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GEOCHEMISTRY AND HYDROTHERMAL ALTERATION
AT SELECTED UTAH HOT SPRINGS
(Revised)

by

W. T. Parry, Nancy Benson, Charles D. Miller

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GEOCHEMISTRY AND HYDROTHERMAL ALTERATION
AT SELECTED UTAH HOT SPRINGS

W. T. Parry, Nancy Benson, Charles D. Miller

ABSTRACT

Application of Na-K-Ca geothermometry to warm springs in Utah indicates several areas with sufficiently high apparent temperatures to be of interest as geothermal exploration targets. A zone of warm springs in the Bonneville Basin show Na-K-Ca temperatures from 150°C to 233°C. Examination of Great Salt Lake, Bonneville sediment pore water, and Jordan Valley well-water chemistry indicates that mixing a small percent of these fluids with warm spring water can cause substantial errors in Na-K-Ca temperature estimates. Other saline deposits which may influence Na-K-Ca temperature estimates are the Paradox formation in southeastern Utah, the Muddy Creek formation in southwestern Utah, the Arapien shale in central Utah, the Preuss formation in northeastern Utah, and Playa salts in much of western Utah. The Roosevelt KGRA is the most attractive target identified by Na-K-Ca geothermometry.

Hydrothermal alteration, heavy metal distribution and water chemistry provide additional characterization of the Roosevelt system. The hot spring deposits consist of laminated opal and siliceous sinter. Opal also cements alluvium. Hydrothermally altered rocks in outcrop consist of opal, alunite, quartz, sulfur and unaltered detritus. Kaolinite, montmorillonite, and K-mica have been recognized in only one outcrop. Hydrothermal alteration in one core consists of opal, alunite, and native sulfur to 60' (18.3m); kaolinite and alunite to 80' (24.4m); kaolinite, montmorillonite, and alunite to 100' (30.5m); and kaolinite, montmorillonite, K-mica and pyrite to total depth of 217' (66.1m). Host for the alteration is alluvium from the Tertiary pluton to 115' (35.1m) and Precambrian gneiss from 115' (35.1m) to 217'

(66.1m).

Mercury in soil occurs as 10m to 30m wide anomalies up to 3 parts per million immediately over subsurface fractures. Arsenic anomalies reach 0.18 percent. A gamma-ray survey over the opal dome shows low Uranium, potassium, and thorium values over vertically laminated opal.

Chemistry of a cool water seep (25°C) shows Na-K-Ca temperature of 241°C and SiO_2 temperature of 125°C . A Phillips well flowing from below 1500' (457m) shows Na-K-Ca temperature of 262°C , SiO_2 temperature of 262°C , and K of 1.5 times the surface spring value.

The near surface alteration assemblage is best explained in terms of a decrease in pH of near surface fluids as sulfide oxidizes. Increasing potassium and pH with depth indicates that a K-feldspar stable zone may be intersected with deeper drilling.

Geology and alteration have been mapped in the Monroe KGRA. Alteration of the volcanic host rocks includes the following assemblages: kaolinite and jarosite, kaolinite and mixed layer clay, montmorillonite and mixed layer clay, montmorillonite and chlorite. Calcite and gypsum are also abundant. The alteration mineral assemblages have been grouped using multivariate analysis. The chemistry of modern spring water is consistent with the alteration. The relationship between the Monroe Springs and other springs and wells in the vicinity suggest that Monroe is a hot brine diluted by cool nonsaline groundwaters. The source brines must exceed 100°C .

INTRODUCTION

Potentially economic geothermal resources in Utah will likely consist of large volume of hot rock at depth with a circulating ground water medium transferring heat to the surface. The chemistry of the water and the water-rock interaction should provide valuable guides to temperatures in the deep reservoir and the size, if not of the reservoir itself, of the fluid leaks to the surface. Water and altered rock chemistry are thus important exploration tools useful in locating areas of anomalous subsurface temperature and important parameters of resource characterization.

This report details the results of estimation of subsurface temperatures in Utah from published water analyses and new analyses of selected hot springs. Two hot spring areas in differing geological environments have been selected for detailed rock alteration studies. The Monroe-Red Hill area is within an area of Tertiary volcanics and is underlain by a thick Jurassic salt bearing deposit. The Roosevelt area is in Tertiary and Precambrian crystalline rocks. Comparison of rock alteration and water chemistry in these two areas should serve as a model for further exploration and resource characterization in Utah using geochemical techniques.

GEOLOGIC BACKGROUND

Proper perspective for subsequent discussion requires a broad overview of the geological development of Utah. The stratigraphic history has been described in six phases by Hintze (1973). Phase I, the earliest of the six phases, consists of deposition of a thick sequence of shallow-water marine sediments in western Utah in the time interval Late Precambrian to Devonian. Sediments of the age are thin to absent in eastern Utah. Phase I sediments were deposited on older Precambrian crystalline rocks which are only sketchily known.

During Phase II, several local basins developed in which thick sequences of marine sediments accumulated in Mississippian to Permian time. The Oquirrh Basin in north-central Utah contains nearly 30,000 feet of late Paleozoic sediments and the Paradox Basin in southeastern Utah contains a thick sequence of marine evaporites and complementary shelf carbonates up to several thousand feet thick.

Phase III, from Triassic to Cenozoic, sediments accumulated in eastern Utah while western Utah was involved in the folding and thrust faulting of the Sevier Orogeny. Associated clastic sediments accumulated in the shallow sea of eastern Utah.

The structural features of the Laramide Orogeny from late Cretaceous to Eocene, Phase IV, ended marine sedimentation in Utah. Anticlinal uplifts of the Uinta Mountain, San Rafael Swell, Circle Cliffs, and Monument Valley rose and adjoining basins filled with stream and lake sediments.

Extensive volcanic activity characterized Phase V (Oligocene). Extensive ash-flow tuffs, breccias, and lava flows blanketed western Utah. Volcanic and plutonic rocks of the Tintic, Bingham, Marysvale, Cedar City, and Park City mining districts, and the laccoliths of the Henry, LaSal, and Abajo mountains formed during this same time.

North-trending fault blocks in western Utah formed during Phase VI (Miocene to Recent). Also approximately 5000 feet of uplift of the western cordillera occurred during late Cenozoic time. Stream and lake sediments were deposited in valleys between uplifted and faulted mountains. Basaltic volcanism accompanied block faulting in western Utah.

WATER CHEMISTRY

The basic tools for estimation of subsurface temperature from water chemistry have been provided in a series of papers by Fournier and coworkers

(Fournier, et al. 1974a, 1974b; Fournier and Rowe, 1966, Fournier and Truesdell, 1973). Application of the Na-K-Ca geothermometer of Fournier and Truesdell (1973) to published analyses of thermal spring waters in Utah (Mundorff, 1970) has indicated several anomalously high temperatures (Swanberg, 1974). Our own application of the Na-K-Ca geothermometer to the most recent analyses of Mundorff is shown in Figure 1 and Table 1. Most of the thermal springs are in western Utah and many of them are associated with the block faults of Phase VI in western Utah.

The highest apparent Na-K-Ca temperature occurs at Roosevelt Hot Springs, but several apparent high temperatures occur in springs in northern Utah notably at Hooper, Southwest Hooper, Ogden, and Utah Hot Springs (Table 1). Apparent Na-K-Ca temperature for Great Salt Lake brine is 241°C . Mixing of a small amount of such water with warm spring water could therefore produce erroneous or misleading Na-K-Ca temperatures. To further illustrate the problem of Great Salt Lake bring mixing, the Na-K-Ca temperature of warm spring water (Table 1) mixed with a varying volume percent Great Salt Lake brine has been calculated (Figure 2). Warm spring water shows Na-K-Ca temperature of 20 to 24°C ; mixed with 0.1% Great Salt Lake water the apparent temperature 52°C , and mixed with 0.3% Great Salt Lake brine the apparent Na-K-Ca temperature is 150°C .

Examination of chemistry of pore water in Lake Bonneville sediments in Jordan Valley, Utah (Buck and Parry, 1976) also indicates anomalously apparent high Na-K-Ca temperatures of up to 302°C . Ground water moving through these sediments gains dissolved constituents which may produce misleading Na-K-Ca temperatures. The general direction of ground water flow in Jordan Valley is from the Wasatch Mountains on the east through lake sediments to Great Salt Lake. An approximate flow path is shown in Figure 3. Well water chemistry along this flow path shows a trend of increasing dissolved chloride and sodium

Na-Ca-K TEMPERATURES

Figure 1. Na-K-Ca temperatures (Fournier and Truesdell, 1973) of Utah warm springs from analyses of Mundorff (1970).

Table 1. Minimum Temperatures of Hot Water - Rock Equilibration
for Utah Thermal Springs

<u>SPRING</u>	<u>Spring 1 Temp. °C</u>	<u>Na-K-Ca temp. °C²</u>	<u>Silica temp. °C³</u>
Como	25	46	62
Becks	56	152	82
Wasatch	41	148	55
Hooper	48	204	80
S.W. Hooper	32	222	100
Ogden	57	222	97
Utah	57	233	85
Stinking	47	184	93
Crystal	53	186	74
Warm	27	21	51
Uddy	43	164	74
Blue	27	160	-
Big	18	160	32
Grantsville	24	152	75
Russells	22	77	62
Morgans	27	83	62
Fish	22	72	64
Abraham	82	155	110
Gandy	27	36	65
Meadow	41	47	99
Richfield	22	27	43
Monroe (Cooper)	42	183	104
Red Hill	77	187	103
Johnson	25	9	82
Joseph	64	138	123
Roosevelt	55 (1957)	283	213
Radium	32	89	82
Thermo	77	196	137
Veyo	32	36	82
La Verkin	42	190	77
Midway	28	81	65
Lincoln	32	194	65
Diamond Fork	20	58	58
Castilla	40	60	80
Goshen	22	149	58
Sterling	19	51	48

-
1. From Mundorff (1970)
  2. Na-K-Ca Empirical geothermometer of Fournier and Truesdell (1973)
  3. Silica temperatures assuming equilibration with quartz and no steam loss.

Figure 2. Apparent Na-K-Ca temperatures of Warm Spring mixed with varying volume fraction Great Salt Lake brine.

Figure 3. An approximate flow path for ground water in Jordan Valley, Utah.

with distance from the Wasatch Mountains shown in Figures 4 and 5. This increase in dissolved solids parallels an increase in calculated apparent Na-Ca-K temperatures of well water as the present lake is approached (Figure 6). Apparent Na-K-Ca temperatures for other selected natural waters are shown in Table 2. Natural non-thermal waters which have been affected by evaporation, contact with marine evaporite salts, or continental evaporite salts may show anomalous Na-K-Ca temperatures. The effect appears to be most severe with evaporite deposits which contain substantial potassium such as the Salado of southeastern New Mexico or the potassium bearing brines of Great Salt Lake and the nearby Great Salt Lake Desert (Table 2). Saline geologic units in Utah which may contribute to misleading Na-K-Ca temperatures are the Paradox formation in southeastern Utah, the Muddy Creek formation in southwestern Utah, the Arapien shale in central Utah, and the Preuss formation in northeastern Utah. The general distribution of these units is shown in Figure 7. Playa salts in much of western Utah not shown in Figure 7 may also cause erroneous temperature estimates. An accurate and thorough examination of the geology of each spring is required for reliable estimation of minimum geothermal reservoir temperatures and effective use of chemical geothermometers as an exploration tool.

The Roosevelt KGRA is the most attractive geothermal exploration target in Utah identified by available water chemistry. The spring waters (presently no longer flowing) show the highest Na-K-Ca and SiO_2 temperatures of any thermal spring in the state and minimal possibilities for evaporite influence on spring chemistry.

Field determination of Na, K, and Ca using specific ion electrodes was attempted. An Orion specific ion meter and sodium, calcium, and potassium electrodes were used. Sodium was determined reliably and accurately compared to laboratory atomic absorption measurements, but the K and Ca membrane

Figure 4. Variation in dissolved anions in well water with distance along the flow path shown in Figure 3. Water analyses from Marine (1964).

Figure 5. Variation in dissolved cations in well water with distance along the flow path shown in Figure 3. Water analyses from Marine (1964).

Figure 6. Variation in apparent Na-Ca-K temperature of well water with distance from Great Salt Lake in Jordan Valley, Utah.

Table 2. Apparent Na-K-Ca Temperatures
for Selected Natural Waters

<u>Water Source</u>	<u>Na-K-Ca Temperature °C</u>
Pecos River	49
San Rafael River	63
Weber River	24
Sevier River	72
Humboldt River	180
Green River	30
San Juan River	47
Salton Sea	130
Great Salt Lake	241
Seep from Salado Salt (New Mexico)	210
Great Salt Lake Desert Brine	212
Salado Salt Brine	690

SALINE DEPOSITS

SODIUM CHLORIDE, SURFACE
AND SUBSURFACE: Qb,
RECENT; Jp, PREUSS; Ja,
ARAPIEN: Php, PARADOX

POTASH, SUBSURFACE

Figure 7. Distribution of major saline deposits in Utah.

electrodes did not produce stable readings despite repeated attempts both in the field and the laboratory. The source of the instability was not isolated, but if stable Na, Ca, and K electrodes could be obtained, then rapid measurement of these elements can be made in the field and minimum wall-rock-equilibration temperatures estimated.

MIXING MODELS

The chemical thermometers Na-K-Ca and SiO_2 frequently indicate equilibration at temperatures higher than actual spring temperatures. Fournier et al. (1974) suggest that in such cases the spring water may be treated as a mixed water using the method of Fournier and Truesdell (1974) providing the rate of flow is sufficiently large. No meaningful mixing model could be developed for Roosevelt Hot Spring due to extremely small flow rates (1 gallon per minute in 1950 and essentially 0 at present). Mixing models have been attempted for Red Hill and Abraham Hot Springs.

Abraham Hot Springs flow from the floor of the Sevier Desert along the east side of a Quaternary basalt flow (Mundorff, 1970). Estimated flow rate in 1967 was 95- liters per minute at a temperature of 82°C . Na-K-Ca temperature is 155°C and SiO_2 (quartz) temperature is 110°C . Water-well drilling in the area has encountered Lake Bonneville clay and sand with salt water to depths of approximately 30 meters and fresh water below approximately 50 meters. Mixing models assuming conservation of enthalpy and silica with mixing of thermal and cold water are shown in Figure 8. Calculations were made assuming cold water temperatures of 10°C and 20°C and cold water silica (CWS) contents of 15 ppm and 10 ppm. Cold water mixing fractions (CWF) are 0.44 for cold water temperature of 10°C and 0.52 for cold water temperature of 20°C . Hot water temperature indicated is 135°C for 10°C cold water and 145°C for 20°C cold water compared to Na-K-Ca temperature of 155°C .

The Red Hill Spring flows as much as 570 liters per minute at a temperature

Figure 8. Abraham (Crater) Hot Springs mixing model.

Figure 9. Red Hill Spring mixing model I.

Figure 10. Red Hill Spring mixing model II.

This figure is erroneous and has been deleted.

Figure 11. Na-K-Ca temperatures of spring and well water in the Milford, Utah area. Areas of above 50°C temperatures are shaded. Arrows indicate groundwater flow directions. Water analyses from Mower and Cordova (1974).

thus be expected to flow toward Milford. Several Na-K-Ca temperatures near Milford are as high as 70°C and one reaches 90°C. Actual water temperatures are as high as 27°C compared with the normal subsurface water temperature for the area of 14 to 17°C; and silica concentrations are 50 to 90 ppm compared to normal values of 20 to 40 ppm confirming the presence of thermal waters near Milford. Subsurface thermal water from Roosevelt would be expected to flow westward to the Beaver River. Two anomalous Na-K-Ca temperatures obtained northwest of Roosevelt are 153°C and 148°C with silica concentrations of 44 ppm and 56 ppm respectively. The actual temperature of the latter is 12°C and the first was not measured. Such anomalies could be expected down stream from thermal areas, in waters leaking salts from Bonneville sediments, or near separate subsurface thermal reservoirs. The cool springs near the southwest margin of the Mineral Mountains show normal cool Na-K-Ca temperatures from 9°C to 22°C.

ROOSEVELT HOT SPRINGS

The geology of the Roosevelt KGRA is described in detail in other volumes of this series (Nash et al., 1976). The KGRA is located on the western flank of the Mineral Mountains, a fault block produced during Phase VI. Precambrian (?) gneiss and Paleozoic and Mesozoic sedimentary rocks have been intruded by a granitic pluton. Several north-south trending normal faults bound the Mineral Mountains. One of these, the Dome Fault, has localized hot spring deposits and alteration. Modern hot spring flow also occurred along the Dome fault system.

Surface geology and rock alteration at Roosevelt Hot Springs has been described by Peterson (1975). Spring deposits along the Dome Fault consist of opal and siliceous sinter. Opal also cements alluvium consisting of quartz, orthoclase, biotite, plagioclase, and minor accessories from the Tertiary pluton and Precambrian (?) gneiss on the flanks of the Mineral

Mountains immediately to the east. The opal deposits vary from x-ray amorphous, to β -cristobalite, to completely recrystallized material consisting of microcrystalline α -quartz. Host rocks for alteration in surface outcrop are detritus from the Mineral Mountains, Precambrian (?) metamorphic rocks, and volcanic ash.

Procedures used in investigation of alteration include x-ray diffraction, x-ray emission, and petrographic methods. Surface outcrops were sampled at localities shown on Figure 12. Pellets for x-ray emission determination of major elements, arsenic, and lead were prepared by crushing to 1/4 inch size in a small jaw crusher with ceramic plates and pulverizing in the tungsten carbide ball mill (Pitchford Manufacturing Pica Mill). One gram of pulverized sample was mixed with one gram of chromatographic cellulose (Whatman No. CF11) and pressed at 20 tons force into a 1-1/4" diameter disc using aluminum caps (Spex Inc.). Instrumental conditions used for the x-ray emission analysis are shown in Table 3. X-ray diffraction scans were obtained for bulk samples and oriented smears of minus two micron size separates of each sample.

Surficial Alteration and Spring Deposits

The most extensive area of spring deposits occurs at the Opal Dome (Figure 12) where the deposits consist of banded and laminated opal, siliceous sinter, and alluvium cemented with opal. The color of the opal varies from deep red, to pink, gray, green, brown and a variety of other shades. Some deep red opal cement contains hematite (Sample R-6) some of which occurs as cubic pseudomorphs after pyrite and 0.18% arsenic. Analyses of opal cemented alluvium are in Table 4 as analyses R1 through R7 and analysis R9 is of porous siliceous sinter. Smaller deposits of opal and sinter occur in various localities along the Dome Fault.

Figure 12. Surface geology in the vicinity of Roosevelt Hot Springs showing location of surface outcrop samples. Geology from Petersen (1975).

Table 3. X-ray Fluorescence Analysis. Analytical Conditions

<u>Element</u>	<u>Na</u>	<u>Mg</u>	<u>Al</u>	<u>Si</u>	<u>S</u>	<u>K</u>	<u>Ca</u>	<u>Fe</u>	<u>As</u>	<u>Pb</u>
X-Ray Tube Target	Cr	Cr	Cr	Cr	Cr	Cr	Cr	W	W	W
Kilovolts	40	40	40	40	40	40	40	40	40	40
Milliamperes	50	40	40	40	40	40	40	5	40	40
Analyzing Crystal	RAP	RAP	EDDT	EDDT	EDDT	EDDT	EDDT	LIF200	LIF220	LIF220
Detector	Flow Counter	Flow Counter	Flow Counter	Flow Counter	Flow Counter	Flow Counter	Flow Counter	Scint	Scint	Scint
Analytical Line two theta	54.4	44.6	142.7	108.05	74.2	50.35	44.9	57.5	48.9	40.5
Background two theta	56	43	141	110	76.8	51.5	43.5	59	47.9 49.9	41.2
Counting time	100sec.	100sec.	10sec.	10sec.	100sec.	10sec.	10sec.	10sec.	100sec.	100sec.
Path	VAC	VAC	VAC	VAC	VAC	VAC	VAC	AIR	AIR	AIR

Table 4. Chemical Analyses of Outcrop Sample from Roosevelt Hot Springs Area

Sample No.	R1	R3	R4	R5	R6	R7	R9	R13	R14	R15	R16
SiO ₂	72.5	85.0	76.4	72.3	64.6	77.5	92.1***	69.8	44.5	50.8	62.5
Al ₂ O ₃	11.3	4.3	5.7	12.6	14.7	8.5	0.9	9.2	20.3	17.1	15.6
Total Fe as Fe ₂ O ₃	3.4	1.0	1.3	3.9	5.4	2.3	0.2	.4	.4	.9	.3
MgO	0.4	0.3	0.5	0.9	ND	0.7	0.1	1.0	ND	ND	.04
CaO	0.9	0.4	3.0	1.2	0.3	0.5	0.1	4.9	0.2	.04	0.2
Na ₂ O	3.1	1.2	1.8	3.4	1.9	1.3	0.1	.6	0.3	ND	.3
K ₂ O	4.3	1.9	4.0	3.4	4.4	2.7	0.1	4.0	5.3	5.0	6.0
SO ₃	0.30	0.07	0.12	0.25	0.42	0.30	0.15	4.67	18.9	18.1	8.6
Loss on Ignition	2.0	3.4	3.3	2.9	4.2	3.9	5.2	11.8	23.9	23.5	12.9
H ₂ O ⁺							3.48*				
Total **	97.9	97.5	96.0	100.6	95.5	97.5	98.95	101.7	94.9	97.34	97.84
Bulk Density	2.20	1.84	2.33	1.92	1.94	1.81	1.49	<1.00	1.58	1.75	1.98
As ppm	110	7	79	57	1760	92	46	46	ND	ND	ND
Pb ppm	28	19	25	42	28	32	ND	37	66	36	33

*H₂O⁺ by Penfield tube
 **2

*** gravimetric silica

R1, R3, R4, R5, R6, R7 - Detrital K-spar, quartz, plagioclase, biotite cemented with opal.

R9 - siliceous sinter

R13 - Alunite-K-feldspar, opal and possibly volcanic glass

R14, R15, R16, R20, R21, R24a, R24b Detrital k-feldspar, quartz, plagioclase, biotite altered in varying degrees to alunite and opal with occasional native sulfur

R22 - K-feldspar, plagioclase, biotite, quartz altered to kaolinite, K-mica, montmorillonite (nontronite?)

R25 - K-feldspar, plagioclase, quartz, biotite detritus unaffected by hot spring activity. Contains trace of montmorillonite

Table 4 Continued

	R20	R21	R22	R24a	R24b	R25
SiO ₂	65.5	53.1	65.3	70.2	40.5	62.5
Al ₂ O ₃	13.2	17.5	13.3	10.1	19.7	14.9
Total Fe as Fe ₂ O ₃	.4	.5	7.5	7.8	1.1	3.1
MgO	.1	ND	2.4	.7	ND	.6
CaO	1.6	.04	0.6	0.9	0.2	0.9
Na ₂ O	.4	.4	1.2	1.9	.6	3.7
K ₂ O	3.4	5.3	3.1	3.3	6.0	5.6
SO ₃	11.6	14.2	0.15	3.7	21.4	0.12
Loss on Ignition	18.1	20.2	6.9	6.9	27.4	2.1
Total**	102.7	97.04	100.3	101.8	95.5	93.4
Bulk Density	1.61	2.05	1.94	1.89	2.15	-
As ppm	ND	ND	150	16	14	ND
Pb ppm	61	39	ND	30	41	40

Surficial alteration characteristically consists of bleached or iron stained, acid-leached alluvium. Intensity of surface alteration varies from detritus with glassy feldspars present to altered rock retaining only relic detrital texture and consisting of opal, alunite, and a trace of native sulfur. Analyses R14, R15, R16, R20, R21, R24a and R24b in Table 4 are representative of surficial alteration and R25 is an example of unaltered alluvium. Alunite and opal occur as veinlets and patches within the rock matrix and as replacements of feldspar grains. Native sulfur occurs as small individual grains intergrown with alunite and opal and as pore filling in coarse detritus. A second alteration mineral assemblage occurs in one outcrop (R22) and consists of kaolinite, K-mica, and green montmorillonite (nontronite) with plagioclase, and strained quartz. Alunite, native sulfur and sulfides are absent in this assemblage.

Subsurface Alteration

Access to subsurface alteration is provided by core from one drill hole located on Figure 13. Bulk chemical analysis of five foot intervals of Core 1A are included in Table 5 and clay mineralogy is shown in Figure 14. Parent rock for the alteration is alluvium to a depth of 115 feet (35m) and Precambrian (?) gneiss from 115 feet (35m) to 217 feet (66m). Alteration consists of opal and alunite with occasional native sulfur to 60 feet (18.3m), kaolinite and alunite to 80 feet (24.4m); kaolinite, montmorillonite, and alunite to 100 feet (30.5m); and kaolinite montmorillonite, K-mica, and pyrite to the total depth of 217 feet (66.1m). The interval 28 feet to 50 feet is predominantly opal with coarsely crystalline native sulfur.

Kaolinite in the interval 60' to 105' is moderately ordered and preliminary scanning electron microscopy indicates the presence of some material with tubular morphology. No tubular kaolinite was detected below 105 feet.

Table 5. Chemical Analyses of Core from U of U Diamond Drill Hole No. 1A

Sample Interval	20-28	28-38	38-45	45-50	50-55	55-60	60-65	65-70	70-75	75-80
SiO ₂	67.2	94.94*	86.3	92.15*	42.7	36.9	51.9	49.0	45.5	54.0
Al ₂ O ₃	10.1	1.3	2.1	2.4	16.6	18.9	13.6	15.8	18.5	14.7
Total Fe as Fe ₂ O ₃	0.4	0.1	0.2	0.2	0.1	0.1	1.3	1.3	0.7	2.4
MgO	0.0	0.0	0.1	0.1	0.0	0.0	0.0	0.0	0.0	0.0
CaO	0.1	0.0	0.0	0.0	0.0	0.0	0.1	0.1	0.1	0.1
Na ₂ O	0.2	0.1	0.1	0.1	0.2	0.5	0.4	0.5	0.2	0.2
K ₂ O	2.7	0.1	0.1	0.3	4.8	5.2	3.4	2.4	1.0	1.9
S as SO ₃	9.5	0.3	9.7	1.2	19.0	21.7	11.8	9.6	2.4	1.2
S as S			3.9		7.6	8.7				
H ₂ O ⁺		1.55**		2.10						
Total										
As ppm	ND	ND	ND	ND	ND	ND	ND	22	33	96
Pb ppm	365	85	97	81	154	110	92	95	128	45

* gravimetric

** penfield tube

Table 5 Continued

Sample Interval	80-85	85-90	90-95	95-100	100-105	105-110	110-115	115-120	120-125	125-130	130-135	135-140
SiO ₂	49.6	47.1	52.6	48.8	47.6	50.3	57.4	51.8	52.3	51.2	49.3	50.7
Al ₂ O ₃	15.6	13.5	11.0	13.3	10.5	13.9	12.1	15.0	14.4	14.1	12.5	13.9
Fe ₂ O ₃	6.0	3.8	2.8	2.7	466	1.2	3.9	2.9	1.9	2.9	4.4	3.0
MgO	0.0	0.2	0.5	0.3	0.9	0.8	1.1	0.6	0.7	0.8	0.9	0.9
CaO	0.1	0.1	0.3	0.5	0.3	0.4	0.3	0.3	0.3	0.6	0.8	0.8
Na ₂ O	0.4	1.6	1.3	1.4	0.6	1.3	0.3	0.3	0.3	0.4	0.3	0.3
K ₂ O	3.2	5.2	4.0	4.8	4.7	4.7	2.5	4.0	2.8	2.4	3.3	2.6
S as SO ₃	3.3	8.6	5.7	7.8	4.8							
						0.7	1.0	1.0	0.5	1.1	1.7	1.0
As ppm	99	30	15	12	24	11	22	ND	12	22	41	22
Pb ppm	42	68	48	45	40	43	30	58	26	25	52	48

Table 5 Continued

Sample Interval	140-145	145-150	150-155	155-160	160-165	165-170	170-175	175-180	180-185	185-190	190-192	192-197
SiO ₂	52.0	51.8	47.4	58.2	45.5	45.2	61.2	58.3	48.7	57.3	42.3	51.1
Al ₂ O ₃	13.6	13.2	12.5	10.3	14.0	12.2	10.2	12.0	13.0	13.0	15.6	17.7
Fe ₂ O ₃	3.0	2.6	6.3	4.5	4.2	8.7	4.6	2.9	5.6	3.0	4.8	5.1
MgO	0.5	1.0	1.6	1.2	1.5	1.6	1.4	0.7	1.2	0.4	1.0	0.4
CaO	0.4	0.7	0.6	0.4	0.8	0.7	0.4	0.3	0.6	0.2	0.6	0.3
Na ₂ O	0.3	0.3	0.3	0.4	0.4	0.3	0.3	0.3	0.3	0.3	0.4	0.2
K ₂ O	2.6	2.7	3.0	3.7	2.0	2.4	1.7	2.4	2.0	3.3	2.9	2.1
S as SO ₃												
S as S	1.4	0.8	2.2	2.0	1.0	3.8	1.9	1.2	2.4	1.5	1.0	2.5
As ppm	38	24	18	77	26	47	58	96	138	142	103	166
Pb ppm	60	25	36	20	ND	76	94	156	223	230	166	269

Table 5 Continued

Sample Interval	197-201	201-207	207-211	211-217
SiO ₂	54.8	57.8	46.0	52.3
Al ₂ O ₃	12.0	10.6	16.7	15.0
Fe ₂ O ₃	7.2	8.3	5.9	4.8
MgO	1.3	1.1	1.1	1.1
CaO	0.3	0.3	0.2	0.2
Na ₂ O	0.3	0.2	0.3	0.2
K ₂ O	3.4	2.7	4.4	2.4
S as SO ₃				
S as S	3.1	4.3	1.2	0.8
As ppm	57	50	23	95
Pb ppm	93	82	38	154

Figure 13. Surface geology in the vicinity of Roosevelt Hot Springs showing location of soil traverse lines and diamond drill hole. Geology from Petersen (1975).

UNIVERSITY OF UTAH DDH No 1A

Figure 14. Diamond Drill Hole 1A logs. Clay mineralogy shown is relative percent of total clay.

Montmorillonite occurs as a constituent of interstratified clay and as a discrete mineral. Interstratified clay with a basal spacing following vapor glycolation of 13.5A occurs in the interval 85 feet to 105 feet with kaolinite and alunite. Interstratified clay with a basal spacing following vapor glycolation of 27A occurs in the interval 185 feet to 217 feet with kaolinite, K-mica, and discrete montmorillonite. Montmorillonite with a basal spacing following vapor glycolation of 17A occurs in the interval 105 feet to 185 feet with kaolinite and K-mica.

Vertical hydrothermal alteration mineral zoning is distinctly apparent. The surface assemblage alunite-opal grades successively with depth into alunite-opal-kaolinite, then to alunite-opal-kaolinite-mixed layer clay, then to montmorillonite-kaolinite-k-mica. The alteration mineral assemblage can best be explained in terms of the stability relationships shown in Figures 15 to 19 in comparison with the chemistry of water at Roosevelt. Analyses of surface springs, seeps, and deep well water (Table 6) shows that the solutions are relatively low ionic strength (.1 to .2) sodium chloride brines. Hydrogen sulfide analyses are not available, but freshly collected water from Phillips 9-1 well had a strong hydrogen sulfide odor. The surface alteration assemblage containing alunite is best explained as a result of a decrease in pH of near surface fluids as sulfide oxidizes to sulfate as described by Schoen et al. (1974). Increasing temperature, potassium ion activity, and pH of the water with depth would produce the observed alteration zoning. With increasing temperature the value of $\log K/H$ for equilibrium between K-feldspar and K-mica and for equilibrium between K-mica and kaolinite or montmorillonite decreases and the stability field of K-mica increases. At the temperatures and water compositions of the deep geothermal reservoir (Table 6, analysis 6), stability of K-feldspar would be expected though no alteration data are available for confirmation.

Figure 15. Activity diagram showing the stability relations of alunite, K-feldspar, K-mica and Na-Montmorillonite in the presence of quartz and unit activity of water at 100°C.

Figure 16. Activity diagram showing the stability relations of alunite, K-feldspar, K-mica, and kaolinite with quartz and unit activity of water at 100°C.

Figure 17. Activity diagram showing stability relations of K-feldspar, K-mica, kaolinite, and alunite with quartz and unit activity of water at 200°C.

Figure 18. Activity diagram showing stability relations of K-feldspar, K-mica, Na-montmorillonite and alunite with quartz and unit activity of water at 200°C.

Figure 19. Equilibrium constants of K-feldspar-K-mica and kaolinite-K-mica reactions as a function of temperature.

Table 6. Roosevelt KGRA Water Analyses

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)
Na	1840 mg/l	2400	1800	2210	2400	2000	2437	2080	2500
Ca	122	113	107	83	9	10.1	8.0	19	22
K	274	378	280	425	565	410	448	472	488
SiO ₂	173	76	107	364	775	560+	560?	405	313
Mg	25	17	24	-	19	.24	.01	3.3	0
Cl	3210	3800	3200	3800	4800	3400	4090	3810	4240
SO ₄	120	142	70	122	200	54	59	65	73
HCO ₃	298	536	300	-	-	200	180	158	156
S ⁼	<.03								
O ₂	5								
Sr	2								
Li		-	17		18	19	20		.27
B		37	29		45	29	25		38
Br					7				3.3
As					3.5				
Mo					.04				
Ag					.09				
Cu					.03				
Pb					.10				
Zn					.04				
Co					.15				
Ni					.18				
Cr					.01				
Mn					.15				

Table 6 Continued

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)
F	-	5.2	3.3			5	5	7.1	7.5
NO ₃		Trace	Trace			Trace	.1	1.9	11
Total Dissolved Solids		7506	5948			6442	7067	7040	7800
Density					1.004				
Temp.	25%	17 ⁰	28			+260 ⁰ C	205+ ⁰ C	81 ⁰ C	51 ⁰ C
pH	6.5	8.2	6.43			6.5	6.3		7.9
Temperature Estimates									
Quartz Saturation			141			263+	263?		
Amorphous Silica						139+			
Na-K-Ca	241 ⁰ C	247	239	262		294+	273		

- (1) Roosevelt seep. University of Utah Analyst 6-25-75
- (2) Roosevelt seep, Phillips Petroleum Co. 5-9-73
- (3) Roosevelt seep. Phillips Petroleum Co. 8-15-75
- (4) Phillips well 9-1, University of Utah
- (5) Phillips well 54-3, Chemical and Mineralogical services analyst
- (6) Phillips well 54-3, Phillips Petroleum Co. 8-26-75
- (7) Phillips well 3-1, Phillips Petroleum Co. 5-25-75
- (8) Roosevelt Hot Springs, Mundorff (1970) 11-4-50
- (9) Roosevelt Hot Springs, Mundorff (1970) 9-11-57

Heavy Metals - Mercury

A field-portable mercury sniffer (Scintrex HGG-3) was used to test soil gas at Roosevelt for mercury. The detection limit for the device as specified by Scintrex is approximately 40×10^{-12} g. mercury. Soil gas was tested for mercury by first driving a metal rod into the ground removing the rod, inserting the sniffer probe which sealed the hole, extracting soil gas with a small hand pump and injecting the gas into the absorption chamber of the sniffer for analysis. Mercury is detected using an atomic absorption technique with background and interference corrected using absorption of zeeman lines from the mercury lamp.

Mercury was not detected in soil gas in a traverse across the opal dome at about 1300 meters north. Mercury was not detected in soil gas in a traverse down Negro Mag Wash nor in the resort area.

An attempt was then made to determine mercury in the solid phases of the soil. Soil samples were first sieved to remove the +80 mesh fraction which was discarded, then approximately 250 mg (volumetric measure used) of -80 mesh soil was added to a test tube, the tube was stoppered and connected to the mercury sniffer sampling hose. A propane torch was used to heat the soil in the test tube for 30 to 45 seconds. The sniffer pump was operated (1.5 to 2 inch stroke) to inject the vapor into the absorption cell and the meter deflection was read. Three soil sample traverses with samples collected on 100 foot spacing were tested for mercury. Traverse number one was in Negro Mag Wash over an area of sulfur and old spring activity. Anomalous mercury values were detected in three samples near sulfur prospect pits. In a similar procedure anomalous mercury values were detected in a soil sample traverse at the resort area and the opal dome. However, no anomaly was present in more than one or two samples in the 100 foot sample spacing.

The preliminary field survey using the Scintrex portable sniffer was

followed with collection of soil samples on three traverses on 10 meter spacing and a sample depth of 15 to 25 cm. These samples were returned to the laboratory, air dried, sieved, and analyzed using a thin gold film detector (McNerny et al., 1972) made available by Kennecott Exploration Services. A soil profile was also sampled at coordinates 4000 meters north on the baseline (Figure 13). Samples were collected at 5 cm intervals to a depth of 40 cm, dried, sieved and analyzed using the thin gold film detector. The mercury distribution with depth is shown in Figure 20 and Table 7. Mercury increases from 414 parts per billion (ppb) at the surface to a maximum of 1090 ppb at 25 cm depth, confirming the optimum depth of 15 to 25 cm for the remaining collection. Analytical precision was determined by 10 analyses of sample 4000 North 10 West which showed 357 ppb with a standard deviation of 11 ppb.

The mercury analyses of soil samples from the traverses at 1300 North, 4000 North, and 5000 North are shown in Figures 21, 22, and 23 and Tables 8, 9, and 10. Mercury anomalies reach a maximum level of nearly 3000 ppb on the 4000 North line (Table 9), 1300 ppb on the 1300 North line (Table 8), and 639 ppb on the 5000 North line (Table 10). Anomalous mercury values are restricted to a few adjacent samples in most cases and correlate very well with subsurface fractures detected by a variety of geological and geophysical methods (Ward et al., this series). No widespread soil-mercury anomaly was found. Soil samples were collected in a traverse crossing an inferred fault in section 8 (Figure 24), but no anomalous mercury values were detected in this area (Table 11, Figure 24).

Four samples from Thermo Hot Spring were also run for mercury and they contained 3850 ppb, 820 ppb, 6 ppb, and 23 ppb indicating the presence of anomalous mercury in three of the four samples.

DEPTH PROFILE 00E

Figure 20. Mercury content of soils as a function of depth at location 4000 meters north on the University of Utah baseline.

Table 7. Mercury in Soils from the Roosevelt KGRA,
4000 North Line Depth Profile at 00 East

<u>Depth Cm</u>	<u>Hg ppb</u>
0	414
5	511
10	909
15	804
20	960
25	1090
30	884
35	1030
40	974

Figure 21. Mercury in soil on the 1300 North traverse.

Figure 22. Mercury in soil on the 4000 North traverse.

Figure 23. Mercury in soil on the 5000 North traverse.

Table 8. Mercury in Soils from the Roosevelt KGRA,
1300 North Line

<u>Meters</u>	<u>Hg ppb</u>	<u>Meters</u>	<u>Hg ppb</u>	<u>Meters</u>	<u>Hg ppb</u>
20E	39	600W	131	860W	109
10W	72	606W	435	870W	43
40W	75	646W	83	880W	35
70W	55	660W	425	940W	109
100W	77	670W	150	950W	619
130W	278	680W	111	960W	1190
160W	65	690W	116	960WDUP	1300
190W	57	690WDUP	147	970W	152
220W	74	700W	188	980W	223
250W	107	710W	132	990W	48
280W	87	720W	116	990WDUP	10
310W	258	730W	194	1000W	74
340W	304	740W	534	1030W	57
370W	332	740WDUP	472	1050W	80
370WDUP	446	750W	98	1060W	80
400W	36	760W	1410	1090W	87
430W	529	770W	123	1120W	43
460W	160	780W	149	1150W	69
490W	218	790W	65	1180W	60
500W	282				
510W	191	800W	404	1210W	64
520W	100				
530W	94	800WDUP	442		
540W	155				
550W	200	810W	87		
560W	153	820W	61		
570W	172	830W	27		
580W	228	840W	137		
590W	99	850W	64		

Table 9. Mercury in Soils from the Roosevelt KGRA,
4000 North Line

Meters	Hg ppb	Meters	Hg ppb	Meters	Hg ppb	Meters	Hg ppb
600E	47	120E	118	10W	355	270W	227
570E	40	120E	118	10W	353	300W	136
540E	46	110E		20W	138	330W	124
510E	52	100E	116*	20W	372	360W	128
480E	29	100E	120	30W	364	390W	76
450E	27	90E	113	30W	287	420W	75
420E	130	90E	122	30W	325	420WC	90
390E	90	80E	66	30W	309	450W	52
		80E	67	40W	244*	450WC	90
		70E	170	40W	228*		
		70E	155	40W	252*		
		60E	166*	50W	287		
360E	158	60E	135*	60W	294	480W	70
330EC	401	60E	178*	60W	261		
330E	177	60E	164*	60W	161		
		60E	2940	60W	159	510W	66
		60E	2960	60W	121*	540WC	65
		50E	881	60W	169*	540W	85
300E	275	50E	959	60W	160*		
		40E	728				
		40E	763	70W	154	570W	85
300E	292*	40EC	987	80W	112	600W	72
270E	270	30E	501				
240E	65	30	473	90W	136		
230E	40	20E	2110	100W	73*		
230EC	50	20E	2520	100W	65		
220E	61	10E	1530	110WC	938		
210E	223	10E	1720	110W	578		
		10E	1520				
200E	67	00E	509*	120W	178		
190E	149	00E	495*	130W	86		
180E	97	00EC	1010	140W	87		
		00E	1160	150W	176		
170E	327	00E	1140	160W	306*		
160E	1120	00E	1210	160W	271		
160E	2380	00E	1040	170WC	470		
160E	1760			170W	442		
150E	168	10W	397	180W	106		
		10W	359	190W	115		
140E	102	10W	371	200WC	139		
		10W	339	200W	153		
		10W	351	200W	171*		
		10W	314	210W	122		

<u>Meters</u>	<u>Hg ppb</u>	<u>Meters</u>	<u>Hg ppb</u>	<u>Meters</u>	<u>Hg ppb</u>	<u>Meters</u>	<u>Hg ppb</u>
130E	111	10W	356	240W	622		
		10W	353				

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*Collected Oct. 1975. Stored in glass bottles. All others collected June 1975 and stored in porous olefin bags.

C - Coarse fraction

Table 10. Mercury in Soils from the Roosevelt KGRA,
5000 North Line

<u>Meters</u>	<u>Hg_ppb</u>	<u>Meters</u>	<u>Hg_ppb</u>
320E	42	1070E	65
350E	58	1100E	43
380E	80		
410E	84		
440E	52		
470E	146		
500E	50		
530E	92		
560E	87		
590E			
590E	53		
620E	122		
650E	164		
680E	85		
710E	58		
740E	92		
770E	55		
800E	63		
830E	95		
860E	99		
890E	101		
920E	90		
950E	292		
980E	639		
1010E	85		
1040E	72		

Figure 24. Mercury in soil on the traverse in Section 8.

Table 11. Mercury in Soils from the Roosevelt KGRA.

Time Crossing Inferred Fault in Section 8, T27S, R9W

<u>Meters From Fault</u>	<u>Hg ppb</u>	<u>Meters From Fault</u>	<u>Hg ppb</u>
150E	25	70	39
140E	21	80	60
130E	29	90	69
120E	43	100	31
110E	88	110	52
100E	24	120	83
		130	35
90E	65	140	96
80E	56	150	25
70E	113		
60E	74	Thermo Mud	820
50E	32	Thermo #4	23
40E	37	Thermo #1	630
30E	93	Thermo Mud	3850
20E	42		
10E	48		
0E	56		
10W	46		
20	55		
30	41		
40	34		
50	16		
60	60		

Heavy Metals - Arsenic and Lead

Arsenic and lead were determined on all outcrop samples, core from U of U diamond drill hole No. 1A (see Figure 12 for location) and soil samples from the 4000 meter north line by x-ray emission analysis using a procedure modified from Fabbe and Espos, 1972. Analytical conditions are shown in Table 3. Due to direct superposition of lead L-alpha and Arsenic K-alpha, a correction for lead was necessary. The ratio Pb L-beta to Pb L-alpha was measured using synthetic standards containing varying amounts of Pb but no As. Lead L-beta was measured, the intensity of Pb L-alpha was calculated using the predetermined ratio and subtracted from the total counts at the Pb L-alpha-As K-alpha two theta position to give arsenic counts. Arsenic standards were prepared by adding known amounts of arsenic to pulverized Little Cottonwood Canyon quartz monzonite. Arsenic and lead content of outcrop and core samples are in Tables 4 and 5. No arsenic and only background lead was detected in soil samples. One outcrop sample contained 1760 ppm arsenic (R6) and the maximum arsenic in core was 166 ppm. Maximum lead was 356 ppm in one core sample. Geothermal reservoir fluids contain 3.5 ppm arsenic 0.1 ppm lead (Table 6, analysis 5).

Gamma Ray Spectrometry

An exploration, portable, four channel gamma ray spectrometer was used to determine U, Th, K, and total gamma ray counting rate for various rock types in the Milford area. The data reveal differences in rock type, metamorphism, alluvium, and hot spring deposits. Optimum detector height above ground, counting time, and precision were determined on one granite outcrop. A height of 400 cm and four minute counting time were sufficient. Data collected on granite, metamorphic rocks, alluvium, and volcanic rocks are shown in Table 12.

Table 12. Gama Ray Spectrometric Measurements of various Rock Types
in the Milford Area

Rock Type	SITE	AVE. TOTAL Counts	AVE. K Counts	AVE. U Counts	AVE. TH Counts
Granite	1	15475	1437	299	191
Granite	1	15599	1466	297	190
Granite	1	15438	1458	294	212
Granite	T18	15209	1395	305	182
Granite	2	13796	1346	255	122
Granite	16	14197	1627	246	76
Granite	4	12633	1227	226	130
Granite	5	12798	1280	200	88
Granite	Ave for all	14393	1404	265	149
Granite		$\sigma = 1220$	$\sigma = 124$	$\sigma = 39$	$\sigma = 52$
	Ave Site 1	15504	1453	297	198
		$\sigma = 88$	$\sigma = 15$	$\sigma = 3$	$\sigma = 12$
Alluvium	1-15	12673	1153	239	112
Alluvium	16-27	10450	1105	176	82
Alluvium	28-38	10746	1136	178	107
Alluvium	1-38	11417	1133	202	101
Metamorphic Rocks	16	9115	848	178	94
	N	12060	998	263	185
	M	16878	1208	419	304
	K	17483	1428	390	295
Volcanic					
Obsidian	6	14137	1335	287	125
Obsidian	T2	15999	1584	332	122
Lt. Gray	7	12782	1165	265	121
Basalt	A1	6165	531	108	66
Perlite	T1	14313	1446	290	129
Perlite	T21	16308	1333	381	72
Perlite	T22	17870	1447	447	229
Pumice	T20	14563	1190	355	186
Pumice	T23	17830	1461	454	243

A traverse was made across the opal dome at 1560 meters north and the results are shown in Figures 25, 26, and 27. The gamma ray counts show depletion of U, Th and K near the center of the opal dome with values rising to normal counts as the proportion of alluvium to spring deposits increases.

Geothermal Exploration

Several geochemical-mineralogical characteristics of potential geothermal resources in the Great Basin of Utah have been examined in detail in preceeding pages including warm spring chemistry, cold spring and well chemistry, alteration mineralogy, and heavy metal distributions. These characteristics provide exploration guides of varying utility. The most effective regional scale exploration guide is chemical geothermometry of water, but careful geological evaluation of each chemical thermometer is required. Hot spring deposits, altered rocks and heavy metal distributions are useful exploration guides on a local rather than a regional scale. The presence of siliceous sinter, various types of opal, advanced argillic alteration, anomalous mercury, lead, and arsenic are typical of many geothermal areas and are considered favorable indicators of a nearby geothermal resource. The extent and nature of alteration can be used to determine the extent of fluid leakage to the surface and to estimate subsurface water chemistry parameters and alteration. Water salinity presence of conductive clays and sulfides are important geochemical-mineralogical parameters for evaluation of geophysical resistivity data.

Figure 25. Gamma ray total counts in traverse across the opal dome.

Figure 26. Potassium gamma ray counts across the opal dome.

Figure 27. Uranium gamma ray counts in traverse across the opal dome.

MONROE HOT SPRINGS

The Monroe Hot Springs Known Geothermal Resource Area (KGRA) was chosen for a detailed alteration study on the basis of accessible thermal waters, high Na-K-Ca wall rock equilibration temperatures and extensive hydrothermal alteration which is exposed in nearly 1000 feet of relief. The petrology and geochemistry of the surficial alteration were examined in order that boundary conditions might be established for the temperature and composition of the altering fluids. The characteristic pattern of alteration associated with this area was also studied so that it might be compared with alteration prevalent in other proven and prospective geothermal resources.

The Monroe KGRA occupies the eastern flank of the Central Sevier River Valley, immediately to the east of the town of Monroe, Utah, (Figure 28). Warm springs and seeps outcrop for approximately a mile, both north and south of Monroe.

The Central Sevier River Valley is bounded to the east by the steep, deeply dissected monocline of the Sevier Plateau and to the west by the Pavant Range. The Antelope Range, a series of low hills, crosses the valley about ten miles south of Monroe.

The Bullion Canyon volcanics of Miocene age comprise the oldest volcanic rock in the area. They are composed of pyroclastics in the lower part, overlain by thick porphyritic latite flows. The total thickness is estimated at 4000 feet (Callaghan, 1961). Latite, basaltic andesite and rhyolite of the Dry Hollow, Joe Lott and Mount Belknap formations, all of Pliocene age, overlies the Bullion Canyon volcanics.

Quartz monzonite intrudes the Bullion Canyon volcanics at Monrovia Park, three miles southeast of Monroe. This stock, about one mile in diameter, is

Figure 28. Index map of the Monroe KGRA.

typical of other Miocene intrusives that have invaded the Bullion Canyon volcanics in the Marysvale district.

Tertiary volcanics are underlain by sedimentary rocks. The Jurassic Carmel formation, an equivalent of the Arapien shale, locally underlies the Bullion Canyon volcanics and is composed of stratified salts, lenticular gypsum beds and thin bedded limestones and shales. This formation attains a maximum thickness of 2500 feet in the Central Sevier River Valley between Gunnison and Richfield before it disappears under the Tertiary volcanic cover (Hardy, 1952).

This region is typified by high angle normal faulting. The Sevier fault, from which the warm springs issue, terminates the west dipping monocline of the Sevier Plateau. The trace of this fault is mostly buried beneath alluvium. Other major range front faults in this locality are the Elsinor fault and the Dry Hollow fault, which is associated with the Joseph Hot Springs area. The Antelope Range is the expression of an anticlinal cross-structure which extends eastward from the Tushar Mountains and terminates against the Sevier fault.

Several major landslides have formed along the shoulder of the Sevier Plateau. The largest of these is the Thompson Creek Slide which involves the Dry Hollow formation immediately adjacent to the Monroe KGRA.

Figure 29 shows the disposition of the Monroe KGRA relative to the Sulphurdale and Roosevelt Hot Springs KGRA's. The linear arrangement of these three areas may reflect a major deepseated cross-structure which is associated with mineralization and hydrothermal activity in central Utah. The youngest volcanic rocks outcrop in the vicinity of the Roosevelt Hot Springs KGRA; however late Tertiary rhyolitic activity is typical of the

Figure 29. Regional map showing Roosevelt, Sulphurdale, and Monroe KGRA's and major volcanic units.

entire region. The Mount Belknap rhyolite is the most extensive of these young acid volcanics.

Callaghan and Parker (1961) mapped the geology of the Monroe quadrangle which includes all of the area within the Monroe KGRA. They recognized two mappable units within the Bullion Canyon volcanic formation. They did not differentiate the lower pyroclastic sequence. Callaghan (1938, 1939) has described the stratigraphy and mineralization of the Marysvale volcanic sequence in two papers.

An area of three square miles adjacent to the Monroe and Red Hill springs was mapped during a six week period in the late summer of 1975. Mapping was done on an areal photo base on a scale of 1:4800. General geology, hydro-thermal alteration, and sample locations were included in the map. This information was later transferred to a topographic base (U.S.G.S. topographic quadrangle). The geologic and alteration maps appear in Figures 30 and 31 respectively.

Stratigraphy

The pyroclastic member of the Bullion Canyon formation which outcrops over most of the KGRA has been divided into four mappable lithologic units. Unit A is comprised of volcanic conglomerate, agglomerate, breccia, tuff and a few thin flows. The conglomerate contains angular to subangular heterogeneous pebbles and cobbles up to six inches in diameter. Fresh outcrops of a dark chocolate brown conglomerate are separated from leached light gray exposures by an irregular contact resembling a solution boundary. This contact is clearly visible in Order Canyon (location 302) on Figure 4. Clasts are frequently a bright green as a result of propylitic alteration. The conglomerate fractures across clast boundaries. It weathers into gentle

Figure 30. Geologic map of the Monroe KGRA.

Figure 31. Alteration map of the Monroe KGRA. Numbers refer to the sample collection sites referred to in the list.

slopes. The clastic texture may not be distinct on weathered surfaces. A thin bedded, arenaceous gray tuff, which intervenes in the conglomerate has been mapped as unit 'A'.

Agglomerate lies above the conglomerate. It forms gray cliff faces and weathers into massive blocks. In Order Canyon it contains chocolate brown stringers and veinlets (location 118). In altered outcrops it may be indistinguishable from conglomerate. Agglomerate grades upward into coarsely brecciated flows. This breccia may be gray to rust colored and outcrops irregularly in narrow ledges.

A rust colored crystal tuff is irregularly exposed. It weathers into low hills and gentle slopes, but occasionally forms small ledges. It is considered to represent the uppermost horizon of unit A. Brecciated outcrops of this tuff are common (location 197).

Thin, dark reddish-brown prophritic flows, several feet to several inches in thickness, intervene sporadically between the conglomerate and agglomerate (location 232).

Unit B is composed of interbedded tuff and volcanic conglomerate. The conglomerate is distinguished from that of unit A by well rounded cobbles and boulders and by its distinct bedding which is frequently graded. It fractures along clast boundaries and the rounded cobbles tend to weather out of the matrix. The conglomerate is invariably gray in appearance and forms steep cliffs and ledges. Alternating with the conglomerate is a thin, bedded, arenaceous gray tuff which is indistinguishable from the gray tuff in unit A. Channel fill structures are common among these tuffs and conglomerates. Due to its pronounced bedding, this unit afforded the only reliable measurements of attitudes in the area. The best exposures are available on

'SS' Mountain east of the Monroe spring.

Unit C is a dark purplish gray porphyritic latite flow, believed to be identical to Callaghan's Monrovia Latite member (Callaghan, 1961). It forms prominent shelves and tables in the southern part of the KGRA. It attains a maximum thickness of approximately 150 feet in Sand Canyon (location 51). Broad scree slopes generally lead up to well jointed outcrops. Weathered surfaces may look brown as a result of oxidation. The flows grade continuously into aphanitic dikes.

Unit D is a massive, pink to light purple crystal tuff. It is distinguished from the crystal tuff of unit A by containing accidental fragments of dark brown effusive rock and peculiar white patches, presumably the relic outlines of lapilli. Large biotite phenocrysts may give the rock a salt and pepper appearance. This tuff forms steep hills and rounded cliffs and weathers into large massive blocks.

Aphanitic dikes outcrop in all four pyroclastic units. All dikes appear to be cogenetic with the porphyritic latite of unit C. The most conspicuous dikes have been sampled at locations 1, 13, 39, 20, 22 and 6. All weather to a light tan and many are markedly foliated. "Picture rock" type staining is common in intensely altered zones and the townspeople frequently use dike rock, so altered, for decoration. As a rule, aphanitic dikes are no more than several feet in width while ophitic dikes are fifteen feet wide or better. All dikes trend north to northwest and dip at angles of 30 to 35 degrees to the east and northeast. It is, perhaps, more than a coincidence that a dike intersects each one of the active spring sites. Possibly, the dikes occupy fractures which continue to be important fluid pathways.

A rust colored porphyritic latite flow, Callaghan's Calcic Latite member (Callaghan, 1961), which outcrops to the east of the pyroclastic

member of the Bullion Canyon volcanics has been designated unit E. In this locality the Calcic Latite can be distinguished from the Monrovia Latite by its deep rust color and by its more coarsely porphyritic texture. It forms broad ridges with gentle slopes which are generally littered with fine skree.

The only other formation which outcrops in the area is the Dry Hollow formation. In this locality it is comprised of dark brown porphyritic andesite flows. It can be distinguished from the Calcic Latite only with difficulty. The Dry Hollow formation forms a steep cliff at the northern edge of the area.

The pyroclastic units vary dramatically in thickness across the KGRA and often rest disconformably upon one another. The implication is that most, if not all, the pyroclastic units were deposited on an irregular terrain. In particular, the porphyritic latite of unit C appears to have flowed along old drainages. A striking disconformity between units A and D located east of "SS" Mountain testifies to the degree to which the terrain was dissected at the time of the deposition of the massive crystal tuff. In addition, the peculiar conglomerates and tuffs of unit B, whose clasts have the appearance of alluvially worked sediments, projects a youthful terrain.

A small outcrop of unsorted and poorly consolidated conglomerate outcrops at location 192, Figure 31. It contains extremely angular clasts of volcanic rock, some of which approach a foot in diameter. From all appearances this is a very young conglomerate and probably belongs to the Sevier River formation.

Spring Deposits

With the exception of perennial warm spring seeps, the warm springs in the vicinity of Monroe are confined to two large tufa mounds. Red Hill Spring

issues from a fissure in the northern-most and smaller of these mounds. It possesses the highest temperature and flow rate of any spring in the area. The flow, which is strong and constant, has been estimated at 200 gallons per minute (Mundorf, 1970). The Monroe mound is over 500 yards in length and contains myriad transient springs and seeps. The total flow from the mound, however, is comparable with the Red Hill Spring. The owner of the property has trenched the mound to increase and direct the flow. The temperatures of these springs varies between 48° and 69° centigrade, compared to 77° centigrade at Red Hill Spring. Perennial seeps outcrop at all the tufa deposits along the range front. The ground in these areas is often warm, even when there is no visible flow. Ground temperatures as high as 35° centigrade were measured at the mouth of Sand Canyon.

Structure

All hydrothermal alteration is confined to the pyroclastic member of the Bullion Canyon volcanic formation. The major faults which bring these rocks into contact with the Calcic Latite member and the Dry Hollow formation terminate the thermal area on the east and north respectively. Together these two faults define a block, containing the pyroclastics, which is displaced upward relative to the adjacent terrain. This block is characterized by intense fracturing and closely spaced echelon normal faults. In many cases it was impossible to elucidate these detailed structures. Wherever displacements are apparent, faults have been mapped. These traces, while not in and of themselves sufficient to reconcile all the geology, are adequate to characterize the pattern of faulting in the KGRA.

Three sets of faults and fractures can be generalized; one trending northwest, a second northeast and a third north. The first two sets are responsible for the significant displacements in the area. The patterns of

hydrothermal alteration, however, are controlled by the latter two sets. In the vicinity of the Red Hill Spring and throughout the central part of the area, travertine veins and fractures have a remarkably uniform trend north and steep dips of between 30 and 40 degrees. At Red Hill Spring the most intense alteration occurs at the intersection of west dipping fractures and the major northeast trending faults. Echelon faults at location 175, near the spring, dip 30 degrees to the east. The distribution of hydrothermally altered outcrops between locations 187 and 82 highlights a well developed set of north trending fractures. The aphanitic dikes in this area also reflect this local fracture pattern.

In the southern half of the area the north trending fractures dwindle to insignificance and northeast trending fractures dominate. The most extensive alteration found in the KGRA occurs in this region. Here the dikes, many of which are doloritic, do not follow the local fractures but dip to the northeast. In Bertherson Canyon, south of the mapped area and separated from it by a major fault, northwest trending fractures become important in controlling hydrothermal alteration. Despite the changing trends in structure, the Bullion Canyon volcanics appear to maintain a reasonably uniform southwestern dip between Order and Monrovia Canyons.

The regions mapped as quaternary landslide (Qls) on the range front are disrupted outcrops of units A and B. This rock is apparently draped over hidden range-front faults. Slumping of large volumes of this rock has produced chaotic attitudes and contacts in these areas. Travertine veins are densely distributed in this confused terrain. An arcuate trace represents the inferred range fault(s) associated with this structure.

Large slump blocks of unit B occur on the slope south of location 202. The clusters of tilted junipers which have grown from these blocks testify

that this is a contemporaneous feature.

The massive Thompson landslide, which occupies the Dry Hollow formation, outcrops at the extreme northeastern corner of the KGRA. It has little bearing upon the structure or geology of the area.

The Miocene intrusive episode that resulted in the emplacement of the quartz monzonite stock at Monrovia Park may have generated the north and northwest trending fractures and faults. The occupation of many of these structures by latite dikes is a convincing argument. The structural complexity of the central part of the area may be the result of the intrusion of a stock at depth beneath the KGRA.

According to Callaghan (1961) the onset of range front faulting is contemporaneous with the deposition of the Sevier River formation which is Pleistocene in age. The most intense alteration is confined to these northeast trending faults and fractures. Therefore, the beginning of the major hydrothermal episode is presumed to have been within the last several hundred thousand years. The thermal system may not have changed significantly and the present day spring geochemistry might reasonably be expected to reflect the geochemistry of the recent past. The heat source responsible for the hydrothermal solutions which have generated an advanced argillic alteration assemblage might not yet be dissipated.

Alteration

The most striking aspect of the alteration within the KGRA is yellow, earthy patches that stand out against the light purple to light gray volcanic host rock. The nearly total involvement of the rock in the southwest corner of the area is clearly visible from the town of Monroe. These patches are generally distributed in distinct linear patterns and for the most part

trend northeast, following the major range front faults. The patches are characterized by white, tan and yellow mottling and invariably contain dendritic veins of gypsum. Large euhedral selenite crystals can commonly be collected from the weathered soils. Calcite veins are less common but are, nonetheless, a frequent feature of these areas. Primary textures may be partially or completely obscured by kaolinite or calcite replacement. The rock is usually banded by hematite stain. The enclosing gray to purple rock shows little variation, except in color, from the patches. It is usually punky and foliated, like the patches, and frequently has undergone extensive carbonate or kaolinite replacement. Aphanitic dikes are generally more resistant than doleritic dikes when involved by this alteration. Where large patches do not occur, yellow bands, several centimeters to several feet in width may surround fractures.

Yellow patches and bands may be the expression of either argillic or advanced argillic alteration, for these two are indistinguishable in outcrop. There is a tendency for argillic outcrops to be more weather resistant and to contain more vein calcite. These criterion are not to be trusted, however.

Less common but more dramatic are argillically altered outcrops which are distinguishable by their bright red, black and white coloration. These outcrops are confined to the northern half of the area and are most common along the fault connecting locations 128 and 189. Rock so altered is an aggregate of quartz, kaolinite and iron oxide. It is unresistant to weathering and is found in deeply dissected exposures. A good example of this alteration is available at location 187. The Red Hill tufa mound, which takes its name from the bright red imparted by contained hematite, appears to be genetically related to these outcrops.

Propylitically altered rock can, in some cases, be recognized by the bright green specks or patches it contains. The green is not produced by chlorite but by montmorillonite, presumably the variety nontronite. Occasionally, propylitic outcrops may be indistinguishable from those of argillic zones. Also, many of the freshest outcrops in this area contain propylitic mineral assemblages. In any case, propylitic alteration is not well developed in this area.

A general pattern of alteration can be discerned in which iron rich argillic assemblages are associated with north-south structures while sulfate dominated argillic assemblages are associated with northeast-southeast-southwest structures. Implied in this pattern is the possibility that the iron rich hydrothermal solutions may have anteceded the sulfate solutions.

The hydrothermal alteration at Red Hill Spring is characterized by west dipping travertine veins, succeeded to the east by zones of kaolinite and calcite replacement. The kaolinite replaced rock is tan to yellow and frequently contains masses of pure white kaolinite. The calcite replaced rock is gray and may contain specks of green montmorillonite. This rock usually has a salt and pepper appearance due to kaolinized andesine phenocrysts. Both zones are very earthy. Where this zonation is obscured by weathering, yellow soil containing abundant small selenite crystals outcrop instead. The flanks of the low hill adjacent to the spring are mantled by gray montmorillonitic "popcorn" alluvium. A similar but less distinct pattern of alteration occurs in the vicinity of Monroe Spring.

Petrography

Hydrothermally altered areas as well as adjacent host rock were sampled and the sample locations recorded on the base map (Figure 31). The specimens

were then examined under a binocular microscope. Based on this examination a collection of 95 representative specimens was assembled. A portion from each of these was pulverized in a Spex shatter box for four minutes. X-ray diffraction patterns were then made for powder pack mounts of each specimen. A Phillips Electronics diffractometer with a copper source and a graphite monochromator was used. Two-micron clay fraction mounts were also made by peptizing with calgon, separating the two micron fraction with a centrifuge, and smearing a centrifuged slurry on frosted glass slides. Thirty such preparations were made and x-ray diffraction patterns recorded for slides which were air dried, glycolated, dehydrated at 300⁰ centigrade, and heated to 700⁰ centigrade.

The composition of each of the 95 specimens was reported in terms of percent total pattern intensity attributable to the major peak of each phase. This data was then stored on a fastrand data file for use in multivariate analysis.

Relying on the x-ray diffraction data, a representative group of 51 specimens was chosen to be made into thin sections. All sections were stained for potassium feldspar. If dolomite was present in the x-ray pattern of a specimen, dolomite was also stained. The sections were examined with a conventional petrographic microscope. Phases which had previously gone undetected by x-ray diffraction but were recognized in thin section were added to the data file nominally at one percent total pattern intensity.

Monroe KGRA

Andesine is the most frequently encountered plagioclase feldspar, with oligoclase or labradorite occurring in less than ten percent of the outcrops sampled. It appears both as large tabular phenocrysts in crystal tuff,

agglomerate and porphyritic latite, and as small laths in the groundmasses of agglomerate, and porphyritic and aphanitic latites. Phenocrysts zoned toward alkali feldspar are common, the most extreme examples being andesines rimmed by potassium feldspar.

Andesine typically alters to calcite, kaolinite or montmorillonite. Occasionally sericite is also found replacing andesine. As a rule the lath-like crystals alter much more readily. The larger phenocrysts alter from their interiors outward and frequently alkali feldspar rims will be preserved unaltered in otherwise kaolinized samples.

Apatite occurs as euhedral hexagonal crystals and is found primarily in crystal tuff or porphyritic latite. It is commonly found as inclusions in fresh and altered biotite phenocrysts and appears to be closely associated with the occurrence of that mineral. It is unaltered in all specimens with the exception of 13II where it alters to gypsum.

Large tabular phenocrysts of biotite with perfect hexagonal cross-sections are abundant in crystal tuff. In agglomerate, however, biotite may appear as small clastic fragments and narrow laths. It is altogether absent from both porphyritic and aphanitic latites. In most specimens the biotite cleavages are warped or broken.

Hematite rims and carbonate are frequently seen splitting the cleavages of biotite phenocrysts even in the freshest outcrops. Biotite is most commonly replaced by montmorillonite, sericite or hematite. However, potassium feldspar and chlorite also have been observed replacing biotite.

Two varieties of hornblende have been observed. One is green, moderately pleochroic and has an extinction angle of 12° . This mineral, perhaps ferrohastingsite, occurs at sample specimens 302 and 14. The first occurrence is in a volcanic conglomerate and the second in aphanitic latite. In both

specimens andesine is unaltered although the hornblende is being replaced by carbonate. The other variety, common hornblende, is brown, displays strong pleochroism and has an extinction angle of 15° . It is known to occur only in crystal tuff of unit D, where it also is partially replaced by carbonate (specimen 216).

Elsewhere relic phenocrysts of hornblende, recognized by remanent cleavages, are abundant. Montmorillonite or kaolinite rimmed by hematite in a distinct zonal pattern typically follow hornblende in a volume for volume replacement. Potassium feldspar, gypsum, calcite and quartz, also rimmed by hematite have also been observed as a replacement of hornblende. Examples of these types of alteration can be found in samples 82II, 145, 128 and 197I respectively. In intensely argillically altered rock, hornblende phenocryst outlines are preserved by hematite ghosts.

Clear sanidine phenocrysts although rare, have been observed in several crystal tuff specimens (specimens 202, 216, and 82I). In addition a few small perthites have been recognized (specimen 128). Elsewhere potassium feldspar, revealed by thin section staining and x-ray diffraction, occurs in the microcrystalline groundmass of specimens.

A group of specimens, 79, 148, 6II, 13I, 155, and 188, display andesines which are mottled or rimmed by potassium feldspar. Although in general this potassium feldspar appears to be primary, in some instances it may be the result of exchange reactions with a hydrothermal fluid. Potassium feldspar is extremely resistant to alteration and persists even in intensely argillized regions.

Quartz occurs both as monocrystalline clasts and polycrystalline aggregates in pyroclastic rocks. Quartz with rounded, blebby outlines is typical of crystal tuff. Polycrystalline quartz is common in agglomerate or conglom-

erate, as are arenaceous patches of quartz sand. Primary quartz is unaffected by hydrothermal alteration.

Aragonite is the only phase precipitating from the Red Hill and Monroe springs. The knobby growths of this mineral which line the spring channel are probably precipitated by algae.

Deep red, blebby secondary biotites are common in many propylitically altered rocks. They form in patches of hematite (specimen 197I) or inside partially altered primary biotite phenocrysts (specimen 158). In all cases the primary biotites, although altered, are not completely obliterated.

Primary biotites have been partially altered to hematite, kaolinite, and saginitic rutile. They are characterized by a cross hatching of rutile intergrowths. The occurrence of saginitic biotite appears to be limited to argillically altered rocks in the northwestern quadrant of the KGRA (Figure 5). These specimens also exhibit considerable carbonate replacement.

Calcite and dolomite are ubiquitous as replacement minerals. The two carbonates are frequently found together but more generally one occurs to the exclusion of the other. There is, however, no special significance attached to the occurrence of dolomite versus calcite.

Plagioclase and hornblende are most readily replaced by carbonate, although biotite and alkali feldspar may also be attacked. In many specimens carbonate occurs as patches or veinlets cross-cutting rock textures. Hematite often rims such patches.

Carbonate is an abundant phase in both propylitically and argillically altered rock. However, its occurrence in advanced argillic zones, or zones containing sulfate minerals, is highly restricted. Nonetheless, outcrops which have sustained extensive carbonate replacement will frequently adjoin advanced argillic zones (locations 6, 100 and 175). Minor amounts of micro-

crystalline calcite may occasionally be found in rocks which contain jarosite and gypsum. In this association it is usually thickly mantled by hematite.

Throughout the area old fault planes are coated with calcite and fractured outcrops are filled with carbonate. The highest concentration of these carbonate veins is along the range front and adjacent to landslide areas. Calcite which is brightly banded with iron oxides can be found immediately south of the Red Hill tufa mound at the bottom of large wash.

Chlorite is of rare occurrence, being of minor importance in all but a few specimens, 51, 131, and 300. In 131 it occurs as a fracture filling in biotite phenocrysts. This chlorite, which is characterized by radiating fibres, is probably not a direct replacement of biotite. Magnesium chlorite is assumed to predominate based upon the intensity of the 14 angstrom x-ray peak following heating of clay slides to 700°C.

Goethite occurs in the tufa mound at Red Hill together with hematite and frothy calcite. The iron minerals form dense black veins through the calcite and stain the tufa a bright red. Hematite is considered to be a replacement of goethite. The occurrence of goethite requires that either the tufa was deposited at temperatures below 40° centigrade (Berner, 1969) or that cool, iron rich solutions permeated the porous tufa at some time after its deposition.

Gypsum occurs in veins and as fracture fillings in all argillically altered areas. It is not unusual to find gypsum and calcite together as at location 140 where large euhedral calcite and selenite crystals are intergrown. The occurrence of gypsum as a replacement mineral, however, is limited to advanced argillic zones. In this environment it is invariably observed in association with kaolinite and jarosite which it replaces. Gypsum most

frequently appears as irregular patches, cross-cutting rock textures, or veinlets. However, it is also found replacing phenocrysts of plagioclase or hornblende (specimens 155 and 6II). Patches of gypsum are typically rimmed by hematite.

Hematite in the form of disseminated blebs is common in most specimens. In addition, hematite accompanied by calcite, gypsum, kaolinite, montmorillonite or sericite replaces phenocrysts of biotite and hornblende in all but the freshest rock. Chlorite, jarosite, and hematite are the only iron bearing alteration minerals.

The matrix of agglomerate southeast of Red Hill Spring is substantially replaced by hematite as is also the matrix of some of the conglomerate in unit A. The iron bearing solutions which infiltrated these permeable rocks are associated with the north trending fractures in this region.

Jarosite, the iron end member of the alunite solid solution series, is diagnostic for the advanced argillic alteration regime. Identified on the basis of x-ray diffraction analysis, jarosite is never visibly crystalline. In most specimens, gypsum contains a feathery, high birefringent substance which is presumed to be jarosite. Jarosite and gypsum may have formed contemporaneously; however, gypsum is believed to be a replacement of jarosite in most cases. The possibility that jarosite is itself a replacement of potassium alunite is impossible to evaluate. The metastable association of jarosite with minor quantities of calcite is occasionally observed.

Montmorillonite is not present in great abundance. Nevertheless, its occurrence serves as an important diagnostic for the propylitic alteration zone. Its presence, although difficult to verify in thin section, is confirmed by x-ray diffraction. Two varieties with (001) basal spacings of approximately 12.5 and 13.4 angstroms when air dried were recognized. These

varieties are considered to be sodium montmorillonite and calcium montmorillonite respectively. The bright green appearance of many propylitically altered outcrops suggests that the variety nontronite predominates. Sodium montmorillonite has a very limited occurrence being confined to the vicinity of the Monroe and Red Hill springs (specimens 142, 148, 144, and 184). It is undoubtedly the result of post formational cation exchange with calcium deficient sodium chloride solutions. Sodium montmorillonite, however, is not stable with respect to the present spring solutions.

Montmorillonite occurs, for the most part, in relatively fresh rocks or rock that has undergone calcite replacement and is common as a replacement mineral after biotite and hornblende. It is frequently accompanied by hematite.

Massive replacement montmorillonite occurs only in the vicinity of location 138 where Dry Hollow porphyritic andesite has been replaced by calcium montmorillonite.

Kaolinite is the hallmark of argillic alteration, and is without question the most common alteration mineral in the KGRA. Massive replacement kaolinite outcrops along fractures immediately to the east of Red Hill Spring. These deposits may be white or stained red and gray by iron oxides. The textures of these rocks have been nearly obliterated with only hematite ghosts remaining. Intensely kaolinized rocks frequently contain sulfate minerals. Kaolinite occurs as a replacement of all primary phases.

Dense, jet black fracture fillings at locations 197 and 143 are comprised in part of manganese hydroxides. At location 143, adjoining the Monroe tufa mound, the fracture fillings are predominately calcite. At 197, however, the fracture fillings are nearly pure $Mn(OH)_4$ with subordinate halite. These minerals are not replacement phases and apparently bear no relationship to the

alteration of the host rock which is argillically altered in the first case and propylitically altered in the second. Several springs with high manganese concentrations are found in the Central Sevier River Valley. One, the Joseph Hot Spring, is closely related chemically to Monroe and Red Hill Springs (Young, 1965). The source of the Manganese is open to speculation. Manganese minerals have also been reported at the Abraham Spring KGRA by Callaghan (1938). The Abraham spring waters are believed to be heated in proximity to a shallow basaltic dike.

Sericite, while never occurring in great quantities, is an ubiquitous phase in propylitic, argillic and advanced argillic zones. It occurs as a replacement mineral after all primary phases. Coarsely crystalline potassium mica, presumed muscovite, is less common but has been observed replacing biotite (specimen 187II). At sample location 1, a white mica with an unusually low 2V (5°), possibly paragonite, occurs as patches inside calcite and is an apparent replacement of alkali feldspar. Muscovite after feldspar was also observed in specimen 129.

Two varieties of mixed layer clay were identified by x-ray diffraction. Both varieties have unglycolated (001) basal spacings of approximately 10.8 angstroms. One, however, swells when glycolated resulting in a broad reflection between 11 and 17 angstroms. Upon heating to 300° centigrade both clays collapse to a 9.8 angstrom basal spacing. On the basis of this data, the mixed layer clays are presumed to contain interlayered potassium mica and montmorillonite. Because these clays cannot be distinguished in thin section, it is likely that much of the microcrystalline clay identified as sericite is in reality mixed layer clay.

Altered rocks of this area have been subjected to little if any silicification. Several large boulders of microcrystalline, milky white quartz

outcrop at location 124. These boulders are enclosed by slightly dolomitized volcanic breccia and contain inclusions of mottled, dark red, hematite bearing quartz. No remanant feldspar has been detected by x-ray diffraction. It is, therefore, unlikely that this outcrop is silicified volcanic rock. a narrow fracture filling of this same milky quartz occurs at the intersection of two faults near Red Hill Spring (location 211). Microcrystalline quartz occurs as amygdules in porphyritic latite but is not common.

There is a higher incidence of polycrystalline quartz in altered rock suggesting that some of this may be secondary in origin. Secondary polycrystalline quartz is observed occupying a relic hornblende phenocryst in sample 187I. It therefore seems likely that quartz could be a product of the alteration of aluminosilicates by silica saturated solutions.

Tiny crystals of native sulfur between 10 and 50 microns in length, and identified as native sulfur are frequently found disseminated in the matrices of argillically altered rock. At sample location 1 euhedral acute bipyramidal sulfur crystals appear as inclusions in patches of calcite which are rimmed by hematite. Rounded, oblong crystals are clustered in calcite which has replaced hornblende in specimen 128. Hematite penetrates the calcite along relic amphibole cleavages.

Primary Textures

The rocks outcropping in the KGRA can be divided into two textural groups. The first includes porphyritic latite flows, doleritic latite dikes, and aphanitic latite dikes. The pyroclastic rocks include crystal tuffs, bedded tuffs and agglomerates. A third group of rocks, conglomerates and breccias, contain clasts from both groups. A remarkable degree of variety is displayed within each classification making correlation of outcrops very difficult. The problem is further complicated by the widespread hydrothermal

alteration in this area.

The most extensive outcrops of porphyritic latite flows occur in units C and E. These units can be differentiated on the basis of size of andesine phenocrysts and on the basis of magnetite content. Unit E is more coarsely porphyritic and is characteristically a deep rust red.

The porphyritic latite of unit C contains large andesine and hornblende phenocrysts. The largest of these are four and two millimeters in length respectively. Hornblende phenocrysts, however, are in general smaller than this, being 500 microns in length or smaller. The groundmass is a mosaic of tiny plagioclase laths, measuring approximately 50 microns in length. A few corroded biotites, less than one millimeter across, are sparsely distributed. All specimens display distinct flow structures. The appearance of porphyritic latite in thin section is more akin to the aphanitic than the doloritic latite found in the dikes.

Aphanitic dikes are characterized by andesine and hornblende phenocrysts in a microcrystalline groundmass containing potassium feldspar. The andesine occurs as fine laths, approximately 120 microns in length. Hornblende phenocrysts, of the ferrohastingsite variety, are scattered through the matrix. These phenocrysts attain lengths of one or two millimeters. Settling structures are usually marked. Vesicular outcrops are also common. This rock grades smoothly into the porphyritic latite of unit C.

Doloritic dikes generally exceed 15 feet in thickness. They contain phenocrysts of andesine and frequently biotite. The andesine occurs as tabular phenocrysts, approximately one millimeter across. Plagioclase laths in the groundmass with maximum lengths of 500 microns are probably alkali feldspar. Settling structures are frequently observed in these groundmass feldspars. Biotite phenocrysts, when present, are intermediate in size, averaging about 700 microns. Biotite occurs as both tabular crystals and in shafts.

Monocrystalline quartz clasts occur sporadically and rarely exceed 200 microns. They may have been introduced via incorporation of country rock. In altered outcrops this rock may be indistinguishable from pyroclastic rock.

Two distinct crystal tuffs occur. The first, in unit A, is characterized by andesine phenocrysts which are approximately two millimeters across, and narrow biotite shafts ranging in length from several microns to one millimeter. Hornblende phenocrysts average 500 microns in length. Quartz clasts are about the same size.

The second tuff, from unit D, has larger andesines which commonly exceed three millimeters in length. Tabular biotites, approximately two millimeters across and frequently with warped or broken cleavages, are abundant. Smaller biotite laths also occur. Deep red hornblende phenocrysts, also two millimeters in length, are thickly distributed. Large rounded clasts of monocrystalline quartz, one or two millimeters in diameter, also occur. Unlike the tuff of unit A, this rock contains many accidental rock fragments. The crystallinity of the matrix varies, resulting in a patchy appearance when viewed in outcrop. These patches are presumed to have resulted from the incorporation of lapilli.

Both tuffs have a distinctly clastic texture, with many of the phenocrysts broken or fractured. Their matrices are composed of fine fragments in a microcrystalline groundmass.

Bedded tuff occurs as thin interbeds in both units B and A. It is characterized by a fine clastic texture made up of remarkably well sorted fragments of plagioclase, biotite, quartz and magnetite. The sorting may be the result of air fall or, perhaps alluvial action. The fragments are angular and uniformly 200 to 500 microns in diameter. The matrix is microcrystalline

potassium feldspar. In outcrop this tuff has a distinct arenaceous appearance.

Agglomerate is texturally similar to doleritic latite in that it contains an aphanitic matrix comprised of plagioclase laths, 100 microns in length or less. Flow structures, preserved in these laths, are often observed. The agglomeratic texture is distinctly nonuniform compared to the effusive textures, however. Andesine, biotite, quartz and hornblende can vary dramatically in size, habit and distribution. Phenocrysts do not exceed one millimeter in most outcrops, however. Quartz occurs both as monocrystalline clasts and polycrystalline aggregates. Agglomerates often contain irregular pebbles of effusive volcanic rock and may grade into breccia.

Alteration Assemblages

The alteration assemblages observed in the KGRA bear a strong resemblance to alteration assemblages in the oxidized zones of many hydrothermally altered areas. A zonal pattern of alteration can be generalized for these areas which contains, in succession outward: 1) an alunite-quartz zone 2) an argillic zone, which may be subdivided into potassium mica-kaolinite and mixed layer potassium mica-montmorillonite subzones and 3) a propylitic zone, which may be subdivided into montmorillonite-vermiculite and chlorite subzones (Hemley, et. al., 1969). The alunite-quartz zone is generally highly siliceous and may contain dickite, pyrite or diaspore. In the Monroe KGRA only the argillic zone is well developed. However, jarosite, an iron alum isomorphous with alunite, does occur in association with kaolinite and sericite and is a common feature in the most intensely altered areas. Outcrops of rock which have been substantially altered to montmorillonite and chlorite are rare (locations 138 and 51). More commonly

montmorillonite occurs as a minor mineral at the extremities of hydrothermally altered areas or in association with kaolinite and mixed layer clays in argillically altered regions. Using the zonal model of Hemley as a guideline, the following zones are proposed for Monroe (Figure 32).

- Advanced Argillic: contains kaolinite and jarosite
- Argillic: contains kaolinite and mixed layer clay but does not contain montmorillonite
- Propylitic: contains montmorillonite and mixed layer clay
- Advanced Propylitic: contains montmorillonite or chlorite to the exclusion of kaolinite and mixed layer clay (rare)

Carbonate is frequently associated with argillic, propylitic and advanced propylitic assemblages. The peculiar hematite-kaolinite alteration associated with the north trending structures of the central part of the KGRA constitute a special case of argillic alteration.

An example of a low temperature hydrothermal area with important parallels to the Monroe KGRA is the Nevada Goldfield district. The Goldfield district is important for its quartz-gold vein mineralization. In the peripheral oxidized zone of this area, kaolinite, frequently massive, occurs with quartz, alunite and pyrite. Native sulfur and diaspore are minor accessory minerals. Potassium mica, (sericite), occurs sparsely. Gypsum is widely distributed in veins and patches both above and below the water table. Calcite is abundant in the propylitically altered areas. Of particular interest is the association of coarsely crystalline alunite with microcrystalline calcite, kaolinite and sericite in some altered rhyolites (Ransome, et. al., 1909).

The alteration relationships in the Monroe KGRA, therefore, are consistent with the peripheral oxidized zone of a low temperature, sulfide dominated hydrothermal system.

Figure 32. Variation of mineral abundance in altered rocks at Monroe shown as percent of total x-ray diffraction peak intensity indexed to kaolinite.

Multivariate Analysis of Rock Alteration

Because of the nearly universal superimposition of alteration assemblages, the subtle differences between these assemblages from outcrop to outcrop, and the feeble correlations between the occurrence of most phases, statistical techniques were adopted to determine more objectively the basic structure of the data. Three basic approaches were utilized: principle component analysis (PCA), cluster analysis of the raw data, and cluster analysis of the data after it had been transformed into factor scores on the principle axis selected by PCA. Clustering was conducted in both the Q mode, that is clustering of samples, and in the R mode, that is clustering of components.

The data consisted of the composition of each of 95 samples which was reported in percent total pattern intensity attributable to 15 recurrent phases: quartz, plagioclase, orthoclase, calcite, dolomite, hematite, jarosite, gypsum, apatite, biotite, kaolinite, montmorillonite, chlorite, and two mixed layer clays. Thirty of the samples for which clay fraction x-ray diffraction was available were analyzed separately as well as with the remainder of the data. Because the data was normalized and because it was not standardized across the variables, bias was unavoidably introduced into the analysis. The result of analyzing data presented in this fashion is to enhance the effect of the variance attributable to the most abundant phases and to generate spurious negative correlations between minor and major phases. Because of the nature of the data, however, standardization, that is insuring that each variable populations has a mean of zero and a standard deviation of one, was not considered to be a meaningful operation. The distortion due to this phenomenon is not believed to have appreciably affected the analysis.

Clustering was based on the degree of association between components or samples as defined by correlation coefficients. In one trial, however, the

data were scaled to a presence-absence format and a simple taxonomic distance coefficient was used. In all, twelve cluster analysis trials were made.

Principle Component Analysis (R Mode)

The variance for the 30x15 array was 15. Ten mutually orthogonal factors were selected which together accounted for 94.2 percent of the total variance. The principle factor loadings (eigenvectors) for the transformations are reported in Table 13. Reference to that table shows that the first factor, accounting for 25.7 percent of the total variance, heavily weights the contribution of plagioclase, quartz, kaolinite, montmorillonite, jarosite, apatite, and biotite. Quartz, jarosite and kaolinite enter with negative contributions. This is a forceful statement of the close association of the latter minerals and their strong negative correlation with the primary phases plagioclase, biotite, and apatite, as well as montmorillonite. This is the anticipated result if a central argillized zone passes outward to a weakly propylitized zone. Factor one, therefore, may be visualized as an index of reaction progress. The groups obtained by cluster analysis are, for the most part, predicated on this factor.

Factor two is characterized by positive contributions from orthoclase mixed layer clay I and gypsum, and by a negative contribution from calcite. It contains 14.1 percent of the total variance. It is apparent that the occurrence of calcite and gypsum is not linked to the degree of overall reaction progress, but is influenced by some independent property, presumably the pCO_2 or pH_2S of the system. In addition, orthoclase and mixed layer clay I, both potassium phases, appear to be highly correlated and to be unfavored in the CO_2 regime.

The third factor, accounting for 10.7 percent of the variance, appears

Table 13. Principle Axis Factor Loadings

<u>VARIABLE</u>	<u>FI</u>	<u>FII</u>	<u>FIII</u>	<u>FIV</u>	<u>FV</u>	<u>FVI</u>	<u>FVII</u>	<u>FVIII</u>	<u>FIX</u>	<u>FX</u>
QUARTZ	-.809	-.119	.175	.340	.169	.177	-.105	-.103	.187	-.020
PLAGIOCLASE	.883	.042	.010	-.145	-.048	.315	.052	-.033	-.031	-.028
ORTHOCLASE	.218	.733	-.047	-.268	-.200	-.368	-.141	-.226	-.205	-.019
CALCITE	.096	-.551	-.190	-.347	-.459	-.144	.342	.342	-.005	.141
DOLOMITE	-.050	-.467	-.344	-.44	.614	-.279	.311	-.132	-.002	-.224
HEMATITE	.119	-.197	-.268	.718	-.274	.099	-.241	.223	-.295	-.248
JAROSITE	-.603	.404	.492	.006	-.178	-.001	.278	-.146	-.069	-.178
GYPSUM	-.389	.509	.156	.140	.091	.204	.509	.436	-.105	.023
APATITE	.590	.018	.543	.039	.285	-.228	.011	.226	.206	-.168
BIOTITE	.557	-.059	.493	.246	.007	-.468	-.126	.219	-.019	.004
KAOLINITE	-.746	-.064	.068	.024	.123	-.280	-.303	.96	-.004	.379
MONTMORILLONITE	.633	.125	.016	.462	-.155	.165	.174	-.188	.348	.268
CHLORITE	.237	-.255	.463	-.343	.332	.499	-.234	.075	-.252	.092
MIXED LAYER CLAY I	-.025	.549	-.446	-.266	.112	.133	-.298	.385	.329	-.179
MIXED LAYER CLAY II	-.393	-.435	.360	-.266	-.506	.054	-.135	-.035	.229	-.259
percent variance	25.7	14.1	10.7	9.33	8.49	7.08	6.28	5.24	3.79	3.44

to be a measure of the alteration of biotite and apatite, both highly resistant minerals. The apparent degeneracy of this factor with respect to factor one is not understood.

Factor four, with 9.3 percent of the variance, is dominated by hematite, reflecting the ubiquitous and capricious occurrence of this mineral.

Factor five heavily weights the contribution of dolomite, mixed layer clay II and calcite. This factor could be interpreted as an indicator of the concentrations of Ca^{+2} and Mg^{+2} which is also presumed to be independent of the overall reaction progress. It accounts for 8.5 percent of the total variance.

Factor six contains 7.1 percent of the variance. It reflects the reaction relationship of chlorite and biotite. This is the last factor from which information can readily be gleaned. The remaining factors together account for less than 19 percent of the variance.

In summary, PCA of the 30x15 matrix reveals that a particular sample can be virtually defined by 10 orthogonal vectors as opposed to the 15 variables in the original data set. Therefore, the original set of variables must contain a substantial degree of covariance. This is useful since it means that reaction relationships among the phases have not been totally obscured. Furthermore, positive correlations are inferred for the following mineral pairs: jarosite-kaolinite, quartz-kaolinite, montmorillonite-plagioclase, montmorillonite-biotite, apatite-biotite, gypsum-mixed layer clay I, gypsum-orthoclase, dolomite-calcite, and mixed layer clay II-calcite. Negative correlations are inferred for the following mineral pairs: chlorite-biotite, calcite-orthoclase, calcite-gypsum, kaolinite-plagioclase, kaolinite-montmorillonite, and kaolinite-biotite.

The variance of the 95x15 array was also 15. Ten factors were selected which cumulatively accounted for 88.3 percent of the total variance. The first

factor contains 19.6 percent of the total. The principle factor loadings are reported in Table 14. The similarity of the two data sets lends confidence to the interpretation based on the PCA of the more detailed 30x15 array. However, several important associations were emphasized in this analysis which were not apparent in the previous run. In particular, a positive correlation between montmorillonite and calcite is inferred from the loadings for factor five.

The clarity of the PCA for these and similar data sets would be greatly improved if a factor rotation routine such as the Kaiser verimax scheme were adopted. This procedure optimizes the factor loadings to values approaching one or zero by fixed rotation of the principle vectors. However, due to limited time, funds and expertise, this was not attempted. As it has been utilized in this study, PCA fosters an intuitive feel for the structure of the data and provides useful guidelines for further analysis.

R Mode Cluster Analysis

R mode cluster analysis was employed in order to determine the relationships prevailing among the phases in a more straight forward and graphic fashion. Cluster analysis of raw data for both arrays was carried out. In addition, cluster analysis was conducted on the 30x15 array after the data had been transformed into factor scores following PCA. Factors were weighted according to their percent variance. The results of all three trials were similar (Figure 33). Phases which cluster with a high degree of similarity have strong positive correlations.

Highly correlated mineral pairs are: plagioclase-montmorillonite, orthoclase-mixed layer clay I, quartz-kaolinite, jarosite-gypsum, and mixed layer clay II-calcite. More distantly clustered groups of phases are:

Table 14. Principle Axis Factor Loadings

VARIABLE	<u>FI</u>	<u>FII</u>	<u>FIII</u>	<u>FIV</u>	<u>FV</u>	<u>FVI</u>	<u>FVII</u>	<u>FVIII</u>	<u>FIX</u>	<u>FX</u>
QUARTZ	.727	-.454	.080	.031	-.325	-.029	.015	-.105	.090	-.086
PLAGIOCLASE	-.782	.175	-.024	-.223	-.120	-.398	-.146	-.155	-.016	.128
ORTHOCLASE	-.134	.713	-.014	-.099	-.076	.080	.547	.029	.030	-.155
CALCITE	-.096	-.340	-.150	-.073	.552	-.141	.176	.645	-.246	-.082
DOLOMITE	.007	-.164	-.077	.800	.007	-.079	-.075	-.166	-.409	.137
HEMATITE	-.192	-.549	-.196	-.319	-.336	.138	-.188	.227	.263	.045
JAROSITE	.687	.355	.010	-.365	.009	-.069	-.165	-.061	-.253	.083
GYPSUM	.524	.408	-.125	-.182	-.184	-.065	-.452	.293	-.261	.027
APATITE	-.170	.134	.758	.071	-.060	-.034	-.113	.260	-.071	.278
BIOTITE	-.146	-.147	.488	-.075	-.357	.520	.241	.143	-.244	.103
KAOLINITE	.815	.030	.116	.144	.070	.077	.145	.041	.104	-.193
MONTMORILLONITE	-.130	.211	.011	.061	.518	.674	-.401	-.098	.194	.022
CHLORITE	.011	.002	.699	.104	.168	-.362	-.205	.031	.276	-.325
MIXED LAYER CLAY I	.184	.357	-.228	.467	-.200	-.075	.032	.396	.476	.300
MIXED LAYER CLAY II	.475	-.196	.146	-.280	.411	-.152	.257	-.211	.146	.509
percent variance	19.6	11.6	0.85	8.77	8.13	7.41	6.58	6.15	5.84	4.43

Figure 33. Cluster analysis in the "R" mode.

plagioclase-montmorillonite-chlorite, and mixed layer clay II-calcite-jarosite-gypsum-quartz-kaolinite. These relationships compare very favorably with the deductions which were based on PCA. Unfortunately, cluster analysis is very ambiguous with respect to negatively correlated phases. In presenting an overall understanding of the dynamics of alteration reactions PCS may be more effective.

For one trail the 95x15 array was scaled to a presence-absence format and clustered using a simple taxonomic distance function. This run did not result in meaningful clustering.

As an aid to interpreting the results of PCA and cluster analysis a graphic approach was adopted. The percentage pattern intensities of montmorillonite, chlorite, mixed layer clay I, mixed layer clay II and jarosite were plotted against the index phases plagioclase, kaolinite and carbonate for the 30x15 array. The plots are presented in Figures 34 and 35. Dotted lines have been added in order to emphasize the limiting correlative relationships. They have no statistical significance.

Mixed layer clay I does not occur with sufficient frequency to draw any conclusions. Montmorillonite is stable at high percentages of plagioclase and when present is roughly proportional to plagioclase. It is inversely proportional to kaolinite, never being observed in specimens containing more than 12 percent kaolinite. Chlorite, however, appears to persist at higher percentages of kaolinite, with a crude inverse relationship prevailing with that mineral. Percentages of mixed layer clay II and plagioclase are roughly proportional. Jarosite is clearly incompatible with plagioclase and therefore with montmorillonite as well. Jarosite is stable in the presence of kaolinite. Kaolinite and mixed layer clay II persist at high percentages of carbonate. Montmorillonite and chlorite, on the other hand bear a marked inverse relat-

Figure 34. Variation of alteration mineral abundances plotted against kaolinite and plagioclase. Abundances are relative x-ray diffraction peak intensities.

Figure 35. Variation of alteration mineral abundances plotted against kaolinite, plagioclase, and carbonate. Abundances are relative x-ray diffraction peak intensities.

ionship with carbonate. It appears that the variable quantities of carbonate in most specimens may contribute to obscuring apparent correlative relationships in other plots.

The graphic presentation supports the interpretations which have been based on both multivariate techniques. It also serves to demonstrate the power of these methods. By not being confined to the use of two normally nonorthogonal vectors, PCA and cluster analysis are not subject to interference due to vectors with low covariances and permit a much better look at the fine structure of the data.

Q Mode Cluster Analysis

It was deemed important to gain an understanding of the spatial distribution of different alteration types as well as to determine the internal structure of these groups. Due to the ambiguities involved in grouping the specimens, cluster analysis was again employed. Three approaches were taken: 1) Raw arrays were clustered. 2) Arrays were clustered after the data had been transformed into factor scores and the factors weighted according to their percent variance. 3) Arrays were clustered after the data had been transformed into factor scores and the first factor deleted.

By deleting the factor containing the highest percentage of the total, it was intended to compensate for the bias introduced by normalizing the data. Because most of the variance is attributable to the most abundant phases, removing a high variance factor should enhance the contribution of the minor minerals in the cluster analysis.

Deletion of the first factor resulted in poorly clustered or nonclustered trials. Other trials produced a variety of related clustering patterns which were, for the most part, predicated on variations in kaolinite, plagi-

clase and carbonate. When evaluated on the basis of PCA and R mode cluster analysis, Q mode cluster analysis of raw unscaled data produced the most rational groupings of specimens. The results are presented in Figure 36. A histogram showing the distribution of phases among the clusters is presented in Figure 37.

Four clusters were obtained. Cluster I contains specimens which share in common a low percentage of plagioclase, a high percentage of kaolinite and are more apt to contain jarosite, gypsum or mixed layer clays. Cluster III specimens are intermediate in their characteristics between clusters I and III and are more apt to contain high percentages of carbonate. Cluster IV contains specimens with high percentages of plagioclase but otherwise is poorly defined. This cluster was discarded.

In general cluster I can be identified with acid sulfate alteration, cluster II with carbonate argillic alteration, and cluster III with a weakly propylitic alteration. When plotted on the base map in Figure 11 the specimens of cluster I and II are observed to follow the major northeast trending structures. The specimens of cluster I, however, occupy the most intensely altered areas. A boundary appears to exist near location 124, (Figure 31), north of which cluster I does not penetrate. Cluster II specimens are heavily concentrated along the north trending structures as well. This suggests that some of the specimens in cluster II may represent a separate genetic group. Cluster III is more diffusely distributed. These specimens frequently occupy locations peripheral to the most intensely altered rock. A distinct zonal arrangement, therefore, is apparent which passes outward from an acid argillic zone to a carbonate argillic zone and then to a weak propylitic zone.

The figure consists of three histograms representing similarity coefficients for different rock groups. The y-axis is labeled 'SIMILARITY COEFFICIENT' and ranges from 97 to 100. The x-axis is divided into three main sections: 'GROUP TWO', 'GROUP ONE', and 'GROUP THREE'. Each section contains multiple histograms representing different rock types or stages. The histograms are labeled with specific rock types or stages: 'ARGILLIC', 'ADVANCED ARGILLIC', and 'PROPYLITIC AND ADVANCED PROPYLITIC'. Arrows point to specific data points on the histograms.

GROUP TWO

- 19
- 82 II
- 130
- 208 I
- 94 I
- 24
- 92 II
- 93 II
- 94 I
- 81
- 4 II
- 15 II
- 152
- 191
- 125
- 192 II
- 300
- 92 II
- 82 I
- 123
- 154

GROUP ONE

- 73 I
- 187 II
- 149
- 39
- 153
- 174
- 5 II
- 193 II
- 100 II
- 101
- 102
- 142
- 100 I
- 175 II
- 183
- 187 I
- 184
- 180
- 61
- 190 II
- 22
- 102
- 140

GROUP THREE

- 209 II
- 202
- 206
- 118
- 197 I
- 179 I
- 131
- 92 II
- 112
- 145
- 302
- 190 I
- 14
- 185
- 186
- 180
- 194 II
- 158
- 180
- 23
- 54 II
- 85
- 124 II
- 212
- 128
- 25
- 303
- 148
- 308
- 216
- 221
- 175 I
- 193 I
- 189
- 129
- 220
- 231
- 205
- 232
- 236

103

DISTRIBUTION OF PHASES IN CLUSTER GROUPS

Figure 37. Histograms of mineral distribution in three major clusters.

Figure 38. Distribution of major clusters.

Spring Geochemistry

The water from the Red Hill and Monroe springs was sampled and analyzed following the procedures outlined by Presser and Barnes (1974). The concentrations of the following ions were determined.

Sr^{+2} , Ca^{+2} and Mg^{+2} : A one liter sample of spring water was filtered through a .1 micrometer filter under compressed nitrogen and then acidified to pH 2 with HCl. The ion concentrations were determined using a Perkin-Elmer 303 atomic absorption spectrometer.

Na^{+} : A one liter sample of spring water was filtered through a .1 micrometer filter under compressed nitrogen. The ion concentrations were determined by atomic absorption spectrometry. The results were checked against the activities of sodium measured directly using an Orion select ion probe and pH meter.

Cl^{-} : A one liter sample of spring water was filtered through a .1 micrometer filter under compressed nitrogen. The ion concentration was determined by a Mohr titration. The results were checked against the activities of chloride measured by select ion probe.

SO_4^{-2} : A one liter sample of spring water was filtered through a .1 micrometer filter under compressed nitrogen and treated with chloroform. The ion concentrations were determined using an indirect atomic absorption technique.

S^{-2} : A 100 milliliter sample of spring water was collected and the sulfide fixed by addition of zinc acetate and sodium hydroxide. Specimens were titrated the same day using a iodine-thiosulfate backtitration technique. Results of the titration were checked against the sulfide activities obtained by select ion probe.

HCO_3^- : An untreated sample was titrated with standard HCl to a methyl orange endpoint.

H_4SiO_4 : A one hundred milliliter sample was diluted 1:10 and stored in a plastic container. Silica concentrations were measured the next day by atomic absorption.

pH: Hydrogen ion activity was measured using a glass electrode and pH meter standardized to the temperature of the spring.

Sulfide was below the detectable limit, 10^{-6} molar in both springs. The measured concentrations of the remaining species are reported in Table 15.

The composition of spring gases was also determined. For this determination a gas sample was obtained by displacement of spring water from a two liter glass bottle. The sealed bottle was then connected to a gas train and the contained gases forced through a series of solution traps by compressed nitrogen. Oxygen was determined using a modified Winkler titration and hydrogen sulfide using the same method employed for the dissolved sulfide determination. Carbon dioxide was trapped in standard barium hydroxide solution, which was then back titrated with standard HCl. Only carbon dioxide was detected in each sample.

High temperature dissociation constants and equilibrium constants were taken from Helgeson (1969), tables 4, 11 and 12. Dissociation constants at 25° centigrade for certain aqueous complexes were taken from Garrels (1965). Activity coefficients were calculated using the Debye Huckel relation and deviation function as tabulated in Helgeson (1969), table 2. The assumption was made in these calculations that $B\alpha^0$ equals 1.5 (Nriagu, 1971).

In calculating the activities of the major ions in the Red Hill Spring water, the following complexes were considered; CaSO_4^0 , CaCO_3^0 , CaHCO_3^- , NaHCO_3^0 , and NaSO_4^- . The molality of Mg^{+2} and K^+ were low enough such that complexes containing these ions need not be considered. Since high temperature

thermodynamic data on CaHCO_3^+ , NaHCO_3^0 and NaSO_4^- was not available, the values of dissociation constants at 25° centigrade were used in calculations of ion activities below 75° centigrade. At higher temperatures the activities of CaHCO_3 and NaHCO_3 were not considered and the dissociation constant of KSO_4^- was assumed to be sufficiently similar to NaSO_4^- to be used in its place. The activity of neutral complexes was assumed to be unity. In addition the activity of Na^+ was treated as a constant, as was also pH.

Given these assumptions the activities of the major ions were calculated for a temperature of 75° centigrade, a pH of 6.6 and a molality of Na^+ of .0255. The results of these calculations are in table 16.

It is apparent that any errors introduced by assumptions made about the complexes CaHCO_3^+ and NaHCO_3^0 will be small. The spring water is saturated with respect to gypsum and the quartz at the temperature of the flowing spring and slightly oversaturated with respect to calcite. The calculated $f\text{CO}_2$ is consistent with saturation with respect to $\text{CO}_{2(g)}$ at a depth of approximately 25 feet, assuming a hydrostatic pressure gradient of .028 atmosphere per foot. Therefore a slight oversaturation in calcite is apt to prevail at the surface as CO_2 gas separates from the solution in the last few feet of ascent.

The analysis of Joseph Hot Spring water reported in Table 15 shows that it is chemically very similar to that of the Red Hill Hot Spring and like that spring, saturated with respect to calcite and gypsum at the spring temperature.

Origin of the Monroe Spring Water

In considering the chemistry of the Monroe spring system, the chemical composition of the Red Hill Spring has been used to the exclusion of the Monroe Spring. This is because the chemistry of the two springs are similar

Table 15. Warm Spring Waters in the Monroe Area

	1	1	2
m_{Ca}^{+2}	.0074	.0069	.0070
m_{Na}^{+}	.026	.025	.063
m_{K}^{+}	.0014	.0013	.0017
m_{Mg}^{+2}	.0015	.0015	.0015
m_{Cl}^{-}	.019	.020	.049
$m_{HCO_3}^{-}$	.0076	.0070	.0070
$m_{SO_4}^{-2}$	.010	.011	.013
$m_{H_4SiO_4}^0$	.00068	.00045	.00089
pH	6.1	6.6	6.6
Temp ($^{\circ}C$)	59	77	65

1 analysis by Benson and Miller

2 analysis by U.S.G.S. (Young, 1965)

Table 16. Activities of Aqueous Species in the Red Hill Spring

Temperature 75° C

pH 6.6

<u>Species</u>	<u>Activity</u>

Ca^{+2}	.0019

SO_4^{-2}	.0035

Na^{+}	.021

Cl^{-}	.021

HCO_3^{-}	.0030

CO_3^{-}	8.4×10^{-7}

K^{+}	.001

Mg^{+2}	.0006

CaSO_4^0	.0026

NaSO_4^{-}	.00039

CaHCO_3^{+}	.00010

NaHCO_3^0	.000035

H_4SiO_4^0	.00045

$f\text{CO}_2 = .12$	

$I = .03$	

$\log(a_{\text{Ca}^{+2}} \times a_{\text{SO}_4^{-2}}) = -5.2$	

$\log K_{\text{sypsum}} = -5.3$

$\log(a_{\text{Ca}^{+2}} \times a_{\text{CO}_3^{-2}}) = -8.8$	

$\log K_{\text{calcite}} = -9.0$

$a_{\text{Ca}^{+2}}/a_{\text{SO}_4^{-2}} = .54$	

$a_{\text{Ca}^{+2}}/a_{\text{HCO}_3^{-}} = .63$	

Table 17. Activities of Aqueous Species in the Joseph Hot Spring

Temperature 65⁰

pH 6.6

<u>Species</u>	<u>Activity</u>
Ca ⁺²	.0020
Na ⁺	.052
K ⁺	.0014
Mg ⁺²	.00066
Cl ⁻	.029
SO ₄ ⁻²	.0032
HCO ₃ ⁻	.0030
CO ₃ ⁻²	8.4 x 10 ⁻⁷
fCO ₂ = .10	

$$\log(a_{\text{Ca}^{+2}} \times a_{\text{SO}_4^{-2}}) = -5.2$$

$$\log K_{\text{gypsum}} = -5.2$$

$$\log(a_{\text{Ca}^{+2}} \times a_{\text{CO}_3^{-2}}) = -8.8$$

$$\log K_{\text{calcite}} = -8.8$$

$$a_{\text{Ca}^{+2}}/a_{\text{SO}_4^{-2}} = .6$$

$$a_{\text{Ca}^{+2}}/a_{\text{HCO}_3^{-}} = .7$$

and because Red Hill Spring has the higher temperature and greater flow rate of the two.

The water contained in the hot springs of this area has equilibrated with calcium carbonate before outcropping at the surface. The most likely site for this equilibration is the limy beds of the Carmel formation which directly underlies the Bullion Canyon volcanics.

It is considered unlikely that the spring waters represent fluids which have equilibrated at high temperatures with sedimentary carbonate rock and then risen without further alteration save adiabatic cooling and separation of carbon dioxide gas. The hypothetical compositions of fluids saturated with respect to gypsum and calcite at elevated temperatures were calculated. In order to maintain ionic strengths similar to those at Red Hill without increasing the molality of NaCl, lower pH's and higher carbon dioxide fugacities are required. In addition, similar activity ratios, $a_{Ca^{+2}}/a_{SO_4^{-2}}$ and $a_{Ca^{+2}}/a_{HCO_3^{-}}$, to those at Red Hill cannot be obtained. Therefore, high temperature solutions which have equilibrated with carbonate should differ fundamentally from the observed spring water in major ion distribution as well as pH. Furthermore, due to the elevated fugacities of carbon dioxide in these fluids, separation of carbon dioxide would begin at considerable depths, nearly 1400 feet at 150° C if a hydrostatic gradient is assumed. This would cause the solutions to be significantly undersaturated with respect to calcite at the surface. A high temperature reservoir, therefore, can be the immediate source of the spring water only if complete re-equilibration with calcite at a lower temperature and pH occurs.

Mixing of high and low temperature solutions is considered a plausible origin of the spring waters.

If mixing does prevail, mixing temperatures, taken as the quartz saturation temperatures, should be proportional to the degree to which a hot source solution has been diluted by cold water. Based upon this reasoning, quartz saturation temperatures were plotted versus the concentration of chloride for available groundwater analysis in the Central Sevier River Valley. Chloride concentration was chosen as an index of dilution because it is independent of most mineral equilibria and is not subject to complexing at moderate temperatures and ionic strengths. Silica versus chloride plots of data from Young (1965) are presented in Figure 39. The low chloride solutions are greatly dispersed, but are a tight cluster, represented by open circles, which can readily be distinguished. This group has an average quartz saturation temperature of 85° centigrade and average pH of 7.5. Joseph Hot Springs, with the highest chloride concentration in the population, has a quartz saturation temperature of 128°C and a pH of 6.6. Three high chloride analyses, including Red Hill Spring, plot linearly between these two points. These points are represented by closed triangles. Red Hill and Joseph Hot springs would appear to be the result of mixing of high and low temperature fluids. Assuming that the source solutions are not boiling at depth and that enthalpy in calories is approximately equal to temperature in degrees centigrade, 50 percent dilution of Joseph Hot Springs water is required to produce the Red Hill Spring mixing temperature of 105°C.

The quartz saturation temperatures, calculated from the analyses reported by Young, have been plotted on a map of the Central Sevier River Valley. Also indicated are different groundwater populations as identified by the mixing plot (Figure 40). The map clearly shows a ridge of high silica temperatures which follows the edge of the Sevier Plateau and then bends westward south of Monroe. The high chloride warm springs outcrop along this ridge as do also low chloride waters with anomalously high quartz saturation temperatures (closed circles). Low chloride waters with quartz saturation temperatures approximating 85° centigrade outcrop along the valley floor. Low chloride waters with quartz saturation temperatures below 70° centigrade have no clear spatial distribution and are considered to be derived from shallow aquifers which have no importance to the proposed mixing model. These solutions are represented by crosses on Figure 39.

Figure 40. Distribution and quartz equilibration temperature of Central Sevier Valley waters. Contours on quartz equilibration temperatures.

On the basis of this evidence two sources of the spring waters are proposed. A high temperature, sulfate dominated fluid is hypothesized which has high salinity and low pH. It may be derived from a deep thermal reservoir although this cannot be determined. The Red Hill and Joseph Hot Springs represent fluid derived from this source which has mixed with solutions of the type which outcrops along the valley floor, namely low salinity warm water with neutral pH's. Certain of the high silica, low chloride waters which outcrop near the hot springs may represent solutions derived from this same aquifer but which have been heated in the anomalously high geothermal gradient accompanying the hot springs. It is proposed that mixing may occur in the vicinity of the Carmel shales which serve as an impermeable boundary between the hot and cold aquifers until they encounter fractures associated with the range front fault system.

The summits on the Sevier Plateau east of Monroe are less than 6000 feet above the valley floor. If an additional depth of 2000 feet is assumed to the base of the volcanic cover, meteoric water which has its area of recharge on the Plateau uplands might be expected to fall a maximum of 8000 feet through the geothermal gradient before reaching the base of the spring system. Furthermore, if a normal heatflow of $1.5 \times 10^{-6} \text{ cal} \cdot \text{cm}^{-2} \cdot \text{s}$ is assumed and an average thermal conductivity of $.01 \text{ cal} \cdot \text{cm}^{-1} \cdot ^\circ\text{C}^{-1}$ for the surrounding terrain, the maximum temperatures that can be expected for the system are in the range of 50° centigrade. It seems likely that the source solutions for the springs would have temperatures of 100° to 200° centigrade. Therefore, another source of heat other than the normal geothermal gradient must be considered. The warm waters that outcrop along the valley floor are well within the temperature range than can be accounted for by the geothermal gradient, however.

If mixing, followed by redistribution of the major ions via equilibration with carbonate and sulfate minerals, does account for the present spring water, the Na-Ca-K geothermometer may provide an erroneous estimate of reservoir temperatures. Curiously, however, Na-Ca-K temperatures estimates do reproduce the thermal anomaly observed when quartz saturation estimates were mapped. High Na-Ca-K temperatures in the warm springs may result if K^+ is concentrated in the high temperature brine. Under these conditions Na-Ca-K temperatures should be roughly proportional to dilution of the high temperature brine, with the least diluted waters possessing the highest temperatures.

A few boundary conditions can be placed upon a hypothetical high temperature source fluid. It is not likely that the Joseph Spring represents an undiluted high temperature fluid. Rather, it probably represents a point in the dilution curve along with the Red Hill Spring. Therefore, the temperatures of the fluid must exceed the quartz saturation temperature of the Joseph Hot Spring. Unfortunately, this value is not accurately known at the time of this writing. Based upon the assumed 50 percent dilution of the Red Hill Spring relative to Joseph Hot Spring, it is probably greater than 100° centigrade.

When water was hypothetically equilibrated with calcite and gypsum at a pH of 6.6, a temperature of 70° C and a NaCl molality of .025, it was observed that the activity ratios, $a_{Ca}^{+2}/a_{SO_4}^{-2}$ and $a_{Ca}^{+2}/a_{HCO_3}^{-}$, exceed those observed in the Joseph and Red Hill springs. It was therefore assumed that the composition of the hot springs must be controlled by sulfate introduced from outside the Carmel shales. Calculations revealed that an initial activity of that ion equal to .0025 will result, after equilibration with calcite and gypsum, in activity ratios of .4 and .6 respectively. This value represents a minimum for the mixed solution because of equilibration with carbonate rocks.

Therefore, it seems likely that the source fluid contains elevated sulfate and that this sulfate is derived from the near surface oxidation of dissolved sulfide.

Alteration Geochemistry

The surficial alteration observed in the Monroe KGRA should reflect the chemistry of the altering hydrothermal fluids. The advanced argillic zone represents the most intense surficial alteration. It is assumed that the phases in this zone most accurately reflect the properties of the source solutions and that solutions capable of generating the typical argillic and propylitic assemblages have been derived from these source solutions as a consequence of fluid-rock reactions.

Theoretical activity diagrams depicting the stability relationships among the principle phases at various temperatures and solution compositions were constructed using thermodynamic data contained in Helgeson (1969), tables 3, 7, 8, 11 and 12, and Shomate (1946). Alunite has been incorporated in the diagrams in place of jarosite because the alunite-kaolinite reaction is better understood than the jarosite-kaolinite reaction, because heat capacity data is not available for jarosite, and because of the difficulties introduced by uncertainties in Fe^{+3} activities. The $\log(a_{\text{Fe}^{+3}}/a_{\text{Al}^{+3}})$ ratios required to stabilize jarosite over alunite are moderate, 1.7 at 25⁰ centigrade, if the free energy of jarosite reported by Brown (1970) is used. These values are probably attained locally in response to the dissolution of ferromagnesium aluminosilicates. The theoretical free energy for the reaction:

probably represent an upper bound. Raymahashay (1968) demonstrated a lower free energy for this reaction in natural hot springs. Furthermore, under

Table 18. Thermodynamic Data

SUBSTANCE	STANDARD ENTHALPY		STANDARD ENTROPY		AVERAGE HEAT CAPACITY			HEAT CAPACITY FUNCTION		
	298°	K	298°	K	60°	100°	150°	a	b $\times 10^3$	c $\times 10^{-5}$
natural alunite	-1235600	cal.	76.1	cal.				231	78.6	-67.8
crystalline gibbsite	-357180	cal.	28.02	cal.				8.65	45.6	
kaolinite	-980020	cal.	48.53	cal.				67.9	19.2	-13.7
microcline	-946000	cal.	52.47	cal.				63.8	12.9	-17.1
Ca-montmorillonite	-1367980	cal.	61.2	cal.				84.2	33.6	-20.0
muscovite	-1420900	cal.	69.0	cal.				97.6	26.4	-25.4
H ₄ SiO ₄	-348060	cal.	45.84	cal.	49	49	46			
H ⁺	0	cal.	0	cal.	23	31	33			
H ₂ O	-68315	cal.	16.71	cal.	18	18	18			
Ca ⁺²	-129770	cal.	-13.2	cal.	45	59	60			
K ⁺	-60040	cal.	24.5	cal.	27	35	35			
SO ₄ ⁻²	-217320	cal.	4.8	cal.	-99	-108	-105			

conditions of low temperature and intermediate acidity where the crystallization rates of silicates to non-silicates are very low, "it is conceivable that alunite may be produced well outside of its stability field in the breakdown of feldspathic rock by sulfate solutions" (Brown, 1969). Therefore, the size of the alunite field presented in the diagrams should be considered a practical minimum.

Ca-montmorillonite was included instead of K-montmorillonite since that mineral does not form in the alteration assemblages and since the size of the montmorillonite field would be underestimated otherwise. Since gypsum is a common mineral in advanced argillic assemblages, Ca^{+2} activities were chosen to be the saturation value with respect to gypsum for any given SO_4^{-2} activity and temperature.

Owing to the difficulties inherent in estimating the activities of aluminum species, all reactions were written to conserve aluminum.

Two types of diagrams were calculated. The first is a three dimensional presentation of the stability relationships among the phases as a function of log potassium, sulfate and hydrogen ion activities. These diagrams are particularly attractive because they include, indirectly, the effects of fH_2S . Saturation with respect to both gypsum and quartz is assumed.

The most striking feature of these diagrams is the sensitivity of the kaolinite volume to temperature. Above 150° centigrade the kaolinite volume is almost entirely displaced by that of montmorillonite (Figure 41). Since one of the important characteristics of the argillic and advanced argillic zones is abundant and often massive kaolinite, this is taken as evidence that the altering solutions must not have exceeded 150° centigrade.

Another important observation is that alunite is stable with respect to the other phases at moderate pH's and ion activities. In Figure 42, horizontal

Figure 41. Activity diagram showing stability relations of Microcline, Ca-montmorillonite, alunite, and kaolinite at 150°C with gypsum, quartz and unit activity of water.

Figure 42. Activity diagram showing stability relations of microcline, Ca-montmorillonite, alunite and kaolinite at 60° C with gypsum, quartz and unit activity of water.

planes through the diagram at pH five and six show the range of sulfate and potassium concentrations in equilibrium with alunite for the given pH values. At a pH of five the alunite field intersects the composition of the Red Hill spring water. It is therefore apparent that the presence of jarosite in the advanced argillic zone does not necessarily indicate that the altering solutions were highly acidic.

At a $f\text{CO}_2$ of four, approximately equivalent to saturation at a depth of 150 feet if a hydrostatic pressure gradient is assumed, calcite first intersects the alunite volume for a temperature of 60° centigrade. However, greater pressures, corresponding to depths exceeding 500 feet are required to precipitate calcite with alunite at reasonable ion activities.

A series of isothermal diagrams were also calculated using the activities of silica and pH as the independent variables (Figure 43). These diagrams correspond to vertical lines through the three dimensional diagrams at the potassium and sulfate concentrations of the Red Hill Spring, considered to represent a minimum estimate for the hydrothermal solutions.

It is assumed that the silica concentrations of the solutions will never fall below the quartz saturation values, provided quartz is a participating phase. This assumption is supported by the observation that secondary polycrystalline quartz occasionally forms as a pseudomorph after ferromagnesium aluminosilicates. On the other hand, supersaturation with respect to quartz is likely, provided the solutions have cooled following last wall rock equilibration. The diagrams may be visualized as depicting the reaction of various source solutions possessing quartz saturation temperatures ranging between the saturation value for a particular diagram and a practical maximum, 275° centigrade. For this reason the silica concentrations have been reported both in terms of molalities and in terms of the corresponding quartz saturation temperature.

Figure 43. Activity diagrams showing stability of mineral phases as a function of pH and silicic acid activity. Saturation with gypsum is assumed and activities of sulfate and potassium identical to Red Hill Spring.

Regardless of the temperature at which the reactions are calculated, kaolinite cannot form stably if the solutions entering into the reactions have equilibrated with quartz at temperatures above 150° centigrade. Therefore a maximum limit on the reservoir temperature of 150° centigrade is established. For appreciable kaolinite formation, however, lower temperatures between 50° and 100° are proposed. The kaolinite field would, of course, be enhanced at higher temperatures if lower ion activities than those at the Red Hill Spring are chosen. It is unlikely that this would represent a realistic situation, however.

The difficulty of precipitating muscovite is apparent. It is most likely to precipitate at lower temperatures where the quartz saturation line approaches the muscovite field and at neutral pH's. It may form at high temperatures provided the reactions are confined to a silica deficient environment. This observation is in good agreement with the lack of extensive sericite replacement in the KGRA. The stability field of mixed layer potassium micas has not been estimated. However, it would certainly encroach significantly on both the kaolinite and montmorillonite fields, making it an easily attainable phase. The Red Hill and Monroe Spring waters plot squarely within the kaolinite field as would be predicted based upon the dominant argillic alteration associated with these springs. Measurements of seven major springs occupying the Monroe tufa mound revealed that pH is negatively correlated with temperature. An arrow has been drawn in Figure 14 to indicate the migration of these solutions toward the muscovite field.

The sequence of hydrothermal alteration observed at Monroe could have been generated by a hydrothermal solution not unlike the present spring waters. An initial solution composition approximating that of the present springs with a temperature of 60° centigrade and a pH of five is proposed. Particularly

if the source solution is more concentrated in sulfate and potassium than the present springs an initial pH of five is more than sufficient to account for the occurrence of jarosite. With progressive reaction with salic host rock the composition of the solution would move first into the kaolinite field and then to the montmorillonite-kaolinite field boundary, generating an argillic and propylitic alteration assemblage respectively. Calcite would coexist with kaolinite as a pH of approximately six is reached. Muscovite or sericite would precipitate as late phases, produced by spent hydrothermal fluids with low temperatures and hydrogen ion activities. Superimposition of different alteration assemblages occurs as the hydrothermal system cools and contracts. Gypsum replacing jarosite could have formed as the advanced argillic regime was superceeded by an argillic regime in which jarosite was not stable.

The peculiar association of alunite and jarosite with minor calcite at Goldfield, Nevada and Monroe respectively, is probably a metastable relationship. The volume of calcite is small in both cases. Furthermore, owing to the fact that only very small variations in pH, less than one log unit, are sufficient to move from the stability field of alunite or jarosite into that of kaolinite and calcite, it is surprising that these associations are not more common.

The formation of chlorite appears to represent a special case. Chlorite always forms in close association with biotite but never as a replacement product. Therefore, chlorite must form from solutions that are in partial equilibrium with biotite.

The initial solutions appear to have high potassium activities. For this reason calcium phases are consumed preferentially and potassium feldspar and sericite may persist metastably. However, as Ca^{+2} activity rises in response to the consumption of sulfate, these phases are also consumed.

Summary

The Monroe KGRA represents the peripheral oxidized zone of an active sulfide containing system. The near surface environment represented by the exposed alteration is consistent with sulfate solutions with temperatures between 50⁰ and 100⁰ centigrade. Moderate pH's of five or greater are sufficient to account for the advanced argillic alteration regime. The overall abundance of calcite suggests that high partial pressures of carbon dioxide, generated by reaction of hydrothermal solutions with sedimentary carbonates, has been an important aspect of the system over its entire history.

The spring water is consistent with the alteration found in the KGRA. Its relationship to the other springs and wells in the vicinity suggest that it is a high temperature brine which has been diluted by cool nonsaline groundwaters. The source brines must exceed 100⁰ centigrade and have high concentrations of sodium chloride, sulfate and potassium.

A model is proposed for the system, whereby hot thermal brines are heated at depth and are confined beneath the impermeable Carmel shales. Meteoric water, entering the groundwater system on the Sevier Plateau descends the Sevier monocline through permeable volcanic rock and is slightly heated by the thermal gradient. The two solutions mix at the site of the springs where major faults interrupt the impermeable cap rock. The temperature of mixing has never exceeded 150⁰ centigrade but much higher temperatures are possible for the undiluted brines.

The techniques of multivariate analysis have been applied successfully to the analysis of alterations dynamics. The methods of principle component analysis and Q mode cluster analysis show great promise for sorting out meaningful alteration assemblages and determining reaction relationships where the alteration zones are poorly developed and superimposed. In particular, PCA has been useful in isolating the major variables of the Monroe KGRA.

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