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THE MOUNT PRINCETON GEOTHERMAL AREA, CHAFFEE COUNTY, COLORADO

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Abstract

The Mount Princeton geothermal area is on the west side of the upper Arkansas Valley near Buena Vista, Colorado, along the northern extension of the Rio Grande rift. The area underwent a complex period of Tertiary igneous activity which terminated with the intrusion of the Raspberry Gulch Rhyolite about 22 m.y. age. Faulting, associated with the Rio Grande rift, began in the Miocene and continued to within the last 30,000 years. Surficial thermal manifestations are characterized by zeolitic alteration, which covers approximately 64 km^2 (34.7 sq. mi.), and many thermal springs and wells which have a maximum temperature of 85°C. Chemical analysis of the thermal waters indicate minimum subsurface temperatures of approximately 125°C. Deep circulation of meteoric water, in a zone of anomalous heat flow associated with the Rio Grande rift, may be the heat source for the thermal features of the area.

Location

The Mount Princeton geothermal area is on the west side of the upper Arkansas Valley between the towns of Buena Vista and Salida in Chaffee County, Colorado (fig. 1). The area is along the western flank of the northern extension of the Rio Grande rift zone and includes a portion of Collegiate Peaks area of the Sawatch Range. The area of thermal manifestations is along the eastern flank of the range and is roughly defined by Cottonwood Creek on the north and Brown's Creek on the south. The thermal manifestations are most pronounced along Chalk Creek where the jagged, zeolitized, white Chalk Cliffs rise approximately 500 meters (1,600 ft.) above the Mount Princeton Hot Springs.

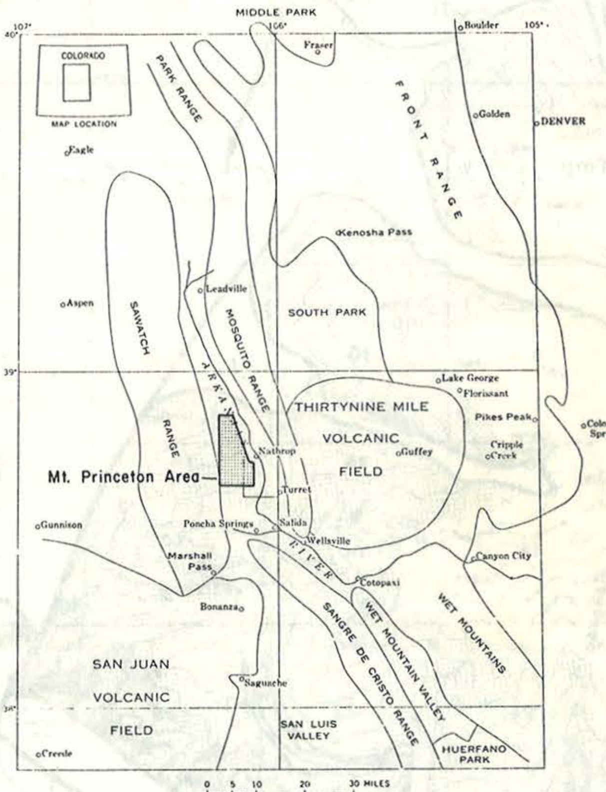

Figure 1. Index map of central Colorado, showing the location of the Mount Princeton area (after Van Alstine and Cox, 1969).

Geology

The oldest rock units in the Mount Princeton area are Precambrian metamorphic and igneous rocks (fig. 2). Paleozoic and

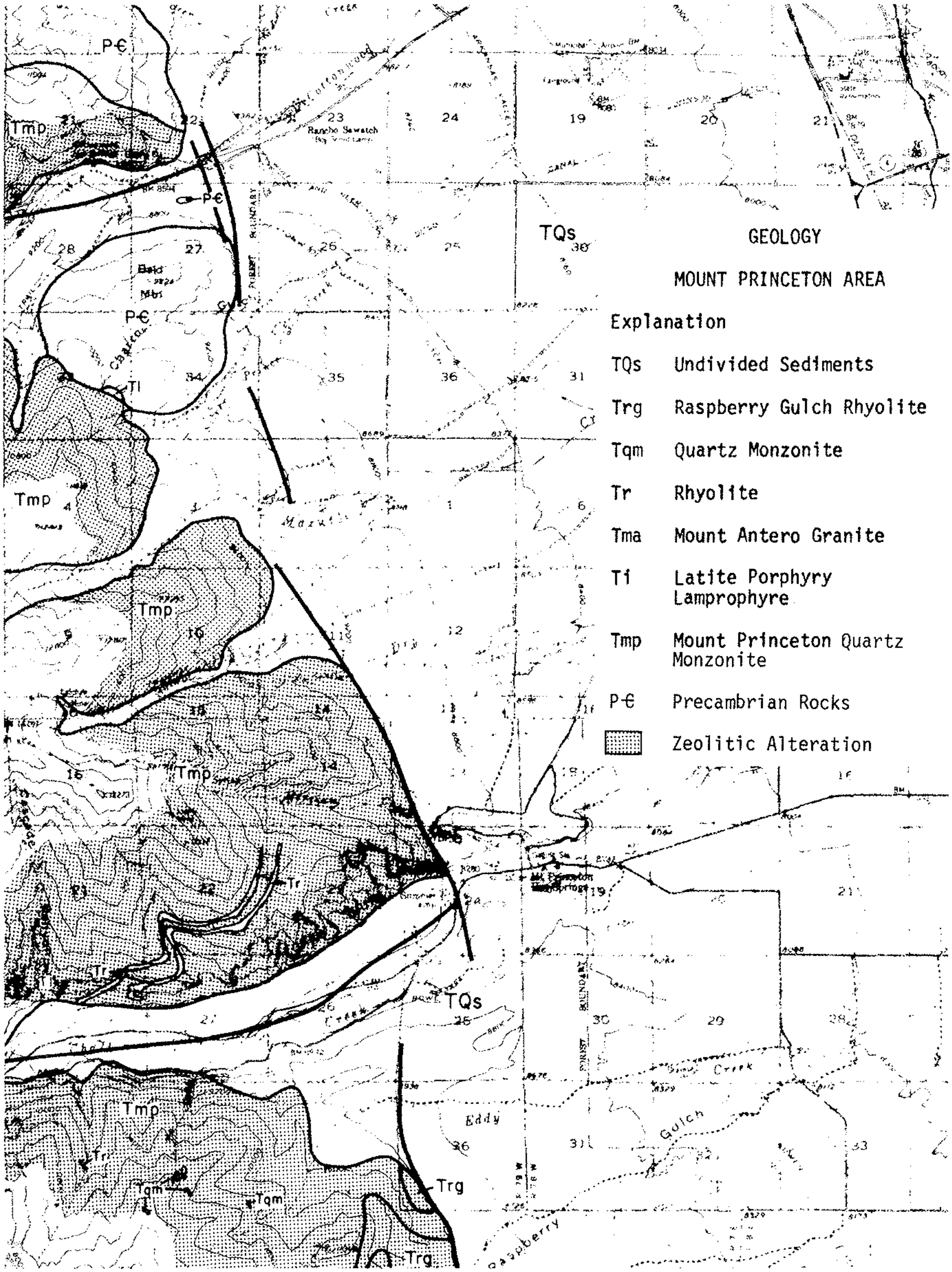

Figure 2. Geologic Map of the Mount Princeton area.

Mesozoic rocks are not present in the area although a thick section of sediments was deposited over the area during this time interval. Apparently, these rocks were removed by the erosional event which followed uplift and the formation of the Sawatch anticlinal structure during the Laramide orogeny.

Tertiary intrusion in the general area may have commenced as early as the Eocene. However, the earliest intrusion in the Mount Princeton area began with the emplacement of the Mount Princeton Quartz Monzonite, dated at 36 $\pm$ 2 m.y. This intrusion of batholithic proportion is the most prevalent rock in the area. It was followed by lamprophyre and latite porphyry dikes, and the Mount Antero granite stock, dated at 30.8 $\pm$ 1.1 m.y.. Afterwards, rhyolite dikes, dated at 25.4 $\pm$ 1 m.y. and small quartz monzonite bodies, dated at 24 $\pm$ 1 m.y., were intruded into the Mount Princeton batholith. These were followed by the emplacement, north of the Mount Antero granite near the headwaters of Raspberry Gulch, of the Raspberry Gulch rhyolite (22 $\pm$ 1 m.y.), the youngest igneous rock in the area.

Volcanic rocks of Oligocene age are common in the Mosquito Range to the east of the Mount Princeton area, and two possible rhyolitic vents, dated at 28-29 m.y., form conspicuous domes across from the mouth of Cottonwood Creek.

Regional uplift and the development of the Rio Grande rift began in the early Miocene. Rifting continued throughout the Miocene and Pliocene, with deposition of basin-fill sediments, and throughout the Pleistocene with the development of pediment gravels and glacial debris. Faulting along the rift is believed to have continued to at least within the last 30,000 years.

Structure

The Rio Grande rift in the Mount Princeton area is characterized by a series of "en echelon" normal faults on the west side of the valley and parallel normal faults on the east side which give the valley a graben structure. Faults on the east side of the valley drop the basin in a series of steps; whereas, the fault system on the west side of the valley consists of a relatively narrow fault zone characterized by a single large displacement. Parallel north-trending faults on the east side appear to have about 300 m. (1,000 ft.) of displacement each (Van Alstine 1969 and Knepper 1974). Zhody and others (1971) believe, on the basis of resistivity measurements, that there are 1,400 m. (4,600 ft.) of valley fill deposits in the graben immediately south of Buena Vista. Considering the topographic relief present in the Sawatch Range, the total displacement on the west side fault system may be as much as 3,000 m. (10,000 ft.).

The gravity data do not show a sharp gradient of the west side of the graben, possibly due to the Mount Princeton intrusive center.

Two major cross faults are projected along Cottonwood and Chalk Creeks. Evidence for the faults, which are covered, are hot spring and alteration pattern location, the linear nature of the two valleys, and the nonalignment of the mountain front at Chalk Creek.

Alteration

Hot spring systems at the base of Mount Princeton have produced an extensive zone of alteration. The hydrothermal system, which is active at present, is probably producing a similar alteration assemblage at depth. Alteration is characterized by the calcium zeolite, leonhardite (Sharp 1970), which may grade into laumontite below surface. In addition to leonhardite; chlorite, illite, epidote, calcite and fluorite are present as alteration minerals.

Zeolitic alteration covers more than 64 km^2 (25 sq. mi.) and has a vertical exposure of 1,000 meters (3,283 ft.). Alteration is strongest at Chalk Cliffs, above Mount Princeton Hot Springs, but a weaker alteration center is concentrated around Cottonwood Hot Springs. The cliffs get their white color from leonhardite and clays which fill fractures to such a degree that they impart a bright white hue to the normally grey host rock.

The zeolitic alteration is strongest in the Mount Princeton Quartz Monzonite but it is also present in Precambrian gneiss, Mount Antero Granite, aplite, rhyolite, and Raspberry Gulch Rhyolite.

Unweathered Mount Princeton Quartz Monzonite shows weak chloritization of biotite and sericitization of feldspars (Limbach 1975). This weak, widespread alteration is probably related to the crystallization of the Mount Princeton intrusive. Zeolitic alteration, related to more recent hydrothermal activity, converts biotite to chlorite; hornblende to calcite and epidote; orthoclase to sericite; and plagioclase to albite. Leonhardite is confined to fracture openings and does not replace other minerals. The calcium for the zeolite comes from the breakdown of hornblende and plagioclase. Chemical analysis of the strongly altered Mount Princeton Quartz Monzonite shows little change from the unaltered quartz monzonite, except for slight additions in iron and water.

The zeolite alteration assemblage forms at high activities of H_2O relative to CO_2. These conditions prevail where hot water has a pH of 8 to 9 (White and Sigvaldson 1963), which fits the hot springs in the Mount Princeton area. Sharp (1970) has suggested that the zeolitic alteration assemblage present at Chalk Cliffs formed within a range of temperatures of 145-220° C and depths of 150-2,000 m. (4,925-6,566.6 ft.) below the surface, based on comparison with active geothermal areas in New Zealand and Iceland.

Thermal Springs

The thermal springs in the Mount Princeton area have been described by George and others, 1920; Lewis, 1966; Sharp, 1970; and Pearl, 1972. Most of the hot wells are used for heating homes, greenhouses, bathing and drinking (fig. 3).

The area contains at least six thermal wells and two thermal springs (Table 1), including Hortense Hot Spring, reportedly the hottest spring in Colorado. Many other thermal seeps issue directly into Chalk Creek, and cannot be counted. Several low-pressure, steam fumeroles are present, about 100 m. (368.3 ft.) to the west of Hortense Hot Spring, in a talus slope at the base of the Chalk Cliffs. The approximate heat discharge for the thermal features of the area, computed as the product of the volume rate and enthalpy of the water in excess of ambient temperature, is seen in Table 1. All the thermal features combine to produce 4×10^6 cal/sec. or enough heat to supply approximately 200 average sized houses.

Table 1. Thermal features of the Mount Princeton area.

Sample Name	T$^\circ$C	Flow l/m	Heat Discharge cal/sec.
Hortense Hot Spring	85	38	4.9×10^4
Younglife Hot Well East	85	379	4.9×10^5
Younglife Hot Well West	67	379	3.7×10^5
Greenhouse Hot Well	68	379	3.8×10^5
Chalk Creek Greenhouse Hot Well	65	1892	1.8×10^6
Jump Steady Hot Well	59	568	5.0×10^5
Mt. Princeton Hot Spring	56	265	2.1×10^5
Deer Ranch Hot Well	38.5	379	1.9×10^5
			4.0×10^6 cal/sec.
			1.6×10^4 BTU/sec.

Analyses of thermal and non-thermal waters of the Mount Princeton area are given in Table 2. Non-thermal waters of the Mount Princeton area generally contain less than 150 mg/l of dissolved solids. Water pH is generally neutral to slightly basic. Bicarbonate is the principle ion followed by silica, calcium, sodium and magnesium. Cold waters contain an average of 22 mg/l of silica. Ice Pond Cold Spring, about 1 mi. (1.6 km) to the northeast of Buena Vista, was chosen to represent background water chemistry.

Thermal waters exhibit basic to very basic pH. Four types of thermal water are recognized:

1. Sulfate-sodium waters, with less than 12 mg/l chloride, may represent steam condensate that has reached equilibrium with quartz monzonite of the Mount Princeton batholith. The

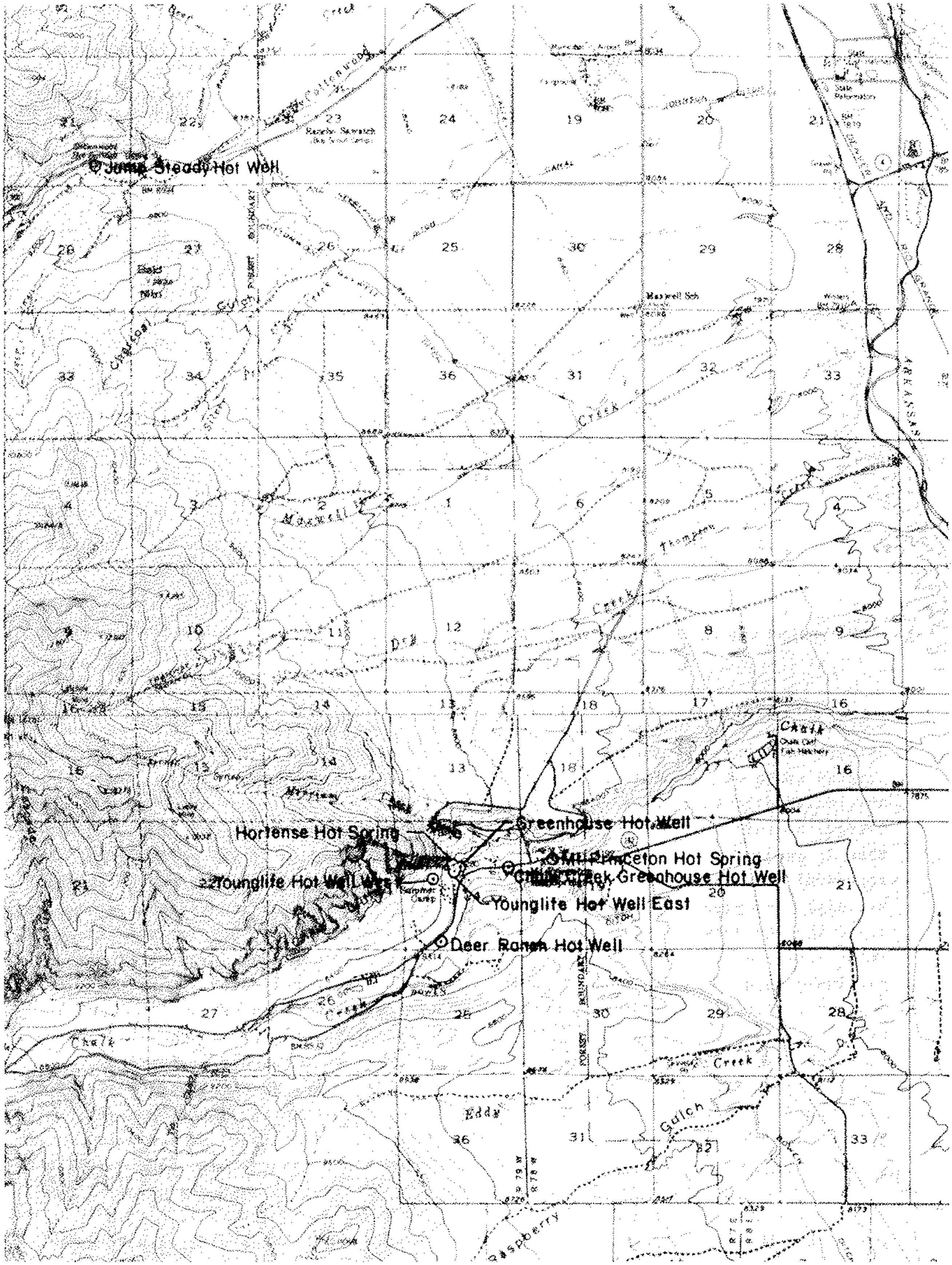

Figure 3. Location of the thermal features of the Mount Princeton area.

Table 2. Chemical analysis of the thermal features of the Mount Princeton area. Units are mg/l unless otherwise noted.

	Hortense Hot Spring	Younglife Hot Well East	Greenhouse Hot Well	Younglife Hot Well West	Chalk Creek Greenhouse Hot Well	Jump Steady Hot Well	Mt. Princeton Hot Spring	Deer Ranch Hot Well	Ice Pond Cold Spring
pH	9.6	9.2	9.1	9.1	8.8	9.2	8.6	8.8	7.6
Cl	8.8	11	6.6	2.2	6.6	28	5.5	4.4	3.0
F	16	15	13	9.3	10	14	9.4	6.2	0.2
HCO_3	46	43	79	44	52	31	59	64	68
CO_3	16	18	20	10	0	24	0	0	0
SO_4	100	90	80	60	70	110	60	40	6
SiO_2	85	80	75	75	65	60	60	45	25
Na	100	80	80	60	60	110	50	40	7
K	4.0	2	2	2	2	2	2	2	1
Ca	15.0	12	7	17	10	50	20	20	18
Mg	0.1	<0.1	<0.1	0.4	0.1	0.3	0.4	0.9	4
Li	0.2	0.1	NA	0.1	NA	0.2	0.1	0.1	NA
B	<1.0	<1.0	<1.0	<1.0	<1.0	<1.0	<1.0	<1.0	<1.0
NH_3	0.4	0.3	<0.1	<0.1	NA	<0.1	<0.1	<0.1	<0.1
TDS	392	351	363	280	276	430	266	223	132
T°C	85	85	68	67	65	59	56	38.5	9
Flow (gpm)	10	100	100	100	500	150	70	100	75
$TSiO_2$°C	125	125	122	122	115	110	111	97	72
TNa/K°C	97	74	64	84	84	45	97	115	230*
TNa-K-Ca°C	75	55	67	47	57	34	43	41	12
Cl/SO_4	0.2	0.3	0.2	0.1	0.3	0.7	0.3	0.3	1.4
Cl/F	0.3	0.4	0.3	0.1	0.1	1.1	0.3	0.4	8.0
Cl/HCO_3+CO_3	0.5	0.6	0.2	0.1	0.4	1.7	0.3	0.2	0.2
Resistivity ohm-m	21.4	24.2	NA	NA	NA	19.2	32.6	36.2	63.7

NA = not analysed
+ = Does not represent true subsurface conditions, i.e. $\frac{\sqrt{Ca}}{Na} > 1$

Hortense and Mt. Princeton Hot Springs, and three hot wells, located south of the Chalk Cliffs, are included in this category.

2. Sulfate-sodium waters with greater than 25 mg/l of chloride probably represent low temperature, low salinity, hot water systems, derived through deep circulation. These waters are saturated with calcium carbonate minerals and deposit varying amounts of travertine. The Jump Steady Hot Well (Cottonwood Hot Spring) characterizes this category.
3. Bicarbonate-sodium waters with less than 7 mg/l of chloride may represent dilutions of sulfate-sodium waters by bicarbonate rich groundwaters. Greenhouse Hot Well and Deer Ranch Hot Well represent this group.
4. Groundwater, as is found at the Ice Pond Cold Spring, is generally rich in bicarbonate.

Chemical analyses of hot spring systems may be used to estimate subsurface temperatures of active geothermal areas. The assumptions made in applying geochemical indicators are summarized by Fournier, White, and Truesdell (1974); Fournier and Truesdell (1973) suggest criteria for selecting the most probable temperature. These techniques, used in determining the subsurface equilibrium temperatures given in Table 2, indicate temperatures on the order of 125°C.

Waters in the Mount Princeton area were analysed for oxygen-18, deuterium, and tritium. In the analysis of the data, deuterium and oxygen-18 have been normalized, relative to standard mean ocean water (SMOW), and are noted as δD and δO^{18}.

Figure 4 shows the variation between δD and δO^{18}, relative to SMOW, for several

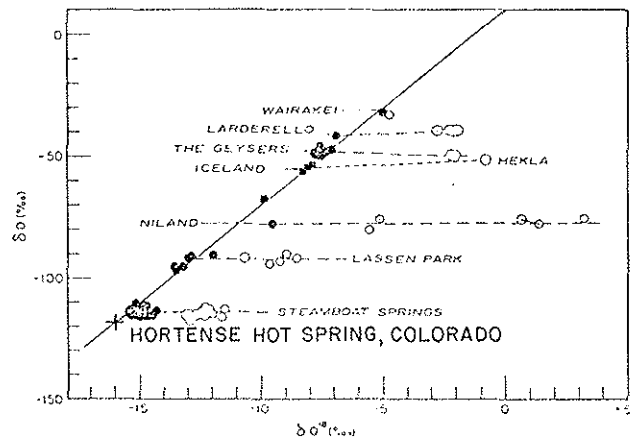

Figure 4. Observed isotopic variations in near-neutral chloride type geothermal waters and in geothermal steam. Solid points are local meteoric waters, or slightly heated near-surface groundwaters. Open circles are hot springs or geothermal water, crinkled circles are high temperature, high pressure, geothermal steam (after Craig 1963).

geothermal areas. The straight line represents the almost world-wide slope for meteoric waters plotted in this way. Deuterium concentrations are constant and equal to local meteoric water. On the other hand, O^{18} concentrations show a characteristic enrichment or shift. The O^{18} shift is due to an isotopic oxygen exchange between groundwater and carbonates and silicates in the rocks. Silicate and carbonate rocks contain O^{18}, ranging from +6 to +30 per mil greater than SMOW. Deuterium generally does not vary from the meteoric concentration because rocks contain negligible hydrogen or deuterium. A strong shift in O^{18} implies a long storage time and/or a large reservoir capacity. A very small shift implies one of two situations: first, temperature-pressure conditions are too low to allow waters to exchange O^{18} with rocks within a relatively short time period, and second, descending meteoric waters are heated and rise so quickly that insufficient time is available for an O^{18} exchange to occur. As shown on fig. 4, Niland waters, which have mingled over a long period of time with carbonate-rich Colorado River sediments, show the greatest shift. On the other end of the scale, Wairakei shows negligible shift which implies that waters descend quickly, stay in storage for only a short time, and then ascend.

Figure 5 is a plot of δD-δO^{18} for selected waters of the Mt. Princeton area. The hot waters show no apparent O^{18} shift. This implies that these hot waters have been in residence with the reservoir rocks only a short time and have not exchanged oxygen isotopes as at Niland.

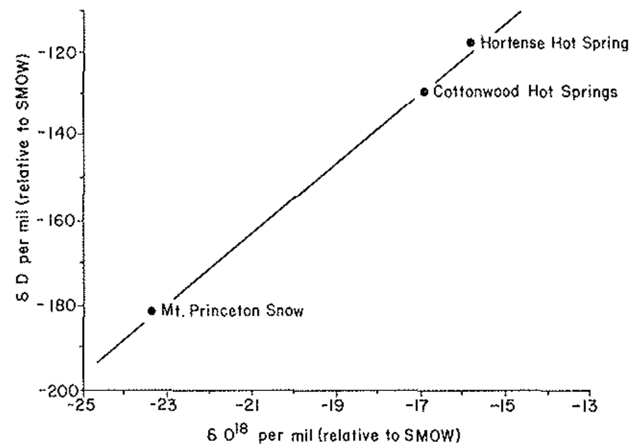

Figure 5. δD versus δO^{18} for waters of the Mount Princeton area.

Tritium analyses indicate an age of 20 to 51 years for Hortense Hot Spring, and 21 to 56 years for the Jump Steady Hot Well. These ages are in good agreement with the suggested youth of the same waters by O^{18} analysis.

Heat Source

The youngest igneous rock in the Mount Princeton area is Raspberry Gulch Rhyolite 22 m.y. old - too old to be the heat source for the Mount Princeton area hot springs - and it is doubtful that any younger igneous intrusions are present at shallow depths. The youngest rocks in the northern part of the Rio Grande rift system are 4.5 - 3.6 m.y. - old olivine tholeiitic basalts in the southern portion of the San Luis Valley. According to Lipman (1969), these basalts were derived from depths of 15-20 km (9.3 - 12.4 mi.). This suggests that a portion of the rift is underlain by an upward protrusion of hot, mantle rocks that forms a zone of abnormally high heat flow. Heat flow measurements along the rift show values of two to three times the average crustal value (Gross 1974). White (1957) states that meteoric water in fractured rocks can circulate to depths of 3,000 m. (10,000 ft.) and the active faulting within the past 30,000 years could have permitted the fracture system at Mount Princeton to remain open to considerable depths. Circulation of meteoric waters to these depths would be more than adequate to produce the hot springs at Mount Princeton and the indicated subsurface temperatures.

Although igneous intrusion into faults of the rift system cannot be completely ruled out, the most likely heat source for the thermal manifestations in the Mount Princeton area is the abnormally high geothermal gradient associated with the Rio Grande rift.

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MT Princeton Chapin (1971)

southern Rocky Mountains were beveled by an erosion surface of moderately low relief in the Late Eocene; thus the disappearance of great thicknesses of Oligocene volcanic rocks across the rift must be ascribed to Late Tertiary uplift and erosion rather than to nondeposition. Assuming an average elevation of 2,000 feet for the erosion surface in Late Eocene time, minimum uplift was approximately 5,000 to 12,000 feet during Late Tertiary time for ranges east of the rift and minimum subsidence was 4,000 to 24,000 feet within the rift. Numerous fault scarps cutting alluvial fans and Pleistocene surfaces indicate that differential movement is continuing.

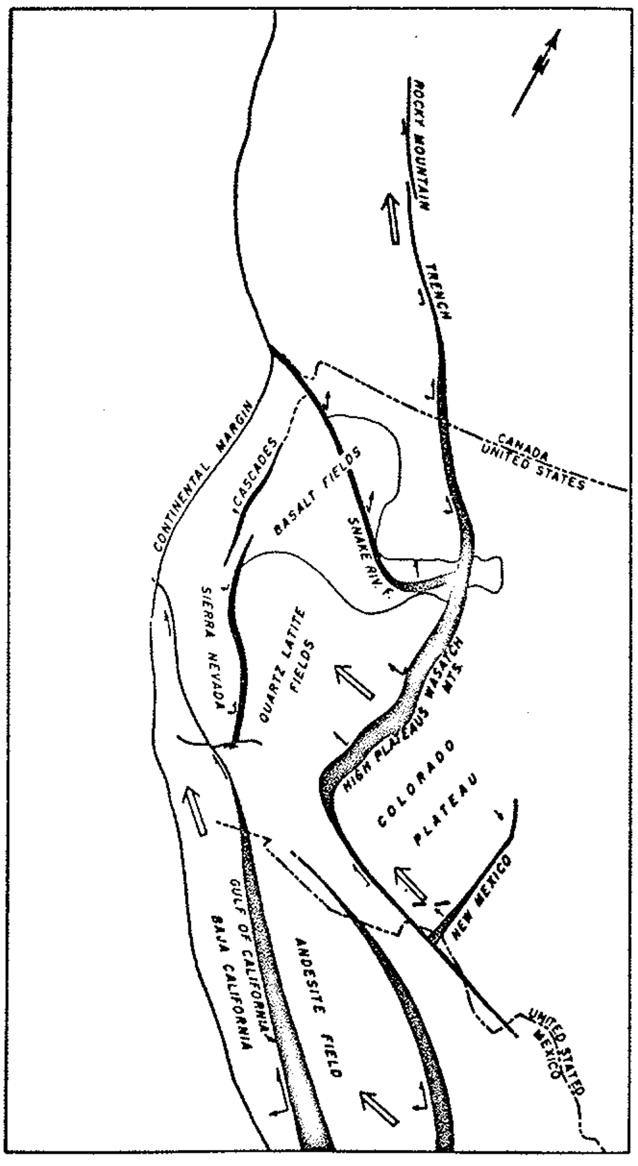

FIGURE 3.

Diagrammatic map exploring the concept of extension and drift affecting western North America. Black lines represent amount of expansion as if localized along a few separations. Small arrows represent apparent vectors of movement; large arrows the apparent resultant direction of the movement. From Eardley (1962, p. 510) with slight modifications along the Rio Grande rift.

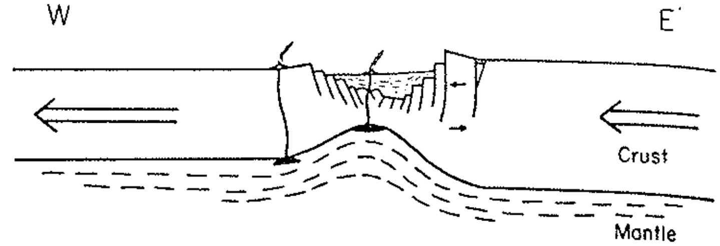

FIGURE 4.

Hypothetical cross-section of the Rio Grande rift. The large arrows indicate direction and relative rate of drift of continental plates. The small arrows indicate a force couple acting on the east shoulder of the rift. Freund (1965, p. 340) has experimentally produced a similar, but symmetric, model of the rifting and "necking" of sand above a convection current in a heavy fluid substratum.

Synthesis of the above observations suggests the following model (see figs. 3 and 4): (1) the continental plate west of the rift is drifting faster than the continental interior (mantle convection may be pulling it over the East Pacific Rise in a "conveyor belt" manner similar to that suggested by Cook, 1962); resultant crustal attenuation formed the Basin and Range province and is splitting the Colorado Plateau block away from the interior; (2) the east side of the rift developed greater structural relief due to riding up of thicker crust onto an upward bulge of mantle material beneath the rift; (3) the west side of the rift is relatively subdued due to crustal stretching accompanied by abundant normal faulting and a tendency to pull the crust away from the mantle bulge beneath the rift; (4) stretching and normal faulting along the west side relieves subcrustal pressure and provides avenues for ascent of magmas and hydrothermal solutions; (5) longitudinal faults along the east side are relatively tight and unconducive to magmatism; westward drift of the interior block against the mantle bulge tends to rotate the fault planes to a near vertical position and may change the sense of movement from normal to reverse; (6) northwestward drift of the Colorado Plateau as suggested by Eardley (1962) causes a slight clockwise rotation against the north end of the rift which tends to keep it tight and relatively free of volcanism; this may also explain the unusually high upthrusting of the Sangre de Cristo horst along the east side of the San Luis Valley; (7) continued widening of the Rio Grande rift in New Mexico appears to be accelerating volcanism and may cause the rift to evolve into a lava-filled trough similar to the Snake River rift.

THE UPPER ARKANSAS GRABEN AND PROBLEMS OF THE NORTH END

North of the San Luis Basin of Colorado, a narrow, north-tapering, sharply defined trough extends for at least 60 miles to the continental divide north of Leadville (fig. 2). That this basin is a graben with a tectonic style similar the Basin and Range province has been recognized by workers in the Leadville area for many years. Emmons, Irving, and Loughlin (1927, p. 97) recognized that Lower Paleozoic strata in the Mosquito Range had undergone at least 8,000 feet of post-ore uplift relative to the same strata in the upper Arkansas Valley at Leadville. They discussed the relative merits of subsidence versus uplift and spoke of the similarity of this faulting to that in the Great Basin. Their cross-sections (pl. 12) show the same progressive stepping down of strata towards the valley axis by numerous longitudinal faults as is characteristic all along the rift. Tweto (1948, pl. 7) showed this structure very well on a cross-section of the upper Arkansas Valley at Leadville and labeled the valley a graben. Both sets of cross-sections show a progressive steepening of the fault planes towards the east boundary of the graben, which culminates in upthrusting of the crest of the Mosquito Range along reverse faults.

Gableman (1952, p. 1608-1609), in a very perceptive discussion based largely on geomorphology, projected the graben structure of the San Luis Valley northward up the Arkansas Valley to Leadville. In 1960, Van Alstine and Lewis (p. B245) described Early Pliocene fossils from a 500-foot section of alluvial fill exposed near Salida. The following year, Tweto (1961, p. B133) named the alluvial fill of the upper Arkansas Valley the Dry Union Formation for exposures near Leadville and showed their continuous distribution along the valley. He also presented evidence for recurrent faulting of unconsolidated sediments continuing into Holocene time; this phenomenon is a nearly universal characteristic of the Rio Grande rift. Chapin and Epis (1964, p. 158) presented evidence for downfaulting of Oligocene ash-flows by as much as 2,100 feet in the Browns Canyon area north of Salida. But the real breakthrough came in 1968 when R. E. Van Alstine (p. C158) demonstrated that the Arkansas and San Luis Valleys are connected by a structural trough containing Late Tertiary sediments. The trough crosses Poncha Pass west of U.S. Highway 285 and is not obvious from that route. In a later paper Van Alstine (1970, p. B46) extended the age of the Dry Union Formation to Late Miocene on the basis of vertebrate fossils identified by G. E. Lewis. On the basis of similarities in structural style, age of alluvial fill, alignment, and physical continuity, the upper Arkansas graben is clearly part of the Rio Grande rift.

How far the Rio Grande rift extends beyond Leadville is uncertain. Geomorphically, it appears to end at about the Continental Divide; however, as Tweto (1968, p. 566) has recently pointed out, scattered *en echelon* north-trending faults extend northward to near the Wyoming line and may be related to the rift. Kelley (1956, 1970) has postulated that the Rio Grande depression, the Colorado Parks, and the Laramie Basin are linked in a right-lateral en echelon system of intermontane troughs. However, the intermontane parks of Colorado are Laramide basins distinct in structural style and age of sedimentary deposits from the post-Laramide grabens of the rift (see summary of Tweto, 1968, p. 563, 566, 567). The contrast is especially sharp between South Park and the adjacent upper Arkansas Valley. South Park is a Laramide basin which received at least 1,000 feet of andesitic extrusives of Paleocene age (57 m.y.) and 8,000 feet of arkosic conglomerates prior to folding and thrusting in the Late Paleocene or Early Eocene (Sawatzky, 1964, 1968). During the Late Eocene, South Park was beveled by a surface of low relief onto which Oligocene eruptive rocks and volcanic detritus from the adjacent Thirtynine Mile field (Epis and Chapin, 1968, p. 56) were emplaced. The only Late Tertiary sediments of appreciable thickness occur in the Antero syncline in the extreme southwest corner of the park. This small basin developed synchronously with the Rio Grande rift as a small subsidiary downwarp along the inflection line of its uplifted eastern shoulder and will be discussed in the following section. In contrast, the area now occupied by the upper Arkansas Basin was situated high on the flanks of the Sawatch anticline until the Miocene when subsidence accompanying regional uplift split the anticline to form a narrow, highly-elongate graben. Streams flowed eastward off the Sawatch anticline in the Late Eocene and were filled with ash-flow tuffs in the Oligocene; these paleovalleys and their volcanic rocks have been tilted to the west and downfaulted over 2,000 feet into the upper Arkansas graben (Chapin, Epis, and Lowell, 1970; Lowell, 1969, and this guidebook). The upper Arkansas Valley then received 500 to 2,000 feet of Late Tertiary sediments which are very similar to those of the Santa Fe Group in New Mexico. Thus, the upper Arkansas graben is the northward extension of the Rio Grande rift, not the Colorado Parks.

INFLECTION BASINS

The sharply uplifted and outward-tilted shoulders of the rift are generally 10 to 30 miles in width beyond which the structural slope is more gentle. The change in slope is rather sharp and marked by an inflection line along which numerous faults, small folds, and local downwarps occur. These smaller structures parallel the rift and were formed contemporaneously with it. They appear to be more abundant east of the rift, probably because of the greater uplift and tilting on that side. Four local downwarps, whose development can be demonstrated to be synchronous with the rift, are the Antero, Pleasant Valley, Moreno, and Estancia basins; others probably exist.

The Antero Basin

The Antero Basin of southwestern South Park is a north-to-northwest-trending syncline, which parallels the upper Arkansas graben about 10 miles east of its rim. Oligocene ash-flow tuffs and volcaniclastic sedimentary rocks of the Thirtynine Mile field have been folded as much as 28 degrees about the synclinal axis (De Voto, 1964, p. 124). Alluvial fan deposits of the Late Tertiary Trump and Wagontongue formations are as much as 700 feet thick in the southern part of the syncline; these beds dip 5 degrees or less (op. cit.). Vertebrate fossils from the Wagontongue Formation have been identified by C. L. Gazin (in Stark and others, 1949, p. 69) and by G. E. Lewis (in De Voto, 1961, p. 168) as species of Late Miocene or Early Pliocene age. Thus, formation of the Antero Basin was approxi-

THE MOUNT PRINCETON GEOTHERMAL AREA

CHAFFEE COUNTY, COLORADO

by

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Abstract

The Mount Princeton geothermal area is on the west side of the upper Arkansas Valley near Buena Vista, Colorado, along the northern extension of the Rio Grande rift. The area underwent a complex period of Tertiary igneous activity which terminated with the intrusion of the Raspberry Gulch Rhyolite about 22 m.y.b.p. Faulting associated with the Rio Grande rift began in the Miocene and continued to within the last 30,000 years. Surficial thermal manifestations are characterized by zeolitic alteration which covers approximately 64 square kilometers and many thermal springs and wells which have a maximum temperature of 85°C. Chemical analysis of the thermal waters indicate minimum subsurface temperatures of approximately 125°C. Deep circulation of meteoric water in a zone of anomalous heat flow associated with the Rio Grande rift may be the heat source for the thermal features of the area.

Location

The Mount Princeton geothermal area is on the west side of the upper Arkansas Valley between the towns of Buena Vista and Salida in Chaffee County, Colorado (Figure 1). The area is along the western flank of the northern extension of the Rio Grande rift zone and includes a portion of Collegiate Peaks area of the Sawatch Range. The area of thermal manifestations is along the eastern flank of the range and is roughly defined by Cottonwood Creek on the north and Browns Creek on the south. The thermal manifestations are most pronounced along Chalk Creek where the jagged zeolitized, white Chalk Cliffs rise approximately 500 meters (1600 ft) above the Mount Princeton Hot Springs.

Geology

The oldest rock units in the Mount Princeton area are Precambrian metamorphic and igneous rocks (Figure 2). Paleozoic and Mesozoic rocks are not present in the area although a thick section of sediments was deposited over the area during this time interval. Apparently these rocks were removed by the erosional event which followed uplift and the formation of the Sawatch anticlinal structure during the Laramide orogeny.

Tertiary intrusion in the general area may have commenced as early as the Eocene. However, the earliest intrusion in the Mount Princeton area began with the emplacement of the Mount Princeton Quartz Monzonite, which is dated at 36 $\pm$ 2 m.y.b.p. This intrusion of batholithic proportion is the most prevalent rock in the area. It was followed by lamprophyre and latite porphyry dikes and the Mount Antero granite stock which is dated at 30.8 $\pm$ 1.1 m.y.b.p. Afterwards rhyolite dikes dated at 25.4 $\pm$ 1 m.y.b.p. and small quartz monzonite bodies dated at 24 $\pm$ 1 m.y.b.p. were intruded

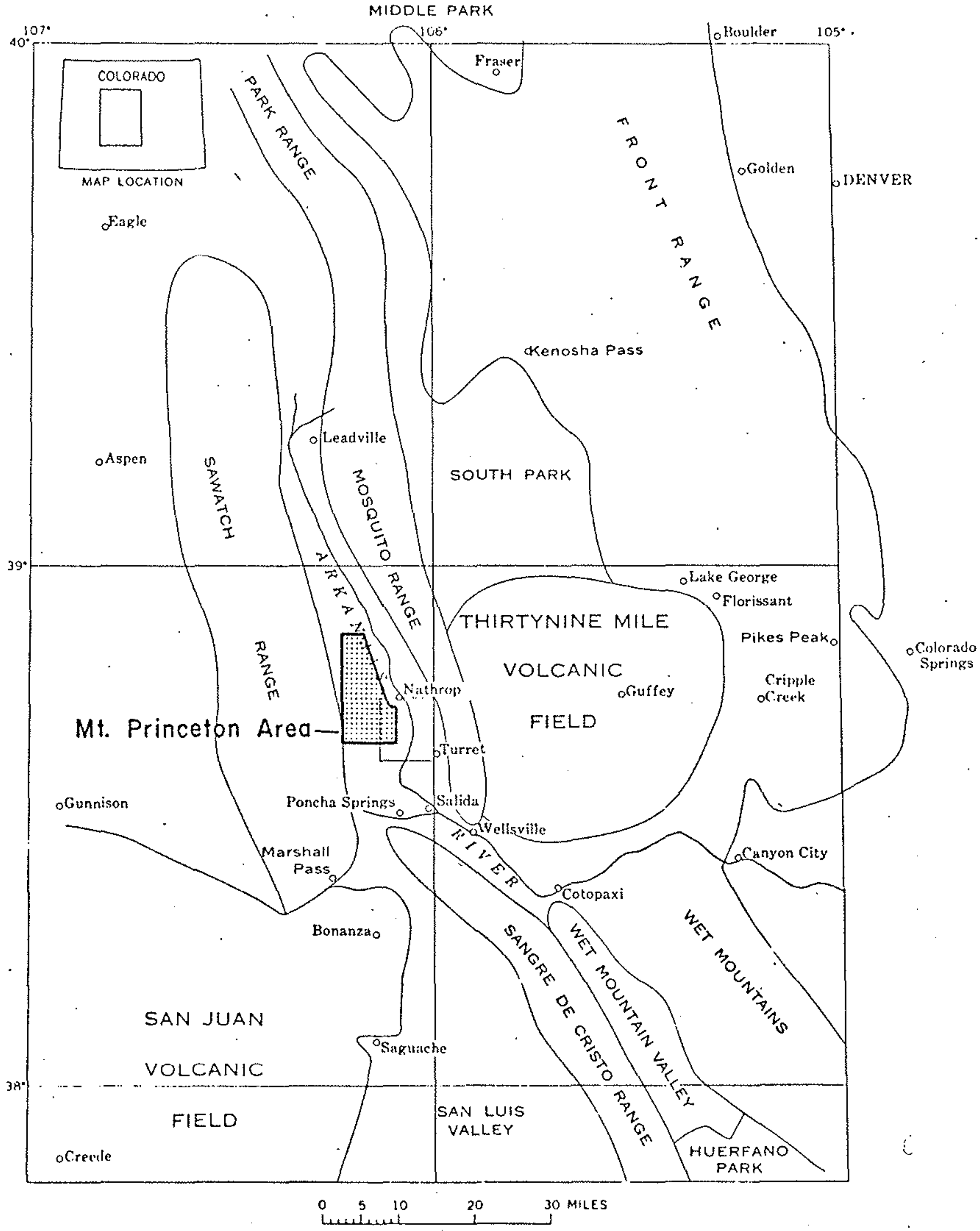

Figure 1. Index map of central Colorado, showing the location of the Mount Princeton area (after Van Alstine and Cox. 1969).

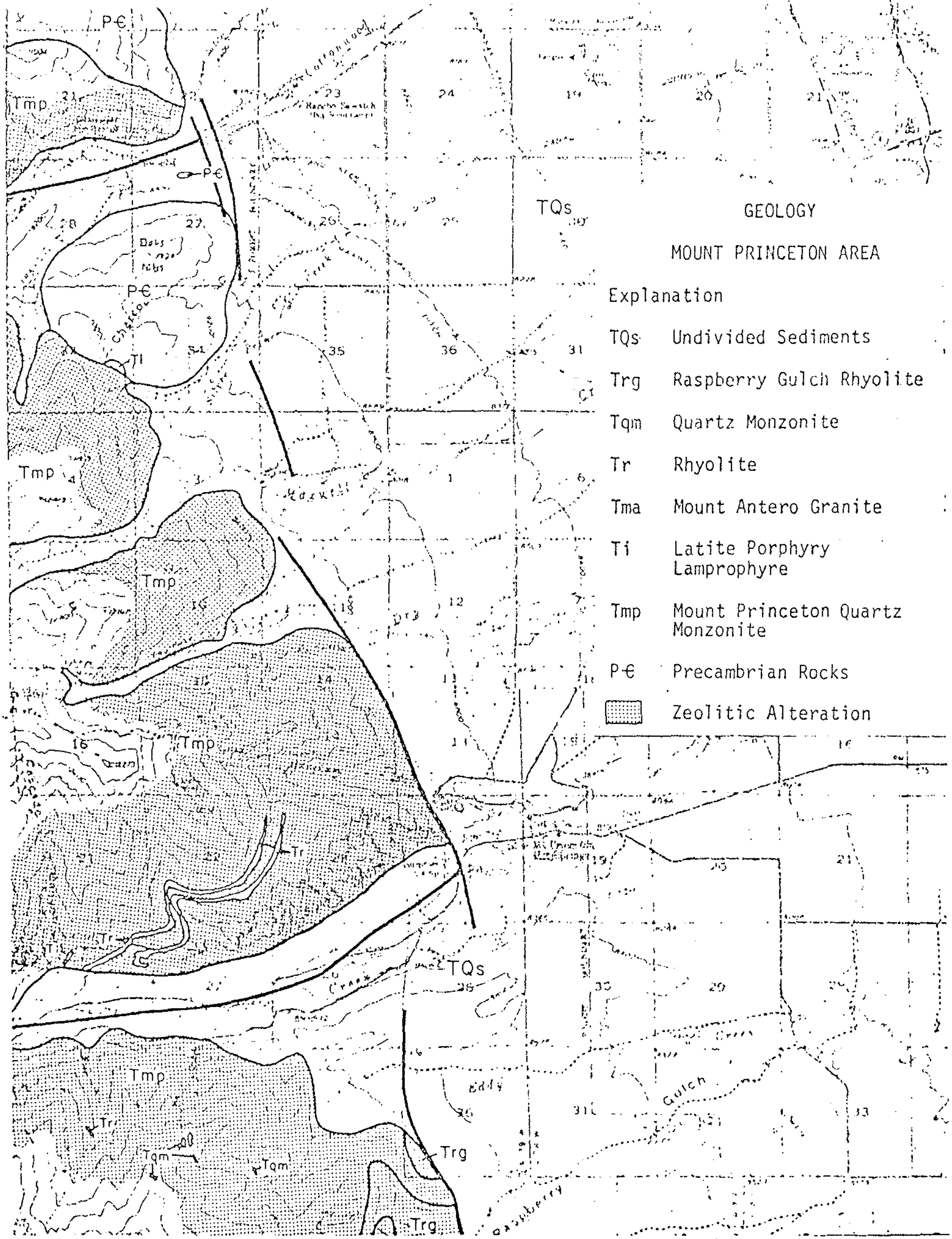

Figure 2. Geologic Map of the Mount Princeton area.

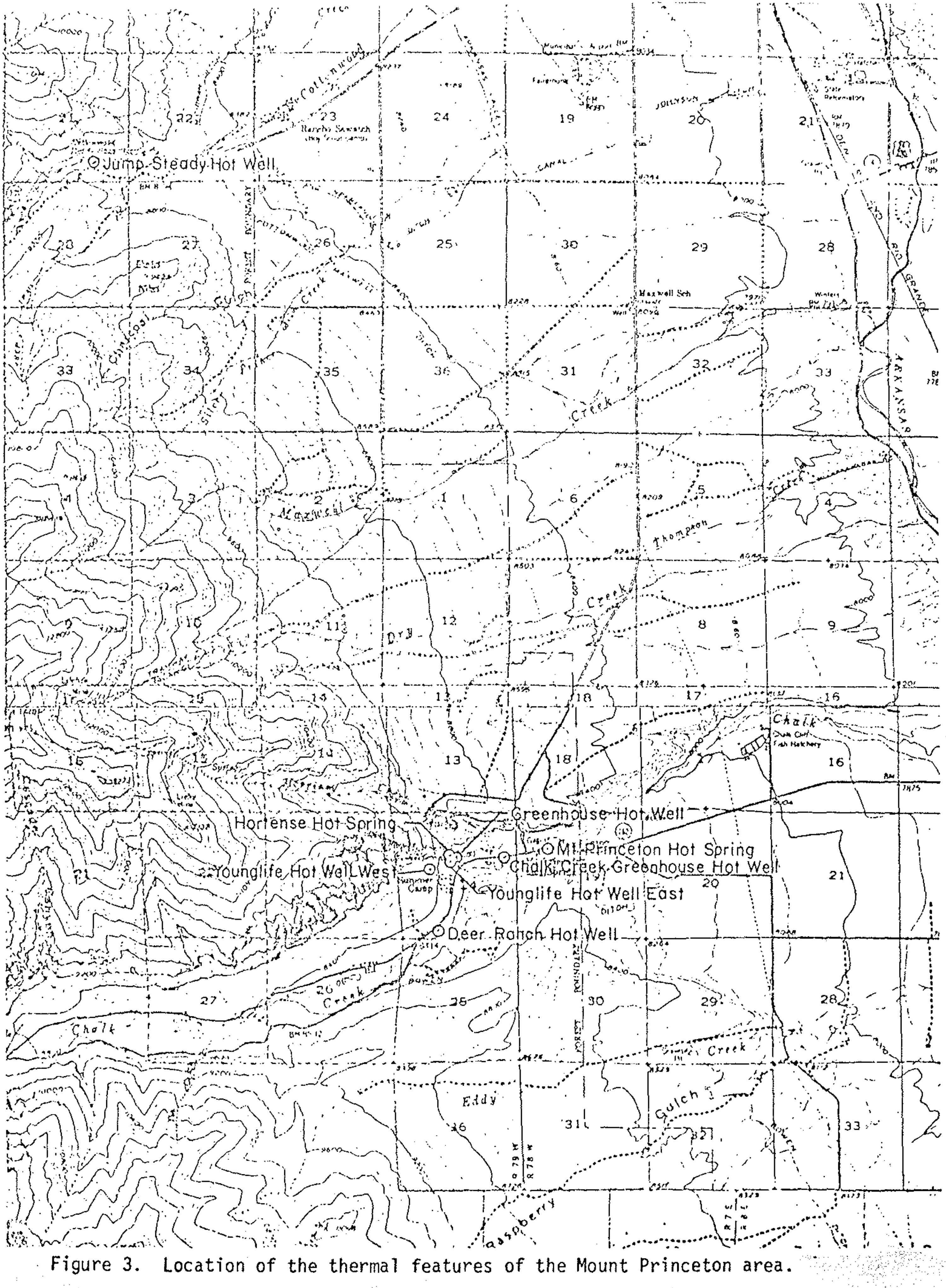

Figure 3. Location of the thermal features of the Mount Princeton area.

into the Mount Princeton batholith. These were followed by the emplacement, to the north of the Mount Antero granite near the headwaters of Raspberry Gulch, of the 22 ± 1 m.y.b.p. Raspberry Gulch rhyolite, which is the youngest igneous rock in the area.

Volcanic rocks of Oligocene age are common in the Mosquito Range to the east of the Mount Princeton area, and two possible rhyolitic vents dated at 28-29 m.y.b.p. form conspicuous domes across from the mouth of Cottonwood Creek.

Regional uplift and the development of the Rio Grande rift began in the early Miocene. Rifting continued throughout the Miocene and Pliocene with the deposition of basin fill sediments, and throughout the Pleistocene with the development of pediment gravels and glacial debris. Faulting along the rift is believed to have continued to at least within the last 30,000 years.

Structure

The Rio Grande rift in the Mount Princeton area is characterized by a series of en echelon normal faults on the west side of the valley and parallel normal faults on the east side which give the valley a graben structure. Faults on the east side of the valley drop the basin in a series of steps, whereas the fault system on the west side of the valley consists of a relatively narrow fault zone characterized by a single large displacement. Parallel north-trending faults on the east side appear to have about 300 meters (1000 ft) of displacement each (Van Alstine, 1969 and Knepper, 1974). Zhody and others (1971) believe on the basis of resistivity measurements that there is 1400 meters (4600 ft) of valley fill deposits in the graben immediately south of Buena Vista. Considering the topographic relief present in the Sawatch Range, the total displacement on the west side fault system may be as much as 3000 meters (10,000 ft).

The gravity data do not show a sharp gradient of the west side of the graben possibly due to the Mount Princeton intrusive center.

Two major cross faults are projected along Cottonwood and Chalk Creeks. Evidence for the faults, which are covered, are hot spring and alteration pattern location, the linear nature of the two valleys, and the nonalignment of the mountain front at Chalk Creek.

Alteration

Hot spring systems at the base of Mount Princeton have produced an extensive zone of alteration. The hydrothermal system, which is active at present, is probably producing a similar alteration assemblage at depth. Alteration is characterized by the calcium zeolite, leonhardite (Sharp, 1970), which may grade into laumontite below surface. In addition to leonhardite, chlorite, illite, epidote, calcite and fluorite are present as alteration minerals.

Zeolitic alteration covers more than 64 square kilometers (25 sq mi) and has a vertical exposure of 1000 meters (3000 ft). Alteration is strongest at Chalk Cliffs above Mount Princeton Hot Springs but a weaker alteration center is concentrated around Cottonwood Hot Springs. The cliffs get their white color from leonhardite and clays which fill fractures to such a degree that they impart a bright white hue to the normally grey host rock.

The zeolitic alteration is strongest in the Mount Princeton Quartz Monzonite but it is also present in Precambrian gneiss, Mount Antero Granite, aplite, rhyolite, and Raspberry Gulch Rhyolite.

Unweathered Mount Princeton Quartz Monzonite shows weak chloritization of biotite and sericitization of feldspars (Limbach, 1975). This weak, widespread alteration is probably related to the crystallization of the Mount Princeton intrusive. Zeolitic alteration that is related to more recent hydrothermal activity converts biotite to chlorite, hornblende to calcite and epidote, orthoclase to sericite, and plagioclase to albite. Leonhardite is confined to fracture openings and does not replace other minerals. The calcium for the zeolite comes from the breakdown of hornblende and plagioclase. Chemical analysis of the strongly altered Mount Princeton Quartz Monzonite shows little change from the unaltered quartz monzonite, except for slight additions in iron and water.

The zeolite alteration assemblage forms at high activities of H_2O relative to CO_2. These conditions prevail where hot water has pH of 8 to 9 (White and Sigvaldson, 1963), which fits the hot springs in the Mount Princeton area. Sharp (1970) has suggested that the zeolitic alteration assemblage present at Chalk Cliffs formed within a range of temperatures of 145-220°C and depths of 150-2000 meters (500-6600 ft) below the surface based on comparison with active geothermal areas in New Zealand and Iceland.

Thermal Springs

The thermal springs in the Mount Princeton area have been described by George, et al (1920) Lewis (1966) Sharp (1970) and Pearl (1972). Most of the hot wells are used for heating homes, greenhouses, bathing and drinking (Figure 3).

The area contains at least 6 thermal wells and 2 thermal springs (Table 1) and includes Hortense Hot Spring which is reportedly the hottest spring in Colorado. Many other thermal seeps issue directly into Chalk Creek, and cannot be counted. Several low pressure steam fumeroles are present about 100 meters (300 ft) to the west of Hortense Hot Spring in a talus slope at the base of the Chalk Cliffs. The approximate heat discharge for the thermal features of the area, computed as the product of the volume rate and enthalpy of the water in excess of ambient temperature, is seen in Table 1. All the thermal features combine to produce 4×10^6 cal/sec. or enough heat to supply approximately 200 average sized houses.

Table 1. The thermal features of the Mount Princeton area.

Sample Name	T°C	Flow l/m	Heat Discharge cal/sec.
Hortense Hot Spring	85	38	4.9×10^4
Younglife Hot Well East	85	379	4.9×10^5
Younglife Hot Well West	67	379	3.7×10^5
Greenhouse Hot Well	68	379	3.8×10^5
Chalk Creek Greenhouse Hot Well	65	1892	1.8×10^6
Jump Steady Hot Well	59	568	5.0×10^5
Mt. Princeton Hot Spring	56	265	2.1×10^5
Deer Ranch Hot Well	38.5	379	1.9×10^5
			4.0×10^6 cal/sec.
			1.6×10^4 BTU/sec.

Analyses of thermal and non-thermal waters of the Mount Princeton area are given in Table 2. Non-thermal waters of the Mount Princeton area generally contain less than 150 mg/l of dissolved solids. Water pH is

Table 2. Chemical analysis of the thermal features of the Mount Princeton area. Units are mg/l unless otherwise noted.

	Hortense Hot Spring	Younglife Hot Well East	Greenhouse Hot Well	Younglife Hot Well West	Chalk Creek Greenhouse Hot Well	Jump Steady Hot Well	Mt. Princeton Hot Spring	Deer Ranch Hot Well	Ice Pond Cold Sprin
pH	9.6	9.2	9.1	9.1	8.8	9.2	8.6	8.8	7.6
Cl	8.8	11	6.6	2.2	6.6	28	5.5	4.4	3.0
F	16	15	13	9.3	10	14	9.4	6.2	0.2
HCO_3	46	43	79	44	52	31	59	64	68
CO_3	16	18	20	10	0	24	0	0	0
SO_4	100	90	80	60	70	110	60	40	6
SiO_2	85	80	75	75	65	60	60	45	25
Na	100	80	80	60	60	110	50	40	7
K	4.0	2	2	2	2	2	2	2	1
Ca	15.0	12	7	17	10	50	20	20	18
Mg	0.1	<0.1	<0.1	0.4	0.1	0.3	0.4	0.9	4
Li	0.2	0.1	NA	0.1	NA	0.2	0.1	0.1	NA
B	<1.0	<1.0	<1.0	<1.0	<1.0	<1.0	<1.0	<1.0	<1.0
NH_3	0.4	0.3	<0.1	<0.1	NA	<0.1	<0.1	<0.1	<0.1
TDS	392	351	363	280	276	430	266	223	132
T°C	85	85	68	67	65	59	56	38.5	9
Flow (gpm)	10	100	100	100	500	150	70	100	75
$TSiO_2$°C	125	125	122	122	115	110	111	97	72
TNa/K°C	97	74	64	84	84	45	97	115	230*
TNa-K-Ca°C	75	55	67	47	57	34	43	41	12
Cl/SO_4	0.2	0.3	0.2	0.1	0.3	0.7	0.3	0.3	1.4
Cl/F	0.3	0.4	0.3	0.1	0.1	1.1	0.3	0.4	8.0
Cl/HCO_3+CO_3	0.5	0.6	0.2	0.1	0.4	1.7	0.3	0.2	0.2
Resistivity ohm-m	21.4	24.2	NA	NA	NA	19.2	32.6	36.2	63.7

NA = not analysed

+ = Does not represent true subsurface conditions, i.e. $\frac{\sqrt{Ca}}{Na} > 1$

generally neutral to slightly basic. Bicarbonate is the principle ion followed by silica, calcium, sodium and magnesium. Cold waters contain an average of 22 mg/l of silica. Ice Pond Cold Spring, about 1 mile to the northeast of Buena Vista, was chosen to represent background water chemistry.

Thermal waters exhibit basic to very basic pH. Four types of thermal water are recognized.

1. Sulfate-sodium waters with less than 12 mg/l chloride may represent steam condensate that has equilibrated with quartz monzonite of the Mount Princeton batholith. Hortense and Mt. Princeton Hot Springs and three hot wells located south of the Chalk Cliffs are included in this category.
2. Sulfate-sodium waters with greater than 25 mg/l of chloride probably represent low temperature, low salinity hot water systems, derived through deep circulation. These waters are saturated with calcium carbonate minerals and deposit varying amounts of travertine. The Jump Steady Hot Well (Cottonwood Hot Spring) characterizes this category.
3. Bicarbonate-sodium waters with less then 7 mg/l of chloride may represent dilutions of sulfate-sodium waters with bicarbonate rich groundwaters. Greenhouse Hot Well and Deer Ranch Hot Well represent this group.
4. Groundwater as is found at the Ice Pond Cold Spring, is generally rich in bicarbonate.

Chemical analyses of hot spring systems may be used to estimate subsurface temperatures of active geothermal areas. The assumptions made in applying geochemical indicators are summarized by Fournier, White, and Truesdell (1974), and Fournier and Truesdell (1973) suggest criteria for selecting the most probable temperature. These techniques were utilized in determining the subsurface equilibrium temperatures given in Table 2 and indicate temperatures in the order of 125°C.

Waters in the Mount Princeton area were analysed for oxygen-18 deuterium and tritium. In analysis of the data deutrium and oxygen-18 have been normalized relative to standard mean ocean water (SMOW), and are noted as δD and δO^{18}.

Figure 4 shows the variation between δD and δO^{18} relative to SMOW for several geothermal areas. The straight line represents the almost world-wide slope for meteoric waters plotted in this way. Deuterium concentrations are constant and equal to local meteoric water. On the other hand O^{18} concentrations show a characteristic enrichment or shift. The O^{18} shift is due to an isotopic oxygen exchange between groundwater and carbonates and silicates in the rocks. Silicate and carbonate rocks contain O^{18} ranging from +6 to +30 per mil greater than SMOW. Deuterium generally does not vary from the meteoric concentration because rocks contain negligible hydrogen or deuterium. A strong shift in O^{18} implies a long storage time and/or a large reservoir capacity. A very small shift implies one of two situations: first, temperature-pressure conditions are too low to allow waters to exchange O^{18} with rocks within a relatively short time period and second, descending meteoric waters are heated and rise so quickly that insufficient time is available for an O^{18} exchange to occur. As shown on Figure 4, Niland waters which have mingled over a long period of time with carbonate rich Colorado River sediments show the greatest shift. On the other end of the scale, Wairakei shows negligible shift which implies that waters descend quickly, stay in storage for only a short time and then ascend.

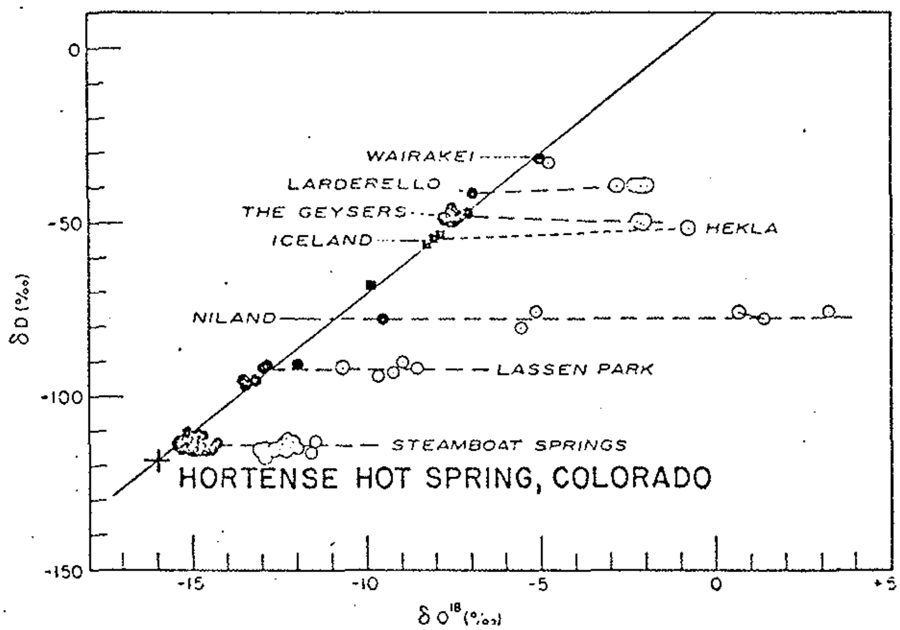

Figure 4. Observed isotopic variations in near-neutral chloride type geothermal waters and in geothermal steam. Solid points are local meteoric waters, or slightly heated near-surface groundwaters. Open circles are hot springs or geothermal water, crinkled circles are high temperature, high pressure, geothermal steam (after Craig, 1963).

Figure 5 is a plot of δD-δO^{18} for selected waters of the Mt. Princeton area. The hot waters show no apparent O^{18} shift. This implies that these hot waters have been in residence with the reservoir rocks only a short time and have not exchanged oxygen isotopes as at Niland.

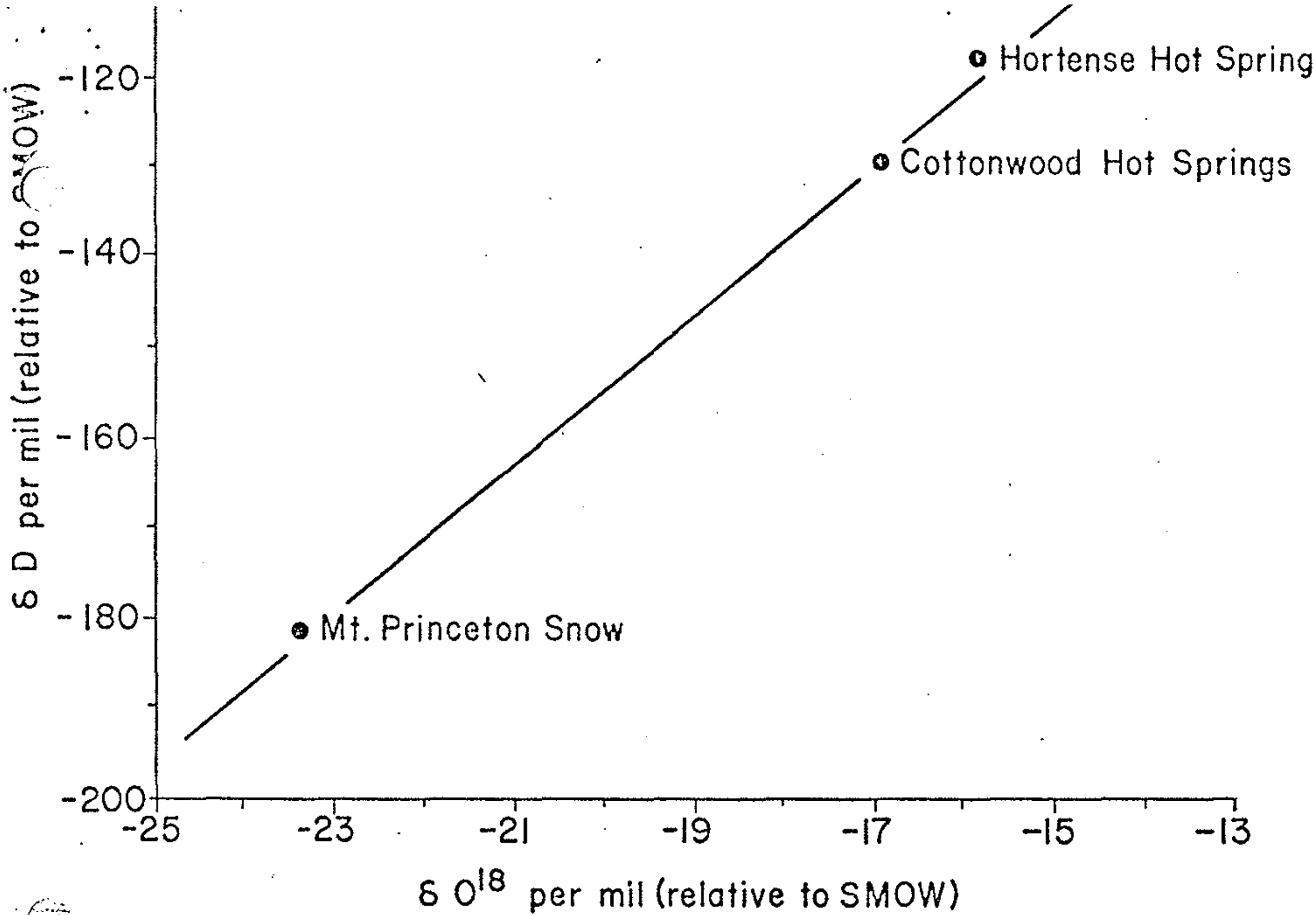

Figure 5. δD versus δO^{18} for waters of the Mount Princeton area.

Tritium analyses indicate an age of 20 to 51 years for Hortense Hot Spring and 21 to 56 years for the Jump Steady Hot Well. These ages are in good agreement with the suggested youth of the same waters by O^{18} analysis.

Heat Source

The youngest igneous rock in the Mount Princeton area is the 22 million year old Raspberry Gulch Rhyolite which is too old to be the heat source for the Mount Princeton area hot springs, and it is doubtful that any younger igneous intrusions are present at shallow depths. The youngest rocks in the northern part of the Rio Grande rift system are 4.5 to 3.6 million years old olivine tholeiitic basalts in the southern portion of the San Luis Valley. According to Lipman (1969) these basalts were derived from depths of 15-20 kilometers. This suggests that a portion of the rift is underlain by an upward protrusion of hot mantle rocks that forms a zone of abnormally high heat flow. Heat flow measurements along the rift show values of two to three times the average crustal value (Gross, 1974). White (1957) states that meteoric water in fractured rocks can circulate to depths of 3000 meters, and the active faulting within the past 30,000 years could have permitted the fracture system at Mount Princeton to remain open to considerable depths. Circulation of meteoric waters to these depths would be more than adequate to produce the hot springs at Mount Princeton and the indicated subsurface temperatures.

Although igneous intrustion into faults of the rift system cannot be completely ruled out, the most likely heat source for the thermal manifestations in the Mount Princeton area is the abnormally high geothermal gradient associated with the Rio Grande rift.

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