

**GEOHERMAL RESOURCE
EVALUATION AT CASTLE HOT SPRING,
ARIZONA
FINAL REPORT**

by

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INTRODUCTION

The purpose of this report is to define the geothermal system that supplies the hot water for Castle Hot Spring. An attempt was made to specify the "reservoir" temperature and to model possible sources of heat. This report summarizes our findings and presents the important supporting information in the various tables and figures. Additional data, including details of the techniques and methodologies used in the investigation, will be recorded in an Arizona State University MS thesis (Satkin, in preparation, 1980). Appendices I and II are two papers resulting from the funded work on Castle Hot Spring.

Our geologic feasibility study consisted of five parts: 1) construction of a detailed geologic map of hot spring area, 2) production of geologic cross sections, 3) calculation of water geothermometry based on repeated analyses, 4) measurement of geothermal profiles in shallow wells, and 5) speculation on models for the heat source.

During the initial phase of this investigation a topographic base map with 5-m contour intervals was constructed by plane-table mapping at a scale of 1:1000. Geologic data were plotted directly on the base allowing the construction of accurate cross sections showing the relationship of the springs to the fault system. Rock samples were collected for potassium-uranium-thorium (KUT), whole-rock, and petrographic analysis. In addition, thermal and non-thermal waters were periodically collected for chemical and isotopic analysis. Thermal gradient measurements were attempted at several shallow wells through the cooperation of Claudia Stone Arizona Bureau of Geology and Mineral Technology (ABGMT).

GEOLOGIC MAP

Geologic Setting

Castle Hot Spring (Figure 1) is the highest temperature hot spring in the Transition Zone northwest of Phoenix (Swanberg and others, 1977). Although the chemistry of the water is similar to other hot springs within the Transition Zone, its purity, high sulfate, and high fluorine (Mariner and others, 1977) are anomalous. The spring is located along the northern margin of a northwest-trending graben of Tertiary volcanic rocks displaced into a Precambrian basement complex to the south of the Bradshaw Mountains (Figure 2). The emergence of Castle Hot Spring and several other related springs is controlled by the fault contact between the basement and volcanic rocks.

Detailed geologic data plotted on the topographic base map were reduced to a scale of 1:2,000 for the final geologic map (Figure 3). Although the geology is fairly complex and alteration is intense near the main spring system, the basic relationships are clearly illustrated. Northeast dipping volcanic rocks south of the fault system are in fault contact with Precambrian basement rocks to the north. Carbonate alteration of the volcanic rocks and surface deposits of travertine occur along the contact. Terrace deposits 40-80 m above Castle Creek indicate filling with Tertiary-Quaternary gravels during an early phase of the hot spring activity. These gravels have subsequently been exhumed forming the present drainage, but the main spring system remains at an elevation well above the present stream level. An extensive area of hydrothermal alteration to the northwest with associated siliceous sinter deposits is additional evidence of an older thermal system at this location.



Figure 1. Location map for Castle Hot Spring.

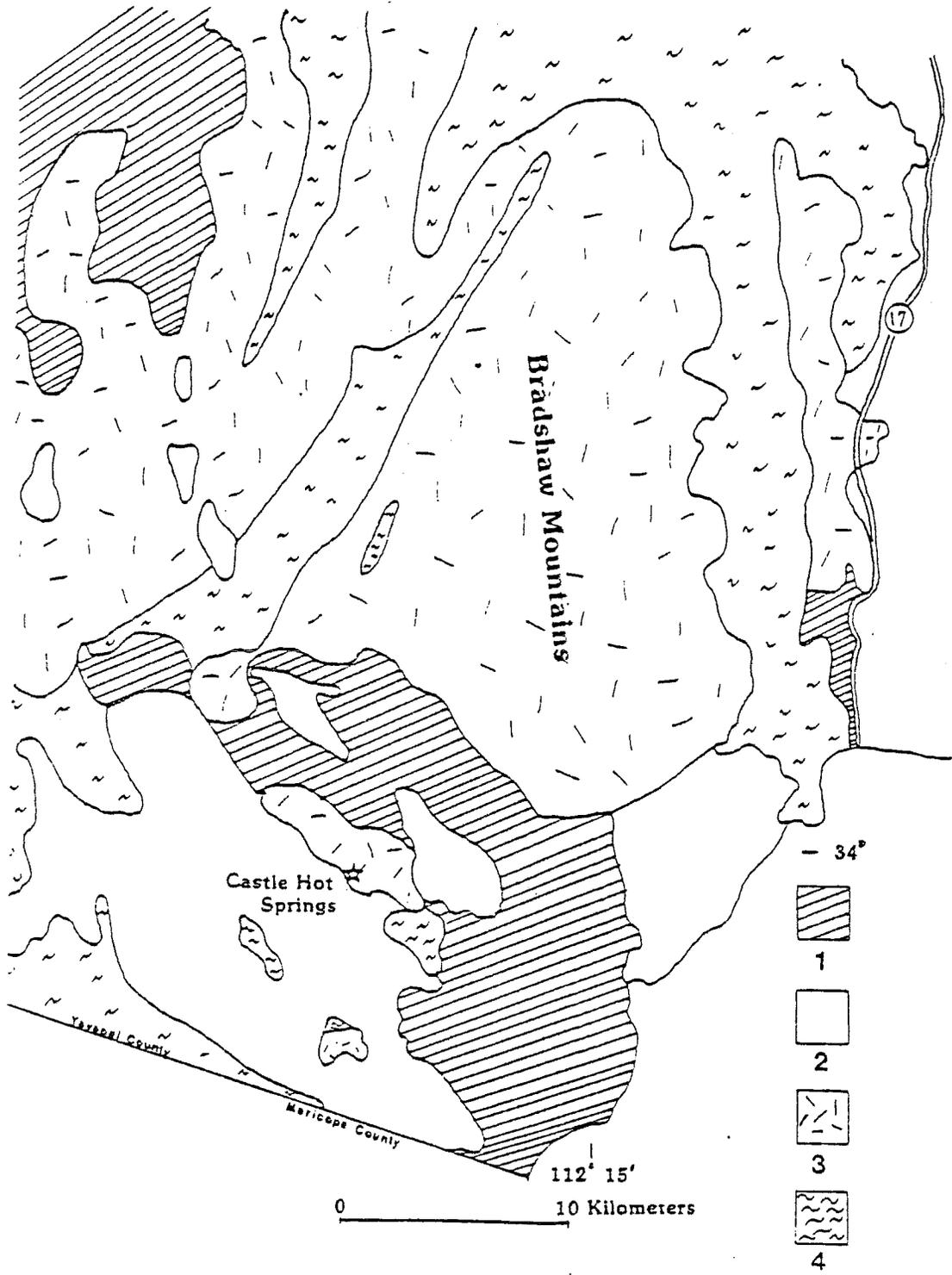
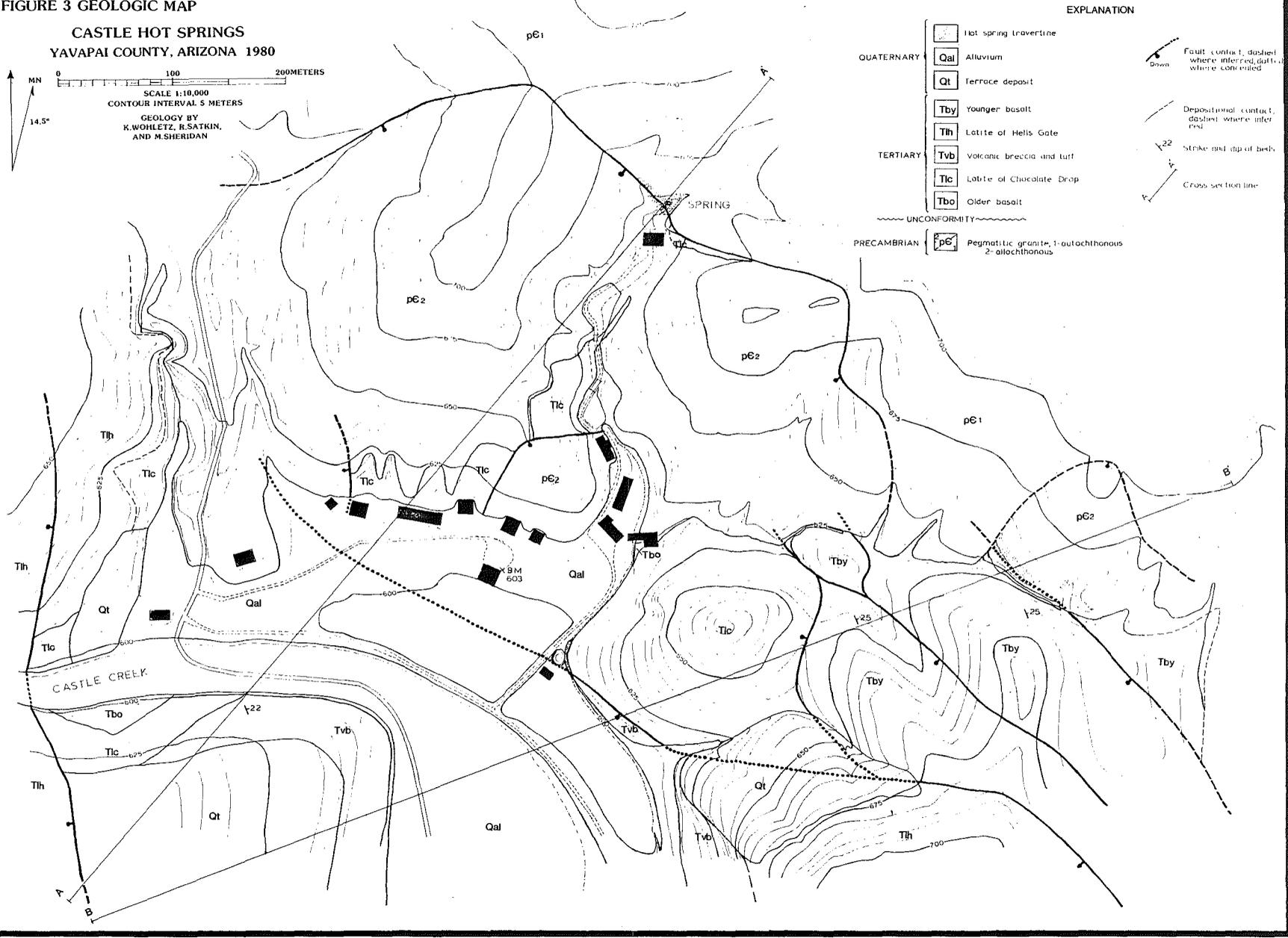


Figure 2. Regional geologic sketch map. 1) Quaternary to Tertiary sedimentary rocks. 2) Tertiary to Cretaceous (?) volcanic rocks. 3) Precambrian (?) granitic rocks. 4) Precambrian metamorphic rocks.

FIGURE 3 GEOLOGIC MAP

CASTLE HOT SPRINGS
YAVAPAI COUNTY, ARIZONA 1980

MN
0 100 200 METERS
SCALE 1:10,000
CONTOUR INTERVAL 5 METERS
GEOLOGY BY
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AND M. SHERIDAN



EXPLANATION

- | | | |
|-------------|---|---|
| QUATERNARY | [Stippled Box] Hot spring travertine | [Dashed Line with 'inf'] Depositional contact, dashed where inferred |
| | [Box with 'Qal'] Alluvium | [Dashed Line with 'inf'] Fault contact, dashed where inferred, dash-dot where concealed |
| | [Box with 'Qt'] Terrace deposit | [Dashed Line with 'inf'] Strike and slip of beds |
| TERTIARY | [Box with 'Tby'] Younger basalt | [Dashed Line with 'inf'] Cross section line |
| | [Box with 'Th'] Latite of Hells Gate | |
| | [Box with 'Tvb'] Volcanic breccia and tuff | |
| | [Box with 'Tlc'] Latite of Chocolate Drop | |
| | [Box with 'Tbo'] Older basalt | |
| | [Wavy Line] UNCONFORMITY | |
| PRECAMBRIAN | [Box with 'pC1'] Pegmatitic granite, 1- autochthonous | |
| | [Box with 'pC2'] Pegmatitic granite, 2- allochthonous | |

Stratigraphy

The area including Castle Hot Spring was recently studied by Ward (1977) who provided a geologic map (scale of 1:48,000), a stratigraphic section, several whole-rock chemical analyses, and the structural and tectonic setting. The generalized stratigraphy of the area (Sheridan and Others, 1979, see Appendix I) consists of Precambrian schists and granitic rocks overlain unconformably by 300-450 m of basaltic lavas and rhyolitic tuffs which are in turn overlain by approximately 120 m of latitic epiclastic breccia and lava flows.

On the basis of detailed field mapping and new chemical analyses the rocks at Castle Hot Spring correlate quite well with the stratigraphy of Ward (1977). Although no new regional map units were found, a new local unit (Chocolate Drop latite), correlated in time and space with the Hell's Gate latite, was designated on the map. This unit has an unusual chemistry (high K_2O , U, and Th) and is strongly altered suggesting an intimate relationship with the present geothermal system. Additional data on rock analyses and petrogenesis are given by Ward (1977) and Satkin (in preparation, 1980).

Alteration

Hydrothermal alteration is widespread within the volcanic rocks in the vicinity of Castle Hot Spring. Carbonitization is the most abundant type of alteration, commonly with 5 to 15 percent replacement of phenocrysts and groundmass by calcite. Calcite stockwork veining and fracture coatings are widespread. Some of the mafic volcanic rocks in the area are stained red, indicating oxidation and hydration of iron-bearing minerals. For example, hornblende phenocrysts in the Chocolate Drop latite have been replaced by opaque masses of secondary iron-bearing oxides and hydroxides.

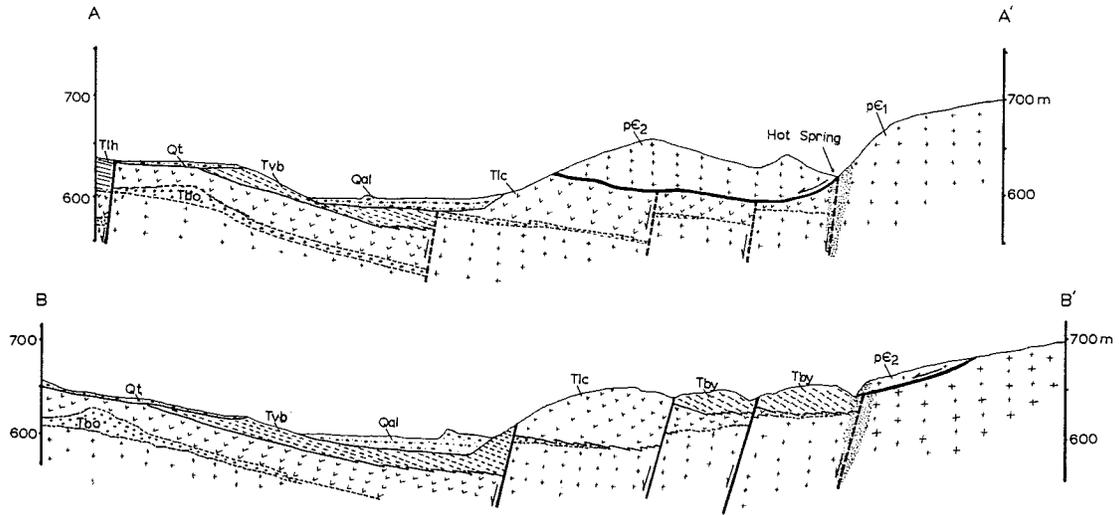
Chemical analyses also show that some of the volcanic rocks have been strongly depleted in Na^+ but augmented in K^+ , Ca^{++} , and CO_3^- . The source of the alteration products along the Castle hydrothermal system is believed to be fluids similar to the present geothermal solutions. This assumption is supported by the deposition of travertine along active springs and the discontinuous outcrops of travertine throughout the map area at locations of extinct hot springs. The intensity of alteration increases to the northwest where widespread argillitization has obliterated the primary characteristics of the volcanic rocks (Ward, 1977). The presence of large siliceous sinter deposits in this area attests to a higher temperature hydrothermal system in the past.

CROSS SECTIONS

The location of Castle Hot Spring and other related thermal springs is structurally controlled by a complex system of basin-and-range faults. The largest displacements noted by Ward (1977) occur along northwest-trending en echelon faults. A north-striking fault of moderate displacement (100-200 m) which is covered by a gravel in Castle Creek (Figure 3) possibly intersects the graben fault system at Castle Hot Spring. In addition there are several faults of low displacement (10-50 m) with northwest or northeast strikes that account for minor adjustments at the graben margin.

Plane table mapping (Figure 3) has documented a low-angle listric fault that displaces an allochthonous block of Precambrian granite on top of the Tertiary chocolate drop latite (Figure 4). The hot spring occur where erosion has exposed the intersection of the listric fault block and the north wall of the main graben. This north wall is cut by a deeply penetrating northwest trending fault system, here termed the Castle fault, that separates Precambrian granitic rocks to the

FIGURE 4 GEOLOGIC CROSS SECTIONS



northeast from Tertiary volcanics on the southwest.

The slump block has been intruded by latite dikes that cut through the listric fault system. Displacements of these latite dikes, which have the same composition and texture as the chocolate drop latite, by later faulting indicates alternating episodes of volcanism and faulting. The main displacement in the Castle fault system must have preceded the chocolate drop latite which fills the main graben. The later basalts and latites are only off-set by minor faults and their distribution is not controlled by the graben.

Several other thermal springs, including Henderson Ranch Spring, Alkali Spring and the Dodd Well, occur along the 2 km, N. 45° W. Castle fault (Figure 5), which coincides with the trend of the main fault system. These springs all emanate at an elevation of approximately 640 m above sea level suggesting an apparent hydrostatic relationship. Several other springs with geothermal potential, Casa Rosa Spring, Dripping Spring, Mud Spring and Spring Valley, also occur along deeply penetrating faults that separate Precambrian granites and schist from Tertiary rocks. The relationship of these later springs to the Castle system is not known.

WATER GEOTHERMOMETRY

Background

Chemical analyses of spring and well waters may be used to estimate subsurface temperatures by applying chemical geothermometers. These thermometers assume that the chemical composition of the water reflects the last temperature of equilibration between the thermal fluid with the surrounding rock. This temperature is generally assigned to represent the reservoir temperature of the thermal system. It must be remembered that for the Castle Hot Spring system the hydrologic

nature of the reservoir is not yet well understood.

The silica geothermometer (Fournier and Rowe, 1966) is one of the most widely applied and reliable geothermometers because it is little affected by precipitation, base-exchange, salinity, and pressure. However, interpretation of the silica geothermometer requires an assumption of the silica phase that controls the dissolved silica content of the thermal water, i.e., amorphous (chalcedony) or crystalline (quartz). For Castle Hot Spring chalcedony was considered to be a more reliable predictor of the silica content; therefore, the chalcedony equation (Table I) was used to estimate the subsurface reservoir temperature.

Geothermometer temperature estimates are considered more reliable if results from several techniques agree. Therefore Na-K and Na-K-Ca geothermometers (Ellis and Mahon, 1967; Fournier and Truesdell, 1973; and Fournier and Potter, 1979) have also been applied for independent temperature estimates. The equations used to calculate geothermometers presented in this report are given in Table I. Several assumptions must be made when using chemical analyses of spring and well waters to predict subsurface temperatures (Fournier and others, 1974). The following assumptions are applicable to the silica, Na-K, and Na-K-Ca geothermometers:

- (1) Temperature dependent reactions between water and rock occur at depth.
- (2) All constituents involved in temperature-dependent reactions are sufficiently abundant (i.e., supply is not a limiting factor).
- (3) Chemical equilibrium is reached at the reservoir temperature.

Table I. Equations for geothermometers

Silica Geothermometers (SiO ₂ in ppm)	
SiO ₂ (quartz, adiabatic cooling)	$t^{\circ}\text{C} = \frac{1522}{5.75 - \log \text{SiO}_2} - 273.15$
SiO ₂ (quartz, conductive cooling)	$t^{\circ}\text{C} = \frac{1309}{5.19 - \log \text{SiO}_2} - 273.15$
SiO ₂ (chalcedony, conductive cooling)	$t^{\circ}\text{C} = \frac{1032}{4.69 - \log \text{SiO}_2} - 273.15$
SiO ₂ (α - cristobalite, conductive cooling)	$t^{\circ}\text{C} = \frac{1000}{4.78 - \log \text{SiO}_2} - 273.15$
Na-K-Ca Geothermometers (Na, K, Ca in moles/liter)	
Na-K-Ca	
B = 4/3 for $\sqrt{\text{Ca}/\text{Na}} > 1$ and $t < 100^{\circ}\text{C}$	$t^{\circ}\text{C} = \frac{1647}{\log(\text{Na}/\text{K}) + B \log(\sqrt{\text{Ca}/\text{Na}}) + 2.24} - 273.15$
B = 1/3 for $\sqrt{\text{Ca}/\text{Na}} < 1$ or $t_{4/3} > 100^{\circ}\text{C}$	
Na-K Geothermometer (Na, K in ppm)	
Na-K	$t^{\circ}\text{C} = \frac{855.6}{\log(\text{Na}/\text{K}) + 0.8573} - 273.15$

- (4) Little or no re-equilibration or change in composition occurs at lower temperatures as the water flows from the reservoir to the surface.
- (5) The hot water coming from deep in the system does not mix with cooler shallow ground water.

Thermal Waters of the Study Area

Non-thermal and thermal springs and wells in the study area were divided into three groups (Sarkin and others, 1980; see appendix II). Group I thermal waters emanate from the Castle fault system and are of the sodium-chloride-sulfate type. Group II waters are from non-thermal springs and wells, and have low concentrations of total dissolved solids. Group III waters issue from perennial springs that have a higher salinity than Group I or II waters.

Because interpretation of geothermal waters is strongly dependent on the quality of the analyses, care was taken to present the best possible data in this report. Analyses of SiO_2 , Na^+ , K^+ , Ca^{++} , Mg^{++} , and Li^+ were made using Varian 1250 atomic absorption spectrophotometer at Arizona State University. A Dionex 10 ion chromatograph was used to analyse F^- , Cl^- , and $\text{SO}_4^{=}$. Total dissolved solids were determined on filtered untreated samples by the residue-on-evaporation method (Rainwater and Thatcher, 1960). Field measurements of pH were made with a Photovolt model 126A pH meter. Temperatures were measured with an Extech 1200 digital thermometer. A comparison of our analyses of Castle Hot Spring water with those of other laboratories is given in Table II.

Group I Waters

The sources of Group I thermal waters are Castle Hot Spring, Henderson Ranch Spring, Alkali Spring, Mesquite Drip; and the Dodd Well

Table II. Comparative Analyses of Castle Hot Spring.

Concentration (ppm)	This study mean	Engineers Testing Laboratories (1975)	USGS Open File Rpt. 77-654 (1977)	USGS Greg Littin (1979)	Rocky Mountain Geochemical (1980)
Temperature °C	50.4	48	46	45	n.d.
pH	7.8	8.1	7.5	7.7	n.d.
Total Dissolved solids	662	706	820	750	645
SiO ₂	61	n.d.	58	64	47
Na ⁺	207	198	200	220	182
K ⁺	5.4	7.4	4.7	5.7	6
Ca ⁺⁺	31	32	33	29	32
Mg ⁺⁺	2.4	2.7	2.4	2.3	1.0
Li ⁺	0.31	n.d.	0.29	n.d.	0.49
F ⁻	8.6	8.0	n.d.	3.1	9.1
Cl ⁻	143	145	n.d.	140	1610
SO ₄ ⁼	205	156	n.d.	220	265
HCO ₃ ⁻	134	163	n.d.	130	176
n.d. = not determined					

(Figure 5). They display a homogenous chemistry (Table III), which suggests that they originate from the same geothermal reservoir. The waters are a sodium-chloride-sulphate type and possess relatively high concentrations of SiO_2 , Li, and F, and low Mg. Underscored reservoir temperatures estimated from silica (chalcedony), Na-K, and Na-K-Ca geothermometers are similar for all Group I waters (Table III). These waters probably arise from a deep reservoir with temperature below 100°C .

Part of our program of water chemistry was to evaluate the consistency of water composition of Group I springs with time. Samples were collected and analysed at approximately one-month intervals for a period of one year at Castle Hot Spring and Henderson Ranch Spring. The small variations noted in the data (Tables IV and V) indicate that rainfall or seasonal temperature variations have no perceptible effect on the chemistry or temperature of the springs.

The discrepancy between the low measured surface temperatures in Group I waters and the higher temperatures estimated from chemical geothermometry suggests that these thermal waters may have cooled either by conduction, by boiling (adiabatically), by mixing with shallow cold ground water, or by combinations of these processes. Adiabatic cooling of Castle Hot Spring is unlikely because the estimated reservoir temperature is not hot enough for boiling to occur. Because of the large flow rate and the quick ascent of water to the surface at Castle Hot Spring, heat loss through conduction is considered negligible. However, conductive cooling is possible for springs with a large flow rate if the water moves laterally for long distances at a shallow level (Fournier, 1977, 1979). For the other waters of Group I, besides Castle Hot Spring, conductive cooling is strongly suggested by: 1) small flow rate; 2) similar chloride

Table III. Selected water analyses and geothermometry of warm springs and wells of Group I discharging along major fault system bounding Precambrian granites and Tertiary volcanics in Castle Hot Spring area. Analyses in ppm (mg/l).

Name	Castle Hot Spring	Henderson Ranch Spring	Alkalai Spring
Location	T8N, R1W, 34, SW $\frac{1}{4}$, SW $\frac{1}{4}$	T8N, R1W, 33, NW $\frac{1}{4}$	T8N, R1W, 33, NW $\frac{1}{4}$, SE $\frac{1}{4}$
Date sampled	May 12, 1980	Jan 9, 1980	Jan 9, 1980
Flow Rate, l/min	879	10	8
Temperature, °C	47.7	29.9	31.2
Field pH	7.9	7.7	7.9
Total dissolved solids	664	818	642
SiO ₂	62	58	71
Na ⁺	203	234	215
K ⁺	5.5	6.8	6.3
Ca ⁺⁺	31	37	16
Mg ⁺	2.5	2.0	0.23
Li ⁺	0.32	0.51	0.42
F ⁻	8.6	7.4	12
Cl ₂ ⁻	143	147	135
SO ₄	206	312	209
Geothermometry			
SiO ₂ (quartz, adiabatic)	112	108	117
SiO ₂ (quartz, conductive)	112	109	119
SiO ₂ (chalcedony)	83	79	90
SiO ₂ (α -cristobalite)	62	58	68
Na-K-Ca* ($\beta=1/3$)	114**	122	128
Na-K-Ca ($\beta=4/3$)	77	82	98
Na-K	80	85	85

*Sample taken from tap

* Na-K-Ca (Mg corrected)

Table III . Continued.

Name	Mesquite Drip	Dodd Well**
Location	T8N, R1W, 33, NW $\frac{1}{4}$, SE $\frac{1}{4}$	T8N, R1W, 33, NW $\frac{1}{4}$, NW $\frac{1}{4}$
Date sampled	Dec 20, 1979	May 12, 1980
Flow Rate, l/min	seep	4-13
Temperature, °C	26.3	23.6
Field pH	8.7	8.0
Total dissolved solids	740	760
SiO ₂	73	63
Na ⁺	264	239
K ⁺	7.4	7.3
Ca ⁺⁺	18	26
Mg ⁺	0.34	0.43
Li ⁺	0.50	0.49
F ⁻	14	8.2
Cl ⁻	153	142
SO ₄ ⁼	243	288
Geothermometry		
SiO ₂ (quartz, adiabatic)	119	112
SiO ₂ (quartz, conductive)	121	113
SiO ₂ (chalcedony)	<u>92</u>	<u>84</u>
SiO ₂ (α -cristobalite)	70	62
Na-K-Ca* ($\beta=1/3$)	127	127
Na-K-Ca ($\beta=4/3$)	<u>103</u>	<u>93</u>
Na-K	<u>82</u>	<u>87</u>

* Sample taken from tap

* Na-K-Ca (Mg corrected)

Table IV. Chemical variation through time of the main Hot Spring's system at Castle Hot Spring.
in ppm (mg/l).

Location	T8N, R1W, 34, SW $\frac{1}{4}$, SW $\frac{1}{4}$				
Date	10/9/79	10/24/79	11/27/79	12/20/79	1/9/80
Temperature, °C	51.3	55.4	54.7	52.7	53.4
Field pH	7.6	7.7	7.9	7.8	7.7
Total dissolved solids	n.d.	640	n.d.	648	n.d.
SiO ₂	60	63	61	60	59
Na ⁺	209	209	208	211	195
K ⁺	5.0	5.5	5.4	5.5	5.6
Ca ⁺⁺	30	34	32	32	29
Mg ⁺⁺	2.4	3.0	2.3	2.6	2.4
Li ⁺	n.d.	n.d.	0.34	0.33	0.32
F ⁻	8.5	9.2	8.5	8.7	8.5
Cl ⁻	147	155	145	141	140
SO ₄ ⁼	212	230	211	211	206
Geothermometry (T°C)					
SiO ₂ (quartz, adiabatic)	110	112	111	110	109
SiO ₂ (quartz, conductive)	110	113	112	111	109
SiO ₂ (chalcedony)	81	84	82	82	80
SiO ₂ (α -cristobalite)	60	63	61	60	59
Na-K-Ca* ($\beta=1/3$)	114	111	117	113	114
Na-K-Ca ($\beta=4/3$)	75	75	76	77	78
Na-K	72	78	77	77	83
n.d. = not determined *Na-K-Ca (Mg corrected)					

Table IV. Continued.

Location	T8N, R1W, 34, SW $\frac{1}{4}$, SW $\frac{1}{4}$				
Date	2/3/80	3/7/80	4/10/80	5/12/80	6/12/80
Temperature, °C	52.1	49.3	47.6	47.7	47.7
Field pH	7.9	7.8	7.9	7.9	7.9
Total dissolved solids	646	n.d.	662	664	692
SiO ₂	59	62	62	62	63
Na ⁺	200	202	222	203	206
K ⁺	5.4	5.6	5.4	5.5	5.5
Ca ⁺⁺	30	30	31	31	32
Mg ⁺⁺	2.3	2.4	2.4	2.5	2.4
Li ⁺	0.31	0.31	0.30	0.32	0.30
F ⁻	8.6	8.5	8.3	8.6	8.5
Cl ⁻	141	138	140	143	142
SO ₄ ⁼	200	196	189	206	200
Geothermometry (T°C)					
SiO ₂ (quartz, adiabatic)	110	111	111	112	112
SiO ₂ (quartz, conductive)	110	112	112	112	113
SiO ₂ (chalcedony)	<u>81</u>	<u>83</u>	<u>83</u>	<u>83</u>	<u>84</u>
SiO ₂ (α -cristobalite)	59	61	62	62	62
Na-K-Ca* ($\beta=1/3$)	115	114	114	114	116
Na-K-Ca ($\beta=4/3$)	<u>77</u>	<u>79</u>	<u>77</u>	<u>77</u>	<u>77</u>
Na-K	<u>79</u>	<u>81</u>	<u>73</u>	<u>80</u>	<u>79</u>
n.d. = not determined		*Na-K-Ca (Mg corrected)			

Table IV. Continued.

Location	T8N, R1W, 34, SW $\frac{1}{4}$, SW $\frac{1}{4}$		
Date	7/17/80	8/15/80	9/11/80
Temperature, °C	47.7	47.8	47.7
Field pH	7.9	7.9	8.0
Total dissolved solids	672	650	686
SiO ₂	59	62	62
Na ⁺	206	208	206
K ⁺	5.2	5.2	5.3
Ca ⁺⁺	32	32	32
Mg ⁺⁺	2.4	2.3	2.2
Li ⁺	0.30	0.30	0.31
F ⁻	8.7	8.6	8.4
Cl ⁻	143	142	140
SO ₄ ⁼	211	204	194
Geothermometry (T°C)			
SiO ₂ (quartz, adiabatic)	109	111	111
SiO ₂ (quartz, conductive)	110	112	112
SiO ₂ (chalcedony)	<u>80</u>	<u>83</u>	<u>83</u>
SiO ₂ (α -cristobalite)	59	62	62
Na-K-Ca* ($\beta=1/3$)	115	115	117
Na-K-Ca ($\beta=4/3$)	<u>75</u>	<u>75</u>	<u>75</u>
Na-K	<u>75</u>	<u>75</u>	<u>77</u>
n.d. = not determined	*Na-K-Ca (Mg corrected)		

Table V. Chemical variation through time of the Henderson Ranch Spring near Castle Hot Spring, Arizona. Analyses in ppm (mg/l).

Location	T8N, R1W, 33, NW $\frac{1}{4}$				
Date	10/9/79	10/24/79	11/27/79	12/20/79	1/9/80
Temperature, °C	30.3	32.5	29.2	29.1	29.9
Field pH	7.5	7.9	7.7	7.7	7.7
Total dissolved solids	n.d.	n.d.	n.d.	798	818
SiO ₂	60	62	60	57	58
Na ⁺	248	234	234	250	234
K ⁺	6.6	7.2	7.3	7.1	6.8
Ca ⁺⁺	39	41	40	40	37
Mg ⁺⁺	2.1	2.2	2.2	2.2	2.0
Li ⁺	n.d.	n.d.	0.55	0.53	0.51
F ⁻	7.4	6.6	7.5	7.4	7.4
Cl ⁻	150	142	150	141	147
SO ₄ ⁼	306	301	299	306	312
Geothermometry (T°C)					
SiO ₂ (quartz, adiabatic)	110	111	110	108	108
SiO ₂ (quartz, conductive)	110	112	111	108	109
SiO ₂ (chalcedony)	81	83	82	79	79
SiO ₂ (α -cristobalite)	60	61	60	58	58
Na-K-Ca* ($\beta=1/3$)	119	124	125	122	122
Na-K-Ca ($\beta=4/3$)	81	81	83	82	82
Na-K	79	88	89	83	85

n.d. = not determined

*Na-K-Ca (Mg corrected)

Table V. Continued.

Location	T8N, R1W, 33, NW¼				
Date	2/3/80	3/7/80	4/10/80	5/12/80	6/12/80
Temperature, °C	28.5	28.6	28.8	29.0	29.3
Field pH	7.7	7.5	7.6	7.3	7.9
Total dissolved solids	782	920	n.d.	900	902
SiO ₂	56	59	59	63	60
Na ⁺	240	285	274	261	260
K ⁺	6.7	8.0	7.2	7.4	6.6
Ca ⁺⁺	38	53	52	45	45
Mg ⁺⁺	2.1	3.6	3.0	2.8	2.6
Li ⁺	0.49	0.60	0.54	0.54	0.52
F ⁻	7.4	6.9	6.9	7.3	7.1
Cl ⁻	142	168	159	158	155
SO ₄ ⁼	315	372	345	325	332
Geothermometry (T°C)					
SiO ₂ (quartz, adiabatic)	107	109	110	112	110
SiO ₂ (quartz, conductive)	107	110	110	113	110
SiO ₂ (chalcedony)	77	80	81	84	81
SiO ₂ (α-cristobalite)	56	59	59	62	60
Na-K-Ca* (β=1/3)	121	119*	118	121	117
Na-K-Ca (β=4/3)	81	82	78	81	78
Na-K	81	82	78	83	76
n.d. = not determined		*Na-K-Ca (Mg corrected)			

Table V. Continued.

Location	T8N, R1W, 33, NW¼		
Date	7/17/80	8/15/80	9/11/80
Temperature, °C	30.0	30.4	29.9
Field pH	7.7	7.5	7.9
Total dissolved solids	844	850	884
SiO ₂	58	59	61
Na ⁺	254	254	253
K ⁺	6.8	6.8	6.9
Ca ⁺⁺	45	46	42
Mg ⁺⁺	2.4	2.5	2.3
Li ⁺	0.50	0.50	0.52
F ⁻	7.4	7.4	7.1
Cl ⁻	153	152	148
SO ₄ ⁼	329	317	321
Geothermometry (T°C)			
SiO ₂ (quartz, adiabatic)	108	109	111
SiO ₂ (quartz, conductive)	109	109	111
SiO ₂ (chalcedony)	79	80	82
SiO ₂ (α-cristobalite)	58	59	61
Na-K-Ca* (β=1/3)	119	118	120
Na-K-Ca (β=4/3)	78	78	80
Na-K	79	79	80
n.d. = not determined		*Na-K-Ca (Mg corrected)	

content but different measured surface temperatures, and 3) decreasing surface temperatures of waters with increasing distance from the main spring system.

Group II Waters

Group II non-thermal waters include Chuck's Well, Menundo Spring, Layton Seep, and Windmill Well (Satkin and others, 1980). They are characterized by low total dissolved solids (380-580 ppm) and low concentrations of SiO_2 , Li^+ , and F^- . The variable geochemistry suggests an origin through mixing of thermal water with near surface aquifers. These waters can be compared with thermal springs by calculation of mixing models. Chuck's Well was selected for periodic analysis throughout the year in order to evaluate the role of climatic variations on the chemistry of Group II waters (Table VI). The similarity of analyses suggests that the chemistry is little affected by ambient temperature or rainfall.

Group III Waters

Sources of Group III waters include Casa Rosa Spring, Dripping Spring, Mud Spring, and the Kent Well. These waters also display a homogeneous chemistry (Table VII) and possess highly enriched concentrations of Na^+ , Ca^{++} , Li^+ , Cl^- and $\text{SO}_4^{=}$, and relatively low Mg^{++} . These waters are characterized by a high salinity that is presumed to be derived from interaction with alluvium and/or dissolution of secondary mineral coatings of halite, gypsum, and calcite found on cobbles and clasts in the alluvium and epiclastic breccia. In spite of the low measured surface temperatures, the silica (chalcedony), Na-K, and Na-K-Ca geothermometers agree quite well (Table VII) and indicate a reservoir temperature between 72-85°C. These waters probably cooled conductively as did the Group I springs.

Table VI. Chemical variation through time of Chuck's Well near Castle Hot Spring, Arizona.
Analyses in ppm (mg/l).

Location	T7N, R1W, 3, SW $\frac{1}{4}$, SW $\frac{1}{4}$				
Date	10/9/79	10/24/79	11/27/79	12/20/79	2/3/80
Temperature, °C	23.3	26.4	22.3	22.1	22.1
Field pH	7.6	7.5	7.5	7.5	7.6
Total dissolved solids	n.d.	582	n.d.	n.d.	585
SiO ₂	54	52	51	51	46
Na ⁺	125	156	137	132	131
K ⁺	3.4	3.7	3.8	3.8	3.9
Ca ⁺⁺	61	66	64	66	68
Mg ⁺⁺	18	20	19	20	21
Li ⁺	n.d.	n.d.	0.16	0.15	0.15
F ⁻	3.7	2.8	3.8	3.8	3.7
Cl ⁻	67	61	81	72	78
SO ₄ ⁼	166	163	170	171	158
Geothermometry (T°C)					
SiO ₂ (quartz, adiabatic)	106	104	103	103	99
SiO ₂ (quartz, conductive)	106	104	103	103	98
SiO ₂ (chalcedony)	76	74	73	73	67
SiO ₂ (α -cristobalite)	55	53	52	52	47
Na-K-Ca* ($\beta=1/3$)	54	54	53	52	51
Na-K-Ca ($\beta=4/3$)	44	48	48	47	47
Na-K	79	72	82	84	86
n.d. = not determined		*Na-K-Ca (Mg corrected)			

Table VI. Continued.

Location	T7N, RIW, 3, SW $\frac{1}{4}$, SW $\frac{1}{4}$				
Date	4/10/80	5/12/80	7/17/80	8/15/80	9/11/80
Temperature, °C	23.5	21.5	26.1	22	21.8
Field pH	7.4	7.7	8.0	7.7	7.8
Total dissolved solids	n.d.	n.d.	595	598	630
SiO ₂	52	46	51	50	52
Na ⁺	151	135	131	134	137
K ⁺	4.1	4.0	3.8	3.5	3.8
Ca ⁺⁺	73	69	73	69	70
Mg ⁺⁺	22	20	21	22	21
Li ⁺	0.16	0.16	0.15	0.15	0.15
F ⁻	n.d.	3.9	4.0	4.0	3.6
Cl ⁻	n.d.	86	77	79	76
SO ₄ ⁼	n.d.	171	183	184	166
Geothermometry (T°C)					
SiO ₂ (quartz, adiabatic)	104	99	103	103	104
SiO ₂ (quartz, conductive)	104	98	102	102	104
SiO ₂ (chalcedony)	74	68	72	72	74
SiO ₂ (α -cristobalite)	53	48	52	51	53
Na-K-Ca* ($\beta=1/3$)	53	53	56	50	53
Na-K-Ca ($\beta=4/3$)	49	48	45	44	46
Na-K	80	85	83	77	81
n.d. = not determined		*Na-K-Ca (Mg corrected)			

Table VII. Water analyses and geothermometry of waters of Group III discharging from Quaternary alluvium and Quaternary epiclastic breccia in fault contact with Precambrian schist in Castle Hot Spring area. Analyses in ppm (mg/l),

Name	Kent Well	Casa Rosa Spring	Dripping Spring
Location	T7N, R1E, 17, SE $\frac{1}{4}$, NE $\frac{1}{4}$	T7N, R1W, 14, NE $\frac{1}{4}$, SW $\frac{1}{4}$	T7N, R1W, 14, NW $\frac{1}{4}$, NE $\frac{1}{4}$
Date sampled	July 27, 1979	Jan 16, 1979	Jan 16, 1979
Flow Rate, l/min		100	2
Temperature, °C	27	18.9	24.6
Field pH	7.6	7.7	7.3
Total dissolved solids	1800	1644	1600
SiO ₂	56	59	59
Na ⁺	460	540	495
K ⁺	11	14	13
Ca ⁺⁺	62	144	138
Mg ⁺⁺	27	1.3	7.0
Li ⁺	-	1.1	1.1
F ⁻	0.9	4.0	3.8
Cl ⁻	110	525	521
SO ₄ ⁼	990	385	372
Geothermometry			
SiO ₂ (quartz, adiabatic)	107	109	109
SiO ₂ (quartz, conductive)	107	110	110
SiO ₂ (chalcedony)	78	80	80
SiO ₂ (α -cristobalite)	57	59	59
Na-K-Ca ($\beta=1/3$)	69	119	120
Na-K-Ca ($\beta=4/3$)	82	85	83
Na-K	72	76	79

Table VII. Continued.

Name	Mud Spring
Location	T7N, R1W, 12, SW $\frac{1}{4}$, SE $\frac{1}{4}$
Date sampled	June 22, 1980
Flow Rate, l/min	2
Temperature, °C	26.2
Field pH	7.6
Total dissolved solids	1656
SiO ₂	55
Na ⁺	443
K ⁺	12
Ca ⁺⁺	120
Mg ⁺⁺	6.9
Li ⁺	1.0
F ⁻	3.6
Cl ⁻	565
SO ₄ ⁼	381
Geothermometry	
SiO ₂ (quartz, adiabatic)	106
SiO ₂ (quartz, conductive)	106
SiO ₂ (chalcedony)	77
SiO ₂ (α -cristobalite)	56
Na-K-Ca ($\beta=1/3$)	121
Na-K-Ca ($\beta=4/3$)	83
Na-K	81

Mixing Models

Several factors point to mixing of hot geothermal waters with cold near-surface water in addition to conductive cooling for the non-thermal Group III waters. The linear relationship of fluoride and chloride concentrations of waters from Castle Hot Spring, Chuck's Well, Windmill Well, and Castle Creek (Figure 6) could be explained by simple mixing. The difference in ionic concentrations between waters from Chuck's Well and Castle Hot spring suggests 42 percent of thermal component. Water from Windmill Well, which is further away, indicates a 23 percent thermal component. Additional evidence for mixing comes from the SiO_2 concentration, which is undersaturated at the measured temperatures. Likewise the higher Na-K-Ca geothermometers, compared with silica geothermometers could be caused by mixing.

If the main hot spring system originates from a homogenous reservoir of higher temperature water that is mixed with cold ground water, the level of steady state mixing of cold ground water with thermal water can be evaluated by a graphical technique developed by Fournier and Truesdell (1974). In addition to the assumptions previously discussed for geothermometers, the use of mixing models involve four more assumptions:

- (1) Initial silica content is controlled by temperature-dependent reactions between the deep thermal water and chalcedony.
- (2) Additional silica is not dissolved or deposited after mixing.
- (3) Enthalpy is not lost by conductive cooling or steam loss before mixing.
- (4) The temperature and silica content of cold springs are similar to the temperature and silica content of the ground water that mixes with the ascending hot water.

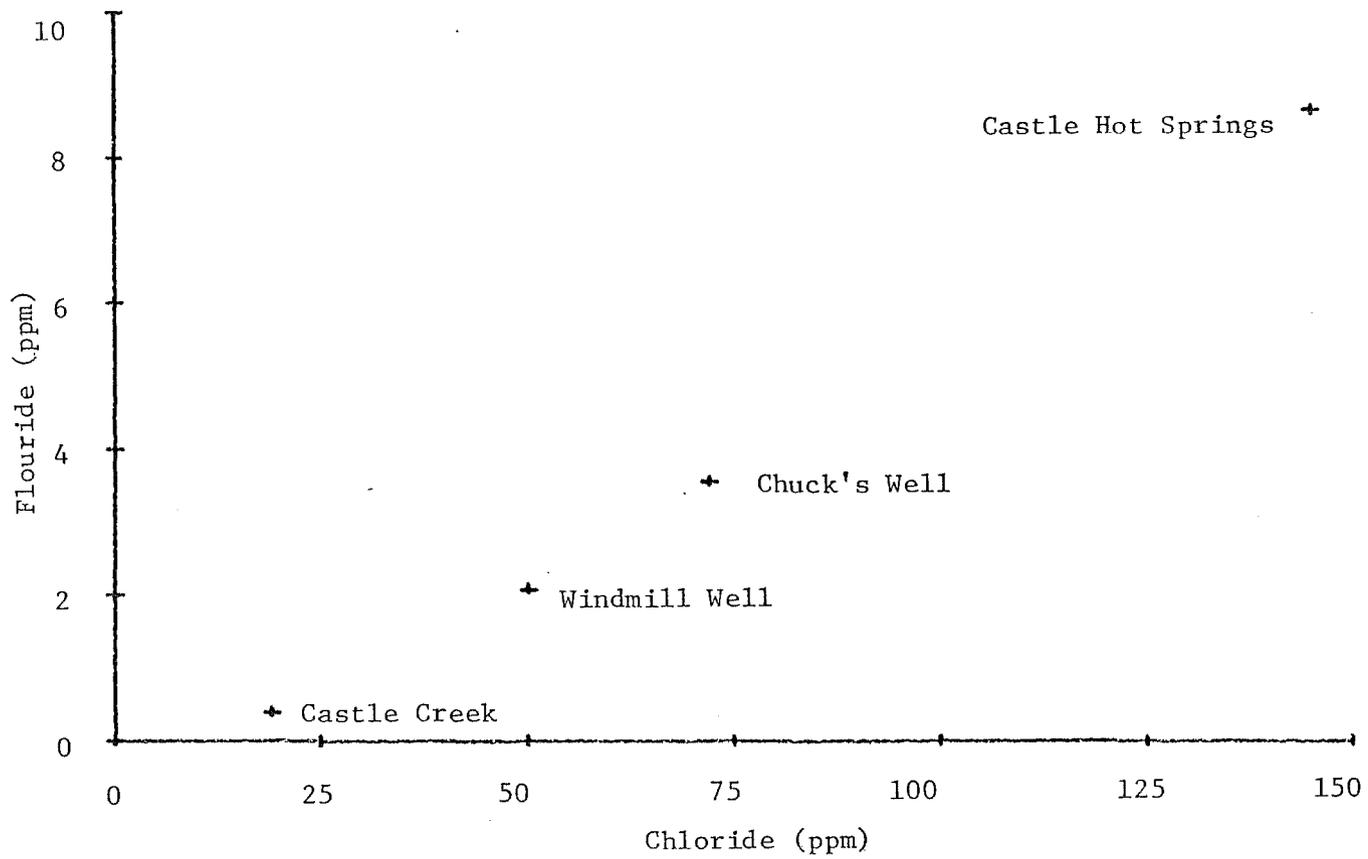


Figure 6. Relationship between chloride and flouride concentrations indicating surficial mixing.

The model is based upon mass balance relationships between enthalpy (heat content) and silica content of the ascending thermal water, cold ground water, and the mixed thermal spring water. Using the measured silica content and temperature of the mixed thermal spring water and cold water, and tabular values of enthalpy and silica at selected temperatures, the temperature of the deep hot water component and the fraction of cold water in the mixed thermal spring may be obtained (Table VIII). The following equations were used:

$$x_t = \frac{E_h - T_{ws}}{E_h - T_{cs}}$$

$$x_{si} = \frac{Si_h - Si_{ws}}{Si_h - Si_{cs}}$$

where:

E_h = enthalpy of hot water at selected temperature (Table IX)

Si_h = silica content of hot water at selected temperatures (Table IX)

T_{ws} = surface temperature at Castle Hot Spring = 51°C

T_{cs} = surface temperature of cold spring = 21°C

Si_{ws} = silica content of Castle Hot Spring = 61 ppm

Si_{cs} = silica content of cold spring = 35 ppm

Values of x_t and x_{si} are plotted versus temperature in Figure 7. The intersection of the two curves gives the estimated temperature of the hot water component and the fraction of cold water. The chalcedony mixing model yields a reservoir temperature of 100°C and a cold water fraction of 62 percent.

THERMAL PROFILES OF EXISTING WELLS

Measurements of geothermal gradients were attempted at several

Table VIII. Calculated values of X_t and X_{si} for chalcedony mixing model solution, Castle^t Hot Spring.

Hot Water Temperature (°C)	X_t	X_{si}
70	.39	-.08
80	.49	.29
90	.57	.49
100	.62	.62
110	.66	.70
120	.70	.76
130	.73	.80

Table IX. Enthalpies of liquid water and chalcedony solubilities at selected temperatures,

Temperature (°C)	Enthalpy (cal/g)	Chalcedony ² (mg/l)
70	70	59.0
80	80	71.7
90	90	86.4
100	100.1	102.9
110	110.2	121.9
120	120.3	142.4
130	130.4	165.4

¹Keenan and others (1969).

²Values generated from:

$$\text{SiO}_2(\text{chalcedony}) = \frac{10^{[-1032/(t + 273.15)]}}{1.665 \times 10^{-5}} - 0.09$$

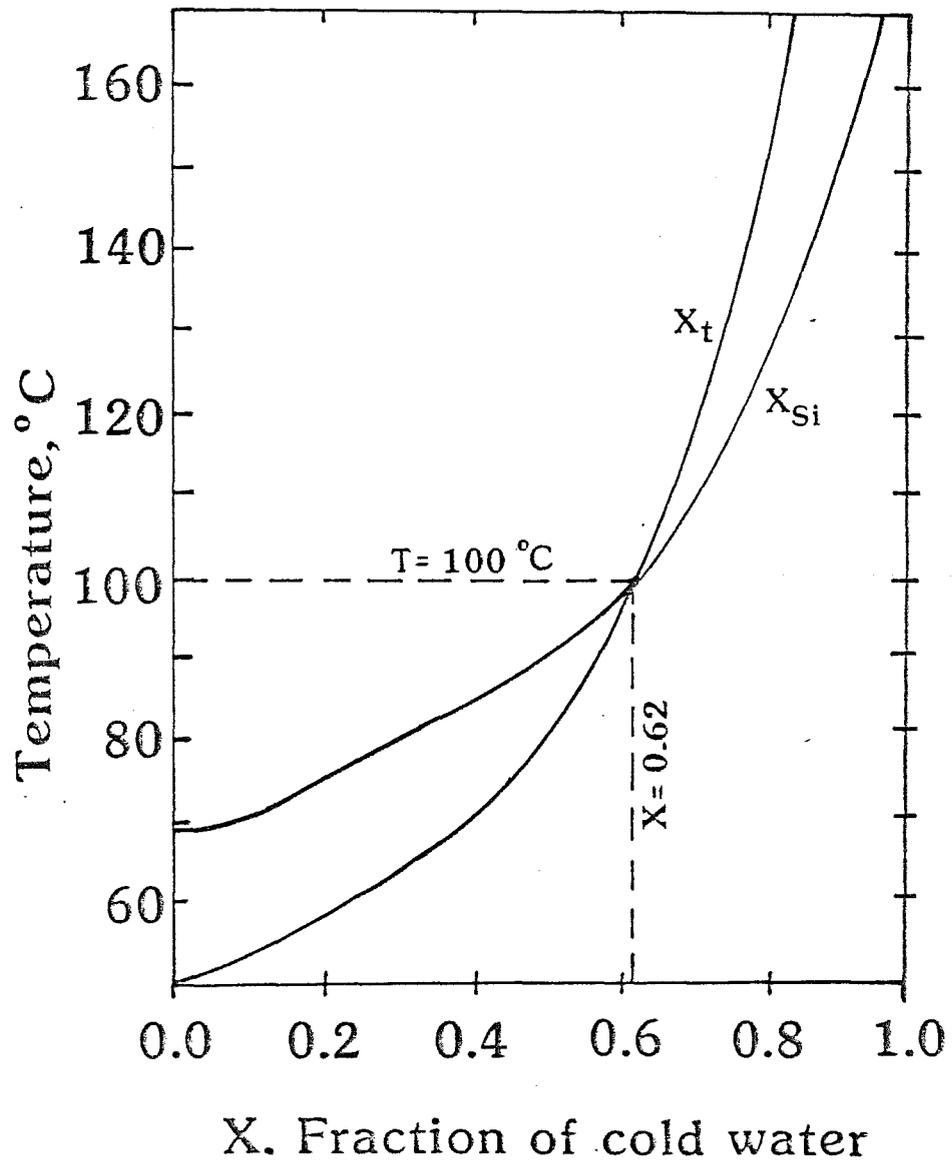


Figure 7. Plot of mixing model data for Castle Hot Spring.

existing wells in the vicinity of Castle Hot Spring. However, the following difficulties severely limited the success of this program:

- 1) Many logged wells were isothermal.
- 2) Poorly marked exploration holes usually could not be found. Those which were found were plugged due to collapse.
- 3) At some wells instrument failure was encountered.

The only successful measurements were compiled for the Dodd Well (Figure 8) where a temperature of 38° was encountered at a depth of 90 m. The very high thermal gradient (approximately 100°C/Km) is nonlinear and indicates an area where warm water is rising toward the surface. The non-equilibrium conditions suspected for this well preclude the determination of the heat flow, although it would appear to be well above normal.

HEAT SOURCE EVALUATION

The three possible mechanisms that were considered as a heat source for the Castle Hot Spring system include recent volcanism, near surface high radioactivity, and above normal geothermal gradient. Although the area was a center for volcanic activity about 20 m.y. to 15 m.y. B.P., residual thermal anomalies from volcanic activity should have dissipated long ago. Without new evidence for young igneous activity, this possible heat source will be discounted.

Analysis of natural radioactivity of the rocks at Castle Hot Springs (Table X) suggests that some units, especially the Chocolate Drop latite, have abnormally high levels of radioactivity. Field measurements gave average radiometric readings of 34 $\mu\text{R/S}$ for the Chocolate Drop latite, compared with a value of 17 $\mu\text{R/S}$ for the Precambrian granite. Results from most samples sent to the University of Texas, El Paso, for KUT analysis are still outstanding, so a quantitative interpretation is not

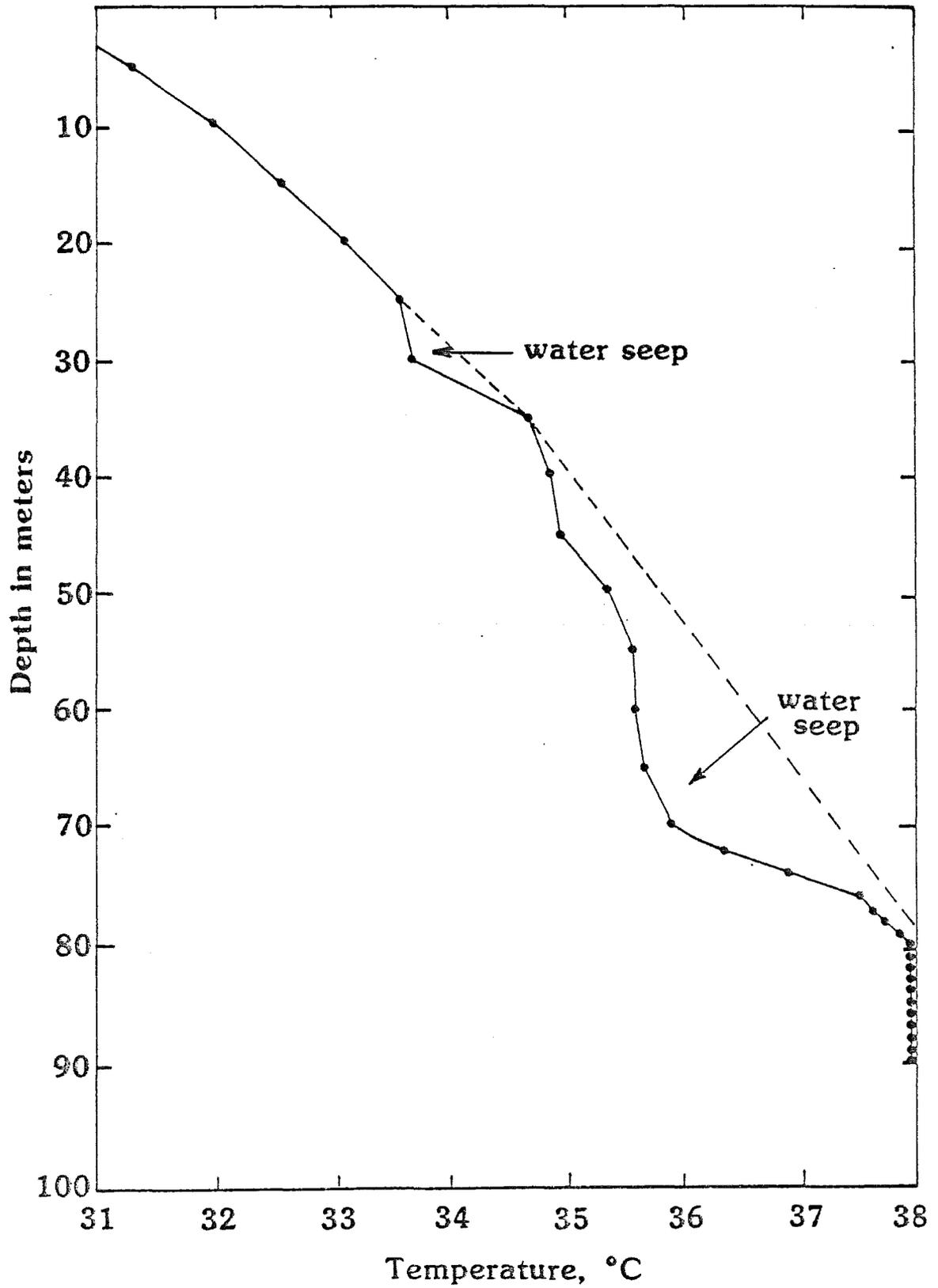


Figure 8. Measured thermal profile at the Dodd Well. The dashed line follows the assumed natural gradient. Irregularities are due to leakage of cold water into the well.

Table X. Radioactive elements in various rock types.

<u>Rock Unit</u>	<u>UR/s</u> ¹	<u>K₂O</u> ²	<u>K</u> ³	<u>U</u>	<u>Th</u>
Precambrian schist	10 (1.0)	n.d.			
Precambrian granite	17 (1.1)	3.9 (0.74)	2.2 (0.77)	1.4 (0.21)	1.5 (0.27)
Basaltic lava	8 (2.0)	1.3 (0.08)			
Rhyolite tuffs & lava	17 (1.1)	6.9 (0.90)			
Volcanic breccia	18 (2.1)	n.d.			
Epiclastic breccia	10 (2.0)	n.d.			
Latite of Chocolate Drop	34 (6.1)	7.1 (1.00)	4.5 (0.13)	19.1 (4.09)	24.7 (3.89)
Latite of Hells Gate	16 (1.3)	3.8 (0.36)	2.3	2.0	7.2
Vitrophyre of Hells Gate	8 (0.5)	2.9			

1) Field radiation measured using a Ludlum model 125 scintillation counter. 2) K₂O values are averages of whole-rock analyses performed at Arizona State University by XRD. 3) K-U-T measurements were performed at the University of Texas at El Paso using gamma-ray spectrometry. n.d. = not determined. Numbers in parentheses are standard deviations of multiple measurements. Some data on K-U-T measurements are still outstanding.

possible at present. The data will be included in the thesis by Satkin (in preparation, 1980).

Because thermal probing within the study area was limited to only the Dodd Well and regional profiles are at too great a distance to be of direct use, a definite statement regarding thermal gradient and heat flow within the study area is precluded. However, the region does exhibit high heat flow so that local areas of high thermal gradients are reasonable. The limited available data rule out a volcanic heat source but make radiogenic heating or deep circulation with heating by the normal geothermal gradient both seem reasonable possibilities.

CONCLUSIONS

The geologic setting of Castle Hot Spring consists of a Precambrian basement of granite and schist that is overlain by a maximum 420-670 m of volcanic rocks of an approximately Miocene age. Toward the end of the volcanic cycle large-scale northwest-trending normal faults formed a series of horsts and grabens. Large listric faults allowed sheets of Precambrian granite to come to rest on top of volcanic rocks along the Castle fault system. This structural setting controls the location of the major high-temperature springs and wells (Group I) in this system.

The thermal system is a steady-state low temperature hydrothermal resource. Chemical geothermometry and mixing models indicate a maximum reservoir temperature of 100°C. The size of the resource is much more extensive than previously recognized. This regional geothermal system may extend along the entire southeastern flank of the Bradshaw Mountains. The heat source is not well specified, but may be due to a shallow source related to radioactive heating or to an abnormally high geothermal gradient. The present information favors heating by the geothermal gradient due to deep circulation.

The hot water rises along the basin-and-range faults and mixes with cold near-surface water. The resultant warm ($\sim 50^{\circ}\text{C}$) water emerges along the intersection of listric faults with the larger basin-and-range system.

Recommendations

Because this low-temperature geothermal resource is of possible regional importance and because it lies in a zone of probable suburban development, the actual extent of the system should be determined. Monitoring for geochemical and geothermal evaluation should continue at the present thermal systems. In addition, new thermal wells and springs should be added to the inventory until the limits of the system are defined. Expansion of the net as in the vicinity of the Kent Well (TIN, RIE, sec 17) should be of primary concern.

The hydrology of the geothermal system also needs further work. The residence time of water in the system should be determined. A single sample of Castle Hot Spring water is being processed at the University of Arizona for radio carbon systematics to be used in age modeling. However, additional data should be collected and other analytical techniques (for example, freon analysis) should be used. The nature of the hot and cold water mixing should be tested with stable isotopes of oxygen and hydrogen in the waters and the hydrothermal alteration minerals. These techniques applied on a regional scale could delineate areas of possible future domestic applications of this large low-temperature geothermal resource.

Because of the questions that still remain regarding the hydrology of the system (size, depth, source of water, source of heat, nature of the aquifer, etc.) we recommend that no drilling be done in the vicinity of existing springs without additional geochemical data to answer some of these important questions. This is especially true for Castle Hot

Spring where such drilling could upset the delicate balance of flow and would probably be of little value for production at this time. On the other hand, drilling, sampling and testing of wells on a regional basis could be of great use in defining the regional extent of this system and might lead to actual production on a reasonable time scale.

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APPENDIX I

THE GEOLOGIC SETTING OF CASTLE HOT SPRINGS, ARIZONA¹

¹In Geothermal Resources Council, Transactions,
v. 3, 1979.

Geothermal Resources Council, TRANSACTIONS, Vol. 3 September 1972

THE GEOLOGIC SETTING OF CASTLE HOT SPRINGS, ARIZONA

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ABSTRACT

Castle Hot Springs, a resort area established in 1891, is presently owned by the Arizona State University Foundation and used as a University conference center. It is located 70 km northwest of Phoenix in an area of several other warm springs. The spring issues an average flow of 1,300 litres per minute with a surface temperature of 46 C. Water geochemistry indicates a subsurface temperature in excess of 100 C. The resource is being evaluated as an energy source for space heating and cooling of the conference center.

INTRODUCTION

Castle Hot Springs is located in a mountain canyon 70 km northwest of central Phoenix (fig. 1). A privately-owned resort has existed on the site since 1891. In 1977 the property was acquired by the Arizona State University Foundation. Since then the facilities have been used as a university affiliated conference center with a capacity of about 50 American plan participants. The spring provides hot water for direct tap use. The hot spring water is also cooled and used for drinking water.

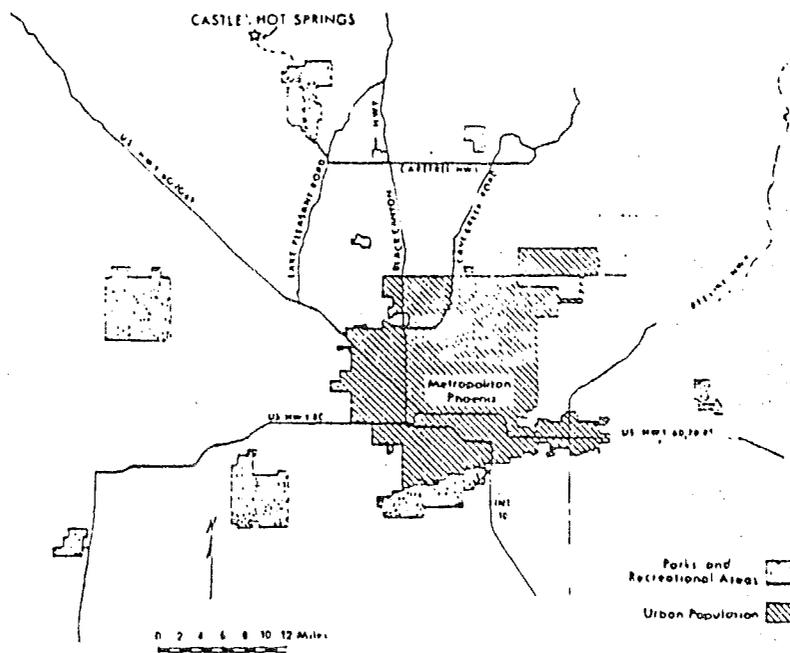


Fig. 1 Location map of Castle Hot Springs.

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A plan for direct-use geothermal development of the resource for space heating and coolings has been developed in cooperation with the ASU Foundation. The first phase of this plan, determination of the detailed geologic framework of the thermal system, is currently in progress. This report provides some information on the general geologic setting of the springs.

DESCRIPTION

Castle Hot Springs is situated at the northern boundary of a Tertiary volcano-tectonic graben that separates Precambrian schists to the southwest from granitic rocks of the Bradshaw Mountains to the northeast. The volcanic rocks were studied by Ward (1977), who provided a geologic map at a scale of 1:48,000, a stratigraphic section, whole-rock chemical analyses, and a structural and tectonic setting. The generalized stratigraphy of the area consists of Precambrian schists and granitic rocks overlain unconformably by 300-450 m of basalt lavas and rhyolitic tuffs which are in turn overlain by approximately 120 m of epiclastic breccia and quartz latitic lavas. A number of other warm springs and warm water wells are located in the hydrologic basin of Castle Creek.

The Castle Hot Springs volcanic field lies predominantly within a NW to NE trending graben which is related to volcano-tectonic subsidence. Normal faulting and tilting characterize structures within the graben (Fig. 2). Castle Hot Springs lies along the trend of the NE graben master fault where displacement is approximately 300 m. Although movement along this and related faults is primarily to Miocene volcanic subsidence, more recent fault movement may be related to local adjustment along Basin and Range faults. Details of the faultline are complex so that current mapping at a scale of 1:1,000 is necessary to locate surface expression of faults that can be projected to depth.

Wide-spread hydrothermal alteration of volcanic rocks in the area is related to hot springs activity. Fountain-like deposits of opaline silica and chalcedony and a silica sinter dike that trends along the northward strike of a fault indicate the position of ancient hot spring activity. A broad area of clay and sericite alteration of surface materials 5 to 10 km north of the hot spring has been extensively explored for base and precious metal ore deposits at depth.

Measured surface temperatures at Castle Hot Springs range between 46° and 50° C with a flow rate of 1300±130 l/min (340 gal/min). The water is unusually pure (Table 1) with only 700-800 ppm total dissolved solids. The Na-K-Ca and silica geothermometers reported in U.S.G.S. Circular 726, U.S.G.S. Open File Report 77-654 as well as geochemical data determined by our laboratory suggest a minimum subsurface temperature of 90° to 110° C. If the model of deep circulation of meteoric water shown in Figure 1 is accurate, deep temperatures could be much higher due to mixing of the deep hot fluid with shallow cooler water.

The heat source for the present hot spring activity is unknown. Although most hot water localities in the Basin and Range Province of Arizona are related to deep, alluvium-filled basins, Castle Hot Springs is in an area underlain by crystalline rocks. Water recharge is from the Precambrian crystalline basement of the Bradshaw Mountains. Drainage is provided in the area by Garfias Wash and Castle Creek, both of which flow intermittently into the Agua Fria River. Location of the springs appears to be controlled by the graben master fault zone which also intersects a series of Precambrian pegmatite bodies that have been mined for lithium and uranium. These pegmatites lie in the recharge area of the springs and may heat groundwater through radioactive decay. Other heat sources such as oxidation of sulfides or the presence of a buried pluton are possible.

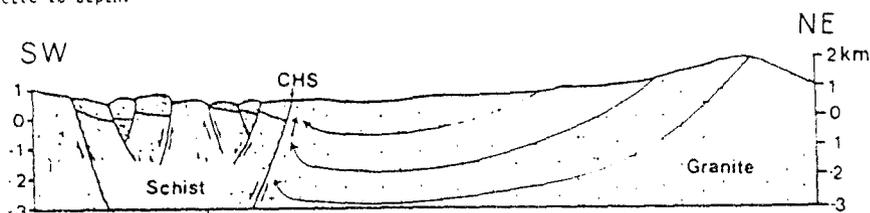


Fig. 2. True scale cross section across the volcano-tectonic graben. CHS shows the location of Castle Hot Springs. Fault displacements shown by short arrows. Groundwater flow lines shown by long arrows.

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TABLE 1: GEOCHEMISTRY AND
GEOTHERMOMETRY OF CASTLE HOT SPRINGS

Dissolved Species	Analysis* (ppm)		
	(1)	(2)	(3)
T.D.S.	801	706	820
CO ₂		1	
HCO ₃		163	
Cl		123	
SO ₄		156	
NO ₃		1.2	
F		6.0	
Mg			2.4
Ca	55		33
Na	191		200
K	8		4.7
SiO ₂	73		58
Li	0.7		0.29
B	0.9c		
T C (surface)	48	48	46
Flow (l/min)		1,300	1,200
pH		8.08	7.45
T C (subsurface)			
SiO ₂	119		109
Na-K-Ca ($\beta=1/3$)			113
Na-K-Ca ($\beta=4/3$)	77	86	71
Na-K	108	101	

* (1) Arizona State University (1977)
 (2) Engineers Testing Laboratories (1975)
 (3) U.S.C.S. Open File Report 77-654 (1977)

CONCLUSIONS

A steady flow of a large quantity of hot water at Castle Hot Springs makes it an attractive geothermal resource for development. On-going work including detailed geologic mapping, thermal logging of existing wells, and water and rock geochemistry; the site for a test well aimed at evaluation of the resource should be determined by the fall of 1980. Deep circulation of meteoric waters through rock with a high geothermal gradient is the strongest hypothesis at this time for explanation of the hot water.

ACKNOWLEDGEMENTS

The Arizona State University Foundation has cooperated with this work and provided facilities at the site for our field parties. Financial support for this work is provided by D.O.E. through the Arizona Bureau of Geology and Mineral Technology Agreement No. De-F107-791D 12009.

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APPENDIX II

WATER GEOCHEMISTRY AT CASTLE HOT SPRINGS, ARIZONA¹

¹In Geothermal Resources Council, Transactions,
v. 4, 1980.

WATER GEOCHEMISTRY AT CASTLE HOT SPRINGS, ARIZONA

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ABSTRACT

A geochemical survey of springs and wells in the Castle Hot Springs area, Arizona, shows that three groups of waters can be distinguished by salinity and chemistry. The thermal waters of Group I range from 640 to 820 ppm TDS, and the waters contain high concentrations of SiO_2 , Li^+ , and F^- . The non-thermal waters of Group II range from 380 to 580 ppm TDS and contain low concentrations of SiO_2 , Li^+ , and F^- . The non-thermal waters of Group III range from 1600-1650 ppm TDS and contain the highest concentrations of Li^+ , Cl^- , and SO_4^{2-} .

The discrepancy between the low measured surface temperature at Castle Hot Springs, and the high temperatures estimated from chemical geothermometry suggest thermal waters may have cooled either by conduction, boiling or mixing. The chalcidony mixing model yields a reservoir temperature of 95°C and a cold water fraction of 56%.

INTRODUCTION

Castle Hot Springs is located 70 km northwest of Phoenix, Arizona (Figure 1). It is presently being evaluated for direct-use geothermal development of the resource for space heating and cooling. The results of hydrogeochemical sampling of thermal and non-thermal springs and wells in the area are presented.

GEOLOGIC SETTING

The geologic setting of Castle Hot Springs has been discussed by Sheridan et al. (1979). Recent detailed geologic mapping has documented a low-angle slump fault displacing an allochthonous block of Precambrian granite on top of a sequence of Tertiary volcanic rocks. The allochthonous block has been altered and is strongly brecciated and jointed resulting in increased permeability.

Mixing of hydrothermal fluids and cold meteoric water may be significant along this low-angle fault. Nielson and Moore (1979) have described a similar geologic situation at the Cove Fort-Sulphurdale geothermal system in Utah. They suggest that the allochthonous rocks may serve as a thermal cap on the system separating a convective ther-

mal regime beneath the low-angle fault from a zone of conductive heat transport and probable fresh water influx above the principal fault zone.

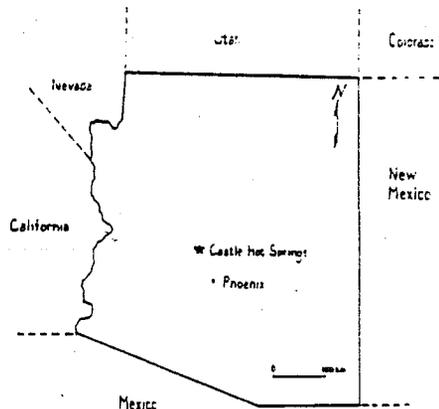


Figure 1. Location map of Castle Hot Springs, Arizona.

HYDROLOGIC SETTING

The thermal waters of Group I display similar physical and chemical characteristics. The waters of Group I are: Castle Hot Springs, Alkali Spring, Henderson Ranch Spring and the Dode Well. The thermal waters occur along a 0.8 km alignment trending N.45°W, which coincides with the trend of a major fault system bounding Precambrian crystalline rocks and Tertiary volcanic rocks. The thermal springs all emanate at an elevation of 658 meters along the same fault system suggesting an apparent hydrostatic relationship. The thermal waters display a homogenous chemistry which indicates they probably originate from the same geothermal reservoir.

WATER SAMPLING AND ANALYTICAL PROCEDURE

An important aspect of this investigation is

Table I. Chemical analyses and calculated reservoir temperatures of springs and wells in the Castle Hot Springs, area, Arizona. Analyses are in ppm (mg/l) unless otherwise noted.

Name, Group (I, II, III) Location	Castle Hot Springs, I T8N, R1W, 34, SW $\frac{1}{4}$, SW $\frac{1}{4}$	Henderson Ranch Spring, I T8N, R1W, 33, NW $\frac{1}{4}$	Alkalai Spring, I T8N, R1W, 33, NW $\frac{1}{4}$, SE $\frac{1}{4}$	Mesquite Drip, I T7N, R1W, 33, NW $\frac{1}{4}$, SE $\frac{1}{4}$	Dodd Well, I T8N, R1W, 33, NW $\frac{1}{4}$, NW $\frac{1}{4}$	
Temperature °C	54.7	29.2	31.2	26.8	23.6	
pH (field)	7.85	7.70	7.85	7.90	8.00	
SiO ₂	61.27	60.42	70.78	71.39	62.69	
Na ⁺	208.03	234.47	214.67	253.88	239.26	
K ⁺	5.42	7.31	6.32	7.38	7.29	
Ca ⁺⁺	32.42	39.72	15.78	17.40	25.68	
Mg ⁺⁺	2.32	2.23	0.23	0.48	0.43	
Li ⁺	0.34	0.55	0.42	0.54	0.49	
F ⁻	8.45	7.45	11.88	12.54	8.19	
Cl ⁻	145	150	135	150	142	
SO ₄ ⁻	211	299	209	228	288	
Geothermometry °C						
SiO ₂ (quartz, adiabatic)	110.93	110.34	117.10	117.47	111.89	
SiO ₂ (quartz, conductive)	111.54	110.85	118.76	119.19	112.67	
SiO ₂ (chalcedony)	82.38	81.63	90.22	90.70	83.60	
Na-K-Ca ($\beta=1/3$)	117.08	124.68	127.50	128.64	127.05	
Na-K-Ca ($\beta=4/3$)	75.94	82.61	97.95	103.32	92.73	
Na-K	77.30	88.86	85.09	84.26	87.34	
Name, Group (I, II, III) Location	Chuck's Well, II T7N, R1W, 3, SW $\frac{1}{4}$, SW $\frac{1}{4}$	Menudo Spring, II T7N, R1W, 14, NW $\frac{1}{4}$, NW $\frac{1}{4}$	Layton Seep, II T7N, R2W, 1, NW $\frac{1}{4}$, SW $\frac{1}{4}$	Windmill Well, II T7N, R1W, 3, SW $\frac{1}{4}$, SW $\frac{1}{4}$	Casa Rosa Spring III T7N, R1W, 14, NE $\frac{1}{4}$, SW $\frac{1}{4}$	Dripping Spring III T7N, R1W, 14, NW $\frac{1}{4}$, NE $\frac{1}{4}$
Temperature °C	22.3	21.8	20.6	20.5	18.9	24.6
pH (field)	7.45	7.55	8.00	7.55	7.70	7.25
SiO ₂	51.03	75.55	39.11	42.22	36.82	30.50
Na ⁺	136.89	25.37	15.30	93.54	539.54	494.52
K ⁺	3.83	1.70	1.62	3.45	13.86	13.28
Ca ⁺⁺	64.47	82.82	88.78	70.15	144.04	137.90
Mg ⁺⁺	19.29	16.37	14.08	22.26	1.27	7.01
Li ⁺	0.16	0.04	0.06	0.11	1.14	1.05
F ⁻	3.83	0.45	0.30	2.11	4.0	3.8
Cl ⁻	81	39.8	11.3	50.2	525	521
SO ₄ ⁻	170	23.7	8.6	122	385	372
Geothermometry °C						
SiO ₂ (quartz, adiabatic)	101.33	119.95	92.92	95.87	90.62	83.65
SiO ₂ (quartz, conductive)	102.76	122.11	90.69	94.08	88.06	80.09
SiO ₂ (chalcedony)	72.91	91.88	60.09	63.61	57.29	48.78
Na-K-Ca ($\beta=1/3$)	109.88	120.60	132.28	115.13	119.30	120.22
Na-K-Ca ($\beta=4/3$)	48.01	10.93	5.44	40.37	84.58	83.15
Na-K	81.80	148.08	193.76	100.40	76.42	79.20

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to test the variation of water chemistry with time. Temperature measurements and water samples collected at each sampling site were taken as close to the source as possible and at the same location throughout the sampling period. Temperatures were measured with an Extech 1200 digital thermometer. The pH was determined in the field on an unfiltered sample with a Photovolt pH meter 126A.

Water samples collected for chemical analysis were analyzed for SiO_2 , Na^+ , K^+ , Ca^{++} , Mg^{++} , and Li^+ on a Varian 1250 Atomic Absorption Spectrophotometer. F^- , Cl^- , and SO_4^{--} were analyzed on a Dionex 10 Ion Chromatograph. Total dissolved solids were determined on filtered untreated samples by the residue-on-evaporation method (Rainwater and Thatcher, 1960).

GEOCHEMISTRY OF THERMAL AND NON-THERMAL WATERS

The thermal waters (Group I) are a sodium-chloride-sulfate type. The waters have relatively high concentrations of SiO_2 , Li^+ , and F^- and

low Mg^{++} (Table I). In contrast, the non-thermal waters (Group II) are enriched in Ca^{++} , and Mg^{++} and have lower concentrations of SiO_2 , Li^+ , and F^- .

Within the non-thermal group of waters a subgroup of waters (Group III) can be distinguished by their high salinity. Both Cass Kosa and Brinn Springs are highly enriched in Na^+ , Ca^{++} , Li^+ , Cl^- , and SO_4^{--} . It is possible that these waters follow a different hydrologic flow pattern. They may derive their high salinity from dissolution of limestones and evaporites that crop out 20 km to the west. A heavy isotopic signature may confirm this suggestion.

The measured surface temperature at Castle Hot Springs ranges between 47.6°C and 55.4°C with a flow rate of 1300 l/min (340 gal/min). The springs were sampled periodically (3-4 week intervals) to test the variation of chemistry with time. It is evident from the chemical analyses listed in Table II that there has been no significant change in the main spring system's chemistry.

Table 2. Chemical variation through time at Castle Hot Springs, Arizona. Analyses in ppm (mg/l).

Date	10/9/79	10/24/79	11/27/79	12/20/79	1/9/80	2/3/80	3/7/80	4/10/80	5/12/80
Temp. °C	51.1	55.4	54.7	52.7	53.4	52.1	49.3	47.6	47.7
pH	7.60	7.65	7.85	7.75	7.70	7.85	7.80	7.85	7.85
SiO_2	59.70	63.48	61.27	60.27	58.68	59.37	61.79	62.01	62.25
Na^+	209.08	208.75	208.03	210.89	195.27	199.64	202.12	221.56	202.93
K^+	4.98	5.49	5.42	5.50	5.55	5.35	5.61	5.39	5.54
Ca^{++}	30.33	34.04	32.42	31.89	29.46	29.78	29.52	31.07	31.46
Mg^{++}	2.36	2.99	2.32	2.63	2.37	2.32	2.41	2.43	2.50
Li^+	n.d.	n.d.	0.34	0.33	0.32	0.31	0.31	0.30	0.32
F^-	8.50	9.16	8.45	8.70	8.53	8.61	8.47	8.31	8.64
Cl^-	147	155	145	141	140	141	138	140	145
SO_4^{--}	212	230	211	211	206	200	196	189	206

Note: n.d. - not determined

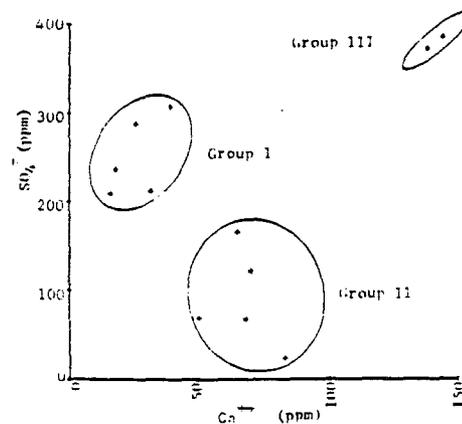


Figure 2. Water chemistry Ca^{++} versus SO_4^{--} .

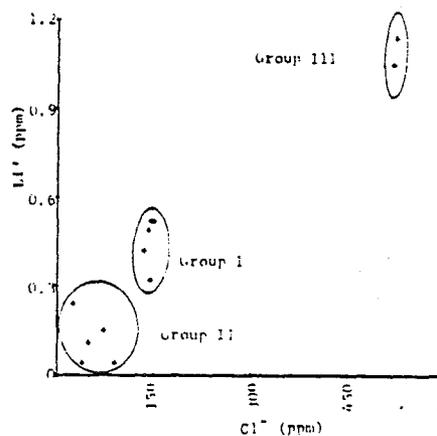


Figure 3. Water chemistry Cl^- versus Li^+ .

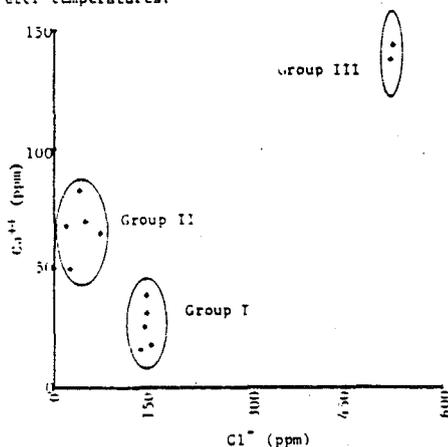
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GEOOTHERMOMETRY

The temperature of the geothermal reservoir at Castle Hot Springs has been estimated by the silica geothermometer (Fournier and Rowe, 1966), the Na-K, and Na-K-Ca geothermometers (Fournier and Truesdell, 1973, Table I). The calculated solubility of chalcedony closely approximates the silica content at Castle Hot Springs. Thus the chalcedony geothermometer yields the most reliable estimate of water temperature at depth.

Large travertine deposits occur near Castle Hot Springs. The deposition of calcium-carbonate will decrease the calcium ion concentration and should yield artificially high temperature estimates. However, the Na-K-Ca geothermometer estimate closely resembles both the Na-K and SiO_2 geothermometer estimates. Travertine may not be deposited during the rapid ascent of the fluid, just at the surface as the dissolved CO_2 bubbles off at atmospheric pressure and lowered temperature.

The chalcedony geothermometer gives an estimated subsurface temperature of 82°C which is above the surface temperature at Castle Hot Springs (51°C). This low surface water temperature may possibly be due to heat loss through conduction, boiling, or mixing. Because of the large flow rate at Castle Hot Springs, heat loss through conduction may be negligible. Cooling the ascending thermal water by mixing with cool groundwater is more probable because numerous intersecting faults may provide passageways. The graphical mixing model solution (Fournier and Truesdell, 1974) using chalcedony as the dissolved silica phase in equilibrium with the hot springs' water yields a subsurface temperature of 95°C and a cold water fraction of 56%. This temperature is similar to the calculated geothermometer temperatures.

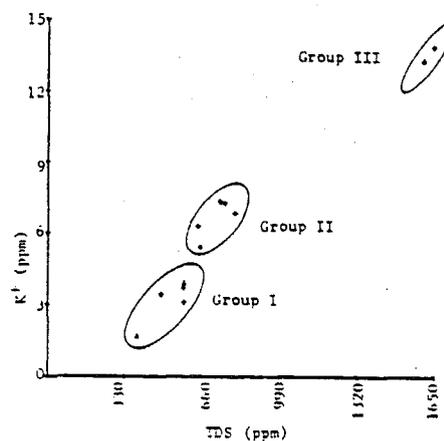
Figure 4. Water chemistry Cl^- versus Ca^{2+} .

ACKNOWLEDGMENTS

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Figure 5. Total dissolved solids versus K^+ .