



Report of Investigations No. 135

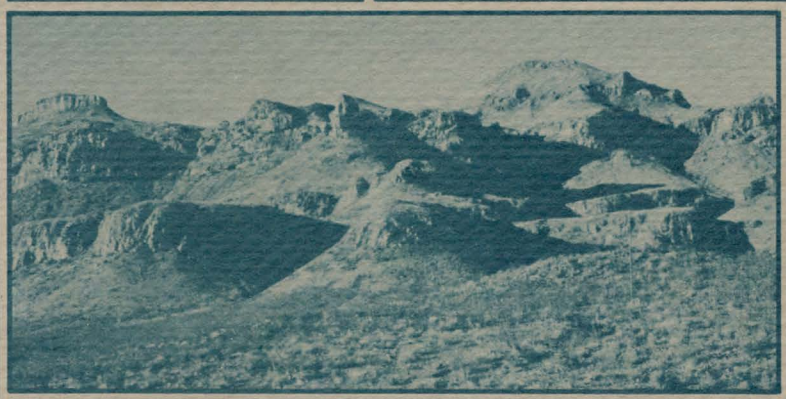
OLIGOCENE VOLCANISM AND MULTIPLE CALDERA FORMATION IN THE CHINATI MOUNTAINS, PRESIDIO COUNTY, TEXAS

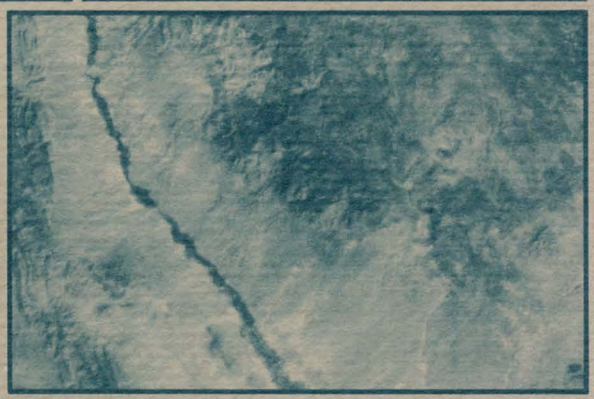
Joseph C. Cepeda
Christopher D. Henry











1983

BUREAU OF ECONOMIC GEOLOGY
W. L. Fisher, Director
The University of Texas at Austin
Austin, Texas 78712



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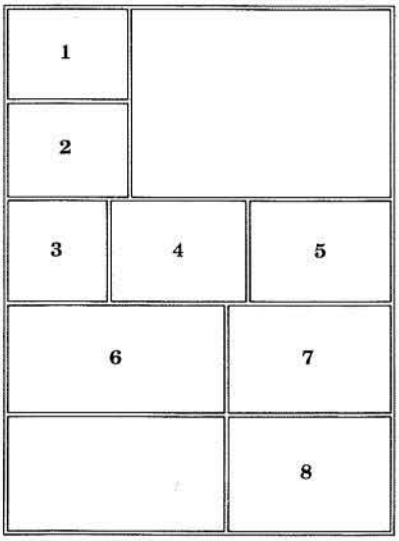
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1. Numerous flows of upper trachyte composing main mass of Chinati Peak.
2. Flow-banded, rhyolitic dome along Oso Creek south of Aguja Peak.
3. Southern caldera boundary with flat-lying lava flows of the Chinati Mountains Group within the caldera in the background and outward-tilted Permian sedimentary rocks in the caldera wall in the foreground.
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ABSTRACT

The Chinati Mountains caldera, which lies in Trans-Pecos Texas in the southern Basin and Range Province, was formed by eruption of the Mitchell Mesa Rhyolite. The caldera (30 km by 20 km) is part of a larger Oligocene volcanic province that includes Trans-Pecos Texas and extends to the Sierra Madre Occidental of western Mexico.

Volcanism in the Chinati Mountains area began several million years before formation of the Chinati Mountains caldera. Rocks of the Morita Ranch Formation, Infiernito caldera, and Shely Group ring the caldera on the south, east, and north. After its collapse, the caldera was filled by rhyolitic to trachytic lava flows and an ash-flow tuff of the Chinati Mountains Group. These include, from oldest to youngest, the lower trachyte, middle trachyte, lower rhyolite, upper trachyte, and upper rhyolite (ash-flow tuff). The Chinati Mountains Group was then intruded by the West Chinati Stock, the resurgent dome of the caldera. Three cycles of rhyolitic to trachytic magmatism, all derived from a zoned magma chamber, are represented by (1) Mitchell Mesa Rhyolite to lower and middle trachytes, (2) lower rhyolite to upper trachyte, and (3) upper rhyolite to West Chinati Stock. Dominant caldera collapse followed eruption of the Mitchell Mesa Rhyolite, but collapse is also associated with rhyolitic eruptions in the second and third cycles. The entire sequence erupted between 32 and 33 mya.

The Chinati Mountains area is the site of one major, inactive silver mine and numerous prospects for silver, lead, zinc, copper, molybdenum, uranium, and fluorite. The Shafter silver district produced 31 million ounces of silver from Permian dolomitic limestones just south of the southern boundary of the caldera. Major prospects are associated with a quartz-monzonite porphyry intrusion (copper-molybdenum) just west of Shafter and with the West Chinati Stock (silver, lead, zinc, copper, and fluorite). All mineralization is probably genetically related to the caldera.

Keywords: ash-flow tuffs, calderas, igneous rocks, K/Ar, mineral deposits, Oligocene, Presidio County, Texas, Trans-Pecos, volcanism.

INTRODUCTION

The Chinati Mountains area is a major Oligocene volcanic center in Trans-Pecos Texas. Available field data, chemical and petrographic analyses, and K-Ar isotopic ages show that the eruption of the Mitchell Mesa Rhyolite 32 mya produced the Chinati Mountains caldera. Subsequently this caldera was filled by more than 1,000 m of rhyolitic and trachytic lava flows and rhyolitic ash-flow tuffs. The Mitchell Mesa Rhyolite is the most extensive ash-flow sheet in Trans-Pecos Texas and is a major unit for correlation in the southwestern part of the Trans-Pecos volcanic field. This report describes the stratigraphy of the Chinati Mountains Group and the geologic history of the caldera and discusses several volcanic sequences that are older than the Chinati Mountains Group.

Regional Setting

The Chinati Mountains lie at the western edge of the Trans-Pecos volcanic field, which was active during middle Tertiary time. They are part of a larger volcanic province that includes rocks of similar age in Mexico and the western United States (fig. 1) (McDowell and Clabaugh, 1979). This volcanic province consists of magmas generated in middle Tertiary time along or above a subduction zone under the western margin of North America.

The Trans-Pecos volcanic field consists of several, mostly rhyolitic, igneous centers of which the Chinati Mountains caldera is the largest identified to date (figs. 2, 3). Other such centers include the Davis Mountains, where a rapid outpouring of igneous material ensued 39 to 35 mya (Parker and McDowell, 1979; Parker, 1972), but only the Buckhorn caldera has been identified (Parker, 1982). Volcanic centers are also located in the Chisos Mountains - Pine Canyon caldera (Maxwell and others, 1967; Ogley, 1979), Paisano Pass (Parker, 1976), Eagle Mountains (Underwood, 1962), Quitman Mountains (McAnulty, 1976), Van Horn Mountains (Henry and Price, in press), Bofecillos Mountains (McKnight, 1968), and the Sierra Rica in Chihuahua, Mexico (Chuchla, 1981). The Cienega Mountains southeast of Shafter (fig. 3) probably represent another, minor source area (Hardisty, 1982).

Areas between volcanic centers contain broad aprons of volcaniclastic sediments that have been reworked from pyroclastic material derived from adjacent centers. A few widespread lavas and ash-flow sheets, predominant among which is the Mitchell Mesa Rhyolite, crop out beyond the immediate vent locality and are interbedded with the volcaniclastic sediments.

The western part of the Chinati Mountains caldera was truncated by Basin and Range faulting, exposing a northwest-southeast cross section of two collapse zones and one major intrusion within a caldera approximately 25 km in diameter. The Shafter mining district is located on the heavily fractured southern edge of the caldera (fig. 3).

Access

The map area is in the western part of Presidio County, Texas. The Chinati Mountains are entirely on private property; permission to travel on ranch roads or to visit geological features must be obtained from the landowners. All-weather paved roads flank the area on two sides. All roads leading from highways to any ranch headquarters within the map area can be traversed by car, but many other ranch roads require a four-wheel-drive vehicle or motorcycle.

History

The Chinati Mountains is one of the least known ranges in Trans-Pecos Texas. Throughout the settlement of the region, the mountains have been bypassed by people and events. Shipman (1926, p. 26) wrote,

> Milton Favor, the first cattle king of the Big Bend got his original herd of cattle together in San Pablo, Chihuahua, Mexico, by trading peloncillas, sweet potatoes, and candles to the indians for cattle in 1849. He moved to the American side of the Rio Grande in the early fifties . . . .
>
> Favor had three ranches near Shafter. Cibolo, the headquarters ranch, was located at the southeast end of the Chinati Mountains . . . Each ranch had a peach orchard and at Cibolo Favor built a still and dispensed peach brandy to all travelers. Cowboys and freighters on the Chihuahua Trail beat a great highway to Cibolo. The owner was noted far and wide for his hospitality.

Indications of metalliferous deposits were discovered at the southeastern end of the mountains by John Spencer, an early rancher, or by his Mexican laborers about 1880 (Ross, 1943). Spencer went into business with a group of army officers at Fort Davis, among them Captain William R. Shafter, for whom the town of Shafter was named. They formed the Cibolo Creek Mill and Mining Company, which was later consolidated under the name of Presidio Mining Company. Production of silver and lead ore began in 1883 and continued until 1942 despite a brief shutdown between 1930 and 1934. Total production was nearly 31 million ounces of silver and about 8 million pounds of lead. The gross value of the output of the Presidio Mine for 1883 through 1940 exceeded $18 million (Ross, 1943).

The Burney Mine area in the West Chinati Stock was first prospected in 1890 (Dennis, 1947). The first production was 1.5 tons of silver chloride ore and 2.5 tons of lead ore in 1935 (McAnulty, 1972). An additional 5.5 tons of fluorspar ore were shipped in 1942 (Dennis, 1947). McAnulty (1972) noted that 5.5 tons of lead-zinc-silver ore were shipped in 1942; no shipments were made thereafter.

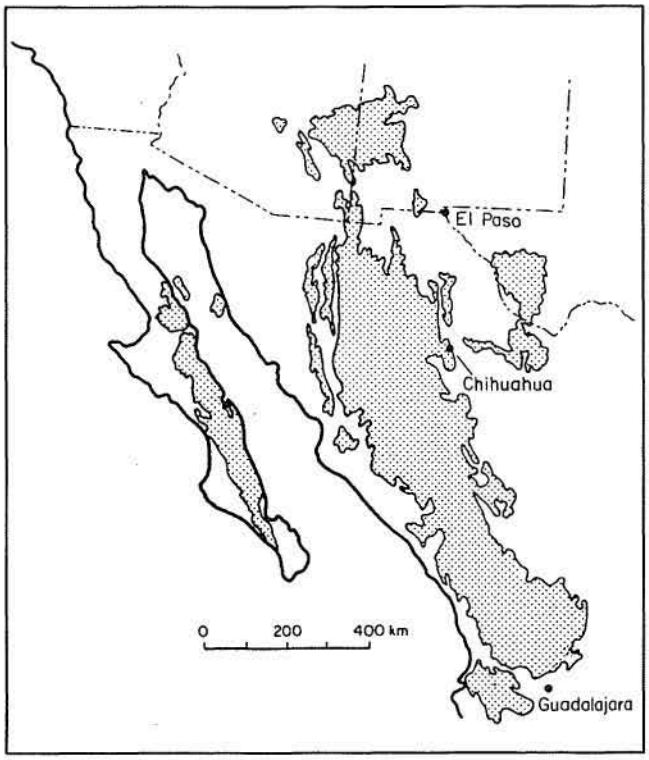


Figure 1. *Approximate distribution of Tertiary volcanic rocks (stippled) in Mexico and adjacent regions of West Texas, New Mexico, and Arizona (from Swanson and others, 1978).*

Previous Work

Figure 4 shows the location of previous geologic studies in and around the Chinati Mountains. Emory's (1857) boundary survey provided the basis for the first geologic map of the region (Hall, 1857). Early investigations of the geology from near our study area focused on the sedimentary Paleozoic and Mesozoic rocks south and east of the map area. These contributions were summarized by Rix (1953), who mapped the 15-minute Chinati Peak Quadrangle. Since that time, Dietrich (1965), who mapped the Presidio area to the south, Maxwell and Dietrich (1970), and Duex and Henry (1981) have attempted to clarify the stratigraphy of the volcanic sequence mapped by Rix and that of the adjacent regions.

Dietrich (1965) described a sequence of basaltic and trachytic flows interbedded with tuffs and conglomerates, which he named the Morita Ranch Formation. This formation includes units that Rix (1953) labeled T1 through T4. Members of the Morita Ranch Formation form the prominent hills south of U.S. Highway 67 near Shafter.

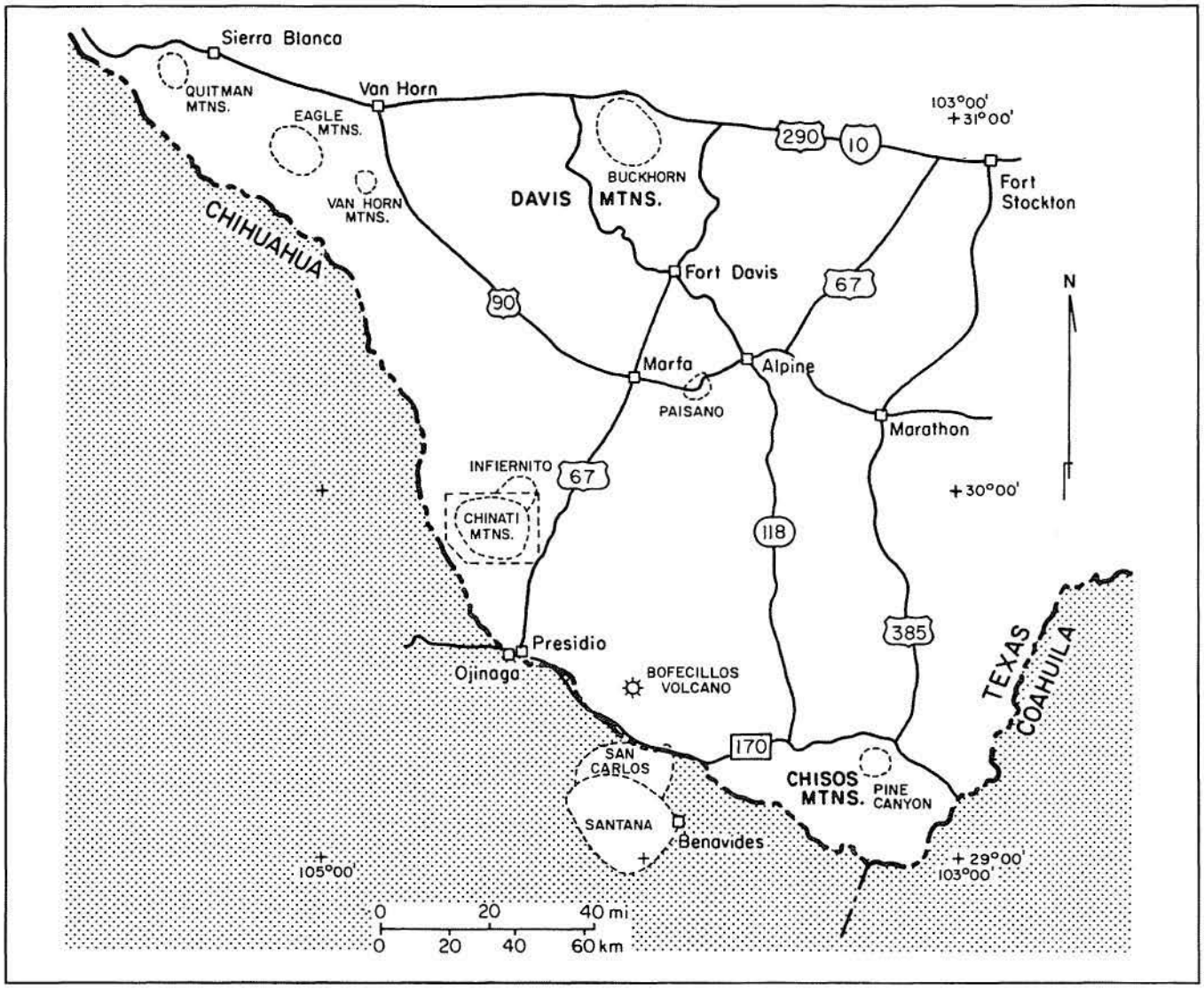


Figure 2. *Known calderas and the Bofecillos volcano of Trans-Pecos Texas and adjacent Mexico. Area of rectangle is depicted in the accompanying geologic map (pl.1).*

Groat (1970) described the depositional facies of upper Tertiary sediments in Presidio Bolson, which forms the western boundary of the map area. The nearest exposed igneous rocks west of the bolson crop out in the Cerros Prietos and Sierra de la Parra in Chihuahua, Mexico. These rocks were mapped and described by Heiken (1966) and Gries (1970). Heiken (1966) described a series of tuffs, one of which he called Brite Ignimbrite (Mitchell Mesa Rhyolite), as being underlain by conglomerates and laharic breccias and overlain by a series of andesitic flows, which he called the Vallecitos Formation.

Amsbury (1958) mapped the Pinto Canyon area north of Chinati Peak, most of the West Chinati intrusion, and the area west of West Chinati Peak. As a result of his field work, he expanded Rix's Chinati Volcanic Series to include 460 m of plagioclase trachyte, tuff, and basal conglomerate. Amsbury also noted that the reddish-brown phase of the Petan Trachyte, the youngest unit of the Vieja Group described by DeFord (1958), might be correlative with Rix's unit T7, which is petrographically similar.

Ore deposits and hydrothermal alteration in the Chinati Mountains were discussed by McAnulty (1972), who presented a brief summary of mining activities in and around the West Chinati intrusion. Duex and Henry (1981), in a report dealing primarily with the Infiernito caldera, also described ore deposits of the Chinati Mountains.

Field Methods

Most of the mapping within the Chinati Mountains caldera was completed during August 1973 and June and July 1974. Additional mapping in adjacent areas

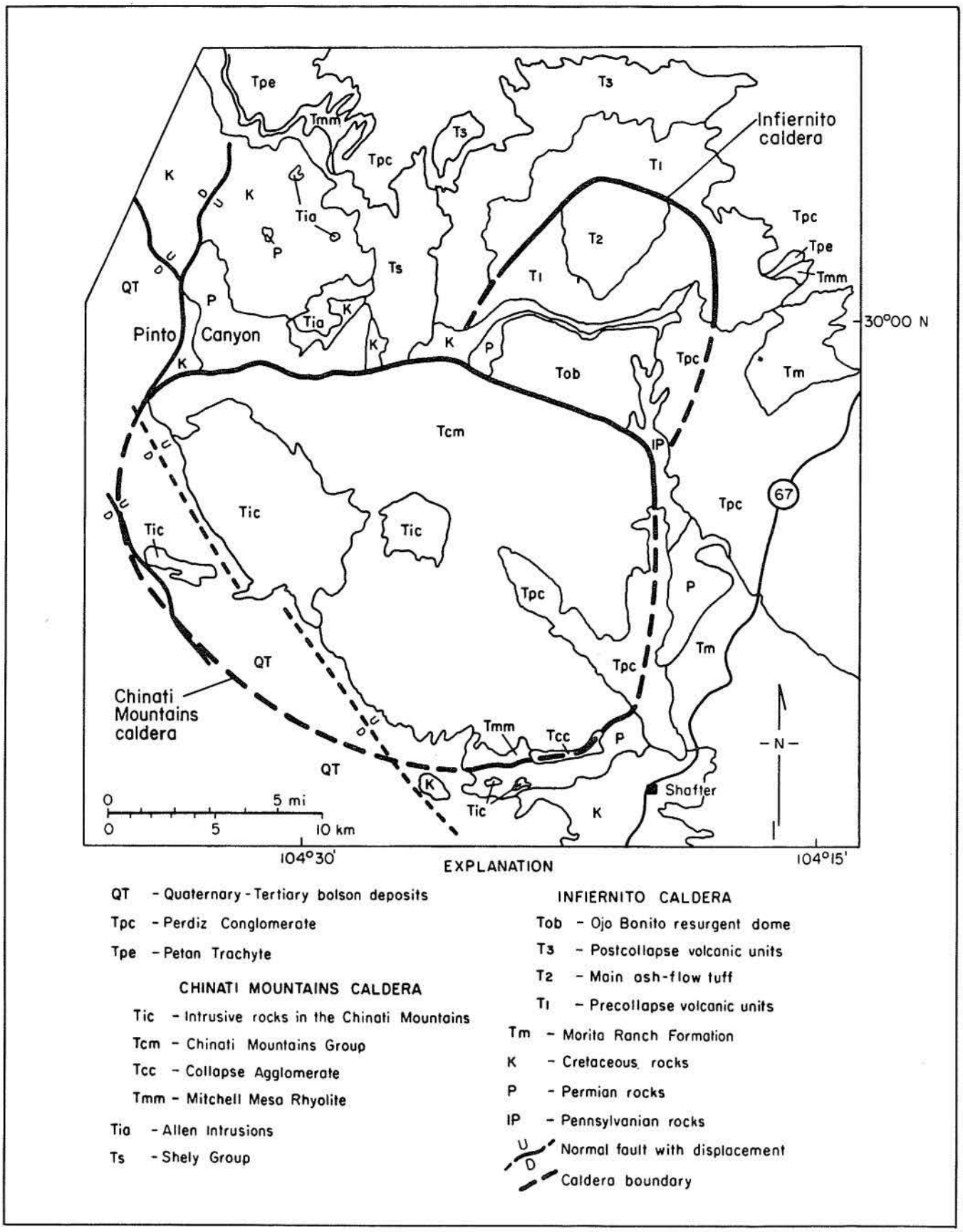


Figure 3. *Generalized geologic map of the Chinati Mountains area (modified from Barnes, 1979, and pl. 1). Location shown in figure 2.*

was done from 1979 through 1982. Geologic contacts were drawn on aerial photographs at a scale of 1:21,000. The compiled map was transferred to U.S. Geological Survey 7.5-minute topographic maps at a scale of 1:24,000 for publication. Stratigraphic sections were measured using a Brunton compass as a hand level.

Nomenclature

Terminology and usage in this report follow standard practice of the U.S. Geological Survey. Volcanic rock names have been applied according to Cook (1965) and the chemical classification scheme of Irvine and Baragar (1971). Plutonic rocks were named according to the IUGS classification (Streckeisen, 1976). Rock colors were determined by comparison with the Geological Society of America Rock Color Chart.

Because individual areas were mapped by different geologists at different times, a hodgepodge of names has developed for the various volcanic units in the Chinati Mountains area. Sequences of volcanic rocks in the individual areas have been termed formations, groups, or series. Many of these groupings incorporate genetically unrelated volcanic rocks and obscure rather than elucidate the relationships. In this report, we attempt to lessen the confusion, but, because our primary purpose is to describe the Chinati Mountains caldera, the attempt is incomplete.

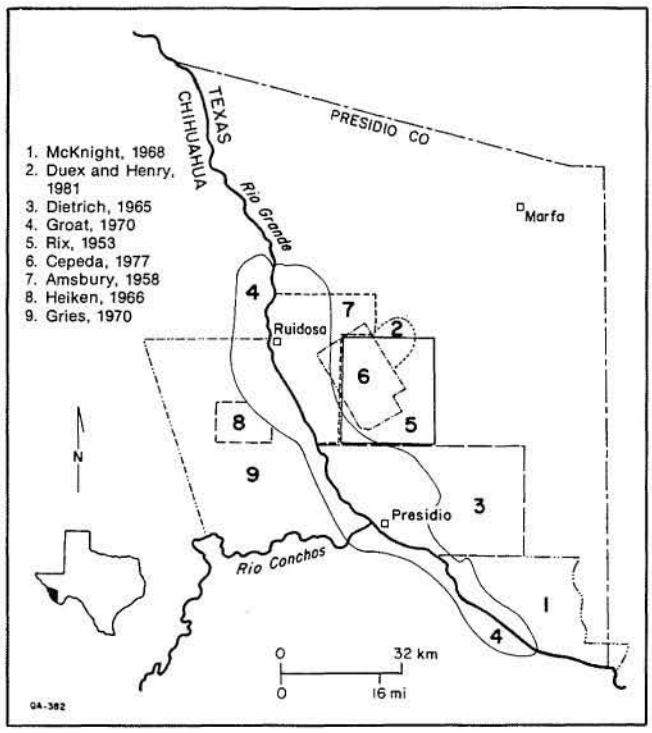


Figure 4. *Index map showing selected previous study areas in Presidio County, Texas, and adjacent Chihuahua, Mexico.*

STRATIGRAPHY

Pre-Tertiary Rocks

The volcanic rocks in the Chinati Mountains area rest on a sequence of upper Paleozoic and Cretaceous sedimentary rocks. Because our primary interest was the volcanic geology of the area, we used the established terminology of Rix (1953) and Amsbury (1958) and relied primarily on their maps of the pre-Cenozoic rocks. However, their subdivision of the Permian and Cretaceous rocks is different and not entirely compatible with each other, especially along the northern border of the caldera where their map areas adjoin. We have modified the previous mapping where necessary and where our field work allowed.

The oldest rocks in the area belong to the Cieneguita Formation of Late Pennsylvanian (Cisco, Canyon, Strawn equivalent) age (Rix, 1953). This unit is more than 600 m thick and consists of quartz-pebble conglomerate and medium- to coarse-grained sandstone with a few lenses of massive limestone (Rix, 1953). It crops out along the eastern border of the caldera.

Overlying the Cieneguita Formation is a Permian sequence divided into the following formations in the Shafter and Infiernito areas: Alta, Cibolo, Ross Mine, and Mina Grande, in order from oldest to youngest. The Alta Formation is by far the thickest unit, consisting of more than 1,700 m of unfossiliferous shale and sandstone. The other three Permian units consist dominantly of limestone and dolomitic limestone with subordinate sand, silt, and shale (Rix, 1953). In the Pinto Canyon area, the Pinto Canyon Formation overlies the Alta Formation and is equivalent to the Cibolo and Ross Mine Formations. The Pinto Canyon Formation is also a predominantly carbonate unit containing minor amounts of chert and clay.

Cretaceous rocks form part of the exposed boundary along the northern margin of the caldera and overlie Permian rocks near the boundary near Shafter. These units are part of the Comanche Series along the northern margin and include, in ascending order, the Yucca Formation, Bluff Formation, and Cox Formation (Amsbury, 1958). The Yucca Formation consists of a basal limestone conglomerate grading upward into red limestone, sandstone, and shale (Amsbury, 1958). The Bluff Formation consists of a very light gray, thick-bedded limestone with interbedded sandstone. The Cox Formation consists of resistant quartzose sandstone, alternating with marl, light-gray limestone, and shale (Amsbury, 1958). In the Shafter area and throughout the area mapped by Rix, the equivalent section includes the Presidio and Shafter Formations. The contact between the two formations falls within the Bluff Formation elsewhere.

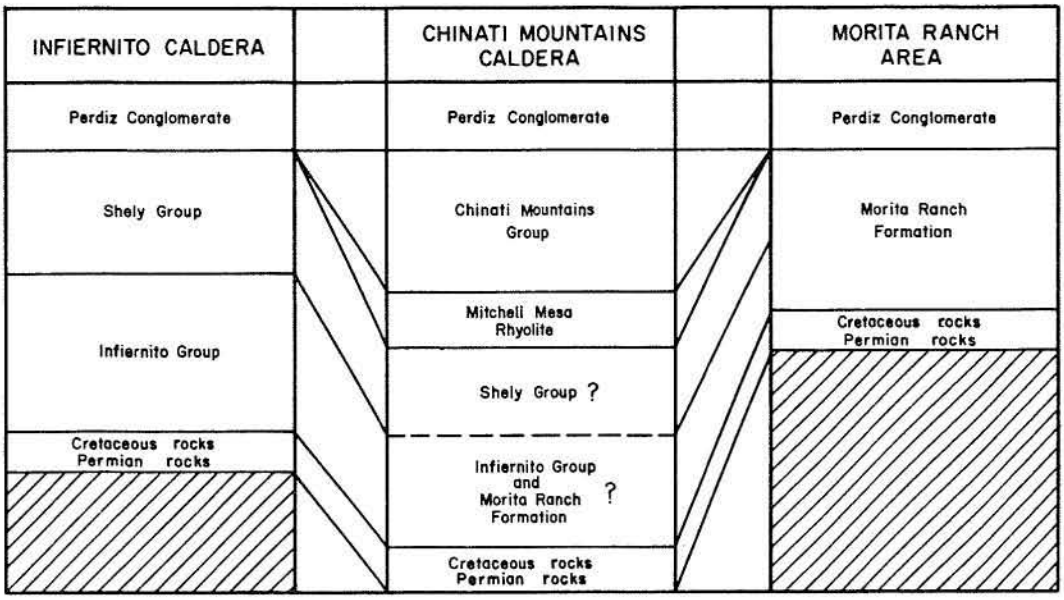


Figure 5. *Known and suspected stratigraphic relations among Tertiary rocks near the Chinati Mountains caldera.*

Tertiary Volcanic Rocks

Several distinct sequences of volcanic rocks overlie, interfinger, or are juxtaposed by faulting in the Chinati Mountains area (fig. 5). Relative age relationships among adjacent strata have generally been established, but some work remains to be done in determining the relative ages of volcanic strata on opposite sides of the Chinati Mountains caldera.

The two oldest sequences exposed on the periphery of the caldera are the Morita Ranch Formation and the rocks of the Infiernito caldera. Both lie on pre-Tertiary rocks and are at least partly overlain by the Tertiary Perdiz Conglomerate. They are both truncated by the Chinati Mountains caldera ring-fracture zone (Duex and Henry, 1981).

Morita Ranch Formation

The Morita Ranch Formation crops out along the eastern and southeastern edge of the Chinati Mountains caldera and extends about 15 km eastward to the Cienega Mountains area. Initially Rix (1953) included the rhyolitic tuffs and basalt and trachyte lava flows in the Buck Hill Volcanic Series. Later, Dietrich (1965) renamed the bulk of this sequence the Morita Ranch Formation, after a locality south of Shafter. Generalized stratigraphy of the Morita Ranch Formation is shown in figure 6.

Several of the units have only local distribution. All the ash-flow tuffs and lava flows below the T4 ash-flow tuff pinch out to the north. The Morita Ranch Formation near the Infiernito caldera consists only of T4 ash-flow tuff overlying tuffaceous sediments and conglomerates. The T1 rhyolites and basalt are particularly restricted; they crop out only within about 5 km of Shafter. Tm3b ash-flow tuff pinches out northeast of Shafter but is extensive to the south and east (Dietrich, 1965; Hardisty, 1982). T4 ash-flow tuff is the most widespread unit of the Morita Ranch Formation and is probably correlative with an ash-flow tuff (Ts6) in the Shely Group to the northwest. T4 ash-flow tuff pinches out north of Shafter; the relative ages of T4 and Tm3b ash-flow tuff and Tm3a lava flows have not been established. K-Ar ages indicate that Tm4 is actually older than Tm3 (McDowell and Henry, unpublished data; see also section on Geologic History).

The Morita Ranch Formation in the Cienega Mountains may have accumulated within a caldera (Hardisty, 1982). Some of the units (T1 and T2) attain large thicknesses, as shown by outcrop and drill-hole data. Also, some rhyolitic lava flows and tuffaceous rocks are clearly of local origin. However, a ring-fracture zone of this postulated caldera has not been observed, possibly because younger gravels cover outcrops of the Morita Ranch Formation east of the Chinati Mountains caldera.

Duex and Henry (1981) noted that the relative stratigraphic positions of the lower part (T1 rhyolitic rocks) of the Morita Ranch Formation and rocks of the Infiernito caldera and the presence of biotite in both rock sequences suggest a possible correlation between the two. Both are older than the Chinati Mountains caldera. A part of the upper Morita Ranch Formation may be contemporaneous with rocks of the Shely Group; for example, Amsbury (1957) and Maxwell and Dietrich (1970) correlated an ash-flow tuff in the Morita Ranch Formation (within T4 of Rix [1953]) and Ts6 ash-flow tuff of the Shely Group. Clearly, T4 ash-flow tuff is younger than rocks of the Infiernito caldera because distinctive clasts from the Infiernito occur in conglomerates that underlie T4. In summary, T1 of the Morita

Ranch Formation may be correlative with rocks of the Infiernito caldera, and T4 ash-flow tuff may be correlative with Ts6 ash-flow tuff of the Shely Group. Other rock units of the Morita Ranch Formation are local and have no known equivalents beyond the Shafter area.

Infiernito caldera sequence

Duex and Henry (1981) recognized a well-defined caldera north and east of the Chinati Mountains; this caldera is older than, and has been truncated by, the Chinati Mountains caldera. Duex and Henry (1981) grouped rocks of the Infiernito caldera into four genetic sequences (fig. 7): precollapse volcanic strata (T1), ash-flow tuffs of the main eruptive event (T2), postcollapse volcanic units (T3), and the Ojo Bonito resurgent dome (Tob) (fig. 3). The volcanic part of this caldera sequence includes rhyolitic and trachytic lava flows; rhyolitic ash-flow tuff, the eruption of which formed the Infiernito caldera; and postcollapse mud-flow breccia and rhyolitic to intermediate lava flows. In the western part of the Infiernito caldera area, rhyolitic to trachytic lava flows (T3) underlie plagioclase trachyte lava flows (Ts2) of the Shely Group, which are thought to be part of the Infiernito volcanic sequence (Duex and Henry, 1981).

The Ojo Bonito intrusion is the resurgent dome of the Infiernito caldera; the intrusion has tilted the first two volcanic units (T1 and T2) and the Cretaceous, Permian, and Pennsylvanian sedimentary rocks radially outward from the intrusion. The intrusion was first mapped by Rix (1953) and named for a spring on its southern boundary. Rix described the intrusion as a biotite-hornblende alkali granophyre; Duex and Henry (1981) showed that it is a quartz monzonite containing 5 to 10 percent quartz and more alkali feldspar than plagioclase.

Shely Group

The Shely Group crops out in the Pinto Canyon area north of the Chinati caldera. Amsbury (1958) described eight volcanic units in the Shely Group, which comprises a complex series of tuffaceous sediment, trachyte, rhyolite, rhyolitic ash-flow tuff, and conglomerate (fig. 7). The non-sedimentary units of the Shely Group are all locally derived and, with the exception of Ts6 ash-flow tuff, are restricted to a north-trending ridge (Shely Rim) that extends from the northern edge of the Chinati Mountains caldera. Closely related to these units in space and time are a series of intrusions (Allen Complex of Amsbury, 1958) composed of rhyolite porphyries, perlites, glassy breccia, and vitrophyre.

Duex and Henry (1981, p. 10) noted that "the transition from intermediate lava flows (plagioclase trachyte of Ts2) to dominantly rhyolitic ash-flow tuffs and tuffaceous sediments (Ts3 to Ts8) represents a marked change in style of eruption and chemical composition." The trachyte is part of a continuous sequence of rhyolitic to intermediate lava flows that compose T3 of the Infiernito sequence. Therefore, Duex and Henry assigned the plagioclase trachyte to the Infiernito caldera sequence. The pyroclastic rocks of the Shely Group may have been derived from the Chinati Mountains caldera but were not derived from the Infiernito caldera. This conclusion is based on the observation that the upper part of the Shely Group thickens and has more flows toward the Chinati Mountains caldera than toward the Infiernito caldera. Also, stratigraphic relationships and preliminary K-Ar ages show that the upper part of the Shely Group is younger than the Infiernito caldera (Henry, unpublished data; see also section on Geologic History). Nevertheless, the Shely Group may have filled a moat developed along the western edge of the Infiernito caldera. Amsbury (1957) thought that unit Ts5 (tuff and conglomerate) was deposited as an alluvial fan in a structurally and topographically low area from northern and western sources. The presence of clasts from the resurgent dome of the Infiernito caldera indicates an eastern source as well. The Shely Group units are truncated by the ring-fracture zone of the younger Chinati Mountains caldera. The southern continuation of the Shely Group is probably buried beneath the volcanic rocks that fill the Chinati Mountains caldera.

We have tentatively assigned a small, isolated outcrop of volcanic rocks along the northern edge of the Chinati Mountains caldera to the Shely Group. The outcrop consists of rhyolitic breccia overlain by a thin, densely welded ash-flow tuff. The volcanic rocks appear to overlie Cretaceous sedimentary rocks. Both rock types dip to the south into the Chinati Mountains caldera; the dip probably resulted from slump into the caldera during or after collapse. The volcanic rocks are overlain by flat-lying lava flows of the upper trachyte of the Chinati Mountains Group. Although these rocks are not petrographically identical to any known sequences and cannot be distinguished by their stratigraphic position, we have assigned them to the Shely Group because of their proximity to other Shely outcrops.

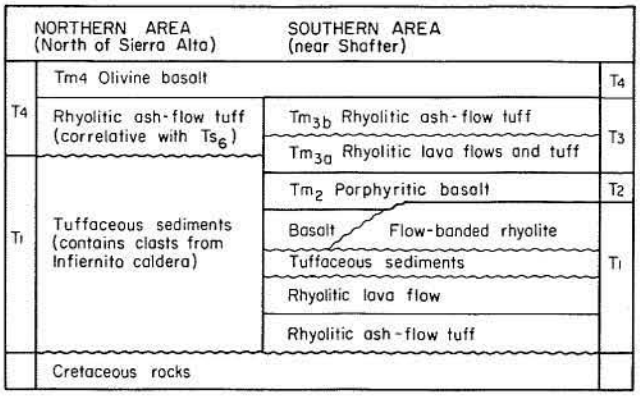


Figure 6. *Generalized stratigraphy of the Morita Ranch Formation near Shafter, derived from Rix (1953), Dietrich (1965), Hardisty (1982), Duex and Henry (1981), and this study. Designations T1, T2, T3, and T4 are from Rix; designations Tm2, Tm3a, Tm3b, and Tm4 are from Dietrich.*

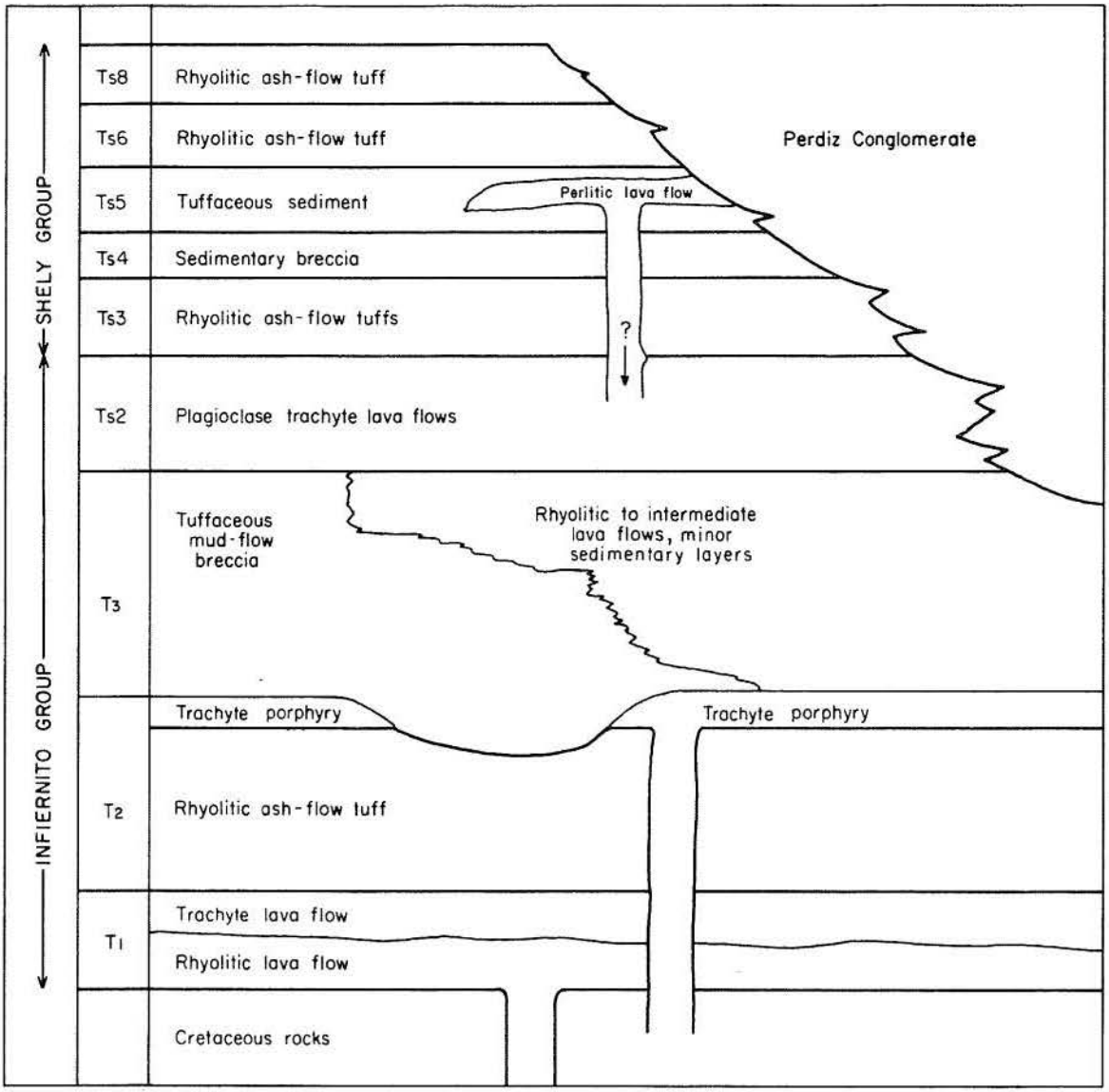


Figure 7. *Generalized stratigraphy of the Shely and Infiernito Groups in the Infiernito caldera area, northern Chinati Mountains, derived from Amsbury (1958), Duex and Henry (1981), and this study. Ts designations are from Amsbury (1958). Ts2 (plagioclase trachyte) was originally mapped as part of the Shely Group by Amsbury; Duex and Henry showed that it is part of the rhyolitic to intermediate flows of T3 of the Infiernito rocks.*

Mitchell Mesa Rhyolite

The Mitchell Mesa Rhyolite is the volumetrically largest and most widespread ash-flow tuff of Trans-Pecos Texas. Its eruption produced the Chinati Mountains caldera. Within the Chinati Mountains caldera, it is exposed only in a small area (~3 km^2) along the southern caldera margin west of Shafter. Here it ponded against Permian rocks exposed in the caldera wall; also, it underlies the lower rhyolite and probably the middle trachyte, the lowest exposed parts of the Chinati Mountains Group in that area. Outside the caldera, it crops out in a broad area completely around, and as far as 70 km from, the caldera (fig. 8). Maximum thicknesses are about 110 m in Mexico approximately 20 km southwest of the caldera (Heiken, 1966) and about 98 m in several areas 10 to 15 km north of the caldera (Burt, 1970). These are the outcrops closest to the caldera. The Mitchell Mesa Rhyolite thins progressively away from the caldera toward the north, east, and southeast. Toward the west and southwest, it probably ponded against the high ridges of the Chihuahua tectonic belt in Mexico but spilled over these ridges or filled valleys in the ridges in a few places.

The Mitchell Mesa Rhyolite is a multiflow tuff. Outcrops near the caldera are made up of more than six individual flows but are all one cooling unit. Only one flow is recognizable in more distal outcrops, but intense vapor-phase crystallization in all outcrops may have disguised some flow boundaries. Burt (1970) recognized

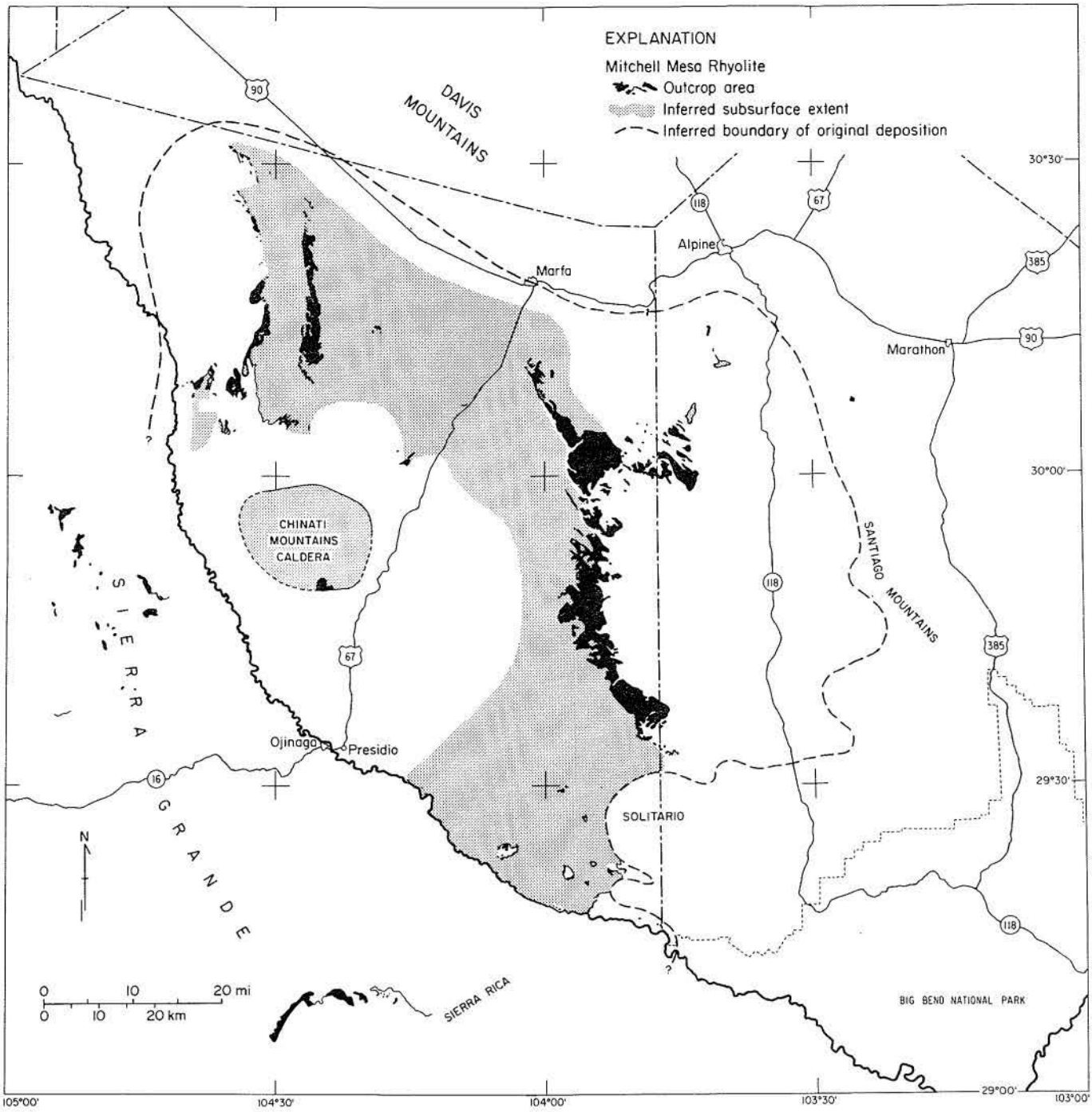


Figure 8. *Distribution of the Mitchell Mesa Rhyolite, Texas and Mexico.*

two cooling units in the thick sections of Mitchell Mesa just north of the Chinati Mountains caldera. However, where we examined these outcrops, the Mitchell Mesa is a single cooling unit, densely welded in the lower half and progressively less welded in the top half. The upper 10 m is non-welded and contains large (up to 20 cm long) pumice blocks that are only slightly flattened. This upper, non-welded zone is preserved where overlying resistant rocks were deposited soon after the Mitchell Mesa formed. For example, north of the Chinati Mountains, the Mitchell Mesa is capped by the Petan Trachyte, a series of resistant lava flows that protected the non-welded tuff from erosion. Poorly welded or unwelded tuff is preserved only locally in distal parts of the ash flow. Within the caldera, the Mitchell Mesa is moderately to densely welded and has undergone extensive vapor-phase crystallization. Because the Mitchell Mesa is poorly exposed and the base is buried, we could not determine the total number of flows present or their thickness.

We found no glass preserved in the Mitchell Mesa Rhyolite. Most of the rock is devitrified; axiolitic devitrification is common and spherulitic crystallization is rare. Vapor-phase crystallization of tridymite and alkali feldspar, and rarely cristobalite, is abundant and has eradicated the original shard texture. A basal, non-welded ash zone escaped devitrification, but ground water has altered the preserved glass to smectite with minor amounts of calcite.

Phenocrysts in the Mitchell Mesa Rhyolite consist primarily of alkali feldspar and a lesser amount of quartz. Densely welded samples contain as much as 25 percent by volume of the two minerals. Quartz is generally bipyramidal and sanidine is iridescent, a characteristic feature of the Mitchell Mesa best developed in the most welded parts. Mafic phenocrysts consist of oxyhornblende, augite, magnetite, and rare biotite. Accessory minerals include zircon, sphene, and apatite.

Burt (1970) summarized available chemical analyses of the Mitchell Mesa Rhyolite. Mitchell Mesa is a high-silica rhyolite, relatively rich in alkalis and low in aluminum, calcium, and magnesium. Two of four analyses listed by Burt are peralkaline with normative acmite. Because all samples are devitrified, the rock may have lost some alkalis and be truly peralkaline. However, sodic pyroxenes or amphiboles have not been found.

The Mitchell Mesa Rhyolite initially may have covered at least 16,000 km^2 in Texas and Mexico. The inferred original distribution shown in figure 8 is based on known outcrops, on reasonably estimated subsurface distribution, and on identification of Oligocene topographic barriers, such as the Solitario Uplift, that would have limited its flow. Because thickness data and outcrops are scattered, it is impossible to estimate precisely the original volume of the Mitchell Mesa. Assuming an average thickness of 50 m over 16,000 km^2 yields a volume of 800 km^3; this value is probably within a factor of 2 of the true volume.

Before recent field work by Henry, the Mitchell Mesa Rhyolite had not been found within the Chinati Mountains caldera. This lack was considered by some to indicate that the Chinati Mountains were not the source. However, even if the rhyolite had not been found within the caldera, it was clear that the Chinati Mountains caldera had to be the source of the Mitchell Mesa Rhyolite because of the nearly symmetrical distribution of the Mitchell Mesa around the caldera and the progressive thinning of the rhyolite away from the caldera. Additional evidence is that the 32.3 m.y. old Mitchell Mesa (McDowell, 1979) is indistinguishable in age from the rocks of the Chinati Mountains Group, which have ages ranging from 32.0 to 32.8 m.y. (table 2). Cepeda (1977) showed that derivation of the Mitchell Mesa Rhyolite from the magma chamber that produced the Chinati Mountains Group is consistent with the major-element composition of the various rocks. Cameron and others (1982) reached a similar conclusion on the basis of trace-element and rare-earth data. Henry and McDowell (1982) showed that the Mitchell Mesa Rhyolite is five times larger, volumetrically, than other ash-flow tuffs of Trans-Pecos Texas. The Chinati Mountains caldera (approximately 400 km^2) is the largest caldera in Texas; its size is consistent with eruption of the Mitchell Mesa Rhyolite from the caldera. If the Mitchell Mesa Rhyolite did not erupt from the caldera, another, equally voluminous, ash-flow tuff must have. No such tuff exists in Texas, however.

The total volume of Mitchell Mesa Rhyolite within the Chinati Mountains caldera is unknown. Only a small outcrop exists, but it probably underlies the entire caldera; however, its thickness within the caldera is also unknown. The area of the caldera is 300 to 400 km^2. If the Mitchell Mesa is 500 m to 1 km thick, the caldera fill has a volume of 150 to 400 km^3. The Mitchell Mesa is comparable to the Bishop Tuff in the Long Valley caldera, California, in that the Bishop Tuff is not exposed within its caldera. As much as 1,000 m of Bishop Tuff may be buried beneath younger deposits within the Long Valley caldera (Bailey and others, 1976).

Table 1. Nomenclature of the Chinati Mountains Group.

Rix (1953)	Amsbury (1957)	This report
T-8 soda rhyolite	Tc6 rhyolite	upper rhyolite
T-7 olivine augite andesine trachyte	Tc5 trachyte	upper trachyte
T-6 soda rhyolite	Tc4 rhyolite	lower rhyolite
		Cieneguita dome and flows
		non-porphyritic domes and flows
T-5 quartz trachyte	Tc3 trachyte	middle trachyte
	Tc2 trachyte	lower trachyte
	Tc1 conglomerate	collapse agglomerate (can occur at any stratigraphic level within this sequence)

Chinati Mountains Group

In the central part of the Chinati Mountains, a sequence of rhyolitic and trachytic lava flows and tuffs more than 1,000 m thick is exposed in the Chinati Mountains caldera. These flows and tuffs are included within the Chinati Mountains Volcanic Series delineated by Rix (1953) and Amsbury (1957). These units, as well as the Cieneguita dome and flows, and the non-porphyritic domes and flows, compose the Chinati Mountains Group as referred to in this report (table 1).

Outcrops of these units encompass approximately 450 km^2. Nomenclature used by Rix (1953), by Amsbury (1958), and in this report is shown in table 1. Rocks of the Chinati Mountains Group compose the volcanic units exposed in the map area whose vents probably also lie in the map area.

The Chinati Mountains Group, as originally described by Rix (1953), includes the following units, from oldest to youngest: T-5 (middle trachyte of this report), T-6 (lower rhyolite), T-7 (upper trachyte), and T-8 (upper rhyolite). Amsbury (1958) added two units below T-5: a basal conglomerate (here interpreted as a collapse agglomerate) and a lower trachyte.

The flows of the Chinati Group have been intruded by granitic dikes and sills. The largest intrusive mass is the West Chinati Stock exposed in and west of San Antonio Canyon (pl. 1). This intrusion has tilted the overlying lower rhyolite and upper trachyte toward the east 10° to 20°. Intrusive rock similar to the West Chinati Stock crops out in the bottom of stream valleys on the eastern side of the mountains where intrusive rock has domed the overlying lower and upper rhyolite. Dikes, ranging in thickness from a few centimeters to several meters, intrude the West Chinati Stock and the volcanic units.

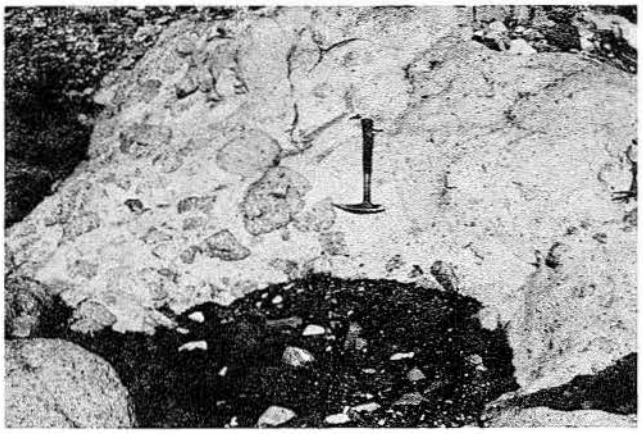


Figure 9. *Volcanic agglomerate (mud-flow deposit) composed of angular to moderately rounded fragments, up to 1.5 m in diameter, consisting dominantly of lower rhyolite, in matrix of finer fragments. Outcrop west of Shafter between cliffs of lower rhyolite (north) and Permian Ross Mine and Mina Grande Formations (south).*

Collapse agglomerate

An agglomerate of coarse, angular to moderately rounded boulders of both pre-caldera volcanic and sedimentary rocks and some rocks of the Chinati Mountains Group crops out along the caldera wall in several places. We interpret these deposits as being shed off the caldera wall after collapse, usually emplaced as mud flows or landslide deposits. The two most prominent outcrop areas are along the southern boundary west of Shafter and along the northern boundary in Pinto Canyon. In several other areas, the outcrops are too thin to show on the geologic map.

The agglomerate consists of both conglomerate and chaotic mud-flow (?) breccia, composed of clasts up to 1.5 m in diameter (fig. 9). Rock types present as clasts are dominantly pre-caldera volcanic and sedimentary rocks and are locally derived from adjacent outcrops. In some places, clasts of rocks that do not crop out nearby indicate the former occurrence of those rocks. West of Shafter, for example, the agglomerate commonly contains numerous clasts derived from the Morita Ranch Formation. From this we infer that the Morita Ranch Formation overlay the Cretaceous and Permian sedimentary rocks there at the time of caldera collapse. At one location along the northwestern caldera boundary in Pinto Canyon, a large (>100 m long) block of Cretaceous Bluff Formation limestone is both overlain and underlain by coarse agglomerate. The block must have slumped off the caldera wall and been incorporated as an exceptionally large clast in the agglomerate. Thus it is comparable to Lipman's (1976) megabreccias. Similar large blocks of Permian and Cretaceous rocks occur along the southern boundary west of Shafter.

The collapse agglomerate does not have a unique position within the stratigraphic column. In most places it overlies pre-caldera rocks in the wall along a steeply dipping contact. However, in the northwest part of the caldera, agglomerate overlies lower trachyte and is in turn overlain by middle trachyte. Also, some clasts are derived from rocks of the Chinati Mountains Group; for example, west of Shafter, presumed mud-flow deposits contain boulders of lower rhyolite. The lower rhyolite must have ponded against the caldera wall, while simultaneously shedding its own debris against the wall. These stratigraphic relations show that the agglomerate was deposited during a long timespan, not just immediately after collapse.

Lower trachyte

The lower trachyte has the smallest outcrop area of any unit of the Chinati Mountains Group. It crops out in low hills west of West Chinati Peak and is best exposed in the stream beds north of the Morales Ranch house. The base of this unit is not exposed because the outcrops are bounded on the west by the West Chinati fault. At the fault the lowest exposed unit is a pale-red (10 R 6/2) conglomerate that underlies the lowest exposed trachyte. This conglomerate lens is at least 10 m thick. The weathered flows are moderate brown (5 YR 4/4) to pale yellowish brown (10 YR 6/2). The color of weathered outcrops and the horizontal jointing, which imparts a platy character to the unit, distinguishes it from the middle trachyte, which overlies it to the east. Estimated exposed thickness of the lower trachyte is 200 m.

Textures range from glomeroporphyritic with a fine-grained groundmass to seriate with large, interlocking feldspar grains and interstitial granophyre. The lower trachyte contains approximately 15 to 20 percent anorthoclase and plagioclase phenocrysts up to 2 mm

long. The groundmass consists of feldspar and opaques. Chlorite and iddingsite aggregates probably represent the altered remnants of clinopyroxene and olivine phenocrysts. Chlorite, epidote, and minor quantities of sericite are alteration products of the groundmass. Apatite is a prominent accessory mineral.

Middle trachyte

The middle trachyte is exposed in a small strip behind the Cibolo Ranch headquarters and in scattered patches to the north, where it is largely covered by an apron of gravel on the eastern side of the Chinati Mountains. Another strip of trachyte is present on the southwestern side, where it is interbedded with conglomeratic tuff and sandstone. The base is exposed at two places: (1) along the upper reaches of Cibolo Creek, 6 km east of Aguja Peak, where the middle trachyte rests on conglomeratic sandstone of the Cibolo Formation, and (2) in the northern part of the map area, where it is underlain by the collapse agglomerate and the lower trachyte.

The measured thickness of 150 m in the canyon west of the Cibolo Ranch house is typical. Widely distributed outcrops of the middle trachyte suggest that it underlies the lower rhyolite over much of the areal extent of the latter unit. However, in most places, the lower rhyolite is the lowest unit exposed. In San Antonio Canyon, lower rhyolite has been intruded by the West Chinati Stock, and a block of the middle trachyte has been rafted up by the intrusion. The presence of outcrops of the middle trachyte on the periphery of the map area as well as in blocks within the West Chinati Stock suggest that the unit is widespread and has been covered by later flows.

The middle trachyte has undergone considerable, probably deuteric, alteration; therefore, fresh samples are extremely difficult to obtain. The weathered rock is moderate brown (5 YR 4/4) to olive gray (5 Y 3/2); fresh surfaces are dark gray (N3). The unit is cut by joints and fractures, and in stream beds these fractures show rusty brown haloes of iron oxide, producing a crosshatched pattern characteristic of all trachyte units. Irregular vugs up to 3 cm in diameter are common and are completely filled by blue or white chalcedony or lined with euhedral, singly terminated crystals of clear quartz up to 2 cm long. Xenoliths are not abundant in this unit. The most common xenoliths are dark, nonporphyritic rock, probably trachyte. These xenoliths are probably fragments of the middle trachyte from earlier flows.

In thin section, the middle trachyte is characterized by 10 to 15 volume percent of subhedral to euhedral, untwinned, plagioclase phenocrysts. The phenocrysts range from 0.5 to 5 mm in the longest dimension. The plagioclase crystals are typically "worm-eaten" by corrosion and in many the trend of the "wormholes" is parallel to cleavage traces or crystal outlines. Thin (less than 0.1 mm) rims of alkalic feldspar encircle most of the phenocrysts. A few phenocrysts exhibit elongate laths of albite- and Carlsbad-twinned plagioclase in the center, mantled by "worm-eaten," untwinned feldspar. These crystals are also zoned. Michel-Levy determinations indicate plagioclase compositions of about An35. Some flows contain a small percentage of large anorthoclase phenocrysts.

Augite (2V=50°, opt. +) occurs as subhedral to anhedral, equant phenocrysts up to 1 mm in diameter. It shows some alteration to opaque minerals and cloudy, fine-grained material along margins and cleavage traces. Augite constitutes approximately 5 percent of the rock.

Opaque minerals, principally magnetite and ilmenite, constitute approximately 5 percent of the rock; many of the grains are larger than 0.1 mm. Apatite constitutes an abundant accessory mineral as stubby microphenocrysts up to 0.4 mm long by 0.1 mm wide. Aggregates of iddingsite (?) and opaque minerals suggest the former presence of olivine. A fine-grained groundmass of feldspar, opaques, and clinopyroxene constitutes the bulk of the rock.

Interbedded with the trachyte in the southwestern part of the map area is an interval of grayish-orange (10 YR 7/4) conglomeratic tuff and conglomerate with alternating beds of very pale orange (10 YR 8/2), crossbedded, tuffaceous sandstone. The conglomerate is made up of cobbles and boulders of fine-grained, leucocratic, porphyritic rhyolite of an unknown source.

Lower rhyolite

The lower rhyolite is the most widespread unit of the Chinati Mountains Group. With few exceptions, lower rhyolite makes up the most distal outcrops of the Chinati Mountains Group. It ranges in thickness from 20 m at the northern edge of the caldera just south of the caldera boundary to a maximum of 450 m at the mouth of Cañon Tinaja Prieta (pl. 1). Prominent cliffs on the western side of the Chinati Mountains are made up of lower rhyolite overlain by upper trachyte (fig. 10). The western edge of the range coincides with the thickest part of the rhyolite.

The lower rhyolite also forms an extensive plain to the east of the mountains, where its thickness probably ranges from 50 to 100 m. The top of the rhyolite and a 2- to 10-m-thick ash-fall tuff above it form a gently dipping plain upon which erosional remnants of the upper trachyte form low hills. The base of the rhyolite is not exposed in this area except at a point 6 km east of Aguja Peak on the western bank of Cibolo Creek where it overlies the middle trachyte.

The lower rhyolite has a variety of stratigraphic relationships with the units above and below it throughout the map area. At rare exposures of the base, the middle trachyte commonly underlies lower rhyolite. The upper trachyte overlies it, and grayish-green ash-fall tuff lies between the two units. The contact between the two units is typically a slope covered with slump blocks of upper trachyte that slid down over the ash-fall layer. In general, the contact is smooth and flat, suggesting that at the time the upper trachyte erupted, flows of rhyolite had produced a smooth, nearly level plain, having perhaps a slight southeastward dip.

The color of the lower rhyolite ranges from moderate brown (5 YR 3/4) to grayish brown (5 YR 3/2) on weathered surfaces and is brownish gray (5 YR 4/1) on fresh surfaces. Euhedral to anhedral anorthoclase phenocrysts (2V=50°, opt. −) up to 4 mm in diameter make up 10 to 15 percent of the rock. The phenocrysts exhibit outer zones or mantles, delineated by microvesicles and strings of glass and mineral inclusions. A chalky white alteration envelopes many grains, some of which are twinned and some of which are zoned. The rest of the rock consists of a fine-grained groundmass and euhedral to subhedral microphenocrysts of feldspar ranging from 0.2 to 1.0 mm in diameter. The nature and abundance (20 to 30 percent) of microphenocrysts suggest that the lower rhyolite consists of lava flows rather than of ash-flow tuffs.

Quartz phenocrysts make up 4 to 6 percent of the lower rhyolite rock; quartz is also present in the groundmass. Opaque minerals, present in the groundmass, constitute about 2 to 3 percent of the lower rhyolite. Small, ragged grains of biotite and light-green clinopyroxene, both 0.2 to 1 mm in diameter, are found in some of the thin sections of this unit. The ragged edges of the grains suggest that they are xenocrysts. The basal flow west of Chinati Peak contains about 1 percent light-green clinopyroxene (2V=65°, opt. +) as phenocrysts and as inclusions within feldspar phenocrysts. The large and abundant anorthoclase phenocrysts are characteristic of the lower rhyolite throughout the map area.

Xenoliths, which are rare in the lower rhyolite, include fine-grained leucocratic syenite and dark trachyte similar to the middle trachyte. Some fragments of sedimentary rocks are present in the basal flow of the lower rhyolite west of Chinati Peak, which suggests proximity to a vent. The base of this flow also has a zone 1 to 2 m thick exhibiting a laminar flow structure. The rock appears to have been glass that has devitrified and recrystallized, especially along shear planes.

The Cieneguita dome near the eastern edge of the map area and just west of the Cieneguita Ranch headquarters (pl. 1) has been mapped separately but is probably related to the lower rhyolite. The dome is approximately 0.5 km in diameter and, together with associated flows, forms a circle 1.5 km in diameter in map view. On the north side of the dome, at least three porphyritic flows are exposed and dip away from the dome at 15° to 20°. These flows are overlapped by horizontal flows of the lower rhyolite. The great similarity of the lower rhyolite to rocks of the dome suggests that the Cieneguita dome is a dissected dome of lower rhyolite lava. To the south the flows of the dome are approximately horizontal and are overlain by flows of the lower part of the upper trachyte.

The Cieneguita dome and flows contain approximately 10 percent alkali feldspar phenocrysts (2V=10° to 45°, opt. −) in a medium light-gray (N 6) groundmass that weathers light brown (5 YR 6/4). The phenocrysts are 0.5 to 6 mm long and show the same marginal zones as those in the lower rhyolite. The groundmass is about 75 percent feldspar, 20 percent quartz, and 5 percent opaques. Epidote is a prominent accessory mineral in the dome and occurs as rounded grains. Biotite forms ragged grains less than 0.1 mm long. These grains are surrounded by opaque minerals and may represent the alteration product of another mafic mineral. Oxyhornblende prisms, altering to opaque minerals, are also present.



Figure 10. *Massive cliffs of lower rhyolite, overlain by upper trachyte on skyline to right, along western front of Chinati Mountains south of San Antonio Canyon.*

A green to yellowish-green ash-fall tuff containing large, tabular feldspar phenocrysts similar to those in the lower rhyolite overlies the rhyolite. This tuff is not commonly exposed because the overlying trachyte slumps down over it. The tuff and underlying rhyolite form a flat, gently sloping plain upon which the upper trachyte was emplaced.

Upper trachyte

The upper trachyte crops out in two major areas within the caldera: a southern belt, which roughly parallels the caldera boundary; and a northern belt, which makes up the main mass of Chinati Peak and also approximately parallels the caldera wall (fig. 11). In both areas the upper trachyte overlies the lower rhyolite.

Two distinct textural variations characterize the upper trachyte. A fine-grained lower part with small, ragged feldspar phenocrysts set in a medium-gray (N5) groundmass is overlain by a rock with larger (0.3 to 1 cm long) phenocrysts in a grayish-red (5 R 4/2) groundmass. The two units are separated by a thin layer of sandstone. Phenocryst content in the lower part is about 5 to 15 percent compared with about 25 percent in the upper unit. The white phenocrysts on reddish-brown weathered surfaces in the more porphyritic phase impart a speckled appearance to the rock.

Petrographically, the upper trachyte is similar to the middle trachyte. Phenocrysts of "worm-eaten" plagioclase have indistinct twinning. The phenocrysts, up to 5 mm long, are cut by a fine network of fractures

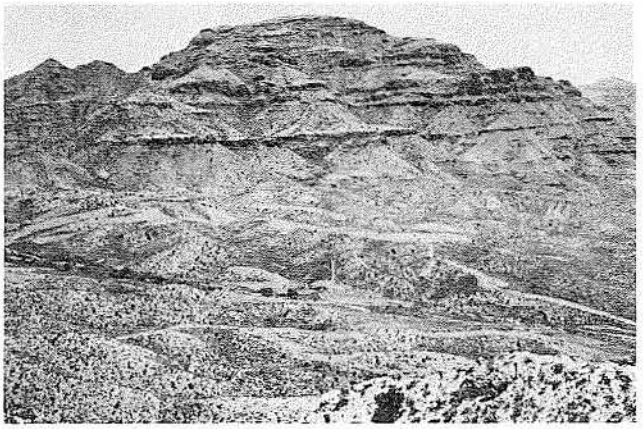


Figure 11. *Numerous flows of upper trachyte composing main mass of Chinati Peak (highest point in middle of photograph). Low hills in foreground are Permian and Cretaceous rocks. Northern caldera boundary is along base of escarpment between rocks of the Chinati Mountains Group and older rocks.*

perpendicular to the long axes of the crystals. Anorthite content ranges from 35 to 45 percent.

Augite, in individual grains up to 1 mm in diameter, forms glomerocrysts more than 2 mm in diameter. The groundmass is a felty to trachytoid mass of feldspar and opaque material with some interstitial clinopyroxene.

Vitrophyres occur at the base of several flows in the lower part of the upper trachyte north of Chinati Peak. The vitrophyres contain approximately 10 percent twinned and untwinned plagioclase phenocrysts up to 2 mm long, and 2 to 3 percent augite phenocrysts. Other phenocryst phases include anorthoclase (2 to 3 percent) and 1 to 2 percent generally equant, opaque grains in a matrix of light-brown glass.

The upper part of the upper trachyte contains 15 to 25 percent albite-, Carlsbad-, and pericline-twinned plagioclase (An30 to An35). These phenocrysts, up to 8 mm in diameter, are set in a matrix of feldspar and opaques. Clinopyroxene phenocrysts, generally less than 0.6 mm in diameter, make up 2 to 3 percent of the mode. Dark patches consisting of a mixture of opaques, chlorite, and calcite probably represent alteration products of primary mafic minerals.

Microphenocrysts of apatite are a prominent accessory in both parts of the trachyte. Amygdules filled with quartz and chalcedony are abundant in both parts near flow tops and constitute 10 to 15 percent of the trachyte by volume.

Several flows constitute the upper trachyte. Eight flows are present in the measured section west of the Wood Ranch house and five in the lower part of the trachyte. Four flows, three of them in the lower unit, make up the upper trachyte section at Lillian Peak. The relation between number of flows, thickness of individual flows, and geographic position is as follows: On the north flank of Chinati Peak, flows of trachyte are characteristically thin and have a generalized stratigraphy in the lower part that consists of 2 to 4 m of black, hard vitrophyre and 5 to 10 m of dense lava, overlain by 2 to 3 m of highly vesicular lava forming the top of the flow. The vesicles are lined with quartz, filled with calcite, or coated or partly filled with yellow and green material that produces no discernible X-ray diffraction peaks. The vesicles are elongate, and the direction of maximum elongation, probably the flow direction, is toward or slightly west of Chinati Peak. To the east and south of Chinati Peak, the number of flows decreases, thickness of individual flows increases, and vitrophyre is absent. This trend is exemplified clearly along a southeast axis from Chinati Peak in the north to Cerro Oroña in the south.

The source of the lower part of the upper trachyte is near Chinati Peak. Two plugs of the upper part of the upper trachyte have been identified and mapped at the head of Cañon Tinaja Prieta. These plugs have vertical flow structures and are feeders to the upper part of the upper trachyte.

The only xenoliths found in the upper trachyte resemble fragments of lower rhyolite. Cobbles of a rhyolitic tuff with opalescent blue feldspar phenocrysts 2 to 4 mm in diameter are present in a sandstone and conglomerate lens. The lens is between two flows in the lower part of the upper trachyte approximately 1 km north of the Cieneguita Ranch house. These cobbles may be fragments of Mitchell Mesa Rhyolite.

In all measured sections, a moderate orange pink (10 R 7/4) to moderate reddish orange (10 R 6/6) tuffaceous sandstone with conglomeratic lenses is present at the boundary between the two parts of the upper trachyte. This sandstone, which is approximately 5 to 10 m thick, represents an interval of sedimentation between the eruption of the lower and upper units of the upper trachyte. No alternation or interdigitation between the upper and lower parts has been observed.

Two flows of basalt (that is, of dark, fine-grained, igneous rock) 30 m thick crop out on the flanks of Aguja Peak, sandwiched between two trachyte flows of the upper part of the upper trachyte. These are the only outcrops of the unit within the map area; this basalt has not been found outside the map area. The rock has a pilotaxitic texture with twinned plagioclase laths and interstitial clinopyroxene and opaques. Plagioclase (An32) phenocrysts as large as 0.6 by 0.1 mm make up about 80 percent of the rock, which suggests that it is not a true basalt. According to the classification system of Irvine and Baragar (1971) and given a normative plagioclase composition of An32 and a color index of 15, the rock is a mugearite. Clinopyroxene phenocrysts up to 0.4 by 0.1 mm constitute 2 to 3 percent of the rock. Elongate and equant aggregates of chlorite probably represent an alteration product of primary mafites. Equant opaques make up 5 to 10 percent of the rock.

The original thickness of upper trachyte throughout the field area is difficult to assess. Major determinants of presently exposed thickness are original topography and subsequent block faulting and erosion. It is possible to determine the original emplacement thickness of the lower part because most outcrops are capped by the

upper part of the unit. The maximum exposed thickness of the lower part, 250 m, occurs southeast of Chinati Peak. The exposed thickness of the upper part varies from 45 m at Cerro Oroña, where it is capped by the upper rhyolite, to 166 m southeast of Chinati Peak in a downfaulted block. At Chinati Peak the upper part forms a thin cap, mostly of vitrophyre, 10 m thick. The thickness of the entire unit varies from 417 m, just east of Chinati Peak, to 275 m at Cerro Oroña, to an estimated 20 to 50 m in its easternmost exposures, where the unit has probably been thinned by erosion.

Amsbury (1957) noted that the upper trachyte might be correlative with the Petan Trachyte, which overlies the Mitchell Mesa Rhyolite in the northern part of Pinto Canyon. The Petan Trachyte is petrographically and chemically (Henry and Cepeda, unpublished data) similar to the upper trachyte. It is possible that flows of upper trachyte actually filled the caldera and spilled out of it to the north to form the Petan Trachyte. However, insufficient evidence exists to document this.

All rocks of the Chinati Group are altered to some extent, and the upper trachyte is the most intensely altered unit within the map area. It has prominent zones of greenish-yellow and yellow in contrast to the brownish-red hues of the lower rhyolite and the grayish green of the upper rhyolite. This color contrast is particularly evident along the cliffs on the western front of the mountains. The feldspar is "worm-eaten" and the groundmass is heavily epidotized.

Non-porphyritic domes and flows

Non-porphyritic domes and flows occupy the low-lying terrain along Oso Creek south of Aguja Peak. The domes, characterized by vertical or nearly vertical flow banding (fig. 12a), delineate an arc, concave to the southwest, from south of Wood Ranch headquarters to east of Cibolo Ranch headquarters (fig. 12b). The non-porphyritic rocks are probably younger than both the lower rhyolite and the upper trachyte because (1) the flows contain clasts of lower rhyolite, (2) the flows overlie porphyritic upper trachyte at a small outcrop in a creek bottom south of the Cieneguita dome, and (3) the flows overlie dark trachyte, probably upper trachyte, still further south of the dome. Thin outcrops of debris-filled upper rhyolite overlie the non-porphyritic flows. Interbedded conglomeratic sandstone and reworked tuff layers within the non-porphyritic flows dip 15° to 20° toward the center of the caldera.

The arcuate trend of the domes probably represents a ring-fracture zone on the boundary of a secondary collapse zone. The dipping tuff layers suggest that collapse was expressed by faulting along the ring fracture and by sagging along a hingeline located outside the dome trend. The amount of collapse depends on the thickness of the upper trachyte within the collapse zone, but may be as much as 200 m, the thickness of the upper trachyte at Aguja Peak.

Textures and contorted flow banding in the domes and flows suggest that the rhyolitic magma was highly viscous. In thin section much of the rock consists of contorted bands of length-fast chalcedony surrounding a mosaic of quartz and feldspar. Accessory minerals include apatite and opaque minerals.

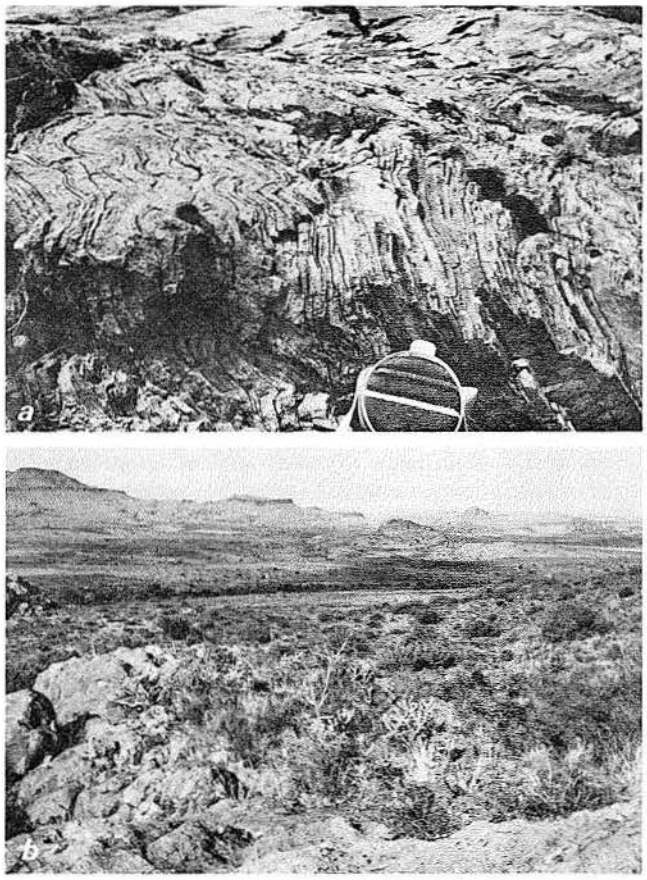


Figure 12. *a. Flow-banded, rhyolitic dome along Oso Creek south of Aguja Peak. b. Domes of non-porphyritic rhyolite along presumed secondary collapse zone.*

Upper rhyolite

The upper rhyolite is a porphyritic gray to green lithic-crystal tuff, the most distinctive rock in the map area. The color of the groundmass and the abundance of small, tabular anorthoclase phenocrysts allow fragments of this unit to be identified readily. Chemical analysis shows that it is a peralkaline, high-silica rhyolite (Cepeda, 1977, 1979; Cameron and others, 1982). It possesses characteristics of both ash-flow tuffs and lavas, as discussed later in this section. With a few exceptions, exposures of this rock are restricted to a small area in the central part of the Chinati Mountains (fig. 13). There the unit is more than 180 m thick and fills a circular depression 4 to 5 km in diameter, which we interpret as a secondary collapse zone within the Chinati Mountains caldera. Thin (30 m), scattered patches of upper rhyolite overlie the non-porphyritic flows in the lowlands to the east; these outcrops are all within a few kilometers of the inferred rim of the

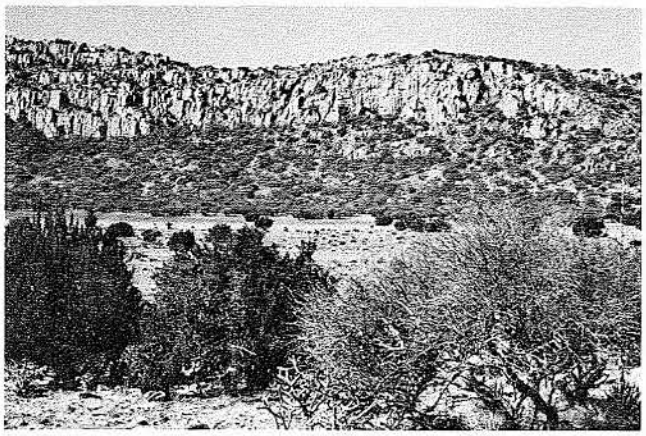


Figure 13. *Distinctly layered cliffs of upper rhyolite within secondary collapse area in central Chinati Mountains, southwest of Wood Ranch. Layers probably represent individual ash flows within upper rhyolite.*

collapse zone. A small exposure of vitric-crystal tuff approximately 15 to 20 m thick appears to rest unconformably on an erosional surface cut into the middle trachyte near the Agua Chano Ranch house. The mineralogy and color of this exposure suggest that it is an erosional remnant of the upper rhyolite.

The thick, central part of the upper rhyolite, which we believe accumulated within its own caldera, contains 15 to 25 percent small (1 to 3 mm) anorthoclase phenocrysts. It also contains about 10 percent resorbed quartz phenocrysts, generally less than 2 mm in diameter, and minor plagioclase phenocrysts. Amphibole prisms up to 2 mm by 0.5 mm make up 1 percent of the rock.

The groundmass, which constitutes the rest of the rock, contains spherulitic aggregates of crystals of dark- to bright-green clinopyroxene, probably aegerine, a mineral common in peralkaline rocks, in a fine-grained mosaic of quartz and alkali feldspar. Some of the aggregates preserve rectangular crystal outlines. Coarse granophyric intergrowths of quartz and alkali feldspar occupy much of the groundmass; the granophyric intergrowths are lens-shaped and may have replaced pumice.

In general, shards within the thick part of the upper rhyolite are not preserved. Recognizable pumice fragments do occur in a few samples along with extremely faint shard outlines. Vague lens-shaped masses, possibly pumice fragments, are common in all outcrops; however, in thin section, these masses are distinguished from the groundmass only by slight differences in texture.

The upper rhyolite contains abundant xenoliths of several types. The most common inclusions are lower rhyolite, dark trachyte, and a leucocratic rock containing spongelike aggregates of dark-green clinopyroxene. Fragments of this leucocratic rock constitute more than 90 percent of the xenoliths in the unit. No outcrops of the parent rock that supplied these xenoliths have been found in the map area. The xenoliths have the same mineralogy as the upper rhyolite and may represent fragments of an intrusive equivalent of the upper rhyolite.

Unlike the thicker outcrops, in which no vitrophyre has been found, the thinner outcrops outside the postulated caldera typically have an irregular lens of greenish-black vitrophyre near the base. Some areas of vitrophyre show laminar flow structure in which shards are unrecognizable. The upper rhyolite above the vitrophyre is densely welded, having axiolitically devitrified, flattened shards, and pumice fragments. Xenoliths are rare. In other respects the rock is petrographically similar to the upper rhyolite within the presumed caldera.

We interpret the overall geologic and petrologic relationships as indicating accumulation of the upper rhyolite mainly in a circular collapse zone produced by its own eruption. Eruption and caldera subsidence were simultaneous. A small part of the rhyolite escaped the caldera to form a thin sheet in canyons cut into the lower rhyolite and on flatlands to the east, where it overlies the non-porphyritic flows.

Evidence of accumulation within a caldera includes (1) the great thickness of upper rhyolite, (2) the dense welding, granophyric crystallization, and destruction of shard texture throughout much of the caldera fill, and (3) the abundance of lithic fragments. Lipman (1975) used similar criteria to characterize near-source and caldera-fill material in the San Juan volcanic field of southern Colorado. Although peralkaline ash-flow tuffs are generally volumetrically smaller than calc-alkaline tuffs (Schmincke, 1974), Lipman's criteria are applicable.

Laminar flow structures similar to those in the upper rhyolite have been found in other peralkaline ash-flow tuffs (Schmincke, 1970). Laminar flow is attributed to complete welding and "reconstitution" by early collapse of low-viscosity, pyroclastic material (Schmincke and Swanson, 1967); the resulting rock has features of a lava flow. Schmincke (1970) suggested that chemical composition, including volatile content, is the main cause of low viscosity and subsequent flow. Features of the upper rhyolite that probably promoted secondary flow are high temperature due to proximity to the source, high volatile, alkali, and iron concentrations, and low aluminum concentrations (Cepeda, 1977; Scarfe, 1977; Henry and Tyner, 1979).

Gonzales rhyolite

The Gonzales rhyolite crops out in two small areas near the southwestern margin of the caldera. The rock is generally similar to the upper rhyolite and may in fact be the same unit. However, the Gonzales rhyolite has been mapped separately because it appears to overlie the middle trachyte and thus seems out of proper stratigraphic position. Either it was faulted down into that position by a part of the West Chinati fault zone or it was deposited in valleys cut into older rocks of the Chinati Mountains Group.

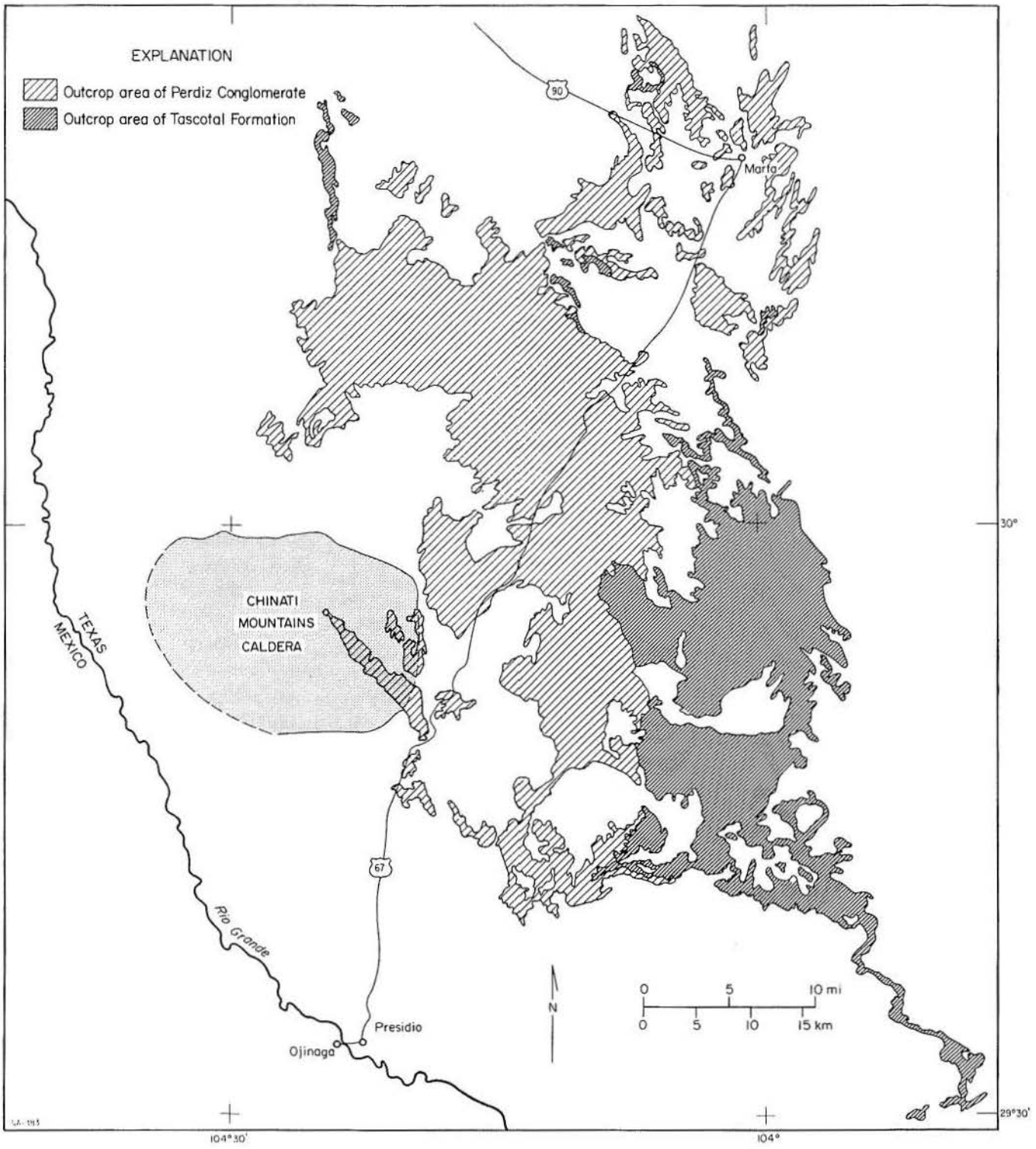


Figure 14. *Distribution of the Perdiz Conglomerate and Tascotal Formation.*

Tascotal Formation

Although the Tascotal Formation does not crop out anywhere in the map area, it is mentioned here because it is an essential part of the Chinati Mountains caldera cycle. The Tascotal Formation consists of tuffaceous sediment derived primarily from the Chinati Mountains (Walton, 1979). It crops out in a north-south-trending belt beginning just south of Marfa, approximately 50 km northeast of the Chinati Mountains, and extending 80 km south to Tascotal Mesa, approximately 40 km southeast of the Chinatis (fig. 14). The Tascotal

overlies the Mitchell Mesa Rhyolite and is in turn overlain by the Perdiz Conglomerate to the south or by the Petan Trachyte to the north.

According to Walton (1979), the major part of the Tascotal Formation was derived from a blanket of tuffaceous material, probably air-fall tuff and nonwelded ash-flow tuff, that erupted from and covered the Chinati Mountains. The tuff was emplaced after eruption of the Mitchell Mesa Rhyolite and after caldera collapse, but its age relative to the Chinati Mountains Group is indeterminate. Walton argued convincingly that the Tascotal Formation was deposited while the Chinati Mountains caldera was still actively producing tuff. Tuffaceous source material is not now exposed in the Chinati Mountains, so either it was stripped off from above the Chinati Mountains Group or it underlies the same rocks in deeper, unexposed parts of the caldera.

The Tascotal Formation has been diagenetically altered by ground water, which dissolved glass and precipitated clinoptilolite, clay, opal, and calcite in its place (Walton, 1979). Undissolved but hydrated glass is commonly preserved at the top of the Tascotal Formation.

Perdiz Conglomerate

The Perdiz Conglomerate overlies the Tascotal Formation in areas distant from the Chinati Mountains (fig. 14). However in the Chinatis, where the Tascotal Formation is absent, the Perdiz overlies a variety of older rocks, including Permian and Cretaceous sedimentary rocks, rocks of the Shely Group, Infiernito caldera, Morita Ranch Formation, and, within the caldera, rocks of the Chinati Mountains Group.

Jordan (1979) showed that the Perdiz was deposited as a series of coalescing alluvial fans composed of coarse debris shed off the Chinati Mountains and adjacent areas. While pyroclastic activity occurred in the Chinatis, the area was blanketed with fine, tuffaceous material. Erosion of this pyroclastic debris provided material for the Tascotal Formation, but erosion generally could not reach the underlying, hard, volcanic rocks. Once pyroclastic activity ceased, erosion cut down to the more resistant volcanic rocks, which were incorporated in the Perdiz Conglomerate. Within and near the Chinati Mountains, the Perdiz Conglomerate contains clasts almost entirely of volcanic rocks, up to 1.5 m in diameter. Most of the clasts are recognizable as being derived from the Chinati Mountains. The clasts become progressively smaller away from the Chinatis, but even as far as 20 km away, clasts average about 20 cm in diameter (Jordan, 1979). The Perdiz contains abundant interbedded tuffaceous sandstone.

We have remapped as Perdiz Conglomerate several areas along the southeastern and eastern margins of the caldera that Rix (1953) mapped as Quaternary deposits. The absolute age of these outcrops is difficult to establish, but the deposits are indurated and unrelated to present-day drainages. Therefore, we conclude that they are Perdiz Conglomerate.

Intrusive rocks of the Chinati Mountains caldera

The West Chinati Stock is the largest and petrographically most variable intrusion in the map area. Zinc, lead, silver, and fluorite mineralization is present in shear zones and joints within the stock. Sills, dikes, and plugs, predominantly granitic (Cepeda, 1977, 1979), intrude both the volcanic and sedimentary rocks of the area and the West Chinati Stock.

Red Hill

The oldest major intrusion probably related to the Chinati Mountains caldera is the quartz-monzonite porphyry at Red Hill west of Shafter, 1 km south of the caldera wall (Price and Henry, 1982). It cuts the Permian Ross Mine and Mina Grande Formations. The intrusion probably predates caldera collapse because fragments of altered porphyry occur in collapse agglomerate within the caldera immediately to the north. Quartz-pyrite veining, sericitic alteration, and supergene weathering obscure the original petrography in exposures of the central intrusion, which contains abundant quartz phenocrysts. Plagioclase, alkali feldspar, and locally biotite are preserved in less quartz-rich peripheral intrusions that were subjected to propylitic alteration characterized by chlorite, epidote/clinozoisite, calcite, sericite, and pyrite. Similar, but less altered, porphyritic dikes and sills occur to the east and southeast.

West Chinati Stock

The West Chinati Stock, the resurgent dome of the Chinati Mountains caldera, is exposed over more than 60 km^2 in the northwestern part of the map area. Rix (1953) originally called the intrusion a laccolith, but mostly discordant contacts show that it is a stock. Its emplacement tilted the lower rhyolite and upper trachyte 10° to 20° toward the east near San Antonio Canyon. The contact between these units is sharp, although small, concordant lenses of intrusive rock occur within the volcanic units nearby. The West Chinati Stock may be younger than the upper rhyolite because granitic sills (described later) intrude between the rhyolite and upper trachyte. Two distinct rock types constitute the intrusive mass: a dark outer zone of plagioclase-rich quartz monzodiorite and a core of leucocratic, porphyritic, hornblende granite. The coarsely crystalline quartz monzodiorite is exposed in San Antonio Canyon, where it is intensely altered with abundant chlorite and epidote; it weathers to spheroidal, granular masses. Here the monzodiorite zone is at least 250 m thick and forms a gently southeastward dipping shell around the core.

On the north flank of West Chinati Peak, which is composed of leucocratic granite, the contact between the granite and the marginal phase dips to the north at a steep angle. The marginal phase is much thinner,

approximately 100 m; it is analogous to a vertical cylindrical envelope around a core of granite.

Roof pendants rafted up by the stock include a large block of interbedded gypsum and limestone from the underlying sedimentary section; above it is a block of dark trachyte similar to the middle trachyte. These rafted blocks suggest that the intrusion is a stock rather than a laccolith. However, the presence of volcanic layers of the lower and middle trachyte dipping toward the stock north of West Chinati Peak suggest that the geometry of the intrusion may be more complex than is apparent at the surface.

The quartz monzodiorite consists of approximately 75 percent equant feldspar phenocrysts with cores of albite-twinned plagioclase (An28 to An30) armored by alkali feldspar. Grains are up to 12 mm in diameter. Quartz forms smaller crystals between the larger feldspar grains and constitutes about 10 percent of the rock. Clinopyroxene (2V=60°, opt. +) grains constitute about 8 percent as equant grains 0.5 to 1 mm in diameter. Hornblende is present in small amounts, as are opaque minerals.

The grain size of the quartz monzodiorite north of West Chinati Peak decreases to less than 5 mm toward the chilled contact with the middle trachyte, and the feldspar is poikilitic with abundant inclusions of equant clinopyroxene grains less than 0.1 mm in diameter. The marginal phase is more alkalic on the north and is probably a quartz monzonite. Amsbury (1957) described the marginal phase on the north flank of West Chinati Peak as a plagioclase-bearing quartz syenite. Interstitial granophyre is present, and a few grains of purple-red to blue-green pleochroic amphibole (probably a sodium-rich amphibole) demonstrate the peralkaline affinity of this rock.

The granitic core of the West Chinati Stock contains 75 percent alkali feldspar grains and 15 percent clear quartz grains. The anhedral feldspar crystals are sericitized, but a core and an irregular rim can be distinguished. Many grains exhibit an albite- and Carlsbad-twinned core surrounded by a sericitized rim of alkali feldspar that is as thick as the core is wide. Ragged grains of light-green to light-brown pleochroic hornblende as large as 1 mm by 0.5 mm make up 3 to 5 percent of the rock. Alteration products of the hornblende include epidote and chlorite. Biotite forms small greenish-brown flakes and constitutes 2 to 3 percent of the rock. Magnetite and ilmenite make up about 4 percent of the granite. Prominent accessory minerals include zircon and euhedral apatite crystals as large as 0.9 mm by 0.15 mm.

The contact between the porphyritic granite and the quartz monzodiorite phase is gradational over an interval of several meters. It can be traced across the Sierra Parda, where it dips to the southeast (pl. 1). Near West Chinati Peak, the contact can be mapped on aerial photographs because of the contrast between the light-colored core and the dark, marginal phase.

To the west, in Presidio Bolson on the down-dropped block of the stock, porphyritic, flow-banded rock with xenoliths and a fine-grained groundmass resembles the marginal rock of the intrusion. Its texture may be due to chilling along the contact or to piercing of the roof of the stock, which allowed some of the magma to rise to the surface as flows. The fracturing of the roof may be related to the shear zones within the marginal rock described by McAnulty (1972). If this rock formed by chilling along the contact, the outcrops define the westward limit of the West Chinati intrusion. Outcrops of porphyritic granite in the bolson are not finer grained than their equivalent in the main body.

Microgranite dikes and plugs intrude the quartz monzodiorite throughout San Antonio Canyon and the Sierra Parda. The dikes range in size from a few centimeters wide and a meter long to several meters wide and hundreds of meters long. The mineralogy is similar to the granite core of the stock. Plugs are also abundant throughout San Antonio Canyon and are as much as several tens of meters in diameter. The contact between the microgranite and quartz monzodiorite commonly contains a floating aggregate of feldspar megacrysts, presumably torn from the coarsely crystalline wall rock by the intrusion. A plug of granophyre several meters wide was also mapped at the head of San Antonio Canyon.

A large, vertical dike 20 to 30 m thick strikes northeast from West Chinati Peak and cuts into flows of the middle trachyte and one flow of the lower rhyolite. It is composed of feldspar (75 percent) and quartz (15 percent). The tabular alkali feldspar phenocrysts range up to 6 mm long. The dike also contains a small amount (3 to 4 percent) of light-green clinopyroxene. Accessory minerals include apatite and zircon.

Plagioclase trachyte intrusions

Plagioclase-phyric, light-olive-gray to dark-gray, trachyte intrusions crop out along the northern caldera boundary. They appear to be associated with the middle trachyte and may represent feeders for it. Distribution of these intrusions along the caldera boundary suggests that the boundary controlled their emplacement.

Dikes and sills

Granitic sills intruded between the upper trachyte and the upper rhyolite and between the lower rhyolite and the upper trachyte. Contact relations, however, are obscure, and thicknesses are difficult to determine. These sills are located on the east side of the mountains south of the Wood Ranch house and at the mouth of Cañon Tinaja Prieta.

Flow-banded dikes and plugs

Dikes and plugs with contorted flow banding intrude the upper trachyte, lower rhyolite, and, in one location, the collapse agglomerate in the extensive plain east of Chinati Peak and north of Wood Ranch. These

intrusions are characterized by their gray to blue color and their flow banding, which is similar to the reddish, non-porphyritic domes and flows to the south. The blue color is imparted by small bluish-purple pleochroic grains of sodic amphibole, probably arfvedsonite. Chemical analyses show that they are indeed peralkaline (Henry, unpublished data). Alkali feldspar phenocrysts are present in some outcrops. The dikes range up to 20 m thick and the plugs are as much as 100 m in diameter. A swarm of dikes and plugs in the northernmost part of the caldera is roughly parallel to the caldera boundary, suggesting that their emplacement was controlled by the caldera fracture system.

GEOCHEMISTRY

Because the primary purpose of this report is to describe the geology and evolution rather than the geochemistry of the Chinati Mountains caldera, chemical evolution is discussed only briefly here. Our discussion is intended to provide a general picture of the composition of the rocks of the Chinati Mountains Group, including their regional setting. This discussion is based on the works of Cepeda (1977, 1979), Cameron and others (1982), and Henry and Price (unpublished data). The reader should refer to them for additional interpretation and for the actual chemical data.

All studies show that the rocks of the Chinati Mountains Group exhibit a continuous range in SiO_2 concentration from about 55 to 77 percent. As are other rocks of Trans-Pecos Texas, these rocks are relatively alkalic (or alkali-calcic, in the classification of Peacock [1931]). Chemical analyses confirm the field and petrographic observations that the Chinati Mountains caldera has gone through three cycles of magma differentiation, dominated by fractional crystallization, and eruption. The cycles are: (1) Mitchell Mesa Rhyolite to lower and middle trachyte, (2) lower rhyolite to upper trachyte, and (3) upper rhyolite and blue dikes to West Chinati Stock. Each cycle represented tapping of a differentiated, zoned magma chamber. The more rhyolitic components accumulated at the top of the chamber and erupted first, commonly as ash-flow tuffs. They were followed by the progressively more mafic rocks (to trachytes), which erupted as lava flows. Calculations by Cepeda (1977) show that the fractionation process is dominated by crystallization of plagioclase, along with anorthoclase, pyroxene, amphibole, and biotite.

Major oxide (Henry and Price, unpublished data) and trace and rare-earth element data (Cameron and others, 1982) show that the first two cycles, although spanning the same SiO_2 range, are chemically distinct from the third. The third cycle has generally lower MgO and higher Y, Nb, and Zr concentrations and differentiated to abundant peralkaline rocks. The significance of these differences is unclear. The rocks were all derived from the same magma chamber but possibly some differentiating event or influx of new magma occurred between the second and third cycles.

The composition of the Chinati Mountains Group is similar to that of other volcanic centers (Henry and Price, unpublished data) in the western part of Trans-Pecos Texas. Barker (1977) defined this western area as the metaluminous belt and distinguished it from a more alkalic eastern belt, which also includes feldspathoidal rocks. Rocks of the Chinati Mountains Group seem to fit into a general eastward increase in alkalinity in middle Tertiary igneous rocks from the Sierra Madre Occidental, through Chihuahua, and into Trans-Pecos Texas (McDowell and Clabaugh, 1979). The origin of this trend and the origin of the rocks in a plate tectonic context is the subject of considerable debate beyond the scope of this report.

STRUCTURAL GEOLOGY

The rocks exposed in the Chinati Mountains exhibit the results of three episodes of deformation: Late Cretaceous - Paleogene Laramide-style deformation, which affected the Paleozoic and Cretaceous rocks; middle Tertiary doming and caldera collapse associated with volcanism; and late Tertiary Basin and Range faulting.

Laramide-Type Folding

The Chinati Mountains occupy the boundary area between the intensely deformed Chihuahua tectonic belt to the west and the relatively stable Diablo Platform to the east. Cretaceous rocks in the Chihuahua tectonic belt were tightly folded, overturned, and cut by numerous thrust faults during Late Cretaceous - early Tertiary (Laramide) time (Gries and Haenggi, 1970). Equivalent rocks on the Diablo Platform are flat lying. In the boundary zone just north of the Chinati Mountains, Cretaceous and underlying Permian rocks are gently folded into the Loma Plata anticline and subsidiary Pinto Canyon anticline (Amsbury, 1958). Both anticlines are truncated by late Tertiary normal faults. South of the caldera, Permian and Cretaceous rocks are locally buckled and thrust faulted (Rix, 1953; Ross, 1943) but the outward (away from the caldera) dip of these rocks is due to caldera development.

Caldera Collapse

Dominant structural features in the map area are the normal faults or the eroded scarps of faults that

delineate oval or circular depressions interpreted as caldera collapse zones (fig. 15; pl. 2). Collapse zones include the main Chinati Mountains caldera and several intracaldera collapse structures.

Chinati Mountains caldera

The Chinati Mountains caldera is somewhat elliptical, having a major axis approximately 30 km northwest to southeast and 20 km northeast to southwest. The caldera collapse zone was not a single fault but rather a series of crudely concentric faults of various displacements. The caldera boundary shown in plate 1 and figure 3 is probably near the major break at any location but is not the actual fault. The boundary is an eroded scarp that formed a wall against which volcanic rocks of the Chinati Mountains Group accumulated. In plate 1, the boundary is the outer limit of outcropping Chinati Mountains Group rocks. The lack of deep embayments in the caldera boundary suggests that there was little erosion of the boundary at the currently exposed level; thus, our mapped boundary is near the actual caldera fault.

In the northern part of the caldera, represented by the Shely fault delineated by Amsbury (1958), the caldera boundary is characterized by (1) a pronounced unconformity where upper trachyte and lower rhyolite overlie Cretaceous and Permian rocks, volcanic rocks of the Shely Group, and volcanic and intrusive rocks of the Infiernito caldera, and (2) intrusions associated with the middle trachyte of the Chinati Mountains Group. The contact between the volcanic rocks of the Chinati Mountains Group and the older rocks is well exposed along much of the northern boundary; the contact dips steeply southward, into the caldera, which indicates that the volcanic rocks filled a steep-sided depression. Lenses of breccia and conglomerate (collapse agglomerate) of clasts derived from the older rocks are present in places along the contact. These deposits are interpreted as being material derived from the rim of the caldera. The contact is depositional and there has been no movement across the contact. The contact undoubtedly represents an eroded scarp along the caldera fault but not the actual fault.

The presence of intrusions along the contact suggests that the actual fault is near the exposed caldera wall. Presumably the intrusions rose within a zone of weakness, the actual caldera fault, and the fault is buried by the younger volcanic rocks of the Chinati Mountains Group. This interpretation is illustrated by the cross sections in plate 2. The actual fault should have dropped the same rocks that are exposed north of the fault down into the caldera. Neither these rocks nor the base of the Chinati Mountains Group is exposed within the caldera.

The northeastern edge of the caldera boundary is marked by the contact of the lower rhyolite with rocks of the Infiernito caldera, including the Ojo Bonito intrusion, which Duex and Henry (1981) interpreted as being the resurgent dome of the Infiernito caldera. A prominent lineament that parallels the caldera boundary mostly within the Ojo Bonito intrusion may be part of the caldera fracture system. From here all the way around to the southwestern edge of the caldera, the boundary is marked by the lower rhyolite abutting against older rocks.

South of the Infiernito caldera, the Chinati Mountains caldera boundary is placed at the contact between the lower rhyolite and the Pennsylvanian Cieneguita Formation or the Permian Cibolo Formation, which dip away from the caldera at angles of 15° to 20°. Streams have incised along this contact. The lower rhyolite exhibits chaotic flow banding adjacent to the Cibolo Formation, probably the result of the buttressing effect of the caldera wall where the lavas ponded against it.

The caldera wall is buried for several kilometers along its southeastern margin by gravels that Rix (1953) mapped as Quaternary terrace deposits. We favor an alternative hypothesis in which the gravels are remnants of the Perdiz Conglomerate, a coalesced alluvial fan of debris shed off the Chinati Mountains after volcanism ceased (Walton, 1979).

The southern edge of the caldera emerges from beneath the gravels about 3 km northwest of Shafter. From there to the edge of Presidio Bolson, a distance of about 7 km, the caldera wall is well exposed. The presence of numerous normal faults in the Shafter mining district suggests that, in this area, caldera collapse occurred along a complex fault system.

The caldera wall just northwest of Shafter is marked by nearly flat lying, lower rhyolite, deposited against the Permian Ross Mine Formation, which dips away from the caldera, toward the southeast, at 15°. The contact dips steeply into the caldera, as it does along the north rim of the caldera. The lower rhyolite and Permian rocks are in contact for about 1 km; southwest of there, they diverge. The caldera boundary is marked by a trough eroded into collapse agglomerate deposited against the former caldera wall. Rix (1953) mapped most of this zone as being intrusive rock but, in fact, there are only scattered intrusions, some of them sills in the Permian rocks. The rest of the material is a chaotic mudflow (?) breccia, composed of clasts up to 1.5 m in diameter. Presumably the lower rhyolite ponded against the caldera wall, which consisted in places of Permian and Cretaceous rocks and, in other places, of breccias piled against the wall. Erosion of the lower rhyolite, in turn, provided clasts to the breccias. A few altered, porphyritic dikes cut the breccia.

The location of much of the ore at Shafter is controlled by east-west- and northeast-trending normal faults that cut the Permian and Cretaceous rocks (Ross, 1943). The faults generally parallel the caldera boundary in a zone as far as 4 km south of the boundary, and are dominantly down to the north. We interpret these faults as being part of the caldera collapse system, although most of the collapse must have occurred within the mapped boundary. The control of mineralization by these structures suggests that the Shafter deposits are genetically related to the Chinati Mountains caldera, as is discussed later in more detail.

Near the Ross Mine 6 km west of Shafter, large (up to 400 m) blocks of the Cretaceous Presidio Formation and

the Permian Mina Grande Formation are stratigraphically out of place relative to the south-dipping Permian Ross Mine Formation to the south. These blocks either are faulted, down to the north, along an east-west-trending fault that is part of the caldera collapse zone or are slumped into the caldera from the caldera wall after collapse. In the latter case, the blocks are extremely large examples of the collapse agglomerate and are analogous to the megabreccias described by Lipman (1976).

The caldera wall is buried to the west beneath upper Tertiary sediments of Presidio Bolson. However, minor outcrops of volcanic rocks and Cretaceous sedimentary rocks are exposed within bolson fill in places, which suggests that the western edge of the caldera extends at least a few kilometers into the bolson. A warm (30°C) spring occurs within the bolson and adjacent to a rhyolite hill that is probably part of the Chinati Mountains Group. Spring water may rise from depth along caldera-related or Basin and Range fractures.

Older rocks around the entire exposed circumference of the caldera dip outward at as much as 40°, although most dips are 15° or less. Although the pre-Tertiary rocks and possibly even some of the older Tertiary rocks were deformed during Laramide folding, most of this outward dip probably resulted from pre-collapse doming of the caldera area. Given that the dip is radially outward, follows the caldera boundary, in places involves Tertiary volcanic rocks, and decreases in Cretaceous rocks southward away from the caldera, the dip is thought to be related to the caldera. We believe that the rising magma chamber, which eventually produced the Mitchell Mesa Rhyolite and the Chinati Mountains Group, domed the preexisting rocks.

Intracaldera collapse zones

The dominant structural features within the Chinati Mountains caldera are several inner collapse zones. The most prominent is a circular, 5-km-diameter zone in the south-central part of the caldera. Normal faults on the boundary of this zone dip approximately 60° inward, and displacements approach 200 to 300 m. At the north boundary of the collapse zone, lower rhyolite flows dip 15 to 20° toward the fault on the upthrown side. On the downthrown side, flat-lying beds of the upper rhyolite lie at a slightly lower elevation. The trace of the ring fracture is covered by a thin veneer of upper rhyolite in many places.

Another collapse zone is partly delineated by an arcuate zone of non-porphyritic domes near the eastern margin of the caldera. Displacement along the trend marked by the domes may exceed 200 m, but the extent and size of this collapse is unknown because the zone is covered by the younger upper rhyolite.

Measurement of stratigraphic sections along the length of the mountains indicates a consistent variation in thickness of the lower rhyolite (fig. 15). This change documents a thickening of the lower rhyolite in the central part of the mountains and suggests an earlier period of sagging or faulting in which an elongate depression formed. One flow 20 m thick is present in the northernmost exposure of rhyolite. The lower rhyolite thickens southward to 460 m in Cañon Tinaja Prieta, where it is composed of at least five flows. Because the base is not exposed here, 460 m is a minimum thickness. The unit thins to the south to less than 300 m near Cerro Oroña. Although the thickest measured accumulations of lower rhyolite are west of the upper rhyolite collapse zone, the two collapse zones may be concentric.

Exposures of the lower and middle trachyte are too restricted to determine possible thickness variations. The upper trachyte appears to have attained its maximum thickness near Chinati Peak.

Basin and Range Faults

Normal faulting beginning in late Oligocene or early Miocene time (approximately 23 mya; Dasch and others, 1969; McDowell, 1979) has cut Trans-Pecos Texas into a series of horsts and grabens. The Chinati Mountains and Presidio Bolson form one such horst-graben system. Basin and Range faulting has continued to the present.

Both Rix (1953) and Amsbury (1958) mapped the West Chinati fault along the prominent escarpment separating the Chinati Mountains from Presidio Bolson. The fault is not well exposed, but its presence can be inferred from the linearity and sizeable relief of the escarpment, from the disappearance of volcanic rocks of the Chinati Mountains Group across the scarp and the appearance of other igneous rocks, from the great thickness of basin fill on the downthrown side, from the large gravity anomaly associated with the scarp (Mraz and Keller, 1980), and from the presence of Quaternary scarps within the bolson west of the presumed West Chinati fault. Displacement on the West Chinati fault could be more than 1 km.

West-Trending Structures

Many apparently unrelated structures exhibit a pronounced westerly trend. Examples include clastic dikes in the bolson fill west of the West Chinati Stock, mineralized shear zones in the stock, the Shely fault and associated faults, microgranite dikes in San Antonio Canyon, flow-banded dikes and plugs that crop out east of Chinati Peak, and faults in the Shafter area.

Udden (1904) noted that the country rock in San Antonio Canyon is cut by a red granitic rock that formed west-trending, dike-like ridges. Baker (1934a), in his description of the geology of the Chinati Mountains, noted that the vertical shear zone at the Burney Prospect has horizontal slickensides suggestive of strike-slip motion. Drilling (Dennis, 1947) revealed that the vein dips to the south at depth.

The pronounced west-trending structural grain may be indicative of an underlying and older structural weakness in the crust. Although the orientations of the Shely fault and of the faults at Shafter can be explained by relating these faults to the northern and southern

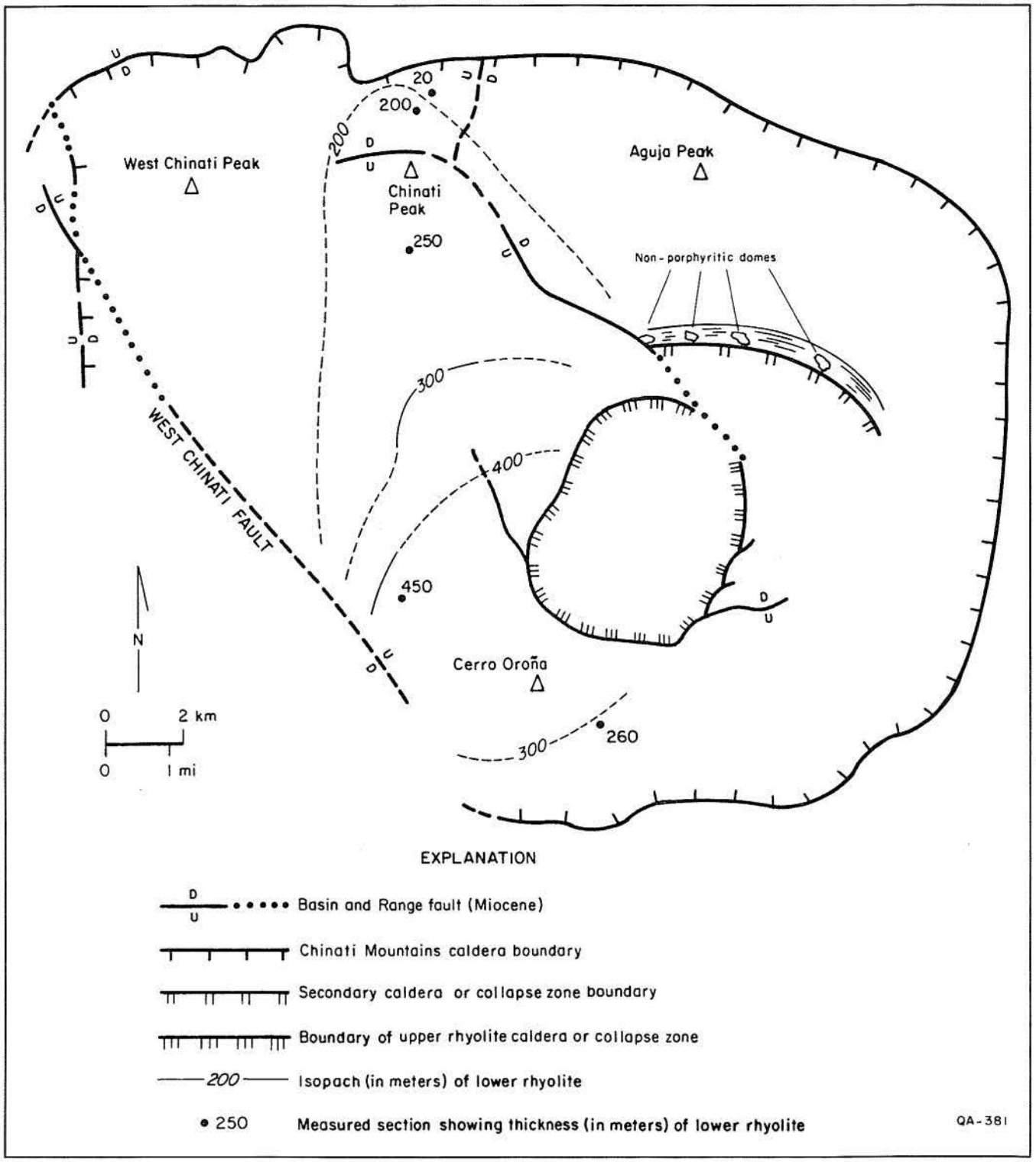


Figure 15. *Structural features of the central Chinati Mountains.*

margins of the caldera, other structures must have had a different origin. Baker (1934b) believed that most of the silver-lead fissure vein mineralization in Trans-Pecos Texas is associated with west-trending structures. The intersection of the west-trending zone and the prominent north to northwest structural trends of Trans-Pecos Texas (Barker, 1977; Woodward and others, 1975; Dickerson, 1980; Muehlberger, 1980) may have influenced the position of the Chinati Mountains caldera.

Water

Water is an important commodity in Presidio County, which receives less than 20 inches of rainfall a year. The local economy, which is based on ranching and some farming, requires water for livestock and household consumption.

Potable water in amounts more than adequate for ranching can be obtained from wells in stream gravels, in fractured volcanic rocks, and in bolson fill along the western margin of the Chinati Mountains. On the Chinati Ranch, electric pumpjacks have replaced the venerable windmills as a source of power for these wells.

Seeps and springs maintain surface flow in Pinto Creek and Cibolo Creek throughout most of the year. Subsurface flow, in stream gravels, occurs year round in Cañon Tinaja Prieta. Other arroyos flow only after heavy rains.

Geothermal Potential

The geothermal potential of this area has been investigated by Henry (1979), who measured temperatures and estimated flow of the numerous hot springs and wells along this part of the Rio Grande Valley. The only hot spring in the map area is at the Agua Chano Ranch (formerly the Gonzales Ranch); the spring discharges 30°C water. This spring is near the southwest boundary of the Chinati caldera, where it is largely buried beneath bolson fill.

Henry (1979) concluded that almost all hot springs occur along normal faults that act as conduits for the rise of hot water from depth. Geothermal water is simply meteoric water that has circulated to a sufficient depth to be heated and then returned to the surface. The source of heat is the Earth's normal heat flow; the geothermal gradient in this area is about 40°C/km, a gradient about twice that found to the east in the cratonic part of Texas. The high gradient is due to thinning of the crust during Basin and Range extension and faulting. All igneous rocks in Trans-Pecos Texas are too old to supply any heat to hot springs.

The highest measured temperature of hot-spring water near the map area is 90°C at Ojos Calientes in Chihuahua, Mexico, approximately 30 km northwest of the Chinati Mountains. Temperatures as high as 80°C at depths of about 900 m have been measured in several wells drilled for hydrocarbon exploration in Presidio Bolson or nearby. Temperatures of geothermal water, estimated using chemical geothermometers, range as high as 160°C, but most are less than 100°C.

Geothermal water in the Rio Grande area is not hot enough to generate electricity using presently available technology. The water, however, is hot enough for space and industrial heating, and the artesian flow of several wells is probably enough to support such use. However, the small population makes the use of geothermal water for purposes other than bathing and irrigation unlikely in this area.

Base and Precious Metals

Lead, zinc, silver, copper, and molybdenum mineralization occurs at several localities within the Chinati Mountains. Many or all of the deposits are probably genetically, as well as spatially, associated with the Chinati Mountains caldera.

The Shafter mining district, the largest silver producing area in the state (Ross, 1943), is located on the southern margin of the Chinati Mountains caldera. Sphalerite, galena, and argentite occur in veins and mantos in Permian and Cretaceous limestones where the rocks have been intensely fractured along and just outside the caldera boundary. From the late 1800's to 1942, the Shafter mines produced more than 31 million ounces of silver, then valued at $18 million. Recent exploration has delineated additional reserves.

The Shafter ores were probably introduced by hydrothermal solutions derived from shallow igneous intrusions and deposited by reaction with dolomitic limestones. Although the timing of mineralization has not been established, the deposits are probably related to caldera activity for the following reasons: (1) Some exposed igneous rocks (for example, Red Hill discussed later) also show the effects of hydrothermal alteration (Ross, 1943; Price and Henry, 1982). (2) The deposits are controlled by faults that parallel the caldera boundary and appear to be part of the ring-fracture system. (3) The overall zone of deposits at Shafter parallels the southern margin of the caldera; no deposits have been found more than a few kilometers from the caldera boundary.

Further work is needed to determine if an older caldera is present east and south of the Shafter district. Field work by Hardisty (1982) indicated that the Morita Ranch Formation, which crops out at Shafter and southward, probably represents the cauldron facies of a caldera.

Disseminated copper-molybdenum mineralization occurs in the intrusion at Red Hill west of Shafter (McAnulty, 1976; Price and Henry, 1982). The intrusion is a sericitized quartz-monzonite porphyry having abundant stockwork quartz-pyrite veinlets. It probably predates caldera collapse because altered fragments of pyritic porphyry occur in collapse agglomerate immediately north of the intrusion. The stockwork-veined intrusion and the Permian sandstone next to the intrusion contain anomalous concentrations of copper and molybdenum. Garnet skarn developed in limestone of the Ross Mine Formation is enriched in copper and zinc. Small amounts of silver occur in the stockwork-veined rocks and in the skarn. In fact, the intrusion may be related to the hydrothermal system that is responsible for the Shafter silver deposits. The potential for molybdenum mineralization probably exists in other parts of the caldera, as suggested by the high molybdenum content (up to 12 ppm) of the upper rhyolite (Cepeda and others, 1981).

Numerous prospects and small abandoned mines occur within the West Chinati Stock along the western side of the Chinati Mountains (McAnulty, 1972), and

sporadic exploration and mining have occurred at least since the 1890's. Lead, zinc, and silver minerals and fluorite occur in fissure veins at the Burney Prospect and the San Antonio Mine. Similar deposits are scattered throughout the West Chinati Stock. Chrysocolla and copper carbonates coat joints and fractures in a variety of rock types related to the West Chinati Stock in outcrops west of the West Chinati fault in Presidio Bolson. The potential for economic copper deposits at depth in down-dropped parts of the West Chinati Stock is unknown. All of the deposits appear to be related to resurgent doming of the Chinati Mountains caldera.

The potential for minable base and precious metal deposits in the Chinati Mountains is excellent. Numerous known occurrences and areas of alteration and apparent association with caldera igneous activity suggest that other parts of the caldera are favorable for deposits.

Uranium

Concentrations of uranium as high as 0.34 percent U_3O_8 (Amsbury, 1958) occur in fracture zones in porphyritic rhyolites of the Allen Intrusions in Pinto Canyon. Uranium deposits could also occur in the upper rhyolite or associated rocks in the Chinati Mountains (Cepeda and others, 1981), or in basin fill of the Presidio Bolson (Duex and others, 1980; Henry and others, 1980).

Uranium in the Allen Intrusions occurs as the secondary U^{+6} minerals, autunite, metatorbernite, and tyuyamunite (Reeves and others, 1979) and as uraniferous Fe-Mn oxyhydroxides (Henry and others, 1980). The uranium occurs in fracture zones, which probably formed during cooling of the shallow intrusions. The highest grades are at or near the surface and probably result from supergene enrichment of more pervasive, lower grade deposits. Exploratory drilling by Wyoming Minerals in the late 1970's confirmed that numerous fracture zones contain uranium, to depths of at least 60 m, and that the fracture zones are smaller and of lower grade at depth.

Henry and others (1980) interpreted the origin of the uranium as follows: (1) release of uranium from rocks of the Allen Intrusions or associated tuffaceous rocks of the Shely Group, (2) transport in oxidizing ground water into and through the fracture zones, (3) accumulation either by precipitation involving oxidation-reduction reactions or by adsorption by amorphous Fe-Mn oxyhydroxides, and (4) relatively recent weathering and oxidation of U^{+4} minerals or the oxyhydroxides with formation of U^{+6} minerals.

Henry and others concluded that although sufficiently high grades exist for economic deposits, the overall tonnage of deposits is probably too low and too scattered for mining. The area could be more favorable if hydrothermal, pitchblende-bearing veins occurred; however, their existence is unlikely.

The peralkaline upper rhyolite has been enriched relative to other rocks of the Chinati Mountains Group in the trace elements U, Th, Mo, Be, Li, and F either by fractional crystallization or by thermogravitational diffusion (Hildreth, 1979) or by both processes. Igneous rocks of Trans-Pecos Texas generally are enriched in these elements relative to typical calc-alkaline rocks (Henry and Price, unpublished data); for example, a vitrophyre of the upper rhyolite contains 13 ppm U, 12 ppm Mo, and 1,300 ppm F (Cepeda and others, 1981). Devitrified and granophyrically crystallized samples of the upper rhyolite contain lower concentrations of the six elements than does the vitrophyre. Cepeda and others (1981) concluded that crystallization of the thick, volatile-rich rock released the elements in a hydrothermal fluid and that hydrothermal uranium deposits could occur in the Chinati Mountains, especially as uraniferous fluorite.

Another potential site of uranium deposits is Presidio Bolson. Presidio Bolson is filled with detritus shed from the Chinati Mountains; surface- and ground-water flow is from the mountains into the bolson. Igneous rocks of the Chinati Mountains generally have moderately high uranium concentrations (Duex and others, 1980). If significant amounts of uranium were released to ground water, the water and uranium would have accumulated in Presidio Bolson, which was a closed basin until integration of the Rio Grande during the Pleistocene. If reducing conditions or other environments that concentrate uranium occur in the bolson fill, economic uranium deposits could have formed. Initial exploration should evaluate the oxidation-reduction potential of ground water and sediments in the bolson and the concentrations of uranium and associated trace elements in the ground water.

GEOLOGIC HISTORY

Pre-Tertiary Geologic History

The late Paleozoic history of the Chinati Mountains is characterized by deposition of marine sedimentary rocks, principally carbonates, unfossiliferous shale, and sandstone. After a long period of erosion that extended through most of the Mesozoic, Cretaceous limestone, sandstone, and shale were deposited on the Paleozoic section. Late Cretaceous - early Tertiary deformation (Laramide) produced broad folding in the area.

Mid-Tertiary Volcanic Activity

The Tertiary geologic history of the region is complex. Because the areas west and east of the Chinati Mountains are covered by gravel, we examined adjacent areas to the north and south. Three sequences of older rocks, the Shely Group on the north, the Infiernito sequence on the northeast, and the Morita Ranch Formation on the south, flank the Chinati Mountains (fig. 3). All three sequences were emplaced during the Oligocene.

Morita Ranch Formation

The Morita Ranch Formation (Dietrich, 1965) crops out near Shafter along the southeast margin of the caldera. It consists of basalt, rhyolite, tuff, and ash-flow tuff, which Dietrich divided into four units. Reconnaissance mapping (Hardisty, 1982) in the Cienega Mountains Quadrangle suggests that additional units are present there.

A lower part of the Morita Ranch Formation, consisting of rhyolitic lava flows, ash-flow tuff, and tuffaceous sediment, is petrographically similar to rocks of the Infiernito caldera and may have been derived from there. A preliminary K-Ar age of an ash-flow tuff in T1 is 37 m.y., similar to ages of Infiernito rocks (Henry, unpublished data). Alternatively, these rocks may have been derived from a source now buried beneath the Chinati Mountains caldera. In any event, T1 and rocks of the Infiernito caldera are the oldest igneous rocks in the Chinati Mountains.

Upper parts of the Morita Ranch Formation disconformably overlie the lower part and are considerably younger. Ash-flow tuff of Rix's (1953) T4 is 34.9 m.y. old (McDowell, unpublished data) and, as discussed previously, is probably correlative with Ts6 ash-flow tuff of the Shely Group, which is the same age (Henry, unpublished data). If so, these two units compose one of the largest local ash-flow tuffs of the area. Its source is probably within the area of the Chinati Mountains caldera and its eruption may have led to some caldera collapse. However, it is too old to be a part of the actual Chinati Mountains sequence.

Preliminary K-Ar ages of the Tm3a lava flow and the Tm3b ash-flow tuff are about 33 m.y. (Henry, unpublished data). Although the ages seem inconsistent with the age of T4, the rocks are not in contact, and no relative age relations have ever been established. Our new data indicate revision is necessary. McAnulty (1976) interpreted the Morita Ranch Formation as a cauldron facies, suggesting that the younger Chinati Mountains caldera formed at the junction between two older, roughly contemporaneous calderas. Additional work by Hardisty (1982) in the Cienega Mountains Quadrangle appears to substantiate this origin of the Morita Ranch Formation, although no evidence of a caldera boundary has been identified for this sequence.

Infiernito caldera sequence

The Infiernito caldera sequence erupted from a caldera centered on the northeast boundary of the Chinati Mountains caldera. Preliminary K-Ar isotopic ages (Henry, unpublished data) indicate that the Infiernito caldera is about 37 m.y. old, and, as noted, this sequence is contemporaneous with the lower part of the Morita Ranch Formation. Rocks of the Infiernito caldera are clearly older than both the Shely Group and the upper part of the Morita Ranch Formation. The Shely Group overlies rocks of the Infiernito caldera, and clasts eroded from the caldera occur within conglomerates of the Shely Group. Similar clasts occur within conglomerates that underlie Tm4 of the Morita Ranch Formation. The Infiernito caldera is also older than the Chinati Mountains caldera, which has truncated the southern part of the Infiernito caldera.

Shely Group

Amsbury (1957) subdivided the Shely Group into eight units consisting of plagioclase trachyte, tuffaceous sediment, rhyolite, and rhyolitic ignimbrite. The Shely Group overlies rocks of the Infiernito caldera and may have been deposited in a moat along the eroded western margin of the Infiernito caldera. Rocks of the Shely Group were erupted from a local source, probably now buried beneath the Chinati Mountains caldera, and filled topographic lows in the area. The Mitchell Mesa Rhyolite overlies Ts6 ash-flow tuff near the northern limit of the Shely Group. Clasts of the Mitchell Mesa do not occur in any of the sedimentary units of the Shely Group. K-Ar ages of the Shely Group range from about 37 m.y. on lower parts to 34 m.y. on upper parts (Henry, unpublished data). The entire Shely Group is thus older than the Mitchell Mesa Rhyolite, and the Shely Group is definitely older than the Chinati Mountains caldera, which truncates it on the south.

McAnulty (1976) and Cofer (1980) suggested that the Shely Group represents a cauldron facies that accumulated in the same basin as the Chinati Mountains Group. This interpretation is inconsistent with the stratigraphic and structural relations in the northern Chinati Mountains.

Chinati Mountains caldera

Eruption of the Mitchell Mesa Rhyolite led to collapse of the Chinati Mountains caldera. A large but unknown thickness of Mitchell Mesa Rhyolite probably accumulated within the caldera. The down-dropped caldera floor was covered by lava flows of the lower and middle trachyte (and possibly by still older, but unexposed, volcanic rocks) with interbedded lenses of conglomerate and breccia adjacent to the caldera margin. Trachyte intrusions associated with the middle trachyte were emplaced along the northern boundary of the caldera.

The caldera was then filled by alternating sequences of trachytic and rhyolitic volcanic rocks. Alternation of trachytic and rhyolitic volcanism can be correlated with episodic secondary caldera collapse. The lower and middle trachyte blanketed the caldera floor and smoothed out any uneven topography. The formation of a secondary collapse zone, in which the lower rhyolite accumulated, signaled another episode of collapse within the Chinati caldera. Then another period of trachytic volcanism (upper trachyte) ensued, creating maximum accumulation near Chinati Peak. After the eruption of the upper trachyte, a third episode of collapse occurred over a 5-km-wide zone during the eruption of the upper rhyolite and non-porphyritic domes and flows in the central part of the map area. Blue dikes and plugs with peralkaline affinities

Table 2. *K-Ar ages of rocks of the Chinati Mountains caldera.*[A]

Sample	Rock	Mineral[B]	%K	$\%^{40}Ar^*$	$^{40}Ar^*$ ($\times 10^{-6}$ scc/gm)	Age (m.y. $\pm 1\ \sigma$)
SL2	West Chinati Stock	H	0.973 0.978	63 49	1.22 1.23	32.0 ± 0.7
MGF-038	Blue dike	K	5.42 5.42 5.43	94	7.01 6.74	32.3 ± 0.7
Chin 431	Upper rhyolite	H	1.08 1.08	17 19	1.28 1.47	32.5 ± 3.1
Chin 445	Gonzales rhyolite	K	4.79 4.83	91 92	6.20 6.08	32.6 ± 0.7
Chin 209	Lower rhyolite	K	5.34 5.31	93 94	6.93 6.77	32.8 ± 0.7
Tmm	Mitchell Mesa Rhyolite	K	[C]	[C]	[C]	32.3 ± 0.7[C]

$\lambda\beta = 4.963 \times 10^{-10}\ yr^{-1}$; $\lambda_{E\cdot E} = 0.581 \times 10^{-10}\ yr^{-1}$; $^{40}K/K = 1.167 \times 10^{-4}$
[A]Ages determined by Fred W. McDowell and Christopher D. Henry
[B]H = hornblende; K = K-feldspar
[C]Age is average and standard deviation of 18 K-feldspar determinations; individual data are available upon request.
*Radiogenic

intruded the lower rhyolite and upper trachyte in the area north of the Wood Ranch house and are probably associated with the peralkaline upper rhyolite.

The West Chinati Stock and related intrusive rocks were emplaced after the eruption of almost all of the Chinati Mountains Group, but their stratigraphic position relative to the upper rhyolite is not firmly established. Chemical data (see section on Geochemistry) indicate that the stock must have followed the upper rhyolite. These intrusions arched the overlying volcanic rocks and may have formed minor flows if the magma broke through the overlying volcanic layers. At this time, volcanic activity ceased in the Chinati Mountains area.

The entire cycle of activity, from eruption of the Mitchell Mesa Rhyolite to emplacement of the West Chinati Stock and eruption of the late peralkaline rocks, took place within the resolution of K-Ar dating (table 2). Errors in the ages are such that none of the ages is distinguishable. Probably the entire cycle lasted no more than 1 m.y.

The caldera model of Smith and Bailey (1968) has become the standard with which most calderas are compared. This caldera model lists seven distinct events, but it can be cyclic, with all or part repeated. It is worthwhile to review the general sequence proposed by Smith and Bailey and to compare that sequence with the known events in the Chinati Mountains caldera. Table 3 lists the seven general events and the examples, evidence, or interpretation of

Table 3. *Comparison of Smith and Bailey (1968) caldera model with events in the Chinati Mountains caldera.*

Stage	Smith and Bailey model	Evidence in Chinati Mountains caldera
I	Regional tumescence and generation of ring fractures	Radial outward dip of pre-caldera rocks; normal faults parallel to southern and northern caldera boundary. Intrusion of Red Hill stock and related sills?
II	Caldera-forming eruptions	Eruption of Mitchell Mesa Rhyolite.
III	Caldera collapse	Formation of large, subcircular basin in which Chinati Mountains Group accumulated.
IV	Pre-resurgent volcanism and sedimentation	Emplacement of lower trachyte, middle trachyte, and lower rhyolite and formation of upper trachyte volcano near Chinati Peak. Very little sedimentary material in exposed caldera.
V	Resurgent doming	Intrusion of West Chinati Stock; uplift of Chinati Mountains Group at least through upper trachyte.
VI	Major ring-fracture volcanism	Minor intrusions of blue dikes and plagioclase trachytes in northern part of caldera; some sills and dikes outside southern boundary of caldera may have been emplaced at this time.
VII	Hot spring and solfataric activity	Evidence is not preserved.

similar events in the Chinati Mountains. It is necessary to elaborate on the events because activity in the Chinati Mountains was cyclic, because the relative timing of some events is unknown, and because the events in the Chinati Mountains do not exactly follow the idealized model.

The Chinati Mountains underwent two additional collapse or subsidence events beyond the dominant collapse associated with eruption of the Mitchell Mesa Rhyolite. The three collapses are associated with the three most silicic and presumably most differentiated volcanic rocks: the Mitchell Mesa Rhyolite, the lower rhyolite, and the upper rhyolite. Each event may represent development of a differentiated cap within the magma chamber, eruption of the cap, and subsequent subsidence.

The relative timing of some events is not established. The West Chinati Stock is not in contact with the upper rhyolite; it is not known which was emplaced first, but chemical data indicate that the stock is younger than the upper rhyolite. Likewise, the timing of several intrusions cannot be established beyond what they are observed to intrude. For example, the blue dikes are at least as young as the upper trachyte. They are probably contemporaneous with the upper rhyolite because both are peralkaline. However, this assumption cannot be proven.

The Chinati Mountains caldera contains little sedimentary material; tuffaceous sediments are particularly lacking. Nevertheless, it is well established that the Chinati Mountains produced abundant tuff after eruption of the Mitchell Mesa Rhyolite. The Tascotal Formation is tuffaceous sediment that was eroded from around the Chinati Mountains and deposited as broad, coalescing alluvial fans (Walton, 1979). Deposition of the Tascotal Formation was probably contemporaneous with volcanic activity within the caldera. Perhaps tuffaceous material did accumulate in upper, now eroded, levels of the caldera. The original caldera rim was certainly much higher than the currently exposed rim. Much material must have been removed from the caldera, and tuff, being non-resistant to erosion, would have been eroded most rapidly.

Late Cenozoic Events

Basin and Range faulting cut the igneous sequence along generally north to northwest trends, probably beginning about 18 to 23 mya (Dasch and others, 1969) after all igneous activity in the Chinati Mountains had ceased. At that time, debris derived from the newly formed mountain blocks began filling the down-dropped areas, including Presidio Bolson to the west. Later, the Rio Grande, a through-flowing stream, initiated the dissection of the bolson fill, a process which continues to the present.

SUMMARY

The Chinati Mountains caldera is the largest caldera in Trans-Pecos Texas. It is the source of the Mitchell Mesa Rhyolite, the largest and most extensive ash-flow tuff of Texas. Volcanism in the Chinati Mountains area began with eruptions of lava flows and ash-flow tuffs from the Infiernito caldera, which may also have been the source of some of the oldest rocks of the Morita Ranch Formation. After these earliest events, volcanic rocks of the Shely Group and upper part of the Morita Ranch Formation were erupted from unidentified, local sources before formation of the Chinati Mountains caldera.

Eruption of the Mitchell Mesa Rhyolite led to collapse of the elliptical, 30-km by 20-km Chinati Mountains caldera. Most collapse was along a crudely concentric and nearly continuous fault zone. The fault is represented by an eroded fault scarp where rocks of the Chinati Mountains Group, which fills the caldera, abut against older rocks exposed in the caldera wall. Minor normal faults, parallel to but outside the caldera boundary, are probably also part of caldera collapse. Coarse breccias, agglomerates, and conglomerates were deposited at the edge of the caldera from debris shed off the oversteepened caldera wall.

The caldera was filled by a repetitive sequence of rhyolitic to trachytic lava flows, ash-flow tuff, and intrusive rocks. Following the Mitchell Mesa Rhyolite, the lower and middle trachytes are the oldest exposed parts of the Chinati Mountains Group. They are porphyritic lava flows with phenocrysts of plagioclase, anorthoclase, and augite in a groundmass of feldspar, clinopyroxene, and opaque minerals. The lower rhyolite is a series of porphyritic lava flows containing anorthoclase and minor quartz phenocrysts in a groundmass of alkali feldspar, quartz, altered mafic minerals, and opaque minerals. The lower rhyolite filled an elongate sag or secondary collapse zone within the caldera and reached a thickness of at least 450 m. The upper trachyte consists of two distinct, textural variations: a lower part with 5 to 15 percent, small plagioclase phenocrysts, and an upper part with 25 percent, larger plagioclase phenocrysts. Both parts also contain augite phenocrysts in a groundmass of feldspar, clinopyroxene, and opaque minerals. The upper trachyte formed a large stratovolcano in the north-central part of the caldera but probably filled and possibly even overflowed the caldera. The Petan Trachyte may be the equivalent rock outside the caldera.

Eruption of the upper rhyolite formed a small, 5-km-diameter, secondary caldera in the central Chinati Mountains. The upper rhyolite is a peralkaline ash-flow tuff with anorthoclase and quartz phenocrysts in a devitrified to granophyric groundmass, which commonly contains sodic clinopyroxene. In many places the upper rhyolite exhibits laminar flow structure. A

series of peralkaline, rhyolitic dikes, containing minor alkali feldspar phenocrysts and sodic amphibole crop out in the northern part of the caldera and are probably related to the upper rhyolite.

The Chinati Mountains Group was intruded and uplifted by the West Chinati Stock, the resurgent dome of the caldera. It is zoned from a quartz-monzodioritic margin to a granitic core. Numerous dikes and plugs of microgranite cut the main intrusion.

The Chinati Mountains were also the source of the Tascotal Formation and Perdiz Conglomerate, which form alluvial fans generally east of the caldera. The Tascotal Formation consists of tuffaceous sediment, derived from erosion of a blanket of air-fall and ash-flow tuff that formerly covered the Chinati Mountains. The Perdiz Conglomerate consists of coarse debris eroded from the Chinati Mountains once the blanket of tuff had been stripped off.

K-Ar ages show that all of the igneous activity of the Chinati Mountains caldera occurred between 32 and 33 mya. Earlier activity of the Shely Group, Morita Ranch Formation, and Infiernito caldera extends back to 37 to 38 mya.

Several ore deposits and minor prospects are genetically related to the Chinati Mountains caldera. The most significant is the Shafter silver district, which produced 31 million ounces of silver and minor lead and zinc, dominantly from Permian dolomitic limestones along the southern edge of the caldera. A quartz monzonite porphyry intrusion west of Shafter contains disseminated copper-molybdenum mineralization and is also at the south edge of the caldera. Several fissure vein deposits containing silver, lead, zinc, and fluorite occur in the West Chinati Stock. Oxide copper prospects also exist in rocks of the West Chinati Stock in Presidio Bolson at the western edge of the caldera.

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