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GEOLOGY OF THE LAS VEGAS AREA

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INTRODUCTION

The purpose of this paper is to describe the geology in the vicinity of Las Vegas, New Mexico. The study area is here defined as being bordered on the south by the village of Romeroville, on the east by the Pecos Arroyo River, on the north by the northern termination of a large mesa which is located approximately 1 mile to the northwest of Storrie Lake, and on the west by the village of Hot Springs.

The first part of the report deals with the geomorphology of the Las Vegas area, including an examination of the area's meteorology which strongly influences the discharge of the two principal streams that flow through the region. These streams have been responsible for the formation of many of the area's physiographic features. One of them, the Gallinas River, has cut a deep valley into the Sangre de Cristo Mountains and has exposed a sequence of rocks ranging in age from Precambrian to Cenozoic. The stratigraphy of these rocks will be examined together with the area's paleogeography and an investigation of the structural evolution of the Las Vegas area. A description and possible explanations of the origins of the Montezuma hot springs will be presented.

GEOMORPHOLOGY

The Las Vegas area straddles the boundary between the High Plains and Southern Rocky Mountain physiographic provinces. It is situated on the top of the Las Vegas Plateau, which is bounded on the south and east by the Canadian Escarpment and on the west by the Sangre de Cristo Mountains. The elevation of the area ranges from about 6,400 ft to 6,800 ft; the mountains to the west attain elevations of 1 0,5 00 ft.

The climate of the Las Vegas area is semi-arid but changes to sub-humid at higher elevations to the west of the city. Patches of humid climate are found at the highest elevations. The area receives its maximum precipitation during July and August when the Bermuda High moves into the Gulf of Mexico and pumps (via a clockwise circulation) moist, warm air into New Mexico. Even though Las Vegas is situated in the rainshadow of the Rocky Mountains some precipitation is derived from fall, winter and spring cyclones which are directed over the area by the easterly flowing jet stream.

In the fall and early winter months, the jet stream is located close to the Canadian border and only occasionally dips far enough south, with its accompanying cyclones, to bring any significant precipitation to the Las Vegas area; however, during late winter and early spring, the jet stream moves further south and the region receives snow and/or rain and wind.

The area's moderately high elevation, southerly latitude and interior continental position combine to produce a dry, mild climate without severe climatic extremes. These same factors and the local geography, have resulted in the area having only a limited watershed. There are only two major streams that flow through the region, the Gallinas and the Pecos Arroyo Rivers. The Gallinas is the largest of the two. It has its head waters in Gallinas Canyon (located approximately 5 miles northwest of Las Vegas) and has an average discharge of 14,700 acre ft (Hendrichs, 1964). Its discharge mostly depends upon precipitation but is also influenced by the perennial springs which empty into it, including those at Montezuma.

The Gallinas River has cut a deep, moderately modified V-shaped valley into the Sangre de Cristo Mountains. There is no evidence of glacial activity as is seen in the Mora Range 30 miles north of •Las Vegas. There is extensive evidence, in the form of entrenched meanders and abandoned river terraces, that at least a portion of the Gallinas River has undergone three stages of rejuvenation during Pliocene and/or Pleistocene time.

The Gallinas River flows southeast as it leaves Gallinas Canyon. At the mouth of Gallinas Canyon, some of the stream's water is diverted via a man-made canal to Storrie Lake and eventually to McAllister Lake. Along this route its waters are used for irrigation, recreation and wildlife preservation. It has cut a broad valley in which lies the city of Las Vegas. To the south of the city, the stream merges with the Pecos Arroyo (a tributary of the Gallinas), and their combined discharge has been responsible for cutting what is locally known as the Box Canyon, a deep, steep-sided valley incised into the Canadian Escarpment on the south flank of the Las Vegas Plateau. Due to the limited discharge of the stream and to the limited ground water resources in the Las Vegas area, the city's growth potential is severely restricted unless water can be imported or recycled.

The Gallinas River, as it flows south from the mouth of Gallinas Canyon cuts into an extensive pediment. Remnants of this pediment form the Camp Luna surface as well as the tops of buttes and mesas. The pediment gravel is 40 ft thick on one mesa north of Las Vegas. The pediment probably formed as a result of erosion of the Laramide Rocky Mountains (prior to their late Cenozoic rejuvenation). An attempt has been made to correlate the pediment's gravels with those of the Ogallala Formation. The pediment abuts against the Creston, a prominent north-south trending ridge on the west side of Las Vegas. The Creston is the surficial expression of the steeply dipping, western limb of the asymmetric Las Vegas syncline.

A Chinle Formation strike valley is located to the west of the Creston. Its width depends upon the dip of the Chinle beds. In the Romeroville area south of Las Vegas its dip is moderate and the valley is wide. At Montezuma (north of Las Vegas) the dip is vertical to overturned and the strike valley is very narrow. The ridge which forms the western boundary of the strike valley is capped by the Glorieta Sandstone and/or the Santa Rosa Sandstone.

STRATIGRAPHY AND PALEOGEOGRAPHY

The Gallinas River has exposed a sequence of Precambrian to Cenozoic rocks. The most complete exposure is at the mouth of the Gallinas River (Gallinas River water gap) and in the area immediately west. The only detailed studies of individual rocks units in the Las Vegas area are by Cathey (1973), on the Precambrian geology of the Montezuma area and by

Jacka and Brand (1972) and Bejnar and Lessard (this Guidebook) on the Dakota Group. Other rock units have received only cursory examination.

Precambrian

Thousands of feet of Precambrian rocks crop out in the Las Vegas area and adjacent regions (Baltz and Bachman, 1956; Bejnar, in press). They consist of metasediments as well as gneisses, schists, granites, migmatites, pegmatites, quartzites and amphibolites (Callender and others, this .Guidebook). The original sediments were probably deposited in a geosyncline or basin during the early accretional history of the North American continent. Laramide and later orogenic movements have been responsible for the uplift.

Cambrian, Ordovician and Silurian

The apparent absence of Cambrian, Ordovician and Silurian rocks in the Sangre de Cristo Mountains suggests that this area was part of the northeast trending Transcontinental Arch, a topographic high, during the time represented by these periods. If the area was periodically submerged during this time and sediments deposited on this structure, the rocks must have either been subsequently removed by erosion, are yet to be exposed by erosion, or if exposed have not as yet been identified.

Devonian and Mississippian

Espiritu Santo and Tererro Formations

Baltz (1972) reported the Devonian(?) Espiritu Santo Formation and the Mississippian Tererro Formation in this area. They are both calcareous formations and unconformably overlie Precambrian rocks. An ancient soil profile separates these units from the Precambrian in some parts of Gallinas Canyon. The formations represent the transgression of a sea across the Transcontinental Arch and are the first units to indicate the end of the early Paleozoic high in this area.

Pennsylvanian

Sandia Formation

The Pennsylvanian Sandia Formation unconformably overlies the Mississippian Tererro Formation. The junior author measured 172 ft of the combined Sandia-Devonian/Mississippian rocks at the mouth of the Gallinas Canyon. The top of the Sandia may be faulted out here because Baltz and Bachman (1956) measured 350 ft of the formation in the upper part of Gallinas Canyon. At the mouth of the canyon the unit is a predominantly gray marine limestone with some shale beds. Baltz (1972) described it as containing buff to brown sandstones and conglomeratic sandstones, interbedded sandy shales, shalt' limestones and a few thick limestones. Molds and casts of the plants *Lepidodendron, Sigillaria* and *Calamites* have been found in the sandy units a few miles west of the mouth of Gallinas Canyon. According to Baltz (1965), the sediments of the Sandia Formation in the Las Vegas area accumulated on an "unstable, mainly marine shelf (presumably of the Rowe-Mora basin) which accounts for its marine facies." Much clastic material was derived from a small, northwest trending anticline that lay to the west of Las Vegas. Active uplift and erosion caused the regression and the deposition of continental sediments and associated plant debris. Inactivity allowed a subsequent transgression. Rapidly subsiding portions of the Rowe-Mora basin lay to the north of the unstable shelf and 2,000 ft (Baltz and Bachman, 1956) of Sandia sediments were deposited in it. The Sandia becomes more clastic in the Mora River water gap and many of the sediments were deposited in channels, bays and swamps (Siemers, pers. comm.). Source areas included the northwest trending San Luis Uplift and the northeast trending Sierra Grande Uplift, both part of the Ancestral Rocky Mountains.

Madera Formation

The Madera Formation lies conformably and unconformably on the Sandia Formation and is **137** ft thick at the mouth of the Gallinas Canyon. Baltz and Bachman (1956) noted that its thickness is highly variable and Baltz (1972) measured as much as 2,000 ft of the unit. The thickness increases to over 3,000 ft northwest of Mora. **In** the Las Vegas area, where it is highly calcareous, it contains a rich, diversified marine brachiopod and coral fauna. Like the Sandia Formation, the clastic content increases towards the north, and Baltz (1965) has suggested that the Madera sediments were deposited on an unstable shelf. Siemers (personal comm.) proposed that in the Mora area Madera deposition occurred on a shallow marine shelf, delta plain and alluvial plain.

Permian

The Permian is represented by the Sangre de Cristo Formation, Yeso Formation, Glorieta Sandstone, San Andres Limestone and Bernal Formation. Baltz (1965) suggested that, northwest of Las Vegas, the lower part of the Sangre de Cristo Formation is of Pennsylvanian age (Missourian and Virgilian) age because the upper Madera grades into the lower part of the Sangre de Cristo. Bachman (1953) traced the Yeso (Leonardian) into the upper part of the Sangre de Cristo, north of Mora.

Sangre de Cristo Formation

The Sangre de Cristo Formation is poorly exposed at the mouth of Gallinas Canyon, but Baltz (1972) measured 800-1,300 ft in the mountains west of Las Vegas; and Bachman (1953), as much as 3,300 ft in the Mora River water gap, where its lithology is essentially the same as in the Las Vegas area. The formation consists predominantly of conglomeratic arkosic sandstones and varicolored shales, but also contains some generally unfossiliferous limy members. Although no specific paleoenvironmental setting has been specified, the sediment lithology suggests burial in a rapidly subsiding basin. Baltz (1965) suggested that the source of the sediments included various Ancestral Rocky Mountain elements, including the Sierra Grande uplift, Cimarron Arch, Wet Mountains and possibly the San Luis uplift.

Yeso Formation

The Yeso Formation lies conformably on the Sangre de Cristo Formation. The formation has been faulted out of the Gallinas River water gap area; however, Baltz (1972) measured 450 ft of the Yeso near the Las Vegas area and Bachman (1953) measured 350 ft in the Mora River water gap. The unit thins and grades into the upper part of the Sangre de Cristo at the ghost town of Lucero, 2 miles north of the Mora River water gap. The Yeso in the Las Vegas area is principally an "orange-red to red sandstone, siltstones and shale containing some tan sandstone and thin lenses of gray limestone and gypsum" (Baltz, 1972) whereas it is a "brownish-pink siltstone" in the Mora River water gap (Bachman, 1953). The Yeso Formation was deposited in a marine environment which was part of a north transgressing sea (Bachman, 1953; Baltz and Bachman, 1956).

Glorieta Sandstone

The Glorieta Sandstone lies conformably on the Yeso Formation. It is 93 ft thick in the Gallinas River water gap where a portion of the upper part of the unit has possibly been removed by faulting. Baltz and Bachman (1956) measured 220 ft of the unit at Kearny's Gap, which is located 1 mile south of Las Vegas, and Bachman (1953) measured 180 ft of the unit in the Mora River water gap. The formation is a yellowish-gray to grayish-orange, well rounded, fine- to medium-grained quartz arenite varying from well cemented to friable. It is both massive and bedded with well developed cross-stratification. The sediments of the Glorieta Sandstone were deposited in shallow marine waters and beaches.

San Andres Limestone

The San Andres Limestone, which lies conformably on the Glorieta Sandstone, is only 6 ft thick in the vicinity of the Gallinas River water gap and pinches out completely to the north. Baltz (1972) has dropped the use of the name in the area and his mapped unit may be equivalent to the San Andres Limestone. Baltz and Bachman (1956) suggested that the sediments of the formation were deposited in a marine environment.

Bernal Formation

The Bernal Formation, which lies conformably on the San Andres Limestone, is 118 ft thick at the mouth of the Gallinas Canyon and Bachman (1953) has measured a thickness of 150 ft at the Mora River water gap. In the Gallinas Canyon water gap it consists predominantly of alternating red arenite and lutites. Bachman (1953) suggested that its sediments were deposited in either marine or lagoonal conditions. It has been reported that an amateur collected terrestrial plant fossils at the Gallinas River water gap. The possibility of either a deltaic or a fluvial origin for the sediments of the Bernal in the Las Vegas area should be investigated. Baltz (1972) suggested that a few beds of the upper part of the unit may be Triassic in age.

Triassic

Santa Rosa Sandstone

The Santa Rosa Sandstone lies unconformably on the Bernal Formation and is approximately 323 ft thick in the Gallinas River water gap. It consists of alternating lutites, arenites and rudites. The arenites and lutites are grayish-red to yellowish-gray and the rudites olive-gray to yellowish-gray. The rudites contain well rounded quartz and limestone pebbles with black, carbonaceous, poorly indurated plant remains. Santa Rosa sediments were probably deposited in fluvial environments. Baltz (1965) believed that these sediments were derived from erosion of the San Luis and Wet Mountain uplifts and local positive structures derived from the deformation of Pennsylvanian and Permian strata.

Chinle Formation

The Chinle Formation lies conformably on the Santa Rosa Sandstone and is 1,202 ft thick in Gallinas Canyon, though the upper part of the unit may be missing due to faulting. Bachman measured 640 ft of the unit east of Ocate, 22 miles to the northeast of the Mora River water gap. In Gallinas Canyon, the unit consists of alternating red and reddish-brown to yellowish-gray lutites, arenites and rudites. Baltz (1972) divided the unit into three members in the Las Vegas area including a lower shale member, a middle sandstone member and an upper shale member. The Chinle includes four members, one tongue, and one lentil in the Ghost Ranch-San Ysidro area of north-central New Mexico (O'Sullivan, 1974). The sediments of the Chinle Formation were deposited in fluvial environments, including channels and floodplains, as indicated by fluvial cross-stratification, ancient channels and lithology of the members. There are extensive deposits of petrified wood in the sandstone member. Baltz (1965) believed that the sources of Chinle sediments were the same as for the Santa Rosa Sandstone.

Jurassic

Exeter Sandstone

The pale orange sandstone unit that unconformably overlies the Chinle Formation has been referred to as the Entrada, Ocate and Exeter Sandstone. It is 127 ft thick in Romeroville gap (6 miles south of Las Vegas) and 50 ft thick in the Ocate area. The unit is a fine-grained to medium-grained, feldspathic quartz arenite, which varies from thin to thick bedded with some cross laminations. The sediments of the formation were deposited in a lacustrine environment (including beach deposits) in the Las Vegas area; and in both lacustrine and eolian environments in northeast Arizona and west, north-central and northeast New Mexico; but Bachman (1953) named this unit the Ocate Sandstone for outcrops in the Ocate, New Mexico area. Baltz (1972), who used the name Ocate in Baltz and Bachman (1956), abandoned the term and has correlated the unit with the Entrada Sandstone of northeast Arizona and west to north-central New Mexico. Man kin (1972) preferred to drop both names in favor of the Exeter Sandstone, "in recognition of the nomenclatural priority established by Lee (1902) for a sandstone exposed in the Dry Cimarron Valley along the Canadian Escarpment in northeastern New Mexico." Assuming that this unit can be correlated from northeast Arizona to northeast New Mexico, we agree with Mankin (1972) that the term Exeter Sandstone should be used. We suggest that the unit be divided into two members, the lacustrine Ocate member and the eolian Entrada member. The Ocate member probably derived its sediments from the Entrada member as the lake(s) in which its sediments were being deposited expanded in size.

Todilto Limestone

The Todilto Limestone lies conformably on the Entrada Sandstone and is approximately 21 ft thick at Romeroville gap but pinches out 7 miles north of the mouth of Gallinas Canyon (Baltz and Bachman, 1956). It is a medium to dark gray, thin bedded, fetid limestone, containing limy siltstone and shale. The Todilto Limestone was deposited in the Ocate lake(s) in areas that were not being influenced by extensive terrestrial sedimentation. Mankin (1972) correlated it with a gypsiferous deposit in the Dry Cimarron Valley of northeastern New Mexico and Tanner (1974) correlated it with the Todilto Formation (limestone and gypsum) of west and north-central New Mexico. The unit contains very little gypsum in the Las Vegas area (Baltz, 1972).

Morrison Formation

The Morrison Formation lies both conformably and unconformably on the Todilto Limestone. It is approximately 415 ft thick in Romeroville gap and Bachman (1953) measured 150 ft near Ocate. The unit is divided into three members in the Las Vegas area. The lower member consists of alternating thin beds of claystone, siltstone, dolostone, limestone and quartz arenite, with bentonite and red jasper, and ancient channel deposits. The middle member contains thick quartz arenites alternating with mudstones and quartz rudites, with cross-stratification and channeling. The upper member is a series of alternating claystones, siltstones and arenites, containing channeling and cross-stratification. The Morrison Formation in the Las Vegas area is principally a fluvial unit deposited in channels, lakes and floodplains. The claystone, siltstone, dolostone and limestone beds could represent parts of ancient soil profiles.

Cretaceous

Dakota Group

The Dakota Group paraconformably overlies the Morrison Formation and is 105 ft thick at Romeroville gap and 121 ft thick at the Gallinas River water gap. Approximately 160 ft was measured at Coyote Creek, north of the Mora River water gap. The Dakota Group has been divided into three units in the Las Vegas area. The Lower Sandstone unit is a pale grayish-orange to very light gray, conglomeratic to fine-grained, cross-stratified, quartz arenite. The Middle Shale Unit consists of a silty, fine-grained quartz arenite and a black carbonaceous shale. The Upper Sandstone Unit is a fine- to medium-grained quartz arenite which is light to dark gray and may be characterized by carbonized wood fragments. The Lower Sandstone Unit was deposited on a piedmont plain; the Middle Shale Unit, on a swampy coastal plain; and the Upper Sandstone Unit, in a littoral, beach, lagoon and marine complex (Bejnar and Lessard, this Guidebook).

Graneros Shale

The Graneros Shale lies conformably on the Dakota Group and is approximately 225 ft thick immediately east of the mouth of the Gallinas Canyon. The basal part of the unit consists of a series of alternating thin fine sandstone, siltstones and shale beds. The amount of coarser clastic material decreases towards the top of the unit, which consists dominantly of limy shale beds that grade into the overlying Greenhorn Limestone. The sediments of the Graneros were deposited in a sea transgressing to the west. As the shoreline became further removed from the Las Vegas area, the sediments became finer.

Greenhorn Limestone

There is a gradational contact between the Greenhorn Limestone and the Graneros Shale. The Greenhorn Limestone is 50 ft thick in the Las Vegas area. The unit consists of alternating beds of gray limestone and calcareous shale. The sediments were deposited in the Cretaceous sea when the Dakota shoreline reached its furthermost western transgression. Eicher (1969) estimated that the depth of the Greenhorn sea in eastern Colorado (and presumably in northern New Mexico) was 2,000-3,000 ft. Lessard (1973), working with the Tununk Shale of central Utah, suggested an alternate depth of 300-600 ft and favored a depth in the lower part of the range.

Carfile Shale

The Carlile Shale conformably overlies the Greenhorn Limestone. The thickness of the unit is estimated to be 220 ft. The lower part of the unit is a dark gray shale, and sediments coarsen upward, becoming yellowish-brown, cross-laminated arenites containing shallow water fossils. Carlile sediments document the regression of the Greenhorn sea. As the shoreline in central Utah retreated eastward by prograding deltas, limy Greenhorn sediments were replaced by muds and finally by shallow, marine sands.

Niobrara Formation

Baltz (1972) measured 700 ft of the Niobrara Formation, northeast of the mouth of Gallinas Canyon. He described it as consisting of "gray shale, light brown weathering siltstone, and a few thin beds of limestone." There is no detailed information on its environment of deposition other than that marine fossils have been found in its lower part.

Cenozoic

The Cenozoic is represented by Gallinas River sediments and older pediment gravels. The pediment gravels are 50 ft thick.

STRUCTURAL EVOLUTION

The oldest evidence of structural elements in the Las Vegas area is the great thickness of the Precambrian metasediments, an indication that either a geosyncline or structural basin existed during this time. Cathey (1973) suggested that Precambrian rocks of Gallinas Canyon were "subjected to at least two structural-metamorphic events" which occurred 1.3 and 0.8 b.y. ago.

The Las Vegas area was part of a northeast trending, predominantly positive area, the Transcontinental Arch, from the Cambrian through the Silurian. The arch weakened during the Devonian and the Mississippian and the area that it had occupied became part of the north trending Rowe-Mora structural basin during the Pennsylvanian. The Sierra Grande uplift lay to the east of the area and the San Luis uplift, along with some smaller anticlines were situated to the west (Baltz, 1965). The area continued as a basin, collecting sediment from the Pennsylvanian through most of the Cretaceous until the Laramide orogeny. At this time the Las Vegas sub-basin of the Raton basin as well as the adjoining Sangre de Cristo uplift, were formed by compressional forces which originated in the west and were directed to the east. The Raton basin is a north trending synclinal structure which extends into Colorado; its southern part is the Las Vegas sub-basin, which is separated from the northern Raton basin by the southeast trending Cimarron arch that passes beneath Springer, New Mexico (Baltz, 1965). In the Las Vegas area, the Las Vegas sub-basin is asymmetrical with an overturned western limb.

With continued pressure from the west, the western limb of the syncline was broken by east directed, high angle reverse faults. They have been mapped at the mouth of Gallinas Canyon by Baltz (1972), Bejnar (in press) and Brown (pers. commun.). Each author has a different interpretation of the faulting. Bejnar (in press) offers the following information about the thrust faulting in Gallinas Canyon:

> "these faults have stratigraphic displacements ranging from a few hundred feet on the two easterly ones to over 2,000 ft along the west trending Peterson Reservoir fault which passes through

Peterson Reservoir. (Peterson and Bradner Reservoirs are located immediately to the southwest of the mouth of Gallinas Canyon and to the south and above the village of Montezuma and the abandoned Montezuma Hotel.) Along the Peterson Reservoir thrust fault, the intense displacement has broken the rocks into a 300 ft wide breccia which contains large angular blocks of the adjacent beds. The erosion of this brecciated zone has produced the wide valley now occupied by Peterson Reservoir. The Bradner Reservoir valley was formed as a result of erosion along the Sebastion Canyon and Creston reverse faults."

The reverse faulting was followed (and possibly accompanied) by smaller scale normal faulting. There is also evidence of the presence of northwest trending strike-slip or oblique-slip faulting west of the mouth of Gallinas Canyon. "They are more recent than the reverse faults because they displace the reverse faults" (Bejnar, in press). The presence of badly eroded pediment in the Las Vegas area suggests that the mountains produced by the Laramide orogeny underwent extensive erosion during and following the orogeny. The youthful nature of the Gallinas River valley suggests that the area was rejuvenated in late Cenozoic time. The rejuvenation probably began in the Miocene and has continued since that time.

MONTEZUMA HOT SPRINGS

One of the most interesting geomorphic features in the Las Vegas area are the hot springs located between the villages of Montezuma and Hot Springs west of the mouth of Gallinas Canyon. The hot springs begin approximately 0.5 mile to the west of Montezuma and extend for 1,500 ft along the lower south side of the Gallinas River valley. At least 20 of the hot springs were developed in the 1880's in conjunction with the building of the historic Montezuma Hotel resort (Rienerts, 1966), but most of these have fallen into disuse. Only one bath house is in use at present.

The entire slope over which the hot springs flow is a bog that is thickly covered with perennially green grass. Standing pools of hot water that never freeze have formed in level areas. They contain thick accumulations of algae and bacteria and are inhabited by frogs, snails and other fauna. The average temperature of the springs water is 56°C (Bejnar, 1967). Its chemical composition in ppm: Ca=7.9, Mg=2.2; Na=168, K=168; HCO_3=79; CO_3=15; SO_4=50; Cl=157; F=20; and NO_3=0.1 (Summers, 1965). The concentration of total solids is 534 ppm.

The origin of the hot springs may be the snow and rain of the Sangre de Cristo Mountains which infiltrates porous and permeable rocks, traveling east to the mouth of Gallinas Canyon. As these subsurface waters move east, they percolate to considerable depths until they encounter the wide Peterson Reservoir fault breccia zone east of the hot springs. The breccia contains fine clay-sized sediments, which makes it impermeable to the continued eastward movement of the ground water. Because this brecciated "dam" stretches at least 3 miles to the north and 5 miles to the south of Montezuma, it retains large volumes of water. The Montezuma area is topographically the lowest site to the west of this "dam"; therefore, the backed-up water moves to the surface along joints and faults in the Precambrian rocks.

There are three possible sources of heat. The first is the geothermal gradient. In sedimentary rocks, the geothermal gradient is about 1°C/100 ft, but in faulted and folded areas the gradient may be 10 times as great. Because the area west of the hot springs has been folded and faulted, there is enough geothermal heat to increase the local geothermal gradient to at least 4°C/100 ft. This gradient and an average temperature of 12°C for the surface water requires the water to percolate to a depth of only 1,000 ft to reach the average hot springs temperature of 56°C. The elevation of the hot springs is 6,740 ft whereas the mountains immediately to the west rise to elevations of 10,470 ft. The difference of 3,730 ft would permit the ground water to percolate to a depth of at least 1,100 ft and still have a sufficient gradient to migrate east to the lower elevations of the hot springs area.

The second possible source is heat associated with relatively recently uplifted basement rock. In the Montezuma area this would involve heat from Precambrian rocks that crop out to the west. They were initially uplifted during the Laramide orogeny with additional movement during the Miocene and post-Miocene rejuvenation of the Rocky Mountains. Subsurface water passing through or adjacent to these rocks could have its temperature raised to 56°C.

The third heat source is related to the volcanism that has occurred in northeastern New Mexico during the last 8 m.y. Although the nearest volcanic cone is 25 miles northeast of Montezuma, several lamprophyre dikes cut Cretaceous rocks 6 miles northeast of the hot springs. Magma may extend under and to the west of the hot springs and contribute heat to subsurface water that surfaces at the hot springs.

It is unlikely that the hot springs at Montezuma can be developed as a geothermal resource. A geothermal area must have a large volume of steam, a temperature of at least 400°F and a pressure of 150 psi (Summers, 1968). These conditions are not present at the Montezuma hot springs.

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