" was used loosely as a formational name south of the type locality. His definition of the Mesaverde Formation at D-Cross Mountain included the Gallego Sandstone of this report. Tonking (1957) called the Mesaverde Formation of this report the Crevasse Canyon Formation. Givens (1957), in the area west of Tonking, had observed the lithologic similarities of this stratigraphic interval with the type-section Crevasse Canyon. However, the Crevasse Canyon Formation, as defined by Allen and Balk (1954), includes the strata between the top of the Gallup Sandstone and the base of the Point Lookout Sandstone. As the Point Lookout has not yet been identified within the study area and the Gallup-Gallego correlation is uncertain, this stratigraphic interval is herein referred to as the Mesaverde Formation.

The Mesaverde Formation is exposed in three major portions of the study area. The most extensive outcrops occur in the north-central portion, north of the northeast-trending Tijeras lineament. Here, the extensively faulted Mesaverde Formation is estimated to be approximately 1000 ft (304.8 m) thick. Another fault-abridged portion is exposed east of Jaralosa Creek in the vicinity of Corkscrew Canyon. A short section is exposed along the axis of an anticlinal fold in the southwestern corner of the study area.

The underlying Gallego Sandstone is sharply overlain by about 60 ft (18.3 m) of dark-gray (N3) marine shale west of Jaralosa Creek at Middle Well (sec. 35, T1N, R6W). East of Jaralosa Creek (sec. 6, T1N, R5W), this shale is about 25 ft (7.6 m) thick. Overlying the basal Mesaverde shale is a light-olive-gray (5Y5/2), moderately-sorted, massive marine sandstone which is approximately 40 ft (12.2 m) thick at Middle Well. This medium-grained sandstone exhibits slight upward coarsening and corresponds to the upper of Molenaar's (1974) two barrier bars.

The Tertiary Baca Formation unconformably overlies the Mesaverde Formation. This contact is relatively sharp in the Hot Spot mine vicinity, where the arkosic, lithic-rich sandstones of the Baca Formation are easily identifiable. This contact is also readily identifiable west of Chavez Creek, where the basal Baca unit is a poorly sorted, sandy, heterolithic conglomerate. In the vicinity of Corkscrew Canyon, however, Baca-type arkosic sandstones are interbedded with organic-rich Mesaverde-type shales, siltstones, and sandstones over a stratigraphic interval of approximately 30 ft (9.1 m). The contact was delineated at the base of a limestone-cobble conglomerate (fig. 7).

The Mesaverde is mainly comprised of nonmarine fluvial and swamp deposits. Approximately 200 ft (61 m) of dark-gray (N2) organic shales containing interbeds of coal and sandstone overlie the basal marine section. The coals range in thickness from 1 to 2 ft (0.30 to 0.61 m) and are intercalated with organic shales and thin olive-gray (5Y4/1), very-fine-grained, silty sandstones. Palynological data

Figure 7: Contact between the Baca and Mesaverde formations; Corkscrew Canyon vicinity (NW 1/4, SW 1/4, sec. 36, T1N, R6W). Basal Baca bed is a limestone-cobble conglomerate. Hammer handle is 12.5 in. (31.8 cm) long.

indicate a coastal or deltaic plain environment of accumulation with brackish water influence (M. Chaiffetz in Chapin and others, 1979). Petrographically (Kmv-1), the sandstones are comprised of 87% angular, slightly undulatory, monocrystalline quartz grains. Subrounded chert grains comprise another 11% of the framework constituents. The cementing agent is limonitic cherty silica with a greenish hue. Each unit of shale and coal is overlain by dusky-yellow (5Y6/4)-weathering, fine-grained sandstone. These sandstones are usually thin-bedded and often exhibit small-scale, low-angle, planar crossbeds. They range in thickness from 2 ft (0.6 m) to 10 ft (3.1 m). Approximately eight coal and sandstone sequences occur in the Mesaverde Formation exposures east of Jaralosa Creek and north of Corkscrew Canyon.

Above these units, the moderately indurated sandstones rapidly thicken to 50 ft (15.2 m) or more. Such sandstones form high bluffs along Jaralosa Creek in the north-central section of the outcrop belt. These sandstones weather to a grayish-yellow (5Y8/4) or a grayish-orange (10YR7/4) color; on a fresh surface they are usually moderate yellowish brown (10YR5/4). Framework constituents of these sandstones (Kmv-2) are generally fine- to medium-sand-size, moderately sorted, subangular grains. Extensively sericitized potassium feldspar comprises 45% of the framework grains. Clear, monocrystalline, quartz with straight extinction comprises another 32%. About 10% of these quartz grains are rutilated. Lithic fragments, comprising 14% of the framework constituents, are primarily clay and sericite balls. The cementing agent of the rock is tiny interlocking carbonate crystals. Porosity is very low. Bedding thicknesses range from Ø.25 in. (Ø.64 cm) to 6 in. (15.2 cm). Organic debris is abundant, as are twig imprints. Small- and large-scale, low-angle, planar or trough crossbedding occurs commonly. Sandy, dark-reddish-brown (1ØR3/4), hemispherical nodules often weather out in bold relief. These nodules are commonly 2 in. (5 cm) or less in diameter. A very-resistant, grayish-brown (5YR3/2)-weathering, fine-grained sandstone often caps the nodule-bearing units (fig. 8). The framework grains in this resistant sandstone (Kmv-3) are generally subangular and moderately sorted. Quartz with straight extinction comprises 61% of these framework grains. Inclusions of stubby muscovite laths, rutile, and bubbles occur commonly. Relict syntaxial silica cement occurs on about 1Ø% of these grains. Other framework constituents include 12% sericitized potassium feldspar, 13% organic-rich clay and sericite balls, and 9% subrounded chert. Blocky calcite cement has filled most pores in these rocks. Small- to large-scale, high-angle, trough crossbeds stand out markedly in these beds.

Between these sandstone units are organic shale, lignite, and intercalated silty sandstone units, commonly 5 ft (1.5 m) or less thick. The shales and lignite generally weather dark greenish black (5GY4/1). Silty sandstones,

Figure 8: Cross-stratified "brown bed" in Mesaverde Formation; Hot Spot mine vicinity (NW 1/4, NW 1/4, sec. 18, T1N, R5W). Jacob's staff is graduated in feet.

weathering grayish brown (5YR3/2) to grayish olive (10Y4/2), are gradational with these shale and lignite units, and rarely exceed 3 in. (7.6 cm) in thickness.

The uppermost approximately 100 ft (30.5 m) of the Mesaverde Formation is similar in appearance to the basal coal-bearing section. The coals in this part of the Mesaverde Formation were mined from several adits at the Hot Spot mine (NW 1/4, sec. 18, T1N, R5W). The thickest coal bed here is 5 ft (1.5 m) thick. These coals exhibit rapid lateral thickness variation as well as rapid lithologic gradation into dark-gray (N3), silty shale and pale-yellowish-brown (10YR6/2), very-fine-grained silty sandstone. Vertically, the coals also grade into medium-dark-gray (N4) shale and grayish-olive-green (5GY3/2)-weathering mudstone with 0.25 in. (0.64 cm) thick wavy laminations. Each coal sequence is capped by grayish-olive (10Y4/2), thin-bedded to massive, very-fine-grained sandstone. These sandstones commonly weather to dark yellowish brown (10YR4/2), and contain fossilized twigs and twig imprints. In contrast to the coals of the basal Mesaverde Formation, pollen recovered from coal of the upper Mesaverde Formation indicate isolated upper coastal plain swamp environments (M. Chaiffetz in Chapin and others, 1979).

Associated with the coal-bearing units of both the basal and upper Mesaverde Formation are ironstone concretions. These concretions are blackish-red (5R2/2), and range in size and shape from 1 in. (2.5 cm) elongate chips to 3 in. (7.6 cm) oblate masses. The concretions are always found as float, and often completely mantle the ground surface. Silicified wood was found only in the upper coal-bearing units at Corkscrew Canyon.

Tertiary

Eocene

Baca Formation

Winchester (1920) included the present-day Baca Formation within his Datil Formation. The Baca Formation was separated from the Datil Formation and named for exposures along Baca Canyon (secs. 4, 5, 8, and 9; T1N, R4W) by Wilpolt and others (1946). Their description of this formation, however, was taken from exposures in the Joyita Hills-Carthage area. They described the Baca Formation as consisting of conglomerates, red and white sandstones, and red clays. Clastic material comprising the conglomeratic units was derived from Precambrian quartzite and granite, the Madera Limestone, and the Abo Formation. Tonking (1957) stated that the Baca Formation ranges in thickness from 0 to 700 ft (0 to 213.4 m). Potter (1970) measured a composite section of 695 ft (211.8 m) at Baca Canyon and also designated three informal sections. Snyder (1971) reported a maximum thickness of 2500 ft (762 m) from a Tenneco oil test drilled south of Pie Town, New Mexico. Massingill (1979) measured a 754-ft (230 m)-thick section about two miles (3.2 km) north of the type section.

Gidley (in Gardner, 1910) identified a fossil tooth found in variegated beds unconformably overlying the Mesaverde Formation in the Carthage area as *Paleosyops* of middle Eocene age. Snyder (1970, 1971) reported the discovery of a partial section of a jawbone containing four teeth which had weathered from the Baca Formation in the Datil Mountains (SE 1/4, SW 1/4, sec. 31, T2N, R9W). The jaw was identified by Dr. C.L. Gazin of the U.S. National Museum as being that of Protoreodon pumilus of late Eocene age. Stratigraphic and sedimentologic studies of the Baca Formation, including exposures in the study area, have been conducted by Snyder (1971), Johnson (1978), and Cather (in prep.). These studies have attempted to reconstruct the environments of deposition and transport directions of Baca sediments.

The Baca Formation is exposed in an east-trending elongate belt that is approximately 120 mi (193 km) long by 20 mi (32 km) wide. Other formations which may be partly contemporaneous with the Baca Formation include the McRae, Galisteo, and Raton formations (Lee, 1915; Wilpolt and others, 1946; Tonking, 1957). Kelley and Silver (1952) and Tonking (1957) believe that these formations represent deposits of isolated basins. Snyder (1971) and Johnson (1978) have suggested that the Eagar Formation of eastern Arizona is a lateral equivalent of the Baca.

In the study area, the Baca crops out in a broad swath through the western, central and southeastern regions. Extensive faulting and local folding occur throughout this belt. Smaller outcrops occur in the southwestern corner on the flanks of a small anticline. A maximum thickness of 950 ft (289.6 m) was approximated from structural cross-sections. The Baca Formation crops out as a series of gently dipping discontinuous sandstone ridges and hogbacks which are generally elongated in a northerly direction. Between these ridges are shallow valleys underlain by less-resistant shales and siltstones.

The nature of the depositional contact between the Baca and the underlying Mesaverde Formation shows rapid variation within the study area. Generally, west of Jaralosa Creek, this contact is marked by an irregular pink or white sandy conglomerate. This bed is 4- to 18-in. (10.2- to 45.7-cm)-thick. Clasts are a poorly sorted, well-rounded mix of iron concretions, petrified wood, and siltstone chips, presumably derived from the Mesaverde Formation. Other lithologies include well-polished, multicolored quartzite and silicified siltstone which are more common in the upper conglomeratic units of the Baca Formation. Above the basal conglomerate, the red, white, and yellow units of the Baca are unmistakeable. Slight angular unconformity between the Baca and Mesaverde is also evident at these locations. East of Jaralosa Creek, the basal conglomerate of the Baca is comprised at least in part of rounded concretions and clasts of massive, gray limestone (fig. 7). The abundance of limestone clasts in this conglomerate generally increases southwards to Corkscrew Canyon. South of Corkscrew Canyon, limestone concretions and clasts occur within the Baca over a stratigraphic interval of approximately 30 ft (9.1 m). In this area, Mesaverde-type sandstones, gray siltstones and shales are interbedded with Baca-type sandstones, conglomerates, and red shales. The two formations appear to be conformable.

The contact of the Baca Formation with the overlying Spears Formation is well-exposed on the western faces of eastward-dipping hogbacks along the east bank of Jaralosa Creek (SE 1/4, sec. 13 and E 1/2, sec. 24, T1N, R6W; fig. 9) and north of the Hook Ranch headquarters (SW 1/4, sec. 24, T1N, R6W). This contact is interbedded and gradational over a stratigraphic interval of about 50 ft (15.2 m). Within this interval the Baca Formation is comprised of thinly bedded, medium-grained, moderately sorted, moderate-red (5R6/4), arkosic sandstones. The Spears Formation is comprised of thinly bedded, medium-grained, moderately sorted, grayish-red-purple (5RP4/2), volcaniclastic sandstones and siltstones.

The Baca Formation is usually comprised of sequences of sandstone, siltstone, and shale that average about 15 to 40 ft (4.6 to 12.2 m) in thickness. These units become finer grained from bottom to top; however, the basal sandstones of these sequences have a slight tendency to coarsen upwards. The units become slightly coarser grained toward the middle and upper portions of the Baca Formation where shales and mudstones are less common. The sandstones in the lower half of the Baca Formation weather grayish orange (10YR7/4) to pinkish gray (5YR8/1). Towards the top

Figure 9: Gradational contact between Spears and Baca formations; east bank of Jaralosa Creek across from Hook uranium prospect (SW 1/4, SE 1/4, sec. 3, T1N, R6W). Volcanic fragments occur within the siltstones and medium-grained sandstones between the two prominent sandstone beds, as well as within the lower sandstone bed. Contact is designated above highest Baca-type sandstone.

of the formation, the sandstones weather moderate red (5R5/4) to pale reddish brown (10R5/4).

Bedding thicknesses range from 1 in. to 2 ft (2.5 cm to 0.6 m). High-angle, small- and large-scale trough cross-bedding is commonly (fig. 10). The sandstones commonly contain near their base chips of siltstone as long as 4 in. (10.2 cm) derived from the top of the underlying sequence. Brown wood casts are common throughout the sandstones. Dark-gray (N3) silicified wood occurs throughout the Baca Formation, but it is larger and more abundant in the top-third of this formation. The sandstones are comprised of medium to coarse sand grains; sorting is poor to moderate. A few polished quartzite pebbles are occur within the sandstones. Petrographically (Tb-2), the sandstones are comprised of approximately 50% monocrystalline quartz with undulatory extinction, 18% sericitized potassium feldspar, 13% well-rounded hematitic chert grains, and 10% lithic fragments. The lithic fragments are comprised chiefly of sericitic mudstone.

Near the middle of the Baca Formation, the basal few inches of sandstone beds are often conglomeratic sandstone (fig. 11). The clasts are a poorly sorted, frequently imbricated mixture of well-rounded and polished multicolored quartzite and chert with lesser amounts of granite and red, silicified siltstone. Well-rounded, medium-light-gray (N6) limestone clasts locally occur in these basal conglomerates. Clasts from the conglomerates will often completely mantle the ground surface.

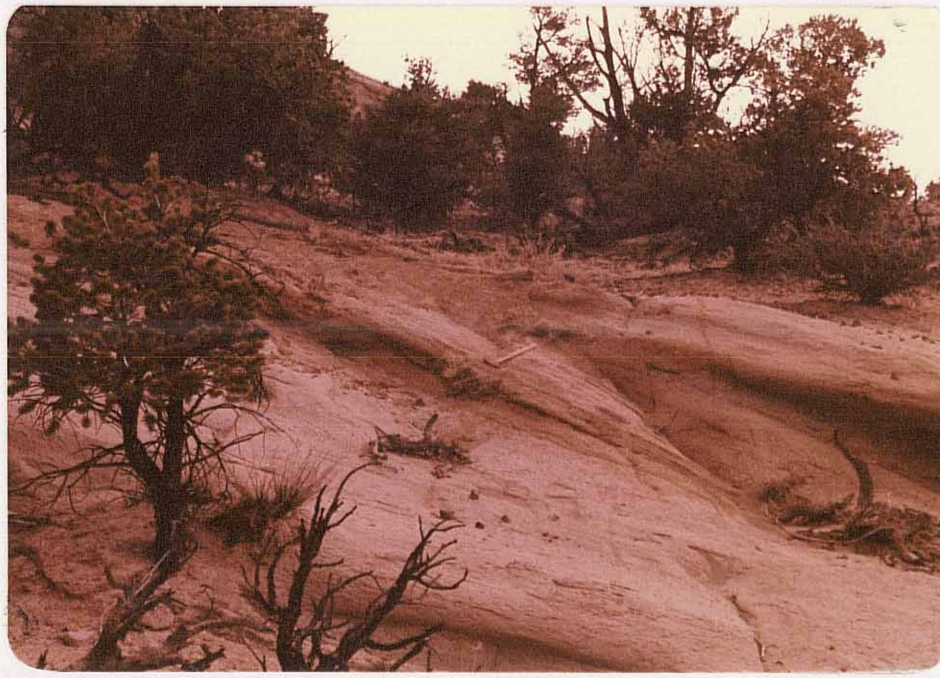


Figure 10: Large-scale, tangential cross-stratification in sandstone of the Baca Formation; south of Corkscrew Canyon (sec. 2, T1S, R6W). Hammer handle is 12.5 in. (31.8 cm) long.

Figure 11: Conglomerate within lower Baca Formation; south of Corkscrew Canyon (sec. 2, T1S, R6W). Hammer handle is 12.5 in. (31.8 cm) long. Clasts are well sorted, well rounded, and imbricated. They are comprised chiefly of multicolored quartzite with some banding or quartz veining. Overlying unit is a bleached, subangular, medium-grained, moderately sorted quartz-feldspar sandstone.

Gradationally overlying sandstones in the lower Baca Formation are dark-reddish-brown (10R3/4) siltstone and shale units that range from 5 to 8 ft (1.5 to 2.4 m) thick. Thin discontinuous pods and lenses of dark-greenish-gray (5G4/1) mudstone occur infrequently below the basal sandstone of the succeeding sandstone and siltstone sequence in the lower-third of the Baca Formation. Towards the top of the Baca Formation, siltstones and sandy siltstones replace shales. The thickest siltstone interval occurs approximately 30 ft (9.1 m) below the top of the Baca Formation and is about 100 ft (30.5 m) thick. It forms the broad valley north of the Hook Ranch headquarters (SE 1/4, sec. 23, T1N, R6W) and presumably underlies Jaralosa Creek in the vicinity of Little Well (secs. 13 and 24, T1N, R6W).

A 3-in. (7.6 cm)-thick, light-gray (N7) nonfossiliferous limestone bed occurs west of Jaralosa Creek (SE 1/4, sec. 26, T1N, R6W). Limestone float of similar appearance occurs near the base of the Baca Formation in the southwestern corner of the study area (sec. 2, T1S, R6W).

The origin of the Baca Formation is presently in dispute. Snyder (1971) and Massingill (1978, oral commun.), on the basis of observations which are similar to those of the author, suggest that the Baca Formation is comprised of cyclic, fining-upward deposits of fluvial and lacustrine origin. In contrast, Johnson (1978) and Cather (1978, written commun.) have described cyclic coarsening-upward sequences which they believe represent deposits of lacustrine deltas. The major difference between these interpretations of depositional environments lies in the interpretation of what constitutes a genetic stratigraphic interval.

Oligocene

Spears Formation

Tonking (1957) subdivided Winchester's (1920) Datil Formation into three members and named the basal member after the Guy Spears Ranch (sec. 8, T1N, R4W). His type section consists of nearly 1350 ft (411.5 m) of volcaniclastic sedimentary rocks. Burke and others (1963) reported an early Oligocene age (37.1 m.y., K/Ar, biotite) for a latite boulder near the top of the Spears in the Joyita Hills. Chapin (1971a) elevated the Spears to formational status.

The Spears Formation is a volcaniclastic apron derived from erosion of the earliest calc-alkalic suite of the Mogollon-Datil volcanic field (Elston and others, 1976). Northeast-trending paleovalleys occurring within the Spears Formation were structurally controlled by the Morenci lineament and indicate source areas to the southwest (Chapin and Seager, 1975). Brown (1972) divided the Spears Formation into two members in the Magdalena area. The lower member is comprised of latitic to andesitic conglomerates, mudflow deposits, and thin, interbedded volcaniclastic sandstones. The upper member is comprised of basal amygdaloidal "turkey track" andesite flows, the tuff of Nipple Mountain, and overlying andesitic lava flows and latitic ash flow tuffs with interbedded mudflow deposits and conglomerates. Massingill (1979) reports a thickness of 1259.5 ft (383.9 m) for the Spears Formation northeast of the study area (sec. 17, T1N, R4W).

Major outcrops of the Spears Formation occur in the south-central portion of the study area, along the eastern and western margins. Maximum thickness of the Spears is estimated at approximately 1000 ft (304.8 m). The lower 100 ft (30.5 m) of this formation crops out in a series of hogbacks; the remainder crops out as low, rounded hummocks.

The basal contact with the underlying Baca Formation is well-exposed on the western face of the gently dipping hogbacks along the east bank of Jaralosa Creek (SE 1/4, sec. 13 and NE 1/4, sec. 24, T1N, R6W) and on the western face of a wedge-shaped fault block north of the Hook Ranch headquarters (sec. 24, T1N, R6W). At these locations the gradational and interbedded nature of this contact over a stratigraphic interval of about 50 ft (15.2 m) can be clearly seen (fig. 11). Within this zone, the Baca Formation is comprised of thin-bedded, medium-grained, moderately sorted, moderate-red (5R6/4) arkosic sandstones. The Spears Formation is comprised of thinly bedded, medium-grained, moderately sorted, grayish-red-purple (5RP4/2) volcaniclastic sandstones.

An upper contact with the Hells Mesa Tuff is poorly exposed on the north face of the hogback west of Abbe Spring (SW 1/4, sec.25, T1N, R5W), and somewhat better exposed in a stream cut east of Jaralosa Creek (SW 1/4, sec.25, T1N, R6W). The contact in each of these places appears unconformable. The Hells Mesa Tuff appears at the latter location to have filled channels eroded into the Spears Formation. Over much of its outcrop extent within the study area, the upper Spears Formation is fault bounded. However, a depositional contact of the Popotosa Formation on the Spears was mapped in a canyon east of Jaralosa Creek (NE 1/4, sec. 25, T1N, R6W). This may record the burial of an earlier fault scarp by Popotosa sediments.

The main body of the Spears Formation consists predominantly of moderately to well-indurated channel conglomerates, mudflow deposits, and lithic sandstones (fig. 12). Bedding ranges from approximately 2 in. (5.1 cm) to 2 ft (0.61 m) in thickness, and at times exhibits low-angle, large-scale, tangential cross-stratification. Outcrop colors range from grayish red purple (RP4/2) to pale pink (5RP8/2). Propyllitic alteration, forming a chlorite-epidote-calcite assemblage, gives some areas of the formation a greenish-gray (10G4/2) color. Fining-upward sequences are common within the channel deposits. Clasts are commonly subrounded and range in diameter from 0.125 to 4 in. (0.32 to 10.2 cm). The clasts are comprised dominantly of latitic to andesitic rocks consisting of felted plagioclase laths in a gray, aphanitic

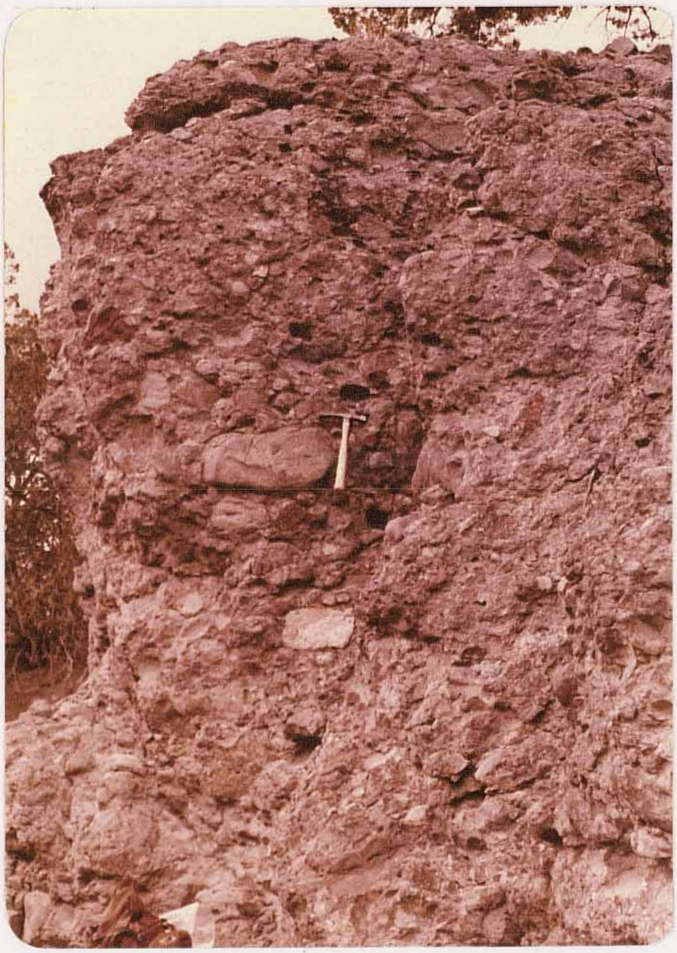


Figure 12: Mudflows in the lower Spears Formation; east of the Hook Ranch headquarters (NW 1/4, NW 1/4, sec. 25, T1N, R6W). Hammer handle is 12.5 in. (31.8 cm) long. Clasts are subangular to rounded, poorly sorted, and heterolithic. They are comprised chiefly of volcanic lithologies with lesser amounts of gray Pennsylvanian limestone, red siltstone (Abo Formation?), and quartzite.

groundmass. Red siltstone clasts, derived from the Abo Formation of Permian age, and clasts of Pennsylvanian limestones and Oligocene basaltic-andesites also occur. Petrographically (Ts-1, Ts-2), the sandstones and conglomerates are very poorly sorted and have low porosity due to high matrix content. The lithic fragments are comprised of argillized plagioclase crystals (average composition An48, average of 20 grains, Michel-Levy method) and green hornblende, extensively replaced by calcite and magnetite, within a groundmass of tiny subparallel plagioclase crystals and hematite-stained microlites. Cementing agents of these lithic fragments are usually present only in small amounts and are chiefly limonite with lesser amounts of patchy calcite. Potassium feldspar is essentially absent from the thin sections examined, with the exception of the groundmass of the lithic fragments.

Dark-greenish-black (5GY4/1) andesite flows form discontinuous outcrops in the upper Spears Formation, especially in the area south of the Hot Spot mine (E 1/2, sec. 24, T1N, R6W). These flows, which Tonking (1957) termed "turkey-track" andesites, are generally porphyritic. The phenocrysts consist of as much as 75% lath-shaped plagioclase, 10% magnetite, and 15% corroded hornblende. The flows generally do not exceed 20 ft (6.1 m) in thickness.

Discontinuous exposures of two tuffs within the Spears Formation were mapped in the study area. The lower tuff (Tst1) is light gray (N7), crystal poor and poorly welded. Petrographically, it consists of 10% rod-shaped pumice fragments, 2% potassium feldspar subhedra, 2% rounded to subangular latitic lithic fragments, 1% magnetite anhedra, and trace amounts of brownish-red biotite. The groundmass is comprised of fine hematite and magnetite, arcuate glass shards, and microlites. Large patches of calcite occur throughout the rock. The upper tuff (Tst2) is grayish-orange-pink (10R8/2) and is found near, or at, the upper contact of the Spears Formation. Petrographically, it consists of 15% corroded plagioclase subhedra (An45, average of 15 grains, Michel-Levy method) with some partial calcite replacement, 5% ragged brown biotite flakes and 3% magnetite anhedra in a matrix of hematite-stained, parallel-aligned arcuate glass shards and microlites. A crude foliation is defined by the glass shards and the biotite phenocrysts. Subrounded andesitic lithic fragments can comprise as much as 5% of the total rock. Each of these tuff units is no more than 15 ft (4.6 m) thick; they occur only as small, isolated remnants within the study area.

The occurrence of mudflow deposits with randomly oriented clasts in an unsorted matrix and conglomerates with low-angle cross-stratification support the interpretation that the Spears Formation represents alluvial-fan deposits. Such features are described by Reineck and Singh (1975) for alluvial-fan deposits in tectonically active areas. The axis of one channel observed in the study area seemingly supports Chapin and Seager's (1975) assertion that transport was directed away from source areas in the Magdalena and San Mateo mountains or beyond.

Hells Mesa Tuff

Tonking (1957) included within his Hells Mesa Member of the Datil Formation, the Hells Mesa Tuff and the A-L Peak Tuff of this report. Subsequent redefinition by Deal (1973) and Chapin (1974) revised the Hells Mesa Tuff to formational status and restricted it to the basal crystal-rich ash-flow tuff. Elston and others (1976) consider the Hells Mesa Tuff to be one of the youngest ash-flow tuffs of a calc-alkalic suite exposed on the Mogollon Plateau. The Hells Mesa Tuff was erupted from the North Baldy cauldron in the central Magdalena Mountains (Chapin and others, 1978). Outcrops of the Hells Mesa Tuff are very widespread and occur northwards to the northern terminus of the Bear Mountains.

Weber and Bassett (1963) obtained a K-Ar date of 30.6 $\pm$ 2.8 m.y. (biotite) from the base of the Hells Mesa Tuff at Tonking's (1957) type section on Hells Mesa. Burke and others (1963) reported a K-Ar date of 32.1 $\pm$ 0.2 m.y. (biotite) from a sample collected from south of Dog Springs Canyon in the Gallinas Mountains (SE 1/4, sec. 7, T1N, R8W), and another K-Ar date of 32.4 m.y. (biotite) from a sample collected from the basal Hells Mesa Tuff in the Joyita Hills.

The Hells Mesa Tuff crops out in the study area principally as steep, rubble-covered hogbacks west of Abbe Spring (SW 1/4, sec. 8, T1N, R5W) and east of Jaralosa Creek (SW 1/4, sec. 25, T1N, R6W). Its unconformable basal contact with the underlying Spears Formation is well-exposed at the latter location in the stream cut between the two hogbacks. Osburn (1978, oral commun.) suggested that a moderate-orange-pink (10R7/4), poorly welded, quartz-bearing tuff observed east of the Hot Spot mine (NW 1/4, NE 1/4, sec. 18, T1N, R5W) represents the base of the Hells Mesa Tuff. This particular rock, however, was not observed at any of the other outcrops of the Hells Mesa Tuff. Maximum thickness of the Hells Mesa Tuff in the study area is about 200 ft (61 m) east of Jaralosa Creek (sec. 25). Both north and south of this location, the Hells Mesa Tuff thins rapidly to zero feet thickness. At the southern terminus of its outcrop, a conglomerate with clasts derived exclusively from the Hells Mesa Tuff was observed. Emplacement of the Hells Mesa Tuff appears to have been controlled by channels cut into the underlying Spears Formation.

The Hells Mesa Tuff is generally a pale-red (10R6/2), densely welded crystal-rich rhyolitic tuff that weathers to grayish-red (10R4/2) angular blocks. Abundant phenocrysts, at times constituting 50% of the total rock, are observable by the naked eye in hand specimen. They include clear to smoky quartz, copper-colored biotite flecks, plagioclase, and sanidine. Pumice fragments comprise approximately 25% of the rock near the base. Thé fragments are round or rod-shaped and impart a moderately good foliation to an outcrop. Pumice content decreases to less than 10% near the top of the formation. Angular lithic fragments of average diameter 1 cm (0.4 in.) and comprised of purplish andesite occur sparsely. Brown (1972) attributes these clasts to the underlying Spears Formation.

Petrographically, clear, angular quartz anhedra to 3 mm (0.12 in.) in diameter increase in abundance from approximately 8% of the phenocrysts at the base to 28% near the top. Plagioclase euhedra of average composition An54 (average of 5 grains, Michel-Levy method) comprise 60% of the phenocrysts near the base. Plagioclase decreases in abundance upwards in the unit to 33% and becomes more calcic (An65, average of 6 grains, Michel-Levy method). Fresh sanidine subhedra increase in abundance from about 17% near the base, to 30% near the top. Pumice fragments are generally devitrified in a spherulitic manner. The groundmass is comprised of glass and spherulites with reddish hematitic pigmentation.

A-L Peak Tuff

Tonking (1957) included the A-L Peak Tuff within his Hells Mesa Member of the Datil Formation. Brown (1972) did the first definitive work on this formation in the Bear Mountains, where he mapped and described two units of what he informally called the tuff of Bear Springs. Deal (1973), and Deal and Rhodes (1976) formally named these rhyolitic ash-flow tuffs for a 2000-ft (609.6 m) section exposed on A-L Peak in the northern San Mateo Mountains and inferred a source in the Mt. Withington cauldron. Smith and others (1974) determined a fission track age of 31.8 ± 1.7 m.y. for the A-L Peak Tuff at the type locality. Chamberlin (1974) mapped three cooling units in the Council Rock district. According to Elston and others (1976), the A-L Peak Tuff is representative of the basal flows of a high-silica alkali rhyolite suite found in the Mogollon Plateau area.

The A-L Peak Tuff occurs as two cooling units separated by approximately 330 ft (100.6 m) of La Jara Peak Basaltic Andesite in the study area. Outcrops occur in the vicinity of Abbe Spring, east of the Hot Spot mine, and east of Jaralosa Creek to the southeast of the Hook Ranch headquarters.

Gray-massive member:

The gray-massive member of the A-L Peak Tuff crops out almost exclusively as a rubble-covered slope in the study area. Platy fragments, commonly 0.75 by 3 in. (1.9 by 7.6 cm), mantle the ground surface. Its basal contact with the Hells Mesa Tuff east of Jaralosa Creek (SW 1/4, sec. 25, T1N, R6W), and with the Spears Formation south of the Chavez Ranch (NW 1/4, sec. 9, T1N, R5W), is unconformable. An upper contact was not observed in the study area due to ground cover. The gray-massive member appears to have filled topographic lows resulting from post-Hells Mesa erosion. It ranges in thickness from about 20 to 300 ft (6.1 to 91.4 m) with rapid local variation.

The gray-massive member is moderate orange pink (5YR8/4) in color and poorly welded at the base; it becomes grayish orange pink (5YR7/2) and densely welded towards the top. Pumice fragments, comprising 10% of the total rock, are oblate- to rod-shaped with long dimensions generally parallel to bedding. Fresh sanidine subhedra are the dominant phenocrysts, comprising as much as 5% of the rock. A few of these sanidine phenocrysts are perthitic. Other phenocrysts, present in trace amounts, include clear quartz anhedra and magnetite subhedra. The latter is commonly replaced in large measure by hematite. Groundmass material is comprised of roughly equal amounts of arcuate glass shards with subparallel alignment and irregular areas of quartz and alkali feldspar developed by vapor-phase crystallization. Andesitic lithic fragments can comprise as much as 3% of the rock. A typical fragment is 3 mm in diameter and composed of plagioclase microlites and magnetite euhedra.

Pinnacles member:

The pinnacles member, interbedded within the La Jara Peak Basaltic Andesite, is well-exposed on Forest Road 123 near Abbe Spring (sec. 8, T1N, R5W) and east of Jaralosa Creek (sec. 18, T1N, R6W). It crops out as a steep cliff or slope. Both upper and lower contacts are disconformable with the La Jara Peak Basaltic Andesite. Blocks of La Jara Peak Basaltic Andesite are caught up in the base of the pinnacles member. Lithic fragments of La Jara Peak average 3 cm in diameter and occur throughout the entire unit. The pinnacles member is 78 ft (23.7 m) thick in the Abbe Spring area, but thins rapidly westward to less than 10 ft (3.0 m) thick. A 1000-foot (304.8 m)-thick accumulation in an area east of Jaralosa Creek may be attributable to ponding in a paleovalley along the Tijeras lineament.

The color of the tuff at the base of the pinnacles member is light brown (5YR5/6), weathering to grayish red purple (5RP4/2). Upwards in the unit, the color of a weathered surface grades into pale reddish brown (19R5/4). Pumice fragments are sandy-textured and elongate, and range from 0.32- to 17.8-cm long. A prominent foliation is defined by these well-flattened fragments. In exposures in Abbe Spring Canyon, sanidine crystals as long as 3.2 cm are gradationally more abundant in a 20-foot (6.1 m) zone beginning 13 ft (4 m) from the top of the unit.

Pumice fragments comprise 10% of the rock. These pumice fragments are seen in thin section to be replaced internally by a 0.1 mm thick layer of axiolites, and by interlayered hematite-stained quartz and alkali feldspar anhedra. Small anhedra of magnetite occur in trace amounts. Sanidine euhedra, exhibiting very little alteration, are approximately 2 mm long and comprise 5% of the rock. Other minor constituents include magnetite anhedra, replaced in part by hematite, and basaltic-andesite lithic fragments. Plagioclase was not observed in the thin section examined. The groundmass is comprised principally of 0.7 mm long, subparallel, devitrified glass shards outlined by hematite dust, and anhedral crystals of quartz and alkali feldspar.

La Jara Peak Basaltic Andesite

Tonking (1957) named the basaltic and andesitic flows overlying his Hells Mesa Member, the La Jara Peak Member of the Datil Formation. The unit derives its name from La Jara Peak, a prominent volcanic neck in sec. 11 (T2N, R5W). Willard (1959) correlated the La Jara Peak Basaltic Andesite with the post-Datil Mangas Basalt in Catron County. This correlation was accepted by Weber who proposed exclusion of the La Jara Peak rocks from the Datil Formation (1963, 1971). Chapin (1971a) reports a whole-rock K-Ar date of 23.8 ± 1.2 m.y. from the east side of the Bear Mountains at Cedar Spring (NE 1/4, NE 1/4, sec. 31, T1N, R4W).

Brown (1972) and Massingill (1979) have mapped the La Jara Peak Basaltic Andesite in the Bear Mountains to the southeast and northeast of the study area, respectively. Chamberlin (1974) attributed the absence of the La Jara Peak rocks in the Council Rock district southwest of the study area to possible damming along a monoclinal fold (pp. 116-117).

Chapin and Seager (1975) postulate that the La Jara Peak Basaltic Andesite was emplaced beginning 4 to 5 m.y. after the inception of block faulting in a closed basin south of the rising Colorado Plateau. They also observe that the present designation "basaltic-andesite" comes from the recognition that these flows are generally higher in silica and potash than true basalts, and have field and petrographic characteristics intermediate between basalts and andesites.

The La Jara Peak Basaltic Andesite occurs within three areas along the eastern margin of the study area, usually as two tongues separated by the pinnacles member of the A-L Peak Tuff. The lower tongue has a thickness of about 350 ft (106.7 m) near Abbe Spring and east of Jaralosa Creek. This tongue rests unconformably on the gray-massive member of the A-L Peak Tuff or on the Hells Mesa Tuff. The upper contact of this tongue with the overlying pinnacles member of the A-L Peak Tuff, as observed near Abbe Spring (secs. 8 and 9, T1N, R5W), is disconformable.

The upper tongue of the La Jara Peak Basaltic Andesite has a maximum thickness of 655 ft (199.6 m) near Abbe Spring and a minimum thickness of 300 ft (91.4 m) east of the Hot Spot Mine. The lower contact of this tongue is disconformable with the pinnacles member of the A-L Peak Tuff. Near Abbe Spring, the base of this tongue contains blocks as much as 3 ft (0.91 m) long of the underlying pinnacles member. Interbedding of the La Jara Peak rocks with the overlying Popotosa Formation was observed at one location. This interbedding occurs through a stratigraphic interval of approximately 50 ft (15.2 m) at a canyon mouth east of Jaralosa Creek (NW 1/4, NW 1/4, sec. 36, T1N, R6W).

The La Jara Peak Basaltic Andesite crops out primarily as steep, rounded, rubble-covered hillslopes often dissected by steep, V-shaped canyons. Individual flow thicknesses are between 3 and 25 ft (0.91 and 7.6 m). Weathered outcrop color ranges from brownish black (5YR2/1) in the massive central zone of an individual flow to dusky red (5R3/4) in scoriaceous autobrecciated zones. Hematitic pseudomorphs of pyroxene averaging 1.5 mm in diameter are a distinctive feature in hand specimens. Almond-shaped vesicles are common and are often filled completely with calcite and/or quartz. Toward the middle of most flows, partings parallel to foliation and spaced 1 in. (2.5 cm) to 1 ft (0.3 m) apart are evident. These parting surfaces exhibit a greenish-black (5G2/1) sheen in hand specimen.

Lenses of moderately sorted, coarse-grained volcaniclastic sandstone, commonly thin-bedded, occur sparsely between flows of the La Jara Peak Basaltic Andesite in the study area. They usually weather a dark-yellowish-orange color (10YR6/6).

Miocene

Popotosa Formation

The Popotosa Formation, the basal formation of the Santa Fe Group in the Socorro area, was named by Denny (1940) for exposures in Arroyo Popotosa along the southeast side of the Ladron Mountains (T2N, R2W). Tonking (1957), unable to differentiate the Popotosa Formation from the overlying units of the Santa Fe Group on the basis of Denny's description, mapped all the Tertiary basin-fill as the Santa Fe Group. The Popotosa Formation and the Spears Formation were miscorrelated in the Bear Mountains and in the western Lemitar and Socorro mountains by Spiegel (1962) and Debrine and others (1963); see also Weber (1963). Bruning's (1973) comprehensive study of the Popotosa Formation in Socorro County corrected these miscorrelations. He interpreted the basin-fill sedimentary rocks along the northwest flank of the Bear Mountains as fanglomerate deposits of the Popotosa Formation. Bruning inferred an early Miocene age for these rocks based on Chapin's (1971a) 24 m.y. date for the interbedded La Jara Peak Basaltic Andesite. From pebble imbrications in exposures of the Popotosa Formation adjacent to the study area, Bruning derived a general flow direction toward the eastsoutheast, indicating source areas in the Gallinas Mountains and the Colorado Plateau. However, from pebble imbrications in basal beds of the Popotosa Formation interbedded with the La Jara Peak Basaltic Andesite, he inferred a southern source of detritus, possibly in the Magdalena Mountains (Bruning, 1973, p. 89).

Chapin and Seager (1975) state that the Popotosa depositional basin was originally about 40 mi (64 km) wide during middle Miocene time, and was subsequently segmented into three parallel 11- to 14-mile (18- to 22-km)-wide basins by uplift and block faulting during latest Miocene or early Pliocene time. The sedimentary fill along the eastern edge of the original Popotosa basin has been largely removed by erosion.

Brown (1972) mapped a facies of the Popotosa Formation in the southern Bear Mountains and Mulligan Gulch graben which he called the fanglomerate of Dry Lake Canyon. These sediments were shed westward off the north end of the ancestral Magdalena Range and consist almost entirely of andesitic detritus derived from the La Jara Peak Basaltic Andesite.

The eastern boundary of this study is the western boundary of the Popotosa outcrop belt. Therefore, only general observations of the basal 100 ft (30.5 m) are herein presented. It should be noted that the outcrops mapped as Popotosa Formation on the geologic map are only indurated rocks. The areas along the eastern boundary mapped as colluvium (Qco), piedmont gravels (Qpm), and pediment gravels (Qpg) may represent, in part, outcroppings of weathered Popotosa Formation.

The Popotosa Formation adjacent to the study area is a moderately to poorly indurated, very-light-gray (N8), weathering to yellowish-gray (5Y8/1), sandy conglomerate. Depositional contacts with underlying formations are rarely exposed in the study area. Exposures of such a contact with the La Jara Peak Basaltic Andesite east of Forest Road 123 (secs. 4 and 9, T1N, R5W), and with the Spears Formation east of Jaralosa Creek (sec. 25, T1N, R6W) show considerable angular unconformity.

Two particularly interesting outcrops should be noted. One of these, along a canyon floor east of Jaralosa Creek (SE 1/4, sec. 25, T1N, R6W) is a poorly indurated conglomerate composed principally of bleached clasts of A-L peak Tuff (pinnacles member?). Osburn (1978, oral commun.) suggested that this might record the position of a buried fault scarp. The other outcrop, at the canyon mouth southwest of the latter location (NW 1/4, sec. 36, T1N, R6W) exhibits lense-shaped bodies of conglomeratic sandstone as long as 8 ft (2.4 m) preserved at the base of La Jara Peak Basaltic Andesite flows. These conglomerates include cobbles derived from both the La Jara Peak Basaltic Andesite and the Hells Mesa Tuff and were mapped as Popotosa Formation.

Elsewhere, the Popotosa Formation is a poorly sorted conglomeratic sandstone with clasts derived principally from the La Jara Peak Basaltic Andesite, A-L Peak Tuff, and Hells Mesa Tuff. Lesser amounts of clasts of "turkey-track" andesite from the Spears Formation, chert, reddish-brown silicified siltstone, crystalline limestone, and rhyolitic tuff also occur. The long diameters of clasts range up to 7 in. (17.8 cm). Outcrops are thin- to thick-bedded and frequently exhibit wedge-shaped sets of low-angle crossbeds as much as 10-feet (3 m) thick. Graded beds are common, so that the tops are often comprised of very coarse-grained, poorly sorted sandstone (fig. 13). The sandy matrix of the conglomerates is generally angular and very poorly sorted. Calcite is the cementing material and sometimes resembles a caliche.

Figure 13: Thin, fining upward beds in the Popotosa Formation; roadcut in Corkscrew Canyon (SE 1/4, NW 1/4, sec. 1, T1S, R6W). Hammer handle is 12.5 in. (31.8 cm) long.

Tertiary Mafic Intrusives

Numerous dikes with orientations ranging from N75W to N25E transect the study area. Chapin and others (1974a,b) ascribe these dikes in the Magdalena area to intrusion of mafic magmas along extensional fault zones related to the Rio Grande rift. The majority of dikes trend between N30W and N10E and many occupy faults. Most dikes intrude strata older than the Spears Formation. However, four dikes intrude the Spears Formation east of Jaralosa Creek (E 1/2, sec. 24, and SE 1/4, sec. 13, T1N, R6W) and one short dike intrudes the Spears Formation in Dove Spring Canyon (SW 1/4, sec. 26, T1N, R6W). Also, the fault bounding the Mulligan Gulch graben is intruded in the northeastern corner of the study area (sec. 4, T1N, R5W) and again east of Jaralosa Creek (NW 1/4, sec. 19, T1N, R5W).

The dikes range in width from 1 to 6 ft (0.3 to 1.8 m) and in length from a few feet to several thousand feet. The longest dike is over 3 mi (4.8 km) long. Dips range from 70 degrees to vertical. In outcrop, the dikes generally form nonvegetated, resistant ridges which are easy to trace in the field and on aerial photographs. In one instance, where a dike intrudes the more resistant sandstones of the upper Tres Hermanos Member of the Mancos Shale west of La Jara Canyon (SW 1/4, sec. 6, T1N, R5W), the dike is weathered to a shallow depression.

Sills were not observed in rocks younger than the lower Mesaverde Formation. The majority of sills intrude shales of the Alamito Well Tongue and the lower marine section of the Tres Hermanos Sandstone in the northeastern section of the study area. One instance of a dike cutting a sill occurs within the Mesaverde outcrop area in the northeastern corner of the study area (sec. 7, T1N, R5W). Also, a dike which becomes a sill was observed along the west bank of Jaralosa Creek (SE 1/4, sec. 1 and NE 1/4, sec. 12, T1N, R6W).

Most of the dikes and sills are pervasively altered to a greenish-black (5G2/1) to dark-greenish-gray (5G4/1) color on fresh surfaces; on a weathered surface they are often stained by limonite to dusky yellowish brown (10YR2/2). A few of the dikes in the northeastern corner of the study area are slightly vesicular. Petrographically, they are comprised of felted plagioclase laths as much as 5 mm (0.2 in.) long (An60, average of 10 crystals, Michel-Levy method), ragged brown hornblende and brown biotite laths altering to magnetite and calcite, highly corroded pyroxene replaced by calcite and hornblende, apatite prisms, and scaly aggregates of light-green chlorite. The aphanitic groundmass is comprised of limonitic, spherulitic material.

Wall-rock alteration zones for these intrusives rarely exceed 1 ft (0.3 m) in width. Intruded sandstones are generally bleached and become slightly micaceous if the sandstone contained appreciable clay matrix. Intruded shales exhibit somewhat wider zones of alteration than do the intruded sandstones. Organic shales, such as are found in the Mesaverde Formation, are often blackened and well-indurated. Brown (1972) states that the pervasive alteration of intrusives and the presence of wall-rock alteration zones are suggestive of hydrothermal alteration subsequent to the emplacement of the intrusives.

Plio-Pleistocene and Holocene deposits

The oldest Plio-Pleistocene deposits occur along the eastern margin of the study area. These unconsolidated piedmont gravels overlie a gently sloping plain, above about 7000 ft (2134 m) in elevation, that was formed by coalescing alluvial fans derived from the Gallinas Mountains. This surface has been dissected by Jaralosa Creek and its tributaries so that it is now separated from the Gallinas Mountains. The gravels are a poorly sorted mix of subangular to rounded cobbles, pebbles, and sand. The principal lithologies of the clasts are Hells Mesa Tuff, A-L Peak Tuff, and the La Jara Peak Basaltic Andesite. Minor amounts of Precambrian granite and quartzite and Madera Limestone are also present. As previously noted, the lithology of these gravels is similar to that of the Popotosa Formation. A colluvial apron along the eroded edge of the piedmont surface makes distinction of these gravels from the Popotosa Formation difficult at some localities.

A remnant of an extensive pediment surface is exposed in the northeastern corner and north-central section of the study area. This surface is between 6740 and 6920 ft (2054 and 2109 m) in elevation and was developed on rocks of the lower Mesaverde Formation. The surface slopes gently northward and was probably graded to an ancestral Rio Salado. The composition of these gravels is much the same as that of the piedmont gravels, except that clasts of the nonvolcanic units are absent. These gravels are between 5 and 10 ft (1.5 to 3.0 m) thick and are usually cemented by white caliche.

Renewed downcutting by the Rio Salado caused dissection of the two previously discussed surfaces and formed the present drainage system. Earlier stages of this drainage may be represented by the older alluvial deposits (Qag) which mantle broad shallow topographic depressions. Terrace development accompanied the establishment of the present drainage system. The terraces are thin veneers of poorly sorted gravels on narrow benches along Jaralosa Creek and Abbe Spring Canyon. The present stream beds are comprised of heterolithic, subrounded, poorly sorted sand and gravel.

Structure

Regional Structure

Structural patterns in and around the study area were developed during late Cretaceous -- early Tertiary (Laramide) compressional folding and late Tertiary extensional faulting. Basaltic-andesite magmatic activity accompanied the early stages of extensional faulting. The Laramide episode of compressional tectonics formed broad uplifts and basins in New Mexico and adjacent areas during the time period of 50 to 75 m.y. B.P. (Chapin and others, 1974b). Elongate stocks, dikes, veins and joint sets of plutons in Arizona indicate an ENE-trending axis of regional compression (Rehrig and Heidrick, 1972, 1976). On the Colorado Plateau, Laramide structures have been described as broad folds, monoclines, and thrust belts (Kelley and Wood, 1946; Tonking, 1957; Kelley and Clinton, 1960; Woodward, 1976). The area of the present-day Magdalena Range was included in a broad Laramide uplift that encompassed most of the area west of the Rio Grande and south of San Acacia (Chapin and others, 1978). The northern and eastern boundaries of this uplift can be roughly established from the unconformable relationship of the Spears Formation with underlying rocks of Eocene to Paleozoic age.

By medial Eocene time, a neutral stress field replaced the Laramide compressional stress field (Chapin, 1974). Erosive beveling of Laramide uplifts continued during the succeeding 7 to 12 m.y. during a hiatus in volcanic activity (Chapin, 1974). Erosion proceeded until a surface of low relief extended throughout southern Colorado and New Mexico (Epis and Chapin, 1973). The Eocene Baca Formation, which unconformably overlies the Mesaverde Formation in the vicinity of the study area, is thought to be comprised of detritus from this episode of erosion (Snyder, 1971).

Widespread andesitic to latitic volcanism began about 37 m.y. B.P. and was followed by the formation of ignimbrite plateaus and numerous cauldron complexes (Chapin, 1974; Chapin and Seager, 1975; Elston and others, 1976). Detritus from the erosion of the early andesitic and latitic volcanic rocks was deposited as the Spears Formation. The Hells Mesa Tuff was erupted from the North Baldy cauldron about 33 m.y. B.P. and formed a broad ignimbrite shield over the Spears Formation. The Hells Mesa is representative of a calc-alkalic suite that was overlapped and succeeded in time on the Mogollon Plateau by a high-silica alkalic suite (Elston and others, 1976). The A-L Peak Tuff is representative of the latter suite in the study area. The change from calc-alkalic, two-feldspar tuffs to high-silica, one-feldspar tuffs and interbedded basaltic-andesite lavas marks the beginning of regional extension in the Magdalena area (Chapin, 1978).

The beginning of bimodal volcanism and east-west extension occurred in the Magdalena area about 32 m.y. B.P.

(Chapin, 1978). Numerous normal faults with an average N10W trend breached Oligocene batholiths, causing widespread pluton and dike emplacement (Chapin and others, 1974b). Uplift and erosion of the Magdalena Range throughout the episode of regional extension is reflected in the lithology and transport directions of the basal Popotosa Formation (Bruning, 1973). Voluminous basaltic-andesite magmas were penecontemporaneously extruded wherever areas of high heat flow were intersected by extensional faults (Chapin and Seager, 1975). This magmatic episode, represented by the La Jara Peak Basaltic Andesite in the study area, overlapped the end of the high-silica alkalic volcanic suite.

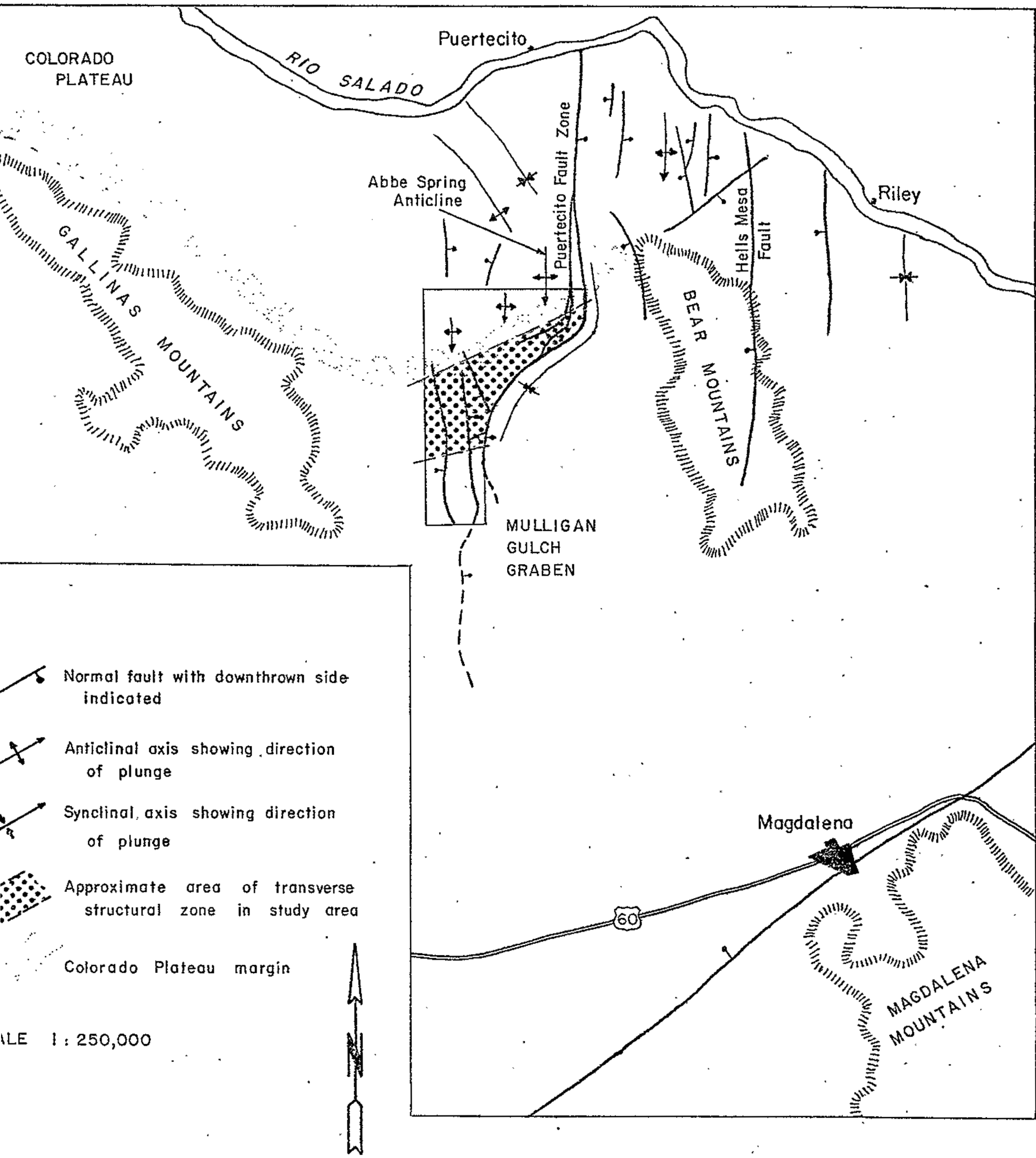
As north-trending normal faults continued to develop, northeast-trending basement faults were reactivated. Many of these are observable today as lineaments on Landsat imagery (Knepper, 1978). These transverse fault zones often absorbed the energy and movement of intersected north-trending faults. En echelon basins of the Rio Grande rift created by such transverse faults developed opposing structural asymmetry (Chapin, 1978; Chapin and others, 1978). The San Augustin graben formed as a bifurcation of the Rio Grande rift along the Morenci lineament, one of these northeast-trending structural zones. Tectonic influence by this lineament in the Socorro-Magdalena area has occurred throughout the past 32 m.y. (Chapin and others, 1978). The Colorado Plateau and the San Mateo and Gallinas ranges began to rise about 24 m.y. B.P. (Bruning, 1973; Bruning and

Chapin, 1974; Chapin and Seager, 1975). Volcanism declined during middle Miocene time (20 to 13 m.y. B.P.; Chapin, 1978). Block faulting accelerated in latest Miocene to Pliocene time. This renewed orogenic activity segmented the original 40-mi (64.4-km)-wide Popotosa depositional basin of the Socorro--Magdalena area into three grabens separated by intrarift horsts (Chapin and Seager, 1975). The Socorro and Lemitar mountains occupy the axis of the original Popotosa basin (Chapin, 1971b). One of these grabens, the Mulligan Gulch graben, lies along the eastern margin of the study area. The depth of this graben varies considerably over its extent (Wilkinson, 1976) and shallows southward across the Morenci lineament (Chapin and others, 1978). Epeirogenic uplift and rifting, accompanied in part by magmatic intrusion, has continued into the present (Reilinger and others, 1978). Figure 14 presents a view of the regional structural setting of the study area.

Local Structure

Folding:

The faulted nose of a southward-plunging anticline (here termed the Abbe Spring anticline, see fig.) is exposed in the northeast corner of the study area (see cross-section C-C'). Erosion has exposed the Chinle Formation in the center of the N10W-trending fold. The maximum dip produced



ure 14: REGIONAL STRUCTURAL SETTING OF STUDY AREA (compiled from Jackson, 1979; Massingill, 1979; Osburn, unpubl.)

by folding seldom exceeds 25 degrees. The center of the anticline is an intruded graben. Antithetic faulting, most notable in the closely spaced faults which cut the eastern limb of the anticline, has little associated fault-block rotation. According to Dennis (1972, p. 313), antithetic block faulting is often produced by keystone collapse of domal structures. In this part of the study area, such antithetic faulting has produced a series of eastward-dipping hogbacks capped by the resistant Dakota Sandstone. A small, tight, southwest-plunging anticlinal fold in the Tres Hermanos Sandstone occurs 0.75 mi (1.2 km) west of the axis of the Abbe Spring anticline. The axis of this small fold trends N15E and can be followed for about 0.75 mi (1.2 km). The fold cannot be followed to the north of the study area due to poor exposures and extensive alluvial cover. A down-to-the-west fault with about 500 ft (152.4 m) of displacement bounds this small anticline on the east. The eastern limb of the fold is approximately 1100 ft (338 m) wide and may represent reverse drag along this fault (see below). To the south the zone of eastward-dipping strata parallels the fault and becomes wider as displacement increases.

Two small anticlinal folds involve at least the basal 100 ft (30.4 m) of the Baca Formation. One of these is shown in cross-section B-B'. The axis of this fold trends N5W and plunges to the south. The crest of this anticline is an intruded horst block and is cut by faults trending oblique to the fold axis. Faulting and lack of exposures on both the east and west limbs prevent the determination of the minimum age of this fold. The other southward-plunging anticline, shown on cross-section A-A', occurs in the southwest corner of the study area. The axis of this anticline trends N25W. The eastern limb is truncated against a major down-to-the-west fault with 1200 ft (365.8 m) of throw. Reconnaissance mapping to the west of the study area suggests that the western limb is also truncated by a down-to-the-west fault.

The Abbe Spring anticline may not involve the Baca Formation and is almost certainly of Laramide origin from its broad and gentle form. The other folds in the study area involve strata of the basal Baca Formation. Faulting and lack of exposures make it impossible to determine whether folding has involved strata younger than this. Since the basal Spears Formation intertongues with the top of the Baca Formation in the study area, it is probable that folding which involves the Baca Formation also involve the Spears Formation. However, this would be at variance with regional evidence that Laramide compressional folding ended by about 50 m.y. B.P. (Chapin and others, 1974b; Elston, 1976) and that the Baca is a post-Laramide formation comprised of detritus eroded from Laramide uplifts (Snyder, 1971). One possible resolution of this problem is that a major unconformity may exist within the Baca Formation. The Baca Formation below this unconformity could have been involved in late-stage Laramide folding, while the sediments deposited above this surface would be relatively undeformed. No dateable fossils from the Baca have been found in the study area. Snyder (1970, 1971) reports an Eocene age for the Baca on the basis of vertebrate fossils collected 20 mi (32.1 km) northwest of the study area. This age is consistent with the other vertebrate age determination for the Baca (Gardner, 1910). However, the author has experienced difficulty in distinguishing the basal Baca from the underlying Cretaceous Mesaverde Formation at some locations. Other workers have reported similar problems in picking the Baca-Mesaverde contact (Tonking, 1957; Anonymous-II, 1963; Snyder, 1971). The basal Baca Formation at some locations may be of late Cretaceous or Paleocene age and these beds may have been folded during the Laramide deformation. At this time, however, the author and other researchers (Snyder, 1971; Johnson, 1978; Massingill, 1979; Cather, in prep.) have not observed such an unconformity.

An alternative, and perhaps better, explanation is suggested by Hamblin's (1965) observations of reverse drag along normal faults on the western Colorado Plateau. Such drag was found to be commonly associated with the downthrown blocks of faults with greater than 100 ft (30.5 m) displacement. Both of the fold axes under discussion roughly parallel major down-to-the-west faults with local displacement in excess of 800 ft (243.8 m). The fold in the southwestern corner of the study area is more difficult to interpret. Its axis roughly parallels a major fault for 1.5 mi (2.4 km), but then appears to diverge westward out of the study area. Exposures in the Spears outcrop belt north of this location and west of this fault are poor and yield diverse strike trends; however, the attitudes recorded suggest that this strata may also be folded anticlinally. North of the same transverse zone (fig. 16), displacement on this fault is greatly diminished and folding is not evident.

A discontinuous zone of steep east-dipping strata occurs adjacent to the western boundary fault system of the Mulligan Gulch graben. This zone is especially prominent in the volcanic rocks. The zone of steeply dipping strata has been recognized southward along the east flank of the Gallinas uplift to south of Highway 60, and may continue along the east flank of the San Mateo Mountains (Chamberlin, 1974). The width of this zone of steep dips in the study area is usually difficult to ascertain due to cover by pediment gravels or colluvium. In the vicinity of Abbe Spring (sec. 8, T1N, R5W), approximately 1500 ft (457 m) of volcanic section is involved in the steep zone of eastward dips. Chamberlin (1974) inferred a late-Oligocene age for the development of a 1-mi (1.6-km)-wide monoclinal hinge zone in the Council Rock area along this graben margin. The development of the eastward-dipping flexure preceded and was modified by basin and range faulting. Areas of decreased dip were ascribed by Chamberlin (1974) to subsequent fault rotation. Drag along the down-to-the-east boundary faults was eliminated as a mechanism because such drag was not observed to extend more than 500 ft (152 m) from faults elsewhere. Instead, Chamberlin (1974) proposed batholithic intrusion and concomitant upthrusting along north-trending basement faults as the mechanism for creating this monoclinal flexure (pp. 99-102).

The author cannot agree with Chamberlin's (1974) mechanism of formation for this zone of steeply dipping strata. This zone closely parallels the faults of the graben-boundary system in the study area and is most likely genetically related to them. The amount of displacement postulated for the faults of this system suggests drag folding as the most likely mechanism of formation. A syncline within the Popotosa Formation east of the study area is subparallel to the graben boundary (Osburn in Chapin and others, 1979) and may also have been caused by fault drag. The age of the monoclinal flexure along the west side of the Mulligan Gulch graben postdates the A-L Peak Tuff (about 32 m.y. B.P.) and predates the basalt of Council Rock (17 m.y. B.P.; Chapin, 1979, oral commun.).

Faulting:

Extensional fault trends in the study area range from N40W to due north, with the majority of faults clustered near N10W. Faulting is particularly severe in the northeast corner of the study area, where numerous step-faults are associated with keystone-collapse of the Abbe Spring anticline. The cross-sections suggest that step-faulting is the dominant structural style in this area. Displacement on any single fault seldom exceeds 100 ft (30.5 m). However, the total structural relief on several parallel, closely-spaced faults with the same sense of motion can frequently exceed 500 ft (152.4 m). Many of the longitudinal faults lose much or all of their relative movement upon crossing northeast-trending structural zones (see next section). Many faults are intruded by mafic dikes. Normal drag on faulted strata is locally associated with the major down-to-the-west faults. Reverse drag is believed to be associated with the major down-to-the-west faults as discussed in the preceding section. Fault planes exposed at the surface dip from 65 degrees to vertical (fig. 15). Hamblin (1965) and Anderson (1971) suggest that normal faults flatten at depth; if so reverse drag is required to fill the space created by the lateral component of fault movement. North-trending faults (except the western graben-boundary fault system) usually have down-to-the-west motion. These faults frequently curve toward the east as they approach the graben margin. This change of fault trend suggests rotation

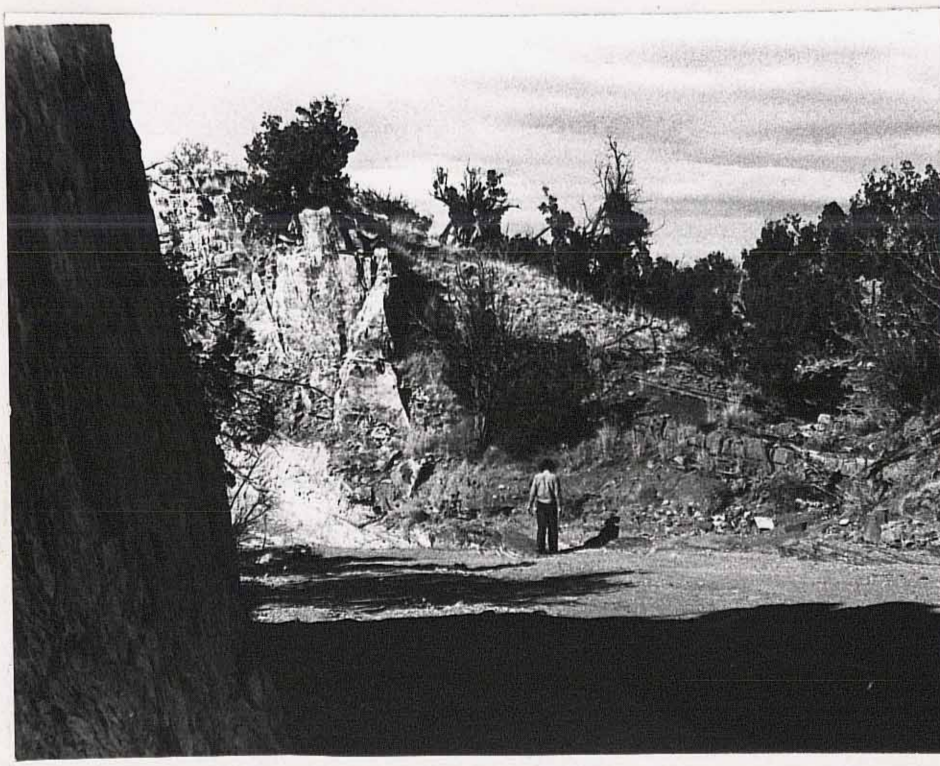


Figure 15: View along fault plane within the Dakota Sandstone; La Jara Canyon (NW 1/4, SW 1/4, sec. 5, T1N, R5W). Dakota Sandstone on the left is faulted against Alamito Well Tongue.

by subsequent movement on the graben boundary fault system. The north-trending normal faults cut formations as young as the Spears and are therefore of Oligocene or younger age.

Faults which cut the volcanic units along the edge of the Mulligan Gulch graben exhibit diverse trends and generally do not extend far into bordering outcrops of the Spears Formation. These faults have displacements of 100 ft (30.5 m) or less and produce jumbled blocks of volcanic rocks. It is likely that these small faults are genetically related to the intersection of transverse structural zones with the Puertecito fault system bounding the Mulligan Gulch graben.

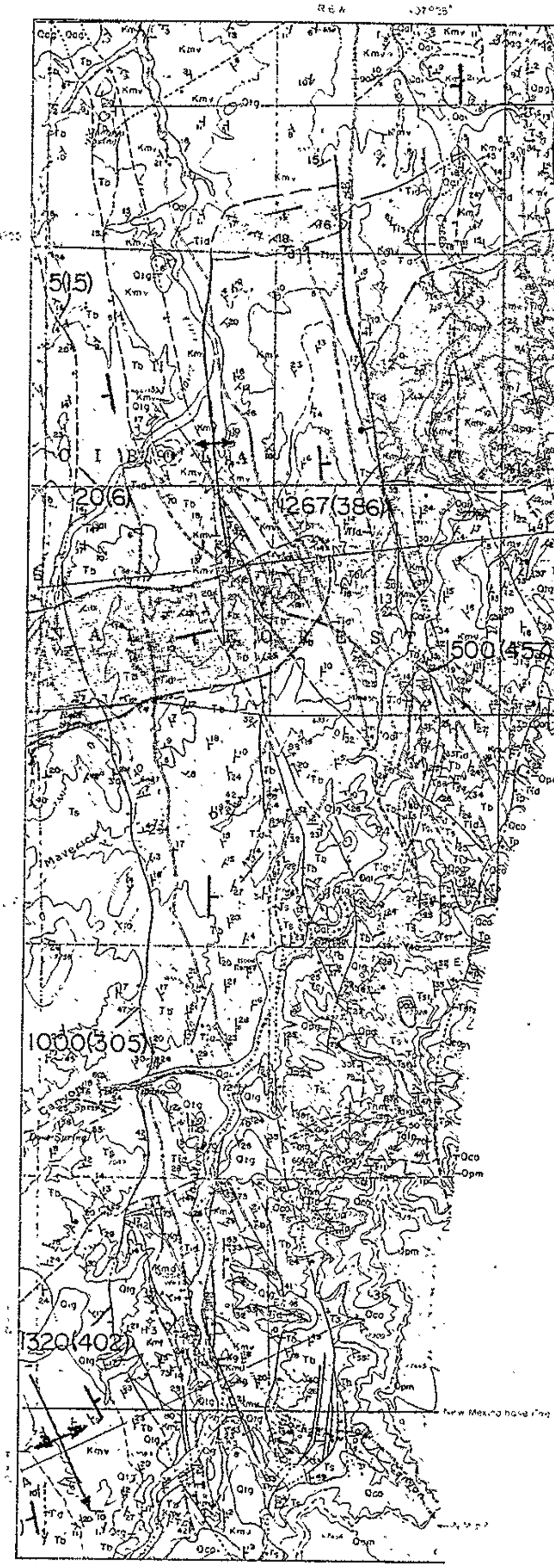
A fault system with predominantly down-to-the-east relative motion defines the western edge of the Mulligan Gulch graben. This graben, which is situated along the eastern margin of the study area, is filled primarily with sedimentary rocks of the Popotosa Formation. Winchester (1920) mapped a short segment of one of these boundary faults to the north of the study area and named it the Puertecito fault. Tonking (1957) also mapped the Puertecito fault north of the study area and estimated a displacement of 50 to 300 ft (15.2 and 91.4 m).

This fault is now recognized to be a wide zone consisting of one to ten faults (see also Jackson, 1979). The Puertecito fault zone is deflected laterally by zones of northeast-trending faults. In the northern part of the study area, the Puertecito fault zone juxtaposes Mesaverde or Baca rocks against upper La Jara Peak Basaltic Andesite. In the southern part of the study area, the Popotosa Formation is faulted down against the Baca or Spears formations. The age of the latest major displacement on this fault system is inferred to be late Miocene, concurrent with the division of the original Popotosa depositional basin into three segments (Chapin and Seager, 1975). Total displacement on the Puertecito fault system is estimated to range from 300 to 3000 ft (91.4 to 914.4 m) in the study area. Displacement decreases in the vicinity of the transverse fault zones where some of the movement of this boundary fault system is absorbed. East of the Hook Ranch (sec. 25, T1N, R6W), the Popotosa Formation unconformably overlies the Spears Formation and the pinnacles member of the A-L Peak Tuff. The Popotosa Formation along this contact is comprised dominantly of clasts of the subjacent lithologies, suggesting that an original fault contact has been buried. A discontinuous zone of steep east-dipping strata on the west side of the Puertecito fault system and a syncline in the Popotosa Formation east of this fault zone (Osburn, 1979) are inferred to represent normal drag.

Transverse structural zones:

The western edge of the Mulligan Gulch graben is deflected to the west twice in the study area by zones of northeast-trending faults. These faults have down-to-the-southeast motion. Volcanic rocks crop out in jumbled blocks along the northeast-trending fault zones adjacent to the western boundary fault system of the Mulligan Gulch graben. East of Jaralosa Creek (SE 1/4, sec. 25, T1N, R6W), an unusually thick section of the Pinnacles member of the A-L Peak Tuff crops out against an ENE-trending fault. This suggests that a fault scarp at the time of A-L Peak deposition formed a structural barrier. The limited and close association of the volcanic rocks with these transverse structural zones in the study area, the rapid northward thinning of these units (Chamberlin, 1979, oral commun.; Laroche, in prep.), and the existence of local unconformities between the Spears Formation, Hells Mesa Tuff and A-L Peak Tuff suggest even earlier structural influence of these zones. At Bird Spring (NW 1/4, sec. 9, T1N, R5W), Osburn (1979, oral commun.) has observed anomalously steep eastward dips in the Popotosa Formation where the northernmost transverse structural zone intersects the graben boundary fault system. This suggests that movement on the transverse fault system continued into Popotosa time.

Away from the edge of the Mulligan Gulch graben, the northernmost transverse structural zone becomes a discontinuous zone of southeastward-dipping strata that separates domains of predominantly east- and west-dipping strata (fig. 16). On color aerial photographs, the area of southeast-dipping strata in the Baca Formation is easily traced. The general trend of this transverse zone is suggested from the southwest elongation of the pediment



Numbers show approximate vertical throw on major faults in ft(m).

SCALE: 1:48,000

Figure 16: MAP SHOWING GENERALIZ
DOMAINS OF CONTRASTIN
STRIKE AND DIP

surface (SE 1/4, sec. 7, T1N, R6W) and from the upturning and termination of the Spears outcrop belt on the western margin of the study area (sec. 23, T1N, R6W).

The transverse structural zones apparently absorb some of the movement of intersecting longitudinal faults (fig. 16). North of Bird Spring (NW 1/4, sec. 9, T1N, R5W), the graben boundary fault system juxtaposes the lower Mesaverde Formation against the upper La Jara Peak Basaltic Andesite, requiring a total displacement of nearly 3000 ft (914.4 m). This displacement rapidly decreases southward. At a good exposure of the graben boundary fault near the Chavez Ranch headquarters (fig. 17) the Gray-massive member of the A-L Peak Tuff is faulted against the Popotosa Formation. Total displacement here is only 733 ft (223.3 m). Displacement decreases southward toward Abbe Spring (SE 1/4, sec. 8, T1N, R5W), and then gradually increases. South of the Hot Spot mine (sec. 18, T1N, R5W), displacement on the boundary fault system again begins to decrease towards the next transverse structural zone; first the Baca and then the Spears formations are juxtaposed with the Popotosa Formation. Other faults intersecting these transverse structural zones show similar changes of total displacement.

Another transverse structural zone may lie to the south of the study area. This is suggested by a westward bend in the boundary fault system, by exposures of the Hells Mesa Tuff (Laroche, 1979, oral commun.), and by the southward plunging of an anticlinal reverse drag fold in the southwest

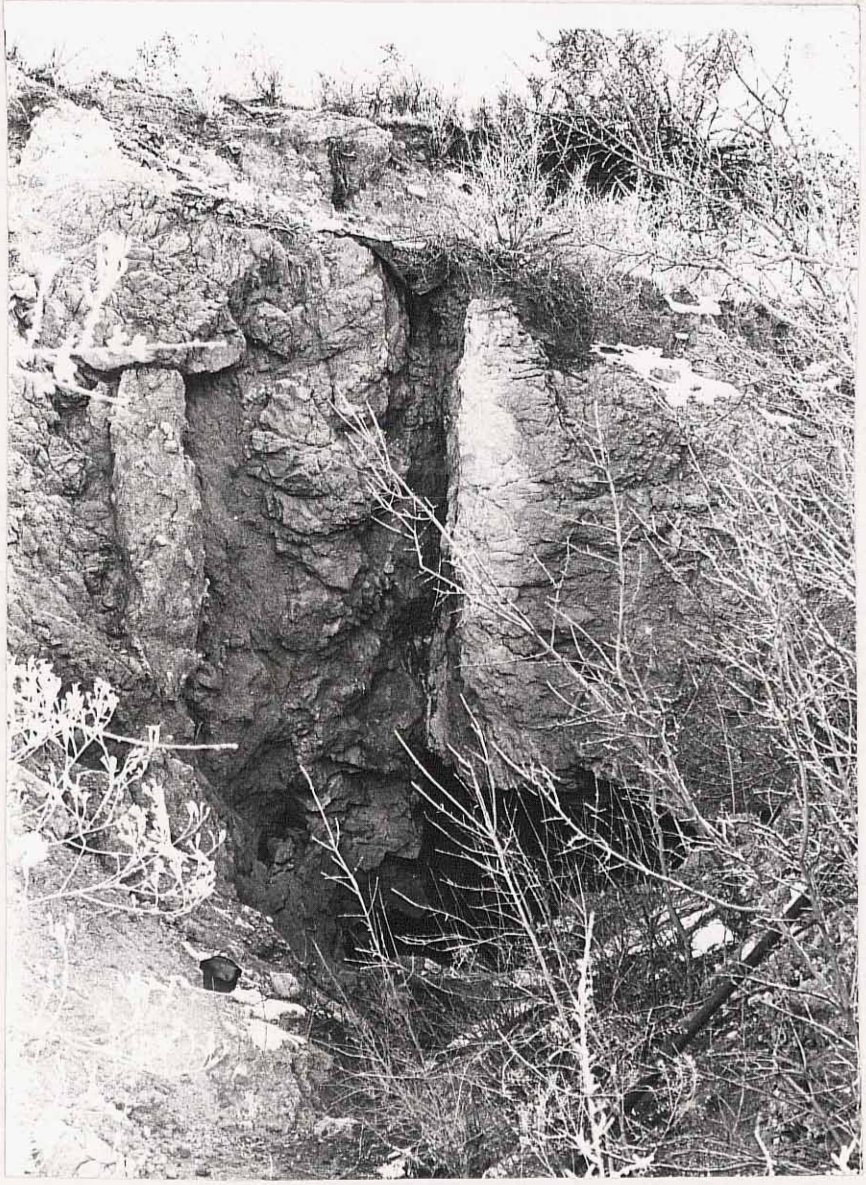


Figure 17: Puertecito fault; exposure behind Chavez Ranch headquarters (NW 1/4, sec. 9, T1N, R5W). Popotosa Formation on left fault against A-L Peak Tuff (Gray-massive member) on right. Excavation provided water for ranch house.

corner of the study area. Other southward-plunging reverse drag folds, discussed previously, terminate against these transverse zones, suggesting that the latest movement on these zones postdates the formation of these folds.

The trend of these transverse structural zones parallels that of a number of other northeast-trending shear zones now recognized to dominate the structural grain of basement rocks in the southern Rocky Mountains (Warner, 1978). Chapin and others (1979) have proposed the name "Tijeras lineament" for the transverse structural zone that crosses the study area. This lineament is recognized in the area northeast of the study area (Massingill, 1979), and crosses the Sandia uplift at Tijeras Canyon. Elsewhere, such transverse lineaments connect en echelon segments of the Rio Grande rift (Chapin, 1978). One of these, the Morenci lineament, crosses the Rio Grande rift at Socorro and has influenced local tectonics for the past 32 m.y. (Chapin and others, 1978). The Morenci lineament is expressed in the Magdalena area by a zone of northeast-trending faults (Brown, 1972; Chapin and others, 1974b), two of which show evidence of left-lateral movement consistent with northwestward drift of the Colorado Plateau (Chapin, 1971b). Chapin and others (1978) cite the following characteristics of transverse lineaments:

> "...these lineaments are deeply penetrating flaws in the lithosphere that tend to "leak" magmas and to influence deformation in the brittle near- surface rocks. undergoing rotation and step faulting in opposite directions...This shear zone is acting as an incipient transform fault connecting segments of the Rio Grande rift." (p.115)

These transverse zones are also characterized by high heat flow and, thus, may be exploration targets for geothermal energy.

Economic Geology

Uranium (see fig. 18 and Table 1)

Regionally, the Baca Formation is considered to offer the best possibilities for uranium mineralization. A small uranium prospect exists north of the Hook Ranch headquarters on the west bank of Jaralosa Creek (sec. 13, T1N, R6W). The mineralization occurs in a lenticular, very poorly sorted, conglomeratic sandstone stratigraphically about 150 ft (45.7 m) below the top of the Baca. Uranium mineralization occurs as haloes around abundant silicified wood. A small amount of this sandstone body was mined during the middle 1950's. Preparations for an *in situ* leaching operation were begun by M.P. Grace in 1974-1975. These preparations included the drilling of several shallow exploration holes in an attempt to delineate the extent of the mineralized sandstone. The project was abandoned before completion due to a contract dispute. To date further exploitation of this small prospect has not been attempted (C.T. Smith, 1978, oral commun.).

Uranium mineralization occurs at several other localities in the study area in conjunction with limonitic wood casts. Anonymous (1959) reports uranium mineralization in purple rocks of the Baca Formation in the Jaralosa Creek area; the uranium is either disseminated in lenticular

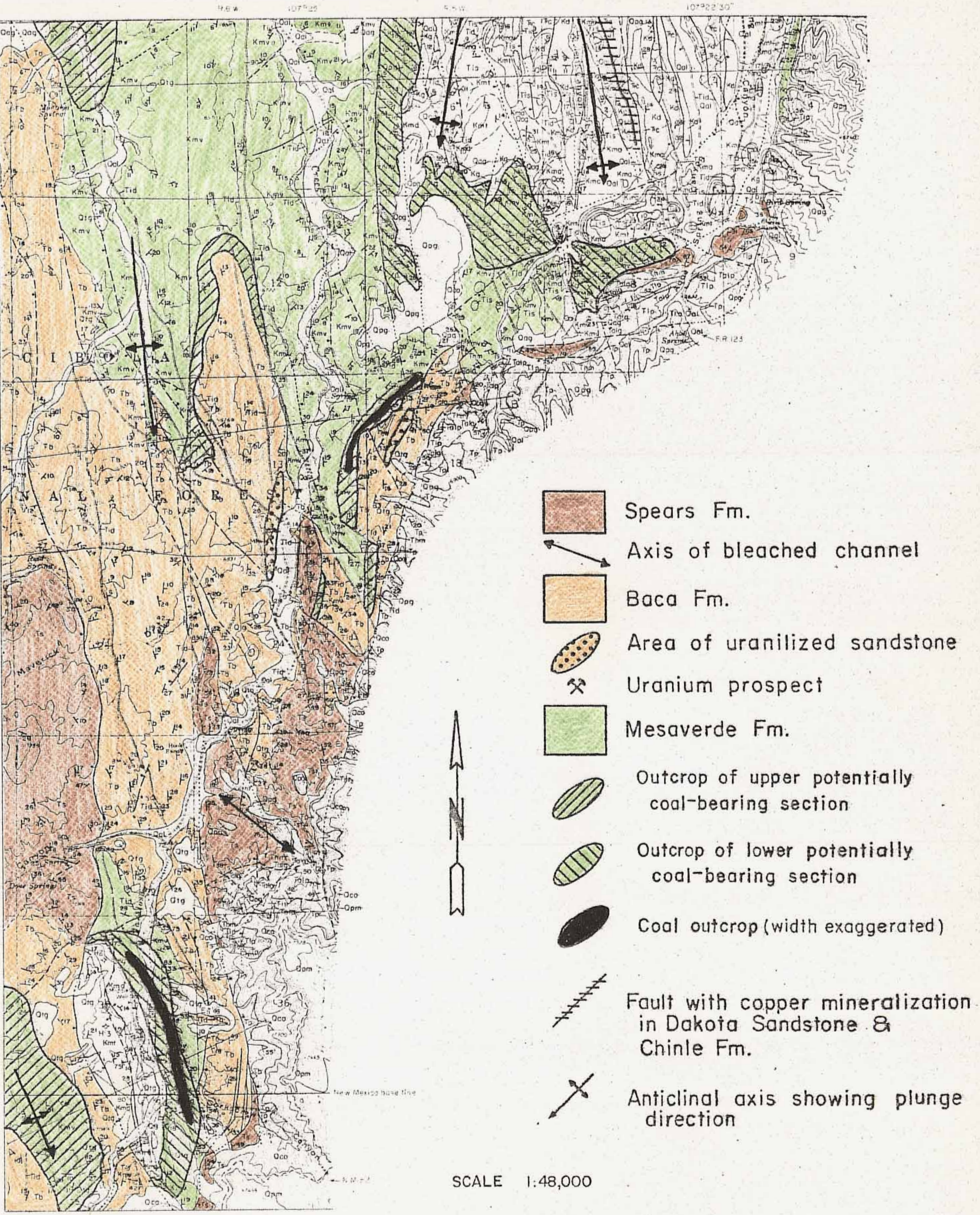


Figure 18: STRATIGRAPHIC INTERVALS FAVORABLE FOR COAL AND URANIUM DEPOSITS

Table 1: Uranium analyses
Data taken from Bachman, Baltz, and Griggs (1957).

Sample	Location	Formation	Eq. U3O8	Ch. U3O8	Ch. V2Ø5
A	18,1N,5W	Baca	Ø.14%	Ø.26%	Ø.1%
B	18,1N,5W	Mesaverde	Ø.ØØ1%	–	–
C	13,1N,6W	Baca	2.Ø%	3.27%	9.21%
D	24,1N,6W	Baca	Ø.24%	Ø.19%	2.98%
E	35,1N,6W	Baca	Ø.13%	Ø.Ø36%	Ø.1%

channel sandstones or associated with carbonaceous debris. He suggests that the purple color may be caused by vanadium compounds. In general, most of the Baca Formation within the study area does not appear to be as favorable for uranium concentration as it appears to be to the east in the Riley area (Chapin and others, 1979). Sandstone bodies are generally thin and lenticular with little organic trash. Reduced shale occurs only as small lense-shaped bodies at the base of sandstones, primarily in the lower part of the Baca Formation. Average grain size increases and sorting becomes poorer towards the top of the Baca, possibly favoring the movement of groundwater.

The top of the Cretaceous Mesaverde Formation may also offer some potential for uranium concentration. Nonuraniferous radioactivity was detected from carbonaceous rocks of this formation locally (Anonymous, 1959, p. 135). The Mesaverde is an aggregate of fluvial sandstones, organic shales, siltstones, and coal exposed over, or underlying, approximately 50% of the study area. In the Red Basin area, west of the study area, uranium occurrences are reported from organic-rich sandstones in the upper portion of this formation (Anonymous, 1959).

The most likely origin for the uranium is leaching from overlying rhyolitic tuffs (Anonymous, 1959). Anonymous suggests, alternatively, that the uranium may have been derived from ascending hydrothermal fluids associated with the eruptive centers of the volcanic units to the south.

Anonymous (1959) states that there is little evidence to support a syngenetic sedimentary origin for the uranium mineralization in this area. The observation of bleached mudflow deposits and conglomerates in channels of the lower Spears Formation east of Jaralosa Creek (sec. 25, T1N, R6W) may indicate passage of mineralized waters. Uranium deposits in the upper Rio Salado area appear to be related to groundwater movement as influenced by structure, permeability, and the presence of organic material (Anonymous, 1959).

Coal (see fig. 18 and Table 2)

The basal 200 ft (61 m) of the Mesaverde Formation is comprised of organic shales, very-fine-grained sandstones and several coal beds. These coals are interpreted to have accumulated in a lower coastal or deltaic plain environment with brackish water influence (M. Chaiffetz in Chapin and others, 1979). At exposures near Corkscrew Canyon, coal beds range from 1- to 2-ft (0.30 to 0.61 m)-thick. These beds exhibit rapid lateral and vertical gradation into organic shales. At least eight of these coal-bearing lithologic sequences occur here; more may be covered. Although this coal-bearing section of the basal Mesaverde is recognized to the north (Jackson, 1979) and northeast (Massingill, 1979), faulting and surficial cover conceal it throughout most of the study area. Chapin and others (1979),

Table 2: Coal analysis

Analysis of sample from Hot Spot Mine (NW 1/4, NW 1/4, sec. 18, T1N, R5W). Analysis provided by United States Bureau of Mines (1976). Values reported for coal sample as received.

Proximate analysis:		
	Moisture	6.6%
	Volatile matter	32.8%
	Fixed carbon	54.2%
	Volatile matter	32.8%
	Ash	6.4%
Ultimate analysis:		
	Hydrogen	5.0%
	Carbon	67.2%
	Nitrogen	1.3%
	Sulfur	0.5%
	Oxygen	19.6%
	Ash	6.4%
Heating value		11,555 BTU

from their regional observations of coal occurrences, suggest that the northeast-trending Tijeras lineament may have influenced coal development in the lower Mesaverde. They observe that coal is more abundant in the lower Mesaverde southeast of the lineament and attribute this to possible differences in subsidence and sedimentation rate across the Tijeras lineament.

Coal from the uppermost strata of the Mesaverde Formation was mined from three adits at the Hot Spot mine (NW 1/4, sec. 18, T1N, R5W). The thickest coal seam here is 5 ft (1.5 m). These coals grade laterally into silty shale and silty sandstone within 0.75 mi (1.2 km); outcrops do not occur elsewhere in the study area. Federal records show that a mining permit was issued to F.L. Dugger on July 15, 1927. Dugger drove a 90-ft (27.4 m) drift into a 46 in. (116 cm) coal bed before encountering a shale roll which thinned the bed to 12 in. (30.5 cm) (Nichelson and Frost, in press). Efforts to continue mining through the roll and to meet the coal bed on the other side were unsuccessful and hampered by financial problems. The prospect produced a recorded 85 tons of coal through June 30, 1931, before the expiration of the permit (Nichelson and Frost, in press). Little remains at the site today. One adit remains open, but it is in very poor condition. The rock outline of a small shelter is also visible near the largest mine dump.

Oil and Gas

Hydrocarbon shows from an unspecified unit in the subsurface were reported from drillhole H1 (sec. 35, T1N, R6W; D. Belknap, 1978, oral commun.). Elsewhere in Socorro County, hydrocarbons are reported from Pennsylvanian strata (Anonymous-III, 1963). Although numerous intrusions occur in the study area, thermal effects upon intruded Cretaceous strata are minor and restricted to narrow zones along the margins of intrusives (M. Chaiffetz in Chapin and others, 1979). In addition, burial depth of Cretaceous strata in the thesis area is very shallow. Thus, it would appear that the potential for oil and gas generation from the Cretaceous units is poor.

Structural fold traps for possible hydrocarbon accumulation are limited to three southward-plunging anticlines. Of these, the two in the northeast and northwest sections are extensively faulted. The small anticline mapped in the southwestern corner of the study area is truncated on both flanks by major NNW-trending faults and cut by two northeast-trending faults of minor displacement. Additional hydrocarbon traps may have been created by the major normal faults mapped in the study area (see fig. 16). Chapin and others (1979) note apparent changes of sedimentation rates across the Tijeras lineament reflected within Cretaceous strata (the Twowells Sandstone, Gallego Sandstone, and the Mesaverde coal beds). They suggest that these changes of sedimentation rates were caused by differential movement on the Tijeras lineament. Similar movement on the Tijeras lineament during Paleozoic sedimentation may have caused facies changes and thickness variations. These, in turn, may have influenced hydrocarbon entrapment.

Other

One fault cutting the Chinle Formation and the Dakota Sandstone in the northeast corner of the study area has an associated narrow zone of copper mineralization (sec. 5, T1N, R5W). This was prospected in two locations by a shallow shaft and an adit. Minerals observed on the dumps include malachite, chalcocite, and bornite. No production records are known. The quantity of dump material does not suggest a very long or extensive production history. The remains of two stone buildings are located nearby. A small exploration pit dug into a similar fault on the eastern side of the Dakota outcrop does not show mineralization. A short chalcocite vein was observed along a fault south of the Hot Spot mine within the Baca Formation (SW 1/4, NW 1/4, sec. 18, T1N, R5W).

The San Andres Limestone and other Paleozoic carbonate units which underlie the study area (Tonking, 1957) may have locally generated CO2 gas during times of high heat flow and igneous intrusion. However no data is available to evaluate this possibility.

Conclusions

1. Stratigraphy

a) Rock units of the following formations, from oldest to youngest, are exposed at the surface within the study area: Chinle Formation, Dakota Sandstone, Alamito Well tongue of the Mancos Shale, Tres Hermanos Sandstone Member of the Mancos Shale, D-Cross Tongue of the Mancos Shale, Gallego Sandstone, Mesaverde Formation, Baca Formation, Spears Formation, Hells Mesa Tuff, A-L Peak Tuff, La Jara Peak Basaltic Andesite, and the Popotosa Formation. The differentiation and nomenclature of these units reflect advances in stratigraphy since Tonking's (1957) reconnaissance map of the Puertecito quadrangle which included the area of this report.

b) Lithologic and paleontologic evidence suggest that the Dakota Sandstone-lower Tres Hermanos Sandstone and the upper Tres Hermanos Sandstone-lowermost Mesaverde Formation record two major transgressive-regressive cycles. The final eastward marine regression from the study area began during deposition of the D-Cross Tongue, and was followed by deposition of the alluvial and paludal deposits of the nonmarine portion of the Mesaverde Formation.

c) The volcaniclastic Spears Formation intertongues with the top of the Baca Formation in the study area. South towards the Magdalena Mountains, the Spears Formation unconformably overlaps successively older Paleozoic strata. Therefore the study area must be located in an area of continuous sedimentation from Baca into Spears time.

d) The Hells Mesa and A-L Peak tuffs are relatively thin, and other tuffs found to the south are missing. In addition, the Hells Mesa Tuff and the gray-massive member of the A-L Peak Tuff fill channels eroded into underlying units. These stratigraphic relationships have not been observed to the south and indicate that the study area was located near the distal edge of these ash-flow sheets. Differential vertical movements along the Tijeras lineament may also have influenced the thickness of the ash-flow sheets.

e) The pinnacles member of the A-L Peak Tuff is interbedded within the La Jara Peak Basaltic Andesite.

2. Structure

a) One broad, southward-plunging anticlinal fold, the Abbe Spring anticline, is exposed in the study area. The Mesaverde Formation apparently is the youngest strata that was folded. This anticline most likely was formed by the late Cretaceous-early Tertiary Laramide orogeny. No other Laramide structures within the study area can be definitely identified.

b) Regional extension related to development of the Rio Grande rift fractured the study area with numerous, closely spaced, predominantly down-to-the-west normal faults. Mafic dikes were emplaced in many of these north-trending faults. The larger of these faults -- those with greater than 500 ft (152.4 m) of vertical displacement -- are approximately paralleled on their downthrown sides by axes of narrow anticlines. These folds, affecting rocks as young as the Spears Formation, probably are caused by reverse drag effects similar to those described by Hamblin (1965).

c) The Mulligan Gulch graben, situated along the eastern border of the study area, was formed during late Oligocene-early Miocene block faulting. It is bounded along its western edge by a complex system of predominantly down-to-the-east faults. The steep eastward dip of strata on both sides of this fault system is unusual for late Cenozoic normal faults and may reflect reactivation of a major structure in the subsurface.

d) The Mulligan Gulch graben margin is deflected westward in two areas along down-to-the-south transverse fault zones. Away from the graben margin, the more northerly of these zones is evidenced by discontinuous regions of predominantly southeast-dipping strata. Across these zones, the regional dip is reversed and fault displacement decreases. These transverse structural zones are believed to represent flexures above basement faults of the northeast-trending Tijeras lineament.

3. Economic

a) The basal 200 ft (61 m) of the Mesaverde Formation contains coal, organic shales, and intercalated thin sandstones. Outcrops of at least eight 1- to 2-ft (0.30- to 0.61-m)-thick coal beds occur in the vicinity of Corkscrew Canyon. The coal beds grade laterally and vertically into organic shales. Locally thick, but discontinuous, coal beds at the top of this formation in similar lithologic packages have been exploited in the past at the Hot Spot mine (sec. 18, T1N, R5W).

b) Exploration efforts for uranium have been concentrated in the Baca Formation based upon characteristics favorable to sandstone-type uranium mineralization. These characteristics are bleached, permeable sandstones containing organic matter and traces of mineralization. The Hook uranium prospect (sec. 13, T1N, R6W) has been developed on a small scale. On the whole, however, the Baca Formation of the study area appears to be unfavorable for discovery of large uranium orebodies because of the lenticularity of the channel sandstones. The top of the Mesaverde Formation in some locations appears to have the same lithologic characteristics which would be favorable for uranium concentration. Many of the sandstones in this formation are more areally extensive than are sandstones of the Baca Formation.

References Cited

Allen, J.E., and Balk, R., 1954, Mineral resources of Fort Defiance and Tohatchi quadrangles, Arizona and New Mexico: New Mexico Bureau of Mines and Mineral Resources Bulletin 36, p. 192.

Anderson, R.E., 1971, Thin skin distension in Tertiary rocks of south- eastern Nevada: Geological Society of America Bulletin, v. 82, p. 43-58.

Anonymous, 1959,; Uranium deposits in the Datil Mountains-Bear Mountains region, New Mexico: New Mexico Geological Society, 10th Field Conference, Guidebook, p. 135-143.

Anonymous-II, 1963, Baca Formation in the area around Socorro, New Mexico: New Mexico Geological Society, 14th Field Conference, Guidebook, p. 100-101.

Anonymous-III, 1963, Oil and gas in Socorro County, New Mexico: New Mexico Geological Society, 14th Field Conference, Guidebook, p. 217- 219.

Brown, D.M., 1972, Geology of the Southern Bear Mountains, Socorro County, New Mexico [M.S. thesis]: Socorro, New Mexico Institute of Mining and Technology, 110 pp.

Bruning, J.E. (1973); Origin of the Popotosa Formation, North-central Socorro County, New Mexico; unpubl. Ph.D. dissertation, New Mexico Institute of Mining and Technology, 132 p.

Bruning, J.E. and Chapin, C.E., 1974, The Popotosa Formation -- a Miocene record of Basin and Range deformation, Socorro County, New Mexico [abs.]: Geological Society of America Abstracts with Programs, v. 6, no. 5, p. 430.

Burke, W.H., Kenny, G.S., Otto, J.B., and others, 1963, Potassium -argon dates, Socorro and Sierra Counties, New Mexico: New Mexico Geological Society 14th Field Conference, Guidebook, p. 224.

Burton, G.C., 1955, Sedimentation and stratigraphy of the Dakota Formation in the San Juan Basin: Four Corners Geological Society Guidebook, p. 78-88.

Cather, S.M., in prep., Sedimentologic study of the Baca Formation, Jaralosa Creek area, Socorro County, New

Mexico [M.S. thesis]: Austin, University of Texas.

Chamberlin, R.M., 1974, Geology of the Council Rock District, Socorro County, New Mexico [M.S. thesis], Socorro, New Mexico Institute of Mining and Technology, 134 p.

Chapin, C.E., 1971a, K-Ar age of the La Jara Peak Andesite and its possible significance to mineral exploration in the Magdalena mining district, New Mexico: Isochron/West, no. 2, p. 43-44.

___________, 1971b, The Rio Grande Rift, part I: modifications and additions: New Mexico Geological Society, 22nd Field Conference, Guidebook, p. 191-201.

___________, 1974, Three-fold tectonic subdivision of the Cenozoic in the Cordilleran foreland of Colorado, New Mexico and Arizona [abs.]: Geological Society of America Abstracts with Programs, v. 6, no. 5, p. 433.

___________, 1978, Evolution of the Rio Grande Rift: Comparisons between segments and the role of transverse structures [abs.]: 1978 International Symposium on the Rio Grande Rift, Los Alamos Scientific Laboratory, New Mexico, p. 24-27.

Chapin, C.E., Blakestad, R.B., Bruning, J.E., and others, 1974a, Exploration framework of the Magdalena area, Socorro County, New Mexico [abs.]: New Mexico Geological Society, 25th Field Conference, Guidebook, p. 380.

Chapin, C.E., Blakestad, R.B., and Loring, A.K., 1974b: New Mexico in the Cenozoic Tectono-Magmatic Framework [abs.]: New Mexico Geological Society 25th Field Conference, Guidebook, p. 380 -381.

___________, Chamberlin, R.N., Osburn, G.R., and others, 1978, Exploration framework of Socorro geothermal area, New Mexico: New Mexico Geological Society Special Publication 7, p. 115-120.

Chapin, C.E., Osburn, G.R., Hook, S.C., and others, 1979, Coal, uranium, oil and gas potential of the Riley-Puertecito area, Socorro County, New Mexico: New Mexico Bureau of Mines and Mineral Resources Open-File Report 103, 33 p.

Chapin, C.E., and Seager, W.R., 1975, Evolution of the Rio Grande rift in the Socorro and Las Cruces area: New Mexico Geological Society, 26th Field Conference, Guidebook, p. 297- 321.

Clifton, H.E., 1969, Beach lamination: nature and origin: Marine Geology, v. 7, p. 553-559.

Cross, W., 1899, Description of the Telluride Quadrangle (Colorado): U.S. Geological Survey Atlas, Telluride Folio 57.

Dane, C.H., and Bachman, G.O., 1957, The Dakota Sandstone and Mancos Shale in the Gallup Area: Four Corners Geological Society, Guidebook, p. 95-98.

Dane, C.H., Landis, E.R., and Cobban, W.A., 1971, The Twowells Sandstone Tongue of the Dakota Sandstone and the Tres Hermanos Sandstone as used by Herrick (1900), Western New Mexico: U.S. Geological Survey Professional Paper 750-B, p. B17-B22.

Dane, C.H., Wanek, A.A., and Reeside, J.B. Jr., 1957, Reinterpretation of a section of Cretaceous rocks in Alamosa Creek valley area, Catron and Socorro Counties, New Mexico: American Association of Petroleum Geologists Bulletin, v. 41, p. 181-196.

Darton, N.H., 1928, "Red Beds" and associated formations in New Mexico, with an outline of the geology of the state: U.S. Geological Survey Bulletin 794.

Deal, E.G., 1973, Geology of the northern part of the San Mateo Mountains, Socorro County, New Mexico: A study of a rhyolite ash-flow tuff cauldron and the role of laminar flow in ash-flow tuffs [Ph.D. dissertation]: Albuquerque, University of New Mexico, 136 p.

Deal, E.G., and Rhodes, R.C., 1976, Volcano-tectonic structures in the San Mateo Mountains, Socorro County, New Mexico, in Cenozoic Volcanism in New Mexico: University of New Mexico Publications in Geology, no. 5.

Debrine, B., Spiegel, Z., and Williams, D., 1963, Cenozoic sedimentary rocks in Socorro valley, New Mexico [abs.]: Geological Society of America, Rocky Mountain section, 16th Annual Meeting Program, p. 22-23.

Dennis, J.G., 1972, Structural Geology: New York, The Ronald Press Company, 532 p.

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

Elston, W.E., 1976, Tectonic significance of mid-Tertiary volcanism in the Basin and Range province: a critical review with special reference to New Mexico, in Cenozoic volcanism in southwestern New Mexico: New

Mexico Geological Society Special Publication 5, p. 93-102.

Elston, W.E., Rhodes, R.C., Coney, P.J., and others, 1976, Progress report on the Mogollon Plateau volcanic field, southwestern New Mexico, no. 3, Surface expression of a pluton: New Mexico Geological Society Special Publication 5, p. 3-28.

Epis, R.C. and Chapin, C.E., 1973, Geomorphic and tectonic implications of the post-Laramide, late Eocene erosion surface in the Southern Rocky Mountains [abs.]: Geological Society of America Abstracts with Programs, v. 5, no. 6, p. 479.

Folk, R.A., 1968, Petrology of sedimentary rocks: Austin, University of Texas, 170 p.

Gardner, J.H., 1910, The Carthage coal field, New Mexico: U.S. Geological Survey Bulletin 381, p. 452-460.

Givens, D.B., 1957, Geology of Dog Springs quadrangle, New Mexico: New Mexico Bureau of Mines and Mineral Resources Bulletin 58, 40 p.

Gregory, H.E., 1917, Geology of the Navajo County: U.S. Geological Survey Professional Paper 93.

Hamblin, W.K., 1965, Origin of "reverse drag" on the downthrown side of normal faults: Geological Society of America Bulletin v. 76, no. 10, p. 1145-1164.

Herrick, C.L., 1900, Report of a geologic reconnaissance in western Socorro and Valencia Counties, New Mexico: American Geologist no. 25, p. 331-346.

Holmes, W.H., 1877, Report on the San Juan district: U.S. Geological and Geographic Survey Terr., 9th Annual Report, p. 237-276.

Hook, S.C., and Cobban, W.A., 1977, Pycnodonte newberri (Stanton) -- Common Guide Fossil in Upper Cretaceous of New Mexico: New Mexico Bureau of Mines and Mineral Resources Annual Report, July 1, 1976 to June 30, 1977, p. 48-54.

Hunt, C.B., 1936, Geology and fuel resources of the southern part of the San Juan Basin, New Mexico -- Part 2 -- The Mount Taylor Coal Field; U.S. Geological Survey Bulletin 860-B, p. 31-80.

Ingram, R.L., 1954, Terminology for the thickness of stratification and parting units in sedimentary rocks: Geological Society of America Bulletin v. 65, p. 937-938.

Jackson, R.A., 1979, Geology of the Puertecito-La Cruz Peak area, Socorro County, New Mexico [M.S. thesis], Socorro, New Mexico Institute of Mining and Technology.

Johnson, B.D., 1978, Genetic stratigraphy and provenance of the Baca Formation, New Mexico and the Eagar Formation, Arizona [M.S. thesis], Austin, University of Texas, 150 p.

Kelley, V.C. and Clinton, N.J., 1955, Fracture system and tectonic elements of the Colorado Plateau: University of New Mexico Publications in Geology, no. 6.

Kelley, V.C. and Silver, C., 1952, Geology of the Caballo Mountains, with special reference to regional stratigraphy and structure and to mineral resources, including oil and gas: New Mexico University Publications in Geology Series, no. 4, 286 p.

Kelley, V.C. and Wood, G.H., 1946, Lucero uplift, Valencia, Socorro and Bernalillo counties, New Mexico: U.S. Geological Survey Oil and Gas Investigations (preliminary), Map 47.

Knepper, D.H., 1978, Analysis of linear features, Rio Grande rift zone, central New Mexico [abs.]: 1978 International Symposium on the Rio Grande Rift, Los Alamos Scientific Laboratory, New Mexico, p. 48-49.

Krumbein, W.C. and Sloss, L.L., 1963, Stratigraphy and Sedimentation (second edition): San Francisco, W.H. Freeman and Company, 660 p.

Laroche, T.M., in prep., Geology of the Gallinas Peak Area, Socorro County, New Mexico [M.S. thesis]: Socorro, New Mexico Institute of Mining and Technology.

Lee, W.T., 1915, Relation of the Cretaceous formations to the Rocky Mountains in Colorado and New Mexico: U.S. Geological Survey Professional Paper 95, p. 27-58.

Massingill, G.L., 1979, Geology of Riley-Puertecito area, southeastern margin of Colorado Plateau, Socorro County, New Mexico [Ph.D. dissertation], El Paso, University of Texas, 301 p.

Meek, F.B., and Hayden, F.V., 1862, Description of new lower Silurian (Primordial), Jurassic, Cretaceous, and Tertiary fossils collected in Nebraska Territory...with some remarks on the rocks from which they were obtained: Proceedings of the Academy of Natural Science, Philadelphia, v. 13, p. 419-420.

Reineck, H.E., and Singh, I.B., 1975, Depositional Sedimentary Environments: New York, Springer-Verlag, 439 p.

Selley, R.C., 1972, Ancient Sedimentary Environments: New York, ornell University Press, 237 p.

Simon, D.B., 1973, Geology of the Silver Hill Area, Socorro County, New Mexico [M.S. thesis], Socorro, New Mexico Institute of Mining and Technology, 101 p.

Smith, C.T., 1955, Uranium occurrences on the Colorado Plateau: Four Corners Geological Society Guidebook, p. 169-176.

Smith, E.I., Aldrich, J.M., Deal, E.G., and others, 1974, Fission track ages of Tertiary volcanic rocks, Mogollon Plateau, southwestern New Mexico: New Mexico University Publications in Geology, no. 8.

Snyder, D.O., 1970, Fossil evidence of Eocene age of Baca Formation, New Mexico: New Mexico Geological Society, 21st Field Conference, Guidebook, p. 65-67.

__________, 1971, Stratigraphic analysis of the Baca Formation, west-central New Mexico [Ph.D. dissertation]: Albuquerque, University of New Mexico, 160 p.

Spiegel, Z., 1962, Preliminary report on the geology and water resources of the Socorro region: unpubl. manuscript in files of Ground-Water Department, RD and D, Socorro, New Mexico Institute of Mining and Technology.

Terry, R.D. and Chilingar, G.V., 1955, Visual estimation of percentage composition: Journal of Sedimentary Petrology, v. 25, no. 3.

Tonking, W.H., 1957, Geology of Puertecito Quadrangle, Socorro County, New Mexico: New Mexico State Bureau of Mines and Mineral Resources Bulletin 41, 67 p.

Warner, L.A., 1978, The Colorado lineament: a middle Precambrian wrench fault system: Geological Society of America Bulletin 89, p. 161-171.

Weber, R.E., 1963, Cenozoic volcanic rocks of Socorro County: New Mexico Geological Society, 14th Field Conference, Guidebook, p. 132-143.

__________, 1971, K-Ar ages of Tertiary igneous rocks in central and western New Mexico: Isochron/West, no. 1, p. 33-45.

Weber, R.E., and Bassett, W.A., 1963, K-Ar ages of Tertiary volcanic and intrusive rocks in Socorro, Catron, and Grant counties, New Mexico: New Mexico Geological Society, 14th Field Conference, Guidebook, p. 220-223.

Weber, R.E., and Willard, M.E., 1963, Field trip 1, western Bear Mountains: New Mexico Geological Society, 14th Field Conference, Guidebook, p. 38-41.

Wells, E.H., 1919, Oil and gas possibilities of the Puertecito district, Socorro and Valencia Counties, New Mexico: New Mexico School of Mines, Mineral Resources Survey, Bulletin 3, 47 p.

Wilkinson, W.H., Jr., 1976, Geology of the Tres Montosa-Cat Mountain Area, Socorro County, New Mexico [M.S. thesis]: Socorro, New Mexico Institute of Mining and Technology, 158 p.

Willard, M.E., 1959, Tertiary stratigraphy of northern Catron County, New Mexico: New Mexico Geological Society, 10th Field Conference, Guidebook, p. 92-99.

Wilpolt, R.H., MacAlpin, A.J., Bates, R.L., and others, 1946, Geologic map and stratigraphic section of Paleozoic rocks of Joyita Hills, Los Pinos Mountains, and northern Chupadera Mesa, Valencia, Torrence, and Socorro counties, New Mexico: U.S. Geological Survey Oil and Gas Investigations Map 61.

Winchester, D.E., 1920, Geology of Alamosa Creek Valley, Socorro County, New Mexico with special reference to the occurrence of oil and gas: U.S. Geological Survey Bulletin 716-A.

Woodward, L.A., 1976, Laramide deformation of Rocky Mountain foreland: geometry and mechanics; in Tectonics and Mineral Resources of Southwestern North America: New Mexico Geological Society Special Publication No. 6, p. 11-17.

Young, R.G., 1960, Dakota group of the Colorado Plateau: American Association of Petroleum Geologists Bulletin, v. 44, no. 2, p. 156-194.

Appendix I
Definitions of descriptive sedimentary terms used in text

Crossbedding:
- high-angle: greater than 15 degrees from bedding plane
- low-angle: less than 15 degrees from bedding plane

- large-scale: greater than 10 feet (3.1 m)
- small-scale: less than 10 feet (3.1 m)

Bedding thicknesses (Ingram, 1954):

very thick bedded:	greater than 1 m (3 ft)
thick bedded:	30 to 100 cm
medium bedded:	10 to 30 cm
thin bedded:	3 to 10 cm
very thin bedded:	1 to 3 cm

Appendix II

Petrography of the sedimentary formations

A. Definitions and code

D : average apparent grain size in mm. (visual est.) using chart of Folk (1968)

Sort: sorting (visual est.)
- vw - very well
- w - well
- m - moderate
- p - poor
- vp - very poor

Ro : roundness (visual est.) using chart of Krumbein and Sloss (1963)
- Ø.1 - angular
- Ø.9 - well rounded

%G : percent framework grains of total rock (visual est.) using chart of Terry and Chilingar (1955)

N : number of grains counted upon which the following percentages are based.

Qm : monocrystalline quartz

Qp : polycrystalline quartz

K : potassium feldspar

Pl : plagioclase

Ch : chert

Li : lithic fragments

T : dominant lithology of lithic fragments
- qz - quartzite
- ms - mudstone
- cs - claystone
- ps - phyllosilicate balls
- sc - quartz-mica schist
- fp - felted plagioclase laths
- ig - igneous

Mi : micas

O : other, including opaques and nonopaques, fossils, and organic debris

Appendix IIA continued

%M : percent matrix material of total rock (visual est.) using chart of Terry and Chilingar (1955)

%C : percent cement material of total rock (visual est.) using chart of Terry and Chilingar (1955)

Min : dominant cement mineralogy
- bc - blocky calcite
- pc - patchy calcite
- s - syntaxial silica
- ph - phyllosilicate
- li - limonite
- ch - chert

%P : percent porosity of total rock (visual est.) using chart of Terry and Chilingar (1955)

tr: trace amount (less than 2%)

Thin-section abbreviations correspond to map symbols.

B. Petrographic data

	Trc-1	Trc-2	Kd-1	Kd-2	Kth-1	Kth-2
D	0.05,0.6	0.13	0.3	0.5,3.2	0.18	0.15
Sort	w,vp	vp	w	m,vp	w	w
Ro	0.3,0.3	0.1	0.7	0.7,0.5	0.3	0.5
%G	25	35	80	82	73	78
N	500	496	500	500	505	450
Qm	10	10	85	50	64	77
Qp	tr	8	5	tr	tr	tr
K	2	40	tr	5	17	12
Pl	tr	tr	tr	tr	tr	tr
Ch	tr	7	3	40	tr	tr
Li	85	24	3	tr	16	7
T	ms	cs	qz	qz	ps	ms
Mi	tr	4	2	tr	tr	tr
O	tr	5	2	4	tr	4
%M	30	50	2	tr	tr	15
%C	45	15	15	15	25	7
Min	bc	pc	s	s	ph	li
%P	tr	3	3	3	2	tr

Notes: Trc-1 is a bimodal, clastic limestone.
Kd-2 is a sandy conglomerate.

Appendix IIB continued

	Kg-1	Kg-2	Kmv-1	Kmv-2	Kmv-3	Tb-1	Tb-2
D	0.18	0.18	0.24	0.20	0.25	0.3	0.35
Sort	w	m	w	m	m	p	w
Ro	0.3	0.3	0.3	0.3	0.3	0.5	0.1
%G	60	35	65	58	50	50	75
N	500	502	496	440	613	476	500
Qm	80	25	87	32	61	68	50
Qp	tr	tr	tr	tr	tr	2	tr
K	14	8	tr	45	12	14	5
Pl	4	7	tr	5	2	3	3
Ch	2	4	11	2	9	10	35
Li	tr	tr	tr	14	13	3	5
T	-	sc	ms	ps	ps	cs	fp
Mi	tr	3	tr	2	tr	tr	tr
O	tr	48	tr	tr	tr	tr	tr
%M	28	15	10	5	7	10	2
%C	10	50	20	35	40	40	20
Min	pc	bc	ch	pc	bc	pc	pc
%P	2	tr	5	2	tr	tr	3

Notes: Kg-2 is a fossiliferous, sandy limestone.
Tb-1 is a nodular sandstone.

Appendix IIB continued

	Tb-3	Ts-1	Ts-2
D	1.0	0.45,1.25	0.6
Sort	m	vp,vp	vp
Ro	0.3	0.1,0.1	0.3
%G	53	70	75
N	532	502	402
Qm	50	tr	tr
Qp	2	tr	tr
K	18	tr	tr
Pl	5	30	20
Ch	13	tr	tr
Li	10	40	70
T	ms	ig	ig
Mi	tr	tr	tr
O	tr	30	10
%M	7	tr	10
%C	40	27	tr
Min	pc	li	li
%P	25	tr	15

Note: Ts-1 is a bimodal rock.

This thesis is accepted on behalf of the faculty of the Institute by the following committee:

Date 7/3/79

SEDIMENTARY ROCKS

HOLOCENE

Qal	Alluvium — *Sands and gravels associated with present drainage*
Qsb	Quaternary slump blocks — *Mostly of Dakota Sandstone*
Qag	Alluvium — *Sand and gravel covering broad, flat areas and dissected by present drainage; often grades into Qal*
Qco	Colluvium

PLIO-PLEISTOCENE

Qtg	Terrace gravels — *Gravel and sand deposited on terraces of present drainage system*
Qpg	Pediment gravels — *Gravel and sand forming a thin covering over broad flat areas between 6740 and 6920 foot elevations*
Qpm	Piedmont gravels — *Gravels with clasts largely of La Jara Peak Basaltic Andesite, A-L Peak Tuff, and Hells Mesa Tuff at elevations greater than 7000 feet*

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MIOCENE

Tp	Popotosa Formation — *White to buff conglomerate, contains clasts of mostly La Jara Peak Basaltic Andesite, A-L Peak Tuff, and Hells Mesa Tuff; interbedded with La Jara Peak Basaltic Andesite at base*

OLIGOCENE

Tlp, Talp, Talg, Tlp	La Jara Peak Basaltic Andesite — *Massive to highly vesicular aphanitic basaltic-andesite flows and associated channel sandstones and conglomerates; interbedded with Popotosa Formation in upper part and with A-L Peak Tuff in lower part*
	A-L Peak Tuff — *PINNACLES MEMBER: Purple and gray crystal-poor quartz-bearing ash-flow tuff containing abundant flattened pumice with sandy texture. GRAY–MASSIVE MEMBER: Gray and orange crystal-poor quartz-bearing ash-flow tuff; pumice is less abundant and exhibits less flattening than in Pinnacles member*

UNCONFORMITY

Thm	Hells Mesa Tuff — *Pink, crystal-rich, quartz-rich ash-flow tuff*

UNCONFORMITY

Tst2, Ts, Tst1	Spears Formation — *Purple to gray volcaniclastic sandstones, conglomerates, mudflow deposits, and siltstones; plagioclase-bearing basaltic-andesite flows present as small discontinuous outcrops in upper part. Tst2: Pink lithic- and biotite-rich ash-flow tuff. Tst 1: Gray crystal-poor ash-flow tuff*

GRADATIONAL, INTERBEDDED CONTACT

EOCENE

Tb	Baca Formation — *Red, pink, yellow, and orange arkosic and conglomeratic sandstones; red siltstones and shale; locally contains abundant silicified wood*

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UPPER CRETACEOUS

Kmv	Mesaverde Formation — *Yellow, brown, and green sandstones; gray and black shales and siltstones; coal; abundant ironstone concretions*
Kg	Gallego Sandstone — *Brownish-yellow to brown, highly bioturbated sandstone; contains* Lopha sannionis *near top*
Kmd, Kmt, Kma	Mancos Shale *Kmd* – D-Cross Tongue – *Dark-gray weathering to olive-gray, fossiliferous, concretionary shale; contains* Scaphites whitfieldi *and* Prionocyclus novamexicanum. *Kmt* – Tres Hermanos Sandstone Member – *Interbedded yellowish-brown thin-bedded sandstone and dark-gray organic-rich shale; contains* Selwynoceras mexicanum *at base and* Coilopoceras colleti *at top.* *Kma* – Alamito Well Tongue – *Dark-gray to medium-gray silty fossiliferous concretionary shale; contains* Pycnodonte newberryi, Sciponoceras gracile, *and* Mammites depressus.
Kd	Dakota Sandstone — *Grayish- to brownish-yellow, iron-stained, cross-bedded quartzose sandstone and conglomeratic sandstone*

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TRIASSIC

Ћc	Chinle Formation — *Red, pink, gray, and green shales and siltstones; yellow arkosic channel sandstones*

IGNEOUS ROCKS

Tid, Tis	Intrusives — *Basaltic and basaltic-andesite intrusions, most altered to green color. Tid – Dikes; Tis – Sills*

Formation contact; dashed where uncertain, dotted where covered

Fault; bar and ball on downthrown side; dashed where uncertain; short dashes where inferred; dotted where covered

Anticlinal axis showing plunge direction — *Synclinal axis showing plunge direction*

13 *Strike and dip of beds* — *Horizontal beds* — 14 *Dip of dikes and faults*

Ephemeral streams — *Road; dashed where unimproved* — *Mine adit*

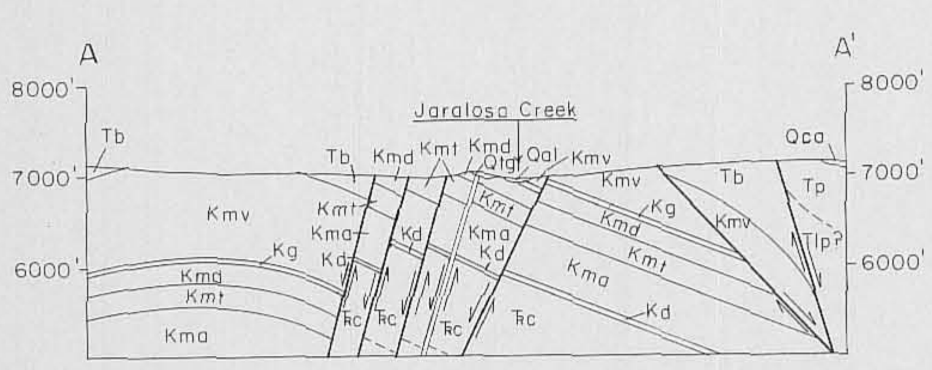
E *Bulldozer excavation* — ×6789 *Point elevation* — *Building* — *Drill hole*

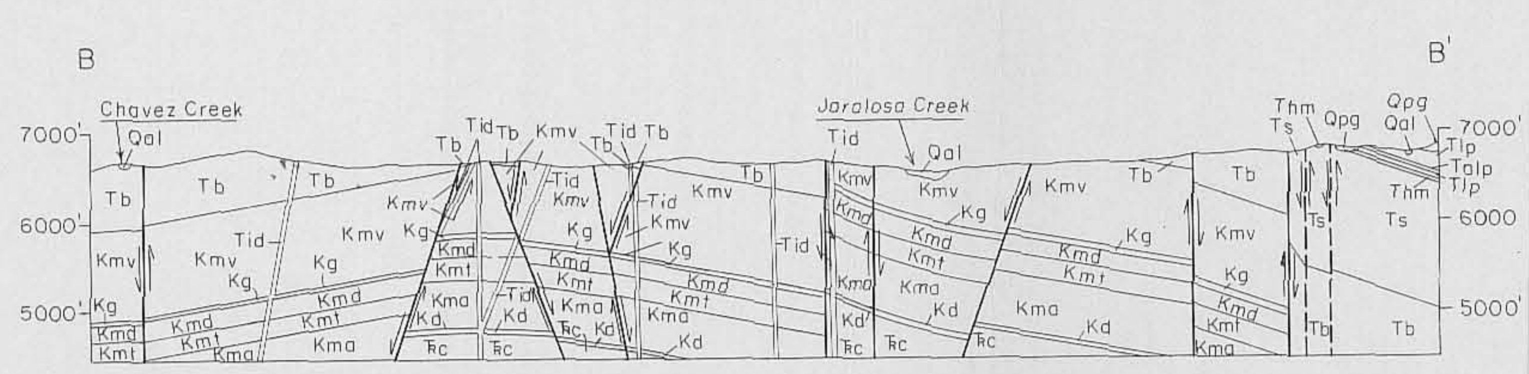
APPROXIMATE MEAN DECLINATION, 1975

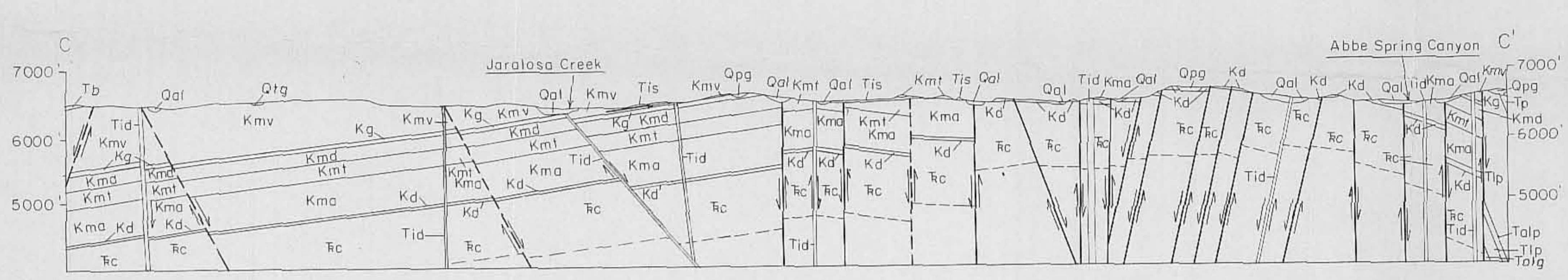
Base map key

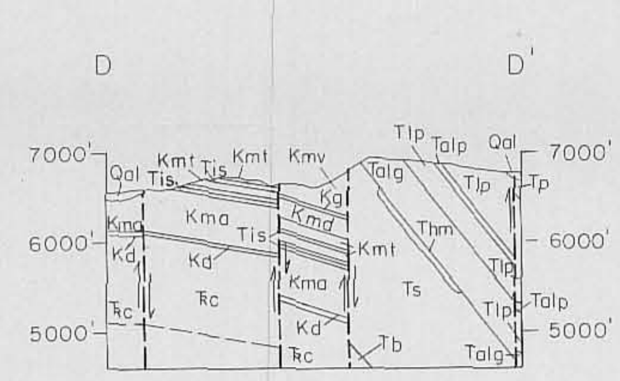
INDEX MAP OF NEW MEXICO

Contour interval: 20ft. except in northeast corner, 40ft.









GEOLOGY OF THE CORKSCREW CANYON-ABBE SPRING AREA, SOCORRO COUNTY, NEW MEXICO

by David L. Mayerson

1979

SCALE 1:24,000

1 mi — 0 — 1 mi

1 km — 0 — 1 km