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A REEVALUATION OF GEOTHERMAL POTENTIAL OF THE WILBUR HOT SPRINGS AREA, CALIFORNIA

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ABSTRACT

In a recent assessment of the geothermal potential of the Wilbur Mining District, it was estimated that a thermal brine at about 150°C is present at depths less than 3 km. The Na-K-Ca geothermometer applied to the four major spring groups in this area gives temperatures ranging from 220 to 240°C. The magnesium corrected geothermometer gives inconsistent temperatures suggesting that Mg enters the water during its passage from the reservoir to the surface. For this reason the Mg correction is not considered appropriate and the fluids are estimated to have originated at 230°C. From the CH_4, H_2S, and CO_2 concentrations in the spring gases at Wilbur Hot Springs and from equations devised primarily for use in steam wells, reservoir temperatures from 227° to 242°C are calculated.

INTRODUCTION

White (1957) suggested that connate water underlies the Wilbur Mining District and that this water is affected by low-grade metamorphism of deep rocks. This could give rise to the peculiar thermal water composition of the four active springs in the Wilbur Mining District: Wilbur Hot Spring, Jones' Fountain of Life, Blanck's Spring, and the Elgin Mine springs. All but the Elgin Mine springs issue from the topographic low (see Fig. 1) of Sulphur Creek. J. M. Donnelly (oral communication, 1979) has mapped a dike of 1.6-m.y. andesitic basalt in the vicinity of Wilbur Hot Springs. This andesitic basalt is undoubtedly too old to be the present day heat source for the springs in the Wilbur Mining District, but it may indicate that magma or hot rock is still present under the Wilbur Mining District.

GEOTHERMOMETRY COMPARISONS

In 1968 a geothermal well, Wilbur #1, was drilled to a depth of 1300 m approximately 1 km southwest of Wilbur Hot Springs, outside of the major thermal activity. White and others (1973) reported that this well erupted water at 140°C. A chemical analysis by Sunoco Energy Development Company of the well water (Table 1) reported ~14,400 mg/L Cl, 1500 mg/L higher than that reported in White and others (1973). Compared

Figure I Generalized map of the Wilbur Mining District, Calif.

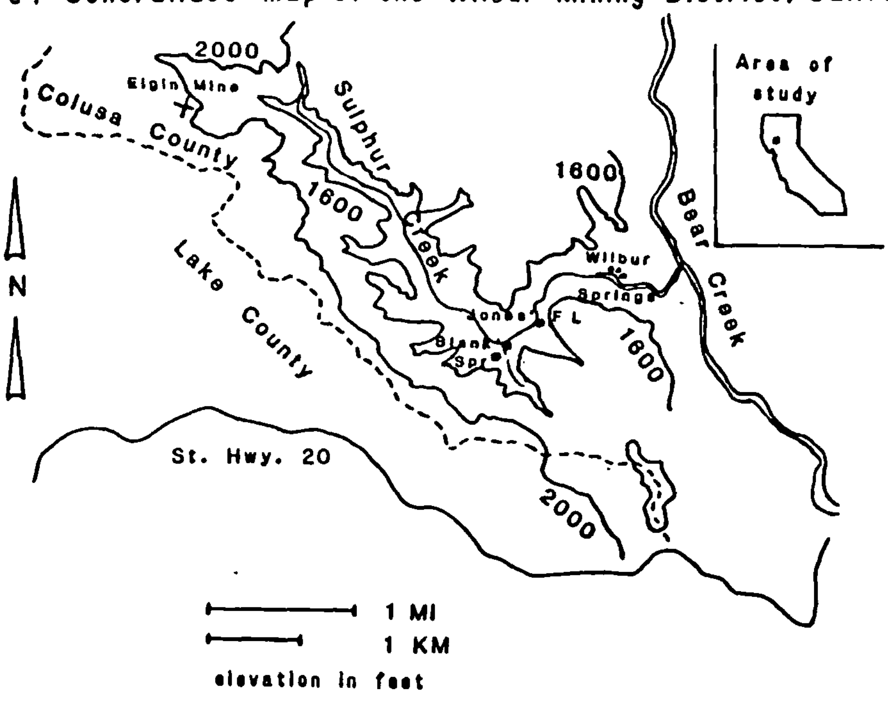

Table 1. Averaged concentrations for spring and well water in the Wilbur mining district[1]

	Wilbur Hot Spring	Jones' Fountain of Life	Blanck's Spring	Elgin Mine Springs	Wilbur #1 Geothermal well	Meteoric water
T °C	52	60	42	61	140	19.5
pH	7.5	7.7	7.8		8.8	6.8
SiO_2	176	89	124	198	133	21
Al	1.8	---	---	---	---	---
Fe	0.17	0.35	0.19	0.17	---	---
Mn	0.04	0.05	0.05	1.0	---	---
Ca	2.5	2.6	3.5	4.8	1	3.4
Mg	45	31	69	28	2	182
Sr	3.6	1.6	1.4	3.7	---	---
Ba	3.1	3.0	1.5	---	---	---
Na	8700	9880	7220	9330	10,000	132
K	408	432	360	540	440	4.1
Li	11.6	10.7	6.9	11	---	0.16
NH_4	294	120	125	243	275	---
HCO_3	6900	5740	6390	7270	5170	1020
CO_3	---	---	---	---	1170	0
SO_4	356(4)	71	180	86	263	185
Cl	9980	11,700	8050	11,550	14,400	83
F	2.4	3.5	2.3	3.2	16	.37
Br	19	34	24	30	---	---
I	12	23	18	25	---	---
B	233	240	150	240		
H_2S	165	---	---	170	148	---
Na-K-Ca	236	232	232	244	244	---
Mg corr Na-K-Ca	84	118	46	148	243	---
SiO_2 Adiabatic	218	---	198	208	---	---
Conductive	---	223	---	---	201	---

[1]Analyses in mg/L

to Wilbur Hot Springs (see Table 1) Wilbur #1 geothermal well contains (1) a higher chloride content (14,400 vs. 10,000 mg/L), (2) a lower sulfate content (260 vs. 360 mg/L) and (3) a much lower magnesium content (2 vs. 45 mg/L).

The magnesium corrected Na-K-Ca geothermometer (Fournier and Potter, 1978) indicates temperatures ranging from 40 to 160°C. However, because the country rock around Wilbur Hot Springs is principally serpentinite, the high magnesium in Wilbur Hot Spring is probably due to serpentine dissolution. The uncorrected Na-K-Ca (Fournier and Truesdell, 1973) temperatures, which range from 220 to 248°C, may be more reasonable. Water from Wilbur # 1 geothermal well has a Na-K-Ca temperature of 244°C (Table 1) and very little magnesium; a correction of only 1°C is calculated.

Due to possible silica addition from serpentine dissolution, severe difficulties are encountered when using dissolved silica concentration in estimating thermal reservoir temperatures. The difficulties include the following: (1) the silica may have already polymerized or precipitated so that direct application of the silica geothermometer (Fournier and Rowe, 1966) will indicate a low reservoir temperature; (2) the thermal water is probably mixed with dilute meteoric water giving rise to the observed spring water compositions and temperatures; and (3) the diluting water or the warm mixed water may contain some silica originating from low-temperature serpentine dissolution. For comparison, the conductive and adiabatic (with assumed subsurface steam loss at 100°C) silica-mixing-model temperatures (Truesdell and Fournier, 1977) of the warm springs are shown in Table 1. Despite all of the possible problems using silica concentrations in springs from this area, the adiabatic mixed-water temperatures are in moderate to good agreement with the Na-K-Ca temperatures. Fournier (1979) indicated that silica reequilibration is more likely to occur than Na-K-Ca reequilibration. The silica concentration in a water sample from Wilbur #1 geothermal well is below that expected in a 220° to 240°C water; however, it may be in approximate equilibrium with quartz at 150°C or chalcedony at 130°C (Truesdell, 1976). The quartz equilibrium temperatures at 150°C and the Na-K-Ca equilibrium temperatures at 230°C in Wilbur #1 are inferred to represent high initial water temperature (230°C) and slow rate of water movement. Ultimately, little confidence can be placed in the temperatures estimated from the dissolved silica concentrations of the springs because of the numerous possible complications.

In an attempt to calculate a third independent reservoir temperature, gas samples were collected and analyzed. Franco D'Amore and A. H. Truesdell (written communication, 1979) have devised a system of equations for geothermal steam wells which quantitatively relate the concentrations of CH_4 and CO_2 and of H_2S and CO_2 measured at the surface to the temperature in the producing zone. Using their equations the reservoir temperatures in Table 2 were calculated for Wilbur Hot Springs. These temperatures are in excellent agreement with the uncorrected Na-K-Ca temperatures from the spring waters.

Table 2.--Gas Analyses of Wilbur Hot Springs[1]

	Wilbur Hot Springs		
Date	12-11-77	12-11-77	8-16-78
Collector	AHT	AHT	JMT
CO_2	54.2	69.3	76.7
H_2S	2.66	2.94	2.92
NH_3	0.622	0.00	0.0323
H_2	9.36×10^{-4}	2.66×10^{-3}	1.08x10
Ar	0.319	0.217	0.188
O_2	4.71	1.26	0.635
N_2	29.0	18.8	15.1
CH_4	2.36	3.33	3.28
C_2H_6	0.00	0.00	0.00
TOTAL	93.87	95.85	98.22
T	242°C	237°C	227°C

[1]Analyses in mole percent. Analyses by Nancy L. Nehring, U.S. Geological Survey.

The Wilbur Hot Springs water composition may result if one part diluting water such as that in Table 1 mixes with two or three parts of thermal water such as that from the Wilbur #1 geothermal well. This diluting water may be unusual because it contains a high magnesium content from dissolved serpentine. Alternatively, the magnesium from serpentine dissolution may not enter the system until after mixing. Another possible scheme is that 230°C water mixes with connate water similar to that described by White and others (1973) to form the ~150°C water in Wilbur #1 geothermal well. This water then mixes with cold meteoric water in different proportions producing the various spring water compositions. This model is not favored because it requires three different water types. Presently, the time at which the magnesium enters the system cannot be determined. The additional sulfate (115 mg/L) probably arises from oxidation of H_2S in the near surface region.

CONCLUSIONS

The brine in the Wilbur #1 geothermal well result from the mixing of deep thermal water of unknown composition at a temperature near 230°C and connate water such as that described by White and others (1973). Alternatively, the connate water may have been heated to near 230°C and then mixed with meteoric water in proportions of 2:1 or 3:1. This diluting meteoric water may contain as much as 180 mg/L Mg. This mixed water may then slowly rise to the surface without appreciable residence in a large reservoir where Na-K-Ca reequilibration could occur. The depth to the 230°C water is unknown.

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BOREHOLE TEMPERATURE STUDIES OF THE LAS ALTURAS GEOTHERMAL ANOMALY, NEW MEXICO

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ABSTRACT

Three phases of borehole temperature studies have been made relative to the Las Alturas geothermal anomaly in southern New Mexico: i) logging of "free" holes; ii) shallow gradient study; and iii) analysis of data from two 300m tests. A maximum temperature of 62.5°C (145°F) has been measured at 300m in one of the tests, and the data indicate that the source of the anomaly is a hydrothermal circulation system. A simple analysis of the temperature data indicate vertical water flow rates of the order of 1×10^{-9} m/s (∿ 1 ft/yr). The borehole temperature data have provided valuable information for delineating, evaluating and characterizing the nature of the source of the anomaly.

INTRODUCTION

Geophysical, engineering and economic studies all indicate that the Las Alturas geothermal anomaly, located adjacent to the city of Las Cruces, New Mexico (Figure 1), is a potentially economic low temperature geothermal resource for direct heat applications at the New Mexico State University campus (Gunaji et al., 1978; Dicey et al., this volume). Borehole temperature studies have been made in and around the Las Alturas anomaly in three phases: i) regional exploration comprising temperature measurements in "free" holes, ii) drilling and measurement of 30m temperature gradient

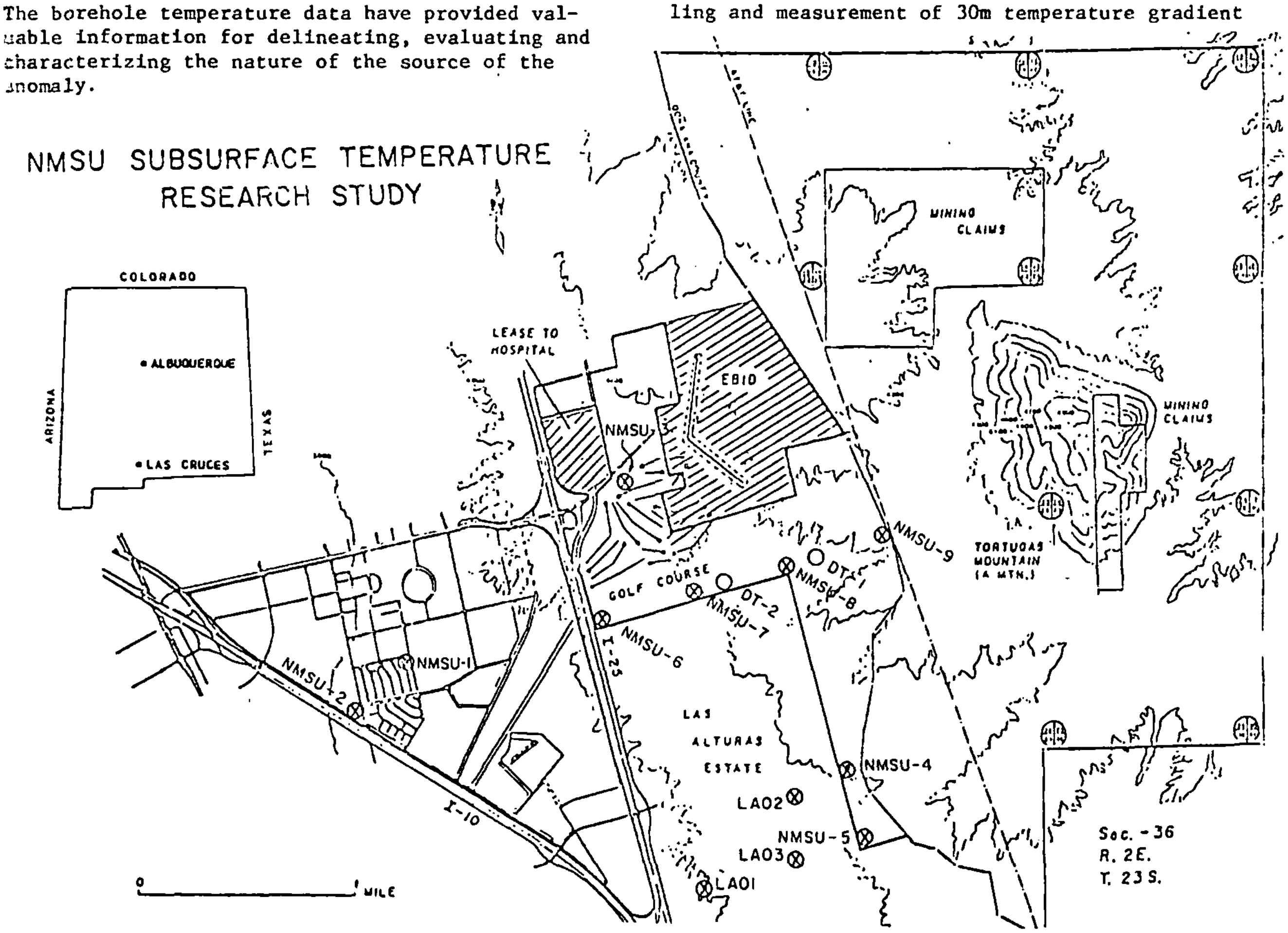

Fig. 1 Index map of that part of Las Cruces, New Mexico, which includes the New Mexico State University campus (to the west) and the Las Alturas area (Wells, labeled 6-9). Also shown are the locations of the boreholes used for temperature studies.

holes over the anomaly prior to the siting of deeper test wells; and iii) drilling and measurement of two 300m test wells. This summary presents the results of the three phases of the temperature study and the interpretation of the subsurface temperature data

REGIONAL BOREHOLE TEMPERATURE STUDIES

There are numerous boreholes in the Las Cruces area, drilled primarily for domestic water supply to depths of a few hundred meters. Eight abandoned water wells were available for temperature measurements in the immediate vicinity of Las Alturas; the locations for six of these (NMSU-1,-2,-3, LA01, 02, 03) are shown on Figure 1. The remaining two wells in the area are located off the map, the J. ABRAMS well being approximately two miles to the NNW, and DA-1 approximately six miles to the east. The temperature data from these eight wells are shown in Figure 2.

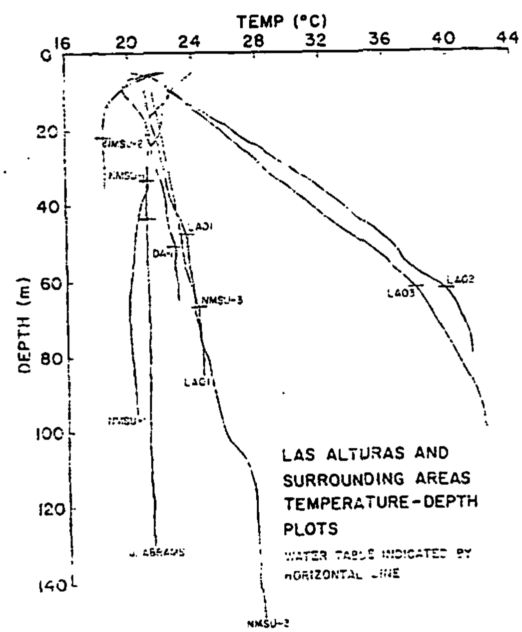

Fig. 2 Temperature data from "free" boreholes in the vicinity of Las Alturas.

The temperature logs fall into three distinct categories: negative gradients, normal gradients, and high gradients. Three wells with negative gradients, NMSU-1, NMSU-2, and J. ABRAMS, are all located to the west of Interstate-25, which approximately divides the irrigated areas of the Rio Grande Valley to the west from the undeveloped areas of the mesa to the east. The data from these wells clearly indicate recharge of groundwater by downward water flow. Three wells east of I-25--NMSU-3, LA01, and DA-1--show reasonably uniform positive temperature gradients of the order of 40°C/km (2.2°F/100 ft) below a depth of 20 m. These gradients are typical of the gradients normally measured in sediments in the Rio Grande Rift. Two wells at Las Alturas, LA02 and LA03, have dramatically higher gradients, of about 300°C/km (16.5°F/100 ft), down to the water table at approximately 60m, below which depth the gradients systematically decrease. These two wells provided the first temperature measurements in the geothermal anomaly and yielded gradients below the water table which indicate that the anomaly is caused by a hydrothermal circulation system. In addition, these two wells provided valuable information for further temperature studies: the high gradients above the water table are essentially established in LA02 and LA03 at a depth of 10 to 20m, which indicates that 30m is an adequate depth for additional temperature gradient boreholes.

SHALLOW GRADIENT BOREHOLES

A program of shallow temperature gradient boreholes was planned to confirm the extension of the Las Alturas geothermal anomaly beneath University land adjacent to Las Alturas, and to provide additional data for site selection for deeper tests. On the basis of the temperature results from LA02 and LA03, 30m was chosen as the depth for the shallow gradient holes.

Two holes, NMSU-4 and NMSU-5, were initially drilled to the east of Las Alturas at the locations shown on Figure 1. The temperature data from these holes, shown in Figure 3, indicate that the anomaly increases to the east of Las Alturas. The measured gradients in NMSU-4 and -5 are 416 and 387°C/km (22.8 and 21.2°F/100 ft), respectively. The anomaly is thereby confirmed to extend to the east beneath University-owned land.

Data from an earlier electrical resistivity study at Las Alturas (Smith, 1977; Jiracek and Gerety, 1978) suggest that the anomaly extends at least one mile to the north of NMSU-4. Based on the resistivity interpretation, the LA02 and LA03 borehole temperature data, and local geologic information, it is thought that the most likely origin for the anomaly is a hydrothermal circulation system controlled by a NNW tending fault. A profile of four boreholes, NMSU-6, -7, -8, and -9, were therefore drilled on an ENE line to the north of the proven temperature anomaly to provide data for deeper test site selection. The locations of the holes are shown in Figure 1. Temperature data from these holes (Figure 3) clearly define the western margin of the anomaly with gradients increasing to the east as follows: NMSU-6, 88°C/km (4.8°F/100 ft); NMSU-7, 320°C/km (17.6°F/100 ft); NMSU-8, 446°C/km (24.5°F/100 ft). The anomaly appears to peak before the eastern hole on the profile, where a gradient of 433°C/km (23.8°F/100 ft) was measured, although this is not absolutely defined. Unfortunately, institutional barriers at this stage prevented the drilling of an additional gradient hole to the east. A tentative interpretation of all the temperature data is that hot water rises along a NNW striking fault to a zone crossing between the two wells NMSU-8 and -9, and then diffuses laterally.

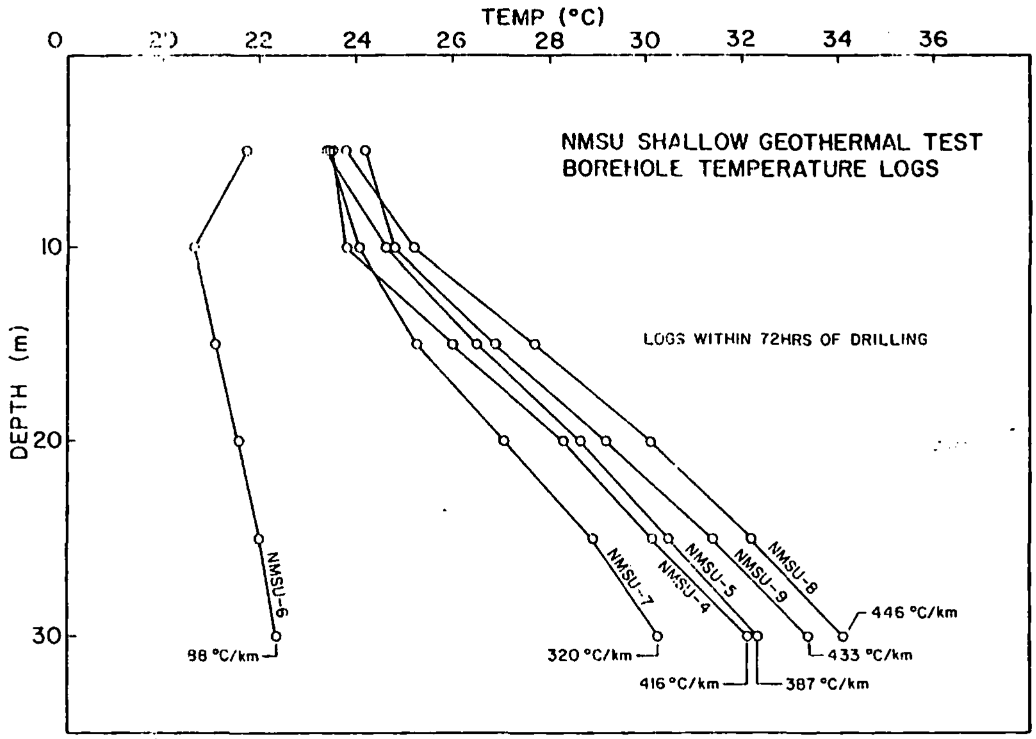

Fig. 3 Temperature data from shallow gradient holes.

300m TEST WELLS

Two 300m test wells were drilled on University land adjacent to Las Alturas under the supervision of L. Chaturvedi of the Civil Engineering Department at NMSU. The wells were sited on the basis of the interpretation of the data from the shallow gradient hole profile, at the locations shown in Figure 1, based on the following logic. The first test well, DT1, was sited 100m east (downdip on the assumed fault) of the apparent peak of the anomaly defined by the temperature gradient profile. With this well it was hoped to intersect and stay in the anomaly source to total depth. The second well, DT2, was drilled 0.4 miles west of DT1 along the profile, between shallow wells NMSU-7 and -8. This site was selected so that if the interpretation of the source of the anomaly as a narrow fault-controlled zone of rising hot water were incorrect, this well would intersect a more laterally extensive source at a greater depth from DT1, but at a closer distance to the potential user, the NMSU campus. If the narrow fault source interpretation were correct, however, DT2 would provide information about the lateral flow from the system, and act as a site for a reinjection well, if required. Wells DT1 and DT2 were drilled in December 1978 and January 1979 and completed as temperature test wells at depths of 300 and 360m respectively with two-inch water filled casing.

Temperature data from the two 300m test wells are shown in Figure 4. The first well, DT1,

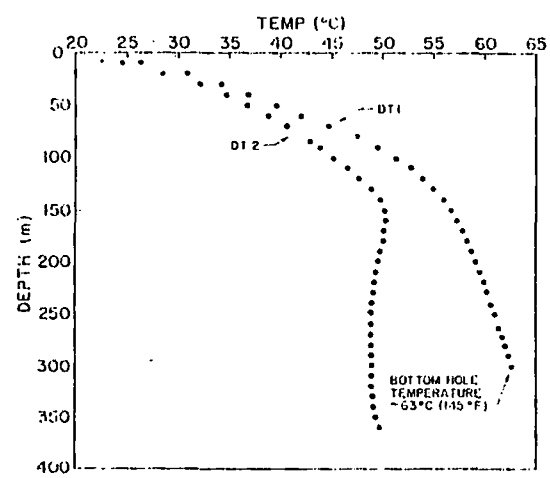

Fig. 4 Temperature data from 300m test wells.

appears to have been drilled either into or very close to the source of the anomaly, and has a positive gradient to total depth, although the gradient decreases all the way down the hole, especially below the water table at approximately 80m. A maximum temperature of 62.5°C (145°F) was recorded at 300m, the base of the well. The second well, DT2, reaches a maximum temperature of 50.8°C (123°F) at a depth of 160m, at which point the gradient becomes negative, indicating a heating of the zone around 160m by a lateral flow of hot water. Below 275m the gradient in DT2 becomes positive again, with the temperature increasing to 49.6°C (121°F) at 360m, the base of the well. These data clearly indicate the source of the anomaly to be in the vicinity of DT1, with a lateral hydrothermal flow towards DT2 to the west. The nature of the circulation to the east of DT1 has not yet been determined.

The curvature in the geothermal gradient in DT1 can be used to estimate the vertical component of water flow in strata penetrated by the well using the relationship

$$q = K \frac{dT}{dz} + \rho CvdT, \qquad (1)$$

where q is the total vertical component of heat flow, K is the thermal conductivity of a zone, dT/dz is the temperature gradient in the zone, ρ and C are the density and specific heat of water respectively, v is the vertical component of water flow velocity, and dT is the temperature drop across the zone. By using temperature gradients and temperature drops across adjacent zones, and assuming q to remain constant, the velocity v can be calculated. Using this technique, vertical water flows of between 0.3 and 1.4×10^{-9} m/s (0.3 and 1.4 ft/yr) have been calculated for the DT1 temperature data. These calculations assumed a density of 1 gm/cm^3 for the effective water density in equation (1). For absolute water flow velocities the numbers given above should be divided by the fractional porosity of the strata (.12 to .30, L. Chaturvedi, unpublished report).

Unlike the measured gradients in DT2 and the six shallow gradient holes, the gradient in DT1 shows some curvature above the water table. This curvature prevents a direct comparison of the shallow gradient in DT1 with the gradients in NMSU-8 and -9, although its temperature at 30m of 34.45°C (94.0°F) is higher than the temperatures in the two flanking holes at the same depth, 34.25°C (93.6°F) and 33.79°C (92.8°F) for NMSU-8 and -9, respectively. DT1 therefore appears to be close to the peak of the anomaly, and the curvature in the gradient above the water table may be due to a significant non-vertical component of heat flow close to the shallow source of the anomaly.

CONCLUSIONS AND FUTURE STUDIES

Borehole temperatures of the Las Alturas geothermal anomaly have partially defined the lateral extent of the anomaly, provided information for the siting of deeper tests, and confirmed the source of the anomaly to be a hydrothermal circulation system. Modelling of electrical resistivity data shows a 5 ohm-m low resistivity zone around the peak of the anomaly delineated by the shallow temperature gradient data (Smith, 1977, Jiracek and Gerety, 1978), but the borehole temperature data have provided the most diagnostic information for delineating and interpreting the nature of the geothermal anomaly.

During the next few months it is planned to drill a deeper test well, up to 750m deep, into the geothermal system (L. Chaturvedi, personal communication). Further shallow gradient holes are planned over the center of the anomaly to provide information for the siting of this well. An extension of the heat flow analysis outlined above indicates a further decrease in the gradient below 300m, with the temperature possibly increasing to as little as 1 to 4°C (2 to 7°F) at 750m, giving a bottom hole temperature in the range of 64 to 67°C (147 to 153°F). This analysis assumes that the vertical water flow rates estimated for the zone from 300m up to the water table are representative of the flow rates below 300m. If this analysis is correct, the source of the hot water for the hydrothermal system could be groundwater circulation down to a depth of a little over 1 km, with a geothermal gradient typical for the area (40°C/km, 2.2°F/100 ft). This extrapolation of the data uses many as yet unproven assumptions, however, which only the drilling of the third deep well can test.

ACKNOWLEDGEMENTS

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CONTINUOUS GRAVITY OBSERVATIONS AT THE GEYSERS: A PRELIMINARY REPORT

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ABSTRACT

A cryogenic gravimeter has been installed at The Geysers to continuously monitor gravity variations at the μgal level. A 38-day record is presented to illustrate the type of information that can be obtained from such an instrument. In addition to information directly related to mass transport within the reservoir, the data reveal a sudden 6 μgal decrease in gravity prior to a local earthquake. We also observe a 5 μgal increase in gravity during a heavy rainfall; however, interpretation of gravity variations at this level is limited by uncertainty in tilting of the gravimeter pier. The potential impact of continuous gravity observations on the study of reservoir characteristics is discussed.

THE CRYOGENIC GRAVIMETER

Although cryogenic gravimeters have been in existence for several years (Goodkind and Prothero, 1968, and Warburton and Goodkind, 1978), this is the first use of such an instrument in a geothermal application. The instrument differs from conventional gravimeters in that mechanical springs and levers are replaced by magnetic fields generated from persistent currents in coils of superconducting wire. These fields support a one-inch diameter superconducting sphere (the gravimeter's only moving part) with a force that does not significantly diminish in time, due to the persistence of superconducting currents. Thus the cryogenic gravimeter does not exhibit the instrumentally produced signal drift which is characteristic of conventional gravimeters. Several years of gravity observations at Pinon Flat in southern California indicate that the total instrumental drift is 0 ± 5 μgal/yr. A planned side-by-side test of two cryogenic gravimeters at Pinon Flat should determine whether this residual variation is of instrumental or geophysical origin.

The short term precision of the cryogenic gravimeter appears to be limited only by the uncertainty with which the contributions from known sources such as earth and ocean tides and atmospheric density variations can be subtracted. The data presented below indicate that this can be accomplished empirically to a precision of ±1 μgal. Spurious tilting of the gravimeter platform can degrade this performance, but tilt will soon be controlled by the addition of an automatic leveling system.

OBSERVATIONS AT THE GEYSERS

A cryogenic gravimeter was installed in February of 1979 in The Geysers steam field at latitude 38° 48' 25" and longitude 122° 48' 50". This site is on a spur of ridge which extends downward from Well Sulphur Bank 19 toward Units 3 and 4. McLaughlin (1974) maps this general area as a quaternary landslide, however, the spur appears to be an outcrop of relatively unfractured Franciscan graywacke. Weathered graywacke slightly uphill of the outcrop was excavated to a depth of 2 - 3 m to expose bedrock, upon which a 3 m high reinforced concrete pier was erected. The gravimeter platform rests on this pier, supported on three points two of which are heavy duty micrometer heads to allow alignment of the gravimeter with the vertical. The active element of the gravimeter is immersed in liquid helium in a Dewar which is fixed to the platform.

The output signal from the gravimeter is in the form of an analog voltage which is filtered electronically for sampling at a variety of rates from 20 seconds to 15 minutes. Barometric pressure as measured by a temperature regulated aneroid barometer is also filtered and monitored at 15 minute intervals. These signals are recorded in the field on digital tape cassettes by a microcomputer-controlled data system, which can also transmit stored data by telephone to our laboratory for inspection.

A sample of raw data together with the results of various stages of its reduction are shown in Figure 1. All four graphs are plotted as functions of time at 15 minute intervals beginning at 067:00:00:00 UTC and ending 38 days later at 105:00:00:00 UTC. The observed gravity signal, Figure 1a, is dominated by smooth tidal variations (the coarseness of this graph is an artifact of the digital plotter: the point density is nearly 1000 points per inch on this scale). The principal component of this signal is the direct gravitational attraction of the moon and sun together with the effects of deformation of the earth's surface due to tidal forces and loading from ocean tides. These tidal effects contain little information relevant to the geothermal reservoir and must be removed before local effects can be observed. This can be accomplished by subtraction of a theoretically generated tide signal, to account for the direct attraction effects, followed by least squares removal of the strongest remaining tidal spectral components, to account for ocean loading effects.

The resulting detided gravity signal is shown in Figure 1b; the upward direction on the plot corresponds to increasing strength of gravity. The detided gravity signal contains variations of up to 13 μgal over a few days; however, most of this is due to large scale atmospheric density fluctuations associated with the motion of weather systems, as is evident from comparison with the barometric pressure record, Figure 1c. The upward direction on this plot corresponds to decreasing pressure, thus the central peak in Figure 1c indicates a strong low pressure system. Superimposed on these major variations are minor oscillations due to the global semidiurnal atmospheric tides and local pressure fluctuations. As demonstrated by Warburton and Goodkind (1977), gravity effects due to major barometric pressure fluctuations can be removed from the detided gravity signal by least squares subtraction of the barometric pressure signal. The fitting coefficient yields an empirical barometric pressure admittance, which in this case amounts to 0.27 μgal/mbar. In other words, as much as 9 μgal of the detided gravity variations in Figure 1b were due to atmospheric effects.

Removing these atmospheric effects produces the residual gravity signal shown in Figure 1d. As in the other plots, a decreasing signal implies decreasing strength of gravity as occurs during mass extraction or surface uplift. Several features which were previously obscured by atmospheric effects are now apparent. Most remarkable of these is the sudden drop of nearly 6 μgal which occured on day 078. Closer scrutiny on an expanded scale reveals that the drop was not instantaneous: the gravity decrease was linear in time over a two and one-half hour period. Moreover, a separate high frequency output channel from the gravimeter indicated no unusual seismic activity or disturbance of the instrument during this period; the decrease was quiet and gentle. However, two hours after this gravity change had ceased, the high frequency channel shows a local earthquake, which, according to Bufe (unpub. data) was of magnitude 3. The sign and magnitude of this gravity event are consistent with an uplift of the gravimeter by approximately 2 cm.

The second noteworthy event in the residual

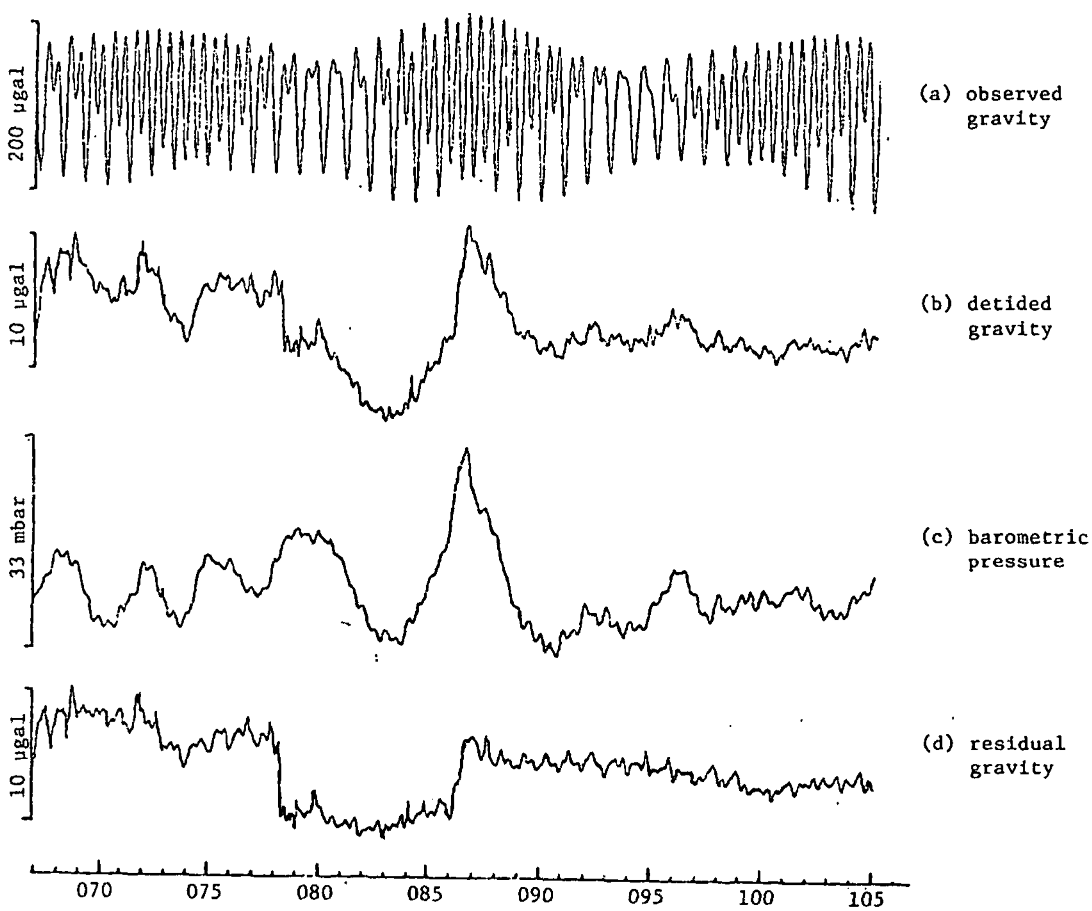

Figure 1. A 38-day data segment from The Geysers, illustrating the extraction of a barometrically adjusted residual gravity signal from the raw gravity and barometric pressure signals.

gravity signal is a rapid but irregular 5 μgal increase beginning on day 086. This increase in gravity occurs over an 18 hour period which coincides precisely with the duration of the only significant rainfall in this 38 day data segment. Analysis of rainfall effects in barometrically adjusted gravity residuals is simplified by the fact that there is a slight decrease in barometric pressure directly associated with the release of mass from the atmosphere. This decrease in pressure equals the weight per unit area of the released mass. If we let Δg represent the gravitational attraction of a sheet of water of thickness equal to the depth of rainfall, then it can be shown that the barometric pressure correction used to produce Figure 1d contributes an amount $-\Delta g$ to the residual gravity signal starting at the time of the rainfall. Thus the net result of rainfall on barometrically adjusted gravity is simply to produce an apparent change by an amount $-\Delta g$ if the sheet of water comes to rest above the gravimeter, or $+\Delta g$ if the sheet lies below the gravimeter. In both cases, these changes vanish as the water in question drains away from the vicinity of the gravimeter. Considering the terrain at The Geysers we would expect a zero change in the present case, or, at most, a change of +2 μgal. Thus the 5 μgal observed change appears to be a spurious effect, most likely due to tilt.

That tilt influences gravity measurements is a direct consequence of the vector nature of the gravitational field. If one assumes that a gravimeter tilted by an angle θ from the local vertical simply measures the component $g\cos\theta$, the theoretical tilt response to lowest order in θ is $-1/2\ g\theta^2$ or -4.6×10^{-4} μgal/μradian^2; i.e., the gravimeter will read a maximum value when aligned with the vertical. This, however, is not the case for the cryogenic gravimeter. The magnetic field geometry currently used to levitate the test mass inside the instrument causes the tilt response to have a larger magnitude and opposite sign compared to the naive estimate. The actual tilt response of The Geysers instrument is $+ 5.9 \times 10^{-4}$ μgal/μradian^2. Thus the gravimeter reads a minimum value when vertical and deviations from the vertical will tend to increase the apparent value of g. Tilts of 40 μradians (1 μgal) would be observable and tilts of 100 μradians (5.9 μgal) would be sufficient to explain events such as seen in Figure 1d. The planned automatic leveling system should limit tilt to less than 10 μradian (0.06 μgal).

In spite of the problems in interpreting the two rapid changes in gravity in this record, the general trend of the gravity residual in Figure 1d shows a slow decrease of 4.5 ± 0.5 μgal over the 38 day segment. Extrapolating this trend yields a rate of decrease in gravity of 43 ± 5 μgal per year, which is in close numerical agreement with the average rate of decrease of 46 ± 7 μgal per year inferred from Isherwood's 1974 and 1977 gravity surveys (Isherwood, 1977). Isherwood showed that this decrease could be explained by the mass deficiency generated by steam production over that two and one-half year period. The close agreement of these two measurements is most likely fortuitous, considering that Isherwood's data represent a yearly averaged effect, that the data were collected during two drought years, and that steam production has changed between 1977 and 1979. The agreement, nevertheless, indicates that the cryogenic gravimeter may be capable of producing results in one month that might take years to accomplish with conventional gravimeters.

GEOTHERMAL IMPLICATIONS

Although the data presented here are insufficient to yield new conclusions at this time regarding reservoir dynamics, they do demonstrate that the cryogenic gravimeter has both the sensitivity and the stability required to produce new results. The continuous nature of gravity observations made feasible by this type of instrument enormously expands the interpretative power of gravity studies. With the addition of a tilt stabilized platform, events which would otherwise be obscured by long term averaging can be detected and events whose contributions to the total gravity change would otherwise be indistinguishable can be separated and identified by their time signatures and correlations with other events.

Even the short data segment presented here, despite its tilt uncertainties, indicates that we will be able to accurately observe the steady decrease in gravity associated with continuous steam production and thus provide the most direct available measure of reservoir recharge, The accuracy of these estimates will be further enhanced by our ability to separate out those sudden effects which appear to be unrelated to mass depletion. It is not unreasonable to expect that meaningful estimates of recharge may ultimately be obtained from as little as 90 days of gravity observations, thereby enabling study of seasonal fluctuations as opposed to multiyear averages.

Furthermore, it may be possible to detect short term mass redistributions within the reservoir that could accompany changes in steam production or changes in reinjection, thereby yielding information regarding the percolation-condensation cycle of steam within the reservoir. In addition, gravity events which are correlated with seismicity could provide clues regarding earthquake mechanisms at The Geysers and their possible relation to reservoir exploitation.

ACKNOWLEDGEMENTS

We thank Richard Dondanville and the staff of the Geothermal Division, Union Oil Company, for their cooperation and assistance, especially during the installation of the site, and Richard Reineman for his technical expertise in fabricating the gravimeter. This work is funded by the United States Geological Survey through the Extramural Geothermal Research Program under Grant USDI-14-08-0001-G-297.

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THE INFLUENCE OF STEAM-WATER RELATIVE PERMEABILITY CURVES ON THE NUMERICAL MODELING RESULTS OF LIQUID DOMINATED GEOTHERMAL RESERVOIRS

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ABSTRACT

Sensitivity analyses for modeling of a hypothetical liquid dominated geothermal reservoir indicate the strong dependence of the results on the assumptions made about the steam-water relative permeability curves. Of significant importance are the critical saturation points for the individual phases and the curvature of plots. The effects are more evident on calculated producing wellbore pressure and projected heat recovery.

INTRODUCTION

The success of numerical modeling for hydrothermal systems depends on the assumptions made about the rock and fluid property data. One piece of information that strongly controls the results of model studies for two phase flow in reservoirs is the assumed values for relative permeabilities.

A review of literature shows that in previously published model studies on geothermal systems, the concept of relative permeability has been treated lightly, perhaps because of lack of information. Relative permeabilities used in the past include systems similar to oil-water models as used by Martin[1], or approximations by simple models such as Corey's[2] as used by Faust and Mercer.[3] Jonsson[4] from his modeling work indicated that relative permeability data had little influence on pressure drop and saturation distribution.

Recently, evidence of actual lab derived relative permeability curves for steam-water systems has appeared in the literature.[5-6] These curves show that the end points, corresponding to the critical water and critical steam saturation, may be much different than the ones used in oil-water or water-gas system.

In this study an effort was made to look at the sensitivity of numerical modeling results to the assumed values of relative permeability data.

DESCRIPTION OF THE MODEL

The numerical model used in this study was a modified version of a program originally developed by Faust and Mercer. Relative permeability curves were furnished to the program through the use of an equation which allowed for selection of a wide range of end points as well as curvatures. The general form of the equation may be shown as follows:

$$K_{rw} = a\,(S_w - S_{wc})^{n_1}$$

$$K_{rs} = b\,(1 - S_w - S_{sc})^{n_2}$$

where S_w = water saturation, fraction

S_{wc} = critical water saturation, fraction

S_{sc} = critical steam saturation, fraction

a, b, n_1 and n_2 are constants.

The model was used for a one dimensional reservoir initially containing hot water with one producing well and no recharge. The heat and mass recovery as a function of time were computed using different sets of relative permeabilities. Table 1 shows some of the specifications used in the model.

Throughout the life of the system, flow toward the wellbore occurs in three distinct periods. The initial period of single phase liquid flow, followed by a two phase liquid-vapor flow and the eventual conversion to one phase vapor flow. In assessing the importance of accurate relative permeability data on modeling results, one must examine the outputs which can be used for history matching purposes. The study of saturation distribution and other profiles in the reservoir as used by Jonsson may be somewhat misleading.

There are several ways to use numerical modeling results for history matching purposes. The results of this study are presented in terms of heat recovery versus time and wellbore producing pressure versus time.

The effect of assumed values for critical saturation of water is shown in Fig. 1. During the two phase flow, the change of critical water saturation from 0.3 to 0.5 could cause significant differences in the performance projection for history matching purposes.

Similar effects for the changes in critical point of the steam is shown in Fig. 2. Variation of the end point for steam has a somewhat smaller effect on estimated heat recovery. For systems with recharge or reinjection there is no need to be concerned about the accurate location of end point of steam relative permeability curve. Effect of the curvature of relative permeability curves on the heat recovery projection is shown in Fig. 3 and is of considerable importance.

A more dramatic effect is seen on the wellbore producing pressure versus time. Significant differences may be observed on the behavior of the pressure curve if the critical water saturation or the curvature of relative permeability curves are varied, Fig. 4-5. The sensitivity of pressure calculations to steam critical saturation and end points are somewhat less.

CONCLUSION

The influence of steam-water relative permeability data on numerical modeling results for geothermal system is significant enough to require more accurate estimates of critical points than that practiced in the past.

Because of ambiguity about the exact nature of steam-water relative permeability curves and contradicting published values about the location of critical points and based on the results of this study, it is imperative that more attention be focused on actually measured relative permeability curves. Field and laboratory derived curves must be carefully examined for establishment of representative relative permeability data.

Table 1. Basic reservoir properties

Pore Volume = 50×10^{12} cc (1.78×10^{9} cu ft)
Production Rate = 20×10^{3} g/sec (160,000 lb/hr)
Permeability = 0.1×10^{-9} cm^2 (10 md)
Initial Pressure = 4.925×10^{7} dynes/cm^2 (714 psi)
Initial Temperature = 236.12°C

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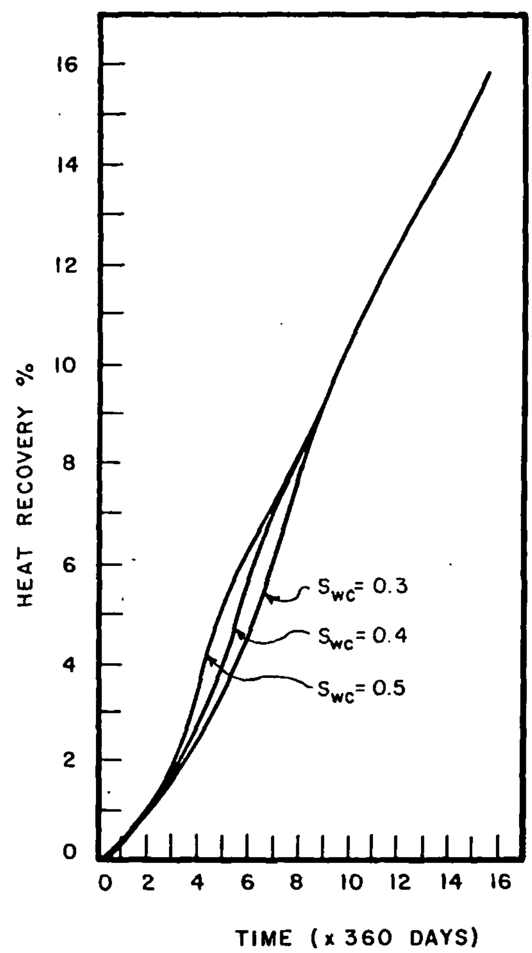

Fig. 1 Effect of S_{wc} on heat recovery vs. time.

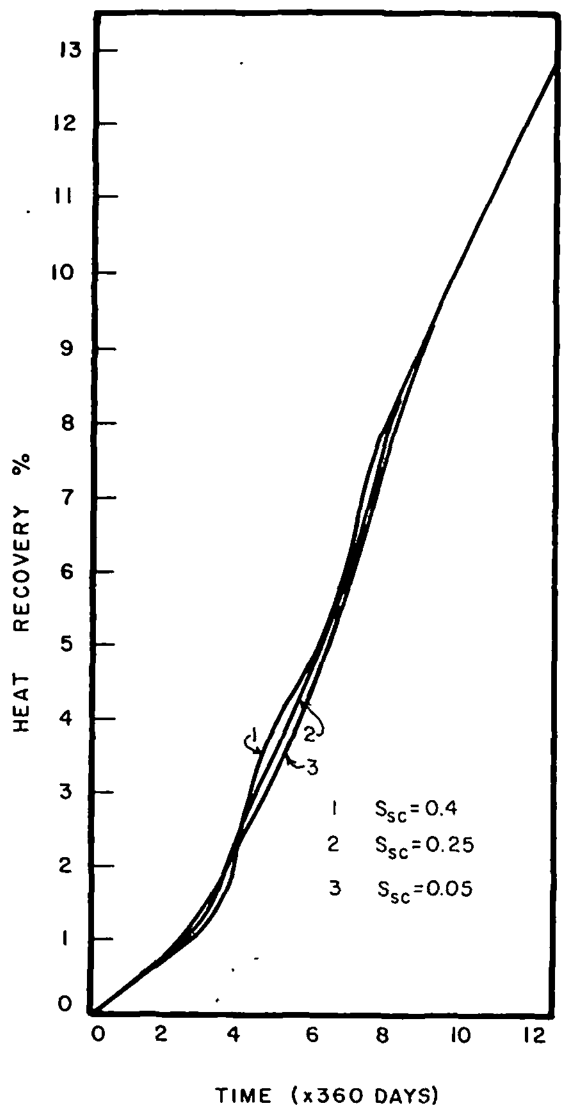

Fig. 2 Effect of S_{sc} on heat recovery vs. time.

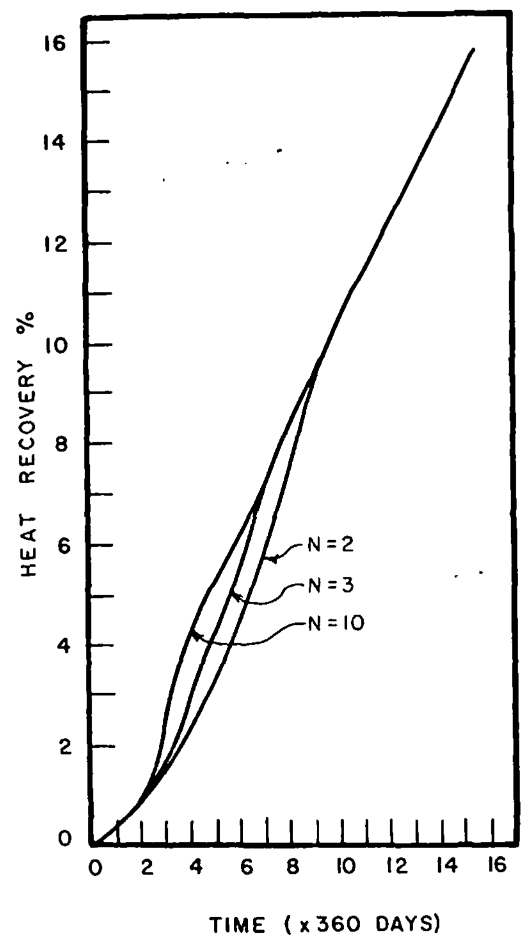

Fig. 3 Effect of curvature on heat recovery vs. time.

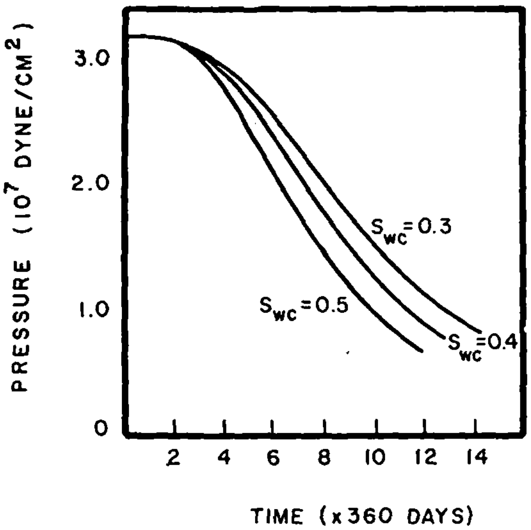

Fig. 4 Effect of S_{wc} on wellbore producing pressure vs. time.

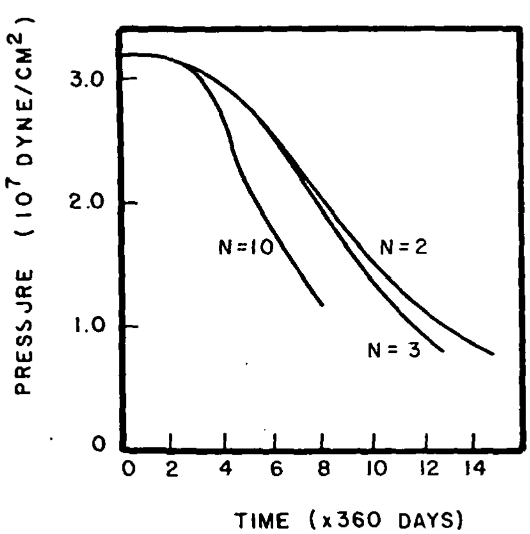

Fig. 5 Effect of curvature on wellbore producing pressure vs. time.

KLAMATH FALLS GEOTHERMAL HEATING DISTRICT

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ABSTRACT

The City of Klamath Falls is proposing to construct a geothermal district heating project. Initially, the system will heat 14 government buildings (Phase I) in the downtown area, subsequently expanded to heat 11 blocks (Phase II), and then to heat the entire 54-block central business district (Phase III). Production wells will be drilled along the east boundary of the City, estimated to supply over 220°F water. A primary 8-inch diameter insulated steel pipeline placed in a concrete tunnel will supply geothermal fluid to a central heat exchange facility at the County Museum Building. Two plate heat exchangers will provide the necessary load for the initial 14 buildings. An injection well is located next to this facility. A closed loop secondary pipeline will supply heat to the 14 buildings at 200°F. This line will consist of buried insulated fiberglass reinforced pipe. The capital cost of the system (Phase I) will be $1.4 million giving an equivalent annual capital, operation, and maintenance cost over a 20-year period of $150,000. Phase II cost of geothermal energy is estimated at $0.29 per therm, whereas the equivalent annual fossil fuel cost is estimated at $0.94 per therm.

INTRODUCTION

The purpose of the 1977 Field Experiment contract awarded under PON EG-77-N-03-1553 to the City of Klamath Falls, Oregon, is to design, construct, and initiate operation of a geothermal space heating district in the central business district of the City. This direct utilization project is for a City-owned and operated system, initially serving 14 City, County, State, and Federal office buildings (Phase I), with initial expansion to serve 11 blocks of commercial buildings along the pipeline route for the 14 buildings (Phase II), with subsequent expansion to commercial buildings on 54 city blocks in the central business district (Phase III). The project will include production wells, injection well(s), transmission lines, controls, and retrofitting equipment for the governmental buildings.

PART I
GEOLOGY, DISTRICT BOUNDARIES, & PRODUCTION FIELD

Geology and Hydrology

The Klamath Falls KGRA is located near the east side and center of the Klamath Basin, a northwesterly-oriented graben. The Klamath Falls urban area is located in the northern and largest portion of the KGRA. The main hot water well area is located adjacent to the eastern fault scarp, over fault blocks that are slightly tilted and raised above the central portion of the graben (Figure 1). The principal geologic formations are lava flows, volcanic breccia (including labilli), locally designated "cinders," and extensive deposits of lacustrine diatomite and tuffaceous siltstones and sandstones.

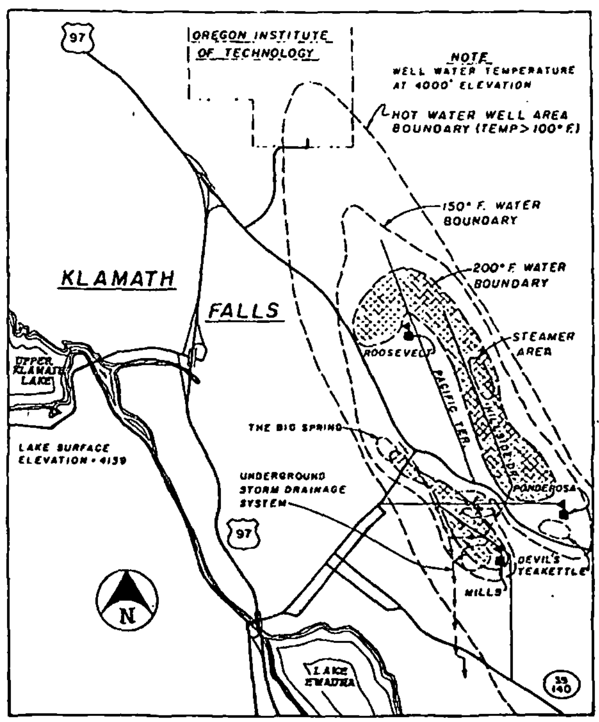

FIGURE 1
Well Water Temperature at 4000' Elevation

In general, the fractured basalts and cinders are highly porous, being capped by a nearly impervious zone of fine grained, lacustrine, palagonite-tuff sediments and diatomite, referred to as the "Yonna formation" and locally as "chalk rock." This formation, Tst on the geologic map, is estimated to be 30 to 150 feet thick in the urban area. It is also interbedded with sandstone or siltstone and fine cinders.

The hot water probably originates from the Cascades to the west and Crater Lake area to the north. Deep circulation of the water is speculated, with the heated water upwelling along the east-side fault in the urban area (Hot Springs District), and then flowing in near horizontal aquifers to the southwest. Contours of static well water elevations indicate a hydraulic gradient of from 0.5 to 3 percent sloping to the west and southwest. In almost all cases, the geothermal water is found in confined aquifer thus producing an artesian pressure head.

Maximum temperatures are found in the vicinity of the main fault, with up to 235°F being recorded. Temperatures tend to decrease down gradient to the southwest, probably due to cooling with time in the aquifer and mixing with colder shallow ground water.

Geothermal wells drilled in the area vary from 90 feet to slightly over 2000 feet in depth. Short term pumping has produced up to 700 gpm (County Museum well), from a single well with 500 gpm (OIT) being pumped on a sustaned basis.

Water levels and well temperatures will vary with time, mainly on a seasonal basis. With usage during the heating season, water levels will drop and temperatures increase. During dry periods (late fall) water levels have also been known to drop markedly, most noticeable in artesian wells. Temperature variations of up to 30°F and water level variations of 20 feet are not uncommon.

The majority of the approximately 500 wells in the urban area use down-hole heat exchangers to provide space heating and domestic hot water, thus only "heat" is extracted from each well. Approximately 55 of the 500 wells pump the geothermal water from the well to be used in surface heat exchangers and then the fluid is disposed of on the surface (storm sewer, sanitary sewer, etc.). The total hot water extracted in the urban area is approximately 2110 gpm in the winter and 340 gpm in the summer. Three cases of successful fluid injection are known in the urban area, all within 500 feet of their production well area.

District Boundaries

Boundaries of the pumping districts for the geothermal distribution system were determined for the urban area. Special attention was given to the boundary for the central business district (referred to as the "Commercial District"), as Phases I, II, and III design were dependent upon the heat load in this district. The remaining districts had their boundaries only approximately located, since their exact heating load was not required for this project. Their location and approximate size would however, be used to locate pipelines and heat exchanger facilities of the Commercial District for possible future expansion. In addition, certain potential well production areas and storage tank areas could be recommended based on district locations.

The boundary of the Commercial District was determined based on four criteria:

1. location of the supply line for the 14 government buildings;
2. the location of private commercial buildings in the downtown central business district;
3. the location of a proposed mini-heating district for 10 church complexes in the downtown area; and
4. consultation with City of Klamath Falls officials.

Based on the above criteria, slightly over 50 city blocks in the downtown area were included in the Commercial District. This area extended from the County Museum (location of the proposed heat exchanger building) to Veterans Memorial Park, varying from 1 to 8 blocks wide and including all but 1 church in the original mini-heating district design. A summary of the heat load calculations of this district for Phase I and II are as follows:

<u>Phase I</u> (14 government buildings)

Peak heat load	15 x 10^6 BTU/hr
Geothermal flow rate	756 gpm (assuming 40°F ΔT)

<u>Phase II</u> (11 commercial blocks)

Peak heat load	28 x 10^6 BTU/hr
Geothermal flow rate	1390 gpm (assuming 40°F ΔT)

<u>Phase III</u> (entire Commercial District)

Building volume	30 x 10^6 ft^3
Unit peak heat load	4.34 BTU/ft^3
Peak heat load	
(space)	130 x 10^6 BTU/hr
+(process)	5 x 10^6 BTU/hr
Total	135 x 10^6 BTU/hr
Geothermal flow rate	6750 gpm (assuming 40°F ΔT)

The remaining district boundaries in the urban area were then located. These boundaries were based on four main criteria:

1. natural topographic features;
2. man-made features;
3. political boundaries; and
4. land use.

Natural and man-made features were the primary controlling items. These included the two lakes, Link River, main ridge lines, the railroad and the "A" canal. Any of these items would be costly to cross. For this reason the local distribution system for a district should fall within these boundaries and only major supply lines would cross them at carefully selected points. City political boundaries did control the district locations to a degree, mainly to simplify future administration.

Figure 2 is a generalized map of the urban area showing the approximate district boundaries. The district areas increase with distance from the central business district, due to the reduction in heating (population) density in the suburban areas. Heating loads for each district have not been determined, however it is estimated that the majority are approximately equal.

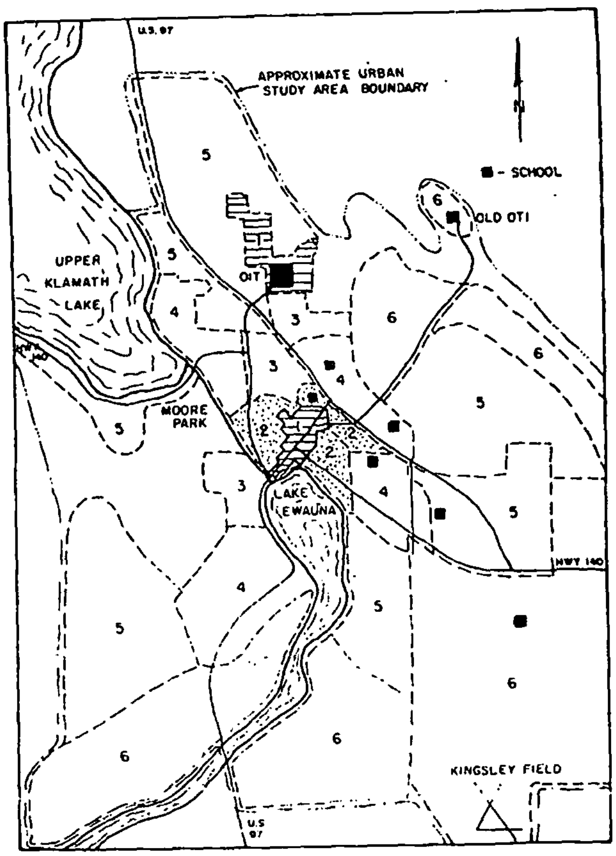

FIGURE 2
Heating District Boundaries

Based on discussions with City planning officials, a priority was given each district as to its probable inclusion into the system. Thus districts labeled "1" on Figure 2 would be developed first, "2" second, and so forth. Districts with priorities "1" and "2" have their boundaries fairly precisely determined for planning purposes, whereas priorities "5" and "6" have more flexible boundary locations due to the possible effects of future growth before these districts come on-line.

A probable development schedule for each of the priorities would be as follows:

District Priority	Time of Development
1	0-2 years
2	2-5
3	5-10
4	10-15
5	>15
6	>20

It should be noted that two of the Districts immediately adjacent to the Commercial District were given a low "4" priority for development. This recognizes the fact that a great portion of these districts are already heated by individual wells, thus some of its heating needs are already being met by geothermal. Future development may be based on either expanding the service load of each existing well (from one to four houses, for example), or by providing a heating district similar to the other districts.

Production Field Locations

Several areas near or in the urban area have potential as production well sites. Each area was evaluated as to certain desirable characteristics, which included:

1. proximity to users so as to minimize supply pipeline lengths;
2. elevation head to provide gravity feed;
3. availability of public land for development; and
4. information on geothermal fluids existing in or adjacent to the site.

Based on the above criteria and geologic information, seven areas have potential to supply the necessary fluids for the near term or future development of the area. In many cases, additional geological, geophysical exploration, and/or drilling need to be performed to verify the existence and characteristics of the resource. Figure 3 is a map to the same scale as the geologic map indicating the seven sites. The numbers also indicate the order of recommended investigation and development of the sites. Site 1 is discussed in detail.

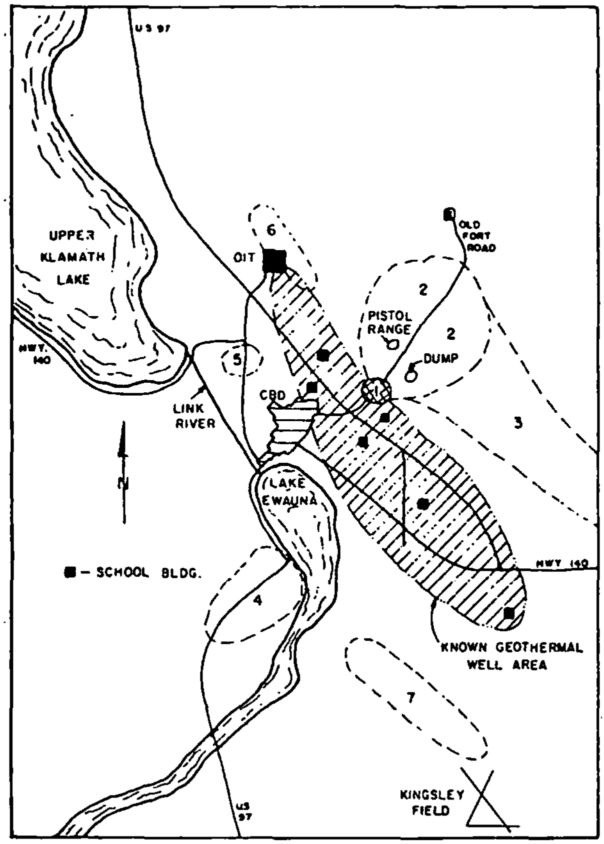

FIGURE 3
Production Fields

Site 1 is located adjacent to Old Fort Road on the City boundary. It is situated approximately on the major fault zone, the source of the majority of the geothermal fluids for the area. Several geothermal wells exist in the area, the majority of which are 200 to 400 feet in depth with with one being 795 feet deep. The hottest temperatures known in the urban area are found here as indicated by one shallow "steamer" and the deep well having a bottom hole temperature of 234°F.

The potential exists for the City of Klamath Falls to acquire wells and/or land in the area for development of the field. The City also has a small triangular piece of land near the intersection of Laguna Street and Old Fort Road that could provide a drilling site for one well.

The site has good elevation head above the central business district and a good easement for a pipeline along streets at a fairly uniform grade. The main disadvantage of the site is that approximately 15 to 20 shallow residential wells exist within 1500 feet of the most likely drilling sites. These shallow wells could be affected by long-term pumping of production wells. The area is also limited as to the total number of production wells that could be drilled due to minimum spacing, however, the demand for the 14 government buildings (756 gpm) could easily be supplied. The total demand for the "Commercial District" (6750 gpm) could probably not be supplied by this field and the availability of land and wells to the City. The location near the main fault zone would give this site the best priority for production wells.

Pump Test

During July, 1978, a pump test was performed on the 795-foot deep well in Site 1. This well has a 14-inch diameter casing to 229 feet and static water level at about 75 feet below the surface. Bottom hole fluid temperature has been measured at 234°F, and dry rock temperatures reported as high as 250°F. It is estimated that this well is drilled into or is very near the major fault zone of the area. Twelve observation wells were selected within 1500 feet of the production well. Unfortunately, no well in the area extended to the same elevation depth as the production well.

Each of four phases of the test lasted 48 hours:

1. pumping without injection (water wasted to the sewer);
2. rebound;
3. pumping with injection; and
4. rebound.

Personnel from LBL and OIT students assisted in the project.

Observations:

1. Maximum production rate was 300 gpm at 100 foot drawdown.
2. Maximum surface flow temperature at the well head was 224°F.
3. There appeared to be no significant effect on adjacent wells when pumping without injection.
4. There appeared to be a measurable effect on adjacent shallow wells with pumping and injecting into a shallow well.
5. Flow of the production well was limited due to caving and filling of the well below the casing (estimated at 82 feet), which was determined after the test.

PART II
HEAT LOADS, WELLS, HEAT EXCHANGERS, AND PUMPS

Heat Load Determination

Commercial District Heat Load. Since it is impractical to do detailed heat loss calculations on each of the downtown buildings, a typical block was chosen and the total downtown area heat loss projected from the values obtained.

The typical block was selected to contain a mix of building construction with particular attention to materials, number of floors in the buildings, display windows, and type of business. The block chosen is bounded by North 7th, North 8th, Main, and Pine Streets. Building construction is primarily masonry (brick) with built up roofs and a good mix of ceiling heights, some of the establishments having been remodeled with lowered ceilings. Businesses include an insurance office, two jewelry stores, two clothing stores (1 men's, 1 women's), an office supply store, two home furnishing stores, a photographer's studio, a restaurant, and an outdoor sporting goods store with office space on the second floor.

The heat loss of each establishment was calculated using ASHRAE recommended procedures and design temperature for the Klamath Falls area. Actual values used were: inside temperature of 60°F (with 10°F night setback during the coldest nighttime periods); and design temperature of 0°F, which will include 99% of the time. Each establishment was inspected, measured, and construction estimated and/or obtained from the building occupants. Heat losses were estimated based on ASHRAE values for the materials and construction involved, with wind speed of less that 7 mph.

Meteorological data was obtained from the U.S. Weather Bureau downtown station for January, 1978, and the degree days for the month calculated. Heat losses for the buildings were again calculated using the January weather data and estimated fuel consumption was compared with actual fuel consumption obtained from building occupants' January fuel bills. Calculated values were 4.9% higher than actual fuel consumption.

From the information thus obtained, a heat load constant of 4.34 BTU/H/ft^3 based on building volume was estimated and applied to building volumes obtained from the City Planning Department.

Fuel consumption and process heat loads were obtained from special case establishments in the proposed district. Domestic hot water was estimated at 10% of the district heat requirements. These loads were added to the loads calculated on the basis of building volume to obtain the total heat load.

Future expansion was provided for by adding approximately 12% to the building volume heat load estimates but no additional high process loads were estimated since zoning restrictions prohibit such future expansion in the area covered by this district.

Total estimated heat load for the downtown district is 135 x 10^6 BTU/hr and will require 6,750 gpm of 200°F water with a 40°F temperature drop.

The heat loads for 10 of the 14 buildings included in the initial district had been estimated using ASHRAE methods for the 1977 county feasibility study. Actual consumption of fuel was compared against estimated consumption for a period of 5 years for that study. Estimated consumption was approximately 8% higher than actual consumption.

The heat load values obtained from the 1977 study were within 15,000 BTU/hr of the values obtained when using the constant of 4.34 BTU/hr/ft^3. The heat load for the Phase I-14 government buildings based on the constant is 15.3 x 10^6 BTU/hr.

Selection of 40 Temperature Drop. Ten of the 14 initial buildings presently have hot water heating systems at average temperatures ranging from 144°F to 190°F. One building is partially heated by heat pumps supplied with 100°F water and partially by 190°F water in finned tube convectors. An electrically heated building obviously will require a retrofit to convert to hot water by placing water coils in the duct plenums.

The county jail has a very old steam system that is inadequate now, but the existing finned tube radiation can be utilized with additional lengths of radiation. Modifications must be made to enlarge the steam condensate return lines.

The Veterans Memorial Building, also steam, has combination radiation and forced air heating. The building has been remodeled and office size reduced several times with installation of new radiation so that there is excess radiation capacity at steam temperatures. Use of 180°F average water temperature will reduce the radiation capacity but this loss can be offset by increasing the coil capacity in the central forced air unit.

The hot water heated buildings, in general, presently have average temperature in the 180°F to 190°F range. Reduction of the average temperature to 180°F will reduce the heating capacity by about 10%. Since the buildings are old and were originally designed with over capacity, the present systems will require little retrofit except in the mechanical room valving and control systems and addition of appropriate valves and controls to accept water from the district supply system.

Use of average water temperatures much below 180°F would require the addition of radiation, fan coils, etc., as appropriate. For instance, use of 170°F average water temperature would reduce the heating capacity by about 23% on the average. This reduction would be excessive in most of the buildings.

Since the heat loss of the insulated pipeline is less than 1°F at full capacity, 200°F heat exchanger outlet, and 160°F inlet temperatures will provide very nearly the 180°F average temperature considered desirable.

Wells. It is anticipated that two production wells and at least one injection well will be required for the initial 14 building district. The production wells will be near the well on Old Fort Road which was tested in July, 1978. The existing well at the museum, just a few feet from the heat exchanger building site, will be used as the injection well.

Production Wells. There are several geothermal wells within a 1,000-ft radius of the proposed production well sites but all are much shallower and have downhole heat exchangers (DHEs) installed so no water production data is available. At the time the test well was drilled in 1961, its production was estimated to be 500 gpm or more, but no pump test was made since it too was intended for DHE operation. Since the proposed production well sites are within a few hundred feet of the test well, the strata and production rates are expected to be similar. Local well drillers estimate that an average 1,000-ft well in that area would produce 500 gpm or more and that an excellent well might produce as much as 700 to 800 gpm.

In order to utilize the expected 500 gpm, 8-inch diameter pumps will be required which in turn requires 10-inch minimum diameter casing at the pump level. Six-inch diameter pumps (with 8-inch diameter casing) could each provide half of the 756 gpm required for the initial 14 buildings, but since the system is intended to be expanded, it is advisable to case the wells to allow for maximum production. This will also allow the use of only one well and pump the majority of the time, and greater efficiency in the use of electrical pumping power.

Assuming strata similar to the tested well will be encountered, the proposed casing program is for 10-inch casing to be set to where hard rock is encountered at 350 to 400 feet and to continue with 8-inch to total depth of 800 to 1,000 feet. There is a possibility that the producing aquifer will be competent and casing not required, but it is more likely that full-depth casing will be required, perforated at the producing levels.

Air or water rotary drilling is the preferred method, thus eliminating the possibility of drilling mud entering the formation, caking, and requiring extensive cleaning operations in order to attain full production. This differs somewhat from the better known geothermal drilling methods where the mud is required to maintain pressures. In those conditions, the downhole pressures prevent mud from entering the formation and the downhole pressure will remove mud when the well is first produced. Where the static level is well below the land surface, mud can be forced a considerable distance into the formation, may cake there and require extensive cleaning to attain full production. Where formation pressures do not force fluid to the surface, the weight of the mud is not required to control the hole.

It must be emphasized that drilling and testing of the production wells is of prime importance, and should be completed as quickly as possible. The design of the entire system depends on the actual temperature and flow production characteristics of the wells and must be considered preliminary until these parameters are known. Pumps, pipeline, heat exchangers, temperature drops, retrofits, etc., can be designed (with attendant cost changes) to provide the required heat with fairly large changes from the well characteristics assumed--but they must be known to provide the basis for the final design. The axiom used in low-temperature direct applications of "never design the system till the temperature and flow available are known" is very true and a good one to follow.

Injection Well. Injection of "used" geothermal water is necessary both from an environmental as well as political point of view. Environmentally, depletion of the reservoir as well as surface thermal pollution are the major considerations. Chemical pollution does not appear to be a serious concern due to the low dissolved solid content (800 ppm) and non-toxic ions (mainly sulfates and carbonates). Political considerations stem mainly from effect on individual residential wells. Temperatures of wells in the urban area generally increase during the winter use period, due probably to less mixing in the reservoir, thus less cooling. Seasonal level changes have also been noted. Water level changes can be tolerated to a degree in wells with downhole heat exchangers (say 10-20 ft), however, owners may object to any noticeable changes (especially lowering). Visual impressions of "waste water and steam" are also of concern to area residents. State regulations do not presently require injection, however, their passage may occur shortly, thus they must be anticipated.

The location of injection wells is a difficult task in the urban area as the local geology and hydrology is not completely understood. The effect of injecting in nongeothermal areas versus known geothermal well areas is also not understood. Several cases of injection are reported earlier in this report, however, these have not been studied in detail. No noticeable effects due to injection have been reported in these cases.

The two greatest concerns of this project are to minimize the effect of production and injection on adjacent wells in terms of level and temperature. Injecting near production wells may preserve the level of the reservoir, but lower the temperature. Injecting away from production wells will eliminate the temperature effects, but will probably not assist in maintaining the reservoir level near the production zone.

Three locations for injection wells have been considered:

1. near the production zone on Old Fort Road;

2. near the County Museum well (or using the actual well); and
3. near the end of the secondary supply line (near the County Courthouse).

Location 1 is best to minimize the effect on adjacent wells in terms of water levels, but may cause premature temperature breakthrough. Location 2 eliminates a return pipeline to the production zone and can use the existing museum well. Temperature effect would also be less due to lower well water temperatures in the injection area (around 180° to 190°F). Location 3 could be considered if the primary heat exchange facility were eliminated, thus eliminating the need for the closed-loop secondary pipeline. Geothermal water would then be delivered directly to all buildings, similar to the Icelandic system. Each building would then be required to supply their own heat exchanger or use the fluid directly. Injection near the end of the line would eliminate the need for a return line, but may effect cold (60° to 90°F) water wells in the area.

LBL recommends the second location, which is the main one considered in this report. The drilling and testing of at least two deep wells in the production area is necessary to better evaluate this recommendation.

Drilling Costs - Klamath Basin. Well drilling costs in the Klamath Basin by cable or rotary rigs up to 3,000 feet are as follows:

$1.00 per inch of diameter per foot of depth in "soft" rock, and
$2.50 per inch of diameter per foot of depth in "hard" rock up to 500 feet in depth.
For every additional 100-foot increment, add $1.00 per foot of depth.

Casing costs can be estimated at $1.05 per inch of diameter per foot of depth. Full-depth casing is assumed for all wells.

Using these costs, which include mobilization and demobilization, and assuming that the production wells encounter the same amounts of "hard" and "soft" drilling that the test well log indicates, drilling and casing of a 1,000-ft well would cost $38,898 with full-depth casing. These costs are expected to rise approximately 10% in the very near future, and do not include costs for drilling mud, additional air compressors if required, foaming agents, etc.

Central Heat Exchangers

There are three basic methods of transferring the heat from the geothermal water to the building air space. These are: 1) Use the geothermal water directly in the building heat emitters; 2) Utilize individual building heat exchangers to transfer heat from the geothermal water to a building closed-water loop which in turn transfers heat to the heat emitters; and 3) Utilize large central heat exchangers to transfer heat to a district closed-water loop and supply the buildings with fresh-heated water to be used in the heat emitters. Each system has its advantages and disadvantages.

Even though the geothermal water in Klamath Falls is relatively pure, the direct use in heat emitters is not recommended, especially in the fan coil units of air handling systems and small tube baseboard units. Experience with direct use at OIT, Shadow Hills Apartments, Kingswood Manor apartments, and several businesses in the East Main Street area indicate that life expectancy of the fan coils may range from 1 1/2 or 2 years to 12 or 15 years. At both Shadow Hills and Kingswood Manor, leaks in fan coil units developed in 2 years or less. Inspection of the units shows corrosion occurring in the units primarily at soldered joints and changes in direction of the tubes at the ends, although a few leaks have occurred in the main body of units where tubes are straight. At OIT, five failures have occurred in the 14 years the units have been in service, again with most near soldered joints, headers, and at the ends of coils. There are no known direct uses of small-finned tube baseboard units, but since the materials and methods of construction are similar, it is assumed they would have similar life times. The exact nature and cause of the failures, and as importantly the difference in life of some units, is not known but is under study by OIT, Battelle N.W. and Radian Corporation in a joint effort. Until the cause of the failures can be determined and/or corrected, direct use is not considered practical.

The usual materials of construction of fan coils are copper tubes with pressed aluminum fins. Other materials which may have longer life are available, but reduce the heat transfer efficiency and are more costly. As far as is known, all units presently installed in any of the air handling units in the buildings to be heated in the district do have copper tubes.

Direct Use. The advantages of direct use are that water would be supplied to the buildings at higher temperature since heat would not be lost in a heat exchanger and the district construction cost would be much less. The main disadvantage is that perhaps few customers would hook up since the reduced cost of heat energy would be offset, or perhaps exceeded by, increased costs of maintenance. Since experience is limited, the cause of failures is unknown, and useful lives of components apparently varies considerably even within individual systems; maintenance and replacement costs are next to impossible to evaluate, and an economic comparison between this system and others is impossible.

Individual Building Exchangers. The second method, that of supplying geothermal water to each building where heat is transferred to a closed loop within the building, is a viable alternative and offers several advantages but in the overall, is more costly due to the economics of size.

Each building would have its small heat exchanger, probably a plate type, with the associated valves and controls to control the pressure flow, and temperature in the small heat exchanger and a pump to return water to the geothermal return line. The normal temperature, flow, and pressure controls would also be required in the building closed-loop system including a water treatment system if so desired.

The individual heat exchange system does not require insulated return lines since all energy to be extracted is taken at the use site, and the control system is somewhat simplified since pressure, temperature, and flow balance across heat exchangers are the responsibility of the building operators. Where the geothermal collection or return is a very simple system, such as direct discharge to storm drainage systems or where the collection system is much shorter than the supply system, this method is probably the most desirable type. Where the collection or return system is essentially the same as the supply system, i.e., disposal is at one end of the system, pipe sizes and pumping requirements will be the same for both supply and return except where elevation head provides advantages.

Advantages:

1. Initial and operation and maintenance costs of the district system are less.
   a. No heat central exchangers.
   b. Simple control system.
   c. Uninsulated return lines.
2. Building operators have the option of direct use or heat exchange.

Disadvantages:

1. Overall initial and operation and maintenance costs are increased but a larger portion is shared by energy users.

Central Heat Exchangers. The third type, centralized heat exchangers, is similar to the initial proposal. (Klamath County Geo-Heating District Feasibility Study, 1976.) Actually, the initial proposal was somewhat of a combination of individual and centralized heat exchangers since two buildings, the museum and fire station were on their own exchanger; the two City buildings were on another exchanger, and the County complex of six buildings on a third exchanger. This type of "mini district" system is viable where groups of buildings are under control of a single entity, i.e., the City, County, or perhaps a shopping center. Where many buildings are under separate control, the cooperation required for the initial installation expenses pro rata seems unlikely. The centralized system is more expensive for the district, but since higher quality energy is delivered to the customer (cleaner hot water), charges for the energy can be increased to offset the costs.

Advantages:

1. Lower initial and operation and maintenance costs to customers which increases the desirability of hookup.
2. Lower total cost when both district and customer costs are considered.

Disadvantages:

1. Higher initial and operation and maintenance costs to district.
   a. Insulated return lines.
   b. More complex controls.

For this district heating system, the central heat exchanger method was selected for the following reasons:

1. Direct use is not desirable due to corrosion.
2. Use of the existing storm sewer system for disposal is impossible.
3. At the present time, disposal appears to be required at one end of the system.
4. Customer hookup in the districts is to be encouraged as the system is expanded.
5. Overall costs (district plus customer) are lower.
6. High-temperature resource is available, so temperature loss across the exchangers is not critical.

Heat Exchanger Type. There are two types of heat exchangers that have proven most satisfactory in geothermal service: 1) tube and shell-straight through with geothermal in the tube side, and 2) plate type.

Tube and shell exchangers were never seriously considered for this application. It is well known that in applications where tube materials other than steel are required, and where close approach temperatures are required, tube and shell exchangers are much more costly. Other major disadvantages are lack of flexibility to accommodate changes in temperature and flow conditions to meet load changes, i.e., additional buildings, difficult and time consuming cleaning when required, greater floor space required, and they are less efficient.

For comparison purposes, specifications of a tube and shell exchanger for application in the initial 14-building design was obtained from a tube and shell exchanger manufacturer.

Geothermal water-tube side - 336 gpm
  219°F inlet 174°F outlet
Secondary water-shell side - 378 gpm
  160°F inlet 200°F outlet
Tube length - 38 ft (would be supplied as 2 units, 19 ft each)
Tube diameter - 5/8 in
Tube material - cupro nickel
Shell diameter - 18 in
Overall length - approximately 23 ft
Price - $36,300 ea. - 2 required

The plate type exchanger is generally considered superior in applications for liquid-to-liquid heat transfer where close approach temperatures are desirable and materials other than mild steel are required. They require little floor space, are easily cleaned, and are much more efficient. Of particular importance in this application is the case of changing exchanger surface area to accommodate changes in flow and temperature conditions by adding or removing plates.

For instance in this application, as individual buildings come on line, plates could be added. Since the plates are in parallel, flow is increased while pressure drops remain the same and inlet and outlet temperatures remain the same. Small changes in secondary loop inlet temperature (or primary for that matter) can be accommodated by adding or removing plates and still maintain outlet temperature. Where large secondary inlet water temperature changes are made (for instance, if all buildings changed from a 40°F T to 60°F T) it may be necessary to series portions of the exchanger. This can be accomplished with external plumbing while using the same exchanger plates and adding a blocking plate between the seried sections.

Design Recommendations. For the initial 14-building district 2 plate type heat exchangers are arranged in parallel with automatic controls to stage the flow in both the pumps and exchangers. Thus, loads up to 50% of peak will be handled by one pump and exchanger, reducing pumping costs and providing maintenance time. One unit will provide the heat required approximately 65% of the time during the Klamath Falls heating season.

The exchangers will be selected to operate at the minimum flows required for domestic water heating during summer months and also allow the addition of plates to handle increased loads while maintaining inlet and outlet temperatures as the district is expanded.

Plate heat exchanger general specifications:

Type - Single pass with 150
316 sst plates EPDM gaskets
Size - 9'3" long x 1'7" wide x 5' high
maximum platage
Geothermal side - 219°F Inlet
176°F Outlet
4.3 psig pressure drop
(1,000 gpm maximum flow with full platage)
350 gpm flow
Secondary side - 200°F Outlet
160°F Inlet
3.7 psig pressure drop
(1,000 gpm maximum flow with full platage)
378 gpm flow
Cost - $14,000 ea. - 2 required

Life of the 316 sst plates is expected to be 30 years or more in the Klamath Falls geothermal waters. Gasket life is expected to be 5 years with frequent cleaning, and gasket cost is $41 each. The unit can be disassembled for cleaning and reassembled in 4 hours by 2 men. Additional plates for future expansion cost, $80 each including gaskets. Estimated maintenance costs--5 years--$6,340.

Pumps

Production Well Pumps. The production well pumps are vertical turbine with variable speed fluid drive. The variable speed drive provides for continuous operation of the pumps to provide constant pressure at the well head and in the supply line under varying flow requirements. Pump discharge pressure is monitored by the fluid coupling control which changes turbine shaft speed to maintain constant discharge pressure from no-flow to full-flow conditions and eliminates the need for storage tanks required with intermittent pump operation. The turbine shaft is always rotating providing lubrication for reduced bearing and shaft wear. The drives have proven successful in the Oregon Institute of Technology and Presbyterian Intercommunity Hospital systems where it is estimated that bearing, shaft, and motor life have been doubled.

Vertical turbine pumps are selected by choosing the most economical combination of pump bowl and number of stages that will produce the desired pressure and flow rates. Attention must be given to efficiency and where variable speeds are a requirement, the pump curve should be as flat as possible to maintain high overall wire-to-water efficiency.

Well Head Pumps.

Vertical turbine with variable speed drive:

Rated flow at 1750 RPM	500 gpm
Column length	350 ft
Column diameter	8 in
Bowl diameter	9 3/4 in
Shaft diameter	1 1/2 in
Number of bowls	11
Discharge pressure	20 psi
Motor (electric)	75 hp
Drive - torque converter type 2% slip at full load.	
Rated Capacity	75 hp
Wire-to-water efficiency	72%
Current cost	$41,488
Number required	2 ea
Estimated maintenance costs:	
Change packing & lubricate, 6-mo. interval	$29
Pull pump, inspect & replace bearings, 3-year intervals	$4,000
Overhaul variable speed drive, 5-year interval	$580

Injection Pump. Since the museum well has been selected as the injection site, and since the well is a flowing artesian well, an injection pump will be required. It is assumed that injection pressures for a good producing well, are approximately equal to the drawdown pressures plus any artesian pressure when that well is used for injection. This has been experienced at Raft River and in Klamath Falls. Additional pumping pressure of about 25% is recommended to account for the difference in temperature when injection temperatures are lower than production temperatures, gradual fouling of the well with continued use, and to account for any injection of debris, scaling, etc.

The drawdown curve of the museum well indicates a drawdown of 50 ft at 800 gpm production. The well normally has a 2-foot to 4-foot artesian head. Injection pressure then should be approximately 54 ft plus 13.5-ft allowance for fouling for 67.5-ft total or 29.3 psi.

For injection, a horizontal centrifugal pump was selected. The pump curve is flat at these low pressures. The most economical method of maintaining NPSH under varying suction flow conditions is to divert a portion of the discharge back to the inlet to maintain NPSH well above the cavitation pressure, rather than a variable speed drive.

Injection pump:	
Horizontal centrifugal	
Rated output @ 1750 RPM, 800 gpm @ 35 psi	
Inlet diameter	5 in
Outlet diameter	6 in
Impeller trimmed to	9 13/16 in
NPSH	8 ft
Motor (electric)	20 hp
Wire-to-water efficiency	71%
Including bypass piping and controls, base, and guards	
Current price	$2,587
Estimated maintenance costs:	
Change packing & lubricate, 6-mo. interval	$20
Inspect pump and replace bearings, 5-year interval	$275

PART III
DISTRIBUTION PIPING NETWORK AND CONTROLS

The Network

The design of the district heating piping network is of vital importance to the economics of the system. There is a trade-off between economics and reliability depending upon the pipe material, insulation, and placement method selected. It is important for this piping network to be arranged according to a predetermined plan, where basic conditions, such as heat demands, production field identification, and siting of wells are decided at an early stage.

Two types of piping systems are required in the Phase I design; an insulated primary pipe to supply the central heat exchanger from the wells; and insulated supply and return pipes providing heat to the buildings from the central heat exchanger (see Figure 4). Both systems will be located in developed residential and commercial areas, requiring underground distribution.

The secondary pipeline is designed to initially supply heat to the 14 government buildings (Phase I), however, it is also sized (8 in) to supply adjacent commercial buildings, on 11 city blocks (Phase II). The primary pipe is sized to supply Phases I and II, however, it will be housed in a concrete duct that will have space available to install a future pipe that will handle Phase III of the project.

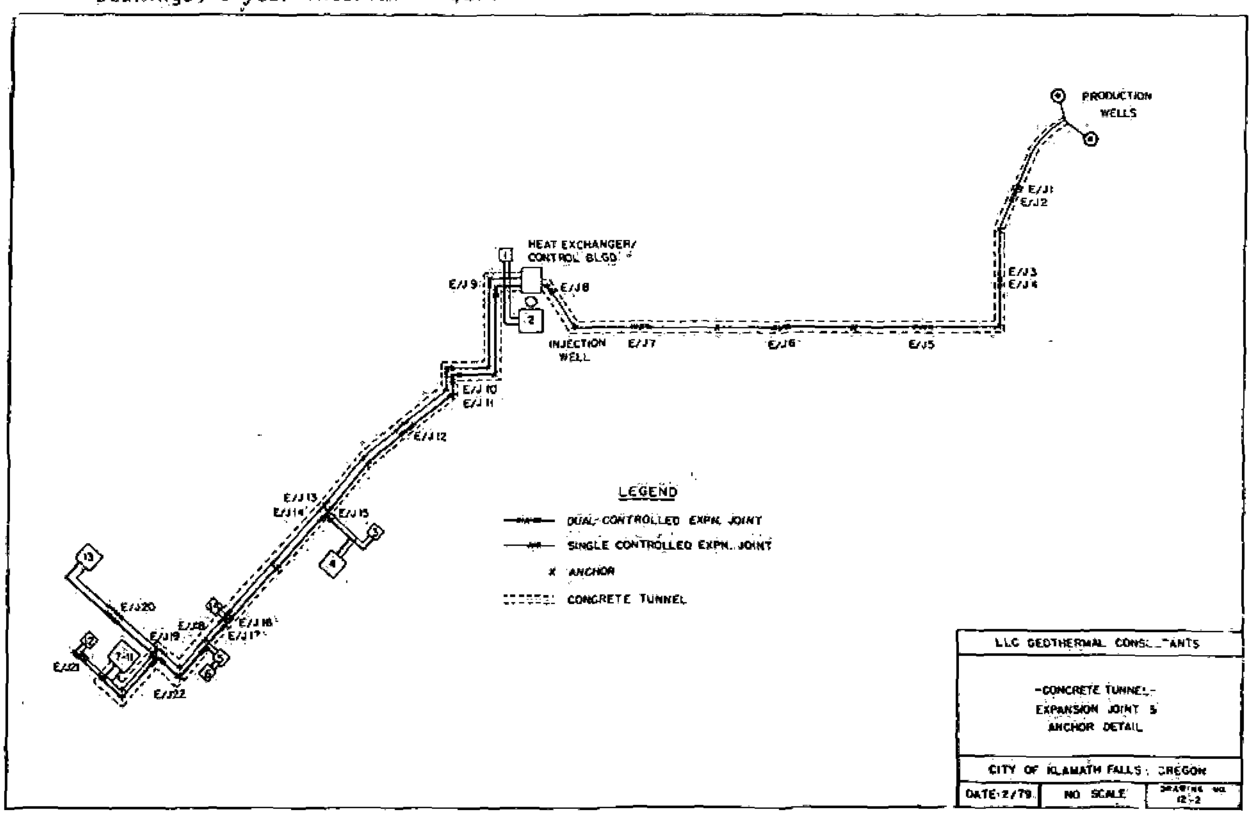

FIGURE 4. Distribution Network

The following four types of piping systems were investigated:

Steel Pipe in a Concrete Tunnel (350°F Maximum Temperature). This type system is essentially a poured-in-place or precast reinforced concrete tunnel with removable concrete lids, which can be used for sidewalks, and removed to provide access for maintenance or installation of future pipes. The insulated carrier pipes are supported on steel rollers which are imbedded into the concrete tunnel. After installation, pipes are insulated with preformed fiberglass or rockwool. Allowance for expansion is by a bellows type expansion joint which is dual acting with an anchor between the bellows. The pipe is also free to expand at elbows with the use of guides. The concrete tunnel is placed on a gravel bed which contains a drain tile. The concrete tunnel is extremely durable, can be used for other utilities, and can be constructed by local contractors. Its main drawback is that it is expensive to construct compared to other type systems.

Concrete tunnel type piping is popular in European district heating systems and the Icelanders who usually use it for pipes larger than 10 inches in diameter, claim there is no better system.

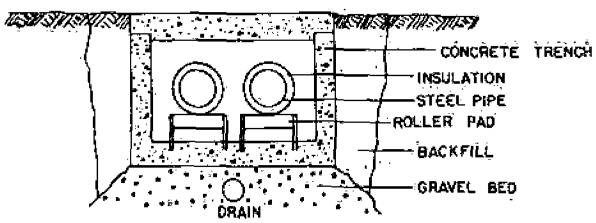

FIGURE 5. Steel Pipe in a Concrete Tunnel

Steel Pipe in a Protective Covering (250°F Maximum Temperature). A single carrier pipe is enclosed in polyurethan insulation with a tight jacket of glassfiber reinforced plastic (FRP) of PVC. Joints of pipe are welded, insulated, and sealed with a joint kit and placed in a bed of sand in the

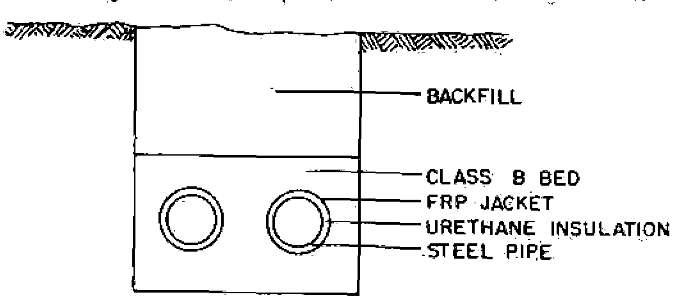

FIGURE 6. Steel Pipe in a Protective Covering

trench. Manholes are constructed between anchors to house the expansion bellows and insulation around elbows is oversized to allow expansion. There are usually eight segments (320 ft) between an anchor and manhole. The manholes are constructed of reinforced concrete with drain provisions. This type system is very durable, resistant to external corrosion, however, susceptible to exterior corrosion. Although costs per lineal foot are comparable to non-metallic pipe, the added cost for manholes and expansion bellows makes it more expensive.

Fiberglass Reinforced Plastic Pipe (FRP) (210°F Maximum Temperature). A single FRP carrier pipe is enclosed in polyurethane insulation with a tight jacket of FRP or PVC. The pipe is a filament wound fiberglass with either epoxy or polyester resin plastic. The epoxy type can handle temperatures up to 350°F and has a low coefficient of roughness, C = 140. The pipe usually does not have any expansion joints, as the coefficient of linear expansion is 8.5×10^{-6} per °F as compared to 12×10^{-6} per °F for steel pipe. Elbows are located in poured concrete thrust blocks in order to hold the pipe in position and allow expansion in straight lengths. A polyester resin pipe of this type failed recently when it pulled apart, mainly at joints, when the pipe cooled. There is available on the market a slip-ring type joint that may be more desirable than the epoxy joined pipes. A check on the filament wound epoxy resin pipe installation, carrying 353°F water, proved satisfactory, however, flashing must never occur in FRP pipe. In areas where flashing conditions are expected, steel pipe sections must be used. The main advantages of this type piping system is its resistance to corrosion and low roughness coefficient.

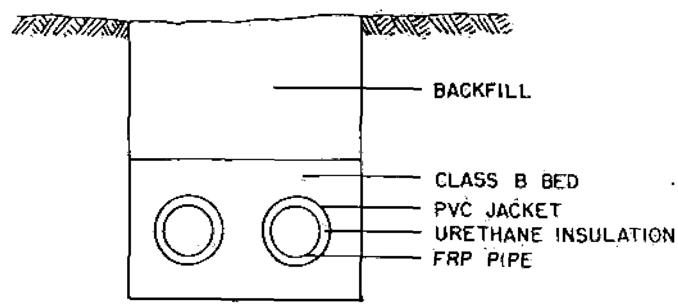

FIGURE 7. Insulated FRP Pipe

Asbestos Cement Pipe (AC) (200°F Maximum Temperature). This type pipe has an epoxy lined AC carrier pipe, polyurethane insulation and an AC

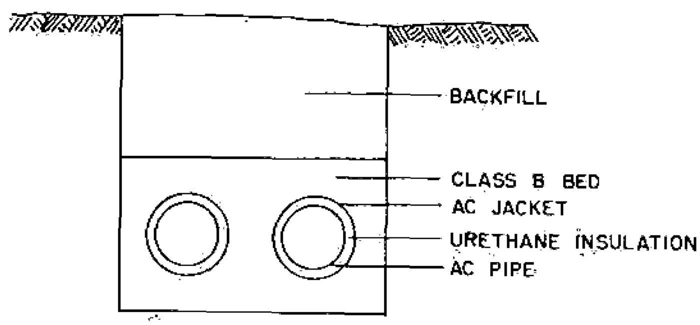

FIGURE 8. Insulated AC Pipe

jacket. The temperature limit of this pipe is 200°F and the only way it can be used for the 200° to 220°F supply line is for the line to be broken (flashed) to atmosphere at the well head. Another alternative would be to use uninsulated AC with special heat resistant slip-rings for the 16-in supply line in the winter when the demand is large and velocities are high and a second 8-inch line in the summer months when the load was low. AC pipe may fracture when a temperature change (cooling) occurs. Expansion occurs in the slip-ring type joints and proper bedding and backfill are important. The main advantage of AC pipe is its resistance to corrosion, however, its cost per lineal foot (16 in) is the largest of any of the four systems investigated. The slip-ring joints are quick to assemble need for expensive expansion joints and manholes.

Piping Recommendation

The distribution network is the most expensive part of the district heating scheme. Therefore, it is very important to build it in the very best way possible.

Steel pipe in a concrete duct was selected for the 4060-ft primary supply pipe carrying 220°F geothermal water. The main advantages of this type system are:

1. access to the pipe for future taps.
2. access for maintenance and repair.
3. other utilities may be readily installed initially or at a future date in the same trench.
4. better assurance that ground water will not come in contact with the pipe.
5. the lid may be used for a sidewalk, with heat radiating from the pipe providing snow removal.
6. the duct may be oversized so that pipes in the future may be added as the district heating system grows. This is especially true in the case of the main supply line.

Disadvantages of the concrete duct system are:

1. it must be watertight to the highest extent possible, because its main task is to protect the steel pipes against external water.
2. high demand for reinforcement of concrete at traffic crossings.
3. in the case of a high ground water level, there is a risk that the duct will rise.
4. costs are high compared to the other systems.

Direct-buried FRP pipe was selected for the secondary closed loop, because of the lower supply temperature (180°F) and it is anticipated there will not be a need for future pipe installations along its route. This type pipe has advantages of having a low-friction factor, corrosion resistance, and does not require special equipment for expansion allowances.

Concerning the choice of a piping system, it must be emphasized that regardless of which of the systems chosen, the quality of construction and execution of the work are of the utmost importance for the supply in a long lifetime.

Controls

The quantities which must be controlled in the district heating network are primarily the fluid flow rate from the production wells to the centralized heat exchanger station and flow rate and temperature in the closed secondary loop supply to subscribers.

The flow in the primary geothermal fluid supply line is regulated by pneumatic butterfly valves (V-1 and V-2) located on the reject side of the heat exchangers which are controlled by outside air temperature (T_1) temperature (T_2) via Receiver-Controller #1 (Figure 9). Closing of control valves V-1 and/or V-2 results in increased pressure in the primary supply line which in turn is relayed to a pressure control regulator located at the production pump, reducing the pumping rate of the variable speed/fluid drive deep well vertical turbine pumps (TP1 and TP2). A reduction in pressure due to opening of valves resulting from a drop in outside air temperature (T_1) and geothermal return fluid temperature (T_2) causes the pressure controller to increase the pumping rate.

The flow in the secondary closed loop is regulated by the temperature and pressure difference between the supply and return lines. The most remote point in the system, at the County Courthouse complex, will be the critical location. In order to provide sufficient heat to subscribers, the pipe temperature loss to this point will be kept to a minimum of .3°F and the pressure to a minimum of 60 psi.

The supply temperature in the closed secondary loop is controlled on the basis of measured outside air temperature (T_1) and heating water return temperature (T_3). Receiver-Controller #2 will activate pneumatic globe valves V-3 and/or V-4 to open when outside air temperature (T_1) and heating water return temperature (T_3) drop. The result is a reduction of pressure in the heating water supply (P_1) and return (P_2) line, causing an increased pumping rate of the variable speed/fluid drive vertical turbine circulation pumps (CP1 and CP2).

Receiver-Controller #3 regulates the pressure in the closed-loop network through the balancing globe valve V-5 when sensing supply pressure (P_1) and return pressure (P_2). This assures that design pressures are maintained to subscribers.

Failures in pumps of pipelines and unusual flow rates, temperatures, or pressures will be monitored by a master flow controller (MC1). This includes the pressures in the pipeline as well as the expansion tank. The master flow controller, under these circumstances, will shut down the pumps in either pipe system and sound an alarm in the heat exchanger/control building. This alarm will be monitored in the Fire Station.

Examples of possible critical situations would be a "fully open" indication from a control valve under low heat load conditions; a reduction in pipeline pressure under high pumping rates (due to

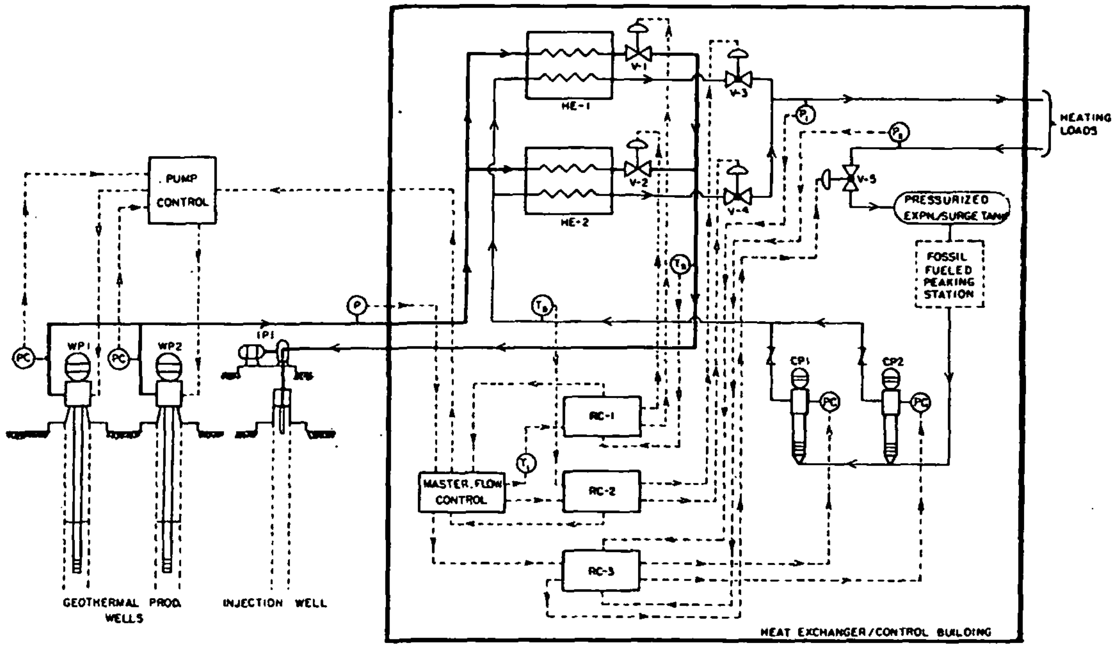

FIGURE 9. Distribution Network Control

a rupture); or a drop in supply temperature (caused by a closed valve or stopped pump).

When the project is expanded to Phase III, a computerized control center will replace all chart recorders and provide BTU calculations for remote building as well as temperature and pressure of heating water supplied and used.

PART IV
LIFE CYCLE AND COST ANALYSIS

For the Klamath Falls district, heating of the central business core Phase II (the eleven-block district) was evaluated due to the fact that this phase will be completed in the very near future.

Cost benefit analysis was based on the annual heat load using geothermal energy as opposed to natural gas. The economic analysis was based on the following assumptions:

1) The economic inflation rate was forecast at 7%. As of this writing, 9% would be more accurate.

2) Inflation rates for conventional energy were obtained from the Oregon Department of Energy as follows:

a. natural gas--5.2% above the economic inflation rate through 1986 and 1.5% above the economic inflation rate thereafter.
b. electric power--2.5% above the economic inflation rate through 1986 and 1.5% above the economic inflation rate thereafter.

3) Cost of capital 6.5% as indicated by the City of Klamath Falls.

4) Current cost of natural gas $0.34394 per therm (rate paid in February, 1979).

5) 85% efficiency for natural gas.

These inflation rates have proven to be very conservative. During the past 3 years, the City has experienced a 26.5% per year increase in the cost of natural gas as compared to 12.2% in our evaluation. The fact that actual inflation rates exceed those used in the study further supports the argument for the geothermal system.

Tables I and II of the report show capital investment for 16 combinations of primary and secondary piping systems.

Life cycle costs were calculated on these piping systems for a 10-year period and appear in Tables III and IV.

A fifty-year life cycle cost analysis was completed on four piping systems and the results are illustrated graphically on the chart following.

Although steel pipe installed in concrete tunnels requires the highest capital investment, the annual maintenance costs were estimated to be considerably lower. Such a system provides easy access, room for future expansion at minimal cost, and reduces maintenance time and cost particularly in conjested business districts.

TABLE I.

Piping Network Costs
(in $)

Primary Supply Pipeline

Secondary Supply Pipeline		16" Steel in Tunnel (I)	8" Steel in Tunnel (II)	16" Steel in Tunnel (III)	8" Steel Buried (IV)
Steel in Tunnel	(a)	726,463	506,175	471,564	282,154
		637,060	637,060	637,060	637,060
		1,363,523	1,143,235	1,108,624	919,214
Steel Buried	(b)	726,463	506,175	471,564	282,154
		490,072	490,072	490,072	490,072
		1,216,535	996,247	961,636	772,226
FRP Buried	(c)	726,463	506,175	471,564	282,154
		329,118	329,118	329,118	329,118
		1,055,581	835,293	800,682	611,272
AC Buried	(d)	726,463	506,175	471,564	282,154
		329,129	329,129	329,129	329,129
		1,055,592	835,304	800,693	611,283

Key
000,000 = primary line cost
000,000 = secondary line cost
0,000,000 = Total pipeline cost
Cost figures based on January 1979 estimates.

TABLE II.

Total Project Cost
(in $)

Primary Supply Pipeline

Secondary Supply Pipeline		16" Steel in Tunnel (I)	8" Steel in Tunnel (II)	16" Steel Buried (III)	8" Steel Buried (IV)
Steel in Tunnel	(a)	1,730,801	1,510,513	1,475,902	1,286,492
		1,990,421	1,737,090	1,697,297	1,479,466
Steel Buried	(b)	1,583,813	1,363,525	1,328,914	1,139,504
		1,821,385	1,568,054	1,528,251	1,310,430
FRP Buried	(c)	1,422,859	1,202,571	1,167,960	978,550
		1,636,288	1,382,957	1,343,154	1,125,332
AC Buried	(d)	1,422,870	1,202,582	1,167,971	978,561
		1,636,300	1,382,969	1,343,167	1,125,345

Key
000,000 = Basic cost
000,000 = 15% Engineering & Inflation costs added

Note: Basic cost = Well costs ($169,772) + pipe costs (Table 7) + heat exchanger costs ($197,506)

TABLE III.

10 YEAR COST COMPARISON OF PRIMARY PIPING SYSTEM
USING CAPITAL RECOVERY AND MAINTENANCE COSTS

16" Steel in Tunnel	
yr 1	$16,168
2	17,299
3	18,510
4	19,806
5	21,193
6	22,676
7	24,264
8	25,962
9	27,779
10	29,724
Total Cost	223,388
Present Value	143,621
Annual Equiv. Cost	$ 21,404

	16" Buried Steel	8" Steel in Tunnel	8" Steel Buried
yr 1	$17,626	$11,265	$10,580
2	18,860	12,054	11,320
3	20,180	12,897	12,113
4	21,593	13,800	12,961
5	23,104	14,766	13,868
6	24,721	15,800	14,839
7	26,452	16,906	15,878
8	28,304	18,089	16,989
9	30,285	19,356	18,178
10	32,405	20,711	19,451
Total Cost	243,534	155,649	146,181
Present Value	156,574	100,069	93,984
Annual Equiv. Cost	$ 23,334	$ 14,913	$ 14,006

	Annual Cost of Buried Steel over Steel in Tunnel	Annual Cost of 16" Steel over 8" Steel	Annual Cost of 8" Steel in Tunnel over 8" Steel Buried
yr 1	$ 1,458	$ 4,903	$ 685
2	1,560	5,246	733
3	1,670	5,613	785
4	1,786	6,006	840
5	1,912	6,426	898
6	2,045	6,877	961
7	2,189	7,358	1,028
8	2,341	7,873	1,100
9	2,505	8,424	1,177
10	2,680	9,014	1,260
Total Cost	20,146	67,740	9,467
Present Value	12,953	43,552	6,085
Annual Equiv. Cost	$ 1,930	$ 6,491	$ 907

Forecasting life cycle costs over a 50-year period leaves much to be desired in regard to accuracy. Data on maintenance costs of pipelines for 50 years is not available due to the lack of experience with such piping systems.

Table VI shows a total cost summary for the project. Table V concludes the study by comparing the annual cost of the geothermal system with the annual cost of continuing to use natural gas over the next 20 years. Using a 6.5% cost of capital, the present value of annual savings exceeds $7,000,000 in 20 years. With a capital investment of $2,000,000, payback would occur in less than 7 years.

The average annual equivalent cost per therm for the 20-year period is $0.29 for geothermal as compared to $0.94 for natural gas.

TABLE IV.

10 YEAR COST COMPARISON OF SECONDARY PIPING SYSTEM USING
CAPITAL RECOVERY AND MAINTENANCE COSTS

Steel in Tunnel	
yr 1	14,178
yr 2	15,170
yr 3	16,232
yr 4	17,369
yr 5	18,585
yr 6	19,885
yr 7	21,277
yr 8	22,767
yr 9	24,361
yr 10	26,066
Total Cost	$195,894
Present Value	125,946
Annual Equiv. Cost	18,770

Buried Steel		FRP		Asbestos Cement	
yr 1	18,318	yr 1	18,734	yr 1	19,862
yr 2	19,600	yr 2	20,046	yr 2	21,252
yr 3	20,972	yr 3	21,449	yr 3	22,740
yr 4	22,440	yr 4	22,950	yr 4	24,332
yr 5	24,011	yr 5	24,557	yr 5	26,035
yr 6	25,692	yr 6	26,276	yr 6	27,857
yr 7	27,490	yr 7	28,115	yr 7	29,807
yr 8	29,415	yr 8	30,083	yr 8	31,894
yr 9	31,474	yr 9	32,189	yr 9	34,127
yr 10	33,677	yr 10	34,443	yr 10	36,515
Total Cost	$253,093	Total Cost	$256,847	Total Cost	$274,425
Present Value	162,718	Present Value	166,417	Present Value	176,434
Annual Equiv. Cost	24,250	Annual Equiv. Cost	24,801	Annual Equiv. Cost	26,294

Annual Cost of Buried Steel Over Steel in Tunnel		Annual Cost of FRP Over Steel in Tunnel		Annual Cost of Asbestos Cement Over Steel in Tunnel	
yr 1	4,140	yr 1	4,557	yr 1	5,684
yr 2	4,430	yr 2	4,875	yr 2	6,082
yr 3	4,740	yr 3	5,216	yr 3	6,507
yr 4	5,072	yr 4	5,582	yr 4	6,963
yr 5	5,426	yr 5	5,972	yr 5	7,450
yr 6	5,806	yr 6	6,390	yr 6	7,972
yr 7	6,213	yr 7	6,838	yr 7	8,530
yr 8	6,648	yr 8	7,317	yr 8	9,127
yr 9	7,113	yr 9	7,829	yr 9	9,766
yr 10	7,611	yr 10	8,377	yr 10	10,450
Total Cost	$ 57,199	Total Cost	$62,953	Total Cost	$78,521
Present Value	36,772	Present Value	40,471	Present Value	50,488
Annual Equiv. Cost	5,480	Annual Equiv. Cost	6,031	Annual Equiv. Cost	7,524

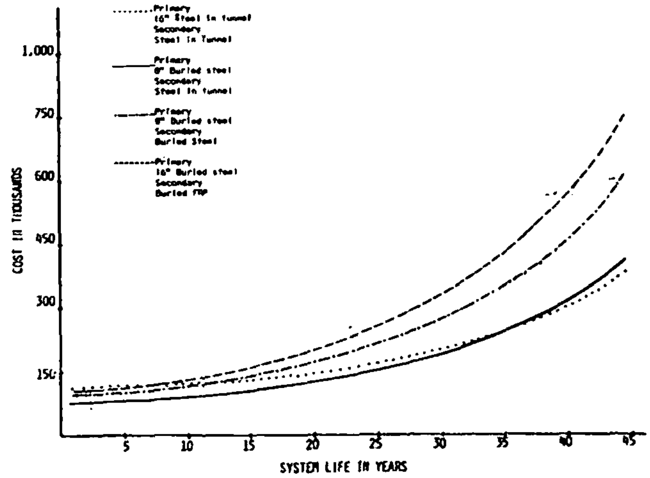

FIGURE 10.

TABLE V.

20 Year Cost Comparison of Natural Gas vs. Geothermal For An 11-Block Area
With A Heat Load of 6×10^5 Therms/Year

Year	NATURAL GAS Present Cost 242,822.00	GEOTHERMAL ELECTRICAL Present Cost 8,492.00	GEOTHERMAL OPERATION AND MAINT. Present Cost 1,090.00	Annual Savings	Present Worth 6.5%	Present Worth 8%
1	272,446.28	9,213.82	1,166.30	262,066.16	246,071.52	242,653.86
2	305,684.73	9,996.99	1,247.94	294,439.79	259,595.58	252,434.67
3	342,978.27	10,846.74	1,335.30	330.796.23	273,849.36	262,596.71
4	384,821.62	11,768.71	1,428.77	371,524.14	288,872.02	273,154.83
5	431,769.85	12,769.05	1,528.78	417,472.02	304,704.83	284,124.44
6	484,445.78	13,854.42	1,635.80	468,955.56	321,391.24	295,521.55
7	543,548.16	15,032.05	1,750.30	526,765.81	338,977.07	307,362.79
8	589,749.75	16,321.80	1,872.82	571,555.13	345,351.44	308,793.45
9	639,878.48	17,722.21	2,003.92	620,142.35	351,845.44	310,230.57
10	694,268.15	19,242.77	2,144.19	672,881.19	358,461.33	311,674.18
11	753,280.95	20,893.80	2,294.29	730,092.86	365,201.38	313,124.31
12	817,309.83	22,686.49	2,454.89	792,168.45	372,067.94	314,580.99
13	886,781.16	24,632.99	2,626.73	859,521.44	379,063.36	316,044.25
14	962,157.56	26,746.50	2,810.60	932,600.46	386.190.08	317,514.12
15	1,043,940.95	29,041.35	3,007.34	1,011,892.26	393,450.55	318,990.64
16	1,132,675.94	31,533.10	3,217.86	1,097,924.98	400,847.28	320,473.83
17	1,228,953.39	34,238.64	3,443.11	1,191,271.64	408,382.83	321,963.74
18	1,333,414.43	37,176.32	3,684.13	1,292,553.99	415,049.80	323,460.38
19	1,446,754.66	40,366.05	3,942.02	1,402,446.60	423,880.84	324,963.80
20	1,569,728.80	43,829.45	4,217.96	1,521,681.39	431,848.66	326,474.02
				TOTAL 15,368,862.44	TOTAL 7,066,112.54	TOTAL 6,046,137.14

TABLE VI.

TOTAL COST SUMMARY

Case IIa
(8", 6" Steel Pipeline in Concrete Tunnel)

Item	Cost
A. Wells and Well Head Equipment:	
1. Production well (2) @ $38,898.	$ 77,796
2. Production well pumps (2) @ $41,988.	83,976
3. Well head buildings (2) @ $3,500	7,000
4. Power hook-up in buildings (2) @ $500.	1,000
Subtotal:	169,772
B. Distribution Piping Network:	
5. Primary supply pipeline (8" steel in concrete tunnel).	506,175
6. Secondary supply pipeline (8" & 6" steel in concrete tunnel, 3" steel buried	637,060
Subtotal:	1,143,235
C. Heat Exchanger Building:	
7. Plate heat exchangers (2) @ $14,000	28,000
8. Control system, wiring, etc. (basic).	44,537
9. Circulation pump (2) @ $13,691.	27,382
10. *Expansion/surge tank*.	5,000
11. Building including installation of equipment. . .	90,000
12. Injection well (museum)	----
13. Injection well pump	2,587
Subtotal:	197,506
Total Equipment and Installation Costs:	$1,510,513
D. Overhead Costs:	
Engineering @ 10%	151,051
Contingency (inflation @ 5% for 6 mos.)	75,526
Total Cost:	$1,737,090

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*This paper is the combined effort of four speakers.

ORE DEPOSITS AS EXPLORATION MODELS FOR GEOTHERMAL RESERVOIRS IN CARBONATE ROCKS IN THE EASTERN GREAT BASIN

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ABSTRACT

The geology of ore deposits in carbonate rocks in the eastern Great Basin gives insight into the geology of present day geothermal reservoirs in carbonate rocks. The Cove Fort KGRA is geologically similar to the Tintic mining district and the geology of the Carlin mine area suggests a carbonate reservoir at the Beowawe KGRA. Carbonate geothermal reservoirs are unique in that fluids moving along fractures may leach calcite and, less often, dolomite and thereby increase porosity and permeability.

INTRODUCTION

Geologists generally recognize that most of the ore deposits of the western U.S. are "fossilized" geothermal systems. However, while some detailed attention has been given to the use of the "porphyry copper model" in geothermal exploration, little attention has been given to other types of ore deposits found in areas being explored for geothermal reservoirs. The purpose of this paper is to point out similarities between ore deposits and geothermal reservoirs in carbonate rocks in the eastern Great Basin and to indicate some of the unique problems and features of carbonate geothermal reservoirs. Two major mining districts and two major geothermal prospects are discussed below. Their comparison is made possible as a result of data from exploratory geothermal wells having been purchased by the Department of Energy and made available through the University of Utah Research Institute (UURI).

The only producing carbonate geothermal reservoirs at this time are the Lardarello and Mt. Amiata fields in Italy where production is from highly permeable, fractured carbonates capped by shaly flysch facies which have been thrust over the carbonates (Celati, et al., 1975). While space does not permit a detailed review of these fields here, a general comparison can be made with the geologic setting in the central Great Basin where siliceous rocks deposited in an early Paleozic eugeosyncline were thrust eastward in late Devonian time over limestones and dolomites which had been deposited in a miogeosyncline. The eastern limit of the prinicipal thrust, which is of significance in the discussion of the Beowawe KGRA, is shown in Figure 1.

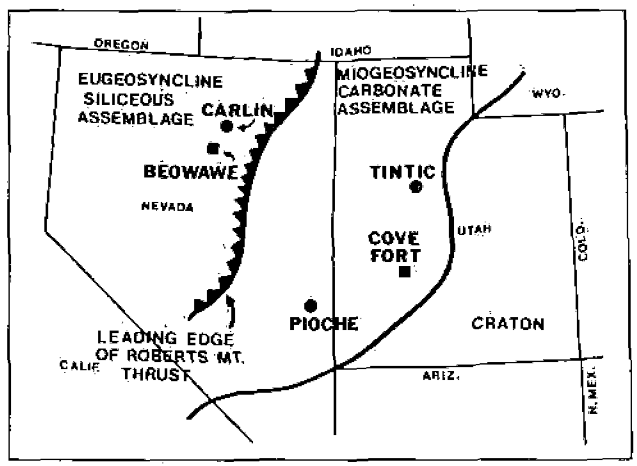

Fig. 1 – Index map with distribution of early Paleozoic facies.

THE TINTIC MINING DISTRICT

The Tintic mining district, located in central Utah (Figure 1), is one of the largest underground mining districts in the Basin and Range province and serves as a model for geothermal water movement in carbonate rocks. The district which includes the Main Tintic and East Tintic subdistricts, has been mined continuously from 1869 until the present; and as of 1976, had produced nearly 17 million tons of lead, zinc, silver, copper and gold ores valued at over $568 million at the time of production (Morris and Mogensen, 1978). The ore bodies of the district consist of both massive, irregular replacements and replacement veins in folded and faulted Paleozoic limestones and dolomites and open space fillings in narrower fissure veins in Paleozoic quartzites and Tertiary igneous rocks. The ore bodies are believed to have been deposited by hydrothermal fluids following a period of major volcanism in the East Tintic Mountains which took place in the Oligocene, ending about 31.5 million years ago. The volcanic activity was centered a few miles south of the present mining district and in the early stages produced a collapsed caldera accompanied by a thick welded tuff. A later composite cone which filled and covered the caldera may have attained a height of 13,000 ft - 16,000 ft. The present day topography of the carbonate rocks may be very similar to that which was for a long period of time buried and preserved by the volcanic rocks (Morris, 1978; personal communication). The temperature of the ore depositing fluids is believed to have been slightly more than 200°C.

The major geologic structures at Tintic include folds and faults related to the overthrusting of the Sevier orogeny, normal faults which preceded the Oligocene volcanism, mineralized faults and fissures associated with the volcanism, and postmineral Basin-and-Range normal faults (Morris and Mogensen, 1978). The ore bodies in the Main Tintic subdistrict occur along five north-northeast trending zones. Within these zones, individual ore bodies are localized at the intersections of obscure fissures and bedding plane faults with other premineral structures and with favorable carbonate beds. Localization of ore at fault intersections is related to increased permeability and the existence of open spaces in fault breccia while the localization in favorable beds is due to chemical reactions involving the hydrothermal fluid and host rock.

Figure 2 is a longitudinal section along one of the major ore zones and may be considered as the fossilized imprint of a geothermal reservoir. The general direction of fluid movement is believed to have been from south to north, away from the center of igneous activity. This is supported by horizontal zonation of both ore and gangue minerals. However, Morris (1968) states that there is evidence that hot waters rose from many centers within the district and extended great distances along bedding plane faults and other minor features. Possible flow paths for the hot waters (author's interpretation) are indicated on the section. Most of the ore bodies shown on the section, horizontal as well as vertical, are pipe shaped.

Five distinct stages of wall rock alteration have been described at Tintic (Lovering, 1949). These are: (1) an Early-Barren stage in which limestone was altered to hydrothermal dolomite; (2) a Mid-Barren, argillic stage in which igneous rocks were altered to clay minerals and the remaining calcite and dolomite cement was leached from syngenetic and hydrothermal dolomites creating "sands" of dolomite grains; (3) a Late-Barren stage chiefly resulting in the silicification of carbonate rocks; (4) an Early-Productive stage characterized by local deposition of sericite, orthoclase, pyrite, barite, jasperoid, and quartz; (5) a Productive stage continuing from the Early-Productive stage but including deposition of ore minerals.

The most significant of these stages, from the standpoint of geothermal exploration are the first two which locally created zones of high porosity and permeability at the expense of converting formerly competent limestone and dolomite to sand. The formation of sanded dolomites at Tintic is believed to have taken place above and below the water table. Above the water table acids were formed by the oxidation of fumarolic gasses chiefly hydrogen sulfide. Below the water table leaching was caused by primary

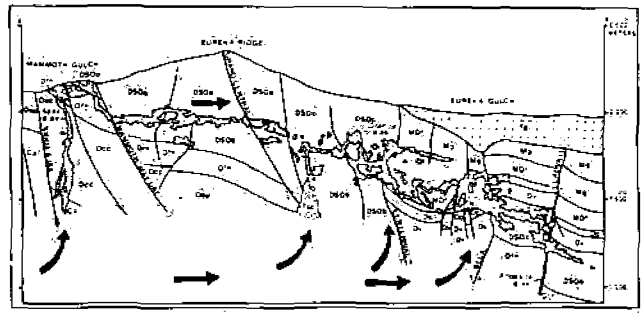

Fig. 2 - Longitudinal section of the Mammoth-Chief ore zone showing interpretted directions of hydrothermal fluid flow. All Paleozoic formations shown are chiefly carbonates. (Modified from Morris, 1978. Used by permission of author.)

sulfuric, hydrochloric, hydrofluoric and other mineral acids. At Cove Fort sanded dolomites have so far been found only above the water table where they contributed to the shallow abandonment of an exploratory well.

From 1918 to 1927, and again from 1942 to 1955, the Chief No. 1 mine in the Main Tintic subdistrict was dewatered by pumping water into one of two natural caverns in limestone found by underground workings. Most of the water was pumped into a cavern located in the adjacent Gemini mine which Morris (1968) describes as an irregular pipelike opening about 35 ft in diameter plunging at 45° into a rectangular chamber about 200 ft below. From 1951 to 1955, an average to 6,600 gpm with peak surges of 10,000 gpm were pumped into the cavern. Various tests with fluorescein dye failed to find any trace of the pumped water either reentering the mine or entering water wells in the adjoining valley. Exploration of the room beneath the cavern revealed no openings large enough for a man to enter. When pumping operations ceased in 1955, water levels in the mine rose 170 ft in 25 hours.

Until July, 1978, large volumes of water at a temperature of 147°F were pumped from the Burgin shaft in the East Tintic subdistrict. These waters were chemically similar to hot springs waters from other locations in Utah and are probably not related to the older hydrothermal ore depositing fluids at Tintic.

COVE FORT

The Cove Fort KGRA, located 100 miles south of Tintic, is the only geothermal prospect in the U.S. for which deep well data on potential geothermal reservoirs in carbonate rocks are publicly available. Well documented data are currently available on three wells drilled by Union Oil Co. to depths of 1,051 ft, 5,221 ft and 7,735 ft; and data should be available in several months on a fourth well reportedly planned for this summer. Geophysical and shallow temperature gradient data are also available.

The geology of Cove Fort is very similar to that of Tintic. Paleozoic strata, chiefly carbonates, were thrust eastward over Mesozoic strata during the Sevier orogeny and buried beneath andesites from a nearby major volcanic center during the mid-Tertiary. A major thrust fault, the Pavant thrust, is exposed in the Pavant Range north of the KGRA. Paleozoic rocks in the upper plate were overturned during thrusting and can be seen exposed above outcrops of Jurassic Navajo sandstone (Crosby, 1959). Volcanism was renewed during the Pliocene and Pleistocene with several rhyolitic eruptions to the west in the Mineral Range and to the northwest in the southern Sevier Desert. The youngest volcanic feature in the area is a Quaternary basaltic cinder cone three miles west of the area currently being explored by Union.

A large, shallow thermal gradient anomaly is known to exist at Cove Fort. Shallow holes drilled by Union and now available from UURI have temperatures of up to 119°F at a depth of 300 ft. The anomaly appears to extend several miles to the north, where Hunt Energy drilled a geothermal test in 1978, and to the northwest in Paleozoic strata in the upper plate of the Pavant thrust. There have been reports that deeper temperature holes drilled on the extensions of the anomaly become isothermal when the water table is encountered at a depth of about 1,000 ft. Apparently a large horizontal flow of warm water is taking place at the water table in the carbonates.

Exploratory drilling at Cove Fort has found large-scale, secondary porosity and permeability in dolomites to a depth of 5,221 ft. Porosity is present in the form of fractures, breccia zones, zones of dolomite sanding and caverns. There may not yet be enough control to establish the continuity of major fractures or the role of folding in the creation of secondary porosity. Water appears to descend rapidly to depth along some structures, be heated, and then return to shallow depths along other structures where it spreads out laterally in permeable carbonates creating a large thermal gradient anomaly. This is basically the model proposed by Morris (1968) for Tintic. The structures associated with water movement at Cove Fort appear to be similar to those found at Tintic and it is likely that additional drilling and subsurface geology at Cove Fort will further add to the similarities between the two areas. The maximum temperature found at Cove Fort to date is about 350°F in a contact metamorphic marble near the bottom of the 7,735 ft well. A weak flow of 43,000 lbs/hr was produced from this zone during a flow test. The temperatures and pressures associated with ore deposition at Tintic are believed to have been greater than those yet found at Cove Fort but were probably similar to those found in other geothermal fields. The problem at Cove Fort appears to be that the carbonates are too porous and permeable, allowing rapid convection and an early cooling of the deep reservoir.

BEOWAWE AND CARLIN

All of the productive wells drilled at the Beowawe KGRA prior to 1974 were completed in an area of fault intersections near the Beowawe Geysers. These wells bottomed in either Tertiary basalts and andesites or underlying siliceous rocks of the Western Assemblage assigned to the Valmy Formation (Ordovician). In early 1974, Chevron drilled a 9,551 ft test which was designed to intersect the Roberts Mountains thrust near its intersection with the Malpais fault. This well, the Chevron-ATR-Ginn No.1-13, encountered two lost circulation zones near TD and was believed to have bottomed in a fault zone. A successful drill stem test was conducted. Geologists familiar with Beowawe have since debated whether the well intersected the Malpais fault zone or fractures related to the Roberts Mountains thrust. Some geologists have also questioned the reservoir potential of the underlying carbonates. The Carlin gold deposit, located 28 miles northeast of Beowawe, can be used as an exploration model to gain insight into the geology of the geothermal reservoir at Beowawe.

At Carlin, microscopic disseminations of gold are found in altered dolomites of the Roberts Mountains Formation (Silurian) in a window of the Roberts Mountains thrust. The Roberts Mountains is variously included in either the Eastern (Carbonate) Assemblage or the Transitional Assemblage in the geologic literature of the region. The geologic section above the ore deposit is reconstructed in Figure 3 which is based on a figure from Hausen and Kerr. Hausen and Kerr state that permeable horizons in the Roberts Mountains Formation provided a favorable environment for mineralizing solutions which had travelled up high angle normal faults and which subsequently leached calcite and deposited clay minerals and silica. Noble and Radtke (1978) state that the temperature of the fluids was 175°-225°C. Unconformably overlying the ore horizon in the mine area is the Popovich Formation (Devonian) which is a gray fossiliferous, medium bedded, locally dolomitic and silty limestone. Hydrothermal fluids invaded the Popovich along its base and along joints and fractures with subsequent calcite leaching and silica deposition, but the fabric of the rock was not totally altered and it is not an important ore host.

The Roberts Mountains thrust separates the Popovich from the overlying Vinini Formation of the Western Assemblage which was thrust eastward into its present position. Hausen and Kerr state that the thrust is about 10 to 20 ft thick and consists of bleached, brecciated and iron stained rock rubble. Carbonate minerals have been leached from the fault zone and from adjacent beds below the thrust. They conclude, and Noble concurs, that the thrust plane, although not an important structure in the localization of ore, did serve as a conduit for migrating solutions over a long period of time. The Vinini Formation at Carlin consists of siliceous shales. In other areas the Vinini usually contains minor quartzites and limestones. Following the main period of hydrothermal activity there occurred a later stage of acid leaching related to boiling of the fluids as at Cove Fort and Tintic (Noble and Radtke, 1978). This resulted in the leaching of dolomite as well as calcite, intense argillic alteration, and precipitation of quartz in the shallower part of the alteration zone. Details of the origin of the Carlin deposit are given in a paper by Radtke, Rye and Dickson (in press). The leaching of calcite at high temperatures has been

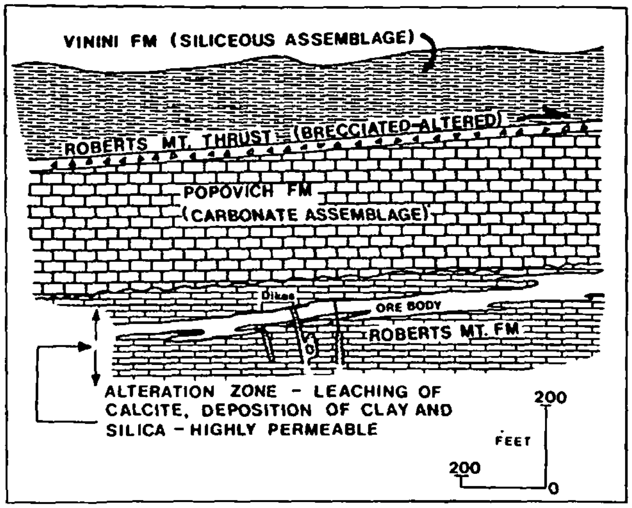

Fig. 3 - Reconstructed section at Carlin Mine. (Modified after Hausen and Kerr, 1968. Used by permission of AIME.)

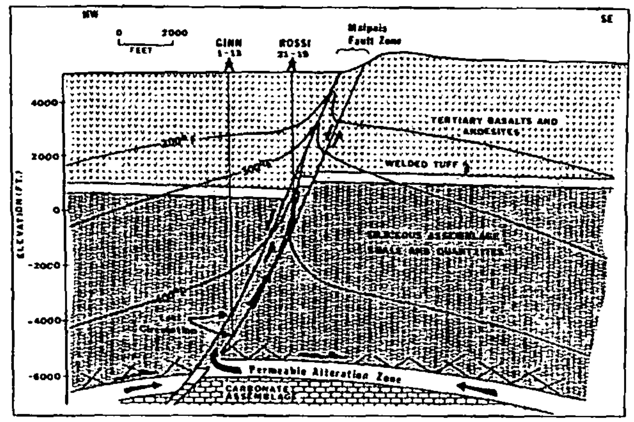

Fig. 4 - Interpretted geologic section at Beowawe.

observed in laboratory studies by Radtke (1979, personal communication) and Tewhey, et al. (1978).

Data from the Ginn 1-13 and a subsequent well, the Rossi 21-19, also drilled by Chevron, are given elsewhere in this volume by M. A. Lane. Figure 4 is a section based on these wells using the geology of the Carlin area as a guide. The temperature regime in Figure 4 is complicated by a flow component normal to the section in the direction of the Beowawe Geysers. Figure 4 shows a deep reservoir at Beowawe in the brecciated rock of the Roberts Mountains thrust and in altered limestones beneath the thrust from which calcite has been leached. For simplicity the potential reservoir zone in the carbonates has been placed immediately below the thrust. However, this zone may be separated from the thrust by beds of less permeable limestone as at Carlin. The depth of the thrust is based on a thickness of about 6,000 ft for the Western Assemblage which is consistent with sections drawn to the south in the Cortez Quadrangle by Gilluly and Masursky (1965). The shales and quartzites of the Valmy Formation may furnish an element which appears to be missing at Cove Fort - a cap rock for the carbonate reservoir.

CONCLUSIONS

Geothermal reservoirs in carbonate rocks differ from other geothermal reservoirs in that secondary porosity and permeability may be enhanced by leaching of the carbonates by the fluids at high as well as moderate temperatures. Unfortunately, carbonate reservoirs appear to lack the self sealing properties of reservoirs in siliceous rocks so that an external cap rock is required for a high temperature reservoir. In carbonate rocks, the greatest potential appears to be for large, moderate temperature reservoirs.

ACKNOWLEDGMENTS

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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 $SO_4^=$.

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

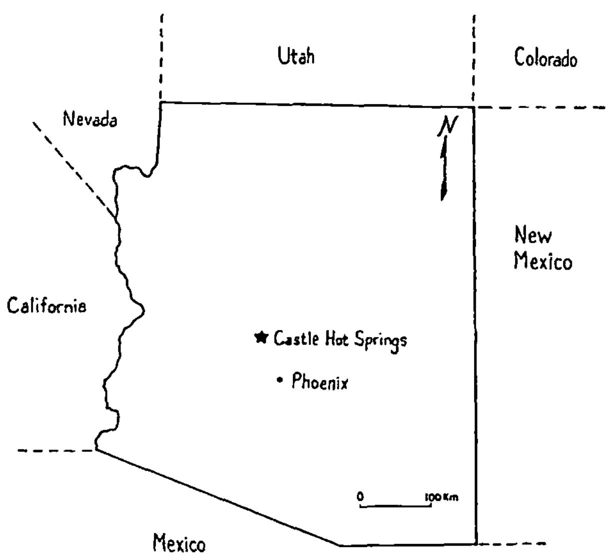

Figure 1. Location map of Castle Hot Springs, Arizona.

HYDROLOGIC SETTING

The thermal waters of Group I display similar physical and chemical characteristics. The waters of Group I are: Castle Hot Springs, Alkalai Spring, Henderson Ranch Spring and the Dodd 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,SW¼,SW¼	Henderson Ranch Spring, I T8N,R1W,33,NW¼	Alkalai Spring, I T8N,R1W,33,NW¼,SE¼	Mesquite Drip, I T7N,R1W,33,NW¼,SE¼	Dodd Well, I T8N,R1W,33,NW¼,NW¼
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^{+2}	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 (β=1/3)	117.08	124.68	127.50	128.64	127.05
Na-K-Ca (β=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,SW¼,SW¼	Menudo Spring, II T7N,R1W,14,NW¼,NW¼	Layton Seep, II T7N,R2W,1,NW¼,SW¼	Windmill Well, II T7N,R1W,3,SW¼,SW¼	Casa Rosa Spring III T7N,R1W,14,NE¼,SW¼	Dripping Spring III T7N,R1W,14,NW¼,NE¼
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^{+2}	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)	103.33	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	93.88	60.00	63.61	57.20	48.78
Na-K-Ca (β=1/3)	109.88	120.60	132.28	115.18	119.30	120.22
Na-K-Ca (β=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

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 Casa Rosa and Dripping 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	51.3	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	143
$SO_4^=$	212	230	211	211	206	200	196	189	206

Note: n.d. - not determined

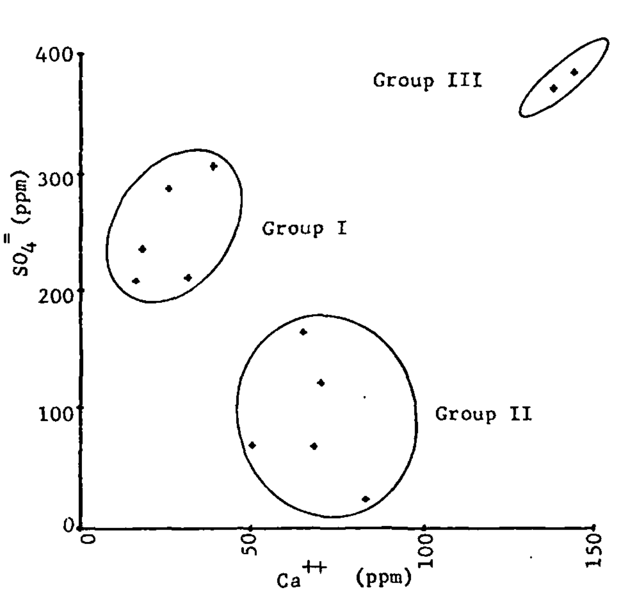

Figure 2. Water chemistry Ca^{++} versus $SO_4^=$.

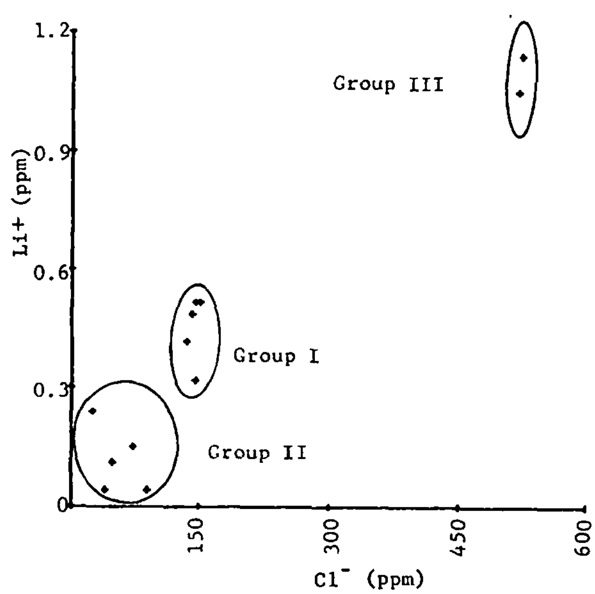

Figure 3. Water chemistry Cl^- versus Li^+.

GEOTHERMOMETRY

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 neglible. 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.

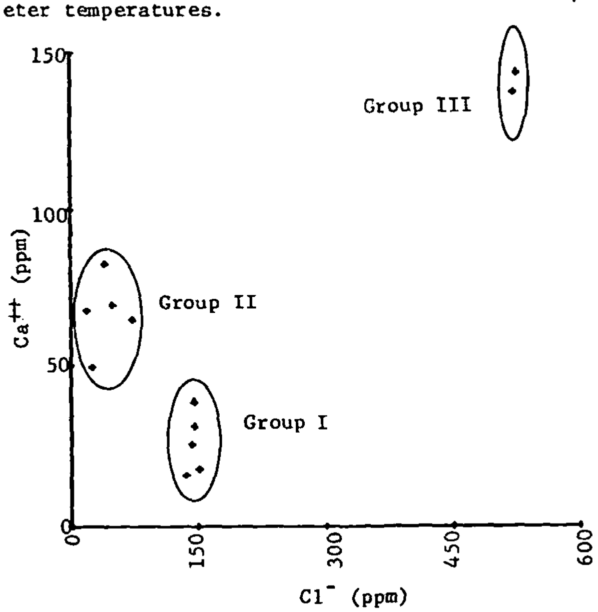

Figure 4. Water chemistry Cl^- versus Ca^{++}.

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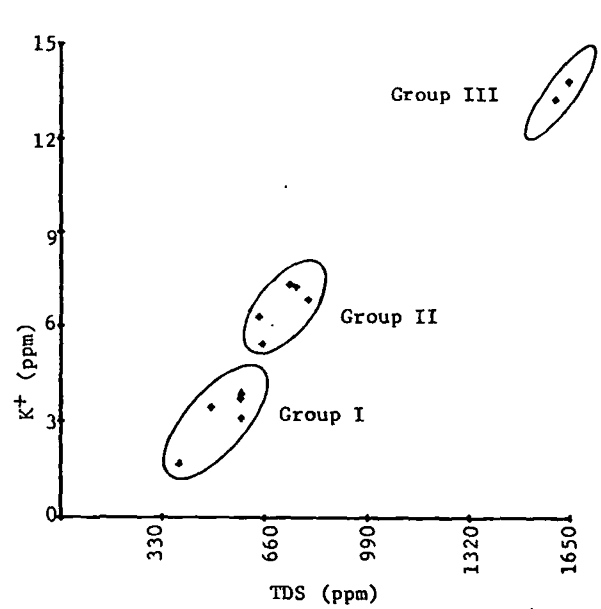

Figure 5. Total dissolved solids versus K^+.

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AN ANALYSIS OF GRAVITY AND GEODETIC CHANGES DUE TO RESERVOIR DEPLETION AT THE GEYSERS, NORTHERN CALIFORNIA

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ABSTRACT

In this paper gravity and geodetic data are combined with reservoir engineering studies to place upper and lower bounds on the volume and pore fluid mass changes within the depleted portion of the steam reservoir at The Geysers. We combined the gravity and temperature data to constrain the changes in pore fluid mass distribution due to fluid depletion, and thus limited the drainage volume to lie between 15 and 25 cubic km. We then modeled the surface geodetic data to determine values of strain between 3. and 8. X 10^{-5} for these drainage volumes. We determined that this strain could be induced either mechanically or thermally, and there is presently no way of distinguishing thermal from mechanical strain.

Since 1974, the average production rate at The Geysers steam field in Northern California (Figure 1) has been nearly 90 million kg of steam per day (Lippman and others, 1977). This large fluid withdrawal rate has caused changes in mass and volumetric strain within the depleted reservoir volume. From 1973 to 1977, time changes in pore pressure, surface strain (Lofgren, 1979), and gravity (Isherwood, 1977) occurred, while the reservoir temperature did not measurably change.

In this paper, the gravity and geodetic data from 1974 to 1977 are combined with reservoir engineering results (Weres, 1977) to determine the pore fluid deficit and strain within the drainage volume. Previously, (Isherwood, 1977; Hunt, 1977) it has been demonstrated that decreases in observed gravity with time reflect mass redistribution and deficits within some depletion volume. By comparing the total mass deficit measured from the gravity flux (found be integrating the gradient of the potential over some bounding surface) with the net mass produced, the recharge was estimated (Isherwood, 1977). Here we combine the gravity changes with well data to constrain the drainage volume between 15 and 25 cubic km. We then analyzed the geodetic data to determine the reservoir strain within the possible range of reservoir volumes. These concepts are summarized in Table 1.

At The Geysers, mapped values of maxima in subsidence, gravity change, pore pressure decline overlap. Figure 2 compares the decreases in observed gravity and pore pressure decay due to steam production. Isherwoods' (1977) previous analysis of this data determined that (1) the gravity changes were too large to be due solely to a deep water table below the producing zone penetrated by the wells, and (2) the gravity flux implied a mass deficit equal to the met mass produced, suggesting negligible recharge.

The lack of a measurable temperature change (plus or minus 3 degrees Celsius) limits the amount of water which has flashed to steam during production to less than 0.5% of the bulk rock volume. Yet reservoir engineering data imply that water flashes to steam to supply the fluid produced by the wells (Weres, 1977). Since the heat energy required in the phase transition from water to steam must come from the rock matrix, the lack of a measurable temperature change limits the change in reservoir liquid content due to steam production to that allowed by the error in the temperature measurements (Weres, 1977; Denlinger, 1979). The steam could come from a deep water table in this case, but this is inconsistent with the magnitude of the time changes in the gravity mentioned above.

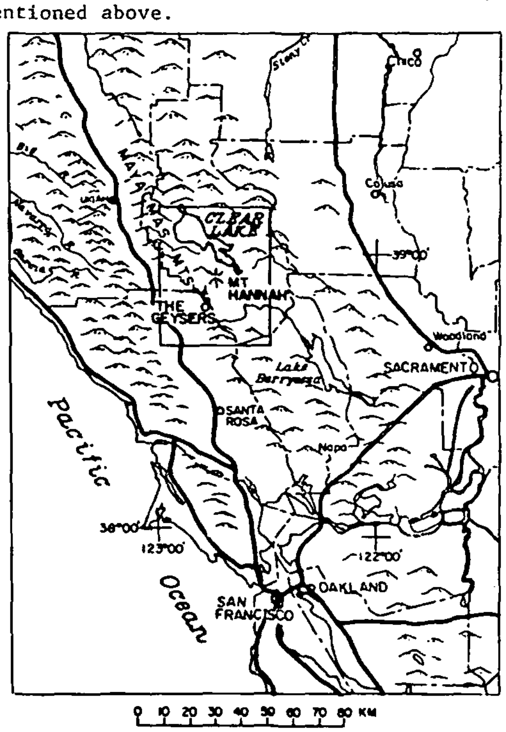

Figure 1. Index map, showing the region of discussion in this paper.

We modeled the gravity data using the thermal constraints above on the mass distribution. The maximum mass change due to water flashing to steam over the bulk reservoir volume, which is consistent with the lack of a measurable temperature change is .004 g/cc. Values of this magnitude were used to model the time changes in the gravity and the results are shown in Table 2. The value of .002 g/cc represents a lower bound in the modeling as the mass distribution is then too diffuse to reproduce the gravity amplitudes shown in Figure 2.

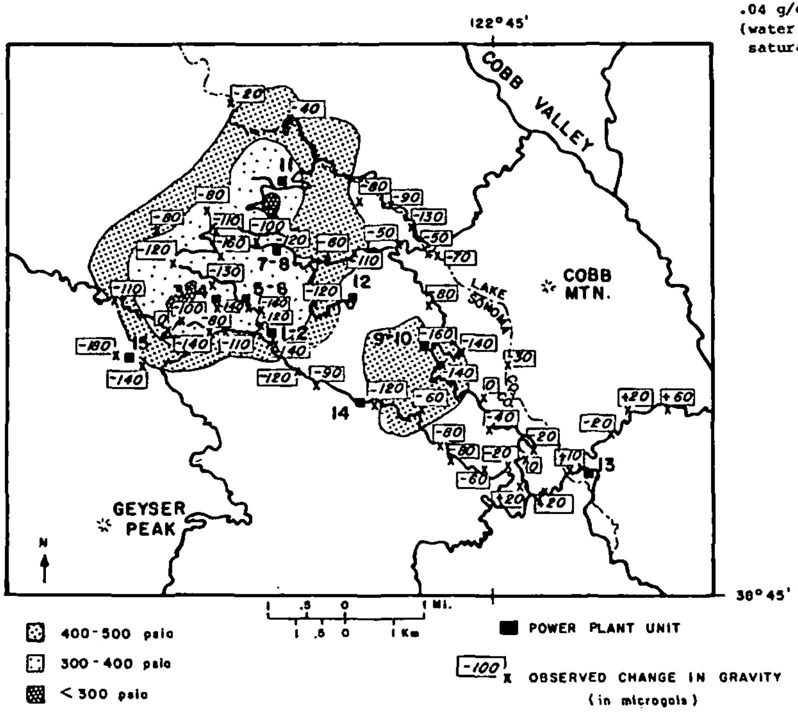

Figure 2. Time changes in surface gravity between 1974 and 1977 and the pore pressure decay near the top of the steam reservoir (from Lippman and others, 1977).

TABLE 1. A STUDY OF RESERVOIR DEPLETION AT THE GEYSERS GEOTHERMAL FIELD.

PHYSICAL CHANGES	SURFACE MEASUREMENT	RESULT OF ANALYSIS
Pore fluid mass deficiency due to production.	Change in gravity with time.	A tradeoff between a depleted reservoir volume and some uniform mass deficiency.
Volumetric strain within the reservoir.	Geodetic measurement of surface strain with time.	For uniform dilatation, a tradeoff occurs between volume and strain for a given maximum strain amplitude.
Change in temperature of reservoir during production, and change in enthalpy of produced steam.	Temperature and pressure measurements of steam within wells.	Limits on the pore fluid mass changes due to water flashing to steam within the reservoir volume.

TABLE 2. RESULTS FROM MODELING OF RESERVOIR GRAVITY CHANGE FOR A CYLINDRICAL VOLUME.

mass deficiency	radius	height	depth to top
.002 g/cc	1.8 km	3.7 km	0.0 km
	2.1 km	2.7 km	0.0 km
.003 g/cc	1.5 km	3.7 km	0.5 km
	1.7 km	2.7 km	0.5 km
.004 g/cc	1.8 km	3.7 km	1.0 km
	2.3 km	2.2 km	1.0 km
.04 g/cc (water saturation)	1.0 km	0.6 km	1.5 km maximum

We also modeled the geodetic strain to determine the volumetric strain within the depleted reservoir volume. At The Geysers, horizontal contraction and surface subsidence measured by Lofgren (1979) overlie the portion of the reservoir volume depleted by steam production, (Figure 3), as shown by pore pressure decay (Lippman and others, 1977).

By modeling the geodetic data, (assuming simple dilatation), we calculated strains up to 10^{-4} within the depleted reservoir volume. By assuming that the reservoir strain is a mechanical response to increased effective stress as the pore pressure decays, we also calculated a bulk or "framework" modulus for the reservoir matrix. The change in effective stress may be calculated from the pore pressure change (which is estimated from data presented by Lippman and others, 1977), and a value for the "intrinsic" bulk modulus of the reservoir rock (Rice and Cleary, 1976). For an intrinsic bulk modulus of the reservoir rock we used lab measurements of the compressional velocity of Franciscan graywacke (Stewart and Peselnick, 1978) combined with values of Poissons ratio form earthquake data (Majer and McEvilly, 1978). The strain we calculated for a given reservoir volume was then combined with the changes in effective stress to produce a value for the bulk modulus of the reservoir matrix. The bulk moduli determined in this way from the geodetic data are listed in Table 3 for several reservoir volumes (which were determined using the gravity and temperature measurements).

Bulk moduli obtained from microseismic monitoring within the reservoir (Majer and McEvilly, 1978) are an order of magnitude larger (bulk modulus about 3.X 10^5 bars), and lie between lab values calculated from Stewart and Peselnick (1978) and calculated bulk moduli for the reservoir matrix. The bulk moduli determined from seismic measurements therefore produce much smaller strains given the observed changes in pore pressure and calculated changes in effective stress.

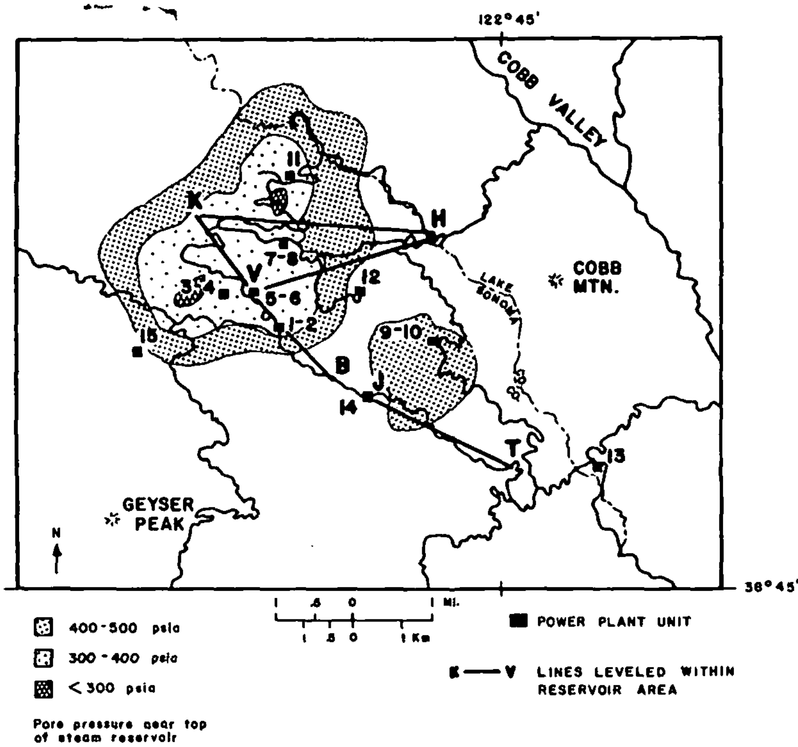

Figure 3. First order leveling lines in The Geysers steam field and the pore pressure decay near the top of the steam reservoir as of early 1977.

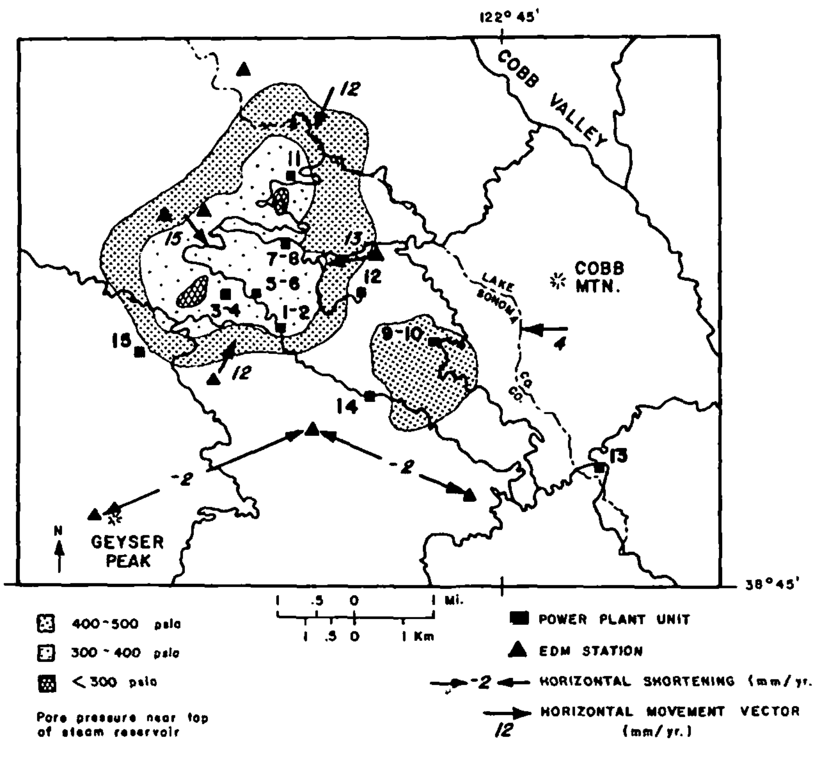

Figure 4. Horizontal movement vectors from high precision geodolite surveys (Lofgren, 1979) and the pore pressure decay near the top of the steam reservoir.

Thus the larger moduli measured at seismic frequencies (averaging about 10 Hz) requires a mechanical model for the reservoir in which there is a significant change of modulus with either frequency or strain amplitude. For seismic stress waves, the small stresses and strains (10^{-8}) are linearly related. But the larger reservoir strain (10^{-4}) may not be linearly related to the changes in effective stress resulting from fluid depletion. This would be especially true if the large strain amplitude resulted from slippage along the numerous fracture surfaces within The Geysers steam reservoir. This slip or creep may be one of the principal reasons our data appears to confirm the commonly observed discrepancy between static and dynamic moduli.

Non-mechanical, thermal contraction of the depleted reservoir volume is an alternative model of reservoir strain at The Geysers. Thermal contraction requires a temperature decline (which at the present is too small to measure) over the bulk reservoir volume, and which occurs independently of mechanical coupling between the pore fluid and the rock matrix. If 0.3% of the bulk reservoir volume is water which flashes to steam, this not only produces a mass deficit of -.003 g/cc, but also decreases the temperature of the reservoir by 1.5 degrees Celsius. Measurements of the thermal expansivity of various rock types (Skinner, 1966) imply that a reservoir strain of 8.X 10^{-5} would correspond to this temperature change, and the strain so obtained agrees with the values calculated from geodetic data (Table 3). Thus a change in liquid content of 0.3% of the bulk reservoir volume, with the inherent temperature change, fits the observed deformation data on The Geysers steam system.

Thus we have shown that two alternative strain mechanisms are consistent with the response of the steam reservoir to fluid depletion at The Geysers. Since both mechanisms are related to fluid withdrawal and pore pressure decay, there are at this time no measurements capable of distinguishing thermal from mechanical strain.

By monitoring the gravity and geodetic strain at The Geysers as the reservoir is depleted, the outward growth of the depletion zone may be studied in detail. Once the fluid is completely depleted in a section of the reservoir, then it

TABLE 3. RESULTS OF MODELING RESERVOIR STRAIN WITH PURE DILATATION.

SHAPE.	radius	height	volume	mass deficiency	strain ($\Delta V/V \times 10^{-5}$) d=0.5 km	d=1.0 km
CYLINDER	1.6 km	4.7 km	37. km^3	.002 g/cc	3.2 ± .3	6.0 ± .6
	1.5 km	3.5 km	25. km^3	.003 g/cc	4.7 ± .5	7.8 ± .8
	1.5 km	2.1 km	15. km^3	.005 g/cc	6.7 ± .7	10.2 ± 1.0

	radius	depth to center	volume	mass deficiency	strain ($\Delta V/V \times 10^{-5}$)
SPHERE	2.1 km	2.5 km	37. km^3	.002 g/cc	4.2 ± .4
	1.8 km	2.5 km	25. km^3	.003 g/cc	6.4 ± .6
	1.6 km	2.5 km	15. km^3	.005 g/cc	10.5 ± 1.0

POISSONS RATIO ASSUMED TO BE 0.25.

USING CYLINDER MODEL, EFFECTIVE BULK MODULUS FROM STRAIN DATA IS

$K = 0.3$ to 0.5×10^5 bars.

SEISMIC REFRACTION DATA (MAJER AND McEVILLY, 1978) IMPLY THAT THE BULK MODULUS IS

$K_s = 2.5$ to 3.0×10^5 bars.

VALUES ABOVE ARE CALCULATED FOR THE PERIOD FROM 1973-75 DURING WHICH THE PORE PRESSURE CHANGE WAS BETWEEN ONE AND THREE BARS.

will be possible to estimate the initial liquid saturation in that portion of the reservoir. If the liquid saturation in one section of the reservoir may be extrapolated to other parts of the steam reservoir, then since the rate of of mass depletion is known, a lifetime estimate may be made for The Geysers steam field.

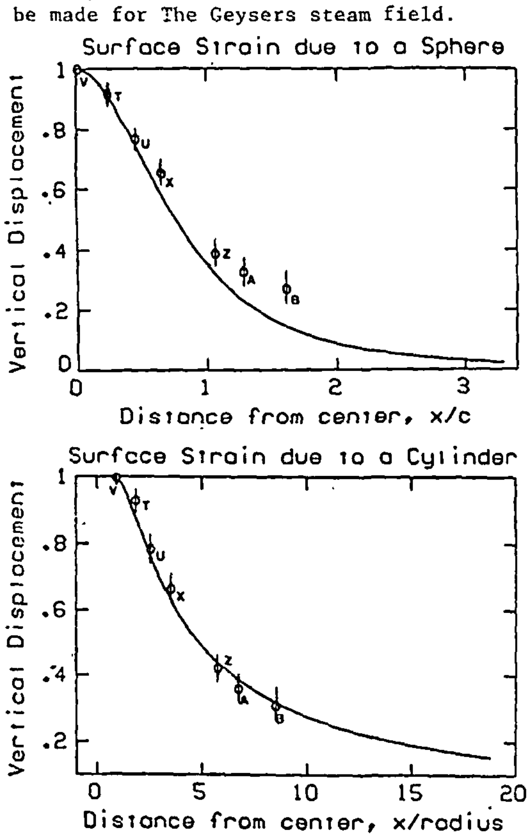

Figure 5. First order leveling data compared to theoretical vertical movement for both a cylinder and a sphere in an elastic half space. Radius of the cylinder is 1.5 km, and the depth to the center of the sphere is 2.5 km.

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GEOCHEMICAL INVESTIGATIONS AT EDGEMONT, SOUTH DAKOTA

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ABSTRACT

Average temperature-depth gradients in wells penetrating the Mississippian Madison Limestone in Edgemont, South Dakota, are 4.7°C/100 m, whereas the average gradient in the Black Hills area is 2.6°C/100 m. Chemical analyses of ground water from forty-one Madison wells and springs were obtained to define the geochemical characteristics of the low-temperature (<60°C) geothermal system.

The results of these analyses show that ground water from wells with high temperature-depth gradients is a sodium chloride-calcium sulfate type, and has concentrations of silica, lithium, and cobalt which are significantly higher than those which are found in all other waters sampled.

INTRODUCTION

Artesian wells penetrating the Mississippian Madison Limestone in Edgemont, South Dakota, exhibit average temperature-depth gradients of 4.7°C/100 m, whereas the average gradient at 24 other Madison wells in the Black Hills area is 2.6°/100 m. This low temperature geothermal area was the focus of a feasibility study (Iszler, et. al., 1979) sponsored by the U.S. Department of Energy, which was designed to evaluate the application of geothermal energy to Edgemont's local school complex.

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The chemistry of low-temperature (<60°C) geothermal waters in the Madison Limestone is not well understood. This investigation was undertaken to define the geochemical characteristics of ground waters from the Madison Limestone in the Black Hills area, South Dakota and Wyoming.

This summary is based on detailed chemical analyses of forty-one ground-water samples from Madison wells and springs in the Black Hills area. Samples from twenty-two sites were collected during the summer of 1978. Additional data were obtained from the U.S. Geological Survey, Water Resources Division, Rapid City, South Dakota, and from Gries (1977).

GROUND WATERS SAMPLED IN EDGEMONT

Five wells within the Edgemont city limits penetrate the Madison Limestone. The water temperatures of three of these wells were monitored daily during six-day pumping tests in February and March 1978. The wells were sampled daily during the first pumping test, and monthly from April to August 1978.

The results of chemical analyses are listed in Table 1. In addition to these analyses, alkalinity, silica, sulfate, and sulfide concentrations were determined in the field. Analytical procedures utilized were those based upon techniques outlined in Standard Methods for the Examination of Water and Wastewater, 14th Edition.

Table 1. Chemical data from ground water in Madison wells, Edgemont, South Dakota

Location	9S-2E-1aa3	9S-2E-1acdb1	9S-2E-1bcd1	9S-2E-2daa1
Well Name	Edgemont-Burlington R.R.	Edgemont-City #1	Edgemont-City #2	Edgemont-City #4
Depth (m)	901	909	970	1055
Temp (°C)	51.7	51.0	51.0	52.5
TDS	800	1210	1192	1254
Li	0.14	0.28	0.28	0.30
Na	150	250	261	270
K	11	17	21	19
Rb	---	0.15	0.15	0.18
Cs	---	0.30	0.30	0.18
Mg	17	32	32	36
Ca	116	116	120	124
Sr	3.1	3.4	3.2	3.4
Fe	---	n.d.	0.01	0.03
Co	---	0.06	0.06	0.06
Ni	---	n.d.	n.d.	n.d.
Cu	---	n.d.	n.d.	n.d.
Ag	---	n.d.	n.d.	n.d.
Pb	---	n.d.	n.d.	n.d.

Table 1. Continued

Location Well Name	9S-2E-1aa3 Edgemont-Burlington R.R.	9S-2E-1acdb1 Edgemont-City #1	9S-2E-1bcd1 Edgemont-City #2	9S-2E-2daa1 Edgemont-City #4
Zn	---	n.d.	n.d.	n.d.
Al	---	n.d.	n.d.	n.d.
S^{2-}	---	3.2	0.04	0.03
HCO_3	181	157	161	177
F	0.8	0.9	1.1	1.1
Cl	139	244	249	284
SO_4	300	424	466	466
SiO_2	34	45	45	43
pH	7.0	6.9	6.8	6.8

All data reported in mg/l
n.d. = not detected
Lowest detectable limits:

Li = 0.02	Sr = 0.02	Ni = 0.05	Ag = 0.0008	Cu = 0.007	Al = 5.0
Rb = 0.13	Co = 0.02	Fe = 0.01	Pb = 0.01	Zn = 0.015	S^{2-} = 0.005
Cs = 0.13					

Water temperatures ranged from 51.0 to 55.6°C during the pumping tests. However, temperatures at each site did not increase significantly and the chemical composition of the ground waters did not vary significantly. Therefore, we concluded that increasing the discharge rate of the wells would not significantly alter either water temperature or water chemistry.

GROUND WATERS SAMPLED WITHIN A 50-MILE RADIUS

Of the forty-one wells and springs sampled in the Black Hills area, eighteen are within a 50-mile radius of Edgemont. The locations of four selected sample sites and the results of chemical analyses are listed in Table 2. Water temperatures ranged from 10.6°C to 53.3°C, and temperature-depth gradients ranged from 1.7°C/100 m to 4.0°C/100 m.

Table 2. Chemical data from ground water for selected Madison wells and springs within a 50-mile radius of Edgemont, South Dakota

Location	8S-5E-20cda	7S-5E-13bdd	10S-2E-3daa1	45N-60W-20dca1
Well/Spring Name	Cascade Springs	Evans Plunge	Black Hills Ordnance Depot #1	Newcastle #1
Depth (m)			1219	804
Temp (°C)	21.5	32.0	53.3	26.1
TDS	2636	1122	1024	266
Li	0.08	0.14	0.25	n.d.
Na	38	84	182	2.1
K	10	14	18	1.9
Mg	95	44	22	29
Ca	549	200	121	62
Sr	6.9	3.2	1.8	0.02
Co	n.d.	n.d.	0.07	n.d.
Ni	n.d.	n.d.	n.d.	n.d.
HCO_3	183	183	156	281
F	1.0	0.9	0.6	0.1
Cl	52	110	308	2.5
SO_4	1490	430	310	57
SiO_2	21	21	34	14
pH	6.9	6.8	7.3	7.6

All data reported in mg/l

TEMPERATURE-DEPTH GRADIENTS

A true geothermal gradient for the Black Hills was not calculated, due to the lack of available heat flow data. Temperature-depth gradients reported in this summary are defined in the following manner:

Temperature of water recorded on-site=t_{H_2O}

Mean annual temperature at or near site of well= t_{MA}

Corrected temperature=$t_{H_2O}-t_{MA}=t_C$

Total depth of well=d

Critical depth (depth to which temperature measurement is affected by mean annual temperature or near surface thermal disturbances)=18 meters (Schoon and McGregor, 1974)=d_{CR}

Corrected depth=$d-d_{CR}=d_C$

Then, the temperature-depth gradient for a single measurement at one well=$t_C/d_C=G_{TD}$.

For most wells, more than one temperature measurement was recorded. In some localities, more than one well exists. For these cases, the temperature-depth gradient was averaged for all measurements at all wells.

This method of calculation assumes the following:

1. Depth is a vertical distance from ground surface.
2. Correction for mean annual temperature is independent of discharge rate.
3. Water is not mixing between aquifers.
4. Temperature equilibrium is maintained in well.
5. Mean annual temperature influences water temperature at surface and influences critical depth as defined above.

CHEMICAL COMPOSITION

The results of standard mineral analyses were plotted on multiple tri-linear diagrams after Piper (1944). Ground waters were classified as either sodium chloride-calcium sulfate type, calcium sulfate type, or calcium-magnesium bicarbonate type. The bicarbonate waters were generally found in wells and springs near recharge areas, and the calcium sulfate waters were found with either increasing distance from recharge areas or in ground waters which were known to be mixing between aquifers. The sodium chloride-calcium sulfate type water was found only in Madison ground waters in the Edgemont geothermal area. The standard mineral composition of these ground waters was found to be significantly different from that of all other sampled ground waters.

Of all the trace elements analyzed, only lithium and cobalt have higher concentrations in Edgemont area waters than in waters at all other sites. Strontium concentrations in Edgemont area waters ranged from 1.8 to 3.4 mg/l. Although cesium and rubidium were detected in Edgemont ground waters, it was not determined whether this is significant as there is no data from other sampled sites. Aluminum, copper, lead, manganese, silver, and zinc were not detected in Edgemont waters, and additional analyses supplied by the U.S. Geological Survey indicated that only trace amounts of these elements, if any at all, occur in ground waters at other sites in the study area.

GEOTHERMOMETRY

Temperatures were calculated using the quartz and chalcedony geothermometers defined by Truesdell (1975) and the Na-K-Ca geothermometer defined by Fournier and Truesdell (1973). Water temperatures calculated using the quartz geothermometer ranged from 9.7°C to 46.1°C above observed temperatures, whereas temperatures calculated using the chalcedony geothermometer ranged from 24.5°C below to 14.0°C above observed temperatures. Water temperature calculated using the Na-K-Ca geothermometer ranged from 41.3°C below to 41.5°C above observed temperatures.

Table 3 lists the results of selected geothermometer calculations. The chalcedony geothermometer yielded the best correlation between observed temperatures and calculated temperatures. This is in agreement with Fournier (1973) who found that chalcedony, rather than quartz, controls the dissolved silica content of waters below 100°C.

Most work with geothermometers has been done for high-temperature geothermal systems. The highest observed temperature in this study was 56°C. No correlations exist between the results of geothermometer calculations and temperature-depth gradients.

ACKNOWLEDGEMENTS

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Table 3. Results of selected geothermometer calculations

Spring/Well	T, °C Observed	T, °C Quartz	T, °C Chalcedony	T,°C Na-K-Ca
Edgemont Burlington R.R.	51.7	84.8	51.8	68.1
Edgemont City #1	51.0	97.1	65.0	87.8
Edgemont City #2	53.3	97.1	65.0	94.8
Edgemont City #4	54.4	95.0	62.7	91.0
Cascade Springs	22.2	65.5	31.4	25.4
Evans Plunge	33.5	65.5	31.4	25.4
Black Hills Ordnance Depot #1	53.3	84.8	51.8	85.1
Newcastle #1	26.1	50.8	16.1	0.3

GEOCHEMISTRY OF ACTIVE GEOTHERMAL SYSTEMS IN THE NORTHERN BASIN AND RANGE PROVINCE

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ABSTRACT

Numerous thermal springs occur in the northern Basin and Range Province due primarily to the structure and high regional heat flow. Dilute to slightly saline (200 to 3,000 mg/L TDS) Ca and/or $Na-HCO_3$ type waters, many associated with travertine, are dominant in eastern and northeastern Nevada. CO_2-charged $Na-HCO_3$ waters are particularly common along the eastern side of the Sierra Nevada from Long Valley north to Bridgeport. Moderately to very saline (3,000 to 35,000 mg/L TDS) Na-Cl type waters predominate near major topographic lows. $Na-SO_4$ type waters are common in western Nevada and in northeastern California. Na-mixed anion waters are common along the north side of the Black Rock Desert in northwestern Nevada and the Alvord Desert in southeastern Oregon.

Measured temperatures in deep geothermal wells in the northern Basin and Range Province are about 14°C cooler than the average temperature calculated using two chemical and one isotope geothermometer on waters discharged by nearby thermal springs and shallow wells (<100m). Isotopic data (δD) for thermal springs in the northern Basin and Range have the same general pattern as modern precipitation. However, the few detailed studies of recharge areas for specific systems have generally found local cold waters to be slightly more enriched in deuterium than water currently discharged by the thermal springs. This probably indicates that the water currently being discharged by the hot springs was recharged during colder time periods, perhaps the Pleistocene.

Introduction

Although chemical analyses of a few hot springs in the northern Basin and Range were published in the late 1800's (Peale 1886, Church 1878), the first detailed studies were begun by Don White at Steamboat Springs in the late 40's and continued into the early 60's (White and Brannock, 1950, White and others, 1964). Research activities exploded in the 70's as interest in the geothermal resource increased. Most thermal springs in the northern Basin and Range Province have now been sampled and one or more analyses are available in the literature. Compilations of chemical data for thermal springs in the various states which include parts of the northern Basin and Range Province have been presented by Mundorff (1970), and Goode (1978) for Utah, Garside and Schilling (1979) for Nevada, Majmundar (in press) for California, and U.S. Geological Survey and Oregon Department of Mineral Industries (1979) for Oregon. Locations and temperatures of thermal springs and wells have also been compiled by Waring (1965) and Berry and others (1980). Figure 1 shows the outline of the northern part of the Basin and Range Province (Great Basin) along with physical features mentioned in the text.

Chemical Composition of Water

Chemically, the thermal waters of the Basin and Range Province range from dilute (<1,000 mg/L TDS) $Na-HCO_3$ or $Ca-HCO_3$ type waters to very saline (10,000-35,000 mg/L TDS) Na-Cl type waters (Table 1). The chemical constituents in thermal waters are derived primarily from the country rock by dissolution of, or exchange with, the rock-forming minerals (Ellis and Mahon, 1964, 1967). Ideally, concentrations of silica, sodium, potassium, calcium and magnesium are controlled by mineral-water equilibria. However, other constituents such as chloride don't attain concentrations sufficient to reach saturation with respect to any chloride bearing minerals.

Na-Cl type thermal waters of moderate salinity occur in northwestern Nevada and west-southwest of Tonopah (Fig. 2). Na-Cl type thermal waters also discharge along the eastern edge of the Basin and Range Province in Utah and southeastern Idaho. These waters result from the extensive reaction of thermal fluids with sedimentary rocks deposited in a marine environment. Sulfate in the Na-Cl type water of Utah is primarily of marine origin based on both the oxygen and sulfur isotopic compositions of dissolved sulfate (Nehring and Mariner, 1979, Cole, 1982). Locally, chemical and isotopic data indicate some admixture with chloride-rich saline-lake or playa waters may have occurred. High chloride concentrations can also be attained by circulation through some granites (Moore and others, 1983), however, the extent of

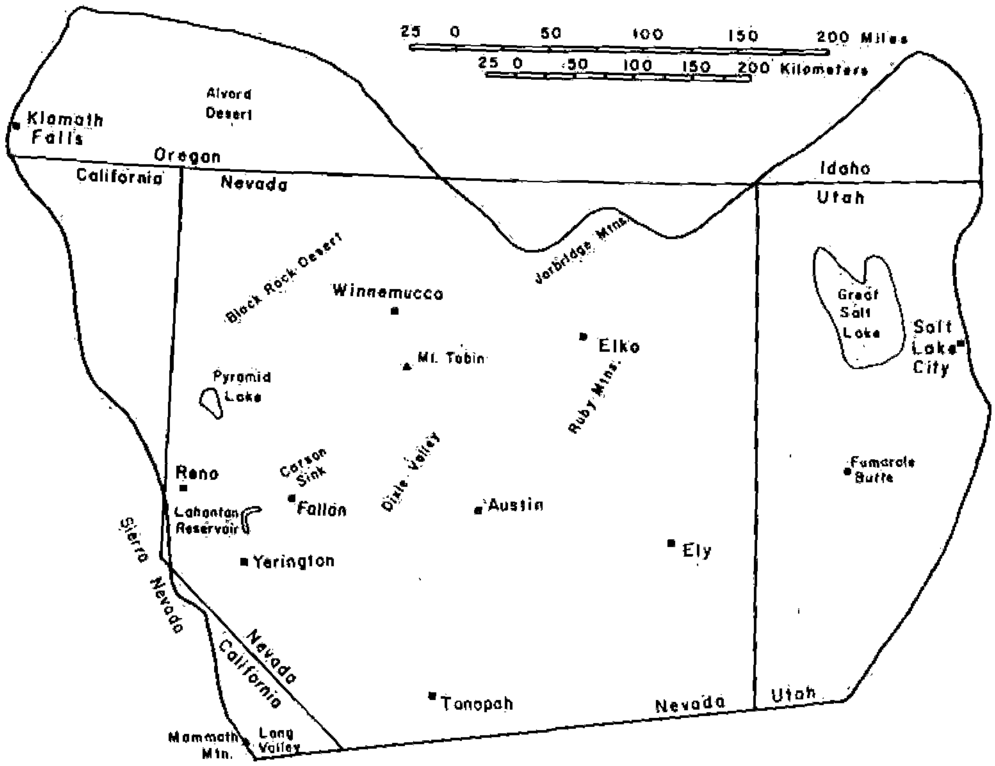

Figure 1. Location map of the major population centers and features mentioned in the text.

this process in the Basin and Range is unknown. High chloride concentrations are an indication that more extensive water-rock reaction has taken place, and this may infer longer flow paths and generally deeper circulation. Most high temperature geothermal wells in the Great Basin have encountered Na-Cl waters.

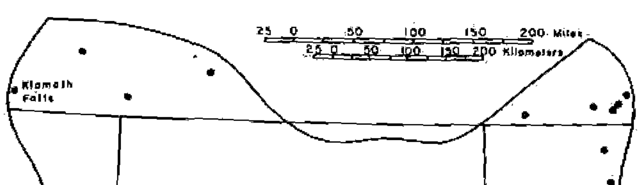

Figure 2. Distribution of Na-Cl waters in the northern Basin and Range Province.

Bicarbonate waters range from dilute to slightly saline depending largely on the source of the dissolved CO_2. If CO_2 is generated at depth and dissolved under pressure in the thermal water, then a high TDS CO_2-charged water will be produced. High TDS CO_2-charged waters are particularly common adjacent to the southern Sierra Nevada. The more saline CO_2-charged waters are slightly radioactive due to their high radium and radon content (Femlee and Cadigan, 1982, and Wollenberg, 1974). Enough calcium is present in most of these waters to form travertine mounds and terraces. Fales and Travertine hot springs, near Bridgeport, California discharge this type of water as does Hyder Hot Springs, in Dixie Valley, Nevada. However, if the CO_2 is dissolved from the atmosphere, or soil gas, then the dissolved carbon concentrations and related total dissolved solids will be very low. Darrough and Soldier Meadows hot springs in central and northwestern Nevada are good samples of dilute Na-HCO_3 waters. Dilute Na-HCO_3 and Ca-HCO_3 type waters are particularly common in central and eastern Nevada (Fig. 3). A dilute Ca-HCO_3 water will be produced if the principal bedrock is a limestone and these waters are common in eastern Nevada where Paleozoic limestones crop

out. The dilute Na-HCO_3 waters develop in areas where the rock contains appreciable sodium silicate or sodium aluminosilicate minerals and only small amounts of CO_2 are available.

Dilute to slightly saline Na-SO_4 (± Cl) waters occur principally in northeastern California and the adjacent part of Nevada (Fig. 4). High sulfate concentrations can result from dissolution of gypsum (anhydrite) in sedimentry rocks or dissolution of minerals such as alunite, jarosite, anhydrite, or pyrite from mineralized zones (Hem, 1970). Oxidation of sulfide to sulfate can also produce large sulfate concentrations and acid sulfate waters (White and others, 1971) but these are rare in the Basin and Range Province. The oxygen isotopic compositions of marine and hydrothermal sulfate are quite different (Longinelli and Craig, 1967, and McKenzie and Truesdell, 1977) thus it is relatively easy to differentiate between them. Sulfates in the waters of the western part of the Great Basin generally range from 0 to -10 o/oo in oxygen-18 (Nehring and Mariner, 1979) and have a possible source in the sulfate minerals associated with ore deposits and gossans. Gypsum and anhydrite deposits in Mesozoic marine rocks are also fairly common in western Nevada, however, only one of the hot springs for which isotopic data is available,

Table 1. Chemical Composition of Selected Thermal Waters of the Northern Basin and Range--continued. [Concentrations are in mg/L, temperatures are in °C; sodium (Na) values followed by K represent sodium plus potassium; dashes (-) indicate no data.]

Name/Location	Temp	pH	SiO_2	Ca	Mg	Na	K	HCO_3	Cl	SO_4	F	Reference
CALIFORNIA												
Alpine County												
Unn. sprs. on the Carson River												
SENESE, sec. 34, T. 11 N., R. 20 E.	84	6.63	178	76	6.5	510	40	388	355	550	6.3	Unpub. data, Mariner and others
Unn. sprs. on the Carson River												
NWSW, sec. 26, T. 11 N., R. 20 E.	65	6.52	110	125	4.0	480.	21	333	295	700	2.4	Unpub. data, Mariner and others
Lassen County												
Amadee Hot Springs												
NESW, sec. 08, T. 28 N., R. 16 E.	96	8.36	98	15	<.1	235	5.7	57	155	280	4.6	Mariner and others, 1976a
Bassett Hot Springs												
NWSE, sec. 12, T. 38 N., R. 7 E.	79	8.53	65	30	<.1	220	3.2	32	93	370	2.0	Reed, 1975
Kellog Hot Springs												
SWSE, sec. 15, T. 38 N., R. 8 E.	78.4	8.63	85	30	<.1	240	5.9	35	110	370	2.6	Reed, 1975
Wendell Hot Springs												
NESW, sec. 23, T. 29 N., R. 15 E.	95.5	8.26	125	20	<.1	280	8.0	53	185	340	4.2	Mariner and others, 1976a
Zamboni Hot Spring												
NWNW, sec. 24, T. 24 N., R. 17 E.	41	9.37	36	2.2	.02	66	.57	74	14	57	2.0	Unpub. data, Mariner and others
Modoc County												
Hot Springs Motel												
NESW, sec. 06, T. 42 N.,R. 17 E.	98	8.40	100	16	<.01	280	5.5	61	200	320	5.1	Reed, 1975
Kelly Hot Springs												
NENW, sec. 29, T. 42 N., R. 10 E.	91.5	8.08	110	20	<.1	250	6.5	47	160	300	2.1	Reed, 1975
Leonards Hot Springs												
NENE, sec. 13, T. 43 N., R. 16 E.	61.8	7.82	110	26	.6	330	8.5	84	220	390	5.2	Reed, 1975
Little Hot Springs												
NWSW, sec. 9, T. 39 N., R. 5 E.	75.7	7.59	87	44	.2	230	5.2	49	120	400	1.9	Reed, 1975
Seyforth Hot Springs												
NWNW, sec. 12, T. 39 N., R. 5 E.	85	7.66	110	28	<.1	300	9.0	63	220	370	5.4	Reed, 1975
West Valley Reservoir (spring)												
NWNE, sec. 29, T. 39 N., R. 14 E.	77.3	7.79	130	19	<.1	330	11.	63	150	510	4.0	Reed, 1975
Mono County												
Benton Hot Springs												
SW, sec. 2, T. 2 S., R. 31 E.	56.5	9.32	63	1.4	<.1	80	1.0	96	22	50	3.8	Mariner and others, 1977
Fales Hot Springs												
SE, sec. 24, T. 6 N., R. 23 E.	61	6.55	114	41	10	560	37	1130	160	260	4.7	Mariner and others, 1977
Long Valley-Hot Creek Gorge												
NE, sec. 25, T. 3 S., R. 28 E.	90	6.6	150	1.6	.1	400	24	549	225	100	9.6	Mariner and Wiley, 1976
Mono Lake - North Shore												
sec. 11, T. 2 N., R. 26 E.	66	7.68	76	13	2.9	430	8.8	454	350	100	4.8	Mariner and others, 1977
- South Shore												
sec. 18, T. 1 N., R. 27 E.	33	6.38	130	120	61	410	34	1560	105	28	.4	Mariner and others, 1977
Travertine Hot Spring												
SW, sec. 34, T. 5 N., R. 25 E.	69	6.73	100	64	18	1100	55	1800	200	920	4.5	Mariner and others, 1977
IDAHO												
Bear Lake County												
Bear Lake Hot Spring												
SW, sec. 13, T. 15 S., R. 44 E.	47.5	6.6	35	210	55	180	61	256	79	800	7.1	Young and Mitchell, 1973
Cassia County												
RRGE-1												
sec. 23, T. 15 S., R. 26 E.	95	8.9	144	58	0.3	505	35	46	896	59	6.2	Nathenson and others, 1982
Franklin County												
Maple Grove Hot Springs												
NE, sec. 7, T. 13 S., R. 41 E.	76	7.3	55	89	24	490	40	491	630	260	1.1	Young and Mitchell, 1973
Wayland Hot Springs												
NE, sec. 8, T. 5 S., R. 39 E.	77	7.0	80	160	16	3100	660	699	5400	50	12	Young and Mitchell, 1973
Oneida County												
Woodruff Hot Springs												
NE, sec. 10, T. 16 S., R. 36 E.	27	7.3	29	130	45	910	87	454	1600	58	.6	Young and Mitchell, 1973

Table 1. Chemical Composition of Selected Thermal Waters of the Northern Basin and Range--continued.
[Concentrations are in mg/L, temperatures are in °C; sodium (Na) values followed by K represent sodium plus potassium; dashes (-) indicate no data.]

Name/Location	Temp	pH	SiO_2	Ca	Mg	Na	K	HCO_3	Cl	SO_4	F	Reference
NEVADA												
<u>Carson City</u>												
Carson Hot Springs, SENE, sec. 5, T. 15 N., R. 20 E.	49	9.3	-	2.6	.4	96	-	92	29	96	-	Worts & Malmberg, 1966
Pinyon Hills Well sec. 23, T. 15 N., R. 20 E.	46	8.6	-	275	3	254	-	41	33	135	4.3	CWRR, 1973
<u>Churchill County</u>												
Dixie Valley Hot Springs SE. sec. 5, T. 22 N., R. 35 E.	72	8.6	115	3.2	.02	190	6.5	133	126	111	16.3	Mariner and others, 1974
Dixie Federal 52-18 NE, sec. 18, T. 24 N., R. 37 E.	boiling	8.27 @32°C	383	4.4	.03	385	36	379	307	127	7.0	Unpub. data, Mariner and others
Brady's Hot Spring (well) SW, sec. 12, T. 22 N., R. 26 E.	-	6.78 @24°C	164	45	.32	850	36	111	1140	320	5.8	Unpub. data, Mariner and others
Eagle Salt Works Spring sec. 34, T. 22 N., R. 26 E.			259	32	2	839	-	99	955	334		Adams, 1944
Soda Lake-Upsal Hogback SW, sec. 28, T. 20 N., R. 28 E.	boiling	7.86 @42°C	160	82	2.1	1000	48	144	1500	360	.6	Mariner and others, 1975
Stillwater Area SW, sec. 07, T. 19 N., R. 31 E.	96	7.57	170	108	1.7	1480	42	90	2200	190	5.0	Mariner and others, 1974
Lee Hot Springs Unsurveyed (39°12'N b 118°43'W)	88	7.4	180	44	.6	450	26	114	380	470	7.9	Mariner and others, 1974
<u>Douglas County</u>												
Hobo Hot Spring SESE, sec. 23, T. 14 N., R. 19 E.	46	8.9	47	6	.7	125	1.7	85	74	109	7.1	Clancy and Katzner, 1975
Saratoga Hot Springs SESE, sec. 23, T. 14 N., R. 20 E.	50	9.0	20	172	-	160k	-	18	39	678	9.0	Clancy and Katzner, 1975
Walley's Hot Spring NE, sec. 22, T. 13 N.,R 19 E.	62	8.8	58	10	.01	145	3.6	68	44	235	4.9	Mariner and others, 1974
<u>Elko County</u>												
Nile Spring SW, sec. 30, T. 47 N., R. 70 E.	43	7.2	31	40	11.5	10	5.6	149	8.7	37	.4	Mariner and others, 1974
Trout Creek Ranch Well NWNW, sec. 23, T. 46 N., R. 69 E.	43	8.3	21	16	5.7	24	5.6	120	2	22	.6	Moore and Eakin, 1968
San Jacinto Ranch Spring NWNW, sec. 23, T. 46 N., R. 64 E.	26	8.1	18	25	8.6	13	3.9	132	3.9	11	.5	Moore and Eakin, 1968
Rizzi Ranch Hot Spring sec. 29, T. 45 N., R. 54 E.	41	7.4	23	29	7.7	110	8.3	380	4.4	36	3.4	Moore and Eakin, 1968
Mineral (Contact) Hot Springs sec. 16, T. 45 N., R. 64 E.	60	9.1	83	1.6	<.01	75	2.2	108	15	45	8.9	Mariner and others, 1974
Wild Horse Hot Spring SESE, sec. 4, T. 43 N., R. 55 E.	54	7.2	40	48	12	130	22	482	14	40	5.2	Mariner and others, 1974
Hot Creek Springs NW, sec. 12, T. 28 N., R. 52 E.	26	7.30	20	46	23.5	10	2.1	228	4.6	27	<.1	Mariner and others, 1974
Hot Creek Springs NW, sec. 34, T. 43 N., R. 60 E.	37.5	6.76	139	62	20	23	17	307	3	51	.78	Mariner and others, 1975
Hot Sulphur Spring NE, sec. 8, T. 41 N. , R. 52 E.	92	7.32	165	9.4	.22	160	15	367	16	54	9.6	Mariner and others, 1975
Wine Cup Ranch Well NWNW, sec. 25, T. 41 N., R. 64 E.	59	8.4	-	49	17	139k	-	426	30	69	-	Rush, 1968b
Hot Lake NNW, sec. 25, T. 38 N., R. 46 E.	18	7.2	57	29	5.8	33	7.0	132	22	34	.4	Unpub. data, Mariner and others
Unnamed spring on Rock Creek SWSW, sec. 1,T. 39 N., R. 47 E.	35	6.87	23	41	11	86	14	330	13	62	2.4	Unpub. data, Mariner and others
Humboldt Wells Area SE, sec. 20, T. 38 N., R. 62 E.	60	6.58	110	78	36	300	30	1210	26	24	6.1	Mariner and others, 1975
Hot Hole NE, sec. 21, T. 34 N., R. 55 E.	56	7.21	65	60	15.5	120	39	490	16	72	1.9	Mariner and others, 1974
Hot Spring near Carlin sec. 33, T. 33 N., R. 52 E.	79	7.6	70	60	15	45	16	335	12	52	-	Mariner and others, 1974
Sulphur Hot Spring NW, sec. 11, T. 31 N., R. 59 E.	93	8.53	210	1.0	.03	135	8.9	274	23	40	17.7	Mariner and others, 1974
Smith Ranch (Unn. spr. - Ruby Marsh) NW, sec. 2., T. 27 N., R. 58 E.	65	8.0	50	45	12	58	14	377	6.5	24	-	Mariner and others, 1974
<u>Esmeralda County</u>												
Alkali Springs NW, sec. 26, T. 01 S., R. 41 E.	60	8.1	-	46	4.6	349k	-	348	68	492	-	Rush, 1968a
Silver Peak (Waterworks) Hot Springs SE,sec. 15, T 02 S., R. 39 E.	40	7.18 @19°C	105	540	71	10000	750	559	17000	410	4.1	Unpub. data, Mariner and others
<u>Eureka County</u>												
Beowawe SW, sec. 08, T. 31 N., R. 48 E.	98	8.98	320	1.0	<.1	230	16	383	69	130	17	Mariner and others, 1974
Hot Springs Point NW, sec. 11, T. 29 N., R. 48 E.	54	6.63	67	53	35	230	58	913	1	7	6.6	Mariner and others, 1974
Bruffey's (Mineral Hill) Hot Spring sec. 14, T. 27 N., R. 52 E.	66	7.0	58	52	16	39	8.7	287	14	27	.7	Roberts, Montgomery, and Lehner, 1967
Walti Hot Springs SW,sec. 33, T. 24 N., R. 48 E.	72	6.47	68	56	12	44	14	264	12	64	2.5	Mariner and others, 1974
Shipley Hot Springs NESE, sec. 23, T. 24 N., R. 52 E.	39	7.2	40	57	21	29	5.9	279	21	35	.2	Eakin, 1962a
Klobe Hot Springs Bartholomae) SE, sec.28, T. 18 N., R. 50 E.	54	9.25	85	1	<.1	64	.7	144	6.3	18	-	Mariner and others, 1974

Table 1. Chemical Composition of Selected Thermal Waters of the Northern Basin and Range--continued.
[Concentrations are in mg/L, temperatures are in °C; sodium (Na) values followed by K represent sodium plus potassium; dashes (-) indicate no data.]

Name/Location	Temp	pH	SiO_2	Ca	Mg	Na	K	HCO_3	Cl	SO_4	F	Reference
Humboldt County												
Cordero Mercury Mine well												
SE, sec. 28, T. 47 N., R. 37 E.	60	-	57	36	10	62k	-	195	34	56	-	Visher, 1957
Bog Hot Springs												
SWNW, sec. 07, T. 46 N., R. 28 E.	54	9.05	57	.2	<.1	81	1.0	138	15	45	1.7	Mariner and others, 1974
Baltazor Hot Springs (well)												
NW, sec. 13, T. 46 N., R. 28 E.	90	7.50	150	10	.1	180	8.2	156	47	230	6.8	Mariner and others, 1974
Howard Hot Springs												
NE, sec. 04, T. 44 N., R. 31 E.	56	9.2	85	3	<.1	88	1.7	127	10	62	-	Mariner and others, 1974
Dyke Hot Springs												
SE, sec. 25, T. 43 N., R. 30 E.	66	8.86	85	1.8	<.1	150	4.3	277	21	82	8.0	Mariner and others, 1974
The Hot Springs												
NE, sec. 20, T. 41 N., R. 41 E.	58	8.0	55	10	8	296	36	881	26	36	-	Mariner and others, 1974
Soldier Meadows Hot Springs												
sec. 23, T. 40 N., R. 24 E.	54	8.53	63	3.1	<.1	74	1.1	98	18	41	12	Mariner and others, 1974
Pinto Hot Springs												
ESE, sec. 17, T. 40 N., R. 28 E.	93	7.14	150	14	.4	330	23	497	160	120	12	Mariner and others, 1974
Double Hot Springs												
sec. 4, T. 36 N., R. 26 E.	80	7.93	105	4.8	.1	180	4.5	265	59	120	10	Mariner and others, 1974
Macfarlane's Bath House Spring												
NW, sec. 27, T. 37 N., R. 29 E.,	75	6.61	82	43	11	1350	28	2050	870	210	2.6	Unpub. data, Mariner and others
Well												
SWSE, sec. 03, T. 37 N., R. 39 E.	70	7.4	-	30	7.1	450	26	1240	14	52	-	Cohen, 1962
Golconda Area												
SE, sec. 29, T. 36 N., R. 40 E.	74	6.53	66	33	6.8	130	22	429	18	56	1.8	Mariner and others, 1974
Hot Pot												
SW, sec. 11, T. 35 N., R. 43 E.	57	6.95	39	18	5.3	660	28	1625	61	155	9.4	Unpub. data, Mariner and others
Hot Spring Ranch (Tipton)												
SE, sec. 05, T. 33 N., R. 40 E.	85	8.36	125	16	.9	200	18	385	41	140	-	Mariner and others, 1974
Lander County												
Buffalo Valley Hot Springs												
SE, sec. 23, T. 28 N., R. 41 E.	49	6.53	80	45	4.9	250	34	813	29	110	4.8	Mariner and others, 1974
" " " " " " " " "	73	-	71	43	5.2	295	33	-	29	104	4.6	Unpub. data, Mariner and others
Hot Springs Ranch (Valley of the Moon)												
NE, sec. 23, T. 27 N., R. 43 E.	53	8.0	40	20	9	118	21	333	21	64	-	Mariner and others, 1974
South Smith Creek Valley												
NE, sec. 25, T. 17 N., R. 39 E.	86	7.72	110	4.8	.06	170	8.4	256	22	102	8.9	Mariner and others, 1974
Spencer Hot Springs												
SE, sec. 13, T. 17 N., R. 45 E.	72	6.49	77	43	9.4	200	36	672	22	51	4.7	Mariner and others, 1974
Unnamed spring near Walti Hot Springs												
Unsurveyed 39°56.6'N by 116°40.8'W	64	6.51	83	66	11	135	42	559	32	58	4.1	Unpub. data, Mariner and others
Lyon County												
Hazen Area												
SW, sec. 18, T. 20 N., R. 26 E.	86	7.05	150	70	1.5	620	38	100	820	400	4.2	Mariner and others, 1975
Wabuska Hot Springs												
SE, sec. 16, T. 15 N., R. 25 E.	94	8.06	110	39	.1	300	14	74	55	620	8.2	Mariner and others, 1975
Hind's (Nevada) Hot Springs												
SE, sec. 16, T. 12 N., R. 23 E.	61	8.65	52	4.5	.01	102	2.5	69	17	169	3.1	Mariner and others, 1974
Wedell Springs												
SW, sec. 07, T. 12 N., R. 34 E.	60	7.83 @50°C	153	13	.2	270	9.2	214	78	315	12	Unpub. data, Mariner and others
Nye County												
Diana's Punch Bowl												
SE, sec. 22, T. 14 N., R. 47 E.	59	7.14	46	50	11	55	15	277	8	59	2.8	Mariner and others, 1974
Darrough's Hot Spring												
sec. 08, T. 11 N., R. 43 E.	95	8.29	98	1.3	.1	110	2.6	152	12	53	14	Mariner and others, 1974
Hot Creek Ranch Springs												
sec. 29, T. 08 N., R. 50 E.	63	8.0	135	51	15.1	197	13.4	545	42	86.4	8	Sanders and Miles, 1974
Upper Hot Creek Ranch Springs												
NESE, sec. 29, T. 08 N., R. 50 E.	67	8.1	161	33	9.5	193	1.4	501	37	64	8.3	USGS, 1977
Warm (Nanny Goat) Spring												
NWSW, sec. 20, T. 04 N., R. 50 E.	61	8.1	60	43	24	175	24	714	32	120	-	Mariner and others, 1974
Pershing County												
Colado (wells)												
SE, sec. 33, T. 28 N., R. 32 E.	60	7.56 @38°C	85	110	6.5	1450	120	199	2400	120	4.6	Mariner, Brook, Swanson and Mabey, 1978
Kyle Hot Springs												
SW, sec. 01, T. 29 N., R. 36 E.	77	6.50	150	95	25.5	540	80	544	770	51	5.7	Mariner and others, 1974
Trego Area												
40° 46'N by 119° 7'W	84.5	7.93	79	11	.4	430	8.6	163	500	180	4.1	Mariner and others, 1976
Leach Hot Springs												
SE, sec. 36, T. 32 N., R. 38 E.	92	7.4	135	8.8	.5	160	13	368	29	53	7.8	Unpub. data, Mariner and others
Humboldt House (Rye Patch)												
SE, sec. 21, T. 31 N., R. 33 E.	-	-	340	43	3.6	1350	240	202	2250	18	6.2	Unpub. data, Mariner and others
art. well	77	-	162	120	7.2	1400	240	380	2300	39	5.3	Unpub. data, Mariner and others
	-	-	440	90	1.0	2000	240	43	3600	90	-	Benoit, 1978
Sou Hot Springs												
SE, sec. 29, T. 26 N., R. 38 E.	70	7.3	64	106	19.8	167	26	324	77	354	5.5	Sanders and Miles, 1974
Hyder Hot Springs												
SW, sec. 28, T. 25 N., R. 38 E.	78	6.77	63	41	10	390	20	926	45	120	8.6	Mariner and others, 1976

Table 1. Chemical Composition of Selected Thermal Waters of the Northern Basin and Range--continued. [Concentrations are in mg/L, temperatures are in °C; sodium (Na) values followed by K represent sodium plus potassium; dashes (-) indicate no data.]

Name/Location	Temp	pH	SiO_2	Ca	Mg	Na	K	HCO_3	Cl	SO_4	F	Reference
Washoe County												
Fly Ranch (Ward's)												
sec. 02, T. 34 N., R. 23 E.	80	7.91	82	31	4.2	340	17	466	240	46	7.0	Mariner and others, 1974
Great Boiling Spring												
NW, sec. 15, T. 32 N., R. 23 E.	86	7.15	165	68	1.2	1400	130	83	2200	400	4.5	Mariner and others, 1974
San Emidio Desert												
Unsurveyed - 40°23'N 119°24'W	89	6.57	205	160	2.2	1400	110	102	2300	220	5.1	Mariner and others, 1976
The Needle Rocks												
NWSWSW, sec. 06, T. 26 N., R. 21 E.	56	8.43	110	260	.1	1100	160	24	1900	340	3.0	Mariner and others, 1974
Lawton Hot Springs												
SWNE, sec. 13, T. 19 N., R. 18 E.	49	9.0	46	6.2	.1	117	5.4	52	57	144	2.5	Cohen and Loeltz, 1964
Moana Springs Area (Biglin well)												
NESE, sec. 26, T. 19 N., R. 19 E.	85	8.0	106	23	.08	236	8.0	86	48	455	5.1	Bateman and Scheibach, 1975
Steamboat Springs												
NE, sec. 33, T. 18 N., R. 20 E.	94	7.19	270	16	.7	680	66	368	837	73	2.1	Mariner and others, 1974
Bowers Mansion Hot Springs												
NW, sec. 03, T. 16 N., R. 19 E.	46	9.36	47	2.9	.2	50	0.6	76	6.3	38	2.8	Mariner and others, 1975
White Pine County												
Cherry Creek Hot Springs												
sec. 6, T. 23 N., R. 63 E.	61	7.77	105	12	.3	150	4.8	380	16	1	1.2	Mariner and others, 1975
Monte Neva Hot Springs												
sec. 24, T. 21 N., R. 63 E.	79	6.35	52	63	21	16	5.6	303	5.0	26	1.0	Mariner and others, 1975
OREGON												
Harney County												
Alvord (Indian) Hot Springs												
SE, sec. 33, T. 34 S., R. 34 E.	78.5	6.90	128	12	2.2	1000	63	1230	770	180	11	Mariner and others, 1974
Crane Hot Springs												
SW, sec. 34, T.24 S., R. 33 E.	78	8.1	83	3.7	.1	170	3.9	208	79	86	9.0	Mariner and others, 1974
Hickey Hot Springs												
sec. 13, T.33 S., R. 35 E.	86	8.31	214	1.0	.1	550	30	814	240	220	17	Mariner and others, 1974
Trout Creek Hot Springs												
sec. 16,T. 39 S., R. 37 E.	52	6.77	105	18	.8	270	10.8	441	24	204	12.8	Mariner and others, 1974
Unn. Sprs. near Hot Lake												
SE, sec. 36, T. 27 S., R. 29 E.	96	7.30	160	14	.3	450	28	382	250	434	7.2	Mariner and others, 1974
Unn. Sprs. near Harney Lake	68	7.26	92	12	1.8	630	13	568	590	140	3.3	Mariner and others, 1974
Lake County												
Barry Ranch Hot Springs												
SE, sec. 27, T. 39 S., R. 20 E.	88	7.76	130	8.8	.1	280	9.0	236	170	240	5.4	Mariner and others, 1974
Crump												
SW, sec. 34,T. 38 S., R. 24 E.	78	7.26	180	16	.2	280	11	155	240	200	4.9	Mariner and others, 1974
Fisher Hot Springs												
NW, sec. 10, T. 38 S., R. 25 E.	68	7.93	77	8.4	1.0	92	7.9	107	56	59	3.5	Mariner and others, 1974
Hunters Hot Springs												
NW, sec. 04, T.39 S., R. 20 E.	96	7.77	140	13	<.1	210	8.5	81	120	260	4.4	Mariner and others, 1974
Summer Lake Hot Springs												
NE, sec. 12, T. 33 S., R. 17 E.	43	8.43	94	2.1	.1	390	4.6	426	280	120	2.2	Mariner and others, 1974
UTAH												
Beaver County												
Roosevelt Seep												
NWSWSW, sec. 34, T. 26 S., R. 9 W	25	5.60	165	110	22	1800	260	298	3150	110	3.5	Unpub. data, Mariner and others
Roosevelt Steam Well												
NWSWNE, sec. 03, T. 27 S., R. 9 W.	-	5.80 @24°C	590	7.0	0.1	1950	400	200	3400	61	5.7	Unpub. data, Mariner and others
Thermo Hot Springs												
sec. 28, T. 30 S., R. 12 W.	89.5	7.98	113	71	10	380	52	360	255	480	6.6	Mariner and others, 1977
Box Elder County												
Crystal (Madsen's) Hot Springs												
SE, sec. 29, T. 11 N., R. 2 W.	56	7.3	26	784	186	13600	654	371	22600	444	19	Mundorff, 1970
Udy Hot Springs												
NW, sec. 23, T. 13 N., R. 3 W.	43	7.9	26	212	55	2690	118	366	4470	90	1.5	Mundorff, 1970
Juab County												
Baker Hot Springs												
SE, sec. 10, T. 14 S., R. 8 W.	85	7.36 @61°C	66	360	54	860	58	150	1550	720	2.9	Mariner and others, 1977
Millard County												
Meadow Hot Springs												
SW, sec. 26, T. 2 S., R. 6 W.	41	7.5	47	433	144	1040	13.8	408	1800	1130	-	Mundorff, 1970
Salt Lake County												
Becks Hot Spring												
SE, sec. 14, T. 1 N., R. 1 W.	56	7.4	32	746	131	4250	156	221	7470	985	3.3	Mundorff, 1970
Toole County												
Wilson Springs												
sec. 33, T. 10 S., R. 14 W.	61	7.4	33	741	224	7090	18	178	11900	1560	4.0	Mundorff, 1970
Weber County												
Ogden Hot Springs												
SW, sec. 23, T. 6 N., R. 1 W.	56	6.87	52	330	7.2	2700	350	196	4950	93	3.3	Unpub. data, Mariner and others
Utah Hot Springs												
SE, sec. 14, T. 7 N., R. 2 W.	57	6.28	38	890	24	5900	790	206	11500	180	3.7	Unpub. data, Mariner and others

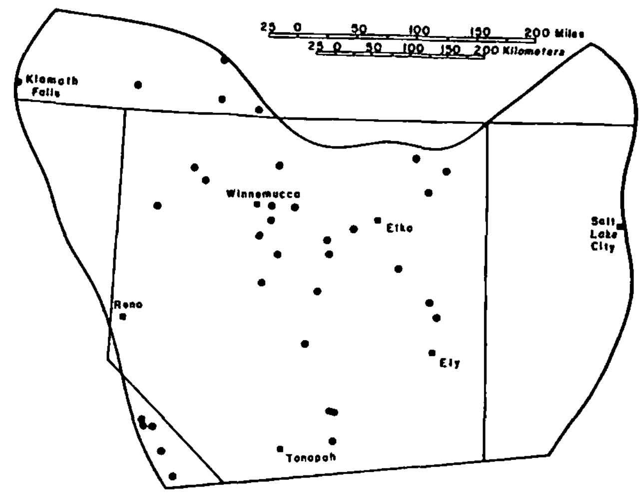

Figure 3. Distribution of Na-HCO_3 waters in the northern Basin and Range Province.

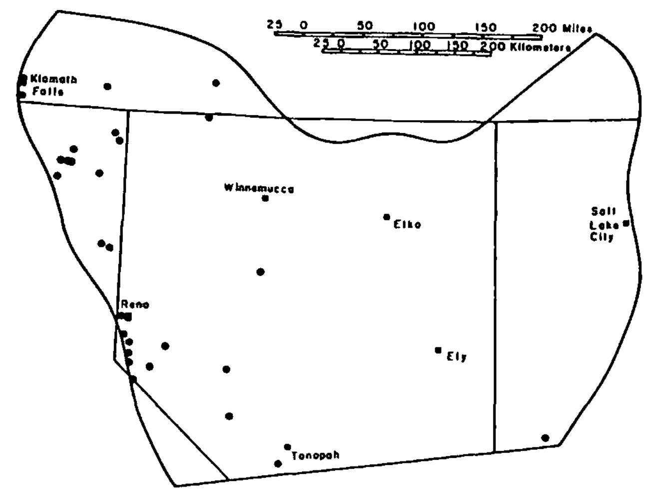

Figure 4. Distribution of Na-SO_4 waters in the northern Basin and Range Province.

discharges sulfate as enriched in oxygen-18 as gypsum of marine origin (+15 o/oo). Either marine gypsum is not available in areas where thermal springs develop or, more likely, the oxygen isotopic composition of the sulfate is being or has been reset in high temperature thermal-environments such as the intrusion of the Sierra Nevada Batholith or the development of mineralized zones. Sulfate-rich waters in some areas, such as northeastern California, are very constant in chemical composition (Table 1). The dissolved sulfate concentration in the water is apparently controlled by the solubility of a common mineral, probably hydrothermal anhydrite.

Mixed anion waters commonly develop in areas where the predominant bedrock or depth of circulation is changing. For instance, the Na-mixed anion waters common in northwestern Nevada west and northwest of the Black Rock Desert, represent a transition zone where rock type and depth of circulation of the thermal waters is changing. Instead of deep circulation in Mesozoic marine strata, which is common to the east and southeast, shallower circulation in predominantly unaltered volcanic rocks of Cenozoic age is more prevalent. The supposition of shallower circulation is evidenced by cooler spring temperatures and lower geothermometer temperatures. Sulfate is a major anion only locally where faults intersect mineralized areas, probably in the Mesozoic rock.

Many of the springs in the Basin and Range Province have been analyzed several times over the last 100 years. These analyses are always slightly different and it is not possible to determine how much of the variation is due to improved analytical techniques and how much to actual changes in the chemical composition with time. Cole (1982) demonstrated that the major constituents of Becks Hot Springs and Wasatch Hot Springs in Utah varied by as much as 15 % during a single year. This variation was caused by dilution on six and three month cycles. Long-term variations certainly occur in many hot springs but these have not been documented.

Chemical Compositions of Gases

Most thermal springs discharge gas along with the water at rates which range from low to high. Chemically these discharges include nitrogen and/or carbon dioxide with lesser amounts of argon, methane, hydrogen, helium, and hydrogen sulfide (Table 2). These gases probably originate from the atmosphere (N_2 and and Ar_2), soil (CO_2), radiogenic processes (He and Ar), and metamorphic or volcanic processes (CO_2). Ratios of nitrogen to argon range from 137/1 to 33/1 although most have N_2/Ar ratios are near the 50/1 expected from an atmospheric source. Organic decay product nitrogen is present in at least one sample (Wedell Spring - N_2/Ar = 131/1). Methane concentrations are low, indicating that breakdown of organic material is contributing relatively little at most springs. Hydrogen concentations are occasionally well above that expected from an atmospheric source and indicate that hydrogen is being generated at depth. Detectable(>0.005 %) hydrogen concentrations occur only where high temperature systems are indicated by geothermometry. Helium concentrations are more than an order of magnitude higher than expected from an atmospheric source in most springs and are almost certainly due to radiogenic decay of uranium, thorium, and/or their daughter products. Carbon dioxide makes up 99% or more of the gas phase in the CO_2-charged slightly to moderately saline Na-HCO_3 ($\pm$ Cl) waters.

Table 2. Compositions of gas discharging from thermal springs, fumaroles and wells in the Northern Basin and Range.

[Compositions reported in volume %]

Name	O_2	Ar	N_2	CO_2	CH_4	C_2H_6	He	H_2	H_2S	Total
CALIFORNIA										
Alpine County										
Unnamed Spr. E. Fk. of the Carson River (9/3/81)	.31	.33	30.51	68.59	.52	<.01	.03	<.005	-	100.29
Lassen County										
Zamboni Hot Springs (9/3/81)	.93	1.26	97.45	.007	.34	<.01	.025	<.005	-	100.01
Mono County										
Hot Creek Gorge (5/29/80)	<.02	.02	1.00	99.83	.02	<.01	.005	.005	<.04	100.88
Fales Hot Springs (11/5/77)	.23	.08	4.27	94.86	<.005	<.05	<.02	<.01	-	99.44
Mammoth Mountain fumarole (7/28/82)	.01	.11	8.27	91.22	.008	<.01	<.005	.010	<.02	99.63
Travertine Hot Springs (11/5/77)	.02	<.02	.15	99.17	<.005	<.05	<0.02	<.01	-	99.34
Unn. Spring S. Mono Lake (11/5/77)	.18	.02	1.05	99.27	<.005	<.05	<0.02	<.01	-	100.52
NEVADA										
Churchill County										
Brady #8 (7/6/79)	.1	1.31	90.14	2.48	2.63	.03	.01	2.94*	<.01	99.54
Dixie Valley well Lamb 4 (5/29/80)	<.02	.04	2.15	97.14	.99	.03	.005	.05	.13	100.80
Elko County										
Hot Sulphur Springs (Tuscarora) (5/30/80)	<.02	.12	3.95	95.54	.34	<.01	.005	.02	-	100.71
Sulphur Hot Springs (Elko) (5/31/80)	<.44	.46	25.31	74.58	.60	<.01	.02	<.005	<.04	100.41
Unnamed Spring on Mary's River (8/11/82)	.03	.91	60.88	37.84	.014	<.01	.170	<.005	-	99.84
Eureka County										
Beowawe (7/31/82)	<.02	1.69	85.40	7.67	4.72	.11	.130	.290	<.02	100.01
Hot Spring near Walti (7/31/82)	.03	.54	27.85	70.71	.47	<.01	.020	<.005	-	99.62
Humboldt County										
Golconda Hot Spring (8/4/82)	.15	.84	39.58	58.09	.97	<.01	.030	<.005	-	99.66
Hot Pot (8/3/82)	.07	.74	48.74	50.06	.16	<.01	.11	<.005	-	99.88
Macfarlanes Hot Spring (8/3/82)	.02	<.02	.09	99.26	.034	<.01	.005	.005	-	99.40
Lander County										
Smith Creek Valley (7/31/82)	.20	1.95	90.31	6.13	1.32	<.01	.050	.045	-	100.00
Mineral County										
Wedell Springs @ (7/30/82)	<.5	.7	92	8.1	.95	<.2	.08	<.05	-	101.83
Pershing County										
Leach Hot Springs (5/--/77)	.03	1.57	73.49	24.53	.90	<.05	.05	<.01	-	100.57

@ Very low pressure sample
* Hydrogen probably from high temperature steam reaction with steel pipe

''Fumaroles'' occur along the west side of Dixie Valley, at Fumarole Butte near Baker Hot Springs in Utah, and at Mammoth Mountain and Casa Diablo in Long Valley, California (Berry and others, 1980). The fumaroles in Dixie Valley in Nevada discharge water vapor mixed with air (Mariner and Evans, unpublished data). The fumarole on Mammoth Mountain discharges mostly CO_2 with minor amounts of nitrogen and traces of hydrogen (Table 2). Chemically the gas discharged by the ''fumarole'' is more like the gases discharged from hot springs in the Hot Creek Gorge part of Long Valley than a fumarole associated with a volcano (analysis of a fumarole on Mt. Hood is included in Table 2 for comparison).

Chemical Geothermometers

The primary use of chemical data in geothermal exploration has been to estimate the temperature of the deep thermal-aquifer associated with a hot spring. The most useful geothermometers for estimating aquifer-temperatures are either the quartz geothermometer of Fournier and Rowe (1966), the Na-K-Ca geothermometer of Fournier and Truesdell (1973), or the Mg-corrected Na-K-Ca geothermometer of Fournier and Potter (1979). Other means of estimating aquifer-temperatures include the sulfate-water isotope geothermometer (McKenzie and Truesdell, 1977), the gas geothermometer (D'Amore and Panichi, 1980), the Na/K geothermometer (Fournier, 1979), and in a very few systems, the solubilities of minerals such as anhydrite (Sakai and Matsubaya, 1974).

Geothermometer temperatures based on the chemical composition of water discharged from hot springs and shallow wells are shown on Table 3. The high temperature areas in the northern Basin and Range Province (>150°C in Table 3) include: Long Valley, Seyferth Hot Springs, unnamed springs on the East Fork of the Carson River in California; Raft River, Idaho; Baltazor Hot Springs, Beowawe, Great Boiling Springs, Hazen, Hot Springs Ranch (Tipton), Hot Sulphur Spings (Tuscarora), Humboldt House, Leach Hot Springs, Lee Hot Springs, Pinto Hot Springs, San Emidio Desert, Soda Lake-Upsal Hogback, Steamboat Springs, Stillwater and Sulphur Hot Springs in Nevada; Alvord Hot Springs, Crump, Hot Lake, Hunters Hot Springs, and Mickey Hot Springs in Oregon; and Roosevelt Hot Springs in Utah. By types of waters, 10 are Na-Cl, 8 are $Na-HCO_3$, 4 are Na-mixed anion waters, and 3 are $Na-SO_4$ waters. '' Successful'' geothermal wells have been drilled at roughly half of the Na-Cl discharging springs but only at one of the $Na-HCO_3$ discharging springs (Table 4). ''Successful'' high-temperature geothermal wells have not been drilled near any of the Na-mixed anion or $Na-SO_4$ springs.

Plots of ^{t}silica and ^{t}Na-K-Ca (Fig.5) show generally good agreement, with relatively few large disparities. Arbitrarily, the quartz

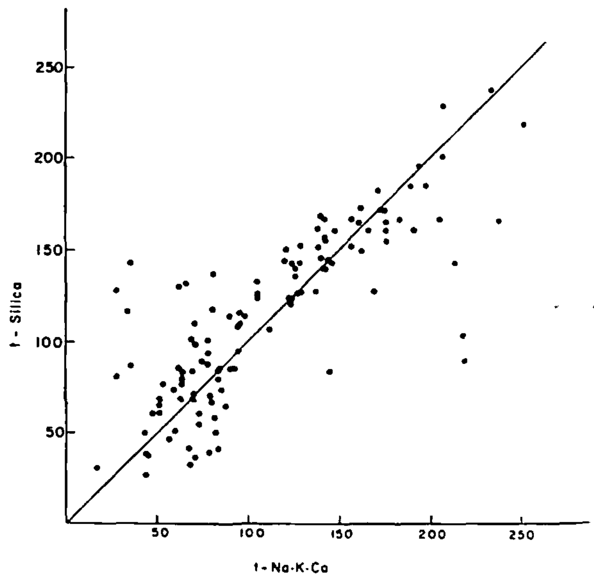

Figure 5. Comparison of temperatures estimated from ^{t}silica and ^{t}Na-K-Ca.

geothermometer was used when the Na-K-Ca geothermometer indicated a temperature of 100°C or more, the chalcedony geothermometer was used when the Na-K-Ca geothermometer indicated a temperature of less than 100°C. The few large disparities occur where waters discharge from silicic tuffs, CO_2-charged waters, waters contaminated with high-chloride saline lake or playa waters, and dilute high-pH waters. Springs issuing from silicic tuffs such as Hot Creek Ranch Springs in Nye County, Nevada and CO_2-charged water such as the water discharged by Travertine and Fales hot springs in California have higher temperatures estimated from the silica geothermometer than from the Na-K-Ca geothermometer. Silica is apparently being taken into solution faster than quartz or chalcedony can be precipitated, supersaturation with respect to quartz or chalcedony is maintained and the quartz (or chalcedony) geothermometer give excessively high subsurface temperature estimates. The magnesium corrected Na-K-Ca geothermometer in these waters generally gives estimated aquifer-temperatures within 25°C of the measured spring temperature. High CO_2 concentrations generally require high temperatures for generation, but these conditions may be very deep (Barnes and others, 1978). Some of the Na-Cl waters issue near saline lakes or playas and may contain some admixed saline lake waters (Utah Hot Springs adjacent to Great Salt Lake is an example). Since the saline water contains almost no calcium or magnesium, abnormally high Na-K-Ca geothermometer temperatures are calculated. The Na/K geothermometer is no better since the proportion of Na to K in the saline water was controlled initially by reactions which included calcium. Finally, the dilute high-pH waters

Table 3. Geothermometer temperatures for springs of the Northern Basin and Range.

[All temperatures in °C.]

Name	Silica	Geothermometers Na-K-Ca	SO_4-H_2O	Measured Surface	Depth
CALIFORNIA					
Alpine County					
Unn. sprs. on the Carson River	172	175		84	
Unn. sprs. on the Carson River	146	140		65	
Lassen County					
Amadee Hot Springs	109	95		96	
Bassett Hot Springs	86	62		79	
Kellog Hot Springs	101	78		78	
Wendell Hot Springs	150	121		96	122
Zamboni Hot Springs	(65)	51		41	
Modoc County					
Hot Springs Motel	110	96	200	98	
Kelly Hot Springs	116	95	198	92	
Leonards Hot Springs	143	124		62	
Little Hot Springs	102	69		79	
Seyforth Hot Springs	143	129	205	85	
West Valley Reservoir	152	129	247	77	
Mono County					
Benton Hot Springs	40	79		56	
Fales Hot Springs	118	81	684	61	
Long Valley-Hot Creek Gorge	161	191	224	90	
Mono Lake - North Shore	94	79		66	
- South Shore	126	28		33	
Travertine Hot Springs	110	71	173	69	
IDAHO					
Bear Lake County					
Bear Lake Hot Springs	55	73		48	
Unn. spring	39	44		42	
Cassia County					
RRGE-1	159	171		145	142
Franklin County					
Maple Grove Hot Springs	77	64		76	
Wayland Hot Springs	125	105		77	
Oneida County					
Woodruff Hot Springs	47	57		27	
NEVADA					
Carson City					
Carson Hot Springs	insufficient data			49	
Pinyon Hills Well	insufficient data			46	
Churchill County					
Brady's Hot Spring (well)	167	157	165	-	188
Dixie Valley Hot Springs	145	144	127	72	
Dixie Federal 52-18	229	207	268	191 gas geot.	
Eagle Salt Works Spring	198	-		-	
Lee Hot Springs	173	162	282	88	
Soda Lake-Upsal Hogback	165	161	127	100	188
Stillwater Area	169	140	177	96	178

Table 3. Geothermometer temperatures for springs of the Northern Basin and Range - continued.

[All temperatures in °C.]

Name	Geothermometers Silica	Na-K-Ca	SO_4-H_2O	Measured Surface	Depth
Douglas County					
Hobo Hot Spring	69	70		46	
Saratoga Hot Springs	31	-		50	
Walley's Hot Spring	80	84		62	
Elko County					
Hot Creek Springs	31	17		26	
Hot Creek Springs	132	66		37	
Hot Hole	86	85		56	
Hot Lake	79	67		18	
Hot Spring near Carlin	90	75		79	
Hot Sulphur Spring	167	183	175	92	117
Humboldt Wells Area	117	34		60	
Mineral (Contact) Hot Springs	127	129		60	
Nile Spring	50	43		43	
Rizzi Ranch Hot Spring	37	71		41	
San Jacinto Ranch Spring	27	44		-	
Sulphur Hot Spring (Ruby Valley)	183	181	161	93	
Smith Ranch	72	71		65	
Trout Creek Ranch Well	33	69		43	
Unnamed spring (Rock Creek)	37	69		35	
Wild Horse Hot Spring	61	73		54	
Wine Cup Ranch Well	insufficient data			59	
Esmeralda County					
Alkali Springs	insufficient data			60	
Silver Peak (Waterworks) Hot Springs	140	142		40	
Eureka County					
Beowawe	196	194	251	98	201
Bruffey's (Mineral Hill) Hot Spring	80	64		66	
Hot Springs Point	87	36		54	
Klobe Hot Springs	69	72		54	
Shipley Hot Springs	61	48		39	
Walti Hot Springs	88	78		72	
Humboldt County					
Baltazor Hot Springs (well)	161	148	158	90	
Bog Hot Springs	65	88		54	
Cordero Mercury Mine well	108	-		60	
Double Hot Springs	140	126		80	
Dyke Hot Springs	128	137		66	
Golconda Area	86	92		74	
Hot Pot	59	81		57	
Hot Spring Ranch (Tipton)	150	162		85	
Howard Hot Springs	71	80		56	
Macfarlane's Bath House Spring	99	71		75	
Pinto Hot Springs	161	176	207	93	
Soldier Meadows Hot Springs	84	64		54	
The Hot Springs	77	54		58	
Well	-	81		70	
Lander County					
Buffalo Valley Hot springs	119	126	140	73	
Hot Springs Ranch(Valley of the Moon)	61	51		53	
South Smith Creek Valley	143	156	143	86	
Spencer Hot Springs	95	95		72	
Unn. spring near Walti Hot Spring	127	129		64	

Table 3. Geothermometer temperatures for springs of the Northern Basin and Range - continued.

[All temperatures in ^{o}C.]

Name	Silica	Geothermometers Na-K-Ca	SO_4-H_2O	Measured Surface	Depth
Lyon County					
Hazen Area	161	166	220	86	
Hind's (Nevada) Hot Springs	74	86		61	
Wabuska Hot Springs	143	146	140	94	108
Wedell Springs	162	139		60	
Nye County					
Darrough's Hot Spring	136	126		95	129
Diana's Punch Bowl	67	80		59	
Hot Creek Ranch Springs	130	62		63	
Upper Hot Creek Ranch Springs	143	36		67	
Warm (Nanny Goat) Spring	81	29		61	
Pershing County					
Colado	128	169		61	155
Humboldt House (Rye Patch Reserv.)	219	252	176	-	156
- artesian well	166	238		77	
Hyder Hot Springs	84	70		78	
Kyle Hot Springs	137	81	154	77	
Leach Hot Springs	155	176	176	92	126
Sou Hot Springs	85	84		70	
Trego Area	124	124		84	
Washoe County					
Bowers Mansion Hot Springs	38	45		46	
Fly Ranch (Ward's)	126	105		80	
Great Boiling Spring	167	205	93	86	
Lawton Hot Springs	84	145		49	
Moana Springs Area (Biglin well)	114	98		85	
San Emidio Desert	185	189		89	115
Steamboat Springs	201	207	207-220	94	208
The Needle Rocks	143	214		56	
White Pine County					
Cherry Creek Hot Springs	114	90		61	
Monte Neva Hot Springs	74	60		79	
OREGON					
Harney County					
Alvord (Indian) Hot Springs	152	157	231	78	
Crane Hot Springs	124	124		78	
Mickey Hot Springs	185	197	273	86	
Trout Creek Hot Springs	140	143	235	52	
Unn. spr. near Hot Lake	165	176	231	96	
Unn. spr. near Harney Lake	133	105		68	
Lake County					
Barry Ranch Hot Springs	152	139		88	
Crump	172	173	202	78	
Fisher Hot Springs	123	123		68	
Hunters Hot Springs	157	143	158	96	
Summer Lake Hot Springs	107	112	189	43	
UTAH					
Beaver County					
Roosevelt Seep	167	142	216	25	208
Roosevelt Steam Well	238	234Na-K	278	208	
Thermo Hot Springs	144	120	142	90	

Table 3. Geothermometer temperatures for springs of the Northern Basin and Range - continued.

[All temperatures in ^{o}C.]

Name	Geothermometers Silica	Na-K-Ca	SO_4-H_2O	Measured Surface	Depth
Box Elder County					
Crystal (Madsen's) Hot Springs	42	84		56	
Udy Hot Springs	42	68		43	
Juab County					
Baker Hot Springs	86	91	22	85	
Millard County					
Meadow Hot Springs	69	63		41	
Salt Lake County					
Beck's Hot Springs	51	82		56	
Toole County					
Wilson Hot Springs	52	60		61	
Weber County					
Ogden Hot Springs	104	218		56	
Utah Hot Springs	90	219		57	

Table 4. Geothermal Systems with Estimated Reservoir-Temperatures >150°C

$Na-HCO_3$ Waters	Na-Mixed Anion Waters	Na-Cl Waters
*Beowawe	Alvord Hot Springs	Crump
Hot Springs Ranch	Hot Lake (Alvord Desert)	Great Boiling Spring
Hot Sulphur Springs (Tuscarora)	Lee Hot Springs	Hazen
	Unn. Springs-Carson River	*Humboldt House
Leach Hot Springs		*Raft River
Long Valley	$Na-SO_4$ Waters	*Roosevelt
Mickey Hot Springs		San Emidio Desert
Pinto Hot Springs	Baltazar Hot Springs	*Soda Lake-Upsal Hogback
Sulphur Hot Springs	Hunters Hot Springs	*Steamboat Springs
(Ruby Valley)	Seyferth Hot Springs	*Stillwater

*Locations of "successful" geothermal wells.

such as Zamboni or Benton hot springs which discharge from granites near the contact of the Basin and Range with the Sierra Nevada contain unusually large silica concentrations due to dissociation of silicic acid (H_4SiO_4 to $H_3SiO_4^-$ and $H_2SiO_4^=$). With one of the computer codes such as SOLMINEQ (Kharaka and Mariner, 1977) the temperature at which the thermal water is in equilibrium with chalcedony or quartz, as appropriate, can be determined. These values are enclosed in parentheses in Table 3.

The apparent agreement between the temperatures estimated from the silica and cation geothermometers in most waters of the Great Basin could be fortuitous. A more important question is, how do the estimated temperatures compare with measured temperatures in geothermal wells? Deep-well temperature data are available for only 15 systems (Table 5). Surprisingly, when measured and estimated temperatures are compared for all 15, the measured temperatures are, on the average, only 14°C cooler than the estimated temperatures. The standard deviation is however,

Table 5. Expected and Measured Temperatures of Geothermal Systems in the Northern Basin and Range.

Name	Temperatures Expected*	Measured
High Discharge Springs		
Beowawe	214	201
Hot Sulfur Springs (Tuscarora)	175	117
Leach Hot Springs	167	126
Long Valley	196	170
Steamboat Springs	205	204
Wendel Hot Springs	136	122
Low Discharge Springs		
Humboldt House	184	156
Roosevelt Hot Springs	175	208
San Emidio Desert	187	115
Wells		
Brady's Hot Springs	162	155
Colado	148	155
Raft River	161	150
Soda Lake	151	188
Stillwater	162	178
Wabuska	143	108

*Average of t_{silica}, t_{cation}, and $t_{SO_4-H_2O}$ when available.

rather large (± 28 C). The estimated temperature for each system is an average which includes values from the quartz or chalcedony geothermometer, as appropriate, the Na-K-Ca geothermometer, and when available, the SO_4-H_2O isotope geothermometer. Surprisingly, the average difference between expected temperatures calculated from geothermometry and the measured temperatures for low discharge rate (<100 L/min) and high discharge rate hot springs are nearly the same (22°C vs 26°C). Normally, high discharge rate springs should give better estimates of subsurface temperature than low discharge rate springs (Fournier and others, 1974). This apparent anomaly may be due, in part, to the small number of systems for which data is available, and, in part, may indicate that the accessible part of the thermal reservoirs may be smaller than previously thought. Long Valley is the best example of an area where recent drilling has shown that the accessible part of the thermal reservoir is smaller than predicted in the last assessment of high-temperature of geothermal resources of the United States (Brook and others, 1978). If geothermal reservoirs are appreciably smaller than anticipated by Brook and others (1978), geothermal development in the northern Great Basin may be seriously curtailed. The six areas where thermal waters were collected from shallow wells (50 - 1000 feet deep) gave remarkably good agreement between estimated and measured temperatures (less than 2°C average difference). The disparity between estimated and observed temperatures is considerably larger for the areas where data is available only from hot springs, measured temperatures in geothermal wells average 24°C cooler than the estimated temperatures.

Isotopic Data

Craig (1963) demonstrated that for high-temperature chloride-rich waters or steam discharges, the deuterium content is approximately equal to that of the local meteoric water. Nehring (1979) reached a similar conclusion in a restudy of the high-temperature chloride-rich system at Steamboat Springs as did Mariner and Wiley (1976) in a study of the Long Valley area. However, other detailed studies of recharge areas for specific systems in the northern Basin and Range Province have generally had difficulty locating cold spring waters in the adjacent highlands which were as depleted isotopically as water discharged by the thermal springs. Welch and others (1981) concluded that the water discharged at Leach Hot Springs was paleo-water which recharged during a colder time-period since it was more depleted in terms of deuterium than any of the water discharged by cold springs in the adjacent mountain ranges. Summit Spring (Table 6; δD = -126.8 o/oo), a perenial spring on the north side of Mt. Tobin near Leach Hot Springs, is as depleted in deuterium as the thermal water discharging at Leach if you assume that the small oxygen shift (+0.3 o/oo $\delta^{18}O$) observed in the cold spring water is due to nonequilibrium evaporation (Fig. 6). This evaporation could have taken place prior to recharge, or in an unconfined aquifer. Although Mt. Tobin is the highest mountain in the region, and could be the recharge area for Leach Hot Springs, this interpretation forces the data to the limits of credibility, only a small summit area on Mt. Tobin exists as a catchment basin and, most damaging, two travertine depositing springs, Buffalo Valley Hot Springs east of Mt. Tobin and Hyder Hot Springs in Dixie Valley south of Mt. Tobin are even more depleted in deuterium (-132 per mil and -133 per mil, respectively).

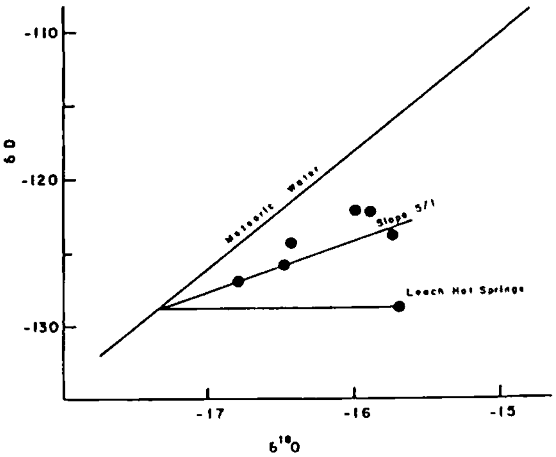

Figure 6. Isotopic composition of thermal and cold waters in the Leach Hot Springs Area.

A detailed study of the Tuscarora Area (Hot Sulphur Springs) at the west end of the Jarbridge Mountains (Bowman and Cole, 1982) also did not find any cold springs as depleted in deuterium as the hot springs. Most of the cold spring water however, appear to have undergone some nonequilibrium evaporation. A study of Ogden and Utah hot springs near Great Salt Lake by Cole (1982) also showed that the thermal waters were more depleted in deuterium than nearby cold springs.

Are these cases the exception or the rule? When the deuterium contents of hot springs in the northern Basin and Range Province (Table 6) are plotted and contoured, the map produced (Fig. 7) is similar to the map for meteoric waters of North America as presented in Taylor (1974). However, based on the map of Taylor (1974), most precipitation in the northern Great Basin should range in deuterium composition from about -110 to -130 o/oo. Most of the hot springs however are in the -120 to -140 o/oo range (figure 7), and a few near Elko are even more depleted (-145 o/oo δD). Appealing to the presence of regional aquifers is not valid since the nearest area with precipitation as depleted as -145 o/oo is in western Montana. Since the isotopic composition of precipitation is a function of temperature, $\delta D = 5.56t - 98.5$ (Dansgaard, 1964), the water being discharged by most of the hot springs in the northern Basin and Range Province apparently recharged during times of colder climate (2 to 3°C colder than at present). Although there have been recent fluctuations in mean annual air temperature with a maximum about 1930, and several recent periods of colder climates (1900-1800, 1700-1575, and 1525-1400) the temperatures were not cold enough to account for the 15 o/oo difference in deuterium observed in most of the Basin and Range. The work of Dansgaard and others (1969) and Johnson and others (1970) has shown that the isotopic composition of precipitation changed from more depleted values to roughly modern values abruptly about 10,000 years ago at the ''end'' of the Pleistocene. Thus most of the water currently discharged by hot springs in the northern Great Basin, apparently recharged at least 10,000 years ago.

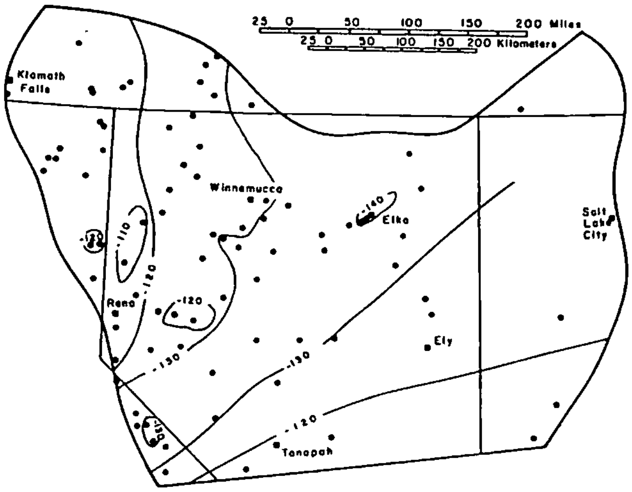

Figure 7. Contour map of the deuterium composition of thermal springs in the northern Basin and Range Province.

Unfortunately, times of circulation for geothermal waters are generally unknown, due in part to the difficulty and uncertainty in interpreting tritium or carbon-14 values. Tritium has a half-life of only 12.3 years and generally concentrations of tritium in thermal waters are less than 0.1 TU. This merely shows that the waters are more than 60 years old (Barnwell, 1963; Wilson, 1963). Carbon-14 has a half-life of 5,670 years and should be much more useful than tritium. However, carbon-14 readily exchanges with reservoir carbon at high temperatures (Craig, 1963).

In contrast to deuterium, which generally does not change in concentration during passage through a geothermal system (boiling excepted), the oxygen isotopic composition of the water

Table 6. Isotopic data for thermal springs of the northern Basin and Range.

	del D	del 18O
CALIFORNIA		
Alpine County		
Unn. sprs. E.Fk. Carson River		
(84°C spring)	-126.5	-15.56
(65°C spring)	-125.3	-15.54
Lassen County		
Amadee Hot Springs	-120.0	-
Bassett Hot Springs	-115.1	-13.54
Kellog Hot Springs	-115.5	-14.09
Wendell Hot Springs	-120.8	-14.04
Zamboni Hot Springs	-118.1	-15.34
Modoc County		
Lake City Mud Eruption	-113.0	-14.79
Kelly Hot Springs	-115.1	-13.54
Little Valley Hot Springs	-116.9	-14.20
Hot Springs Motel (well)	-117.0	-13.81
Menlo Hot Springs	-112.3	-15.30
Seyforth	-121.2	-14.05
Mono County		
Benton Hot Springs	-135.5	-17.46
Fales Hot Springs	-132.8	-17.46
Long Valley (Hot Creek Gorge)	-120.3	-14.83
- LDMW	-115 to -130	
Mono Lake - North Shore	-126.6	-16.91
- South Shore	-126.9	-15.69
The Hot Springs	-137.3	-16.29
Travertine Hot Springs	-139.3	-16.64
NEVADA		
Churchill County		
Brady's Hot Spring (well)	-121.2	-14.22
Dixie Federal 52-18	-133.9	-14.72
Dixie Valley Hot Springs	-126.1	-15.89
- LDMW	-120.0	-15.22
Lee Hot Springs	-125.8	-13.34
Stillwater Area (well)	-110.0	-12.36
Douglas County		
Walley's Hot Springs	-119.5	-15.55
Elko County		
Hot Creek	-126.7	-16.28
- LDMW	-121.4	-15.69
Hot Creek Springs	-135.7	-17.40
- LDMW	-128.9	-16.20
Hot Hole	-144.7	-15.31
Hot spring near Carlin	-132.7	-16.64
Hot Sulfur Spring (Tuscarora)	-138.6	-16.65
Humboldt Wells	-134.7	-
- LDMW	-122.1	-15.81
Mineral (Contact) Hot Spring	-139.0	-17.61
Nile Spring	-139.1	-18.24
Smith Ranch Hot Spring	-132.8	-16.24
Sulphur Hot Spring (Ruby V.)	-130.1	-16.09
- LDMW	-124.6	-16.87
Sulphur Hot Spring (Elko)	-145.9	-17.67

Table 6. Isotopic data for thermal springs of the northern Basin and Range--continued.

	del D	del 18O
Esmeralda County		
Silver Peak (Water Works spr.)	-118.2	-13.50
Eureka County		
Beowawe	-130.0	-14.76
Hot Springs Point	-136.1	-15.97
Klobe Hot Springs	-127.9	-16.28
Walti Hot Springs	-129.8	-16.87
Humboldt County		
Baltazor Hot Springs	-125.3	-15.26
Bog Hot Springs	-124.3	-15.30
Double Hot Springs	-128.8	-15.93
Dyke Hot Springs	-128.0	-16.29
Hot Pot	-136.7	-16.70
Hot Springs Ranch (Tipton)	-131.4	-15.74
Howard Hot Spring	-127.1	-16.17
Macfarlanes Hot Springs	-127.2	-12.54
Pinto Hot Springs	-129.2	-14.48
Soldier Meadows Hot Springs	-129.2	-16.56
The Hot Springs	-134.6	-16.44
Lander County		
Buffalo Valley Hot Springs	-131.6	-15.85
	-135.2	-13.61
- LDMW	-117.3	-14.95
Hot Springs Ranch (Valley of the Moon)	-127.8	-16.28
Smith Creek Valley	-130.4	-16.68
Spencer Hot Spring	-135.8	-16.01
Unn. Spr. (Grass V. nr Walti)	-134.8	-16.73
Lyon County		
Hazen Area	-121.5	-13.30
Hinds (Nevada) Hot Springs	-123.2	-16.01
Wabuska Hot Springs	-129.7	-15.38
Wedell	-131.9	-15.90
Mineral County		
Soda Springs	-130.3	-16.13
Nye County		
Diana's Punchbowl	-124.9	-16.24
Darroughs Hot Springs	-122.5	-15.50
Pershing County		
Colado	-125.5	-14.01
Hyder Hot Springs	-133.2	-15.66
Humboldt House - deep well	-130.6	-14.64
- shallow well	-127.2	-14.09
- LDMW	-119.9	-15.25
Kyle Hot Springs	-130.0	-15.50
- LDMW	-121.1	-14.71
Leach Hot Springs	-128.6	-15.70
Summit Spring - LDMW	-126.8	-16.80
Trego Hot Springs	-124.5	-14.40

Table 6. Isotopic data for thermal springs of the northern Basin and Range--continued.

	del D	del 18O
Washoe County		
Bowers Mansion Hot Springs	-102.3	-14.79
Fly Ranch	-120.7	-14.72
Great Boiling Springs	-100.5	-10.83
San Emidio Desert	-108.3	-12.05
Steamboat Springs	-116.7	-12.16
The Needle Rocks	-106.5	- 6.33
White Pine County		
Cherry Creek Hot Springs	-127.8	-16.20
Monte Neva Hot Springs	-127.8	-16.68
OREGON		
Harney County		
Alvord (Indian) Hot Springs	-123.6	-13.23
Crane Hot Springs	-133.3	-16.17
Mickey Hot Springs	-124.3	-13.42
Pike Creek - LDMW	-108.4	-14.05
Trout Creek Hot Spring	-127.4	-16.17
Trout Creek- LDMW	-115.3	-15.50
Unn. Sprs. near Hot Lake	-125.4	-14.36
Unn. H. Sprs near Harney Lake	-128.5	-
Lake County		
Barry Ranch Hot Springs	-119.4	-13.72
Crane Creek- LDMW	-101.2	-13.40
Crump	-115.5	-13.28
Deep Creek - LDMW	-106.6	-13.46
Fisher Hot Springs	-117.0	-
Hunters Hot Springs	-119.0	-14.32
Summer Lake Hot Springs	-115.0	-13.32
Malheur County		
Unn. Spr. nr. McDermitt	-134.6	-16.95
UTAH		
Beaver County		
Roosevelt Seep	-113.0	-12.95
Thermo Hot Springs	-118.3	-14.32
Steam Well at Roosevelt H.S.	-115.9	-12.99
Juab County		
Crater (Baker, Abraham) Hot S.	-126.3	-16.0

changes due to exchange with minerals in the confining rock (Craig, 1963). In the northern Great Basin, dilute $Na\text{-}HCO_3$ waters and $Na\text{-}SO_4$ waters generally have less oxygen enrichment than Na-Cl waters (Figs. 8, 9, and 10). Bicarbonate-rich waters, however, occasionally have very large oxygen shifts (up to 4 o/oo; Fig. 8). Carbonates in limetsones are usually +20 to +30 per mil in $\delta^{18}O$ while silicates in most igneous rocks range from +5 to +15 o/oo. Larger oxygen shifts are observed in water associated with limestones than in waters associated with silicate rocks because of the concentration difference and the faster exchange rates between carbonates and waters. The lack of correlation between oxygen shift and amount of dissolved solids in the Na-Cl waters is an indication that although these waters generally occur where most water-rock reaction has taken place, the chloride concentration is at least, in part, a function of chloride availability.

The oxygen isotopic compositions of a few sulfates in thermal waters of the Basin and

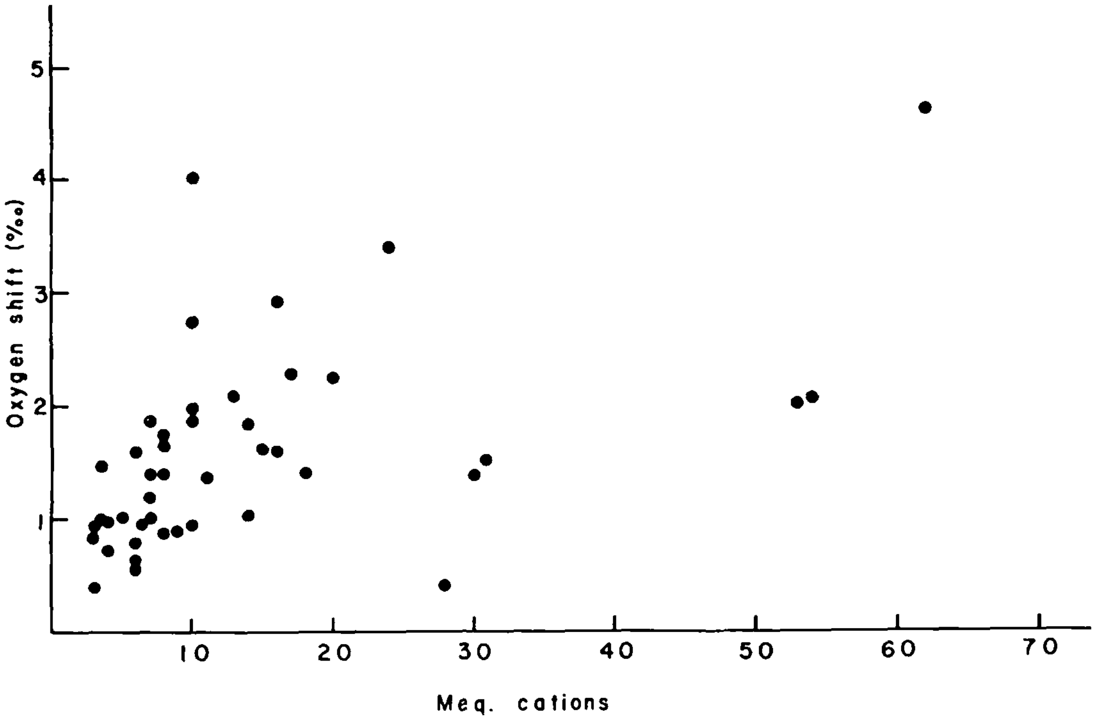

Figure 8. Oxygen shift of HCO_3-rich thermal water as a function of milliequivalents cations (specific conductivity).

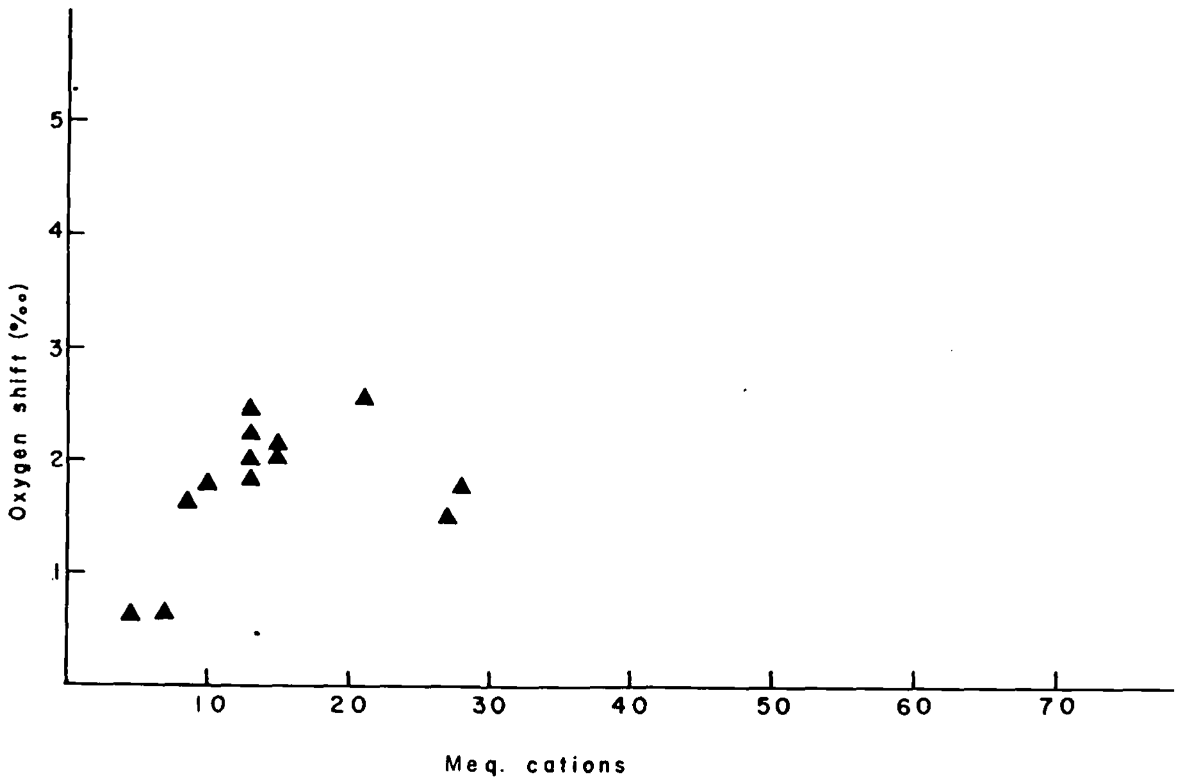

Figure 9. Oxygen shift of SO_4-rich thermal waters as a function of milliequivalents cations (specific conductivity).

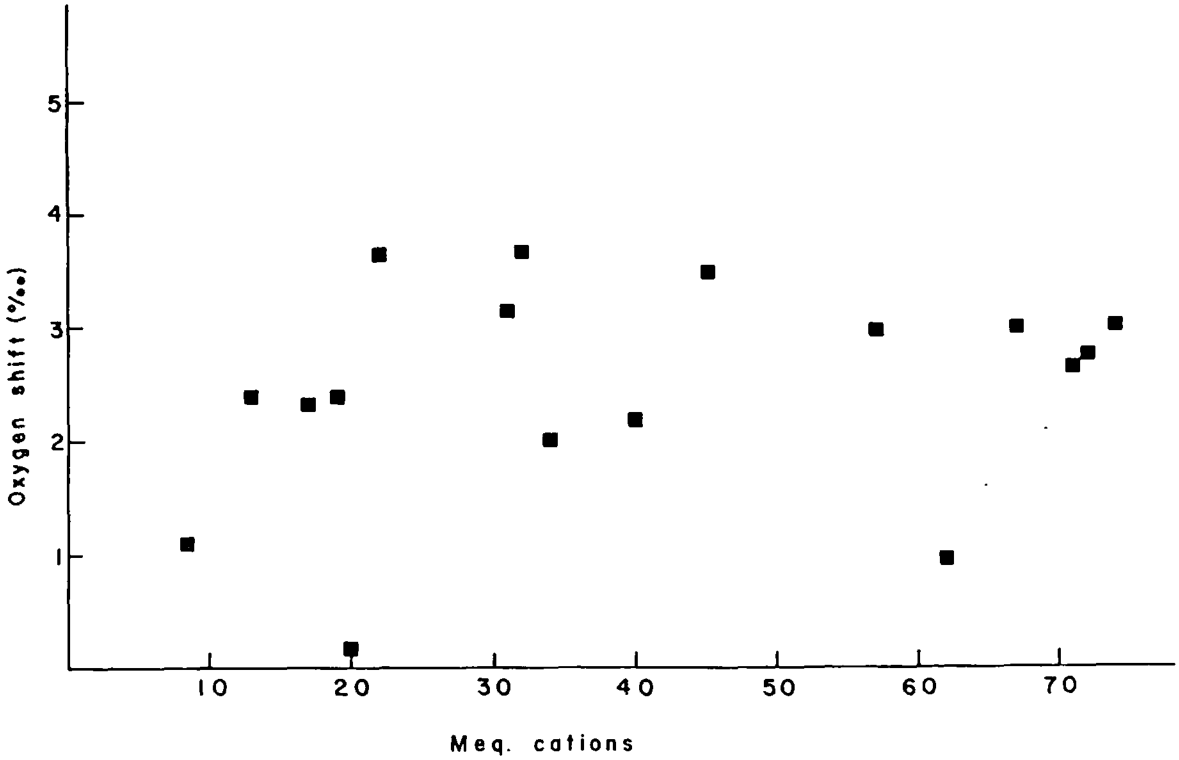

Figure 10. Oxygen shift of Cl-rich thermal waters as a function of milliequivalents cations (specific conductivity).

Range Province have been reported by Nehring and Mariner (1979). Our intent was to estimate the thermal-aquifer temperature using the sulfate-water isotope geothermometer of McKenzie and Truesdell (1977). The sulfate-water isotope geothermometer generally gives calculated temperatures that are slightly hotter than those obtained from the quartz or Na-K-Ca geothermometers. However, in some areas, the sulfate-water isotope geothermometer gives temperatures considerably lower than the measured surface temperatures or considerably higher than the temperatures estimated from the other geothermometers (Table 3). It is possible for sulfate to be dissolved from the country rock after the thermal fluid leaves the thermal aquifer or the thermal fluid may mix with sulfate-rich nonthermal water before it discharges at the surface. This added sulfate probably will have a different original isotopic composition and could significantly change the isotopic composition of the total sulfate in the discharge.

In the northern Basin and Range Province, the sulfate-water isotope geothermometer indicates aquifer-temperatures which are generally 50 to 100°C higher than those estimated from the silica or cation geothermometers. A possible explanation for these higher apparent temperatures is that the depleted sulfate is from dissolution of minerals formed during previous high temperature hydrothermal or metamorphic events. However, these minerals must be situated along the flow path from the reservoir to the surface or the residence times of fluids in the thermal reservoir must be very short. The latter possibility does not appear likely due to the old apparent age of most of the waters. However, in mineralized areas, there is the possibility of dissolving isotopically depleted sulfate which does not have enough time to attain equilibrium with the dissolving fluid. For example, pickeringite (ideal formula $MgAl_2(SO_4)_4\ 22H_2O$) from a site near Lahontan Reservoir west of Fallon was −6.53 o/oo $\delta^{18}O$. Dissolution of such a mineral without concomitant reequilibration would result in excessively high apparent SO_4-H_2O isotopic equilibrium temperatures. Sulfate minerals are often associated with ore deposits in western Nevada and so isotopically depleted sulfate is readily available.

Alternatively, although the sulfate-water isotope temperatures are no closer to the measured temperatures in the deep wells then the temperatures calculated from the cation or silica geothermometers, the apparent sulfate-water equilibrium temperatures could be correct. Calculations with the computer code SOLMNEQ (Kharaka and Barnes, 1973, as modified by Kharaka and Mariner, 1977) indicate saturation with respect to anhydrite ($CaSO_4$) at temperatures near those estimated from the ulfate-water isotope geothermometer in several f the systems in northeastern California (Table). This may be an indication that the emperatures calculated from the sulfate-water sotope geothermometer are accurate in this rea. The cooler temperatures estimated from he quartz and Na-K-Ca geothermometer may ndicate that chemical equilibrium was pproached in a shallow aquifer at temperatures ear the spring temperature but the isotopic omposition remained unchanged. Anhydrite aturation temperatures (Table 7) for Travertine ot Springs, Hot Lake (Oregon), Stillwater, Soda ake-Upsal Hogback, Hazen and Alvord Hot Springs re also reasonably near the sulfate-water iso- ope equilibrium temperatures. The differences n temperatures estimated from theoretical nhydrite saturation (175°C) and sulfate-water sotopic data (127°C) at the Soda Lake-Upsal ogback area may indicate that reequilibration r mixing has taken place in a shallow aquifer ince deep wells have encountered temperatures ore than 50°C hotter. Lee Hot Springs, located outh of Fallon, has an apparent anhydrite aturation temperature of only 173°C, almost 00°C cooler than the sulfate-water isotopic quilibrium temperature. The sulfate at Lee ust be from a near surface hydrothermal mineral ource. At the other extreme, Abraham Hot prings in Utah had a sulfate-water isotopic quilibrium temperature less than the measured pring temperature. Apparently, the sulfate discharged at Abraham Hot Springs is of marine origin (initially about +15 o/oo in $\delta^{18}O$) and it never attained, isotopic equilibrium with the thermal water.

<u>Summary</u>

Thermal waters in the Basin and Range Province range from dilute $Na-HCO_3$ and $Ca-HCO_3$ waters to very saline Na-Cl waters. The most saline Na-Cl waters occur near Great Salt Lake or near the sinks and playas of northwestern Nevada. Slightly saline CO_2-charged $Na-HCO_3$ waters are common near the Sierra Nevada. $Na-SO_4$ ($\pm$ Cl) waters occur in northeastern California and western Nevada. Sulfate in these waters may be from sulfate minerals, initially deposited during previous hydrothermal events.

Meteoric waters in the probable recharge areas for most hot springs in the Northern Basin and Range Province are generally not as depleted in deuterium as the waters currently discharged by the hot springs. This difference is largest for waters associated with travertine and is almost certainly an indication that the thermal waters recharged during times of colder climate, probably the Pleistocene.

Measured temperatures in deep wells are, on the average, 14°C cooler than expected from the chemical geothermometry on waters from nearby

Table 7. Anhydrite saturation temperatures and sulfate-water isotope equilibrium temperatures for thermal waters of the northern Basin and Range.

Name of Sample	T - Anhydrite Saturation	T - SO_4-H_2O
California		
Fales Hot Springs	310	184
Hot Springs Motel (Surprise Valley)	211	200
Kelly Hot Springs	193	198
Seyferth Hot Springs	189	205
Travertine Hot Springs	190	173
West Valley Reservoir (Hot Spring)	220	247
Nevada		
Hazen (Hot Springs)	185	220
Lee Hot Springs	173	282
Soda Lake - Upsal Hogback (well)	125	127
Stillwater (well)	180	177
Wabuska (shallow well)	176	140
Oregon		
Alvord Hot Springs	275	231
Spring near Hot Lake (Alvord Desert)	215	230
Utah		
Abraham (Baker) Hot Springs	159	22
Thermo Hot Springs	172	142

hot springs and shallow wells. The measured temperatures in the geothermal reservoirs are, on the average, 22°C cooler than expected when spring waters are used to estimate the deep aquifer-temperature, however measured temperature are only 2°C lower than expected when waters from shallow wells are used to estimate the temperature of the deep aquifer.

Anhydrite saturation temperatures are similar to sulfate-water isotope equilibrium temperatures for the more saline thermal waters of the northern Basin and Range Province. However, some systems in northeastern California have aquifer-temperatures of 200 to 220 C based on anhydrite saturation and SO_4-H_2O isotopic equilibrium temperatures. These temperatures are roughly 100°C above the temperatures estimated from silica or Na-K-Ca geothermometers.

The oxygen-18 enrichment of Na-HCO_3 thermal waters generally increases as the total dissolved solids increase. Concentrations of dissolved solids in Na-Cl waters generally are higher than either Na-HCO_3 or Na-SO_4 waters. Although Na-Cl waters are generally more enriched in oxygen-18, no correlation seems to exist between their oxygen-18 enrichment and amount of dissolved solids.

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BOREHOLE GEOPHYSICAL TECHNIQUES FOR DEFINING PERMEABLE ZONES IN GEOTHERMAL SYSTEMS

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ABSTRACT

Borehole electrical geophysical methods have considerable potential for helping to define hot and permeable zones in geothermal systems. Borehole geophysics differs from geophysical well logging and has a much greater area of search around a borehole. Very little developmental work has taken place in borehole electrical methods to date. At UURI, we have been developing computer methods to model various electrical arrays for borehole configurations. We plan to compare the several possible survey methods and then design a field system based on the method that appears from the computer studies to be optimum.

From our studies to date we tentatively conclude that the cross-borehole method produces larger anomalies than does the single-borehole method; cross-borehole anomalies using a pole-pole array are smaller than those for a dipole-dipole array; the cross-borehole mise-à-la-masse method produces larger anomalies than does other cross-borehole methods; and, the anomalies due to a thin structure are generally much smaller than those for a sphere, as is to be expected.

INTRODUCTION

The key problem worldwide in development of hydrothermal resources appears to be more in locating permeable zones than in locating high temperatures. Grindly and Browne (1976) note that of 11 hydrothermal fields investigated in New Zealand, all of which have high temperatures (230°C to 300°C), five are non-productive chiefly because of low permeability. Three of the eleven fields are in production (Wairakei, Kawerau and Broadlands) and in each of these fields permeability limits production more than temperature does. Hot but unproductive holes have been drilled at many of the major geothermal areas in the world, including The Geysers, Roosevelt Hot Springs, Coso, and Meager Creek, to name a few.

Permeability can be primary or secondary. Primary permeability in clastic rocks originates from intergranular porosity and it generally decreases with depth due to compaction and cementation. In volcanic sequences, primary intergranular porosity and permeability exist, but greater permeability exists in open spaces at flow contacts and within the flows themselves. Primary permeability in crystalline igneous rocks is generally very low. Secondary permeability occurs in all rock types in open fault zones, fractures and fracture intersections, along dikes and in breccia zones (Brace, 1968; Moore et al., 1985). Changes in permeability come about through mineral deposition in open spaces or by leaching by the thermal fluids.

Although none of the geophysical methods maps permeability directly, any geological, geochemical, or hydrological understanding of the factors that control the permeability in a geothermal reservoir can be used to help determine geophysical methods potentially useful for detecting the boundaries and more permeable parts of a hydrothermal system. At UURI, we have been developing electrical borehole techniques to detect and map permeable zones in the subsurface, especially fractures.

BACKGROUND--BOREHOLE GEOPHYSICS

It is important to understand the differences between geophysical well logging and borehole geophysics. In geophysical well logging, the instruments are deployed in a single well in a tool or sonde, and the depth of investigation is usually limited to the first few meters from the wellbore. Well-logging techniques have been developed by the petroleum industry over a period of half a century and have been applied with variable success by the geothermal industry. The major adaptations to the geothermal environment are the requirements of high temperature tools and the different interpretation required for hard rock (volcanic, igneous) lithologies. Other differences include a strong emphasis in geothermal exploration on fracture identification and the effects of hydrothermal alteration upon certain log responses. Much research remains to be done in order to understand fully the responses of various well logs in geothermal reservoirs and their typically fractured, altered, commonly igneous and metamorphic host rocks. In spite of the relative lack of knowledge of well-log response in geothermal reservoirs, several logs or log combinations have been used successfully to investigate such properties as lithology, alteration, fracturing, density, porosity, fluid flow and sulfide content, all of which may be critical in deciding how and in what intervals to complete, case, cement or stimulate a well (Glenn and Hulen, 1979; Keys and Sullivan, 1979; Sanyal et al., 1980; Glenn and Ross, 1982; Halfman et al., 1982).

By contrast, borehole geophysics refers to those geophysical techniques where energy sources and sensors are deployed (1) at wide spacing in a single borehole, (2) partly in one borehole and partly on the surface, or (3) partly in one borehole and partly in a second borehole. Thus, we speak of borehole-to-surface, surface-to-borehole and borehole-to-borehole surveys. The depth of investigation is generally much greater in borehole geophysical surveys than it is in geophysical well logging.

Only one of the several borehole geophysical techniques, namely vertical seismic profiling (VSP), has been developed to any extent. The petroleum industry has funded relatively rapid development of VSP over the past several years.

VSP

Vertical seismic profiling (VSP) can be done using both P- and S-wave surface sources (usually mechanical vibrators) arranged circumferentially around a well. Direct and reflected seismic waves are detected by strings of down-hole geophones clamped to the wall of the well or by hydrophones. VSP has been used mainly to trace seismic events observed at the surface to their point of origin in the earth and to obtain better estimates for the acoustic properties of a stratigraphic sequence. Oristaglio (1985) presents a guide to the current uses of VSP.

Borehole Electrical Techniques

Borehole-to-borehole and borehole-to-surface electrical methods appear to have considerable potential for application to geothermal exploration. In a benchmark introductory paper, Daniels (1983) illustrated the utility of hole-to-surface resistivity measurements with a detailed study of an area of volcanic tuff near Yucca Mountain, Nevada. He obtained total-field resistivity data for a grid of points on the surface with current sources in three drill holes, completed a layered-earth reduction of the data, and interpreted the residual resistivity anomalies with a 3D ellipsoidal modeling technique.

The borehole electrical techniques, however, are in general poorly developed. One reason for this is that there are a large number of ways that borehole electrical surveys can be performed and it has been unclear which methods are best. At the same time, computer algorithms to model the several methods have not existed so that it has not been possible to select among methods prior to committing to the expense of building a field system and obtaining test data.

R&D PROGRAM AT UURI

The objective of our program is to develop and demonstrate the use of borehole electrical techniques in geothermal exploration, reservoir delineation and reservoir exploitation. Our approach is:

1. Develop computer techniques to model the possible borehole electrical survey systems;
2. Design and construct a field data acquisition system based on the results of (1);
3. Acquire field data at sites where the nature and extent of permeability are known; and,
4. Develop techniques to interpret field data.

To the present time, we have made considerable progress on item (1) above and we are now at such a point that item (2) could be started.

Our research staff has consisted of the following personnel: Stanley H. Ward, Project Manager; Luis Rijo, Professor of Geophysics, Universidade Federal Do Para, Brazil (on 2-year post-doctoral leave at U of U and UURI); F. W. Yang, Peoples Republic of China (visiting scholar); J. X. Zhao, Peoples Republic of China (visiting scholar); Craig W. Beasley (doctoral candidate U of U, awarded MS degree); Richard C. West (MS candidate at UU). Additional technical support has been provided by Philip E. Wannamaker, Howard P. Ross and Phillip M. Wright of UURI and by Gerald W. Hohmann of U of U. Project costs for Rijo, Yang and Zhao have been minimal because these scientists have been supported by their governments. Thus, a great deal has been accomplished at minimal cost while supporting the education of several students. The remainder of this paper will discuss the significance of our research to date.

COMPUTER MODELING OF BOREHOLE ELECTRICAL METHODS

Computer techniques for modeling borehole electrical geophysics have largely been lacking, especially for three-dimensional (3D) cases. Figure 1 indicates conventional usage of the terms 1D, 2D and 3D in geophysical interpretation. In the 1D case, also called the "layered earth" case, the physical property of interest (resistivity for this study), varies only in the vertical direction. In the 2D case, physical property variations in the vertical and one horizontal dimension are allowed, and the anomalous body illustrated has the same shape in and out of the paper for infinite distance. In the 3D case, physical property variations are specified in all three space dimensions. Obviously, the real earth is only occasionally 1D in nature in geothermal areas. The usual case is for physical properties to vary in all three dimensions in the earth, the 3D case. However, the mathematical formulations for electrical anomalies of bodies increase greatly in complexity from the 1D case to the 3D case. This accounts for the fact that in order to begin our task of applying borehole electrical techniques to delineation of permeability, we were required to develop original mathematical formulations of the problem.

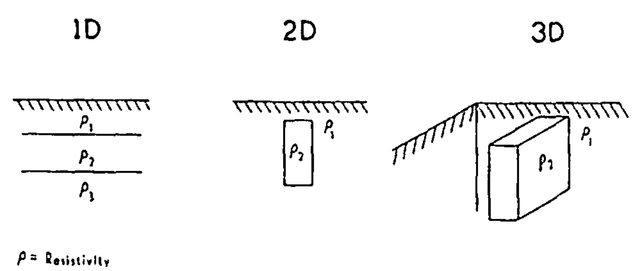


FIGURE 1
Illustration of the meaning of the terms 1D, 2D and 3D in geophysical modeling.

Thick-Body Studies

Prior to 1982, only three published papers considered computer modeling of downhole electrodes for three-dimensional bodies. Daniels (1977) studied six buried electrode configurations and plotted normalized apparent resistivity or apparent polarizability against such configuration parameters as 1) source and receiver depth, 2) depth/bipole length, 3) receiver distance from body, 4) depth of body, and 5) distance of source and receiver from body center. Snyder and Merkel (1973), computed the IP and apparent resistivity responses resulting from a buried current pole in the presence of a buried sphere. Their plots are center-line profiles for normalized apparent resistivity and normalized IP response. Dobecki (1980) computed the effects of spheroidal bodies as measured in nearby single boreholes using the pole-pole electrode array. These three studies are obviously very limited in terms of the problems of defining permeability in geothermal systems.

In 1982, Newkirk (1982) from our group published a study of downhole electrical resistivity with 3D bodies. Using a numerical modeling technique described by Hohmann (1975), theoretical anomalies due to a three-dimensional body composed of simple prisms were computed. The results were presented in terms of 1) the potential, 2) the apparent resistivity calculated from the total horizontal electric field and 3) the apparent resistivity calculated from the potential. Two electrode configurations were considered for each model. Each configuration consisted of a pair of electrodes, where one of the electrodes was remote and the second electrode was located either in the body, for mise-à-la-masse or applied potential, or outside the body, simulating a near miss. Newkirk's computer program was used by Mackelprang (1985) of our group to compute a catalog of models due to bodies that might be of interest in detection of thick fracture zones.

Figures 2a and 2b show the conventions used by Newkirk (1982) and Mackelprang (1985) in calculations of the effects of 3D bodies. The bodies are buried in a homogeneous earth and two of many options for a downhole point electrode are illustrated. Figure 3a and 3b illustrate anomalies on a surface resistivity survey produced by a narrow conductive body buried at a depth of 7 units with the electrode in the body (Fig. 3a) and off the end of the body (Fig. 3b). The peanut shaped anomaly shown in Figure 3a is particularly characteristic on surface resistivity surveys with the borehole electrode in the body.

One basic shortcoming of Newkirk's (1982) algorithm is that it does not apply when the anomalous body becomes thin, i.e. to the case of delineation of fractures or thin fracture zones. To address this important problem, the thin-body studies described in the next section have been undertaken.

Thin-Body Studies

These studies are aimed at targets simulating fracture zones which are thin relative to their other two dimensions. For the most part, we have standardized the aspect ratios of the target dimensions at 10:10:1. While the effect of varying the contrast in resistivity has been examined,

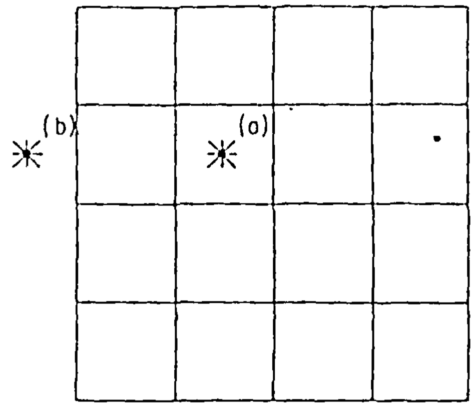


FIGURE 2a
Plan view of standard model.

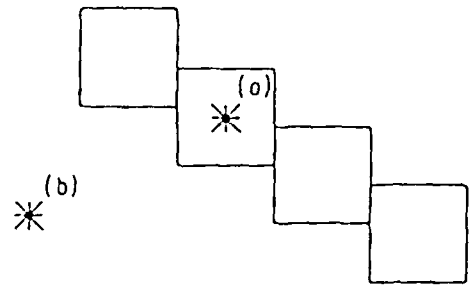


FIGURE 2b
Cross-section view of standard model.

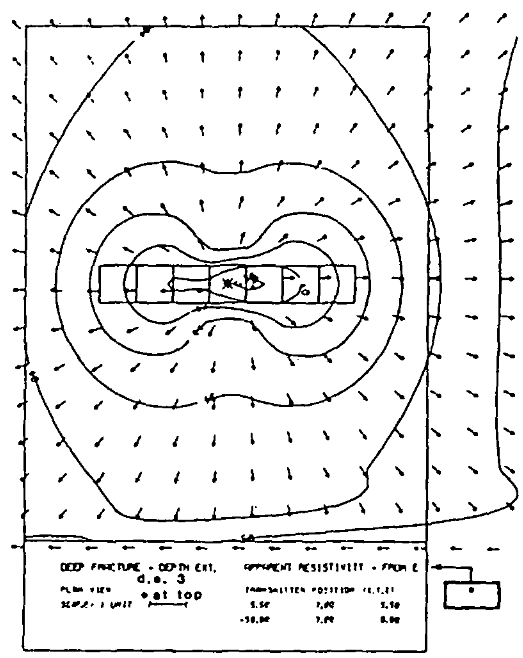


FIGURE 3a
Surface resistivity anomaly due to deep fracture with downhole electrode in body.

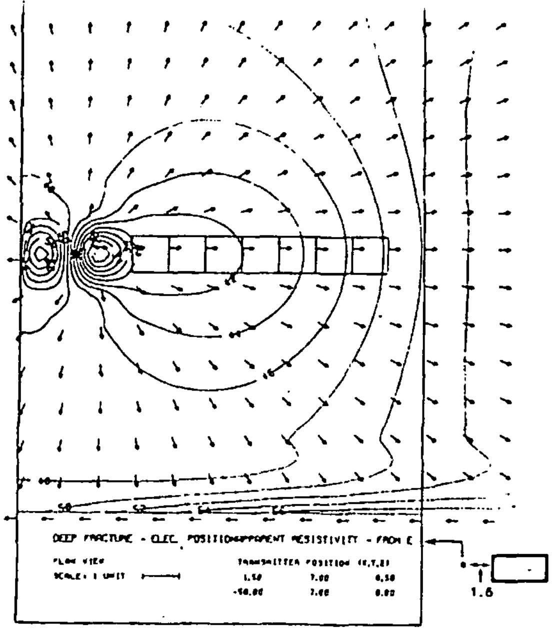


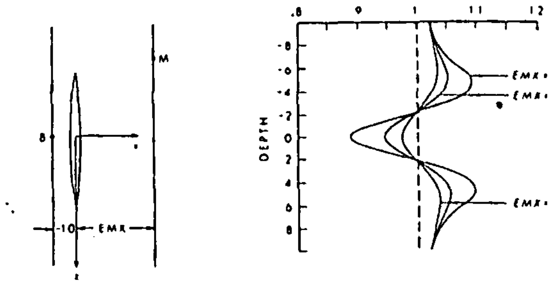
FIGURE 3b
Surface resistivity anomaly due to deep fracture with downhole electrode at side of body.

most of the results are for the case of a fracture zone ten times more conductive than the host rocks.

Four numerical techniques have been utilized in the studies; three have been applied with the D.C. resistivity method. The techniques applied to the resistivity problem are (1) a 3D surface integral equation (Yang and Ward, 1985a,b), (2) a 3D volume integral equation (Beasley and Ward, 1986), and (3) a 2D finite element method (Zhao et al., 1985). A solution for the time domain EM method has also been obtained which uses a 3D volume integral equation formulation (West and Ward, 1985). Elaboration on these four approaches is given below.

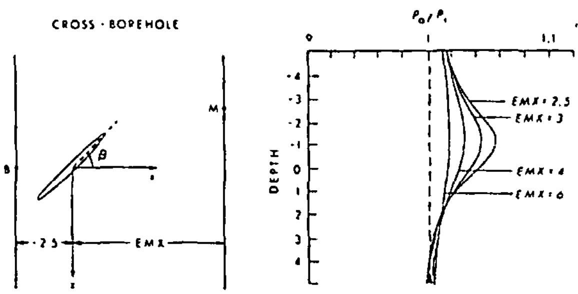
Yang and Ward (1985a,b) present theoretical results relating to the detection of thin oblate spheroids and ellipsoids of arbitrary attitude. The effects of the surface of the earth are neglected and the body is assumed to be enclosed within an infinite homogeneous mass. The surface of the body is divided into a series of subsurfaces, and a numerical solution of the Fredholm integral equation is applied. Once a solution for the surface charge distribution is determined, the potential can be specified anywhere by means of Coulomb's law. The theoretical model results indicate that cross-borehole resistivity measurements are a more effective technique than single-borehole measurements for delineating resistivity anomalies in the vicinity of a borehole.

Figure 4a shows cross-borehole resistivity responses of a vertical conductive fracture zone between two boreholes. The electrode configuration is the pole-pole array with electrode B fixed and electrode M moving in the second borehole. Several curves are plotted depending on the distance between the fracture and the second borehole. The larger anomalies occur when the second



Electrode Configuration	Body Size	Angles	Resistivity Contrast
Fixed Source Moving Electrode SBX=-10 SBY=SBZ=0 EMY=0	a=6(z) b=2(y) c=0.2(x)	$\alpha = 0°$ $\beta = 90°$ $\gamma = 0°$	$\frac{\rho_2}{\rho_1} = 0.10$

FIGURE 4a
Downhole cross-borehole resistivity anomalies for vertical fracture showing effect of varying distance from fracture to second borehole.

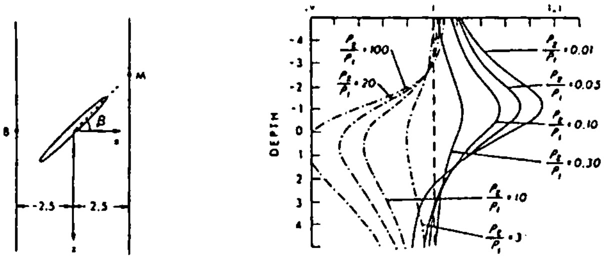


Electrode Configuration	Body Size	Angles	Resistivity Contrast
Fixed Source Moving Electrode SBX=-2.5 SBY=SBZ=0 EMY=0	a=2 b=2 c=0.2	$\alpha = 0°$ $\beta = 45°$ $\gamma = 0°$	$\frac{\rho_2}{\rho_1} = 0.10$

FIGURE 4b
Downhole cross-borehole resistivity anomalies for dipping fracture showing effect of varying distance from fracture to second borehole.

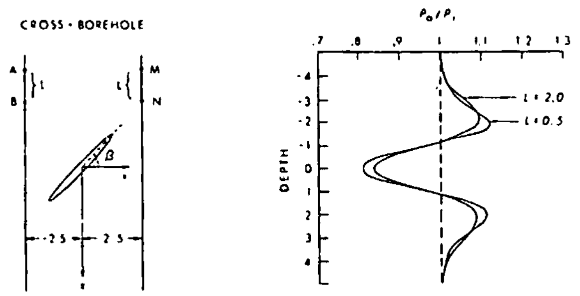
borehole is nearer to the fracture zone. Figure 4b shows anomalies for the same situation as Figure 4a except that now the fracture dips toward the first borehole. Figure 4c shows the effect of varying the resistivity contrast between a dipping fracture and the host medium. As expected, the large contrast cases produce the largest anomalies. Figure 4d shows the change in anomaly shape for the dipping fracture when four electrodes are placed downhole instead of two (compare with Fig. 4b, EMX = 2.5). By study of a large suite of such graphs as these, the comparative capabilities of the various possible cross-borehole arrays can be determined.

The volume integral equation approach of Beasley and Ward (1986) incorporates a half-space formulation, i.e. the earth's surface is not neglected. As with the surface integral equation technique of Yang and Ward (1985a,b), the volume integral equation method requires that only inhomogeneities be discretized. Any number of inhomogeneities of differing sizes and physical properties can be accounted for by this algorithm. Inhomogeneities are discretized into rectangular cells whose size may vary in each of



Electrode Configuration	Body Size	Angles	Resistivity Contrast
Fixed Source Moving Electrode SBX = -2.5 SBY = SBZ = 0 EMX = 2.5 EMY = 0	a = 2 b = 2 c = 0.2	α = 0° β = 45° γ = 0°	$\frac{\rho_2}{\rho_1} < 1$ $\frac{\rho_2}{\rho_1} > 1$

FIGURE 4c
Downhole cross-borehole resistivity anomalies for dipping fracture showing the effect of varying resistivity contrast between fracture and host medium.



Electrode Configuration	Body Size	Angles	Resistivity Contrast
Moving Bipole Source Moving Bipole Receiver SAY = SBY = 0 EMY = ENY = 0	a = 2 b = 2 c = 0.2	α = 0° β = 45° γ = 0°	$\frac{\rho_2}{\rho_1} = 0.10$

FIGURE 4d
Downhole cross-borehole resistivity anomalies for dipping fracture showing the effect of dipole length for downhole electrodes.

the three directions. The fact that targets must be comprised of rectangular or cubic cells means that dipping bodies must be simulated by cells arranged in a staircase fashion. Section and plan views of computed apparent resistivities are the end product of this algorithm. The algorithm is flexible in that it permits a buried electrode to be placed either inside (mise-à-la-masse) or outside (near-miss) the body. The dip of the body and the location of the energizing electrode within it were both varied. The maximum depth at which a body could be located and still produce a detectable anomaly on surface surveys was found to be dependent, as expected, upon the position of the buried electrode and upon the contrast in resistivity between the body and the host. It was found that locating the buried electrode just outside the body did not significantly alter the results from those when the electrode is embedded in the inhomogeneity.

Figures 5a, 5b and 5c show representative results from Beasley and Ward (1986). Each figure is a vertical section through the earth with contours of the resistivity anomaly. A borehole can be placed anywhere on this figure and the resistivity curve that would be observed in such a

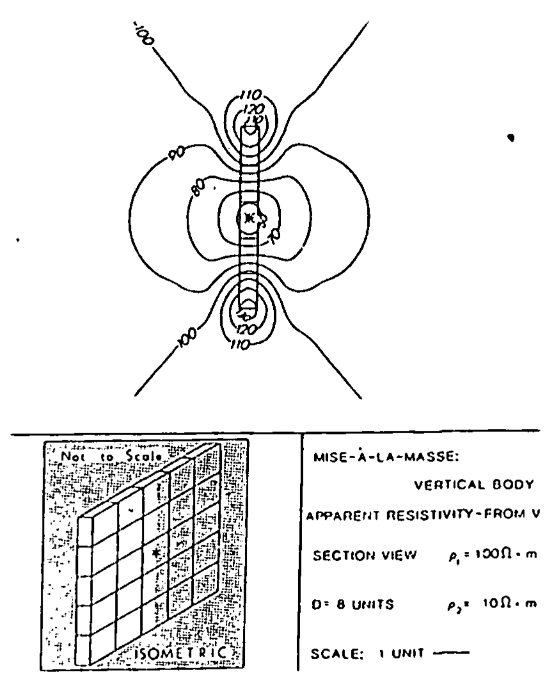


FIGURE 5a
Subsurface resistivity contours for a vertical permeable zone with an imbedded downhole current source.

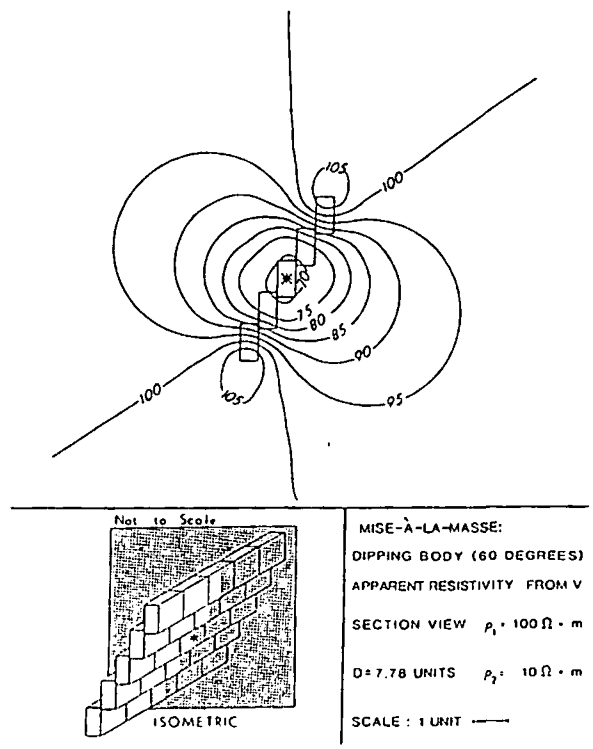


FIGURE 5b
Subsurface resistivity contours for a dipping permeable zone with an imbedded downhole current source.

borehole with a single downhole potential electrode would be given by the intersection of the borehole with the contours. The downhole current electrode source is shown by the star.

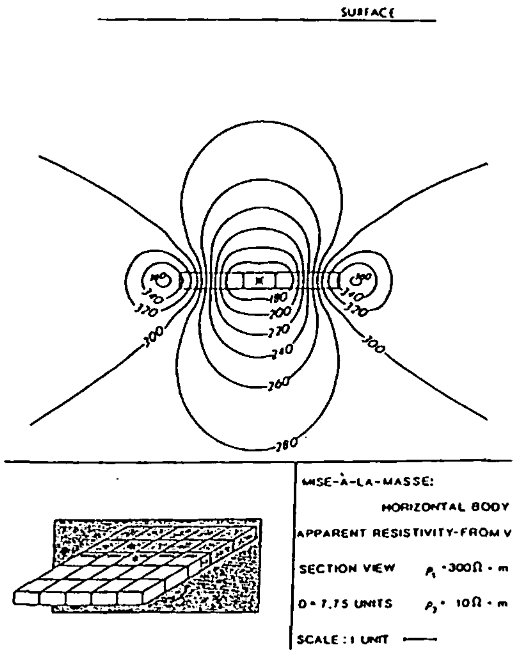


FIGURE 5c
Subsurface resistivity contours for a horizontal permeable zone with an imbedded downhole current source.

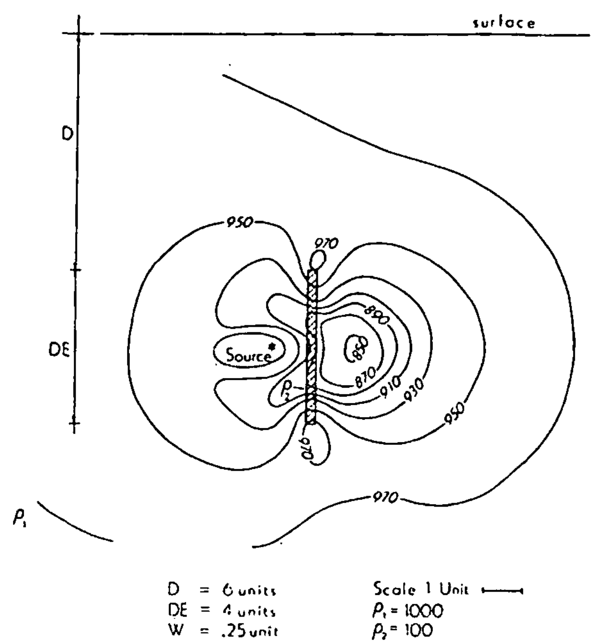


FIGURE 6a
Subsurface resistivity contours for a vertical permeable zone with current source to the side.

Our most versatile algorithm for the borehole resistivity method is the 2-D finite element algorithm used by Zhao et al. (1985). The versatility of this algorithm arises from the fact that the entire subsurface is discretized. Since triangular elements are used for discretization, dipping bodies are readily handled. The algorithm also accomodates a layered-earth host environment. This algorithm was used to evaluate signal-to-noise ratio for various types of noise.

Figures 6a and 6b show typical results from

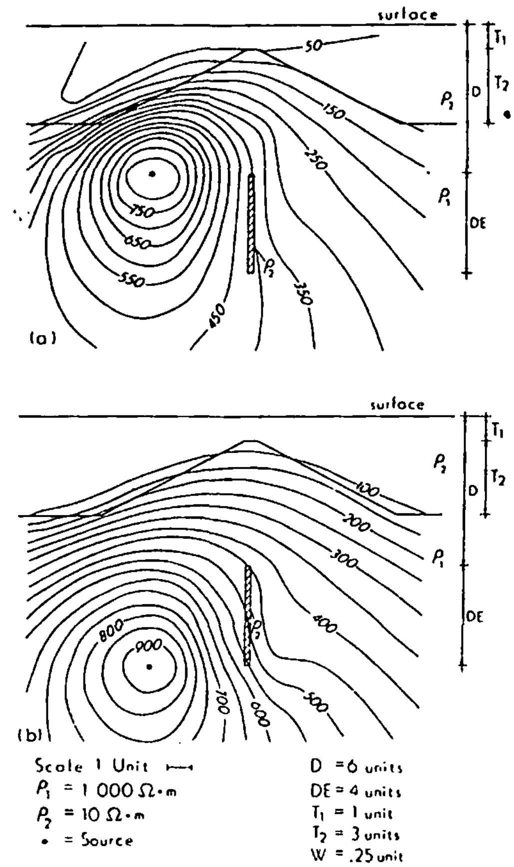


FIGURE 6b
Subsurface resistivity contours for a vertical permeable zone beneath geologic structure with varying positions of the downhole current electrode.

Zhao et al. (1985). Figure 6a shows subsurface resistivity contours in section for a vertical fracture with a current source outside the body. This plot is similar to those given by Beasley and Ward (1986) in Figures 5a, 5b and 5c. Figure 6b illustrates how subsurface topography due to geologic structure affects results. Note that the anomaly due to the fracture is obscured to a great extent by the resistivity pattern created by the contact. This is due in part also to the relatively large distance of the fracture from the downhole current source, shown by the star. A current source in a borehole closer to the fracture would cause a much clearer anomaly.

All computations by Yang and Ward (1985a,b) and Zhao et al. (1985) were performed on an HP9826 desk top computer with 1.6 Mbytes of memory. The algorithm used by Zhao et al. (1985) is currently being extended to 3-D. It is probable that the HP9826 will accomodate the 3-D version. If so, these modeling programs could easily be used in the field with no need to return to a large computing facility.

From the above studies we tentatively conclude the following: the cross-borehole method produces larger anomalies than does a single-borehole method; the cross-borehole anomalies using a pole-pole array are smaller than those for a cross-borehole dipole-dipole array; the cross-borehole mise-à-la-masse method produces larger anomalies than for the other cross-borehole methods; and, the anomalies due to a thin sheet were generally much smaller than those for a sphere, as is to be expected.

Using a 3-D integral equation algorithm developed by San Filipo and Hohmann (1985), West and Ward (1985) performed a model study to evaluate the time-domain electromagnetic (TDEM) response of a horizontal conductive body (fracture zone) imbedded in a half-space. Simplifying assumptions in the algorithm allow modeling only of bodies with two vertical symmetry planes with sources directly above or below. The source transmitter is a large square loop located on the surface of the earth. Receivers are located in boreholes at various locations in the vicinity of the body. Responses are computed at 60 time steps at intervals of 0.4 ms for a total data window of 24 ms. EM field decay curves and plots of decay versus depth are obtained for all three components of the primary, secondary, and total responses. The results are expressed in terms of percent difference plots, and are still under study at this time.

Surface-to-borehole EM in which a large transmitter is coaxial with the well and a downhole detector is run in the well may provide useful information on the location of conductive fractures intersecting the wellbore. Whether this technique will work in cased wells and whether a "crack" anomaly can be distinguished from a stratigraphic conductor are topics under study.

The above discussion outlines our research to date. Other current research involves a model study using the VLF (very low-frequency) method as well as developing a borehole inversion scheme using the finite-element technique. Inversion of the 3D integral equation is also being investigated. An inversion scheme which can incorporate multi-array data is an ultimate goal. Interpretation of complex borehole field data from geothermal sites may then become a reality.

DISCUSSION

The problem of selecting an appropriate borehole electrical system is quite complex. Variables include where to place the electrodes, i.e. how many on the surface and how many down each borehole, and whether to use direct-current galvanic resistivity, which each of the above figures illustrate, or some alternating current, electromagnetic scheme. It is clear that the computer based study of these questions is cost effective in helping select and design an optimum field system.

Our current opinion is that the more data one can collect the better one should be able to characterize the subsurface. We have therefore been making a preliminary investigaton of the design of a system for obtaining both borehole-to-borehole and borehole-to-surface data simultaneously. Such a scheme is conceptually illustrated in Figure 7. We believe we are nearing the stage when a field system can be designed with the very real hope of yielding much more subsurface information than can be realized by presently available systems.

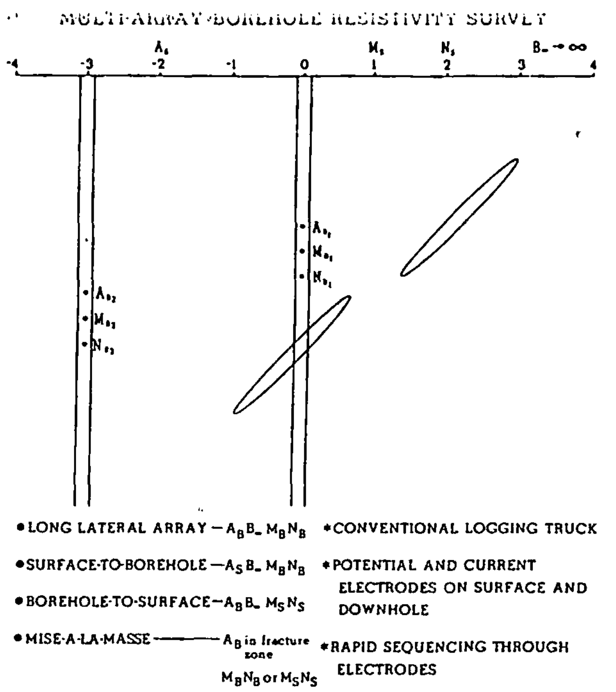


FIGURE 7
Conceptual illustration of a multi-array borehole resistivity system.

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A E I O U: ACCELERATED EXPLORATION for INTEGRATED and OPTIMAL UTILIZATION
A Strategy for Geothermal Resource Development at Department of Defense Installations

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ABSTRACT

A program is currently being conducted by the Division of Earth Sciences, UNLV, and the Geothermal Utilization Division, China Lake Naval Weapons Center (and funded jointly by the U. S. Navy and U. S. Department of Energy), at the Marine Corps Air-Ground Combat Center, Twentynine Palms, California, as an example of accelerated development of geothermal resources at Department of Defense installations. The focus of this program is to assess the potential for development of low-temperature geothermal resources for space heating applications. Decisions are based on data derived from geologic and geophysical surveys, temperature gradient holes, environmental issues, and engineering and economic studies.

There are several important differences between this and previous studies. The most important is that the geothermal reservoir data are known and not assumed. In addition, selected base heat loads are considered as separate items and specific environmental issues are identified in areas of greatest anticipated activity. Recommendations are also made for reservoir confirmation, retrofitting existing structures, and co-locating new structures within the areal extent of the geothermal resource.

INTRODUCTION

Since 1978, members of the Division of Earth Sciences (DES) at the University of Nevada, Las Vegas, have performed assessment studies of geothermal resources throughout the west. The approach has been based on an efficient application of capital resources to obtain a multi-perspective data base that integrates natural resource characteristics, economic qualifications, technical possibilities, and environmental liabilities.

This paper describes an integrated program for development of low- to moderate-temperature geothermal resources that was jointly conceived by the U.S. Navy, the U.S. Department of Energy, and the University of Nevada, Las Vegas. The program consists of the four major aspects of development: 1) exploration; 2) resource definition; 3) environmental considerations; and 4) optimal resource utilization (fig. 1).

Although military installations throughout the United States require vast amounts of energy, traditional marketing strategies of alternate energy resources are not entirely applicable to most mission-oriented bases. There are two important reasons for this. First, the primary function of the active military base is to supply or support the National Defense. Any extracurricular activity that either damages, compromises, or adversely affects this fundamental mission is wholly unacceptable. Second, most energy conservation plans rely largely on economics; market competition and variations in energy prices and availability are the fundamental requirements in the business world. The military first recognizes the mission, then the cost. It is mainly for these two reasons that, although many bases are capable of geothermal resource development, no active base in the United States uses a geothermal resource to offset as much as a single BTU. Indeed, considerable effort has been directed toward geothermal resource development on active bases for more than a decade. Although some resources have been found, there has been no subsequent development. It is clear that if geothermal resources are to become an important factor of the military's energy formula, future effort must focus on active and accelerated integration of exploration, development, and utilization.

Working closely with both military and civilian personnel is mandatory and may result in identification of potential applications not normally considered for commercial development. Providing key military personnel with timely, detailed plans for activities in both restricted and non-restricted areas is a requirement for a safe, efficient assessment program. This requires that the first three elements outlined above be tailored to the base characteristics.

Experience has shown that identification of geothermal resources at military installations does not necessarily result in resource utilization. A moderate temperature (95°C) geothermal resource was identified at the Hawthorne Army Ammunition Plant in western Nevada in 1981 (Trexler and others, 1981). Development of that resource has been slow due to the uncoordinated activity by civilian contractors and government agencies who are not familiar with optimal development of geothermal resources.

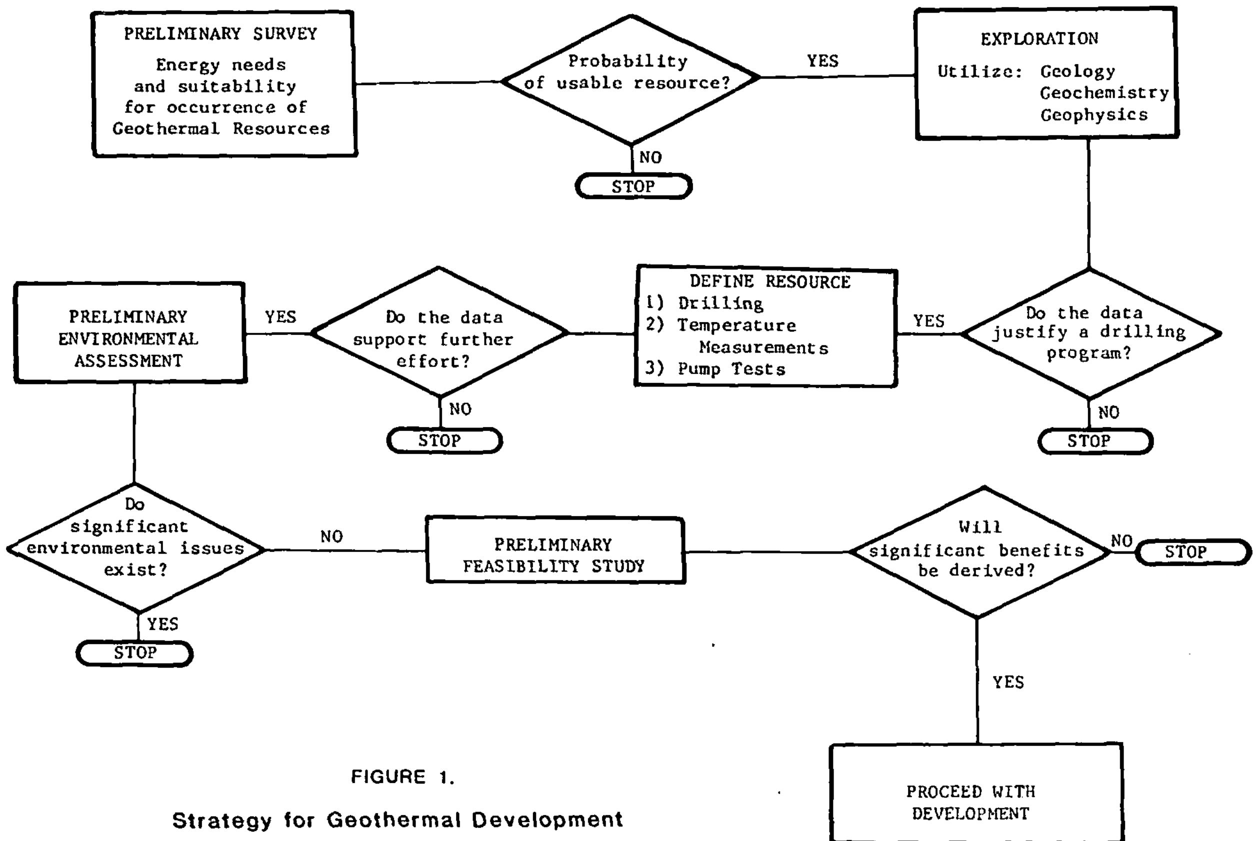

FIGURE 1.

Strategy for Geothermal Development at Department of Defense Installations

The integrated program approach that is currently underway at the Marine Corps Air-Ground Combat Center (MCAGCC), Twentynine Palms, California, demonstrates the benefits of resource development by the directed efforts of agencies familiar with military operations and geothermal energy.

PRELIMINARY SURVEY

Interest in development of geothermal resources beneath the MCAGCC was stimulated by the report of a well, 122 m deep, with a water temperature of 73°C, located 3.6 km southeast of the Center's boundary. on or off the Base?

Warm ground water in the Twentynine Palms area has been known for at least 30 years. Wells drilled for domestic water north of the city of Twentynine Palms have reported temperatures of 40-73°C. Higgins (1980) reported 3 wells ranging in temperature from 48°C to 63°C. The approximate boundary of the geothermal area in the vicinity of Twentynine Palms was described in Leivas and others (1981) as extending approximately 15 km in an east-west direction and 6 km north-south.

The Center encompasses approximately 2,600 square kilometers of the southern Mojave Desert. The administrative and housing area is located 8 km north of the city of Twentynine Palms, California (fig. 2). Large buildings such as offices, barracks and classrooms are heated by a central boiler plant employing a low pressure steam and distribution system. Individual and multiple family housing employ individual gas-fired forced air heating systems.

An expeditionary air field (EAF) is located at Camp Wilson, approximately 10 km northwest of the Center's administrative area. The only permanent structures at Camp Wilson are 14 shower and lavatory buildings. Water is heated by fuel oil.

The annual expenditures for heating oil and natural gas for the entire Center were $2,050,000 for fiscal 1983 (Facilities Engineer personnel per. comm., 1983).

EXPLORATION

Geophysical exploration was performed by the Geothermal Utilization Division, Naval Weapons Center, China Lake. Gravity and magnetic surveys indicated a geologic structure, the Bullion Mountain fault, trending northwest-southeast beneath the MCAGCC administrative area and trending southeasterly towards the geothermal well mentioned above. Other geophysical anomalies tended to confirm the

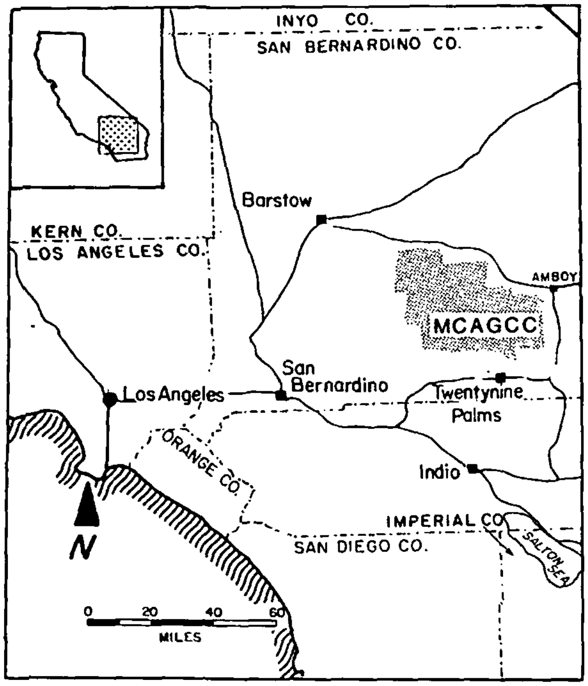

Figure 2. Index map of MCAGCC, Twentynine Palms, California

‹istence of northwest trending structures subpar- llel to the Bullion Mountain fault. These are rom east to west: 1) Mesquite Lake fault; 2) Sur- rise Spring fault; and 3) Emerson Copper Mountain ault system (fig. 3).

:SOURCE DEFINITION

Seven sites for temperature gradient drilling :re selected based on geophysical surveys and ·oximity to the Center's administrative area. The ·illing program was supported by the Navy and the S. Department of Energy, as a cooperative program reement, and supervised by the Division of Earth iences, University of Nevada, Las Vegas (Trexler d others, 1984). A pre-drilling conference was ld at MCAGCC to appraise base personnel of our tent to proceed with temperature gradient drill- g and to ascertain what restrictions would be aced on drilling operations.

The drilling plan specified drilling to a pth of 304 m or bedrock, which ever came first. nce no wells had been drilled at the Center to a pth of 304 m or greater, blowout prevention uipment was required on the first hole. Hole No. (fig. 3) was located adjacent to a housing area d as near to the suspected trace of the Bullion untain fault as possible. Quartz monzonite bed- ck was encountered at 201 m; drilling continued 268 m. Maximum mud return temperature recorded ring drilling was 27°C.

Hole No. 2 was located 1.37 km southwest of le No. 1, perpendicular to the strike of the llion Mountain fault. This location is near the Mesquite Lake fault which was considered to be a favorable controlling structure for geothermal fluid migration.

Hole No. 2 was completed to a depth of 304 m without encountering bedrock. Maximum mud return temperature of only 27°C suggested that if a geo- thermal resource was present it was very deep.

The third drill site was located approximately half-way between temperature gradient hole No. 1 and temperature gradient hole No. 2 (fig. 3) along the trend of the gravity anomaly and 2.1 km to the north of temperature gradient hole No. 2. This location would confirm if the Bullion Mountain fault (gravity anomaly) was the controlling struc- ture for the geothermal fluids.

This hole was completed to 335 m and maximum mud return temperatures were 30°C. These data con- firmed that the Bullion Mountain fault, in the vic- inity of the Center's administrative area, was not the controlling structure for the migration of geo- thermal fluids.

After analyzing the results of drilling, it was decided by DES and Navy personnel to drill dif- ferent structural blocks on the Center to determine which faults controlled the migration of geothermal fluids.

Temperature gradient hole No. 4 was located immediately east of the Bullion Mountains (east of the Bullion Mountain fault, fig. 3), to ascertain if the geothermal fluids reported south of the Cen- ter were controlled by faults on the east side of the Bullion Mountains. Bedrock was encountered at 271 m and drilling was terminated at 280 m. Maxi- mum mud return temperature was 29°C at 280 m which indicated that the geothermal fluids are not in this structural block.

At this point, DES and Navy personnel agreed to drop two remaining primary sites near the admin- istrative area and focus on other secondary sites west of the Bullion Mountain fault. This was done in an effort to locate the controlling structures for the geothermal fluids. These two additional sites were chosen on opposite sides of the Surprise Spring fault. A major logistical problem surfaced because these sites are located on training ranges with restricted access. Temperature gradient hole No. 5 was drilled while permission to enter the training area was obtained.

Site 5, is located 3 miles west-northwest of the Center's administrative area. It is situated between the Mesquite Lake fault on the east an Sur- prise Spring fault on the west. Maximum mud return temperatures were 34°C, indicating the presence of geothermal fluids. The hole was to be drilled to 335 m, however, a bit change was required at 287 m and, upon tripping back into the hole, circulation could not be recovered. Subsequent attempts to recover circulation failed and temperature gradient hole No. 5 was completed to 287 m.

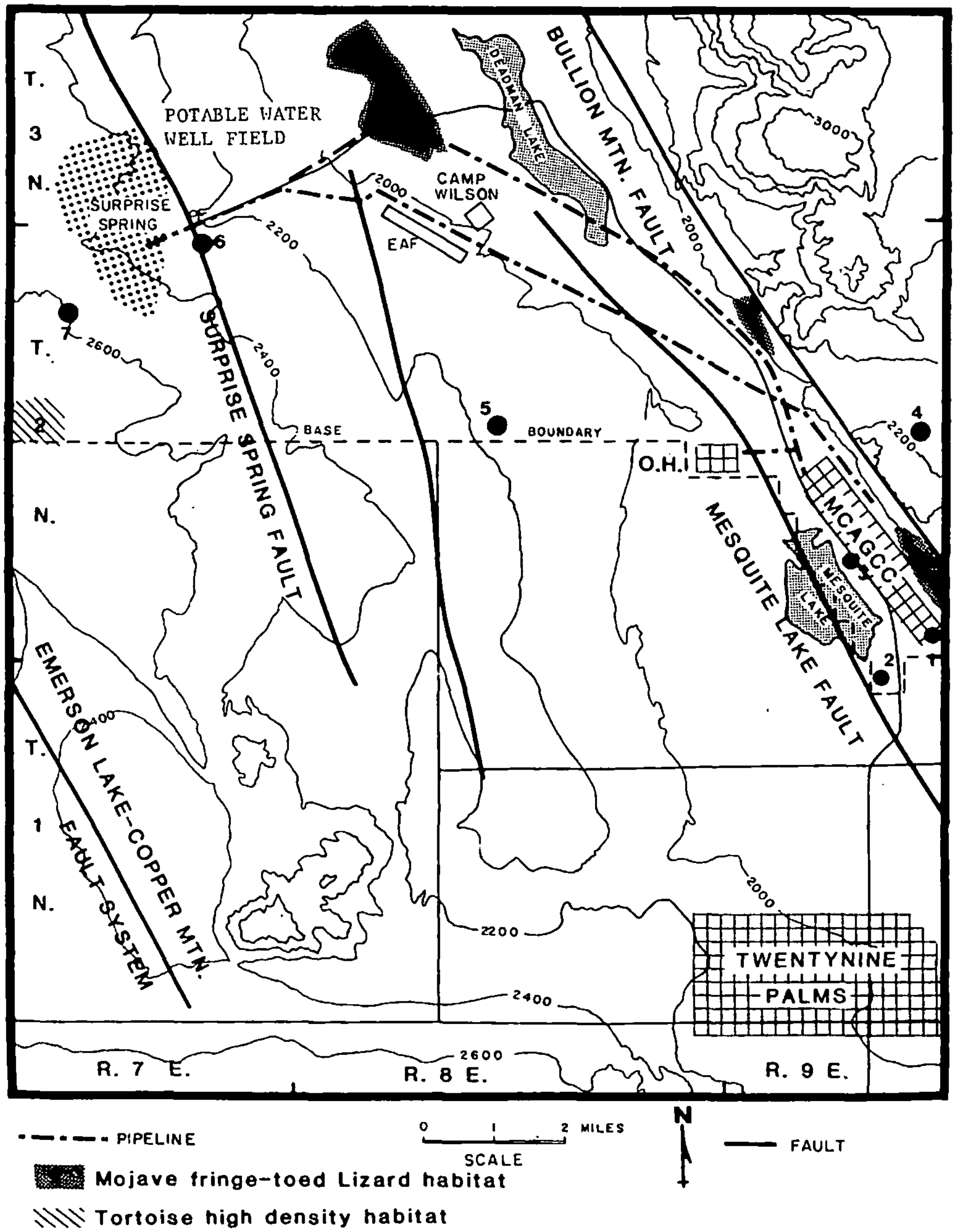

Figure 3. Composite map showing important geologic, geothermal, and environmental features at MCAGCC

Once permission to enter the training area was received, hole No. 6 was drilled to a depth of 335 m. Maximum mud return temperatures were 39.4°C, however, after termination of drilling and prior to trip-out, the mud return temperature increased 1.4°C in 20 minutes during circulation.

Hole No. 7 is located west of the Surprise Spring fault (fig. 3). The hole was completed to 323 m and the maximum mud return temperature was only 23°C. The low mud return temperatures tentatively indicated that geothermal fluids were migrating up the Surprise Spring fault and flowing east.

All drill holes were cased with 6.35 cm T. & C. iron pipe capped on the bottom and filled with water. The holes were back-filled with cuttings and a cement seal was placed from ground surface to 3.3 m.

:MPERATURE GRADIENTS

Temperature gradient measurements were made on ebruary 13th and 14th, and February 27th and 28th, 984, two and four weeks after the termination of ne drilling program. Temperature measurements ere made at 6 m intervals.

A maximum temperature of 32.6°C was measured : 268 m in hole No. 1. The temperature gradient alculated over the interval from 61 m to 244 m as 1.3°C/100 m. Hole No. 2 had a BHT of 29.7°C nd a gradient of 2.7°C/100 m. A similar tempera- ure gradient of 2.7°C/100 m was measured in hole o. 3. The temperature gradient in hole No. 4 was .6°C/100 m which is quite similar to holes 2 and .

The temperature gradients in holes 1 through 4 robably reflect the regional background tempera- ure gradient for this portion of the Mojave block, hich is 2.5 to 3.0°C/100 m.

A maximum temperature of 51.6°C was measured : 287 m in hole No. 5 (fig. 4). The temperature radient calculated in the interval between 110 m nd 287 m was 8°C/100 m. As shown in figure 4, the radient remains positive at the bottom of the ole. Hole No. 6 had the highest measured tempera- ure of all holes drilled during this phase of geo- hermal development at MCAGCC. A maximum tempera- ure of 67.1°C was measured at 335 m. The tempera- ure gradient below 275 m (fig. 4) is 3.3°C/100 m nd probably reflects the convective gradient in ne geothermal reservoir.

Hole No. 7, located west of the Surprise pring fault, has a maximum temperature of 33.9°C : 323 m and a gradient of 3.8°C/100 m.

IVIRONMENTAL FACTORS

The ultimate development of geothermal re- ources at the MCAGCC will require an acceptable ethod of fluid disposal and will have an impact pon the desert ecosystem. Although the absolute agnitude of the environmental impact is not pre- ently known, selected fluid disposal options can : discussed in terms of the impact they will have 1 the major environmental issues on the base. A echnical report completed in April, 1984, des- ribed the fluid disposal options available at the enter (Flynn and others, 1984).

Four fluid disposal options, identified as echnically feasible at MCAGCC, included surface sposal on existing playas, fluid injection, irri- ation, and sewage disposal. Figure 5 shows a sug- ested utilization rationale that includes all four nd that may be easily accommodated by the existing ase structure.

Nine major environmental issues were also lentified and the ramifications of each, with espect to geothermal fluid utilization, were dis- ussed. The nine issues and pertinent comments are resented in Table 1.

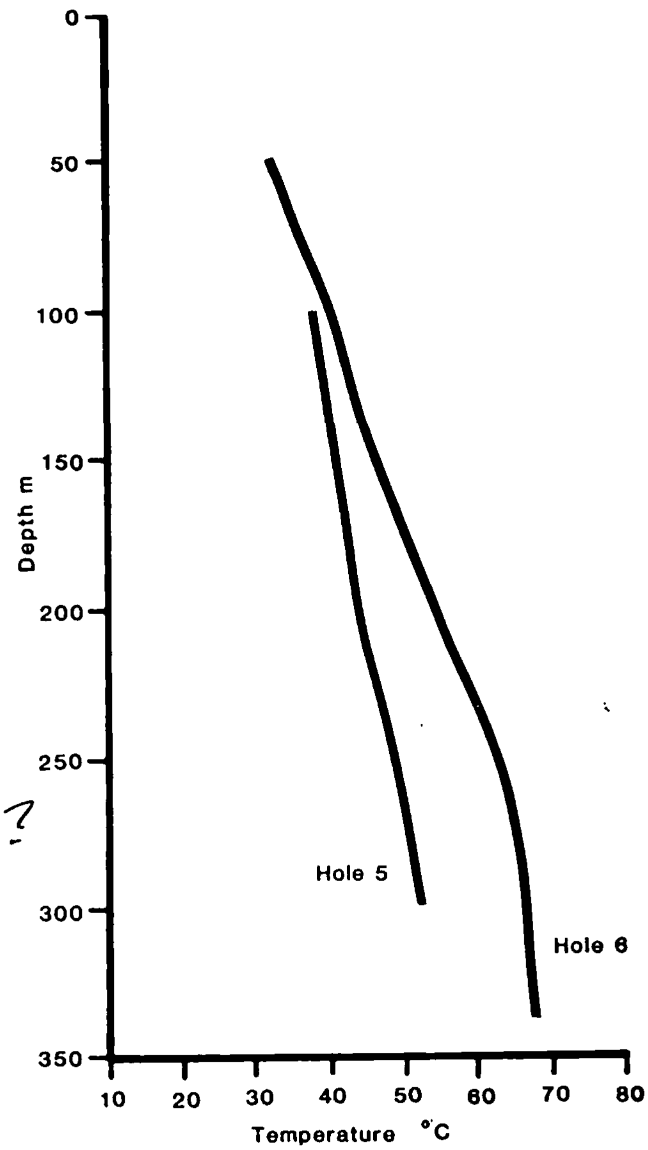

Figure 4. Temperature gradient profiles of holes 5 and 6

No environmental issues were identified that would preclude development of geothermal resources at MCAGCC. The total impact is estimated to be equivalent to the impact of the existing potable water well field and associated pipelines.

In addition to the obvious fuel savings, several ancillary benefits will accrue from the development:

1) reduce stress on potable water aquifer
2) enhance vegetation and tree growth with irrigation
3) increase bacterial digestion efficiency (sewage)
4) mitigation of dust from Deadman Lake Playa

Table 1.

Environmental Issue	Site Characteristics
1. Land use	Well field and fluid distribution system will be in training area - present potable water distribution system is located along roads in training area. Proposed surface disposal on playa (Deadman Lake) represents area of minor concern.
2. Fish, wildlife, vegetation, endangered species of plants and animals	There are no species of fish within the study area. Although some sensitive species have been identified surrounding the base, the prospects of geothermal utilization and surface disposal (on playas) represents no more hazard than present activities associated with training. The habitats of two sensitive species, indigenous to the area, have been identified and will not be seriously affected by proposed development.
3. Water quality	There are no permanent surface waters within the study area that can be used as a source of potable water. Geothermal waters are likely to contain slightly high concentrations of fluoride and boron.
. Air quality	Although geothermal fluids for direct-use rarely contain appreciable amounts of non-condensable gases, a chemical analysis is warranted.
5. Hot springs	There are no thermal springs presently flowing within the study area.
6. Physical geology a) subsidence	The geothermal reservoir rock at MCAGCC, Twentynine Palms, is nearly identical to the unconsolidated formations that produce non-thermal drinking waters. Although 35 feet of drawdown has occurred, there have been no reports of subsidence within the well field.
b) induced seismicity	This is generally associated with deep, high-pressure injection and is not likely to be a problem.
7. Noise	The area is presently used as an air-ground combat training center. Also, no residential, recreational or breeding areas are adjacent to proposed production area.
8. Socioeconomics	Will likely reduce the cost of heating at Mainside. Secondary application may also reduce amount of fluids pumped from non-thermal aquifers. An economic feasibility study is presently being conducted.
9. Archaeological/ cultural resources	Archaeological surveys have been successfully used to locate and isolate sensitive cultural areas (i.e., Surprise Spring) within the study areas. Proposed development will not affect sites.

GINEERING FEASIBILITY

Estimated temperatures of the geothermal uids at a depth of 610 m near hole No. 5 range om 80°-85°C based on the observed temperature adient. The primary uses for fluids at these mperatures are space heating and domestic hot ter. These uses employ existing technology and mmercially available equipment.

Cost effectiveness is a primary concern at the nter. The costs for a new geothermal heating stem include a production well, piping system, sposal system, and end-user heating retrofits. :h of these costs increase as the service area expands. Critical to the geothermal system is the location of the production well in close proximity to the heat load.

Relative to the known geothermal reservoir, Ocotillo Heights, which is composed of 250 family housing units, is the closest existing large heat load. A preliminary cost estimate for converting Ocotillo Heights (O.H. on fig. 3) to geothermal heating from a source at hole No. 5 is presented in Table 2.

The estimated offset natural gas consumption is 150,000 therms per year or $90,000/year in natural gas costs. This gives a simple payback period

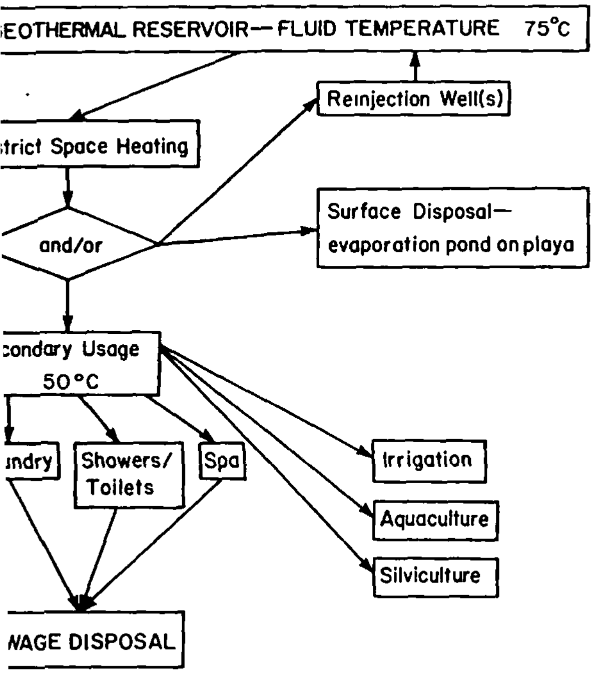

Figure 5.

Direct Utilization of Geothermal Energy
MCAGCC Twentynine Palms, CA.

Table 2.

,000 ft. production well & pump	$ 150,000
0,000 ft. 8" insulated pipe $25/ft.*	500,000
,000 ft. 6" insulated pipe $30/ft.	60,000
50 retrofits @ $1,200/unit	300,000
0,000 ft. disposal line @ $4/ft.**	40,000
Estimated Total	$1,050,000

*Installed on surface
**Buried

2 years. If the geothermal well can be located cent to Ocotillo Heights, the conversion cost 500,000 less and the corresponding simple pay- time is 6 years.

If the new construction is located in the nity of hole No. 5, then these new buildings d be ideal candidates for geothermal space ing. Supply line costs will be minimized and rofit" costs would be limited to the cost dif- ntial between heat exchangers and conventional ices.

CONCLUSIONS AND RECOMMENDATIONS

This report demonstrates the utility of integrating data from those well defined parameters that most influence the success of geothermal resource utilization. The temperature, depth and approximate areal extent of a low-temperature geothermal resource (70°C) was determined on the basis of data derived from geological, geophysical, and temperature gradient hole drilling surveys carried out by the Geothermal Division at China Lake Weapons Center and the Division of Earth Sciences, UNLV. Data from those studies were used to develop use-scenarios that included heat and water utilization in a framework that was consistent with existing military operations and environmentally beneficial.

Data are presently being collected that will help determine the engineering and economic feasibility of offsetting all or part of the Center's energy demand with geothermal heat. A report by Bakewell and Renner (1982) included an economic analysis of using geothermal fluids for MCAGCC which was based on assumptions which have been found to be totally misleading. The important data are listed in Table 3:

new data made the difference

Table 3.

Resource Character	Bakewell & Renner 1982	Trexler and Others 1984
Location	unknown	between #5/6 on map
Temperature	63°C	70°C - 85°C
Depth	90 m	350-600 m

The conclusion that the attractiveness of geothermal utilization is sensitive to co-locating the resource and end use is correct. The report differs, however, in assuming the location of the resource, in ignoring optional uses for the fluids, and for not considering separating isolated heat loads from the entire base heat load.

The principal recommendation of this report is to define the eastern-most limit of accessible and usable geothermal fluids by drilling temperature gradient holes. A series of 2-3, 600 m holes in the area of Ocotillo Heights and west will provide the required data. Following this, a pump test on a properly sited well will complete the resource definition phase of the program.

better than gp to define area

Detailed engineering and economic feasibility studies using the most accurate resource data would then be warranted. Preliminary estimates show that economic benefits may be realized within 6 years if the Ocotillo Heights residential area is retrofitted for space and water heating. More significantly, new construction located at the site of the geothermal reservoir would achieve a payback in a shorter time period if geothermal heating systems were included during construction.

ACKNOWLEDGEMENTS

Many people participated in the successful completion of this project and the authors would like to acknowledge their contributions. John Crawford and Bill Holman, of the San Francisco Operations Office of the U.S. Department of Energy, provided pertinent suggestions during site selection review and the drilling phase of the program. Geothermal Utilization Division, Naval Weapons Center, China Lake, personnel were instrumental in conceiving the project, providing geophysical data used in drill site selection, and assisting in temperature gradient measurements and site restoration. Individuals of the Geothermal Utilization Division who provided this support are Al Katzenstein, Jack Neffew and Ted Mort. Carl Halsey's critical review of the drilling report showed that the characterization of specific mission contraints must utilize and be functionally interwoven with the preliminary qualification limit.

Logistical support during activities at the Marine Corps Center was provided by Lt. Col. C.E. Schaffer and the staff of the Installations Division, specifically Staff Sgt. William Flummerfelt, who provided assistance in acquiring necessary permits for drilling activities at the MCAGCC, and baseline data for the fluid disposal study. Stu Hammonds helped secure data for the engineering feasibility study, and Charles Miles and David Holmes of the Naval Civil Engineering Laboratory were instrumental in funding this program.

We would also like to acknowledge the drilling contractor, Fred Anderson and Son Exploration Drilling, Inc., for a job well done and for their patience while drill site locations and depths were changed because of data acquired during drilling.

Without the assistance and support provided by these individuals and organizations, this project would not have been completed.

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Finally, many thanks to U.R. Bearmatt for keeping the lines of communications at MCAGCC open. 10-4.

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GEOCHEMICAL EXPLORATION OF THE CALISTOGA GEOTHERMAL RESOURCE AREA, NAPA VALLEY CALIFORNIA

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ABSTRACT

Chemical analysis of well waters in the upper Napa Valley, near the city of Calistoga, California suggest that the upwelling and localization of fluids with temperatures up to 135°C, may be related to faulting or fracturing along the geographic axis of the Napa Valley. Calculated temperatures from chemical geothermometry are always higher than measured temperatures, with maximum values often 50°C higher than sampling temperatures at well point. Two shallow, subsurface systems of moderately-high temperatures were detected using waters with locally high chloride values (Cl 180ppm). Maximum reservoir temperatures may exceed 150°C. Mixing of the geothermal water with shallow, cool groundwaters is indicated by intermediate concentrations of Cl, F, and B, by ternary molality plots of Cl, B and HCO_3, and by a trilinear diagram of major cations and anions.

INTRODUCTION

The Calistoga geothermal area, located near the head of the Napa Valley, in northern California is a shallow hydrothermal convection system of low-to-moderate temperature (Fig. 1). The hottest wells—The Geysers at 135°C, and wells near Pacheteau's spa at 120°C—are coincident with the geographic axis of the valley (Fig. 2).

Tertiary volcanic rocks, resting unconformably on rocks of the Jura-Cretaceous Franciscan assemblage form the most prominent surficial exposures in the highlands around Calistoga and along the margins of the valley. The volcanic rocks which consist predominantly of ash flows, welded and partially welded tuffs, and tuff breccia, agglomerate and rhyolite are considered to be part of the Sonoma Volcanic Field of Upper Plicene age (Sarna-Wojcicki, 1976; Fox, 1983). The youngest published date on this volcanic sequence, obtained from the summit of Mt. St. Helena seven miles north of Calistoga, is 2.9±.2my (Mankinen, 1972, p. 2065). Further to the northeast, lies the Clear Lake Volcanic field which has been active into the Holocene (Fox, 1983). Two large negative gravity anomalies centered both north and south of Calistoga maybe related to the geologically young volcanism. Although the cause of the anomalies is uncertain, Youngs and others (1980) suggested that underlying the southern anomaly could possibly be an elongate intrusive mass. Although little additional information is available concerning the northern gravity anomaly, the area is immediately southwest of the Clear Lake volcanic field. Thus, both the northern and southern anomalies may well represent shallow magma chambers that were the source of the late Pliocene-Pleistocene volcanic sequences in the Calistoga-Clear Lake region. It is thus possible that residual heat from either of these chambers is the driving mechanism responsible for the geothermal activity at Calistoga.

Rocks underlying the Napa Valley have undergone gentle folding and faulting. Major pre-Pliocene northwest trending fault zones in Franciscan rocks have been mapped to the northwest of the Napa Valley by Fox and others (1973), but the Sonoma Volcanic sequence and Quaternary alluvium mask any pre-Sonoma faulting that may be present within the upper Napa Valley. Minor faulting of the Sonoma Volcanic sequence however, was noted just to the north of Calistoga, as well as some relatively large-scale faults of probable normal displacement approximately five to six kilometers southwest of Calistoga (Fox and others, 1973; Fox, 1983).

Fox and others (1973) have also indicated the presence of a major northwest-trending thrust fault along which Franciscan rocks have been thrust over rocks of the Sonoma Volcanic field at an angle of 20° to 30°. This fault is a major feature of the western limb of the Napa Valley syncline. Its eastern terminus has been interpreted as possibly being coincident with the current axial plane of the Napa Valley (Youngs and others, 1980). The eastern terminus of this thrust may represent the major structural discontinuity underlying the upper Napa Valley as suggested by Waring (1915), Faye (1975), and Taylor (1981), enabling geothermal fluids to rise into the alluvial and fractured volcanic aquifers beneath Calistoga.

Geophysical studies (Youngs and others, 1980) undertaken by the California Division of Mines and

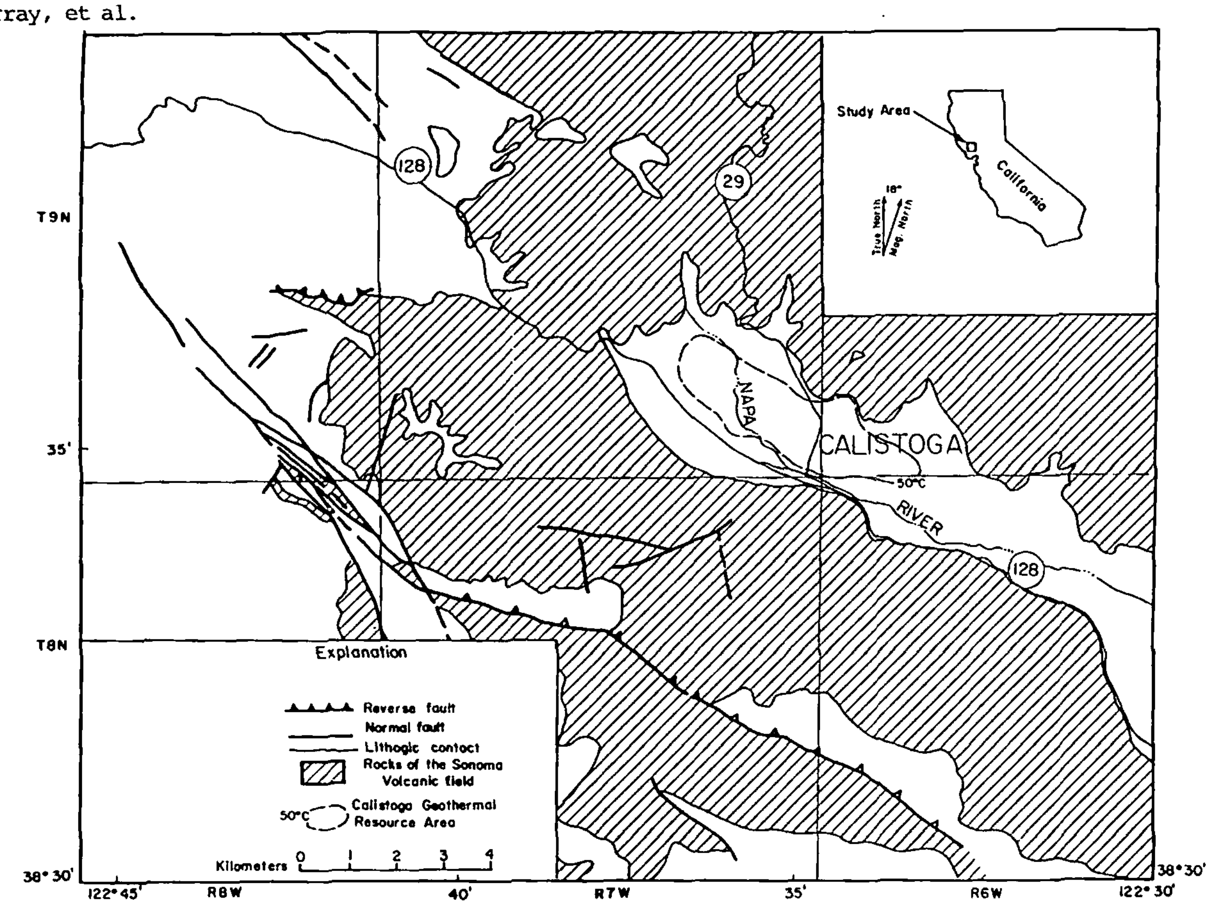

Figure 1. Map showing the general location of the Calistoga Geothermal Resource area (Modified from Fox and others, 1973).

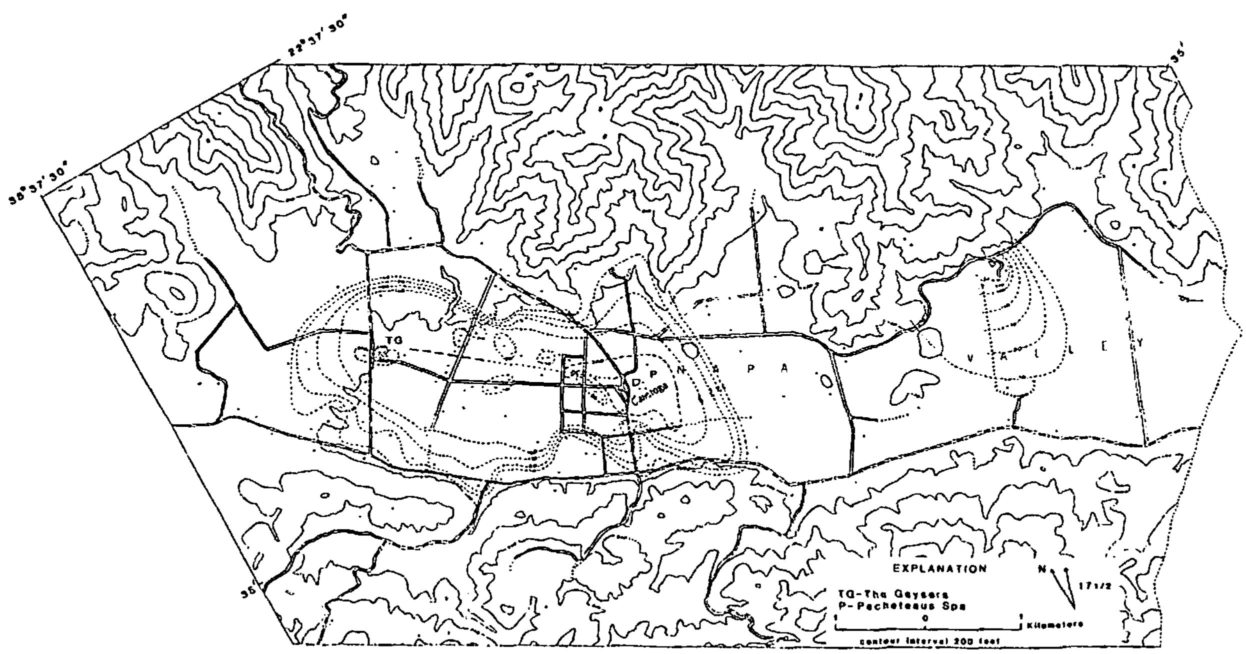

Figure 2. Isochloride lines outlining the Calistoga geothermal anomaly. The hottest wells in the area--135°C and 120°C--occur at The Geysers (TG) and Pacheteous Spa (P). The zones appear to line along an inferred fracture zone which coincides with the geographic axis of the valley.

Geology (CDMG) within the Napa Valley-Calistoga area indicated several areas of possible, but inconclusive evidence of faulting associated with areas of known geothermal waters (Taylor, 1981).

In particular, seismic refraction and electrical resistivity surveys (Youngs and others, 1980) have detected possible zones of hot water that maybe associated with this faulting near The Geysers. The resistivity sections indicate an irregular but somewhat elongate distribution of the conductive zones. Some of the lines, running perpendicular to the axis of the valley, do not show any lower depth limits for the resource near the center of the valley, perhaps suggesting the upwelling of fluids. In this model, the hottest water wells would be expected along the fracture zone, or geographic axis of the valley, while progressively cooler water, caused by mixing with surface water, would be encountered in wells located closer to the margins of the valley, away from the fracture zone. Our field and geochemical work has shown that geothermal waters do indeed rise near the geographic axis of the valley, and further, that these waters have a distinctively different chemistry than water from the margins of the valley. As a result, water chemistry can be used to isolate and trace the geothermal system throughout the Napa Valley - Calistoga area.

GEOCHEMISTRY

We have analysed a variety of geochemical parameters to form a preliminary assessment of the low-to-moderate temperature resource at Calistoga. Water chemistry for selected water wells within the Calistoga area shown in Table 1. Chemical geothermometry results for several of these wells are presented in Table 2. The equations used to calculate the geothermometers are shown in Appendix A. Two shallow, localized, subsurface systems of elevated temperatures are indicated by geothermometry, using water with high chloride values (Cl 180ppm). A subsurface aquifer of intermediate depth generally yields maximum values approximately 50°C higher than the sampling temperatures registered at shallow depths. A deeper aquifer is indicated by Na-K-Ca geothermometry to have a temperature range of 100° to 135°C. Several of the calculated temperatures however, are much higher, perhaps suggesting the potential of still greater temperatures at depth. Recent deep drilling in the greater Calistoga area by the CDMG and the California Energy Commission however, indicate only moderate increases in temperature with depth, are expected unless the wells are located directly along the fracture zone shown in Figure 2. For example, in two test wells drilled by the CDMG to depths of 787 and 836 feet,

Sample Number	Depth meters	Depth feet	Sampling temp. °C	pH	Na	K	Ca	Mg	ppm SiO_2	Li	Cl	B	F	HCO_3	SO_4	Fe
M-002-80	54.9	180	26	7.05	176	16	13	6	36	1.12	193	10.5	8.9	230		
1-M-004-80	61.0	200	33	7.10	174	8	16	6	15	0.50	183	10.2	6.5	200	-	2.70
2-M-015-81	97.6	320	38	7.16	183	5	7	2	54	1.39	188	11.5	9.0	189	-	5.01
3-M-010-81	189.0	620	51	7.90	204	6	8	1	36	1.18	189	12.5	7.3	217	-	8.78
4-M-018-81	64.0	210	36	8.36	170	4	14	4	30	0.60	188	11.2	4.3	207	-	0.32
5-G-016-80	3.7	12	12	7.75	13	L	22	4	36	L	7	0.2	L	75	11	0.11
6-G-001-80	91	300	91	7.00	198	8	5	1	55	1.58	202	9.7	10.5	147.9	-	0.07
7-G-009-81	58	190	135	8.50	206	9	2	0.5	56	1.95	201	9.8	11.5	-	-	-
8-G-012-80	60	198	81	7.20	202	7	2	0.5	61	1.81	190	9.4	11.0	151	-	0.08
9-G-016-80	123	400	37	6.50	184	4	28	2	20	1.12	190	9.8	8.5	193.7	-	0.56
10-G-020-80	59	193	116	7.75	190	7	30	1	60	2.05	201	9.9	11.0	154.8	-	0.08
11-G-025-80	61	200	61	6.97	179	10	22	4	39	1.53	188	9.9	9.1	166.2	13	0.12
12-G-037-80	125	410	52	6.70	188	7	8	1	31	1.75	195	10.1	10.0	172.3	-	0.21
13-G-044-80	61	200	40	6.65	180	12	11	4	35	1.14	194	10.2	7.0	203	-	-
14-G-058-80	61	200	74	6.65	191	4	10	1	42	1.40	191	9.4	8.5	165.5	-	0.14
15-G-096-80	72	235	85	8.05	192	6	10	0.5	41	2.02	187	8.8	10.7	203.6	19	-
16-G-097-80	46	152	95	8.40	222	10	4	0.5	56	2.26	219	11.1	12.3	121.2	12	0.05
17-G-105-80	76	250	44	7.40	200	7	6	0.5	59	1.59	191	9.7	9.1	219.6	-	0.12
18-G-111-80	63	207	104	7.85	205	9	6	0.5	54	2.09	212	10.3	11.5	174	24	0.01
G-112-80	-	-	30	6.40	239	7	10	2	38	2.28	198	8.9	8.8	279.8	-	-
19-G-115-80	61	200	35	7.00	143	10	6	3	34	0.93	76	8.1	4.2	316.4	11	0.43
20-G-116-80	57	187	41	6.95	238	7	8	3	37	1.48	191	12.6	3.1	359.9	-	0.59
21-G-122-80	38	125	15	6.98	29	4	17	7	25	0.05	12	0.6	0.8	180.7	-	2.20
22-G-143-80	73	240	27	6.88	138	5	14	8	24	0.05	130	8.0	1.2	275.7	15	1.50
23-G-181-80	46	151	20	5.99	121	4	19	18	16	0.06	107	8.3	1.7	225.2	-	0.07
24-G-201-80	64	210	64	6.33	173	11	13	5	32	1.20	191	10.1	7.5	229.5	-	-
25-G-206-80	55	180	55	6.35	186	7	16	2	32	1.46	229	9.8	10	152.9	-	-
26-G-006-80	91	300	16	6.25	13	2.5	9	3	31	0.05	6	0.125	0.1	61.8	10	0.025
27-G-077-80	73	240	19	6.80	12	2.5	51	21	15	0.05	32	0.125	0.1	193	20	0.12
28-G-083-80	54	210	19	5.98	10	2.5	30	7	10	0.05	5	0.125	0.1	112.1	13	0.68
29-G-089-80	42	139	22	6.18	11	2.5	13	3	12	0.05	5	0.125	0.1	41.9	-	0.15
30-G-121-80	79	260	15	6.90	26	4	15	5	31	0.05	9	0.125	0.5	125	-	0.26
31-G-135-80	134	440	19	6.52	61	9	14	7	29	0.27	9	0.5	0.3	231.8	50	0.19
32-G-150-80	65	212	20	6.42	107	11	15	9	25	0.27	108	5.5	0.3	264.8	-	1.02
G-165-80	49	160	20	6.00	17	7	16	9	29	0.05	7	0.125	0.2	165.8	-	-
33-G-169-80	107	350	40	6.10	54	2.5	20	3	22	0.05	5	0.125	0.3	242.5	31	-
34-G-173-80	146	480	22	6.32	55	2.5	16	20	16	0.28	37	1.8	0.2	253.9	21	0.76
35-G-175-80	57	186	21	6.20	40	12	21	11	34	0.05	7	0.125	0.2	164.7	68	6.72
G-195-80	73	240	22	5.64	12	4	12	5	32	0.05	8	0.125	0.1	106.8	-	-
36-G-196-80	61	200	19	6.18	41	10	26	8	22	0.05	7	0.125	0.3	236.4	-	-
37-G-205-80	55	180	55	8.00	21	9	15	9	30	0.24	29	0.3	0.7	216.6	12	-

Table 1. Water chemistry for selected water wells in the greater Calistoga area.

Sample number	Sampling temp.	Crist. (1)	Chalcedony (2)	Chalcedony (3)	Quartz (4)	Quartz (5)	Na - K (6)	Na - K (7)	Na - K - Ca (8)	Na - K - Ca (8')	Na - K - Ca (9)	(10)	w/ Mg (11)	w/ Mg (11')	Na-Li (12)
		t°C													
M-002-80	26	37	58	64	87	90	209	185	180	138	47	202	50	50	213
1-M-004-80	33	4	25	33	53	60	158	127	144	104	66	158	47	53	143
2-M-015-81	38	55	77	81	105	106	126	92	128	107	75	132	66	69	232
3-M-010-81	31	37	58	64	87	90	131	96	131	112	75	135	108	106	204
4-M-018-81	36	30	51	57	79	83	118	83	116	82	69	123	59	69	159
5-G-016-80	37	14	35	42	63	69	114	78	109	68	65	119	-	-	208
6-G-001-80	91	56	78	81	106	106	150	118	149	137	78	154	102	100	238
7-G-009-81	135	57	79	82	107	107	155	122	161	173	85	159	-	-	257
8-G-012-80	81	61	81	85	111	111	140	107	149	160	85	146	-	-	251
10-G-020-80	116	60	82	85	110	110	144	111	131	85	64	145	-	-	274
11-G-025-80	61	60	62	67	91	93	172	142	151	104	65	169	82	84	245
12-G-037-80	52	31	52	58	81	84	145	112	141	117	74	148	112	108	255
13-G-044-80	40	35	57	62	86	89	185	157	165	131	69	181	60	61	213
14-G-058-80	74	44	65	70	94	96	112	76	115	91	73	118	106	-	228
15-G-096-80	85	42	64	69	93	95	134	100	132	106	73	138	-	-	271
16-G-097-80	95	57	79	82	107	107	157	126	158	156	81	160	-	-	267
17-G-105-80	44	59	81	84	109	109	141	108	141	126	77	145	-	-	237
18-G-111-80	104	55	77	80	105	106	155	124	152	137	76	158	-	-	267
G-112-80	30	39	61	66	89	92	130	96	132	115	76	136	85	86	259
19-G-115-80	35	35	56	61	85	88	188	161	169	137	70	185	52	52	215
20-G-116-80	41	38	60	65	88	91	111	96	134	121	78	136	55	56	210
21-G-122-80	15	21	41	50	72	77	246	230	168	57	41	224	36	-	-
22-G-143-80	27	21	42	48	70	75	141	110	131	87	65	145	27	.39	-
23-G-181-80	20	7	27	35	55	62	137	104	123	72	61	139	-	-	37
24-G-201-80	64	32	53	59	82	85	181	153	161	121	68	178	52	55	222
25-G-206-80	55	32	53	59	82	85	145	113	136	96	68	147	103	102	236

Table 2. Chemical geothermometry for selected water wells in the greater Calistoga area.

maximum recorded bottom hole temperatures were only 124°C and 118°C respectively (Taylor, 1981). Both test wells were located at horizontal distances of less than 1,000 feet away from The Geysers (Fig. 2), where temperatures of 135°C were reached at 192 feet. Although the temperature logs of both of these wells show a continuous increase in temperature with depth, the lower maximum bottom hole temperatures indicate that the hottest water must be confined to the fracture system. If Na-K-Ca calculated temperatures from only those wells located directly along the geographic axis of the valley are considered, then higher, average reservoir temperatures are obtained, and it becomes reasonable to infer a temperature in excess of 150°C for the deeper aquifer.

Water chemistry in the Calistoga area can be used to characterize and distinguish between waters of a geothermal and nongeothermal origin or the geothermal component of water in wells with a mixed origin. All geothermal waters analyzed were classified as sodium chloride type and can be easily distinquished by concentration ranges of B (8ppm), Cl (180ppm), F (7ppm), and Na (170ppm). Fresh water on the other hand is characterized as bicarbonate type with relatively high concentrations of SO_4, Mg and Fe. Recent tritium isotope studies, completed on a well located on the southwest side of the valley, indicate that the geothermal compenent of the water sampled, is a least 32 years old.

The ratio Cl/B has been useful in distinguishing between geothermal and nongeothermal waters in a particular area (Ellis, 1970; Hull and Elders, 1984). Since boron and chloride act as soluble elements which are not controlled by temperature and pressure-dependent chemical equilibria, these ions will have a constant Cl/B ratio in waters of a geothermal origin, while waters of a nongeothermal origin tends to be more variable. Mixed waters also have characteristic water chemistries, but depend upon the relative contributions of each component end-member. Figure 3 gives the molecular proportions of chloride, boron and bicarbonate in a range of waters from the Calistoga area. The homogeniety of the geothermal water is evident; the hottest wells have analyses which plot at the low HCO_3 end

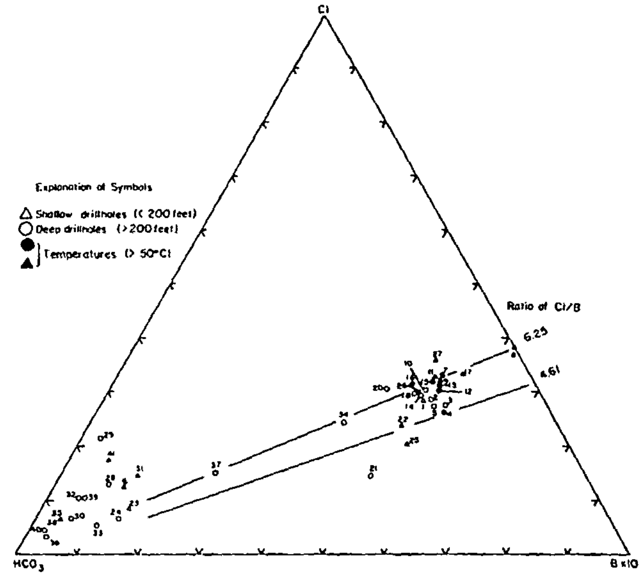

Figure 3. Molecular proportions of chloride, boron and bicarbonate in a range of waters from the Calistoga area.

of the constant Cl/Bx10 lines. The close grouping of points along the 6.25 Cl/Bx10 line is suggestive of a single aquifer, or a source that is chemically very homogeneous. The minor scattering of points about the 6.25 Cl/Bx10 line indicates local contamination. Of all the wells evaluated in the greater Calistoga area, those with the highest concentrations of B and Cl are located along the geographic axis of the Napa Valley, approximately coincident with the projected strike of the subsurface fault shown in Figure 2. This relationship between geothermal fluids that have a high boron content, and discharge along fault or fracture zones, has been used successfully elsewhere in California to both, locate faults and to trace waters transported upward along these shear zones (Barnes, 1970; Hull and Elders, 1984).

Within the Calistoga aquifer the highest TDS values, ranging from 600 to 900ppm also occur along the geographic axis of the upper Napa Valley. These moderately-high mineralized waters coincide with chloride values ranging from 183 to 229ppm (Table 1). Both the chloride anomaly and an outline of wells showing a high TDS content are notably linear and are located away from the borders of the valley as would be expected if the anomalies were structurally controlled (Fig. 2).

A decrease of chloride values and TDS, along with a general decrease in well temperature and pH occur at the border of the valley. The appreciable difference between the mid-valley waters in Calistoga and the border waters can be explained by the mixing of hot mineralized water with cool meteoric waters. Fournier (1976) indicates that the composition of a mixed water is likely to exhibit marked non-equilibrium between water and rock. High chloride, mid-valley waters are compared in Figure 4 to waters from the

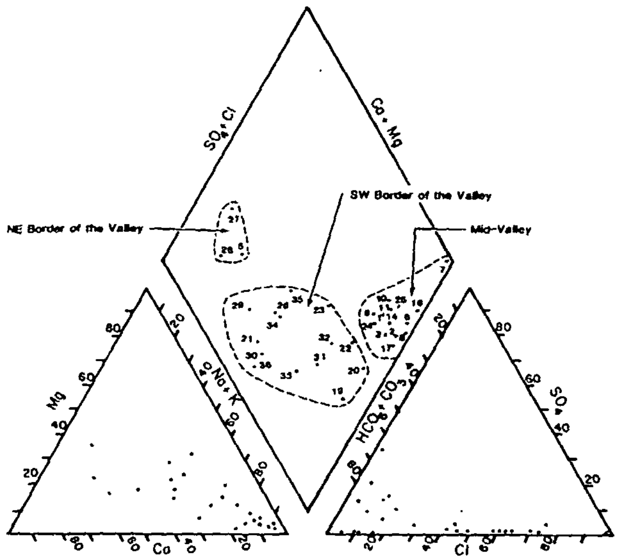

Figure 4. Trilinear diagram showing three distinct water types in the greater Calistoga area.

northeast and southwest border of the valley. The proportions of cations and anions in the groundwater from a variety of wells in the upper Napa Valley, exhibit three distinct water types when plotted on a trilinear diagram (Piper, 1953). Water from the southwestern border areas of the valley can be easily distinguished from mid-valley water by a more variable, but generally lower chloride compositions, and increased HCO_3, SO_4 and Fe concentrations. Water from the northeastern border areas, as typified by wells 5, 27, and 28 are even lower in Cl with higher concentrations of Ca and Mg. Well 7, plotting within the mid-valley, high-Cl range, most likely represents the true chemical characteristics of the deeper geothermal reservoir. This water is high in Cl (201ppm), B (9.8ppm) and F (11.5ppm) and low in Fe (0ppm), SO_4 (0ppm) and Mg (0.5ppm). It has a calculated Na-K-Ca temperature of 159°C. The scatter associated with the remainder of the analyses plotting within the NaCl-type range probably reflect minor mixing with surface waters.

CONCLUSIONS

The source of the geothermal component of waters in the upper Napa Valley may be related to one of two possible shallow plutonic bodies located both north and south of the city of Calistoga. Geophysical and geochemical evidence suggests the presence of a subsurface fracture zone approximately coincident with the geographic axis of the valley. The fracture zone appears to act as a conduit for the upward migration of fluids. The hotter wells in the Calistoga area are distributed along the geographic axis of the valley, have similar Cl/B ratios and are high in Cl, B, F and Na. These wells also indicate higher temperatures by geothermometry, implying a deeper aquifer source.

As the geothermal water seeps upward along the fracture zone it migrates laterally towards the margins of the valley, and gradually becomes enriched in Fe, SO_4 and HCO_3 by mixing with the cool, near surface groundwater. A comparison of these geochemical indicators on trilinear diagrams suggests various degrees of mixing between the geothermal waters and shallow fresh groundwater, at different locations throughout the greater Calistoga area, and can thus be used to trace and map the geothermal waters.

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APPENDIX A
Equations for Geothermometers

				Source
(1)	Cristobolite	$t°C = \frac{1000}{4.78-\log SiO_2} - 273.15$		(1)
(2)	Chalcedony	$t°C = \frac{1112}{4.91-\log SiO_2} - 273.15$	25-180°C (conductive)	(2)
(3)		$t°C = \frac{1264}{5.31-\log SiO_2} - 273.15$	100-180°C (adiabatic)	(2)
(4)	Quartz	$t°C = \frac{1309}{5.19-\log SiO_2} - 273.15$	0-250°C (conductive)	(1)
(5)		$t°C = \frac{1522}{5.75-\log SiO_2} - 273.15$	100-250°C	(1)
(6)	Na - K	$t°C = \frac{1217}{1.483+\log(Na/K)} - 273.15$	100-300°C	(3)
(7)		$t°C = \frac{933}{0.993+\log(Na/K)} - 273.15$	25-250°C	(1)
(8)	Na - K - Ca	$t°C = \frac{1647}{2.24+\log(Na/K)+B\log(\sqrt{Ca}/Na)} - 273.15$	4-340°C	(4)
		$B = 4/3$ for $\sqrt{Ca}/Na$ 1 & t°C 100 $B = 1/3$ for $\sqrt{Ca}/Na$ 1 & t°C 100		
(9)		$t°C = \frac{-22200}{\log(Na/K)-6.3\log(\sqrt{Ca}/Na)-64.2} - 273.15$	$t < 100°C$	(5)
(10)		$t°C = \frac{1416}{\log(Na/K)+0.055\log(\sqrt{Ca}/Na)+1.69} - 273.15$	$t > 100°C$	(5)
(11)	Mg correction for Na-K-Ca	$dt_{Mg}°C = 10.66-4.741R+325.87(\log R)^2 -1.032\times10^5(\log R)^2/T-1.968\times10^7(\log R)^2/T^2 +1.605\times10^7(\log R)^3/T^2$ $R = ((Mg/(K+Ca+Mg)) \times 100$	$T = K°$ $5 < R < 50$	(6)
(12)	Na - Li	$t°C = \frac{1000}{\log(Na/Li)+0.38} - 273.15$		(7)

conc. in ppm for eq. (1) thru (7); Molar for eq. (8,9,10,12); equivalents for (11).

sources: (1) Fournier (1977); (2) Arnorson (1983); (3) Fournier (1979); (4) Fournier & Truesdell (1973); (5) Benjamin & others (1983); (6) Fournier & Potter (1979); (7) Fouillac & Michard (1981).

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A GEOCHEMICAL MODEL OF THE CALISTOGA GEOTHERMAL RESOURCE NAPA VALLEY, CALIFORNIA

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ABSTRACT

The Calistoga Geothermal Field consists of a shallow, moderate temperature resource located at the head of the Napa Valley in northern California. Like many near-surface low-to-moderate temperature (<150°C) hydrothermal resources being developed in the United States, the Calistoga Field, is attributed to fault-charged, hydrothermal circulation. The Calistoga aquifer is recharged by thermal water upwelling along a central linear fracture system trending parallel to the axis of the Valley. Modelling of the field is accomplished by mapping the concentration of specific chemical tracers which are characteristic of the Calistoga geothermal fluid. Because of the shallow depth and moderate temperature of the field, the Calistoga area is very attractive for the development of direct-use projects.

INTRODUCTION

The Calistoga Geothermal Field, located near the head of the Napa Valley in northern California, is a shallow, fault-charged hydrothermal convection system of moderate temperature (Fig. 1). Two localized, shallow subsurface systems of elevated temperatures are indicated by a variety of geochemical parameters (Fig. 2). Chemical analyses of water obtained from 139 wells in the greater Calistoga area have been used to characterize and distinguish between waters of a geothermal and non-geothermal origin as well as water with a component of mixing (Murray and others, 1985). All geothermal waters analyzed were classified as sodium chloride type and can be easily distinguished from fresh water by elevated concentrations of Cl>180ppm, B>8ppm, F>7ppm and Na>170ppm. By noting the locations of water wells displaying the highest concentrations of these elements, the position of the deep water source, the boundary of the hot water area, subsurface geothermal flow patterns and the extent of dilution of the thermal water rising to the surface may be accurately determined.

DISCUSSION

Elements of a conceptual geochemical model of a geothermal resource must include a heat flux at the base of the system, a source of water and interconnected hydraulic conductivity. Hydrothermal convection systems readily develop in areas where there is a residual heat supply related to relatively young volcanism. Located immediately southwest of the Clear Lake Volcanic Field, which had activity into Holocene time, the upper Napa Valley is bounded by and underlain by Tertiary pyroclastic rocks that range in age from 3.0 to 9.0 my (Sarna-Wojcicki, 1976). In addition, recent evidence has suggested that the Napa Valley may have been subjected to minor volcanism as recently as perhaps 5000 years ago (Murray, 1986).

The heat source or "driving mechanism" for the hydrothermal convection system at Calistoga is probably the residual heat from the magma chamber(s) that were the source of these late Pliocene-Recent(?) volcanic extrusives.

The second requirement for a hydrothermal convection system, a source of water, is provided by meteoric water coming into contact with this residual heat source and then ascending along fault or fracture zones in the greater Calistoga area. The existence of faulting in the upper Napa Valley has long been suggested as the cause for the hot water seepage at Calistoga. Waring (1915), for example, speculated that faulting was chiefly responsible for the hot water resource; later, Faye (1973;1975) also inferred the existence of a fault aligned with the topographic axis of the upper Napa Valley to supply the geothermal fluids to Calistoga.

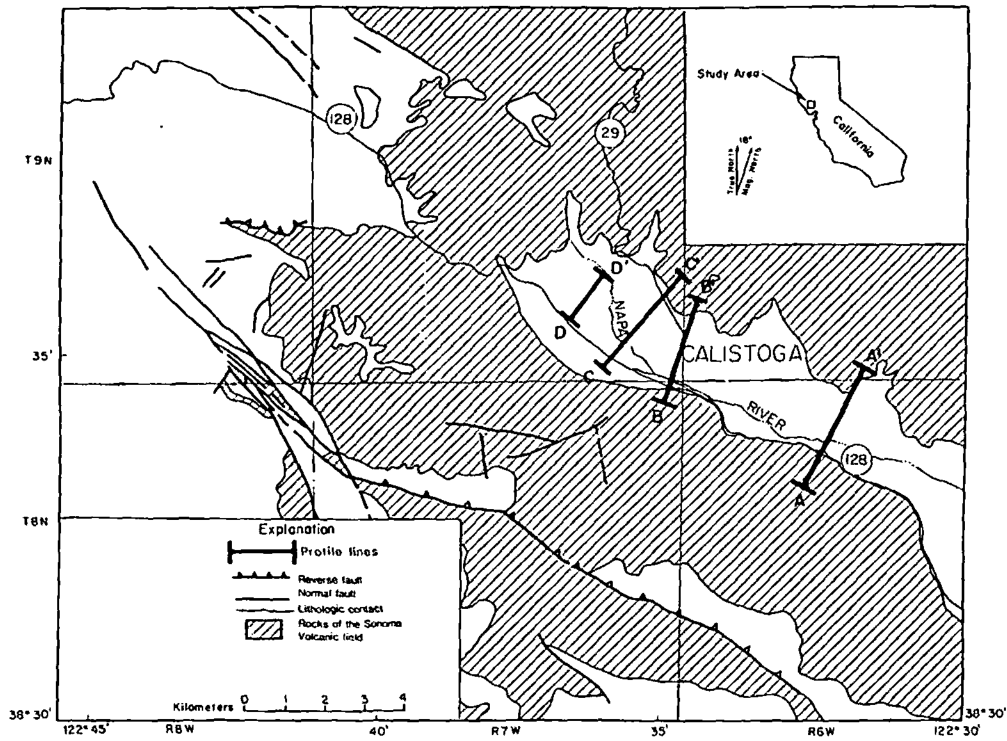

Figure 1. Map showing the general location of the Calistoga Geothermal Field in the upper Napa Valley, California. Cross-valley profile lines show location of wells used in the composite diagram shown in Figure 3.

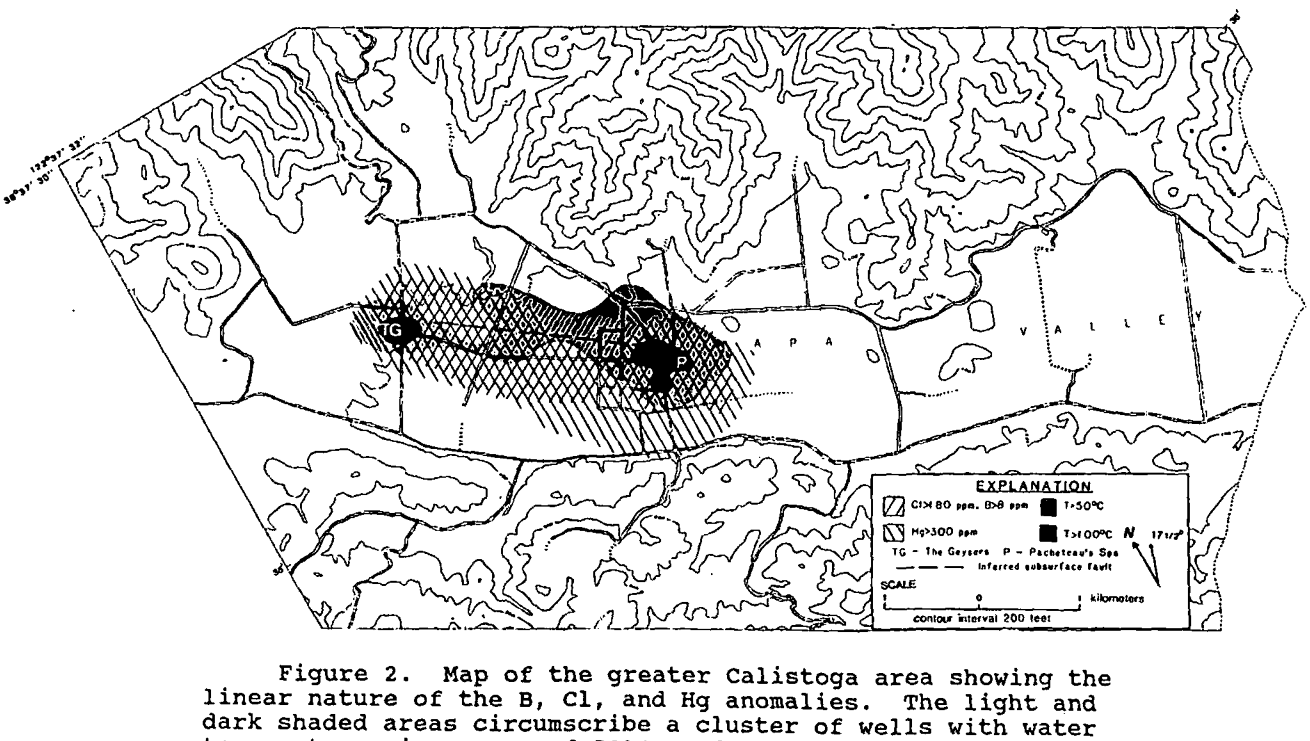

Figure 2. Map of the greater Calistoga area showing the linear nature of the B, Cl, and Hg anomalies. The light and dark shaded areas circumscribe a cluster of wells with water temperatures in excess of 50°C, and 100°C respectively.

Geophysical studies by Youngs and others (1980) at Calistoga, and an initial assessment of the geothermal resource at Calistoga by Taylor and others (1981) also indicated several areas of possible faulting associated with the areas of known geothermal resources, but were unable to confirm the actual existence of a fault or fracture zone. Finally, recent physical evidence developed during a test drilling program has also tended to support this hypothesis, but without confirmation (Murray, 1986).

The mapping of geochemical anomalies (areas of anomalously high concentrations of Cl, B, F, Na, and Hg) however, has been successfully employed elsewhere in California to not only locate faults but to trace waters transported upward along these shear zones (Barnes, 1970; Hull and Elders, 1984). The application of these same techniques to the Calistoga Field has resulted in a good correlation between geophysical data and the occurrence of the hot water resource. High chloride concentrations, in particular, have been used to locate the source of the deep hot water zones, with the highest chloride values representing the location of probable upwelling of thermal fluids from depth (Hull and Elders, 1984). In addition, Chloride, along with B, F and Hg can also be used to delineate the actual boundary of the geothermal resource.

Figure 2 is a composite diagram combining areas of anomalously high Cl, B, and Hg along with an outline of wells discharging water in excess of 50°C. The notably linear appearence of this diagram tends to confirm that the upwelling of geothermal fluids is localized along a vertically permeable fault or fracture zone trending parallel the axis of the upper Napa Valley. The upwelling geothermal fluids can also be recognized by a localized increase in the potentiometric surface, coincident with wells displaying the highest values of B and Cl. The location of the two areas with the highest surface discharge temperatures, the California Geyser at 135°C and Pacheteau's at 121°C are also aligned along the projected trace of the fault or fracture zone (Fig. 2). Both of these areas produce hot water from wells drilled to depths of 192 and 201 feet respectively.

A CONCEPTUAL MODEL

A conceptual model of the Calistoga resource based upon water chemistry and soil mercury profiling is shown in figure 3. The model is similar to other shallow hydrothermal convection systems found throughout the Basin and Range and Cascade Range provinces of California, Nevada and Oregon. Hot water flows up the vertical fault and spreads laterally into a relatively thin aquifer under pressure. The rock matrix above the aquifer is assumed to be sufficiently permeable to allow vertical mixing with the overlying fresh water aquifer (Murray, 1986). The fluid then flows through the aquifer losing heat by conduction to the caprock and basement.

A corollary of this model is the development of a distinctive temperature reversal below the aquifer. Figure 4 is a schematic diagram which illustrates the evolution of this temperature gradient with time. In the case of Calistoga, the graph can also be used to predict the temperature profile at any given location away from the fault.

Figure 5a is a series of temperature gradient profiles from five geothermal wells located in the upper Napa Valley along profile line D-D'(Fig. 1). The wells were logged originally by Occidental Geothermal Inc. and the data published by Youngs and others (1980). Figure 5b is an interpretive diagram showing the relationship of the wells to the geothermal aquifer and their relative distance away from the central fault or fracture zone. As shown in Figure 5a, two of the temperature profiles, wells 6 and 9, show a distinctive temperature reversal with depth, at 164 and 190 feet respectively. The profiles from other wells, while displaying no reversal, can be easily interpreted in terms of distance from the fault or fracture zone, or the depth of the well relative to the geothermal aquifer. For example, wells 7 and 8 show a continuous increase in temperature with depth, and are thus probably located within the permeable fracture zone. Whereas, wells 6 and 9 penetrated through the

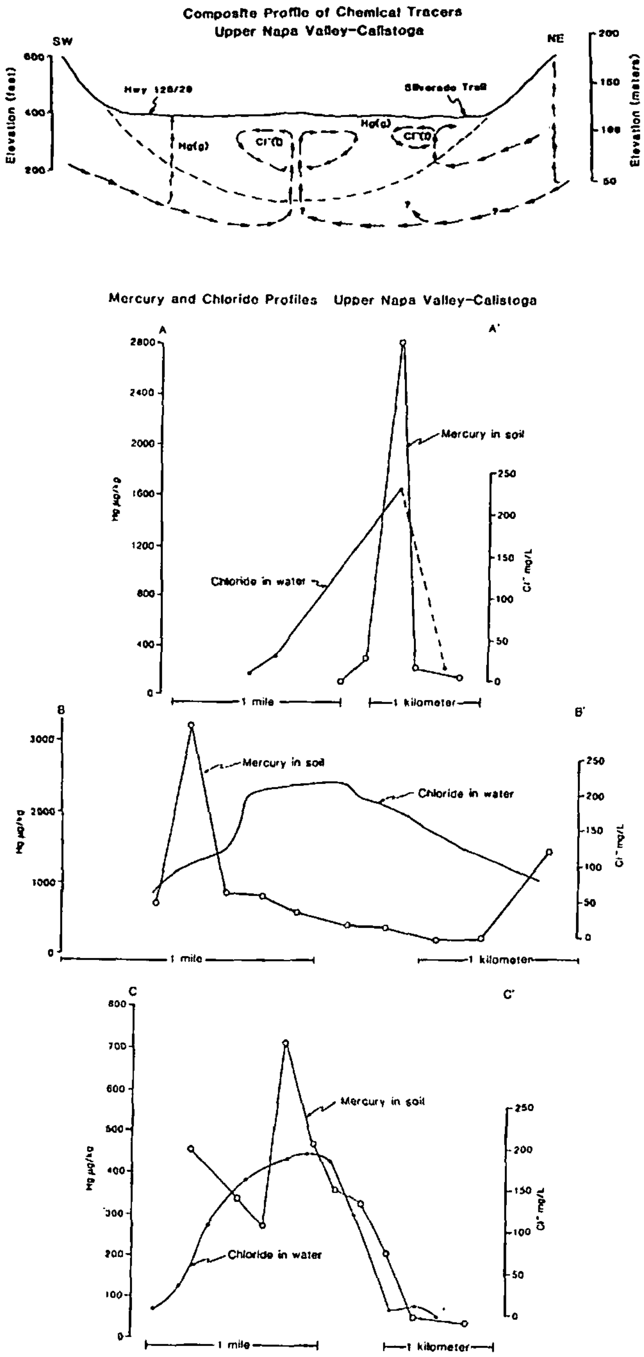

Figure 3. Conceptual model of the Calistoga Geothermal Field based on the concentrations of Cl and Hg. The lower cross-sections show the measured values of Cl and Hg along three separate traverses across the upper Napa Valley (see Fig. 1).

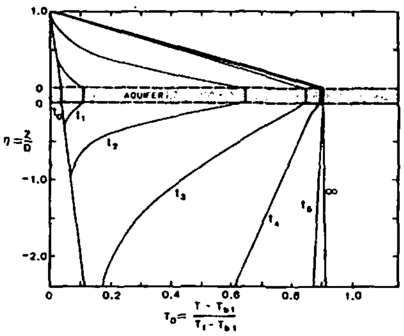

Figure 4. A schematic diagram illustrating the evolution of fault-charged hydrothermal systems with time. Modified from Benson and others (1981).

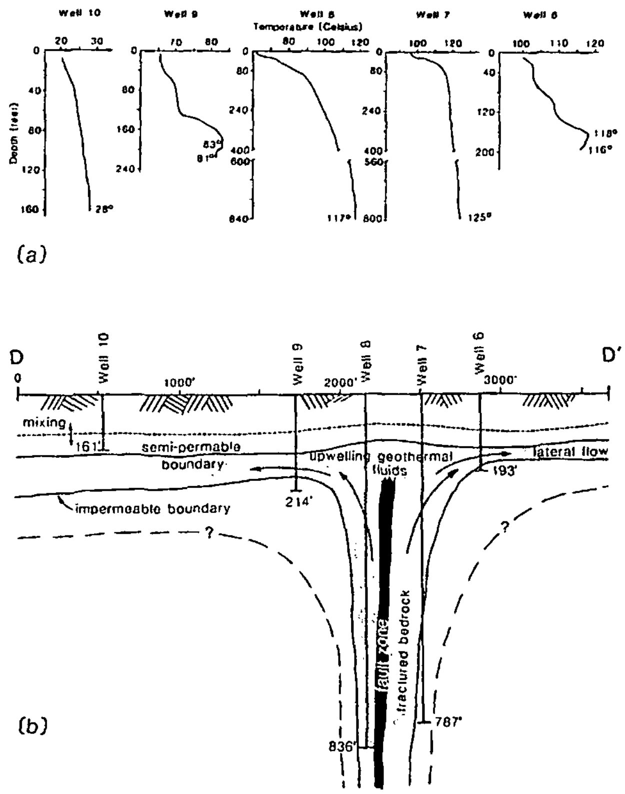

Figure 5. (a) Five temperature gradient logs selected from a series of geothermal wells drilled by Occidental Geothermal Inc. (b) Diagrammatic cross-section across the Napa Valley illustrating the relationship between the temperature gradient profiles shown in (a) and the geothermal aquifer.

;eothermal aquifer into a zone of :ooler temperature. Well number 10, on :he other hand, shows no change in :emperature gradient with depth and lischarges water of a much cooler :emperature than the other wells nentioned. This suggests that well 10 s probably not of sufficient depth to ntersect the main body of the eothermal aquifer (Youngs and others, 980).

RESERVOIR VOLUME ANALYSIS

The Calistoga geothermal reservoir s a complex, heterogeneous system of ock and water. The temperature ontour of 50°C, indicated on Figure 2, ircumscribes a cluster of water wells ithin the greater Calistoga area that ischarge water in excess of 50°C below depth of 200 feet. The boundary line as drawn to enclose the wells, then odified slightly to fit geophysical nd geochemical evidence as ppropriate. The boundary as drawn, erves as an estimate of the lateral xtent of the geothermal aquifer, an rea of approximately 5.79 square iles.

Youngs and others (1980) vertically ivided the upper Napa Valley into four ıbsurface zones. From top to bottom ıese are: (1) alluvial sediments com the surface to 120 feet in depth, 2) alluvial sediments below 120 feet ıd extending to the top of the ıderlying southwest dipping pyroclas- .c beds (a cumulative thickness proaching perhaps 1400 feet on the uthwest side of the valley).) impermeable pyroclastic material mposed wholly of volcanic ash and h-flow tuff; and, (4) saturated roclastic material and interbedded diments underlying zone 3.

Zone 2 appears to represent the main dy of the Calistoga reservoir. The erage thickness of the aquifer was timated by Youngs and others (1980) be approximately 640 feet. This ickness represents a volume of rock d water above 25°C and is considered conservative estimate based on the ility of the aquifer to transfer heat conduction, with time, to the rrounding rock matrix. The actual servoir thickness, as shown in gure 3, is probably relatively thin, rhaps no more than 100 feet, based on mperature gradient profiles and bsurface water chemistry.

A reservoir volume can thus be calculated from the boundary determinations and an estimated average thickness. From these calculations, a conservative, steady state aquifer yield of approximately 13,500 to 20,500 acre feet or 4.4×10^9 to 6.6×10^9 gallons has been estimated. Current withdrawal of water from the resource by bottlers of mineral water and spa operators is estimated at 55×10^6 gallons per year. Assuming no recharge to the system and no further development, the reservoir would thus be expected to last approximately 100 years.

Geochemical mapping, however, has demonstrated that recharge of the resource is taking place along a central fault or fracture system, suggesting that further development of the resource is feasible. The rate of natural charge and the rate of downstream discharge (surplus water) currently leaving the the system, however, is not known and cannot be accurately determined. Thus an accurate prediction of reservoir longevity is not possible at this time. The determination of reservoir depletion, however, can be made through careful well monitoring, and/or numerical modelling techniques. Detectable drawdown of the potentiometric surface and seasonally corrected declines of chemical tracers such as chloride, in selected wells, will serve as a measure of the rate of reservoir depletion.

he thinks he has proved the fault w/ geochem.

CONCLUSION

Geochemical mapping of the Calistoga Geothermal Resource Area has led to the development of a conceptual model which satisfactorily explains the chemical and thermal characteristics of the reservoir. Anomalously high values of Cl, B, F, Na and Hg were used to locate a central fault or fracture system of high vertical permeability through which water flows to the surface.

A volumetric analysis was completed based on geochemistry, geophysics and the distribution of thermal wells displaying temperatures in excess of 50°C. Recharge versus withdrawal rates suggest that the resource could undergo reasonable development without detriment; however, accurate depletion rates must be based on subsequent well monitoring and/or numerical modelling studies.

ACKNOWLEDGEMENTS

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ANALYSIS AND INTERPRETATION OF THERMAL DATA FROM THE BORAX LAKE GEOTHERMAL PROSPECT, OREGON

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INTRODUCTION

The results of geothermal exploration at the Borax Lake geothermal prospect, Harney County, Oregon with emphasis on interpretation and thermal modeling of temperature gradient data are presented in this paper. The total heat loss of the Borax Lake system is calculated and compared to other Basin and Range geothermal systems. Thermal models are developed to test the hypothesis that the location of geothermal activity at Borax Lake is controlled by the structures and/or stratigraphic section associated with the buried horst block. In addition, downward continuation modeling is used to estimate the subsurface configuration of isotherms ranging from 160 to 190°C. The Borax Lake area of southern Oregon is located in the northern part of the Basin and Range province. It lies in the Alvord Valley, a north-trending, complex graben between the horst blocks of the rugged Steens Mountains to the west and the Trout Creek Mountains to the east (Figure 1).

The rocks exposed in the ranges consist primarily of middle to late Cenozoic volcanic rocks, including the thick sequence of Miocene Steens basalt, which is time equivalent to the Columbia River basalt of northern Oregon and southern Washington. The basalt unit is underlain and overlain by more silicic volcanic units. Volcaniclastic rocks are interbedded with the volcanic rocks. Alvord Valley is postulated to be the site of the Pueblo caldera (Rytuba and McKee, 1984), the source of the major ash flow tuff unit found in the Trout Creek Mountains. The tuff has a volume of greater than 290 km² and age of 15.8 m.y. The upper part of the geologic section in the Alvord Valley consists of several hundred meters of late Cenozoic lacustrine and fluvial units deposited in the subsiding graben. Very little information is available on the nature and composition of rocks beneath these shallow units.

A number of geophysical surveys have been conducted in the vicinity of the Borax Lake geo-

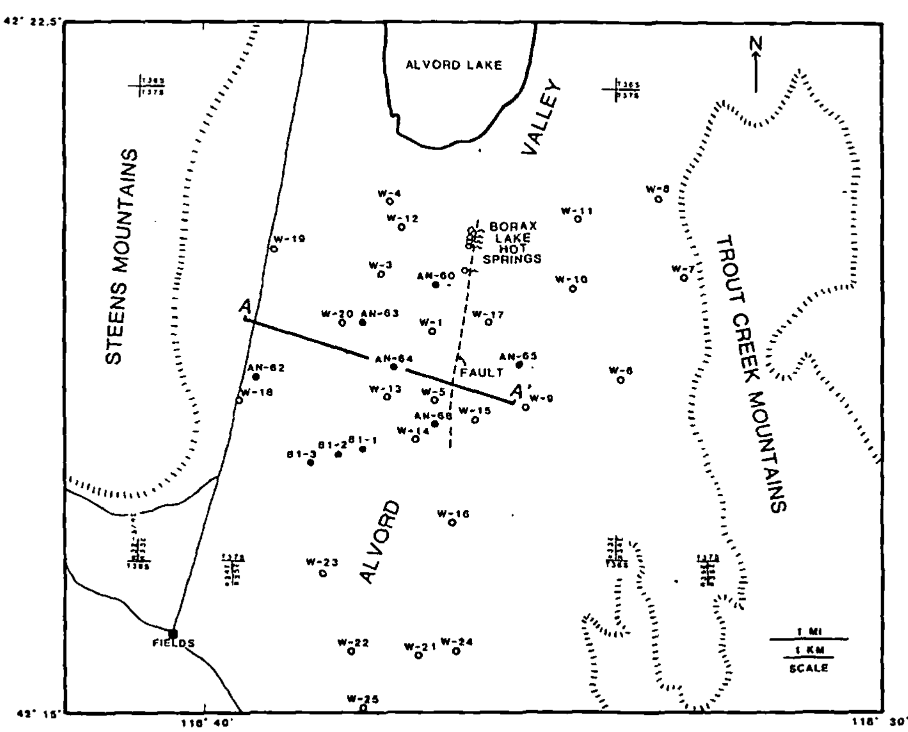

Figure 1. Location map of Borax Lake area. The circles represent drill holes less than 76 m deep; the dots represent drill holes deeper than 76 m.

thermal system. Both gravity (Cleary and others, 1981, and unpublished Anadarko reports) and seismic reflection surveys (unpublished Anadarko reports) indicate that a buried, north-northeast trending horst block exists beneath the geothermal area. The location of the fault that bounds the eastern margin of this horst block coincides with the location of the hot springs at Borax Lake (Figure 1). The horst block becomes narrower and has less relief to the north, and may not be present north of the south end of Alvord Lake.

Geothermal activity is found at two other locations in Alvord Valley. Mickey Springs are approximately 45 km northeast of the Borax Lake Hot Springs. Alvord Springs emerge from a fault along the eastern front of the Steens Mountains at a locality about 25 km north-northeast of the Borax Lake Hot Springs. The regional heat flow in this part of Oregon is 60-100 mWm^{-2} and the regional geothermal gradient is 40-60°C/km (Blackwell and others, 1978), based on sparse data. Since no Pliocene or Quaternary volcanic rocks are found in the area, there is no evidence that a volcanic heat source is responsible for the geothermal activity in Alvord Valley. Thus the heat source for the geothermal systems appears to be the natural heat flow of the region. The minimum depth of circulation of ground water required to reach the highest observed temperature in the Borax Lake system of 160°C (320°F) would be 3-4 km, and the minimum depth to reach the inferred reservoir temperature of 190°C (375°F) would be 4-5 km. This situation is analogous to the Basin and Range province of northern Nevada which is the site of several major geothermal systems such as Desert Peak, and Dixie Valley. These geothermal systems are described by Edmiston and Benoit (1984) and by Benoit and Butler (1983).

DISCUSSION OF THERMAL DATA

The thermal modeling presented in this report is based on data collected during the exploration process (Gardner and others, 1980; and Nosker and Nosker, 1981). Information from interpretations from the gravity and seismic reflection surveys is incorporated into the structural features of the geothermal models.

A location map of the temperature gradient holes in the area is shown in Figure 1. The average geothermal gradient for the depth interval of 61 to 76 m in each hole based on the reports is shown on Figure 2. There are two different sets of temperature-depth data. The holes with the W designation were obtained from Union Oil Company. This series of holes was drilled to a nominal depth of 76 m. Another series of holes was drilled by Anadarko to a nominal depth of 150 m. The geothermal gradient contours shown in Figure 2 are based on computer contouring of the data by GSI, Inc. An area of over 20 km^2 (15 mi^2) has gradient values of at least 100°C/km (5.5°F/100ft), approximately twice the regional average gradient, and an area of 2 km^2 (1.5 mi^2) has gradient values greater than 340°C/km (19°F/100ft).

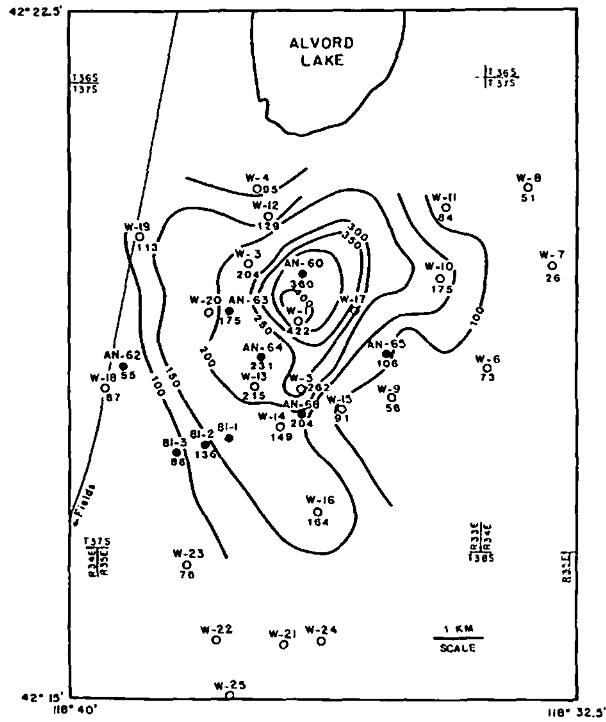

Figure 2. Map of geothermal gradients for the depth interval 61 to 76 m. Computer contoured by GSI, Inc.

A summary string plot of the temperature-depth curves from selected holes is shown in Figure 3. The typical background gradients in the valley sediments range between 35 and 60°C/km. Examples are W-6, 9, and 25 and AN-62. Most of the curves are linear, indicating conductive heat flow. Temperature-depth curves for holes W-13, W-17, and 81-1 overturn or are strongly curved, indicating lateral flow in shallow aquifers (Ziagos and Blackwell, 1986). Holes AN-68 and W-5 are strongly nonlinear; the nonlinearity may be due to major changes in thermal conductivity with depth or to water flow effects. The holes with nonlinear gradients generally lie over the buried horst.

Only two holes are available that exceed 150 m in depth, one drilled by Anadarko (AN-64R), and one by Union (W-1). AN-64R was drilled to 347.5 m, but was plugged back to 250 m before completion because of pressure in excess of hydrostatic in a fractured basalt at 347.5m. Also shown in Figure 3 (as an inset) is a temperature-depth curve for the Union deep hole W-1. The curve plotted for W-1 does not represent equilibrium, but is the one measured the longest time after completion of drilling.

The temperature-depth curves in both of the deeper holes are not characteristic of either conductive or convective heat flow alone. Convective effects clearly influence the temperature-depth curves in these deep wells, but the curves do not become isothermal or have negative gradient sections. Since the drilling and completion history of both of the deep holes is complicated, the curves are open to different interpretations.

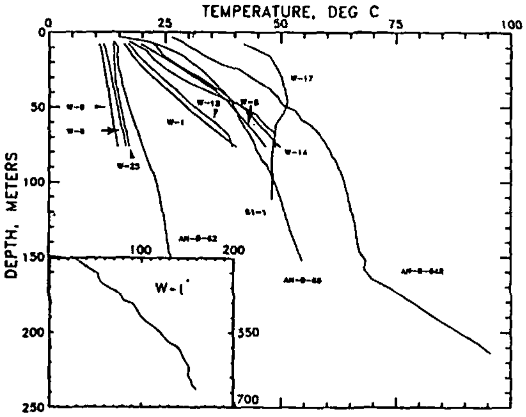

Figure 3. Temperature-depth plot of selected holes. The non-equilibrum W-1 log was made 106 hours following completion of drilling.

As part of this study, the thermal data were reinterpreted. The temperature-depth data were reanalyzed and interpreted geothermal gradients used in subsequent modeling are shown in Figure 4. Heat flow values corresponding to the redetermined gradients were calculated, but since the shallow thermal conductivity does not vary, the heat flow and geothermal gradient contours are identical in shape. No terrain corrections to the gradients are necessary because of the very low relief in the valley. The gradients shown in Figure 4 were contoured by hand to serve as a second anomaly pattern for interpretation in addition to that shown in Figure 2 and are referred to as the "deep" gradients (see discussion below).

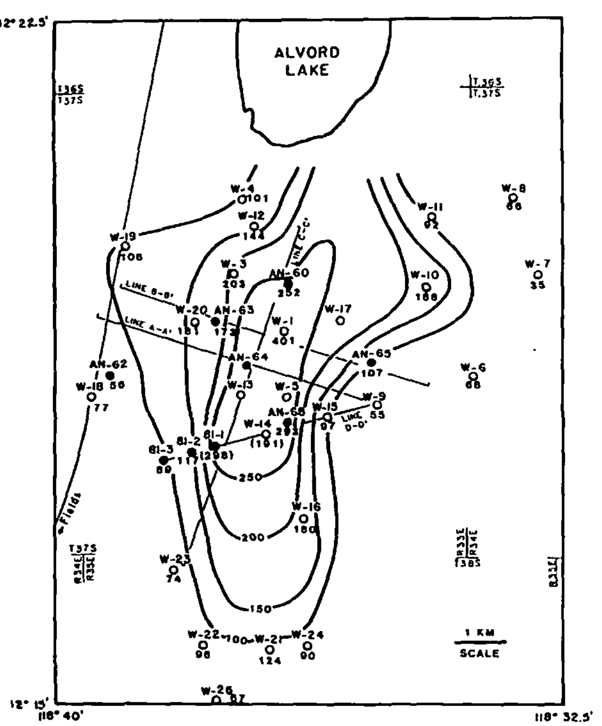

Figure 4. Map of "deep" geothermal gradients. The high gradients due to shallow leakage over the horst were not used in preparing the map (see text). These contours were used in the downward continuation modeling.

HEAT LOSS

One of the characteristic parameters of a geothermal system is the rate of heat loss i.e., the excess heat above the background transported through the system by the convecting ground water. The flow rate of the Borax Lake hot springs is approximately 3500 l min^{-1} and the exit temperature is 96°C (205°F) according to Brooks and others (1979). The estimated reservoir temperature ranges from 165°C (329°F) based on the quartz adiabatic geothermometer, to 176°C (349°F) based on the Na-Ca-K geothermometer, to 191°C (376°F) for the "best in situ reservoir temperature" (Brooks and others, 1979). The actual heat loss, assuming a temperature drop of 81°C (the exit temperature minus the surface temperature), is 2.0×10^7W. The overall heat loss of the system, assuming that the temperature field of the system is in steady state, is calculated using the reservoir minus the surface temperature. That value is 4.3×10^7W if the reservoir temperature is 190°C. The difference in the two values should be the heat lost to the surroundings during flow of the water through the system.

The heat lost by conduction can be calculated by integrating a heat flow contour map for the prospect. This calculation gives a value of 4.2×10^6 W above the assumed background of 80 mWm^{-2}.

The addition of the actual convective heat loss and the conductive heat loss for the prospect gives a total heat loss for the Borax Lake system of 2.7×10^7 W. This value compares to the value calculated from the assumed reservoir temperature and the observed flow rate of 4.3×10^7 W. Since either method of calculating the total heat loss could easily have an error of ±20%, the agreement of the two figures is satisfactory. A good estimate of the heat loss of the prospect is $3-4\times10^7$ W.

This heat loss is equivalent to the total heat loss due to the regional heat flow over an area of 50 km^2. The heat loss compares to values that range from 10^5 to 10^7 W for geothermal systems in the Basin and Range province such as Grass Valley, Nevada, Roosevelt, Utah, Desert Peak Nevada and the geothermal systems in the Imperial Valley in California.

FORWARD THERMAL STRUCTURE MODELS

As a second stage in the thermal interpretation the thermal structure in the valley was evaluated using forward modeling. In this approach the configuration of the heat source causing the anomaly was assumed and theoretical heat flow profiles calculated for comparison to the observed anomaly. The technique used was a two-dimensional finite difference solution. The structure of Alvord Valley has been explored by the seismic reflection technique and by gravity. A simplified cross section based on the interpretation of the geophysical data along profile A-A' on Figures 1 and 4 is shown in Figure 5.

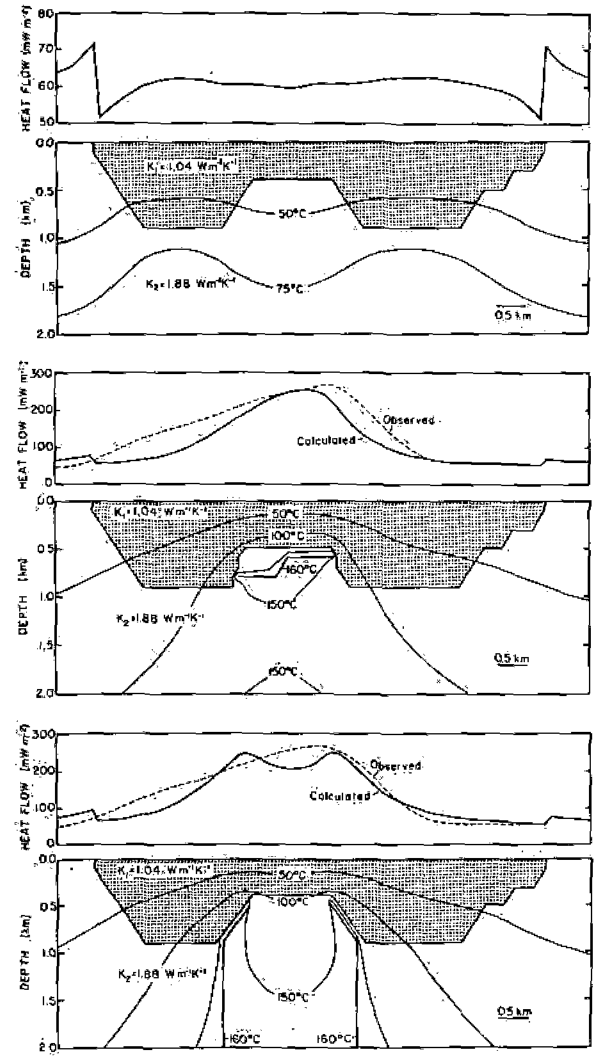

Figure 5. Forward numerical models. The contact between the bedrock and the valley fill (stippled area) is generalized from seismic data along section A-A'. The generalized heat flow based on the gradient data from Figure 4 is shown as the dashed line on Figures 5b and 5c.

The model shown in Figure 5a is designed to investigate the magnitude of thermal conductivity refraction effect. The nature of this effect and its influence on the thermal conditions in the Basin and Range setting are discussed by Blackwell and Chapman (1977) and Blackwell (1983). The thermal conductivity refraction effect is caused by differences in thermal conductivity between the low-conductivity, valley-fill sediments and the higher-conductivity volcanic rocks. The regional heat flow used in this model was assumed to be 60 mWm^{-2}, and the thermal conductivity values assumed are shown. The assumed surface temperature in all of the models was 15°C.

Based on this background, conductive-heat-flow model, the surface heat flow would be depressed valley over the horst by a few percent. In addition, larger heat flow anomalies of ± 15% would be present at the contact between the volcanic bedrock of the ranges and the valley sediments. The temperature at a given depth in the valley would be higher than the temperature at that same depth in the volcanic bedrock because of the difference in thermal conductivity. The temperature in the horst block would be intermediate between the valley and the range.

Even though the geothermal system is due to transfer of heat by convection, in many real geologic situations, conductive modeling can be used to evaluate the geometry of the "reservoir". For the conduction modeling to apply, the fluid circulation must be confined to discrete paths rather than circulating freely through a "homogeneous, porous medium." These discrete paths can be treated as boundaries of known temperature, and their geometry can be inferred from conductive models of the surface heat flow or subsurface temperatures outside the circulation paths. The basis of this approach is discussed by Blackwell and Chapman (1977) and by Brott and others (1981). The approach is particularily useful in the early stages of prospect evaluation when a minimum of subsurface information needed for convective modeling is not available.

Two forward models of possible isotherm configurations associated with a geothermal system (Figures 5b and 5c) were calculated. In the model shown in Figure 5b, the top of the system was approximated as a faulted horizontal plane of constant temperature (160°C) at depths of 400 and 800 m. In the model shown in Figure 5c, the heat source was approximated by two fault zones at a constant temperature of 160°C. The location of the two fault zones was assumed to coincide with the edges of the horst. The assumed temperature is conservative because it is the temperature actually observed in hole W-1. The temperature of the heat source (reservoir) could be higher (and the sources correspondingly deeper) without a major change in the isotherms calculated away from the heat source as long as heat transfer is conductive throughout the region outside the paths of circulation modeled as isothermal surfaces.

The actual heat flow anomaly is asymmetrical along cross section A-A', and neither of the heat flow curves calculated from the finite difference models match this asymmetry. When compared to the heat flow anomaly predicted from the models, the actual anomaly is steeper on the east side and less steep on the west side. Thus if taken literally, the actual source would be shallower or have a steeper dip on the east side than either model configuration. On the west side the actual source would be deeper, or have a shallower dip than assumed in the models.

The model with a near-horizontal source (Figure 5b) has gradients that are too steep on both sides and does not match the width of the observed anomaly. Thus the source would appear to be broader than the model configuration. The model of the system as two discrete fault zones (Figure 5c) has a heat flow anomaly that shows two discrete peaks, in contrast to the heat flow anomaly from the actual system. However, the calculated width of the anomaly matches the observed anomaly width better than the single source model. The calculated anomaly shape also matches the observed anomaly shape closely on the east side of the profile.

The model with a source within the horst (Figure 5b) appears to give the better fit to the overall shape of the observed anomaly. However, the discrete heat flow peaks predicted by the two-fault model (Figure 5c) may be obscured by shallow horizontal flow, and/or the gradient data may be sparse enough that such rapid lateral variations are not resolved.

DOWNWARD CONTINUATION MODELING

Finally an inverse modeling technique was used to evaluate the possible subsurface thermal condition based on the observed surface pattern. The downward continuation technique (Brott and others, 1981) was used in this part of the study.

Gradients from the 60-150 m depth range were plotted and contoured (Figure 4). Information on the structural geology of the area, provided by a seismic reflection survey across the thermal anomaly, was used to guide the location of the geothermal gradient contours where control was sparse. The resulting observed profile across the thermal anomaly along cross section A-A' is shown on Figure 6. The gradient was determined at equal intervals (600 m), by interpolation. The selection of data spacing is equivalent to low pass filtering of the observed anomaly (see Li and others, 1982, for a different approach). The data set is to sparse to allow more complicated filtering techniques. The equally spaced gradient data were then input into the downward continuation modeling program to determine the depths of the 150, 175, and 190°C isotherms.

The very high gradients (greater than about 250°C/km) found in some of the shallow holes over the horst (see Figure 2) must be related to the shallow circulation of hot water derived from a deeper source. If the gradients of 350°C/km found in some of these shallow holes were projected to depth, the predicted temperatures would be much higher than those actually observed in the two deepest holes. One of the major assumptions associated with the downward continuation technique is that no heat sources exist between the surface and the "target" reservoir. The existence of shallow leakage violates this assumption. The only drill hole that can be directly used to characterize the geothermal gradient due to the deeper source is W-1, the 600 m deep hole. The geothermal gradient associated with the deeper source estimated from W-1 is approximately 250°C/km, which is well below the peak values of over 400°C/km associated with the shallow leakage. So in order to model the temperatures at depth due to the deeper heat source, a maximum gradient of 250°C/km estimated from W-1 was assumed to be the gradient over the horst (site of holes W-1, W-5, W-13, W-14, AN-64, AN-68, 81-1).

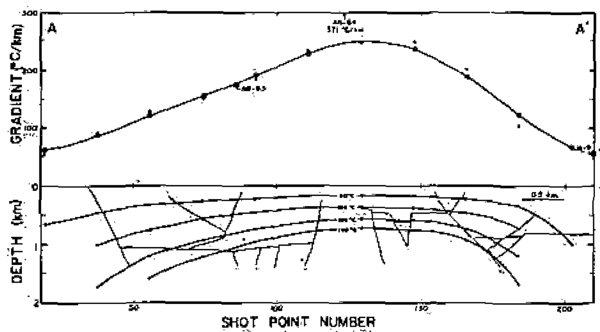

Figure 6. Isotherms along cross section A-A' calculated by the technique of downward continuation. The isotherms are superimposed on the structure along the section based on the gravity and seismic interpretations. The temperature gradients used in the interpretation and the fit of the calculated gradients are shown also.

The inferred 190°C (375°F) isotherm shown on Figure 6 is approximately 800 m (2400ft) below the surface at its shallowest point. The calculated shape of the isothermal contours corresponds with the NNE trending horst. The fault on the east side of the horst as determined from the seismic data almost coincides with the calculated position of the 150-190°C isotherms. Of course, any of the isotherms could satisfy the surface thermal anomaly, and without direct measurement of the reservoir temperature, there is no unequivocal way to associate one isotherm with the reservoir. Nonetheless, given the temperature measurements in the hole W-1, one interpretation would be that the 160°C isothermal contour as mapped by interpolation using the results shown in Figure 6 might represent the outline of the reservoir. If the acutal reservoir temperature is 190°C the system would be slightly deeper.

The downward continuation model does not account for variations in thermal conductivity, thus the motivation for the forward modeling. Based on this modeling, the thermal conductivity refraction effect should not greatly influence the position of the isotherms at depths less than 900 m. The model given here is based on the budget was cut back on 29th.

She is very disappointed that she got a severe cut in work scope due to budget-planning, etc. Holds PARS responsible.

Hydrodrills vs Turbos - letter - She really will not debate -- her engs says O&G can't 450° for hostile environ -- feels can extend temps rise by 50°.

assumption that temperatures will continue to increase with depth, which is consistent with the data from W-1. However, a sudden decrease of temperature with depth in the lower portion of the shallow system may occur, and our model would no longer be valid. If convective fluid motions occur in the valley sediments then the major assumption on which the interpretation is based breaks down. The final limitation is that there are large parts of the area that are too sparsely sampled for the horizontal gradients in thermal gradient (heat flow) to be well enough determined to constrain the continuation.

CONCLUSIONS

In spite of the limitations of the modeling several important conclusions about the Borax Lake geothermal system can be reached:

1) The Borax Lake system is a major Basin and Range geothermal system with a heat loss of $3\text{-}4\times10^7$ W. The heat source is the regional heat flow of the area.

2) The system is associated with a horst block in the center of the valley as delineated by interpretation of gravity and seismic studies.

3) Temperatures of 160°C (320°F) are measured in a 600 m (2000ft) hole in this system and geochemical thermometry suggests temperatures of 190°C (395°F).

4) Based on the downward continuation modeling an area in the subsurface at a depth of approximately 900 m (300ft) with a size of at least 6 km^2 (2.5 mi^2) and possibly as large as 15 km^2 (6 mi^2) has temperatures in excess of 160°C (320°F).

5. The close match between the observed and calculated anomalies along the east side of the thermal profiles is strong evidence that the fault bounding the east side of the horst is one of the major controls on the geothermal system.

The geothermal system is clearly associated with the horst. The nature of the association could be of at least two different types, which cannot as yet be resolved. The circulation could be controlled by permeability in the volcanic basement rocks or by permeability associated with fracturing and faulting relating to the formation of the horst. The surface leakage is associated with the fault that bounds the horst block on the east side so some circulation is obviously associated with faults. While the model of the source as confined to the horst seems to be the preferred one, the geometry of the thermal structure within the horst cannot be determined on the basis of available data. The actual area that might be suitable for geothermal exploitation cannot be determined until deeper drilling can resolve the nature of the flow controls. Nonetheless the Borax Lake geothermal system appears to be a major resource area and to have potential for exploitation.

ACKNOWLEDGEMENTS

The authors thank Anadarko Production Company for support for this study and for permission to publish this paper.

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Sue P (522-3356) in Idaho Falls

1. Sue tried to get us Thursday -
   showed up - 800P @ Sun Valley - no outside access from S.V.
2. DLN called Sunday - her initial impression plan ok because in accord w/ comm Hq plan.
3. Her biggest disturbance - drill budget 534 ⇒ 334 -- lost 200K in few weeks. WLF says have M.etc to do such --
4. Sue P has real probs of not getting a complete budget each update -

5. my Sect
   (1) Response Section is realistic but not not -- realistic -- Holding up
   ⟹ talk to her

6. Concerned that basic vendor costs have gone up 5%

7. Magma 60-65 Ē Jan - in Albuquerque
   Call Joel

8. got call DLN 4:45 P. -
   agreed on a compromise - everyone feels ok about it.
   Prob. - DLN playing it very conservatively -- she talked him into taking part of coring to drill -
   If take 70°, go to 12,500 ft, then this saves 2 days drilling -- put this into deepening rig hole.
   Power - take care of DLN -- I let him get us in a bind. She holds WRST responsible -- got into Sept -- we are already committed. DLN has come 180° -

FRACTURE DEFINITION: FORMATION, PRESERVATION, DELINEATION

Dennis L. Nielson, Joseph N. Moore, Phillip M. Wright
University of Utah Research Institute

INTRODUCTION

In our experience, all high-temperature geothermal production occurs from fractured rocks. However, the mapping and definition of these fractures have proven to be difficult. The Reservoir Definition Program at UURI has concentrated on the development and application of geological, geochemical and geophysical techniques which aid in the definition of fractures in the geothermal environment. The locations of important faults and fractures can often be predicted by placing the geothermal occurrence within a regional structural model that relates faults and fractures to causative stress fields. These provide temperature and compositional information on pathways which have been sealed through hydrothermal alteration as well as those which are presently active. Fluid inclusions frequently also can provide a convenient method for determining the temperature of a resource before temperatures within geothermal wells have a chance to re-equilibrate. Since surface geophysical methods lack the precision necessary for fracture detection at reservoir depths, borehole geophysical methods are being investigated.

Our geological and geochemical studies have concentrated on down-hole samples from active high-temperature geothermal systems. The following documents briefly the studies which have been completed in these systems. Following this, our borehole geophysical modeling studies are briefly described.

VALLES CALDERA

Studies in the Valles caldera have utilized samples which were donated by UNOCAL as well as samples recently acquired through Continental Scientific Drilling program.

Hydrothermal brecciation along structural zones may be an important process in enhancing permeability. A model relating the pressure controls of boiling to the pressure required for hydrofracturing has been developed to explain the process of hydrothermal brecciation. Data from fluid inclusions confirm that the boiling process was the causative mechanism for brecciation.

Studies in the Valles have also concentrated on intracaldera sandstones as guides to evolution of the caldera complex. Sandstone marker horizons are currently being investigated in detail to modify and improve the intracaldera geologic history. These studies have already revealed the following new information. Sandstone horizons, along with enclosing ash-flow tuff sequences, have been tilted up to 50^{o} from their originally horizontal configurations. This may be the result of caldera resurgence, or it could reflect post-depositional slumping or gravity sliding. The sandstones clearly have served as thermal aquifers, second only in importance to faults, fractures and breccia zones. They are intensely altered. Pumice fragments were initially very porous, but have since been clogged with hydrothermal illite, fluorite, quartz, calcite and chlorite. Chlorite is definitely the latest secondary phase to be deposited; it has been observed forming delicate microcrystalline rosettes encrusting euhedral fluorite crystals.

The alteration of these sandstones illustrates an important reservoir concept. Although these rocks originally had a high permeability, they have been effectively sealed through hydrothermal alteration. The continued maintenance of fluid pathways in such rocks requires fracturing to re-open areas sealed through hydrothermal alteration.

In addition, these studies have demonstrated that stratigraphy is an extremely important structural tool. Analysis of subsurface samples has defined the location of major structural zones which were produced during the formation of the caldera, but which have since been buried by the products of the volcanic eruptions.

COSO

The need to develop better models of the permeability variations and fluid flow patterns in a high-temperature fractured reservoir prompted us to initiate detailed sampling of core from the Coso geothermal system. Thermal gradient hole 64-16 was chosen for the initial work because of the extensive amount of geothermal veining in it. This well is located approximately two km from the main production area.

Two stages of geothermal alteration are apparent in 64-16. The earlier stage is characterized by silica and pyrite deposition in the intensely brecciated rocks penetrated in the upper half of the well. The second stage is characterized by the deposition of calcite. Geothermal alteration of the granitic wallrock has resulted in the formation of a highly ordered mixed-layer illite-smectite with 10-20% smectite.

Fluid inclusion heating and freezing measurements have been performed on calcite from the well. The inclusions are two phase and at room temperature consist of a small vapor bubble (10-20% by volume) and a low-salinity liquid. No evidence of a separate gas phase or of boiling has been observed in the inclusions. Homogenization temperatures of fluid inclusions are nearly identical to the present equilibrated borehole temperatures, and range from 150°C at 326 feet to 200°C at 826 feet. Homogenization temperatures near the base of the well average 165°C. The salinities of these inclusions average 4500 ppm equivalent NaCl (59 measurements).

Calcite has been isotopically analyzed from two depths in 64-16. By combining the homogenization measurements with the isotopic fractionation factors for calcite-water, the O-18 content of the hydrothermal fluids can be determined. The results indicate that the fluids are 7-8 per mil heavier than the local meteoric water and are similar to the reservoir fluids. Additional isotopic analyses are in progress. Because the isotopic shifts in the fluids are a function of the water-rock ratios and hence permeability, it may be possible to map variations in permeability across the field using calculations of this type.

MEAGER CREEK, B. C.

Drilling for geothermal fluids at Meager Mountain, in southwestern Canada, has provided an opportunity to study hydrothermal processes and fluid flow beneath an active strato-volcano. Drill holes have encountered temperatures as high as 264°C in altered crystalline basement rocks that act as the geothermal reservoir. Petrographic, mineralogic and trace-element studies have been used to establish the paragenetic relationships among the several thermal events that have affected these rocks. These relationships indicate that fault and fracture zones, steeply dipping dikes, and hydrothermal breccias related to recent volcanic activity have focused the upward movement of the geothermal fluids.

Four chemically distinct groups of thermal fluids occur at Meager Mountain. Three are NaCl in character and are associated with a well-defined thermal anomaly on the southern flank of the volcano. The fourth group consists of $NaHCO_3(SO_4)$ fluids that represent steam-heated groundwaters. The fluids range from low-temperature and very saline with moderate isotope shifts to high-temperature and moderately saline with large isotope shifts.

The chemical and isotopic compositions of the NaCl waters show that little mixing of the different fluid types has occurred. In contrast, extensive fluid mixing is a common feature of highly productive geothermal systems in other volcanic terrains. In these productive systems, mixing is generally believed to result from convectively driven fluid flow in rocks with high permeabilities. We suggest that the lack of mixing at Meager Mountain reflects fluid flow through a few discrete fracture zones in low permeability rocks and that fluid movement is driven by topographically-controlled head differences. Calculated water/rock weight ratios, based on oxygen and deuterium isotopic shifts of the reservoir fluid, range from 0.005 at 123°C to 0.022 at 126°C.

SALTON SEA

The Salton Sea geothermal field is located in an active rift zone where greenschist-facies metamorphism is currently taking place. The field is capped by low-permeability rocks that control the distribution of fluid and heat flow to a depth of several hundred meters. Chemical, petrographic, and fluid-inclusion data from two high-temperature wells show that the composition of the brines, fluid flow patterns, and thermal characteristics of the caprock changed as the geothermal system evolved. The caprock in these wells consists of two layers. The upper 250 m is composed of impermeable lacustrine claystone and evaporite deposits. The lower layer consists of deltaic sandstones. During the initial development of the geothermal system, downward percolating waters deposited anhydrite in the sandstones, reducing their permeabilities. Homogenization temperatures of fluid inclusions in anhydrite define a conductive gradient through the caprock. Temperatures reached 245° C near its base at a depth 335 m. The salinities of the brines ranged from 7 to 24 equivalent weight percent NaCl.

Subsequent incursion of high-temperature brines into the sandstones resulted in potassic alteration, deposition of base metal sulfides, and dissolution of anhydrite and calcite. The final stage in the evolution of the caprock records the initiation of fracture permeability. During this stage, veins containing quartz, barite, and base metal sulfides formed at temperatures ranging from 180°C to 240°C.

SURFACE AND BOREHOLE ELECTRICAL MODELING

Although none of the geophysical methods maps permeability directly, any geological, geochemical, or hydrological understanding of the factors that control the permeability in a geothermal reservoir can be used to help determine geophysical methods potentially useful for detecting the boundaries and more permeable parts of a hydrothermal system. At UURI, we have been developing electrical borehole techniques to detect and map permeable zones in the subsurface, especially fractures.

It is important to understand the differences between geophysical well logging and borehole geophysics. In geophysical well logging, the instruments are deployed in a single well in a tool or sonde, and the depth of investigation is usually limited to the first few meters from the well-bore. Logs such as the gamma-ray, acoustic induction and borehole televiewer logs are commonly applied in geothermal work. By contrast, borehole geophysics refers to those geophysical techniques where energy sources and sensors are deployed (1) at wide spacing in a single borehole, (2) particularly in one borehole and partly on the surface, or (3) partly in one borehole and partly in a second borehole. Thus, we speak of borehole-to-surface, surface-to-borehole and borehole-to-borehole surveys. The depth of investigation is generally much greater in borehole geophysical surveys than it is in geophysical well logging.

The borehole electrical techniques are in general poorly developed. There are a large number of ways in which borehole electrical surveys can be performed and it has been unclear which methods are best. At the same time, computer algorithms to model the several methods have not existed so that it has not been possible to select among methods prior to committing to the expense of building a field system and obtaining test data.

The objective of our program is to develop and demonstrate the use of borehole electrical techniques in geothermal exploration, reservoir delineation and reservoir exploitation. Our approach is:

1. Develop computer techniques to model the possible borehole electrical survey systems;
2. Design and construct a field data acquisition system based on the results of (1);
3. Acquire field data at sites where the nature and extent of permeability are known; and,
4. Develop techniques to interpret field data.

To the present time, we have made considerable progress on item (1) above and we are now at a point where item (2) could be started. We are working on item (4) concurrently with (1) since they are closely related.

To date, we have not had enough funding to build an appropriate field survey system. However, we have recently negotiated an agreement with Utah International, a large minerals mining company, for the use of their well logging equipment. With minor modification, this equipment can be used to test the borehole electrical methods we have been developing. Utah International will also provide technician and other in-hand support to the project. This represents a major breakthrough for us, and we are grateful to them.

Currently in our theoretical work, the finite element method is being applied to the development of an algorithm capable of modeling (in the forward sense) the electrical response of a two-dimensional (2-D) earth excited by a three-dimensional (3-D) point source. So that 2-D formulation can be applied, the problem is solved in the Fourier-transform domain using a source with the strike direction transformed out. The solution obtained using this source is then inverse Fourier transformed to obtain the solution for the 3-D source. Finding an efficient and accurate method of performing the inverse transform is the task presently at hand. Ultimately, the program will enable both surface and borehole modeling of complex 2-D earth structures for multiple electrical arrays.

SUMMARY

The Reservoir Definition Program at UURI approaches the problem of fracture definition in complex geologic environments through a multi-faceted approach. This approach emphasizes prediction of the formation of permeability by analysis of stress fields. It emphasizes processes along fractures which will either maintain permeability or destroy it through alteration processes. And it emphasizes delineation of fractures through geological, geochemical and geophysical modeling.

CALDERA RESERVOIR INVESTIGATIONS PROGRAM

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ABSTRACT

Caldera environments are young volcanic environments in which are often found the type of high-silica volcanic rocks that are believed to indicate a large magma chamber in the subsurface. Such a magma chamber would provide a heat source for geothermal systems. Thus, caldera environments are fruitful places to look for geothermal energy. From the geoscientific viewpoint, there are a great many questions remaining to be answered about caldera environments. This is especially true in evaluating the geothermal potential in particular volcanic areas, in locating geothermal systems in these areas, and in siting wells to intersect production zones. The objective of the Caldera Reservoir Investigations Program is to develop analytical and interpretive tools for industry to use in locating and evaluating geothermal reservoirs within young volcanic regions.

During the past two years, the program has concentrated on the Cascades region of the northwestern U.S. DOE has been performing cost-shared research with industry consisting of coring in specifically chosen areas and in obtaining geophysical well logs down hole as well as physical and chemical properties of the core. These data are being compared to surface geological, geochemical and geophysical data for the purpose of developing and verifying new analytical tools and testing existing tools. Results to date indicate that better tools are needed for use in conjunction with surface electrical geophysical surveys because some of the low-resistivity zones found from surface surveys correlate with low-temperature hydrothermal alteration rather than selectively pinpointing high-temperature positions of geothermal systems. A second important result is the measurement at three sites of the depth to which cold surface water circulates, which is the minimum depth that industry must drill to obtain reliable heat-flow measurements.

BACKGROUND

General Considerations

The heat source for most high-temperature geothermal systems is a body of molten or recently cooled rock in the subsurface which has been injected from great depth into the upper crust of the earth. During such intrusion processes, it is common for some of the molten magma to make its way to the surface and vent as volcanos in the form of flows, airfall tuffs and other types of deposits. Thus, areas containing young volcanic rocks (less than about 1 million years old) are generally favorable for the occurrence of geothermal systems.

Some volcanic areas contain only basaltic magma, a low-silica magma that is low in viscosity and can therefore flow from great depth in narrow fracture zones. Such basaltic areas do not necessarily indicate the existence of a magma chamber close enough to the surface to form a geothermal resource. Other volcanic areas contain felsic volcanic rocks which are higher in silica content than basalts and which are very viscous. Felsic magmas can not flow from depth into the crust through narrow fractures because of their high viscosity but tend to move upward as fairly large bodies through magmatic stoping or forceful injection. The existence of felsic rocks in volcanic deposits on the surface is, thus, an indication (but not proof) of a large magma body in the subsurface at depths between about 2 and 10 km. Sometimes a felsic magma body will come near enough to the surface to degas precipitously, resulting in an explosive eruption of a large volume of material. The May 18, 1980 eruption of Mt. St. Helens was a small-scale example of such an occurrence. Subsequent collapse of the surface into the volume previously occupied by the magma body may occur, resulting in a nominally circular depression known as a caldera. Calderas are taken to indicate that a large silicic magma body existed at depth, that some of the magma may still be in place, and that a great deal of heat has been brought into the shallow subsurface along with the magma body. Thus, caldera environments are prime areas for the occurrence of geothermal resources.

Calderas and volcanos are rather large geologic features. The primary problem in locating geothermal reservoirs associated with these features is in finding zones of open permeability near enough to the magma chamber that circulating water can be heated to temperatures above 150 deg C. Such zones of hydrothermal circulation are usually quite restricted in size compared to the volcanic features with which they are associated. Because of the high cost of drilling, industry can not afford to use the drill rig indiscriminantly as an exploration tool. Sites for exploration drilling must be carefully chosen to maximize chances for success. Thus, the techniques of geology, geochemistry and geophysics are used to help select the best test drilling locations.

Each geologic environment has its own set of exploration problems, and techniques that work well in some environments do not work well in others. The geologic processes of formation and evolution of geothermal systems in the caldera environment are not well understood at the present time. It is difficult for industry to predict which of the many exploration techniques should be used to find geothermal systems in caldera environments.

The Cascades Region

For the past two years, DOE has been performing cost-shared research under the Caldera Reservoir Investigations program with geothermal developers in the Cascade region of the northwestern U.S. The Cascade range is being formed by a chain of active volcanos that stretches from Lassen Peak in northern California, through the great volcanos of Oregon and Washington to Mt. Meager in western British Columbia. Nearly two dozen active volcanos attest to the large amount of heat being transported into the shallow crust in this area. Yet, in spite of the many obvious heat sources, high-temperature hydrothermal systems have been found at only four sites in the Cascades -- Lassen Peak and Medicine Lake, California; Newberry Caldera, Oregon; and, Mt. Garibaldi, British Columbia (see Figure 1). The discovered, high-temperature systems at Lassen and Newberry are not candidates for development because of the environmental sensitivity of these areas.

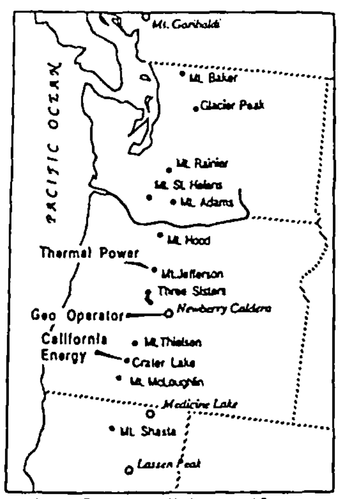

FIGURE 1
Volcanos of the Cascades

One of the reasons for the small number of discovered resources is believed to be the downward flow of a large volume of cold water from rain and snow melt. Downward movement of meteoric water apparently dominates the hydrologic regime to unknown depth. At some depth, the waters begin to move laterally, sweeping away the thermal effects of underlying hydrothermal systems. The occurrence of surface geothermal manifestations is one of the primary tools for exploration, and in the absence of such manifestations, exploration is blind. In addition, heat flow observations, another of the primary exploration tools used by industry, are difficult and expensive because a hole must be provided which penetrates the zone of downward meteoric water flow and intersects a conductive thermal gradient representative of thermal conditions at depth. Because the near-surface hydrologic conditions predominate to unknown depth, perhaps of the order of 1 km, heat-flow holes are expensive to drill. Therefore, not many representative heat flow measurements exist for the Cascades. Thus, the industry is forced to rely on such indirect methods for detection of geothermal systems as geophysics and geologic mapping. Improvements are needed in tools and techniques for locating geothermal systems in the Cascades and other caldera environments, and industry supports the DOE research efforts to make these improvements.

OBJECTIVE

The objective of the Caldera Reservoir Investigations Program is to develop analytical and interpretive tools for industry to use in locating and evaluating geothermal reservoirs within young volcanic regions. To date, the Program has concentrated its research efforts mainly in the Cascades region.

APPROACH

DOE's approach to research in the Cascades region has been to cost-share with industry the acquisition of core samples and geophysical well logs from coreholes sited within specific geologic environments. The core samples are subjected to mineralogical and chemical analysis, with emphasis on determining the nature of fractures and the nature and genesis of any hydrothermal alteration minerals. Such studies help to determine the structural and chemical controls on permeability within and adjacent to geothermal reservoirs. Geophysical measurements are also made on the cores for comparison with the geophysical borehole logs. Both the core geophysical measurements and the borehole logs are compared with surface geophysical observations. This type of work helps to determine exactly what conditions in the subsurface cause the surface geophysical responses that are observed, and serves to calibrate and evaluate various surface and borehole geophysical tools.

To date, three coreholes have been completed and a fourth has been temporarily suspended due to winter snows. Table 1 shows data on the research holes and Figure 1 shows their general locations. The two GEO holes were cored and logged in the Newberry caldera area with GEO Operator Corporation, a subsidiary of Geothermal Resources International, as DOE's partner. The hole designated as Thermal CTGH-1 was cored with Thermal Power Company as DOE's partner. Thermal has performed as the operator on a joint Thermal-Chevron agreement for this work. This hole was drilled in the Clackamas area on the north slopes of Mt. Jefferson, near Breitenbush Hot Springs. The hole shown as CECI MZI-11A is currently suspended for the winter, but it is being cored with California Energy Company, Inc. as DOE's partner in the Mazama area, on the southeast slopes of the Crater Lake caldera. Table 2 shows the geophysical well logs that have been obtained in these holes to date.

CORING SUMMARY

	GEO N-1	GEO N-2	THERMAL CTGH-1	CECI MZI-11A
Spud date	6/24/85	6/7/86	6/7/86	8/12/86
Completion Date	10/20/85	8/1/86	9/2/86	(drilling temporarily stopped)
Drilling Contractor	Tonto	Tonto	Boyles	Longyear
Core Recovery	>90%	>90%	>90%	....
Coring Rate	69ft./day	67ft./day	66ft./day	$61/ft.
Total Depth	4550 ± ft.	4002 ft.	4800 ft.	(1354 ft.)
Public Domain Data	0-4000 ft.	0-4002 ft.	0-4800 ft.	....
Completion	1 1/2" iron pipe to T.D.	1 1/2" iron pipe to T.D.	HX rods to 4203 ft. open to T.D.	....

TABLE 1

GEOPHYSICAL WELL LOGS

	GEO N-1	GEO N-2	THERMAL CTGH-1	CECI MZI-11A
Temperature	X	X	X	0-1329'
Caliper	X	X	4100'-4800'	-
Gamma Ray	X	X	X	-
Neutron	-	-	X	-
Gamma-Gamma Density	-	-	775'-900'	-
Spontaneous Potential	X	-	4200'-4798'	-
Resistivity	X	-	4200'-4799'	-
Induced Polarization	-	-	4200'-4799'	-
Laterolog	-	-	4200'-4798'	-
Induction	X	-	-	-
Acoustic	X	X	4225'-4425'	-
Acoustic Fraclog	X	X	-	-

TABLE 2

The research team comprises scientists from government agencies, federal contractors and the three companies so far involved. Program management and technical direction is provided by the Idaho Operations Office as well as by Headquarters. The companies are performing their own analyses of the data, and in some cases have published the results. The federally funded researchers have formed a good working relationship with the industry participants in determining the topics and priorities for research. The University of Utah Research Institute is responsible for taking DOE's share of the core samples and data generated by the project and placing it in the public domain. UURI also performs lithologic logging of the core, hydrothermal alteration studies, analysis of the geophysical well logs and physical and chemical measurements on the core samples. The Oregon Department of Geology and Mineral Industries (DOGAMI) is mapping the geology in the vicinity of some of the coreholes. Southern Methodist University performs precise temperature logging of the holes and laboratory thermal-conductivity measurements on the core for the purpose of determining the heat flow. The U.S. Geological Survey performs detailed analysis of the hydrothermal alteration minerals and has been analyzing fluid geochemistry in the area. The University of Wyoming is performing lithologic and major-element geochemical work on the cores from the Newberry area using funding from the National Science Foundation.

RESULTS

In this paper, I will discuss two examples of the research results to date.

Electrical Resistivity Anomalies

Surveys designed to measure the electrical resistivity of rock at depth are common tools in geothermal exploration. Rock resistivity is lowered by saturation with the high-temperature, commonly briny fluids in hydrothermal systems, yielding a contrast with rocks of higher resistivity exterior to the system. Surface electrical surveys at Newberry have been described by Bisdorf (1983), Fitterman (1983) and Fitterman et al. (1985) of the U.S. Geological Survey. Figure 2 shows the results of a time-domain electromagnetic survey reported in the latter two references. The interpretation shows an upper layer about 800 m thick and of resistivity 200 to 500 ohm-m overlying a conductive horizon whose parameters are shown on Figure 2. These data indicate that electrically conductive horizons, having a resistivity about one order of magnitude lower than the layer above, underlie a substantial portion of the Newberry area, continuing well outside the caldera. The cause of the portions of this geophysical anomaly which lie outside the caldera was unknown prior to the research coring at GEO-Newberry N-1. However, it was known that a research hole drilled by the U.S. Geological Survey in the caldera, at the location shown for Newberry-2 on Figure 2, encountered a high-temperature hydrothermal system and constituted a discovery (Sammel, 1981). Presumably the low-resistivity anomaly inside the caldera is due to the high-temperature hydrothermal system and its associated alteration.

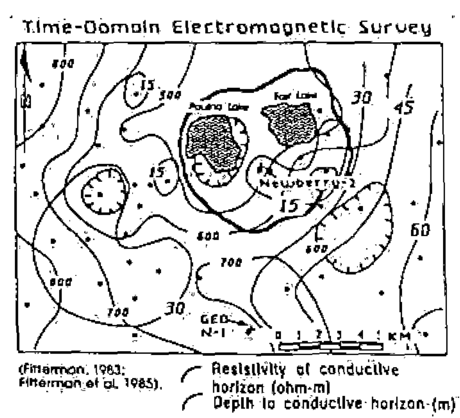

FIGURE 2
Newberry Caldera Electrical Geophysical Survey Results.

Geophysical well logs were run in GEO N-1, and a portion of these logs is shown in Figure 3. This figure shows a selected portion of the electrical induction log along with a simplified lithologic log and data from a separate temperature survey. The parameter of the induction log is electrical conductivity in siemens per meter, which is the reciprocal of resistivity in ohm-m. Deflections to the left on this log indicate higher conductivity, i.e. lower resistivity. Conductive zones can be seen near 2830 ft, 2890 ft, 3110 ft, 3350 ft, 3430 ft, 3470 ft, 3490 ft, 3600 ft, 3670 ft, 3710 ft, 3830 ft and 3880 ft. These conductive horizons correlate with altered volcanic ash and tuffaceous units.

Several of the altered zones which exhibit low resistivity on the geophysical well logs were chosen for mineralogical study (Wright and Nielson, 1986). It was found that the chief alteration mineral is calcium smectite, a clay mineral. Hydrothermal alteration produced the smectite from volcanic glass in the ash and tuff units, and the chemical reactions responsible are known to proceed at temperatures about 100 deg C and possible somewhat lower. The formation of clay minerals in the rock lowers the resistivity because clays can carry electrical current as loosely held ions migrate along their surfaces. We have concluded, that the cause of the surface electrical geophysical anomaly, in the N-2 area, is the low-temperature hydrothermal alteration observed in certain volcanic units below about 2800 ft. This result indicates that surface resistivity surveys must be carefully interpreted in order to avoid confusion between areas of low-temperature alteration and high-temperature alteration, and that other screening tools are needed by industry to avoid drilling low-temperature zones.

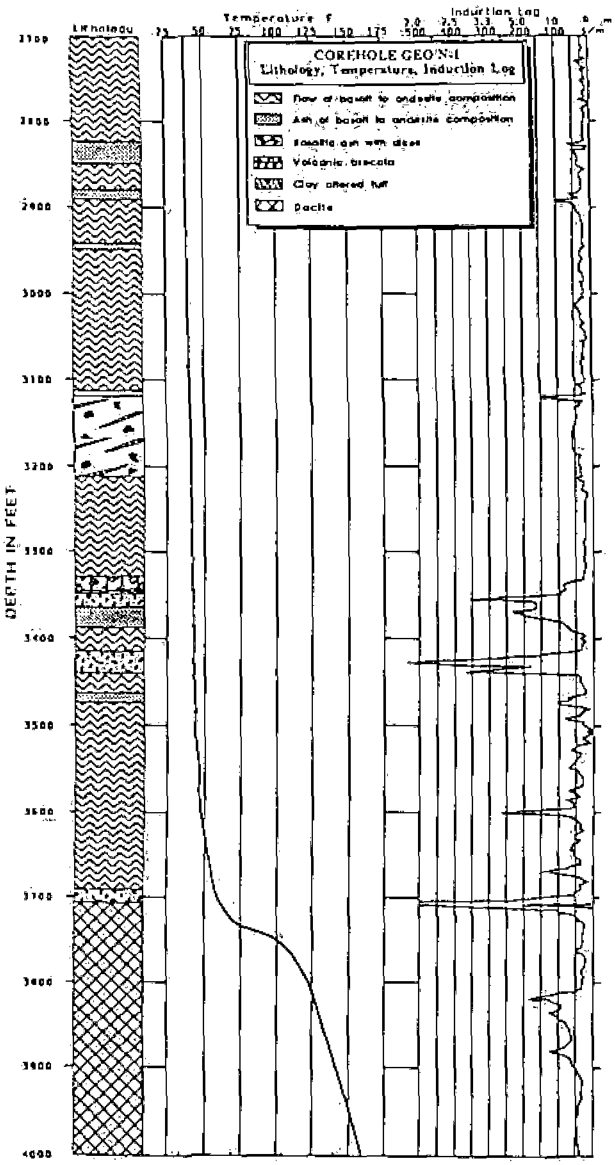

FIGURE 3

Research hole CTGH-1 was cored on the north slopes of Mt. Jefferson, an active volcano in Oregon about 50 miles north of Newberry. The industry participant in this research is Thermal Power Company. The corehole was sited to intersect a subsurface area of low resistivity as indicated on a telluric-magnetotelluric (T/MT) survey done previously in the area by Chevron Geothermal. The corehole found a section of successive volcanic flows of predominately basaltic and basaltic-andesite composition to the total depth of 1463 m (4800 ft). Electrical geophysical logs were run in the uncased portion of the hole, below about 1280 m (4200 ft). The results of this logging are given in Figure 4, which shows the 16-in normal and 64-in normal resistivity logs. The predominate resistivity observed is about 5 ohm-m, but there are occasional increases in resistivity to more than 75

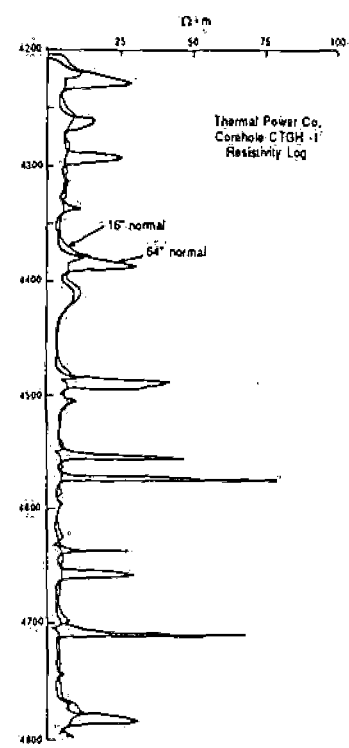

FIGURE 4

ohm-m. The logs probably do not represent the true value of the high resistivities because the high-resistivity zones are so narrow. Minealogical studies and laboratory measurements of resistivity were undertaken on core samples from both the high-resistivity portions and the low-resistivity portions. The results are shown on Table 3. We see that typical laboratory resistivity values are 11 to 16 ohm-m for samples that correspond to logged resistivities of 5 ohm-m. The laboratory measurements were made at room temperature whereas the downhole temperature was measured at 80 to 95 deg C for this portion of the hole. Higher in-situ temperature would lower the resistivity by a factor of 0.3 to 0.4 for the downhole temperatures observed. We conclude that the laboratory measurements on core samples are consonant with the geophysical well logs. The lower half of Table 3 shows the results of mineralogical analyses on two core samples. Here again we found that the sample having lower resistivity contained an appreciable amount of smectite, a low-temperature hydrothermal alteration mineral. We have tentatively concluded that, in the Mt. Jefferson area also, low-resistivity anomalies found by surface resistivity surveys do not necessarily indicate the presence of high-temperature geothermal systems at depth.

THERMAL POWER CTGH-1

SUMMARY OF CORE MEASUREMENTS

Sample Depth (ft.)	D (cm)	L (cm)	Porosity (%)	ρ_t (Ωm) in ρ_w = 1Ωm	ρ_t (Ωm) in ρ_w = 5Ωm	ρ_t (Ωm) extrapolated to ρ_w = 10Ωm
4228	4.7	12.3	0.5 (?)	245	236 (?)	240
4450	4.7	11.9	10.8	11.5	15.0	17
4625	4.7	8.9	14.8	16.1	26.4	36
4733	4.7	12.3	16.3	12.6	16.5	22

ρ_w = measured resistivity of saturating solution $\rho_t = a \rho_w \phi^{-bm}$ (Archie's Law)

X-RAY DIFFRACTION

Sample Depth (ft.)	quartz	plag	clino-pyrox	mordenite	smectite	celadonite	amorph
4226	1	41	10	0	0	0	46
4733	2	47	10	3	13	Tr	25

TABLE 3

More research work needs to be done on the comparison between resistivity anomalies caused by low-temperature alteration and those caused by high-temperature alteration that seem to occur in several areas of the Cascades. Surface electrical surveys have formed an important part of industry's exploration strategy in other areas, and we need to learn to use this valuable tool in the caldera environment, too. We plan to study methods of distinguishing these two anomaly sources. One possible approach to this problem would be to use the induced polarization method to try to distinguish the different alteration mineral suites that are associated with high-temperature and low-temperature hydrothermal alteration.

Near-Surface Hydrologic Regime

We stated above that the porous volcanic rocks of the Cascade range allow the deep infiltration of rain and snow-melt water. This downward movement of cold water suppresses surface manifestations of underlying geothermal systems and causes heat-flow measurements to be unrepresentative of the deeper thermal regime. One of the objectives of the research coring is to obtain data on temperature vs depth that indicate the depths one would typically have to penetrate to in order to get beneath this near-surface, cold hydrologic regime. Figure 5 shows the temperature data logged by David Blackwell of Southern Methodist University for the two Newberry coreholes and the Mt. Jefferson corehole.

GEO-Newberry N-1, shows a nearly constant temperature from the surface to a depth of about 1000 m. Below this depth, the temperature increases very abruptly from about 8 deg C to

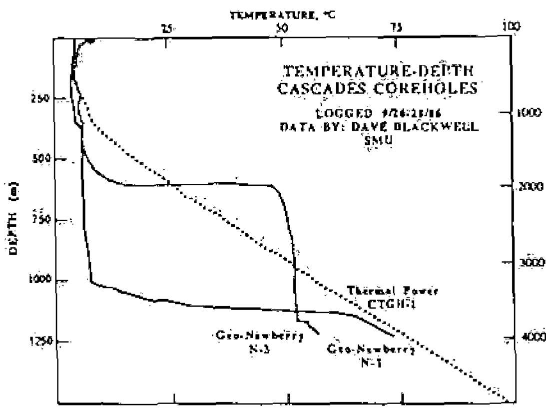

FIGURE 5

about 65 deg C at 1150 m. The gradient then decreases to an average of 100 deg C per km to a depth of 1220 m. This gradient in the lower part of the hole is apparently a conductive gradient and represents upward flow of heat from some unknown depth. It is an anomalously high gradient since the average for the earth in the West is perhaps 35 deg C per km. The thermal profile for N-1 shows that the zone of downward percolation of cold water extends to 1000 m, and it is apparent that heat-flow holes would have to be drilled below this depth in any exploration program by industry in this area.

GEO-Newberry N-3 shows a somewhat different temperature profile. Temperatures are low and nearly constant from the surface to about 600 m. Then the temperature increases abruptly to 47 deg C at 625 m. Below this, the temperature is nearly isothermal at about 50 deg C to a depth of 1164 m, where it begins to increase again. The very bottom part of the hole shows an average gradient of about 70 deg C per km, clearly anomalous but lower than the bottom-hole gradient observed in N-1. In N-3, the zone of near-surface cold water flow is believed to extend to a depth of 625 m, and the isothermal section from 625 to 1164 m is believed to be due to warm water entering the hole in the lower parts of this interval, flowing upward and then exiting in the upper parts of the interval. The bottom-hole gradient may be typical of the deeper thermal regime in the area. Here we see that in order to obtain reliable heat-flow measurements, one would have to drill below 1220 m (4000 ft).

The Mt. Jefferson hole of Thermal Power Co. contrasts sharply with the Newberry holes. At CTGH-1, the temperature is low and slowly increasing from the surface to a depth of about 350 m. Below 350 m, a nearly constant thermal gradient is observed, indicating a conductive regime. The gradient has an average value in the lower part of the hole of 80 deg C per km. If this gradient persists to depth below the bottom of the hole, a temperature of 200 deg C would be encountered at 2600 m (8500 ft). The zone of near-surface cold water downflow extends only to 350 m. Presumably, the use of heat-flow studies in exploration would be much less expensive for industry to carry out in the Mt. Jefferson area than in the Newberry area.

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APPLICATION OF GEOPHYSICS TO EXPLORATION FOR CONCEALED HYDROTHERMAL SYSTEMS IN VOLCANIC TERRAINS

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ABSTRACT

Exploration for concealed geothermal systems in volcanic terrains will require well planned and executed programs to succeed and be cost-effective. The geologic record indicates that large-scale hydrothermal convection systems occur only sporadically around plutons, and so a great deal more than identifying heat sources will be needed. Geophysical surveys can contribute to integrated exploration programs if used properly. This paper discusses some potential applications of geophysics and how it might be integrated into an exploration program.

INTRODUCTION

Exploration for completely concealed geothermal resources has been undertaken to a much smaller extent than has exploration for resources with direct surface manifestation. By concealed resources, we mean those that lack such direct manifestations as hot springs, geysers, fumaroles, mud pots, hydrothermally altered areas or sinter. Many volcanic terrains seem to be characterized by abundant potential heat sources, as indicated by active volcanism, but by a comparative lack of geothermal surface manifestations. This is true for the Cascades province of the western U.S., as has been pointed out by several authors (Brook et al., 1979; Priest, 1983). It is also common in other volcanic terrains. In Ecuador, for example, the Andes mountains are composed of young volcanic features including about 30 active or recently active volcanos. Yet there are very few surface manifestations indicative of large-scale, high-temperature convection systems (Instituto Nacional de Energía de Ecuador, pers. comm.). For the Cascade range and for other volcanic areas such as Japan, it has been argued that high precipitation produces downward and laterally migrating cold water that suppresses primary surface manifestations (e.g., Oki and Hirano, 1974; Priest, 1983). It is also possible that there is a relative lack of high-level magma chambers associated with some andesitic to basaltic volcanic terrains. Lack of magma chambers, of course, implies lack of large heat sources at shallow depth (5-10 km) to power large, high-temperature convection systems. Clearly, it will be necessary to understand these volcanic areas better before reliable conclusions about resource potential can be made.

In considering the relationships between magma chambers and hydrothermal systems, the geologic record from many mining districts tells us that hydrothermal systems typically occur only sporadically around plutons. For example, in the Bingham district, Utah, there are extensive outcrops of unaltered, unmineralized Last Chance stock and its contact with Paleozoic and Mesozoic sedimentary rocks, but only the Utah Copper Stock, a later intrusion, produced a large hydrothermal system (Peters et al., 1966). In the Valles caldera, New Mexico, the known hydrothermal system is associated with the resurgent Redondo Dome, but deep drill holes at nearby Fenton Hill show that no hydrothermal system exists at this location, although temperatures are certainly hot enough to support convection (e.g., Smith et al., 1983). Hydrothermal convection systems can be expected to form around a pluton only if sufficient permeability exists or is developed (Norton, 1984). Some intrusions or some parts of intrusions produce or are otherwise associated with the fracturing needed for convection, whereas others are not. In exploration for concealed hydrothermal resources, we can expect that once a heat source is located, the search for an associated hydrothermal system will have only begun.

Economic discovery of hydrothermal systems, if they exist around known volcanos, will not be an easy task. It is clearly of interest to devise cost-effective strategies for their discovery. While some would prefer to base exploration solely on geology (La Fleur, 1983), we believe that the best approach will be a carefully selected and integrated mix of geological, geochemical, geophysical and hydrological techniques. In this paper, we focus on the potential applications of geophysics in such an integrated program.

CONSIDERATION OF GEOPHYSICAL METHODS

In this section, we discuss some of the problems and promises of commonly used geophysical methods in volcanic terrains.

Thermal Methods

Thermal gradient and heat flow surveys provide basic data about subsurface temperatures, and some program of shallow and deep thermal-gradient holes is applied in most systematic geothermal exploration programs throughout the world. The interpretation of temperature, thermal gradient, and heat flow data and the evaluation of resource potential from these measurements can be quite complex, as discussed by numerous authors (e.g., Sass et al., 1981; Rybach and Muffler, 1981). Drill holes must be deep enough to penetrate the near-surface hydrologic regime, which may be dominated by meteoric recharge and lateral flow of cold water. In the Cascades, this zone may exceed 1 km in thickness. The limitations on the use of thermal methods are generally imposed by the cost of the drilling program.

Because it would be unwise to proceed to a deep production test in the absence of a known temperature anomaly, thermal gradient drilling must provide encouragement where surface manifestations are lacking. If the holes must be 3000 feet deep or more, they will be expensive and therefore limited in number. The maximum amount of pertinent information must be brought to bear on siting thermal gradient holes, consistent with limited exploration resources.

Electrical Methods

Geothermal reservoirs frequently exhibit low resistivities due to high temperature, enhanced porosity, salinity of the interstitial fluid, and alteration of silicate minerals to clays (Moskowitz and Norton, 1977; Ward and Sill, 1984). Thermal brine and alteration may occur predominantly along faults, so these methods may map faults controlling a fractured reservoir. Alternatively, they may map a stratigraphic unit that contains thermal brines and/or alteration. The electrical methods can also map faults, stratigraphy, intrusions, and geologic structure in general, independent of the presence of brine or alteration.

Galvanic Resistivity. This technique can be very useful if significant hydrothermal effects occur no deeper than about 2000 feet. For deeper occurrences, the tradeoffs between depth penetration, loss of resolution of anomalies the size of many geothermal systems and difficulty in performing the survey make the technique of questionable utility. In addition, volcanic areas often have high electrode contact resistance, causing low transmitted current, and precluding deep exploration.

AMT/CSAMT. Most reported AMT surveys are scalar AMT, that is, only one component of electric field and one of magnetic field are measured at once (Hoover et al., 1978). It can be demonstrated that in layered terrains this scheme is adequate for obtaining resistivity structure, but if resistivity also varies in either or both horizontal dimensions, as it invariably will in volcanic areas, scalar AMT is inadequate and is not recommended for exploration for concealed resources in volcanic terrains. For this task, a tensor measurement is needed.

MT. Magnetotelluric instrumentation incorporates the capability to make a tensor measurement, that is, to measure simultaneously both orthogonal electric field components (E_x, E_y) and all three orthogonal magnetic field components (H_x, H_y, H_z). This method is generally considered to be capable of exploration to depths of tens of kilometers, and to be capable of detecting magmas directly. Neither of these attributes is true in all cases. Although magma is conductive due to mineral semiconduction, the amount of contained water substantially affects the conductivity, dry magmas being much less conductive than wet ones (Lebedev and Khitarov, 1964). In geothermal exploration, it is possibly the wet magmas that we seek, however, because they have enough volatile content to produce the fracturing needed for hydrothermal convection. Depth of exploration depends to a certain extent on the near-surface resistivity structure. Also of great importance is the size and other characteristics of the magma body. Newman et al. (in press) have explored conditions under which crustal magma bodies can be detected. They conclude that if the body is isolated, i.e. has broken off from conductive magma at depth, it is more easily detected than if it maintains connective roots to the mantle.

The MT method has been used a great deal in geothermal exploration with generally disappointing results (Ward, 1983). By far the biggest problems appear to be misapplication and inadequate interpretation. Most MT data have been interpreted using one-dimensional inversion to a layered-earth resistivity structure. This method is totally inadequate in most geothermal exploration and usually produces misleading results. Full three-dimensional modeling is needed. The MT method has many subtleties, and must be applied with a great deal of care by geophysicists who are well experienced.

CSEM. Controlled-source electromagnetic methods have been used as alternatives to galvanic resistivity or AMT surveying (Keller and Rapolla, 1974; Keller et al., 1982). A high-powered CSEM system has been developed and reported by workers at Lawrence Berkeley Laboratory (Wilt et al., 1981). The primary limitation of these techniques to date has been that interpretation methods have been limited to the one-dimensional case. Two- and three-dimensional algorithms are now becoming available, but further development is needed.

SP. Spontaneous-potential anomalies over convective hydrothermal systems arise from the electrokinetic and thermoelectric effects, which couple the generation of natural voltages with the flow of fluids and the flow of heat, respectively (Sill, 1983). SP surveys have been used successfully in certain volcanic terrains. On Hawaii, Zablocki (1976) found a large SP effect over the East Rift zone. Although these surveys are relatively inexpensive to run, they are also difficult to interpret in terms of the nature and location of the source area.

General Discussion. There is no wholly satisfactory electrical method for exploration for concealed resources in rugged volcanic terrains. Galvanic resistivity surveys, while relatively easy to run and for which interpretation methods are reasonably well worked out, lack depth penetration. Scalar AMT, which is easy to run and for which highly portable equipment is available, does not provide enough data to resolve the subsurface resistivity structure adequately. The MT method is able to resolve complex structure better, but uses very sophisticated, marginally portable equipment and requires a highly trained crew and complex, sophisticated interpretation. The CSEM methods are relatively easy to run but equipment is only marginally portable and adequate interpretation is only now becoming available. SP surveys are easy and cheap but interpretation is difficult and ambiguous. In view of the relevance of electrical methods to geothermal exploration, development of electrical equipment and techniques specifically for the geothermal environment would seem like a wise research investment.

Gravity Method

Density contrasts among rock units permit use of the gravity method to map intrusions, faulting, deep valley fill, and geologic structure in general. Regional gravity studies may play a major role in understanding the tectonic framework of geothermal systems in volcanic environments such as the Cascade Range. Couch et al. (1981) note that a contiguous zone of gravity lows west of the High Cascades in central Oregon defines major structural trends and delineates fault zones which may localize the movement of hydrothermal fluids. Williams and Finn (1982) report that large silicic volcanos produce gravity lows when proper densities of 2.15 to 2.35 g/cm^3 are used for the Bouguer reduction. Other volcanos produce gravity highs as a result of higher-density subvolcanic intrusive complexes.

Magnetic Method

The locations of faults, fracture zones, intrusives, silicic domes and major altered areas are apparent on magnetic data from many geothermal prospects. Bacon (1981) interprets major structural trends and fault zones from aeromagnetic data in the Cascades. A magnetic low occurs over a part of the hot-spring area at Long Valley, California, and is interpreted by Kane et al. (1976) as due to destruction of magnetite by hydrothermal alteration. Magnetic data can also be used to determine the depth to the Curie isotherm (Bhattacharyya and Leu, 1975, among others). These interpretations are dependent on many assumptions and have serious limitations. It is assumed that long-wavelength negative anomalies due to lithologic changes do not significantly perturb the interpretation, and that the decreased magnetization of crustal rocks at depth is due to temperatures above the Curie point rather than to deep-seated lithologic changes. In addition, because the bottom of a magnetized prism is not accurately determined, accuracy of Curie-point depth can be poor.

Seismic Methods

Earth Noise. There is limited evidence (e.g. Liaw and Suyenaga, 1982) that hydrothermal processes can generate seismic body waves in the frequency band 1 to 10 Hz. Noise also arises in such sources as traffic, trains, rivers, canals, wind, etc. Liaw and McEvilly (1978) have demonstrated that field and interpretive techniques for earth noise surveys require a great deal of understanding and care. These surveys can provide a guide to hydrothermal processes provided that data quality is good and careful interpretation is done.

Microearthquakes. Microearthquakes frequently are closely related spatially to major geothermal systems. Accurate locations of these earthquakes can provide data on the locations of active faults that may channel hot water toward the surface. Microseismic activity is generally episodic rather than continuous, and this characteristic may provide a basic limitation to the technique in searching for or prioritizing geothermal prospect areas.

Teleseisms. If a sufficiently distant earthquake is observed with a closely spaced array of seismographs, changes in P-wave traveltime from station to station can be taken to be due to velocity variations near the array. Traveltime residuals are computed as the observed arrival time minus that calculated for a standard earth. A magma chamber beneath a geothermal system would give rise to low P-wave velocities and hence to late observed travel times (Iyer and Stewart, 1977). While one can speculate that relative P-wave delays are caused by partial melts or magmas, they can also be caused by alluvium, alteration, compositional differences, lateral variations in temperature or locally fractured rock.

Refraction. The seismic refraction and reflection methods can be used to map the depth to the water table, stratigraphy, faulting, intrusions, and geologic structure in general. Seismic refraction has been used mainly as a geophysical reconnaissance method for mapping velocity distributions and, hence, faults, fracture zones, stratigraphy, and intrusions. These data contribute to a better understanding of regional geology and are indirectly of use in geothermal exploration.

Reflection. The seismic reflection method provides better resolution of horizontal or shallow-dipping layered structures than any other method. However, where the structure becomes complicated, diffraction of seismic waves occurs and makes the task of interpreting structure difficult. At Beowawe, Nevada, extensive and varied digital processing was ineffective in eliminating the ringing due to a complex near-surface intercalated volcanic-sediment section (Swift, 1979). This problem is typical in volcanic areas. Denlinger and Kovach (1981) showed that seismic-reflection techniques applied to the steam system at Castle Rock Springs (The Geysers area) was potentially useful for detecting fracture systems with-training and experience in each of the earth science disciplines be used to form the exploration team, even if some outside expertise must be acquired through consulting. This will help reduce misapplication of techniques, faulty survey design and erronous data interpretation and will result in more cost-effective exploration.

Prospect Area Selection (Reconnaissance)

Our preferences in reconnaissance geophysical techniques for volcanic terrains include: remote sensing to help map recently active faults, contacts, and other structures and, aeromagnetic surveys to help map subsurface lithologic changes and structure. Acquistion of satellite imagery is simple though not inexpensive. Outside expertise may be needed for appropriate processing and for interpretation since the average geologist will not have the required skills. High-quality air photos are also available for many areas and should be acquired and interpreted at the same time. If regional aeromagnetic data are not available, one should strongly consider flying a survey at a data density of about one line per mile specifically for reconnaissance purposes.

If regional heat flow studies and gravity data are available, they should be acquired and interpreted, but we would not generally recommend acquisition of such data for reconnaissance purposes. None of the electrical or seismic methods are generally appropriate at the reconnaissance stage, although any available information should be used.

Prospect Ranking

Once a list of candidate prospects is made, one must assign a relative ranking to each prospect. The one geophysical method that should, in our opinion, be applied to each prospect area, is some form of electrical geophysics to rank prospects on the basis of occurrence of conductive materials at depth. The specific electrical technique should be selected on the basis of access and the exploration model for the specific area. Areas which display a resistivity anomaly may further be ranked on the basis of microearthquake or earth noise studies if desired, but we do not strongly recommended it. At this point it will usually be meaningful to drill one or more thermal gradient wells for the primary purpose of determing the shallow hydrologic regime. Magnetics and gravity usually play only a minor role in ranking of prospects.

Prospect Testing

For prospects that pass the screens discussed above, it will generally be true that additional detailed geophysical data will be needed to help site test wells. Assuming that areas with a resistivity anomaly have been ranked high for drill testing, a more detailed electrical survey will probably be needed. One should be guided by the results of previous work in the area in selecting the electrical method and designing the survey. It would be advisable at this time also to perform a detailed aeromagnetic survey of the area at a data density of about 3 lines per mile to help with structural detail, especially that which may be hidden beneath more recent volcanic rocks. It should be mentioned here that ground magnetic surveys in volcanic terrains are usually worthless because of the high geologic noise level. Flying a survey a few hundred feet above terrain performs the filtering needed to help eliminate this noise and is usually cost effective over ground surveys in any case. Gravity surveys may also help site drill holes, and should be given some consideration at this stage. As holes become available, borehole methods should also be considered. With luck, one will soon have the problem of siting a production test, and by this stage the explorationists will probably know which geoscientific methods are useful and which are not at the particular prospect.

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Thermal genesis of dissolution caves in the Black Hills, South Dakota

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ABSTRACT

Jewel Cave (118 km of mapped passages beneath an area of 2.7 km^2) and Wind Cave (70 km beneath 1.8 km^2) are, respectively, the fourth and tenth longest known cave systems and the world's foremost examples of three-dimensional, rectilinear networks of solutional passages. Other caves in the Black Hills are similar. They occur in 90–140 m of well-bedded Mississippian limestone and dolomite. Walls throughout Jewel Cave are lined with euhedral calcite spar as much as 15 cm thick. Wind Cave displays lesser encrustations and remarkable calcite boxwork. Since 1938, opinion has favored cave excavation by slowly circulating meteoric waters in artesian confinement similar to that surrounding the Black Hills.

We believe that the caves were developed by regional thermal waters focusing on paleospring outlets in outlying sandstones. Four sets of criteria are evaluated: (1) morphological—the three-dimensional, one-phase maze form having convectional features is similar to known and supposed thermal caves in Europe; (2) petrographic and mineralogical study of the chief precipitates shows a record of carbonate solution → calcite precipitation consonant with a model of cooling, then degassing, waters; (3) a thermal anomaly at regional hot springs is shown to extend beneath Wind Cave, where basal lake-water samples show chemical and isotopic affinities with the thermal waters; and (4) $\delta^{13}C$ and $\delta^{18}O$ measurements place all suspected paleothermal water precipitates in the domain of thermal calcites reported by others and being deposited at the modern hot springs. Finally, U-series dates show that the Wind Cave deposits are Quaternary and that the cave is still draining. Jewel Cave is truly relict and divorced from the modern thermal ground-water system; its great calcite spar sheets are probably older than 1.25–1.50 Ma.

INTRODUCTION

Jewel Cave and Wind Cave, in the Pahasapa Limestone of the Black Hills, South Dakota, are the world's fourth and tenth longest known caves, respectively. They are the foremost examples of three-dimensional, rectilinear cavern networks. Each displays, in great abundance, types of calcite precipitates that are rare in caves formed by direct infiltration of meteoric water. Most parts of Jewel Cave are encrusted with coatings of euhedral calcite spar that average 15 cm in thickness. Wind Cave displays a variety of lesser coatings and also remarkable wall and ceiling boxwork of composite solutional-depositional origin. Shorter caverns are known elsewhere in the Black Hills, and with few exceptions, they appear to be fragments of network complexes similar to Jewel and Wind Caves. They contain the same exotic forms of calcite.

The Black Hills caves are composed of two or three levels of solution galleries, most of which are disposed in rectilinear arrays. Passage size varies greatly, with no trend to increase downstream of junctions. The multi-storey characteristic makes them true three-dimensional mazes. All levels appear to have developed simultaneously in the same phase or sequence of phases. This is a rare phenomenon. Two-dimensional rectilinear maze caves (that is, one storey, one phase) are common, being formed where mete-

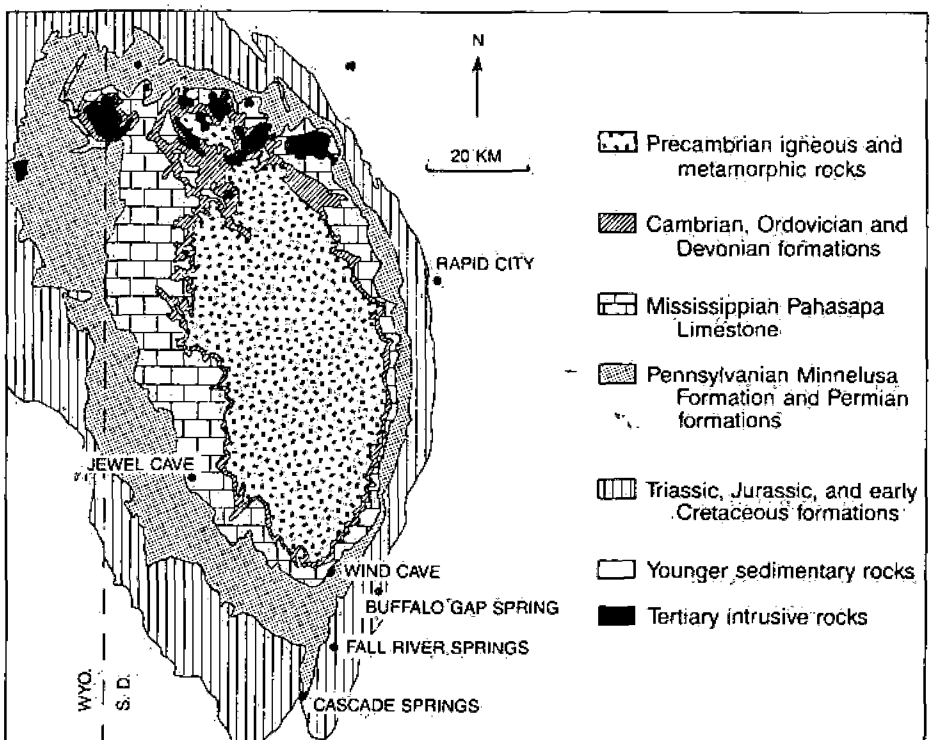

Figure 1. Geologic map of the Black Hills, showing locations of described sites.

Geological Society of America Bulletin, v. 99, p. 729–738, 8 figs., 4 tables, December 1987.

oric waters are guided into a well-jointed and soluble limestone either as uniform infiltration or as periodic flood waters. Multi-level caves with crisscrossing galleries are also common, but only in cases in which each level represents a different phase of development. Normally, these are series of passages developed at successively lower levels, displaying dendritic patterns rather than rectilinear ones; passage size tends to increase systematically downstream of junctions.

The problems posed by the Black Hills caves, therefore, are to explain the development of these multi-level but single-phase solution mazes of exceptional extent and to account for their exotic mineralization. A majority of previous authors have advocated variations of a confined or artesian flow hypothesis with meteoric waters. In this paper, we present evidence that both the dissolution and the mineralization of the caves are the product of rising thermal waters.

Geologic Setting

The Black Hills are a dissected domal structure of Laramide age (Fig. 1). The core consists of rugged mountains of Precambrian igneous and metamorphic rocks that were further intruded by igneous rocks during the early Tertiary. Around the perimeter, there are cuestas of radially dipping Paleozoic and Mesozoic sedimentary rocks, mainly sandstones and shales, breached by a few wind and water gaps. By the end of the Eocene, dissection had advanced close to the modern base levels (Palmer, 1981). Much of the landscape around the perimeter of the Black Hills was then covered by extensive terrigenous sediments of the Oligocene White River Group. There has been renewed uplift and dissection in the later Tertiary and Quaternary.

The lowest sedimentary rocks are 20–70 m of Cambrian to Mississippian sandstones, shales, and arenaceous limestones, resting unconformably on the Precambrian basement (Fig. 2). They are succeeded conformably by the Pahasapa Formation, a platform carbonate of Mississippian age. It is 90–140 m thick in the vicinity of the caves. The lower Pahasapa is massive, limy dolomite with prominent joints, favoring a simple fissure form of passage. Middle strata, in which the principal boxwork is found, include medium-bedded limestones and dolomites, locally highly fractured and brecciated, with some prominent chert beds near the top. Passages are less regular in form, with lower ceiling heights. Upper strata are massive limestones with sparse chert nodules. Passages are well rounded.

The top of the Pahasapa Limestone is a Mississippian paleokarst that has a preserved relief of ~50 m buried by Pennsylvanian sandstones. The paleokarst extends deep into the Pahasapa in the form of filled solutional clefts, sinkholes, and caves. Pennsylvanian detrital fillings vary from collapse breccia to water-laid allochthonous sediments. The modern caves primarily follow later fracture systems but ramify into the palaeokarst cavities, complicating the modern patterns. Reworked pakeokarst fill is a major component of the detrital veneers in the modern caves.

The sandstone cover (100–200 m thick) seals the paleokarst and the caves from overhead penetration by all but diffuse infiltration of meteoric water, except where recent shallow canyons have approached or intersected upper galleries.

Physiography of the Caves

Jewel Cave comprises 118 km of surveyed galleries contained within an area of no more than 2.7 km^2 (Fig. 3). It extends between 1,511 and 1,645 m above sea level in strata dipping a few degrees southwest. The cave is fully drained today, except for a few small perched pools, and is without flowing water. A nearly ubiquitous wall coating of calcite spar is an outstanding feature of this cave (Fig. 4). In the uppermost passages, this encrustation has been partly removed during at least one solutional episode. There is also some local boxwork and a few small occurrences of the travertine (stalactites, stalagmites) typical of most caves. The cave air temperature is ~8.3 °C.

Wind Cave contains 70 km of known passages beneath an area of 1.8 km^2. It extends between 1,120 and 1,265 m above sea level. Its solutional form is very similar to that of Jewel Cave, except that the average cross-sectional dimensions are smaller. Spar coatings on walls are concentrated mainly in the lowest levels and are much thinner than at Jewel Cave. There are many other unusual calcite precipitates, including horizontal fins and false floors that appear to mark growth at former pond surfaces. Normal travertine deposits are rare. The famous "boxwork" of this cave is, in scale, extent, and complexity, probably the finest that has been described. It comprises skeletal structures of vein

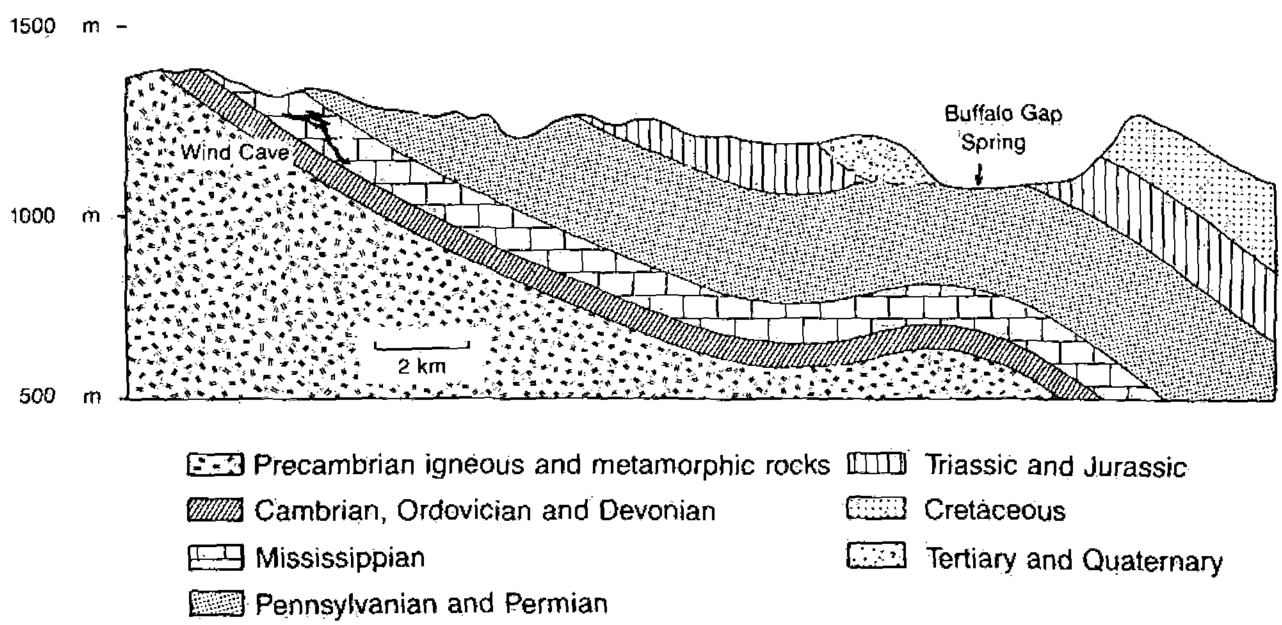

Figure 2. Cross section through the Wind Cave area in the southeastern flank of the Black Hills.

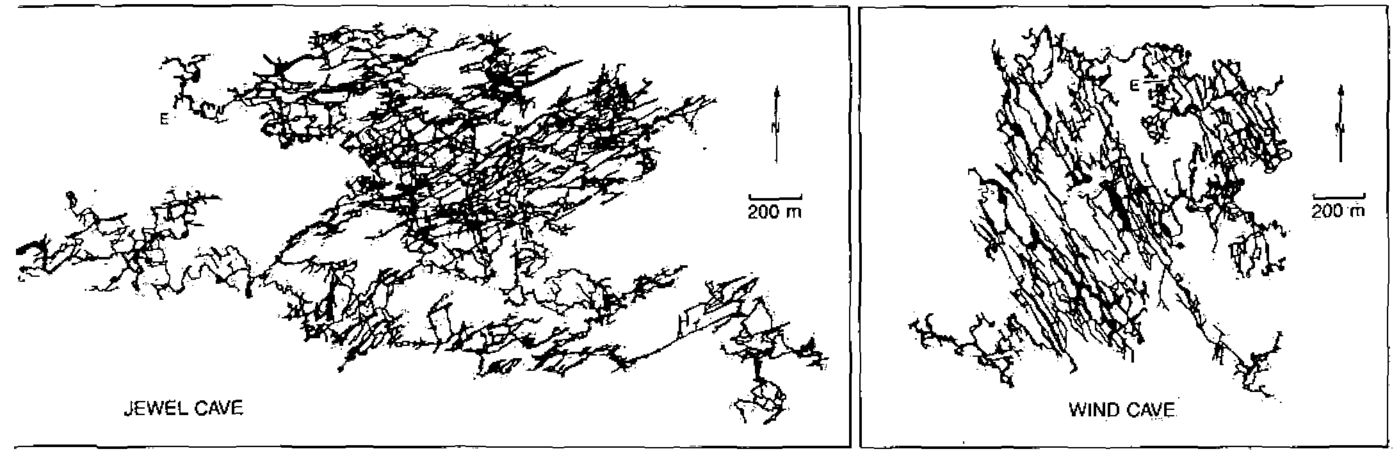

Figure 3. Maps of Jewel Cave and Wind Cave, drawn at identical scales, showing all passages mapped as of 1985. E = natural entrance.

Figure 4. Cross section through calcite wall crust in Jewel Cave, in which it has broken way naturally from underlying silty textured dolomite. Thickness of crust is 15 cm.

alcite with later calcite overgrowths (Fig. 5). 'he vein calcite has resisted dissolution and pro-'udes as much as 1 m from walls and ceilings. 'he cave air temperature is 11 °C in the upper arts, rising to 14 °C in the lowest levels.

ocal Hydrology

The climate in the vicinity of the caves is emi-arid, with a mean annual rainfall between l60 and 420 mm. Radially draining streams that ise in the wetter central Black Hills become nfluent where they cross the sedimentary rocks ind lose much of their water to infiltration, par-icularly into the Pahasapa.

The Pahasapa and its equivalents form a najor confined aquifer in the region. Most of the recharge in the vicinity of the caves appears to be through diffuse infiltration, except where perennial or ephemeral streams sink at rare swallow holes. Ground water flows outward from the Black Hills, much of it emerging at springs along anticlines and fault zones a few kilometres farther out from the limestone outcrops. Some ground water continues down gentle hydraulic gradients into the deep basinal areas beneath the surrounding plains.

The modern caves bear no apparent relation to the present surface topography or stream patterns. They are drained, relict features that have been intercepted locally by shallow canyons that carry seasonal runoff. Wind Cave, however, is located along a paleovalley of possible Tertiary age, now largely abandoned by flow (Palmer, 1981). Jewel Cave is beside Hell Canyon, the major drainage line in the southwestern Black Hills, which carries only intermittent flow. Despite their juxtaposition to major river valleys, there is no clear evidence that the caves were formed by water from these sources.

An important feature of Wind Cave is the presence of permanent lakes at its lowest (downdip) end. Similar lakes are not known in the other Black Hills caves, probably because explorers have not yet found them. The Wind Cave lakes are at 1,120 m above sea level, essentially the same as the static water level in wells that penetrate the Pahasapa Limestone farther downdip. This suggests the existence of a very flat piezometric surface within the limestone. That, in turn, implies that there is high permeability (probably in the form of solutional caves) extending farther downdip below the water level. This piezometric surface is 70 m higher than is Buffalo Gap Spring, 8 km east of the lakes in Wind Cave. The elevation difference represents the head required for ground water to flow upward to the spring through the overlying Minnelusa Formation.

The lakes appear to be stagnant backwaters fed from below. Their level has varied seasonally ~1 m since they were discovered 20 yr ago. Calcite rafts are forming upon them, the calcite precipitating onto dust particles settled on the water surface. Raft formation is indicative of great hydraulic stability and the renewal of a supersaturated solution. Raft debris is abundant as much as 30 m above the modern lakes. It is draped over some helictite bushes, which are fragile, subaerial branching calcite speleothems; very slow, steady rise and fall of the water level is implied by this phenomenon.

A final relevant feature is the occurrence of groups of warm springs at Buffalo Gap, Fall River, and Cascade River (Fig. 1). These springs rise through the Minnelusa Formation where it is exposed as inliers along local anticlinal folds on the dip slope. Water temperatures at points of emergence are 17–26 °C and do not display seasonal variation. Cascade Spring has a mean discharge of 0.6 m^3s^{-1} and is the largest spring of any type in the Black Hills. Most of these springs deposit abundant travertine in their modern channels, which are incised into older alluvial deposits.

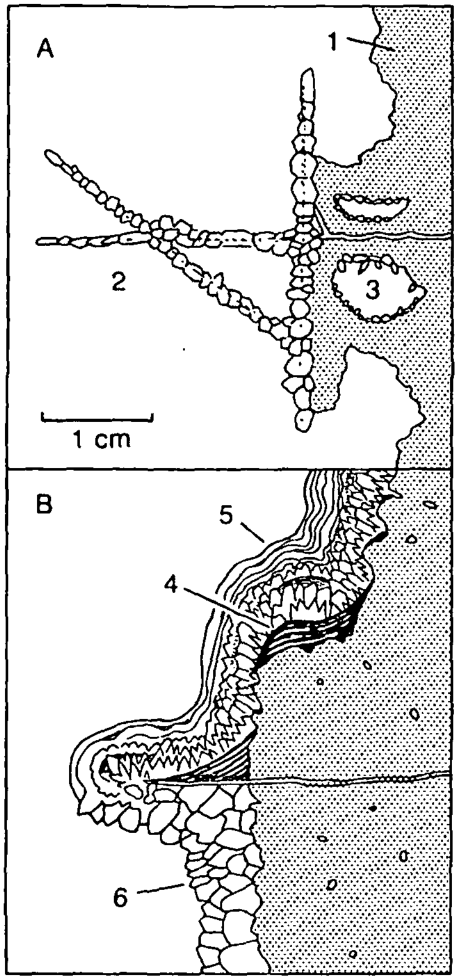

Figure 5. Idealized cross sections through boxwork in Wind Cave. (A) Typical boxwork exposed to weathering in upper passages above the level of calcite wall crust. Most boxwork fins project several tens of centimetres but are attenuated here for clarity. (B) Projecting veins coated with layered calcite wall crust in lower passages. 1 = bedrock (friable in A, competent in B); 2 = boxwork fins, consisting of recrystallized and overgrown pre-cave calcite veins, with ghosts of veins now represented by hematite crystals; 3 = pores in bedrock lined with calcite crystals; 4 = "internal sediment" of detrital carbonate weathered from higher walls; 5 = layered calcite wall crust; 6 = local recrystallized wall crust, the layers of which are faint or absent (in most cases, on undersides of projections).

The lowest point in Jewel Cave lies 15 m above the local water table. The supposed resurgence for ground water of this area is 37 km to the southeast at Cascade Spring (Rahn and Gries, 1973).

Previous Work

Davis (1930) attributed the caves to deep-seated solution of the kind associated with hydrothermal ores on the basis of (a) their equidimensional maze characteristic, suggesting dissolution by slowly flowing waters in the phreatic zone, and (b) the spar coatings of Jewel Cave, which resemble the linings in hydrothermal veins.

Later authors accepted the morphologic evidence of dissolution by low-velocity phreatic water but turned away from the thermal interpretation. The earliest hydrogeologic studies (for example, Darton, 1918) established the existence of a regional, artesian aquifer in the Pahasapa Formation beneath the nearby plains in Wyoming and South Dakota. Tullis and Gries (1938) suggested that the caves were excavated by meteoric water circulating slowly through the aquifer during the Eocene-Oligocene, soon after the uplift of the Black Hills. Howard (1964) developed this concept into a more comprehensive model in which vadose and water-table feeder caves developed updip from the artesian mazes. Cave development was not necessarily tied to Eocene-Oligocene events and could have been much more recent. A problem with Howard's proposal is that only isolated and poorly integrated cave fragments survive in the putative feeder areas.

Deal (1962, 1968) published perceptive studies of the mineral suites in Jewel Cave, recognizing no fewer than seven distinct genetic phases: (1) solution by meteoric waters in a confined aquifer, as above; (2) partial or complete drainage of the caves; (3) return to phreatic conditions, with deposition of the principal deposits of nailhead spar; (4) drainage of certain cavities, as indicated by typical vadose deposits; (5) further complete inundation accompanied by widespread dissolution of nailhead spar in the higher parts of the caves; (6) progressive drainage, with deposition of travertine in the upper cave and of mud in the lower parts; and (7) the modern phase, in which minor travertine deposition continues. The final drainage of the known cave has long been complete; it is a hydrologic relict.

We agree with Deal that the history of dissolution with mineral deposition in these caves has been complicated rather than simple. Two of us (Bakalowicz and Ford) question the strength of the evidence for a vadose phase 2, and clearly, a great problem of Deal's sequence is the integration of an apparent hot-water inundation (stage 3) into what is treated otherwise as alternate filling and emptying of normal meteoric water. White and Deike (1962) used geochemical and mineralogical criteria to suggest pressures of 10–100 atm and temperatures of 150–200 °C during stage 3. White (1982, personal commun.), however, has since accepted that these criteria may be irrelevant and that the minerals in question may have been deposited at much lower temperatures.

Palmer (1975, 1981, 1984) pointed to problems of explaining the caves by a simple confined-flow model. He emphasized that the caves are located in a zone in which flow of water undersaturated with respect to calcite and dolomite is possible to and from the Pahasapa limestones through the overlying Minnelusa sandstone and via flood-water recharge from sinking streams. Examples elsewhere show that network mazes in limestone are commonly formed by either type of recharge.

Wind and Jewel Cave morphology, although unusual, is similar to certain caves of flood-water origin. Palmer (1984) suggested that the major episode of wall coating might be caused by ponding of water in the caves, resulting from the regional Oligocene aggradation. Petrographic evidence suggesting that the caves formed under hydrochemical conditions similar to those of hydrothermal ores, however, has turned Palmer's opinion away from a standard origin (cool water and soil CO_2) for the cave formation.

Presented below are evidences for cave origin and development by ascending thermal water. It is possible that some mixing with cool meteoric waters played a role that is not yet elucidated. The evidence derives from geomorphic features, from A. N. Palmer and M. V. Palmer's petrographic and mineralogical studies, and from isotopic measurements of wall rocks, secondary minerals, and waters by Bakalowicz, Ford, and Miller.

CAVE ORIGIN BY RISING THERMAL WATER

Morphological Evidence

In the Western literature, there is little discussion of modern and relict hydrothermal solution caves. They have been much studied in eastern Europe (Czechoslovakia, Hungary, Poland, and the Soviet Union), however, where Kunsky (1950), Jakucs (1977), Rudnicki (1978), and Dublyansky (1980) have published English or French summaries. Dublyansky listed five criteria that strongly indicate a hydrothermal origin. One of them—composition of exotic precipitates—is considered later in our paper. The other four are morphological.

1. The cave systems lack a genetic relationship to the surface topography.

2. They are largely or entirely devoid of fluvial sediments.

3. The caves in most cases display a three-dimensional rectilinear maze form guided by major fracture systems and, more rarely, by bedding planes. This is indicative of excavation by slowly flowing ascending waters.

4. The highest parts of the caves may display cupola-form solutional pockets dissolved upward into the ceilings. These pockets appear to be convectional in origin. Their form is in most cases different from that of ceiling and wall solution pockets in meteoric-water caves, being more multi-faceted but lacking deep penetration into a guiding joint.

Jewel and Wind Caves meet all four of these criteria very clearly. The first three are noted in our introductory description. Cupola-form ceiling pockets as much as 10 m in height form "The Loft" and other highest places in Jewel Cave. They are best seen in "The Fairgrounds," stratigraphically the highest part of Wind Cave. They are not well developed lower in these caves.

Petrographic and Mineralogical Evidence

Samples of wall rocks and secondary minerals were taken from the caves and nearby outcrops for analysis with petrographic microscope, X-ray, and scanning electron microscope. Samples were obtained under permit from the National Park Service and consisted chiefly of small, detached fragments. A complex sequence of solution, alteration, deposition, and replacement is revealed in the walls of both caves and will be treated in detail in later papers. Only the main aspects pertinent to cave origin are described herein.

The Pahasapa Limestone was subjected to continental weathering late in the Mississippian Period, as noted. In addition to karst forms filled with Pennsylvanian clastics, the carbonate bedrock contains highly fractured and brecciated zones. Wedging features in the breccia indicate an origin due to crystallization and later solution of sulfates. Fractures and breccia interstices then were filled with hematite-rich calcite, as is common during dolomitization. This calcite comprises many of the boxwork veins; most of them are ~100 μm thick, although in breccia zones, some reach several centimetres. The veins show at least two orders of crosscutting that represent different episodes of fracturing. Most are truncated by the solutional paleokarst features, although some extend upward into Pennsylvanian rocks.

The major phase of cave development occurred after the Laramide uplift of the Black Hills. Limestone and dolomite were at first dissolved at approximately equal rates, as shown by somewhat uniform passage enlargement in different lithologies, but during the late stages, dolomite was selectively removed by water that was apparently close to saturation with respect to calcite. Solution of dolomite rhombs created porosity as high as 90% in the cave walls and exposed the calcite veins as resistant fins (Fig. 5). Although the calcite veins are much older than the cave, their exposure as boxwork is the result of cave origin by slow-moving, nearly saturated water quite different from that in karst areas fed by normal surface infiltration.

Iron-rich silica replaced much of the remaining calcite in the porous bedrock. X-ray analysis shows it to vary from opal to microcrystalline quartz. The latter is not uncommon in meteoric-water caves if there are sources of silica. It is an evaporite and thus is limited to frequently wetted patches of rock. In Jewel and Wind Caves, the silica is rather uniformly distributed. This suggests subaqueous deposition, which requires a decrease in either pH or temperature. A small amount of the silica forms meniscus cement, indicating vadose conditions. This may be reworked.

Precipitation of the great calcite spar coatings succeeded silica deposition in Jewel Cave. These crusts average 15 cm thick and contain as many as 20 distinct growth layers (Fig. 4). There are no hiatuses or erosion surfaces between layers. They appear to be cyclic phenomena.

In Wind Cave, the calcite crusts occur as overgrowths on the protruding boxwork veins in the dolomitic middle strata (Fig. 5) and as more general wall cover in the lower cave. They average only a few millimetres in thickness. There are also some pool rim deposits associated with them in the lower cave.

These crusts are subaqueous deposits from water brought to supersaturation either by degassing of CO_2 into air-filled upper caves or by heating. Degassing evidently occurred in Wind Cave. Warming (that was perhaps cyclical) appears necessary to account for the great extent and volume of the encrustation in Jewel Cave.

Modern Geothermal and Hydrochemical Features

Rahn and Gries (1973) studied present geothermal conditions in the Black Hills, including the chemical character of the hot-spring waters. In January 1982, we sampled the hot springs, artesian water, and the different types of water in the caves and obtained the results shown in Table 1. The hot-springs data are essentially identical to those of Rahn and Gries. Cave waters gave results very like those from a larger sampling in Wind Cave by Miller (1979).

TABLE 1. SUMMARY OF SAMPLE WATERS COLLECTED JANUARY 29 AND 31, 1982

	°C	pH	Ca^{2+} (mM/l)	Mg^{2+} (mM/l)	HCO_3^- (mM/l)	SO_4^{2-} (mM/l)	P_{CO_2} 10^{-2}atm	SI calcite	SI dol.	SI gypsum	$\delta^{18}O$ SMOW
Hot Spring	26.7	7.01	2.75	1.03	4.12	6.66	2.5	−0.14	−0.50	−1.28	−16.0
Hot Brook Spring*	17.2	7.80	1.76	0.94	4.20	0.36	0.35	0.39	0.65	−2.04	−14.8
Higher Cascade Spring	20.9	6.75	14.70	2.75	3.80	13.12	3.1	0.04	−0.52	−0.08	−15.4
Lower Cascade Spring	20.4	6.78	14.15	3.03	4.04	20.84	4.1	−0.14	−0.83	0.04	−15.1
Buffalo Gap Spring*	16.9	7.18	13.25	2.10	3.90	11.25	1.5	0.28	−0.15	−0.14	−14.3
Artesian well near Buffalo Gap†	19.0	7.60	1.09	0.49	3.20	0.03	0.56	−0.12	−0.49	−3.19	−11.6
Drip waters in Jewel Cave	9.2	8.30	0.83	1.42	3.86	0.52	0.09	0.45	1.11	−2.13	−12.7
from modern	8.9	8.30	0.85	1.73	4.52	0.13	0.10	0.53	1.34	−2.71	−12.6
speleothems	8.3	8.40	0.65	2.15	5.00	0.32	0.08	0.53	1.53	−2.47	−12.3
Drip water in Wind Cave	9.6	8.07	0.94	0.69	2.92	0.24	0.13	0.14	0.14	−2.36	−11.8
"Calcite Lake" in Wind Cave	12.9	7.93	0.92	0.77	3.49	. .	0.22	0.17	0.29	. .	−12.1
"Windy City Lake" in Wind Cave	14.1	7.90	0.82	0.67	3.00	. .	0.19	0.07	0.08	. .	12.3

*Sampled downstream of spring point.
†Well penetrates Pahasapa Formation.

Rahn and Gries (1973) showed that the Black Hills are characterized by two thermal anomalies. The first is regional, a slightly higher ground-water temperature around the perimeter of the Black Hills uplift. The second is the more sharply defined high-temperature zone around the hot springs. In our data, an artesian well penetrating through the Minnelusa Formation into the Pahasapa Limestone at 2 km from Buffalo Gap hot spring yields a geothermal gradient of 5 °C/100 m. Air and water temperatures at Wind Cave, 7 km from these springs, show a gradient of 3.7 °C/100 m, also rather high. The thermal anomaly observed at the hot springs thus extends beneath modern Wind Cave. There are no comparable data for Jewel Cave.

Schoeller (1962) defined a thermal water as one for which the mean temperature is at least 4° (Celcius) higher than the mean annual surface temperature at the spring. In the southern Black Hills, this implies temperatures above 15–16 °C. Table 1 shows that the hot springs are only feebly thermal but highly mineralized, especially in SO_4^{2-}. Hot Brook Spring (a tributary to Fall River) and Buffalo Gap Spring appear anomalous because they could not be sampled at their bedrock outlets but were sampled downstream after some chemical evolution in the open air in cold weather. The other springs are weakly undersaturated or at equilibrium with respect to calcite and are undersaturated with respect to dolomite. Their calculated P_{CO_2} shows high values, 1.5–4.1 × 10^{-2} atm.

Drip waters in the caves unquestionably represent meteoric infiltration. They are marked by high pH and high Mg^{2+} but very little SO_4^{2-}. They are clearly supersaturated with respect to calcite and are presently depositing stalactites, but they have a relatively low P_{CO_2} (1–2 × 10^{-3} atm). The artesian well water from the Pahasapa Limestone, at a depth of 200 m, is chemically most like the drip waters (P_{CO_2} = 6 × 10^{-3} atm) but is warmed to within the thermal range of the hot-springs anomaly (19 °C). The lake waters of Wind Cave display characteristics intermediate between the drips and the artesian sample. They are best interpreted as local artesian waters that have cooled and degassed in the cave.

Stable Isotope Evidence

Water. Meteoric waters dripping into Jewel and Wind Caves have average $\delta^{18}O$ values of −12.5‰ and −12.1‰, respectively, with respect to SMOW. Yonge and others (1986) have shown that $\delta^{18}O$ of cave drip waters is equal to that of the average annual precipitation in the recharge area. The Jewel and Wind values agree well with the local precipitation values given in Yurtsever and Gat (1981).

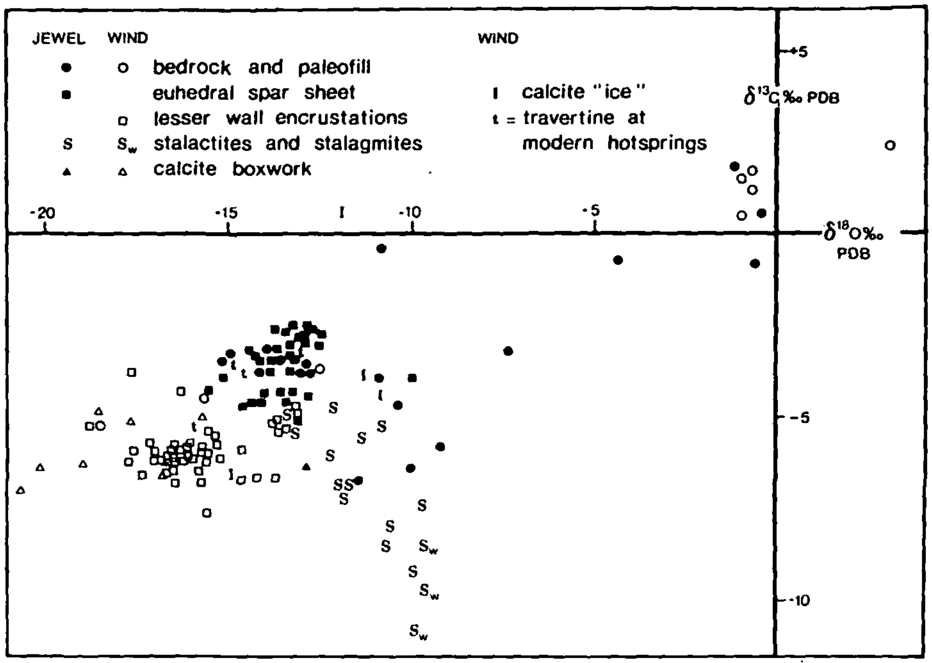

Figure 6. $\delta^{13}C$ and $\delta^{18}O$ per mil values wrt PDB for wall rock, suspected thermal calcites, and normal (meteoric water) speleothems from Jewel and Wind Caves, plus recent and modern hot-springs travertines for Hot Brook and Cascade River, South Dakota.

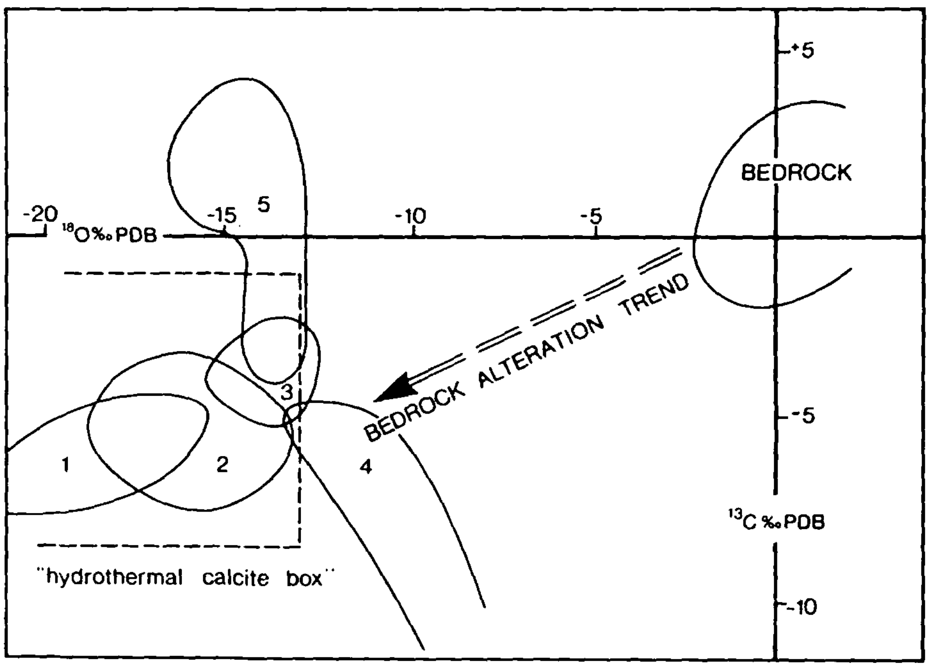

Figure 7. Interpretation of the data plotted in Figure 6, plus data from paleo-hotsprings caves at Budapest, Hungary. Envelope 1 contains all boxwork samples from Wind Cave; 2, all suspected thermal calcite crusts in lower Wind Cave; 3, all euhedral spars from Jewel Cave; 4, normal stalactites and stalagmites from both caves; 5 is the envelope for subaqueous and pool-rim deposits sampled in relict hot-springs caves of Budapest.

TABLE 2. TESTING FOR EQUILIBRIUM OR KINETIC ISOTOPE FRACTIONATION IN JEWEL AND WIND CAVE CALCITES

Site; type	Sample no.	‰ versus PDB $\delta^{13}C$	‰ versus PDB $\delta^{18}O$	Equilibrium precipitation?
Jewel Cave; calcite crusts of supposed hydrothermal origin	JC 0-4	−2.55	−13.52	
	0-5	−2.94	−13.83	Yes
	JC 2-1	−2.60	−13.91	
	2-2	−2.80	−13.92	
	2-5	−2.67	−13.91	Yes
Wind Cave; calcite crusts of supposed hydrothermal origin	WC 0-1	−6.82	−16.54	
	0-2	−6.14	−16.04	
	0-3	−6.55	−16.24	Yes
	WC 8-6	−4.72	−13.76	
	8-7	−4.89	−13.33	Yes
	WC 36-1	−5.88	−16.17	
	36-2	−5.90	−15.92	
	36-4	−5.94	−15.66	Yes
	WC 37-2	−6.52	−17.29	
	37-4	−6.15	−16.95	Yes
	WC 38-1	−6.12	−16.71	
	38-2	−6.36	−16.48	
	38-3	−6.08	−16.23	Yes
Wind Cave; modern calcite "ice"	WC 21	−4.45	−10.94	^{13}C, no ^{18}O, yes
Jewel Cave; normal stalactite	JC 11/3-1	−8.13	−10.58	
	11/3-2	−6.12	−12.19	No

Note: JC 2-1, 2-2, 2-5, and so on = samples measured at fixed intervals along one travertine growth layer.

The hot-springs waters display average $\delta^{18}O$ of −15.1‰, that is, 3‰ lighter than the drip waters. The possibility that this depletion is due to discharge of "fossil" Pleistocene water from cooler climatic phases may be ruled out on quantitative grounds. The depletion must be attributed to isotopic exchange with depleted rocks.

Calcite. To investigate carbon and oxygen isotopic characteristics, 75 samples were collected and 150 analyses made by mass spectrometry. Samples included cave wall rock (ranging from fresh to highly weathered), paleokarst fills, euhedral spars, boxwork and other exotic coatings and modern (meteoric water) speleothems in the caves, plus travertine and water from the hot springs. Results are displayed in Figure 6 and interpreted in Figure 7. Seven samples of the calcites believed to be of hydrothermal origin were selected at random to test for isotopic fractionation. In terms of the criterion of Hendy (1971), all of these were deposited in isotopic equilibrium with the source water (Table 2). One ordinary stalactite that was tested was not in equilibrium.

In ordinary speleothems (that is, deposited from meteoric waters), $\delta^{13}C$ values are about −11 ± 2‰ PDB. In effect, the carbon is a one-to-one mixture between the carbonate rock ($\delta^{13}C$ = 0 ± 3‰ PDB) and soil CO_2 ($\delta^{13}C$ = −22 ± 5‰ PDB). More positive values of $\delta^{13}C$ (as in all suspected hydrothermal precipitates shown in Fig. 6) are explained by precipitation from HCO_3^- which is enriched in ^{13}C with respect to such a mixture. Such CO_2 could be derived from a magmatic source or by high-temperature decarbonation of limestone (Truesdell and Hulston, 1980).

The equilibrium ^{18}O concentration in speleothems is determined principally by the concentration in the source water and the water temperature at time of calcite precipitation. In Figure 6, it is shown that with few exceptions, all suspected hydrothermal calcite deposits in Jewel and Wind Caves plot within the domain of hydrothermal calcites (Fig. 7), as shown by Friedman (1970), Robinson (1975), Barnes (1979), and Hoefs (1980). Travertines from the Black Hills hot springs are also in this domain. Three samples of modern "calcite ice" precipitating on the surface of a lake in the bottom of Wind Cave have $\delta^{18}O$ values in equilibrium with the modern temperature there but are enriched in ^{13}C if compared to normal speleothems in the caves. By contrast, an ancient "ice" sample 30 m above the lakes has more characteristically hydrothermal isotope ratios (−6.64‰ $\delta^{13}C$ and −14.9‰ $\delta^{18}O$ PDB). One small and aberrant sample of spar from Jewel Cave (−10.0‰ $\delta^{18}O$) is now believed to have derived from the paleokarst fill.

The great encrustations of nailhead spar in Jewel Cave display little isotopic variation; their mean value is $\delta^{18}O$ = −13.8 ± 0.7 (1 σ) PDB. If the temperature relationship proposed by O'Neil and others (1969) is used, their depositional temperatures were probably in the range 15 to 35 °C. The mean value of boxwork in Wind Cave is $\delta^{18}O$ = −18.1 ± 1.6 PDB. This corresponds to a temperature range of 30 to 60 °C. Wind Cave boxwork has precisely the same isotopic range as does the hot-springs calcite in Yellowstone Park reported by Truesdell and Hulston (1980).

Modern dripstone and flowstone speleothems that have been deposited by meteoric waters infiltrating into the caves are generally quite distinct. Some Jewel Cave samples are unusually enriched in ^{13}C and depleted in ^{18}O. These may be disequilibrium deposits, as in the example given in Table 2. Alternatively, their feed waters may flow over or through the spar sheets as well as through isotopically depleted country rock, exchanging with these depleted rocks. Wind Cave speleothems are not depleted in ^{18}O; there are no great barriers of spar along the courses of their feed waters. Samples of the wall rocks are also shifted to lower $\delta^{18}O$ and $\delta^{13}C$ values, presumably by exchange and some recrystallization in the same hydrothermal waters.

For comparison, Ford collected samples of spar, lesser crusts, and pool rim deposits from caves at Budapest that undoubtedly are of hydrothermal origin. They display the same depletion in ^{18}O as do the suspected hydrothermal calcites in Jewel and Wind Caves (Fig. 7). Some Budapest samples are notably enriched in ^{13}C. This is probably due to local, high-temperature metamorphism of limestone along some master joints during a Miocene volcanic phase that preceded cave genesis there (Muller, 1987). Budapest cave wall-rock samples also display a strong complementary alteration trend.

In summary, conditions similar to those measured at the Black Hills hot springs today ($\delta^{18}O$ of waters in the range −14‰ to −16‰ SMOW and temperatures of 20 to 40 °C) will readily explain the isotopic composition of most of the exotic precipitates sampled in the caves.

Uranium Series Dating of Cave Calcites

The caves are devoid of flowing water today and are therefore hydrologic relicts. We thus cannot measure directly the conditions that created them. It is possible that they were formed as early as the late Eocene–Oligocene or in the mid-Tertiary and thus in origin, might be fully divorced from any current geohydrologic or geothermal conditions, although the contrary is implied by some of the hydrochemical and stable-isotope evidence already discussed.

Tables 3 and 4 present U-series dates for the cave and hot-spring deposits. The modern rate of deposition of travertine at the hot springs appears to be very rapid; samples are difficult to date with great precision because of detrital thorium contamination, a problem that is encountered in many subaerial tufas. Sample CRO 1 (which was taken from an extensive terrace that is now distant from the modern hot-spring channel of Cascade River), however, can be only a few thousand years in age.

Passages in Wind Cave rise to a maximum height of ~145 m above the basal lakes. Sample WC+ was a small nailhead spar encrustation

TABLE 3. $^{230}Th/^{234}U$ AGES OF SECONDARY CALCITE SAMPLES IN WIND CAVE AND AT CASCADE RIVER

Sample no.	Description	U (ppm)	$^{234}U/^{238}U$	$^{234}U/^{238}U_0$	$^{230}Th/^{232}Th$	$^{230}Th/^{234}U$	Age (Ka)
WC +	Nailhead spar, central cave; ~1,200 m asl	3.57	1.568	. .	33	1.099	>350
WC 7	Wall crust, Boxwork Pit; ~1,180 m asl	2.17	1.587	2.157	540	0.982	242 ± 16
WC 9	Vadose flowstone floor, Boxwork Pit; ~1,180 m asl	3.17	1.580	2.035	134	0.923	205 ± 14
WC 5	Rafts of "calcite ice" draped upon WC 7 and WC 9	3.579	1.663	2.114	60	0.890	185 ± 12
WC 6	Cornice grown upon WC 7; probable pool surface deposit	7.86	1.122	1.189	110	0.775	154 ± 2
WC 10	Wall crust, Rescue Pit; ~1,155 m asl	2.67	1.670	2.284	316	0.974	230 ± 20
WC 12	Cornice, Rescue Pit	0.532	1.538	1.665	7	0.571	76 ± 5*
WC 20	Wall crust; ~1,145 m asl	2.62	1.585	2.101	210	0.957	225 ± 15
WC 14	Wall crust; ~1,135 m asl	1.71	1.736	2.421	47	0.984	234 ± 20
WC 19	Fine-grained calcite crust 10 m above Calcite Lake; shows re-solution	1.51	1.514	1.922	24	0.931	208 ± 23
WC 18	Coarse-grained calcite crust 10 m above Calcite Lake; shows re-solution	1.14	1.459	1.706	350	0.802	153 ± 18
WC 16	"Calcite ice" draped 25 m above lakes	1.62	1.600	1.600	1.3	0.014	1.6 ± 0.4*
WC 17	"Calcite ice" draped 14 m above lakes	1.29	1.773	1.773	5	0.002	0 ± 0.1*
81062-2	"Calcite ice" draped 5 m above lakes	2.19	1.627	1.627	1.5	0.016	0 ± 0.3*
WC 21	"Calcite ice" floating on Calcite Lake; 1,120 m asl	1.41	1.810	1.817	10	0.032	3 ± 0.2*
CRO 1	Oldest tufa terrace below Cascade River hot springs	1.04	1.903	1.906	2	0.029	3 ± 0.2*

Note: $^{234}U/^{238}U_0$ = calculated ratio of these two species at time of co-precipitation in the calcite.

*Age calculated assuming an initial $^{230}Th/^{232}Th$ ratio of 1.25.

TABLE 4. $^{230}Th/^{234}U$ AGES OF SECONDARY CALCITE SAMPLES IN JEWEL CAVE

Sample no.	Description	U (ppm)	$^{234}U/^{238}U$	$^{234}U/^{238}U_0$	$^{230}Th/^{232}Th$	$^{230}Th/^{234}U$	Age (Ka)
JC 1*	Nailhead spar crust, 12 cm thick; whole-rock age	0.37	1.010 ± 0.03	. .	11	1.07	>350
JC 1T	Top 2.5 cm of JC 1	0.36	0.922 ± 0.02	. .	158	1.08	>350
JC 1B	Basal 2.5 cm of JC 1	0.45	1.001 ± 0.01	1.001	186	0.912	265 ± 30
JC 3	Nailhead spar 6 cm thick	0.30	0.998 ± 0.02	. .	107	1.62	>350
JC 7A/T	Nailhead spar crust 6 cm thick; top 1.0 cm	0.17	0.922 ± 0.11	. .	86	1.04	>350
JC 7A/M	As above, central 1 cm	0.33	0.972 ± 0.02	. .	170	1.02	>350
JC 7A/B	As above, basal 1 cm	0.49	0.980 ± 0.02	. .	780	1.02	>350
JC 7B	Nailhead spar sheet 2–5 cm thick, below JC 7A and separated from it by a thin red and black silty layer	0.18	0.986 ± 0.04	. .	130	1.64	>350
RL 18A	Nailhead spar from lowest part of cave; A = base of spar crust	0.28	1.011 ± 0.12	. .	123	0.99	>350
RL 18B	As above; B = top of spar crust	0.18	1.047 ± 0.10	. .	15	1.04	>350
JC 11M	Stalactite drapery (vadose) growing over spar, now fallen from wall; 4–5 cm above base	7.47	1.078 ± 0.01	1.119 ± 0.02	690	0.766	153 ± 18
JC 11T	Stratigraphic top of JC 11; 9–12 cm above base	2.97	1.173 ± 0.02	1.233 ± 0.02	76	0.64	106 ± 6
JC 11TR	Replicate of JC 11T	1.72	1.170 ± 0.02	1.226 ± 0.02	87	0.62	102 ± 7

*This sample displayed normal magnetic polarity; signal very weak.

taken ~80 m above the lakes (Table 3). It indicates that the cave was filled with water to or above this level before 350 Ka. ^{234}U and ^{238}U are far from equilibrium with each other (compare with the Jewel Cave spar samples, Table 4), indicating that the sample is probably much younger than 1.25 Ma.

The remaining samples were collected between the lakes (water table) and Boxwork Pit, 60 m above. Samples WC 7, 10, 20, 14, and 19 are of thin, discontinuous calcite wall crusts occurring throughout this height range. They indicate that between, broadly, 200 and 250 Ka, this lower region of the cave was water filled and experienced slow deposition of calcite everywhere. Crust deposition continued, in the lowest places at least, until ~150 Ka (WC 18).

Between ~200 and 150 Ka, the water table appears to have stood close to the bottom of Boxwork Pit, with some oscillation through a range of several metres or more. This is illustrated by the excellent stratigraphic sequence of dates for phreatic crusts, pool rimstones, and "calcite ice" shown in Figure 8. The dynamic hydrologic conditions implied by this figure appear to be the same as those that now occur at the modern lakes 60 m below.

Abundant deposits of "calcite ice" accreted to dust particles on the lake surfaces and became stranded as the lakes withdrew. The dust nuclei create serious detrital thorium problems for dating, but preliminary results suggest that the pattern of slow lake-level fluctuations superim-

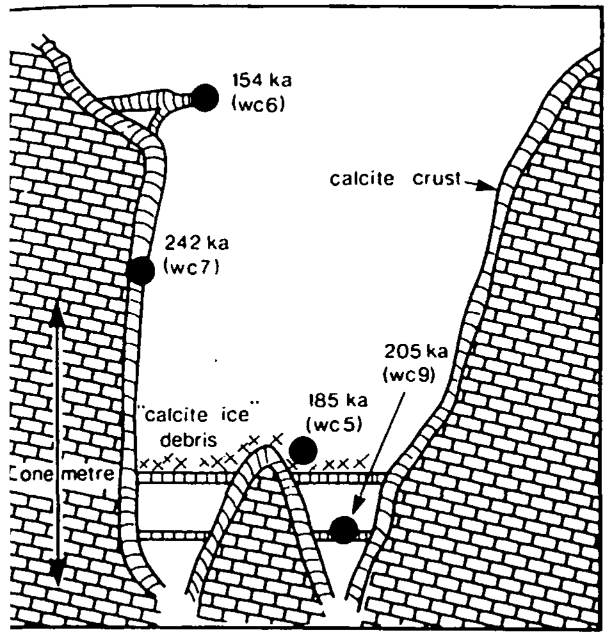

Figure 8. $^{230}Th/^{234}U$ **ages of the general wall crust, a former flowstone floor, a waterline fin, and lake "ice" of calcite at the bottom of Boxwork Pit, Wind Cave. This is from a field sketch; the scale is only approximate. See the text for discussion.**

sed onto a longer term lowering are ntinuing (samples WC 16, 17, 21, and 81062- Table 3).

These U-series results for Wind Cave are ;hly consistent. They show that at least the ver half of the cave has been in a phreatic te with warm-water calcite precipitation dur- ; the past 250,000 yr or less. Most or all of its utional excavation could have occurred im- ıdiately before crust deposition. The lowest e-third of the explored cave (lying below the ıin boxwork zone) has drained as a backwater ce ~200 Ka and continues to do so today. A ›re comprehensive dating study is now in ›gress to elucidate the details there.

Jewel Cave stands 400 m higher in elevation ın does Wind Cave and is much farther from : modern hot springs. It is to be expected that ıas been relict for a longer period, even in its vest parts. This is clearly borne out by U- ies analyses of the great spar sheets. Our sam- ng included spars from high, intermediate, d low sites in the cave. If sample JC1B is set de as aberrant (Table 4), all specimens are ler than 350 Ka in age, the limit of the $^{230}Th/^{234}U$ method. The ratio $^{234}U/^{238}U$ is less ın or equal to unity. This is in marked contrast Wind Cave. We have no means of estimating : initial $^{234}U/^{238}U$ ratio in spar sheets at wel Cave. The lowest modern ratio is 0.922, ›orded in the stratigraphic tops of samples :1 and JC7A. If it is assumed that this ratio rresponds to an age of 350 Ka (it cannot be unger), then a ratio of 1.00 ± .01 would represent a minimum age of 830 Ka. The bases of the Jewel Cave spar sheets are most probably older than 1.0 Ma in age.

The remanent magnetism of sample JC1 was measured. The polarity was normal, but the signals were very weak; no age inferences can be drawn from the data.

As noted, there are very few normal (meteoric water) calcite speleothems in these caves. At Jewel Cave, one of the most massive (and presumably older?) examples had fallen and shattered. Samples taken from its stratigraphic middle and upper parts are 150–100 Ka.

At Jewel Cave, the spar sheets are ancient, and their deposition could have been completed before 1.0 Ma. Much of the cave may have been drained at that time. If that is the case, there has been remarkably little development of overhead infiltration routes into these empty voids during the past 1 m.y. because even comparatively massive stalactites are quite young.

DISCUSSION AND CONCLUSIONS

The evidence we have presented suggests that the large network caves of the Black Hills were formed by CO_2-rich waters that were heated and that ascended through the Pahasapa Formation. Most or all of the thermal precipitates in Wind Cave are of Quaternary age. It is possible that the dissolutional enlargement of this cave was also limited to the Quaternary, but we suspect that it probably began during the Miocene or Pliocene. Formation of Wind Cave is compatible with the geothermal and hydrogeological conditions that exist in its local region today. Waters rose and converged through what is now the cave zone en route to spring points in an adjoining valley. As a consequence of continued surface entrenchment, spring positions have shifted down dip. The valley is now a dissected relict, and the known cave has become a drained backwater. It retains a strong thermal gradient, and final-stage warm-water precipitates (the "calcite ice") are still forming at the modern water table in it, where exploration is terminated.

Jewel Cave is significantly older and more completely relict. It is not related to the hydrogeological conditions prevailing in its region today. Nevertheless, we suggest that it was formed in the same mode as Wind Cave, by thermal waters rising and converging through it toward spring positions in an earlier level of Hell Canyon. The culminating morphological event in Jewel Cave was the deposition of the calcite spar sheets. They are among the greatest known in any cave. We have shown that they are deposited from thermal waters probably at some time before about 1.0 Ma.

The caves display a broad tendency to descend in stratigraphic elevation in the direction of stratal dip. This suggests that the thermal plumes were rising with an updip component. An igneous heat source within the Precambrian rocks is envisioned, related to the igneous activity occurring elsewhere in the Black Hills throughout much of the Tertiary. Recharge to the systems was probably from infiltration of meteoric waters over wide areas, a pattern of circulation that has been documented in many geothermal systems (Dublyansky, 1980; Ellis and Mahon, 1977). The position of the Pahasapa Formation close to the base of the sedimentary sequence plus the greater elevation of the Black Hills make it unlikely that any significant part of the cave discharge consisted of basinal fluids from strata beneath the surrounding plains.

There were three principal modes of cave development: (1) solution of limestone and dolomite at nearly equal rates by water considerably undersaturated with respect to both carbonates, which mode was quantitatively predominant, (2) selective solution of dolomite only, by water near to saturation with calcite, and (3) deposition of calcite from supersaturated water. In an ideal thermal model, all three stages will occur simultaneously in a vertical sequence. At any fixed P_{CO_2}, the saturation concentration of dissolved carbonate in water increases as the temperature decreases. As the thermal water rises and cools, it acquires or retains solutional aggressiveness with respect to both calcite and dolomite, regardless of the initial dissolved carbonate content. If cooling of water is very gradual, however, the system can hover near the saturation value of calcite and dolomite, preferentially dissolving the species that is more soluble under prevailing geochemical conditions.

Decreasing hydrostatic pressure in the rising water may allow partial degassing of CO_2, which sharply reduces the saturation concentrations and causes precipitation of the secondary carbonates. At most sites observed by Ford in the Budapest thermal caves, precipitation was intense down to 2 m below the paleo-water tables and reduced to zero at depths greater than ~10 m. Rapid degassing in well-ventilated hills best explains such sharp zonation. Wind Cave is not so well ventilated (despite its name); slower degassing probably explains its poorer zonation of precipitates. Simultaneous deposition of the spar crust throughout Jewel Cave can be explained by a phase of warming, inducing degassing within the cave zone, or by a protracted backwater phase marked by very slow circulation and degassing.

This model is highly simplified and must be modified to account for details in the history of the individual caves. Water levels appear to have fluctuated in response to local aggradation at springs or to wetter spells during the later Tertiary and Quaternary, within the over-all lowering induced by regional erosion. As higher springs were abandoned, meteoric flood waters may have penetrated by way of them, contributing to cave enlargement.

Finally, the three-dimensional network pattern of the caves must be explained. Their origin requires a way of distributing the solutional capacity of the water rather uniformly between major joints over particular areas of several square kilometres or more. We suggest that regional and diffuse, heated discharge converged upon what became the cave zones, flowing up dip in the Pahasapa and ascending through the lower formations. Cooling of these waters simultaneously throughout the joint nets produced the crucial solutional aggressiveness. Fluctuating head within the evolving cave zones (in response to varying recharge) and mixing corrosion probably played subordinate roles.

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We are deeply indebted to the administrative staff of Wind Cave National Park and Jewel Cave National Monument for permitting access to all parts of the caves and the collection of the rock and secondary mineral samples. The stable-isotope and U-series analyses were carried out in the McMaster University laboratories by Martin Knipf and Nicolette Caesar under the direction of Ford. H. P. Schwarcz discussed and evaluated the results with us; we are also indebted to L. R. Gardner and M. Gascoyne for their comments. Field and laboratory expenses for Bakalowicz, Ford, and Miller were paid from a general operating grant to Ford from the Natural Sciences and Engineering Research Council of Canada. Bakalowicz's North American sabbatical studies were supported by a North Atlantic Treaty Organization Scholarship and the Centre National de Recherche Scientifique, France. Much of the field and laboratory work by A. N. Palmer and M. V. Palmer was supported by the Wind Cave and Jewel Cave Natural History Association. The Speleological Society of Hungary graciously guided Ford through representative caves at Budapest and permitted collection of samples for comparison.

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RADON EMANOMETRY IN GEOTHERMAL EXPLORATION OF VOLCANIC ZONES

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ABSTRACT

Emanometry of Radon on surface of geothermal volcanic areas allows to determinate those subsurface zones with low amount of sealed geologic structures. These non-sealed fractures behave like conduits of Radon flow which is used as a pathfinder of subsurface geothermal fluids. Radon measurements in surface can be made fastly and easily, to a low cost, using an appropiate sampler and a plastic detector; this can record the tracks of alpha particles which are produced by radioactive decay of Radon. Such a sampling methodology has proved to be useful to measure Radon concentrations in México, at the Los Azufres, Michoacán, geothermal field, and at the Las Tres Vírgenes, Baja California Sur, geothermal zone, both in volcanic framework.

INTRODUCTION

Presence of faults or fractures in volcanic zones is one of the most important factors for geothermal exploration; however, it is not enough to detect fracture evidences but these structures must be without filling material along their fracture planes, in order to act as conduits for the probable underground geothermal fluids.

Measurement concentrations of Radon gas on the surface (emanometry), is a good way for taking a decision on what structures can be conductors in a geothermal zone. Basis isnot complicated. Radon is a noble and radioactive gas which originates to depth and raises to surface --with no combination, mixing nor dilution-- using most expeditous ways. In Nature, one of ways is through planes of failure or fracture; therefore, if measurements of Radon concentrations on surface are the greatest, also is detected a subsurface zone with high density of fracturing. These zones are the most attractive ones since the point of geothermal view.

There are some ways for measurement of Radon concentrations on surface. But one --most simple, fast and effective-- is making use of radioactive properties of Radon. Thus, this gas decays by emission of alpha particles whose tracks or impressions can be recorded; these tracks are formed when an alpha particle meets a plastic detector. Tracks have just some 100 armstrongs of diameter, although they can be enlarged and then be evaluated by an optical microscope. Amount of tracks are directly proportional to amount of Radon.

This sampling method was experimentally carried out in the Los Azufres, Mich., geothermal field, and in the Las Tres Vírgenes, B.C.S., geothermal zone, both located at volcanic regions of México. In both cases was determinated the amount of tracks of alpha particles by area unit and by time of exposition; anomalous zones coincided in most cases with high potential geothermal areas. Thereafter, this Radon sampling method has proved to be an useful tool in geothermal exploration, eventhough testing should be extended to other areas under geological survey.

METHOD'S BASIS

Radon has, in Nature, just three isotopes: 219 Rn, 220 Rn, and 222 Rn, which are products from intermediate decay of the radioactive families from 235 U, 232 Th, and 238 U, respectively. 219 Rn, namely Actinon, and 220 Rn, known as Thoron, have very short half lives. These three Radon isotopes are radioactive themselves and decay to isotopic forms of Polonium through alpha emission with distinctive energy, together with gamma emission. Table 1 shows some nuclear properties of Radon isotopes.

ISOTOPE	HALF LIFE	ALPHA ENERGY	GAMMA ENERGY
$^{219}_{86}Rn$	0.03 sec.	7.13 (99%) 6.52 (0.2%)	0.61 (0.2%)
$^{220}_{86}Rn$	51.5 sec.	6.28 (99%) 5.75 (0.3%)	0.54 (0.3%)
$^{222}_{86}Rn$	3.83 days	4.58 (99%) 4.98 (0.08%)	0.51 (0.08%)

Table 1.- Radon isotopes.

When an atom of Radon decays by emission of an alpha particle, it can be recorded by a near detector; then, the particle interacts with the detector and makes on it an atomic damage (track), which remains latent and can be seen only under an electronic microscope. However, that track can be enlarged to an enough size for optical microscope --by chemical attack on detector.

It was tested four kinds of plastic detectors, being all of them polimeric made with cellulose nitrate or acetate and having 15 to 600 microns of width. One of these detectors was most efficient for record of alpha particles tracks; it is commercially known as LR-115 and is made by cellulose nitrate with 15 microns of width. In addition, it was proved several chemical attack conditions for enlargement of alpha tracks (etching); best result was obtained by using of Sodium Hidroxide to 20%, at 50 °C during three hours (Gutiérrez-Negrín and López-Martínez, 1983).

By another hand, Radon isotopes originate to depth; owing its greater half life, 222 Radon is the sole with probabilities for arrive and be detected at surface. But in some cases in situ generation of 220 Radon is probable, owing to very small amounts of Thorium in soils. This 220 isotope also decays by alpha particles which could be record at the plastic detector; thus, it is convenient to reduce the probabilities of recording alpha particles tracks from Thoron.

With that objective, several types of samplers were proved, from classic inverted cup sampler (Fleischer and Likes, 1979) to new sampler designs that were ex profeso made; one of these types, nameley M-5 (Fig.1), was the best. Length of its inner tube --25 cm-- is enough for fast decay of any emanation from Thoron or Actinon before it can get to detector. The intertube place is filling with cotton or wool which absorbs humidity that could be condensed, and disposable plastic cup put on top prevents water infiltrations from surface. The M-5 sampler just needs a little 40 cm depth hole; therefrom, it can be easily placed by just one people without special tools. Furthermore, the same sampler --except its detector-- can be used again for several times.

APPLICATION IN THE LOS AZUFRES GEOTHERMAL FIELD

The Los Azufres geothermal field is located at northeastern portion of Michoacán state, 100 km from Morelia City, in the central part of México (Fig. 2). It extends on an area of 190 square kilometers. In this field drilling includes 45 geothermal wells, with an average of 1500 m depth. About 21 wells are productive and available for production of electricity; whole production now is near of 1300 steam tons per hour, though electric generation in only 25,000 kw.

Stratigraphical sequence is consisting of volcanic rocks, from Upper Miocene andesites to Pleistocenic basalts and pyroclastics deposits, including rhyolites, tuffs, dacites, glassy domes and pumicite tuffs (Gutiérrez-Negrín and Aumento, 1982). Geothermal reservoir is in andesites. It has been considered that structural systems determined geothermal fluids circulation at depth, also size and boundaries of hydrothermal system itself (De la Cruz and Castillo, 1984).

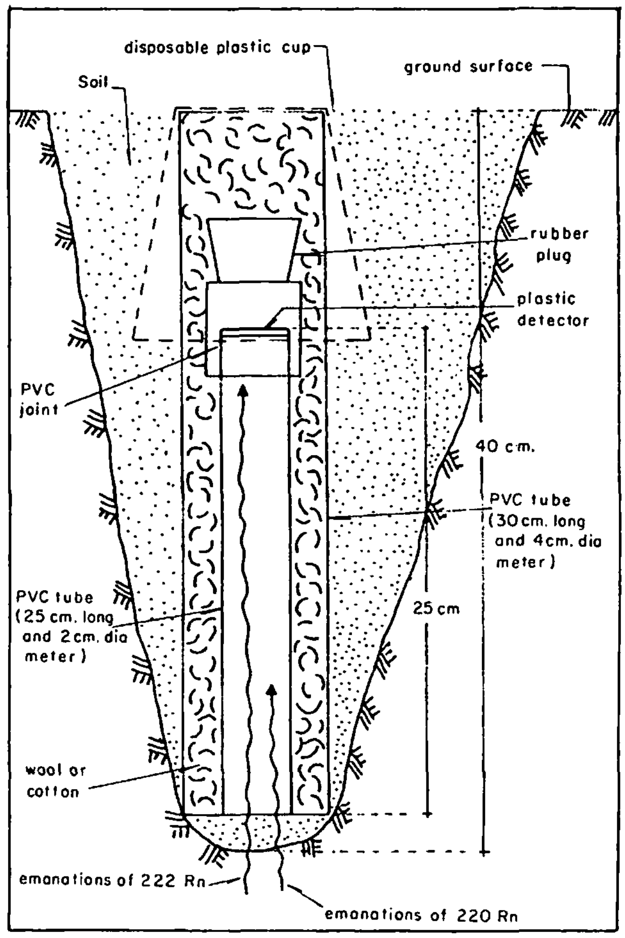

Fig. 1.- Radon sampler M-5.

Radon sampling carried out at southern zone of field; 175 M-5 samplers were placed --with respective LR-115 detectors-- spaced at about 150 m intervals, forming a network with N-S and E-W lines cutting all known structures at that zone. 21 days later, only 153 samplers were gotten --rest was lost. Detectors were etched with NaOH under conditions above mentioned. Then, each detector was computed under an ordinary optical microscope with 470 X enlargement.

Amount of tracks was expressed in relation to area and exposition time, as units of tracks per square centimeter and per hour (t/cm^2h); figure 3 shows an histogram with all values. Average value was 24 t/cm^2h, standard deviation was 16, minimum value was 2 and maximum one 83. It is obvious that these units do not express real quantity of Radon, but determination of its readioactive effects is directly proportional to that quantity. Furthermore, sampling purpose is to determine those areas with more open structures rather than absolute amount of Radon.

Figure 5 pointed out isoconcentration curves

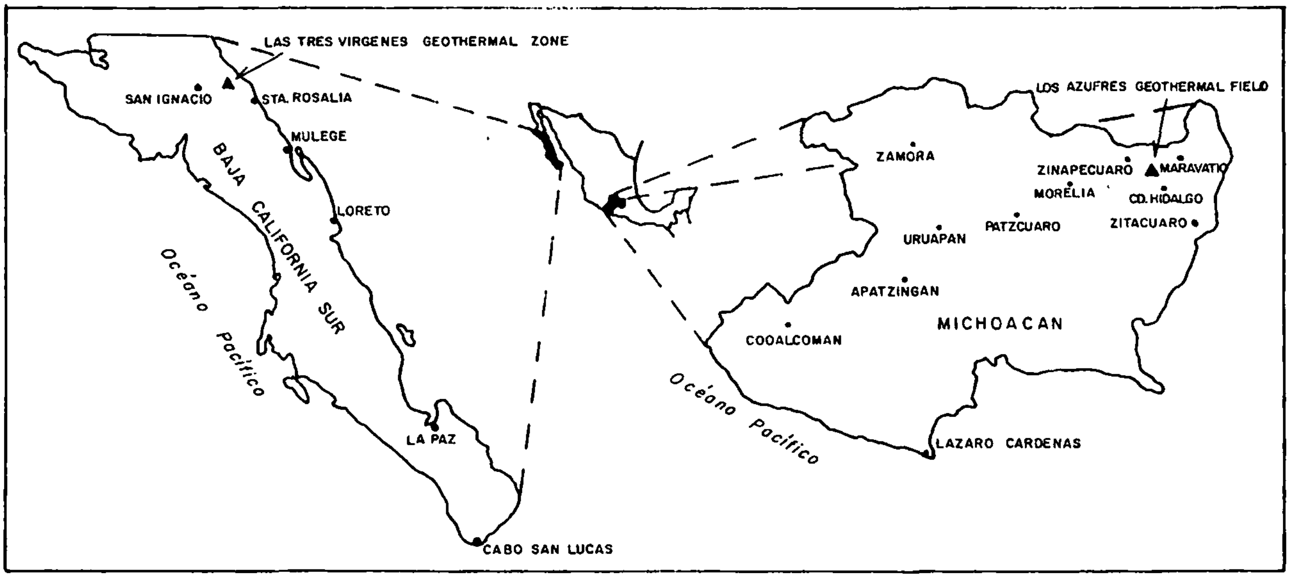

Fig. 2.- Location map showing the Las Tres Vírgenes and Los Azufres geothermal sites.

of t/cm^{2}h --that is, uniform Radon concentration curves--, rebounding some anomalous zones with high concentrations; an anomalous value was that greater than average plus standard deviation, it is to say greater than 40 t/cm^{2}h. Those anomalous zone have general coincidence with location of productive wells and with superficial planes of known structures (Fig. 5); likewise, some unproductive wells are related with lesser than 10 t/cm^{2}h zones, that may be considered as minimum ones. Additionally, if one analize results detaill-ly, one would found interesting data on "conductive" or "no conductive" behavior for each distinctive fault or fracture; this can help for location of next geothermal wells (Gutiérrez-Negrín and López-Martínez, 1983).

APPLICATION IN THE LAS TRES VIRGENES GEOTHERMAL ZONE.

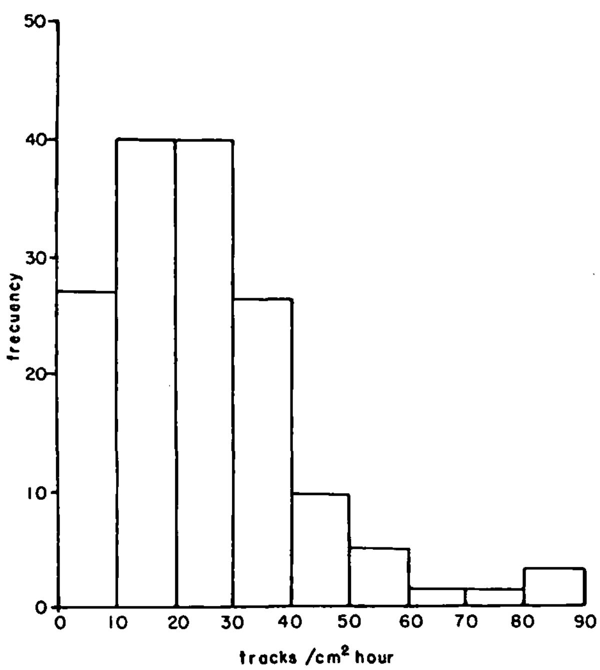

Fig. 3.- Histogram of Radon values (Los Azufres).

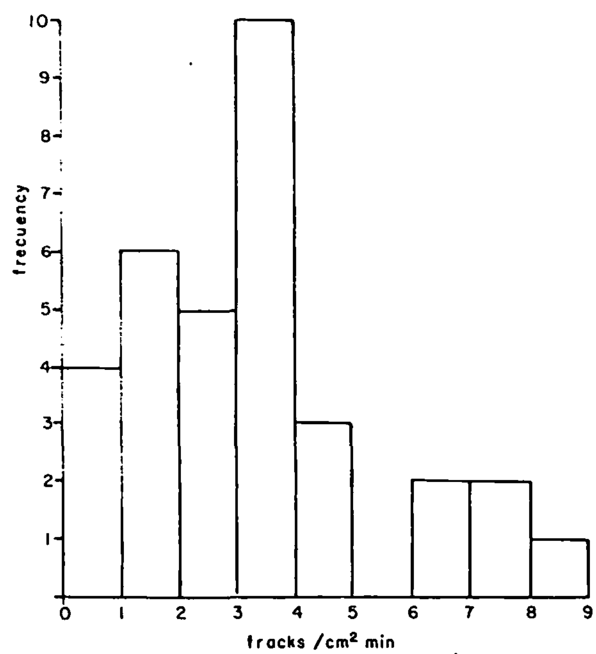

Fig. 4.- Histogram of Radon values (Tres Vírgenes).

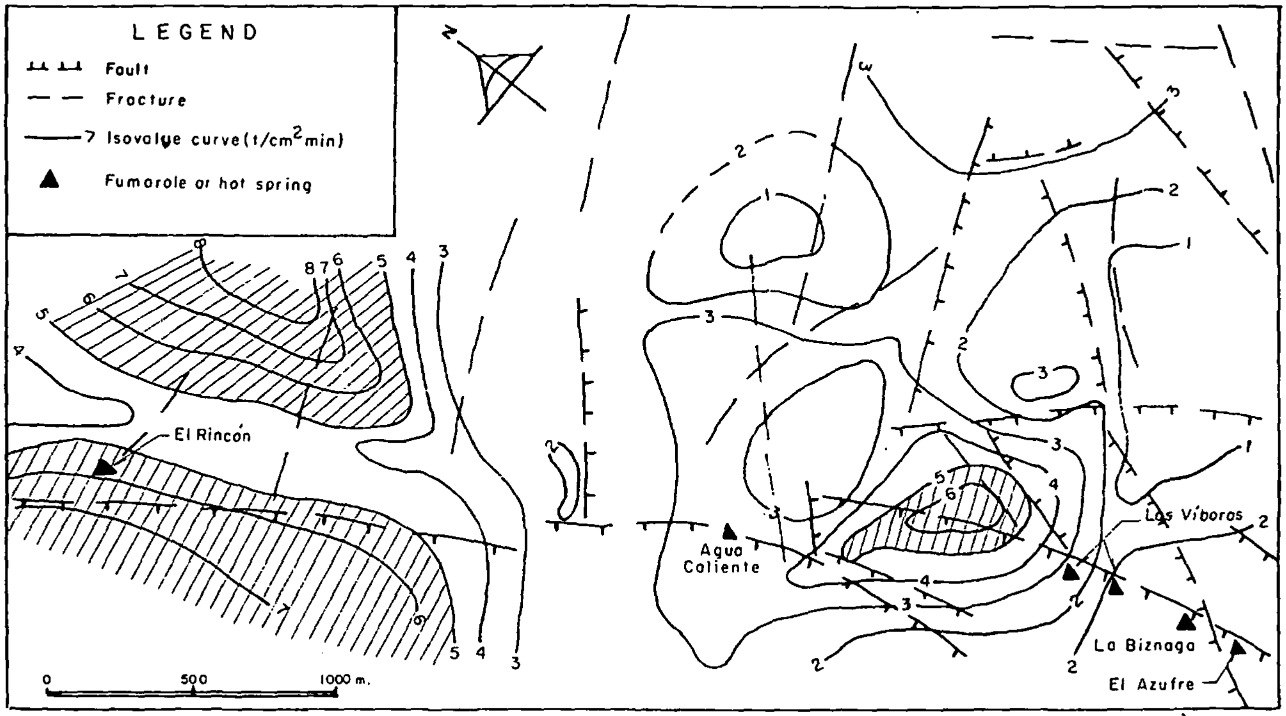

Figure 6.- Map showing high Radon concentration zones in Las Tres Vírgenes.

were expressed in alpha particles tracks per area unit and per exposition time unit-- are directly related with non-sealed fractures and faults that act as conduits.

2. In geothermal volcanic zones faults and fractures driving Radon have high probabilities of driving also underground gethermal fluids. Therefore, determination of superficial zones with anomalous Radon concentrations allows to help for choose the most interesting places for drilling exploration.

3. It is suggested to use the M-5 sampler, easily available due to its low cost, uncomplicated manufacture and good efficiency, and to use the LR-115 plastic detector --which is commercially made in some countries.

4. From the above considerations, is suggested also that this methodology of measurements of Radon concentrations could be included as a routinary method in geothermal exploration of volcanic areas.

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This geothermal zone is located northeasternly of Baja California Sur state (Fig. 2), 35 km from Santa Rosalía town. It is an elongated portion from NW to SE, in a 10 square kilometers area, with nine thermal places --essentially hot springs, fumaroles and hydrothermal alteration zones--, and superficial temperatures between 58 and 98 °C.

Las Tres Vírgenes geothermal zone seems tectonically related with a great transform fault that extends from the California Gulf and penetrates into continent; interaction between this fault and a tectonic weak zone with N-S direction --which is evidenced by alignment of Quaternary volcanoes-- would be responsible for the origin of active geothermal system (Lira et al., 1984). Locally, fossiliferous sandstones from Lower Pliocene are oldest outcropping rocks, begining the volcanic sequence at the top of them. It is represented by andesitic, pumicitic, ignimbritic and dacitic rocks, overlapped by Pleistocenic conglomerates and alluvion; last eruption of near Las Tres Vírgenes volcano happened in 1746 (Mooser and Reyes; Iven; after Lira et al., 1984).

46 M-5 samplers with LR-115 detectors were placed at intervals of 500 m, in a network whose lines were orientated NW-SE and NE-SW. After 30 days, samplers were recovered and their detectors were treated at before mentioned conditions. Notwithstanding, that conditions become too much drastic for this zone, because of its high superficial temperatures and its great emission of gas like H_2S; therefore, concentration of NaOH and temperature were keep up, but etching times were changed, specifically by each detector in order to get a residual width between 3.5 and 4 microns. This residual width was keep up uniform for all detectors, since the amount of observed tracks is a function of that residual width (López-Martínez, 1984).

After track computing, a correction --which was obtained as result of specific test (López-Martínez, 1984)-- was applied in order to compensate tracks destroyed by high superficial temperatures --annealing-- and by emission of H_2S. With that correction, values become greater than those who were obtained at Los Azufres; thence, it was used an another unit: tracks per square centimeter per minute (t/cm^2min). Figure 4 presents an histogram with values of 33 detectors --being the rest lost during etching--; lowest value was 0.2 t/cm^2min, highest one was 8.9; average was 3.3 and standard deviation was 1.8. Thus, those values higher than 4.1 t/cm^2min were considered as anomalous one (Gutiérrez-Negrín and López-Martínez, 1984).

Likewise at Los Azufres, an interpolation with values for each sampling point was made, and isovalues of t/cm^2min curves --equivalent to superficial concentration of Radon-- were drawn. Figure 6 shows these curves and emphasizes three anomalous zones that have values greater than 5 t/cm^2min. These anomalous zones coincide with low resistivity anomalies obtained by an electrical survey at Las Tres Vírgenes. Thence, that zones are most attractive for exploration drilling.

CONCLUSIONS

1. The results obtained by measurements of Radon concentrations at Los Azufres and Las Tres Vírgenes geothermal sites, both in México, show that anomalous concentrations of that gas --which

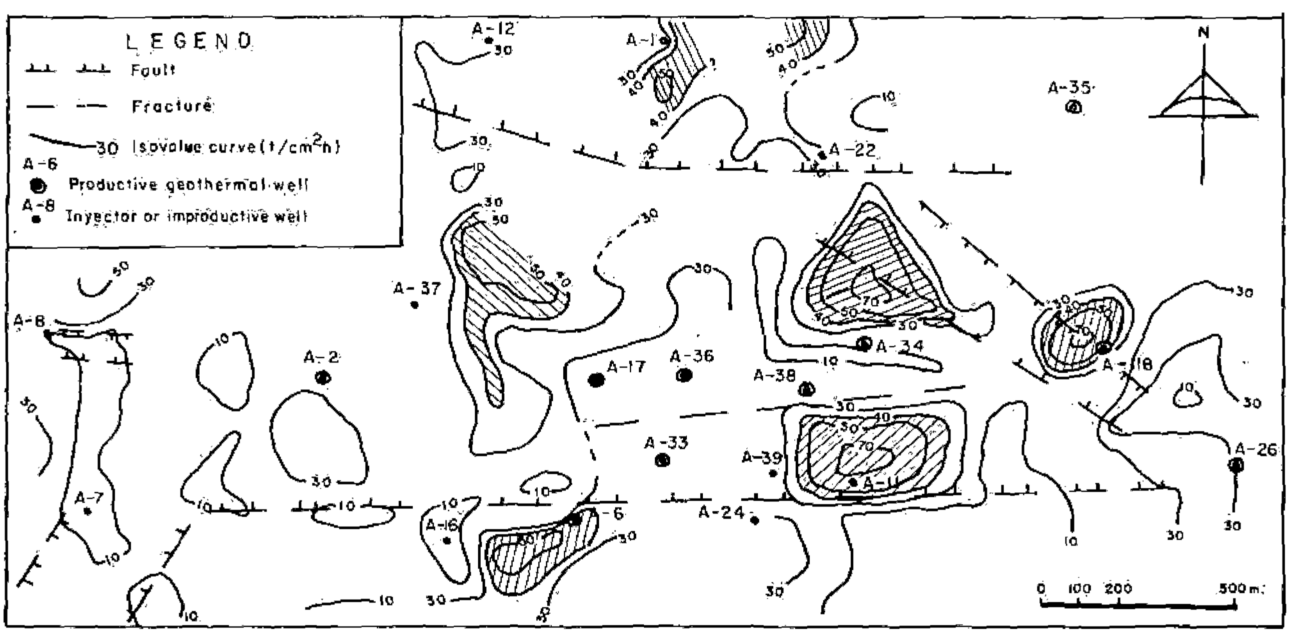

Fig. 5.- Map showing high Radon concentration zones in The Los Azufres geothermal field.

GEOTHERMAL RESOURCES IN THE WILLISTON BASIN: NORTH DAKOTA

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ABSTRACT

Temperatures in four geothermal aquifers in the Williston Basin are in the range for low and moderate temperature geothermal resources within an area of about 128,000 km^2 in North Dakota. The accessible resource base is 13,500 x 10^{18} J., which, assuming a recovery factor of 0.001, may represent a greater quantity of recoverable energy than is present in the basin in the form of petroleum. A synthesis of heat flow, thermal conductivity, and stratigraphic data was found to be significantly more accurate in determining formation temperatures than is the use of linear temperature gradients derived from bottom hole temperature data. The thermal structure of the Williston Basin is determined by the thermal conductivities of four principal lithologies: Tertiary silts and sands (1.6 W/m/K), Mesozoic shales (1.2 W/m/K), Paleozoic limestones (3.0 W/m/K), and Paleozoic dolomites (4.0 W/m/K). The stratigraphic placement of these lithologies leads to a complex, multi-component geothermal gradient which precludes use of any single-component gradient for accurate determination of subsurface temperatures.

INTRODUCTION

Geothermal resources in the Williston Basin in North Dakota occur as thermal waters in at least four regional aquifers, i.e., the Inyan Kara (Cretaceous), Madison (Mississippian), Duperow (Devonian), and Red River (Ordovician). These resources are classified as either moderate temperature resources (150° > T > 90°) or low temperature resources (T < 90°) (Muffler, 1979). Any assessment of these resources must establish the temperature, areal extent, thickness, chemical properties, and hydrologic properties of the aquifers. Previous work by Harris et. al., (1980, 1981, 1983) provides information on areal extent, thickness, and water chemistry as well as temperature data recorded in shallow wells, a few heat flow holes, and a large amount of data recorded as bottom hole temperatures (BHT) in oil and gas exploration wells. The temperature data of Harris et. al., (1983) that are relevant to the thermal aquifers are given as linear temperature gradients calculated from the BHT and mean annual surface temperatures. Those data were used in an analysis of low-temperature geothermal resources in the United States by the U.S. Geological Survey (Sorey et. al., 1983a); and geothermal resources in North Dakota were estimated for two aquifers, the Madison and the Inyan Kara as 7.5 x 10^{18} J. and 2.3 x 10^{18} J., respectively.

Sorey et al.'s (1983a) estimate of geothermal resources suggests a major new energy resource for North Dakota. However, the BHT data used in the resource estimates gave incorrect predictions of subsurface temperatures and the resource was underestimated by about 50 percent. A fundamental problem was that a two-point temperature gradient calculation is inappropriate for the Williston Basin because there are large differences in thermal conductivity among the four principal rock types in the sedimentary section. These rock types and their estimated average conductivities in S.I. units (W/m/K) are: Tertiary clays, silts, and sands, K = 1.6; Cretaceous shales, K = 1.2; Upper Paleozoic limestones, K = 3.2; and Lower Paleozoic dolomites, K = 4.0. Consequently, a typical temperature-depth curve for the Williston Basin is a multi-component curve with slopes differing by as much as a factor of four. Each of the four rock types has a thickness on the order of a kilometer in parts of the basin. A linear temperature gradient based on accurate BHT data from any unit within the basin will give an inaccurate prediction of temperature in any other unit (Figure 1).

Because the thermal structure of the Williston Basin is complex and cannot be represented by linear temperature gradient calculations, the first goal of this project has been to determine accurately the temperatures of the thermal aquifers in the basin. The ultimate goal of this project has been to reassess the resource in the Inyan Kara and Madison aquifers and to extend the resource analysis to include the Duperow and Red River aquifers.

SUBSURFACE TEMPERATURES

Accurate determination of subsurface temperatures should be the first objective in assessing geothermal resources in sedimentary basins. The methods for determining those temperatures have

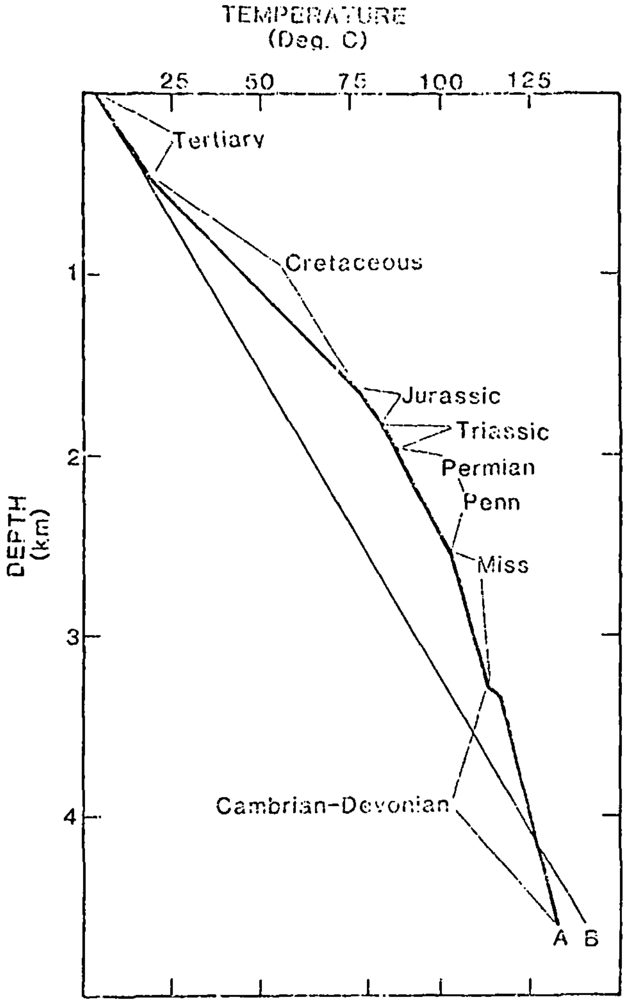

FIGURE 1 - Hypothetical temperature-depth curves for the Williston Basin in western North Dakota. Curve A was computed from heat flow and stratigraphic data. Curve B was taken from bottom hole temperature data.

differed among the various DOE State-Coupled Geothermal Resource Assessment Programs, and the most commonly used method has been to compile and analyze the bottom hole temperature data from oil and gas wells. Other methods that have been used are direct measurement in deep wells and prediction of temperatures from heat flow data. Because the basic quantity sought in exploration for geothermal resources is heat, establishing the most accurate method for determining subsurface temperatures is crucial for geothermal research.

The accuracy of bottom hole temperatures as predictors of subsurface temperatures was questioned in the introduction. In that discussion it was assumed that BHT data accurately represent the temperatures of the formations in which they were recorded. Tests of that assumption are available from studies where other methods as well as analysis of BHT data were used to determine subsurface temperatures. For example, Gosnold (1982) compared data derived from the geothermal gradient map of North America (A.A.P.G., 1976) and equilibrium temperature data in Nebraska. The temperature gradients differ on the order of 10°C/km to 40°C/km and the temperatures differ by about 20°C over the study area. In this case the equilibrium temperatures are categorically higher than the temperatures extrapolated from the BHT data.

The differences between the temperature data sets are due to the data and to the correlation applied to the data. The quality of the data in the oil fields in Nebraska is not good. Analysis of bottom hole temperatures recorded in nine different sections in western Nebraska shows that, in some cases, about 20 percent of the temperatures have the same value regardless of depth or time interval between cessation of mud circulation and logging (Gosnold, Eversoll, and Carlson, 1982). In these cases, it is suspected that the BHT is a guess by the logger rather than an actual record. The time of logging is also suspect in most of the data. In a total of 14,000 records, there are fewer than 100 instances in which recorded logging times are not exactly 1 or 2 hours after circulation ceased. The problem with the correction to the BHT data is that it was based on equilibrium temperatures recorded in wells in the Texas Gulf Coast region. The gross lithologies and the thermal properties of the sediments there are not the same as those in the Cretaceous rocks underlying Nebraska. Consequently, the constants in the correction equation (see Wallace et. al., 1979) do not apply to the rocks in Nebraska.

Uncorrected bottom hole temperatures are, as expected, less close to the equilibrium temperature data than the corrected data. This condition also was demonstrated in the Nebraska project where one of the tasks was to produce a contour map of temperature gradients calculated from uncorrected bottom hole temperature data. That map (Gosnold, 1982) is based on about 14,000 data and vaguely resembles the A.A.P.G. temperature gradient map, but it shows little agreement with the equilibrium temperature gradient map.

The Denver Basin in Nebraska has a multi-component geothermal gradient curve similar to that in the Williston Basin. The geothermal gradient in the shale-rich Cretaceous section is about 50 K/km due to the low thermal conductivity of the shales, i.e. about 1.2 W/m/K (Sass et. al., 1982; Blackwell et. al., 1981). The gradient in the Paleozoic carbonate section ranges from one-third to one-half of that in the

Mesozoic rocks due to the high conductivity of the limestones and dolomites, i.e., about 3.0 W/m/K to 4.5 W/m/K (see Sass et. al., 1981). However, for much of the Denver Basin the BHT data are based on temperatures recorded in the Dakota Group and only one component of the temperature gradient curve influences the data. This observation is most significant. In this case, a two-point temperature gradient curve should apply, yet large differences between equilibrium temperatures and BHT data exist. Therefore, BHT data may not accurately represent formation temperatures even for the case of one-component geothermal gradient areas, and use of BHT data in cases where multi-component gradients do influence the data seems wholly inadvisable.

An alternate method for determining subsurface temperatures is to use a synthesis of heat flow, thermal conductivity, and stratigraphic data. This method is a direct approach to determining subsurface temperatures because it addresses the fundamental variables in the thermal structure of the crust, i.e., heat flow and thermal conductivity. This method was used in the geothermal resource assessment of Nebraska (Gosnold and Eversoll, 1981; 1982) and its accuracy proved to be excellent. Subsequent measurement of temperatures in nine wells at depths ranging from 1.2 km to 1.8 km in the Denver Basin have found actual temperatures to be within 2 degrees of the predicted temperatures.

WILLISTON BASIN

At least four geothermal aquifers lie within the Williston Basin. Accurate determination of their temperatures was the first objective in assessing the total geothermal resource. Because of its better accuracy, the heat flow-stratigraphy synthesis method for determining subsurface temperatures was used in this analysis of the Williston Basin. Consequently, one of the significant results of this study is that it provides another comparison between the BHT and heat flow synthesis methods for assessing geothermal resources.

The data for the Williston Basin include heat flow data from previous studies (Blackwell, 1969; Combs and Simmons, 1973; Scattolini, 1978) and stratigraphic data summarized in the previous geothermal studies in North Dakota (Harris et. al., 1982). Thermal conductivities of rocks at heat flow sites were used as a basis for estimating regional conductivities for gross lithologies. Although thermal conductivity of a specific unit may differ from site to site, the range of variation for one rock type is small compared to the difference in conductivities for different rock types characteristic of the Williston Basin. For example, the range in conductivity for the Paleozoic shales in Kansas is about 0.3 W/m/K (Blackwell et al., 1982), the difference in conductivity between the Pierre shale and the Madison limestone is about 2.5 W/m/K. A constraint on the range of thermal conductivities used is obtained by comparing the predicted temperature-depth plot with the actual temperature logs taken at nearby sites (Figure 2).

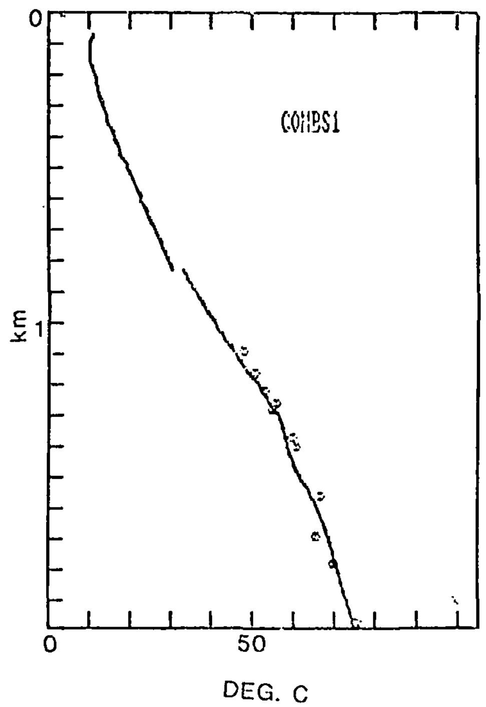

FIGURE 2 - Comparison of an equilibrium temperature-depth log (small dots) with hypothetical temperatures calculated from heat flow (large dots).

In the application of this method in the Nebraska study, stratigraphic data were taken from electric logs for a number of sites within the resource area. However, in this study the data were taken from a series of structure contour maps of the principal rock formations in the Williston Basin (Harris et. al. 1982). These maps permitted establishment of a regularly spaced grid of points for subsurface temperature computations.

Selection of the grid spacing was determined from the spacing of available heat flow data, which is the quantity most likely to vary from site to site. The nature of the temperature field arising from a radioactive basement source is essentially the same as that of a gravitational field arising from different density distributions in the basement (Simmons, 1967). The simple half-width rules and depth rules that apply to gravity data also apply to temperature data, and it is reasonable to assume that lateral variation in heat flow due to differences in basement radioactivity should have its shortest wave lengths on the order of the thickness of the sedimentary cover. For the Williston Basin the ideal spacing of heat flow data would be on the order of 4 kilometers. The actual spacing of data from previous studies (see Scattolini, 1977) ranges from 10 to greater than 100 km and is commonly about 40 kilometers. To form a grid for temperature projections, speculative interpolation of the data is necessary. However, extrapolation of these widely spaced data to a dense grid of 4 kilometers is unjustified; and the least speculative extrapolation seems to be a grid spacing of about 40 kilometers. For the purpose of portrayal on available maps, a spacing corresponding to 4 townships, i.e., 24 miles (38.6 km) was adopted.

RESOURCE ESTIMATES

Temperatures on top of each of the aquifers were projected for each point in the 9x10 grid using the simple equation for one dimensional heat flow

$$Q = K(dT/dZ) \qquad \text{(Eq. 1)}$$

where Q is heat flow, K is thermal conductivity, dT is the incremental change in temperature for an incremental change in depth of dZ. The temperature at any point Z can be calculated by

$$T = To + Zi(Q/Ki) \qquad \text{(Eq. 2)}$$

where To is surface temperature, Zi and Ki are the thicknesses and thermal conductivities of the n overlying layers.

Estimates of the mean accessible resource base were obtained using the method of Sorey et. al. (1983b), i.e.,

$$qR = \rho\, c\, a\, d\ (t - tr) \qquad \text{(Eq. 3)}$$

where qR is the accessible resource base, ρc is the volumetric specific heat of the rock plus water, a is the reservoir area, d is the reservoir thickness, t is the reservoir temperature, and tr is 15°C. This method gives an optomistic estimate for the resource base because of the large temperature drop that is used. However, each use of geothermal waters may require different amounts of heat extraction, and heat exchanger characteristics vary widely among different types and in different applications. Therefore, it may be better to specify a specific reference temperature for purpose of resource estimation and let the potential user make additional estimates based on the data and his particular needs.

The recoverable resource can be calculated from the accessible resource base by considering the hydrologic properties of the aquifers. The general approach of Sorey et. al. (1983b) could be applied to the different aquifers in the basin using available data on their respective hydrologic properties. However, the general conclusion reached by Sorey et. al. (1983b), i.e., that the recovery factor for large sedimentary basis approaches 0.001, serves as a convenient method for making the resource estimate. In fact, applying this recovery factor to the Williston Basin data gives lower estimates for the resource than were obtained by Sorey et. al. (1983a). (See Table 7, pg. 59).

Applying this recovery factor to the data obtained in this study gives estimates for the resource that exceed the estimate of Sorey et. al. (1983a) by about 107 percent for the Inyan Kara and 25 percent for the Madison. The difference for the Madison is due only to the temperature differences used in the calculations. The difference for the Inyan Kara is due to temperature differences and to the size of the area included in the estimate. The extent of the resource area can be calculated by applying the criterion of Reed (1983), i.e., that a resource must have a temperature exceeding Tr, where

$$Tr = T10 + Z(25) \qquad \text{(Eq. 4)}$$

T10 is mean annual surface temperature plus 10°C and Z is depth to resource. The Inyan Kara underlies Cretaceous shales that have a thermal conductivity on the order of 1.2 W/m/K, assuming that the mean heat flow in the basin is 55 mW/m^2 the minimum depth at which the Inyan Kara becomes a resource can be calculated by setting Equation 2 equal to Equation 4 and solving for Z. For the conditions given above, this depth is 720 meters.

CONCLUSIONS

Methodology

The method of estimating subsurface temperatures used in this study is significantly more accurate than is the use of BHT data. Application of the heat flow synthesis method in this study relied on the assumption that thermal conductivities do not vary over the study area. This assumption is not entirely correct. Formation conductivities do vary throughout the basin, but the variation is significantly less than the differences in conductivities between formations. Consequently, errors in calculated subsurface temperature due to variation in formation conductivities are significantly less than errors that result from applying linear gradients extrapolated from BHT data.

The heat flow synthesis method would be best applied where actual conductivities are measured at each grid point. In most sedimentary basins this condition can be met. Most state geological surveys maintain drill core repositories or libraries and numerous samples are available for thermal conductivity analyses. It is suggested that a cooperative effort between the state geological surveys and the geothermal laboratories at several universities and the U.S.G.S. could lead to accurate temperature analyses of most sedimentary basins. It is recommended that this type of project be a major component of any future national geothermal program.

Resources

This assessment of geothermal resources in the Inyan Kara, Madison, Duperow, and Red River aquifers places the accessible resource base in North Dakota at 13,500 x 10^{18} J (Table 1).

TABLE 1

	Mean Temperature	Maximum Temp. °C	Minimum Temp. °C	Reservoir Area (Km^2)	Reservoir Thickness (Km)	Mean Accessible Resource Base (10^{18} J.)
Inyan Kara	51	84	25	128,000	0.091	1,100
Madison	69	117	31	128,000	0.366	6,600
Duperow	81	127	34	126,000	0.100	2,200
Red River	87	138	35	128,000	0.150	3,600

Assuming an estimated recovery factor of 0.001 for geothermal waters and that a barrel of petroleum contains 6.07 x 10^9 J., the recoverable geothermal resource contained within four aquifers in North Dakota is equivalent to the energy contained in 2.22 x 10^9 barrels of petroleum. A surprising result of this study is that the quantity of geothermal energy in the Williston Basin may exceed the energy that is present in the form of oil. The potential impact of this energy resource on the industrial climate of North Dakota should be explored, [in depth]

Technology for utilization of the geothermal resource directly as a heat source and for electric power generation with binary systems has developed to an economical stage. When exploited using both production and re-injection wells, this large energy resource is almost non-depletable and is non-polluting. Some possible uses for the resource are: electric power supply, direct heating supply, lignite drying, grain drying, electric rail systems, vegetable crops in geothermally heated green houses, and fish farming.

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GEOCHEMISTRY AND GEOTHERMOMETRY OF THE DESERT HOT SPRINGS GEOTHERMAL RESOURCE AREA

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ABSTRACT

The Desert Hot Springs Geothermal Resource Area (GRA) is about 14.5 km (9 miles) north of the City of Palm Springs, California. The northwesterly-trending Mission Creek fault borders the GRA on the southwest. Geothermal water is produced from the alluvial deposits underlying the GRA.

Chemical analyses of water from 22 wells throughout the GRA indicate the geothermally heated water north of the Mission Creek fault is high in sodium and sulfate, differing from the water sampled south of the fault, which is high in calcium and bicarbonate.

The results of the study indicate that meteoric water, originating in the San Bernardino Mountains, flows southeasterly toward the GRA along the Mission Creek fault. Geothermometry indicates that the water is heated to temperatures as high as 110°C at depths between 2.7 km and 3.0 km. The geothermal water ascends along fractures near the intersections of the subsidiary Blind Canyon and Long Canyon faults. After cresting in the shallow alluvial rock, some geothermal water flows northeasterly and southwesterly. All of the water, however, eventually flows southeasterly along the direction of regional hydraulic gradient.

INTRODUCTION

Desert Hot Springs, California, "The Spa City", is about 14.5 km (9 miles) north of the City of Palm Springs, California (Fig. 1). Desert Hot Springs is partially within the Desert Hot Springs Geothermal Resource Area (GRA) (Fig. 2). Since 1941, about 200 low-temperature geothermal wells (below 100°C or 212°F) have been drilled in the GRA. Today, about 50 of the wells are used commercially for pools and spas. Water produced from some of the wells reaches temperatures of about 90°C (194°F).

PHYSIOGRAPHIC AND GEOLOGIC SETTING

The Desert Hot Springs GRA straddles a major tectonic plate boundary that also marks the division between the Transverse Ranges and the Salton Trough geomorphic provinces. The GRA is bounded by the Mission Creek fault for 10.5 km (6.5 miles) on the southwest and The Little San Bernardino Mountains on the northeast. The GRA is about 2.8 km (1.7 miles) wide, lying between the fault and the mountains at the head of coalescing alluvial fans. The terrain is gently sloping and rocky.

The GRA is underlain by a wedge of Quaternary alluvium that grades downward into the late Pleistocene Ocotillo Conglomerate (Proctor, 1968). The conglomerate, in turn, overlies Precambrian, igneous-metamorphic rocks called the "San Gorgonio Complex" by Proctor. The complex is part of a roof pendant for the Cretaceous, Southern California batholith that regionally forms the core of the Transverse Ranges. Structurally, the Mission Creek fault is the dominant feature, but two subsidiary faults, the Long Canyon fault and the Blind Canyon fault, trend obiquely northward away from two, hot-water centers in the GRA (Fig. 2).

WATER TEMPERATURES

Isothermal contours of produced water for the Desert Hot Springs GRA were drawn using produced-water temperature data for 54 wells (compiled by R. Proctor, 1968) and on produced water-temperature data measured from about 100 additional wells* (Fig. 2).

GEOTHERMOMETRY

Water samples were collected from 22 geothermal wells. The approximate locations of the 22 sampled wells are shown in Figure 2. (Twenty of the wells are within the GRA; 2 city water-wells are southwest of the Mission Creek fault just beyond boundaries of the GRA.) The concentrations of selected chemicals are listed in Table 1.

Selected cation and anion concentrations were plotted on Langelier diagrams for each sample (Fig. 3). The diagrams clearly show that most of the samples taken from geothermal wells on

* Data for these additional wells were from well owners; local well drillers; and city, county, and state records. The data were field checked randomly. Ambiguous data were not used. Locations for the wells were plotted and measured, and the water temperatures recorded. Due to the confidential nature of some well data, well locations and temperature data were omitted from the isothermal contour map.

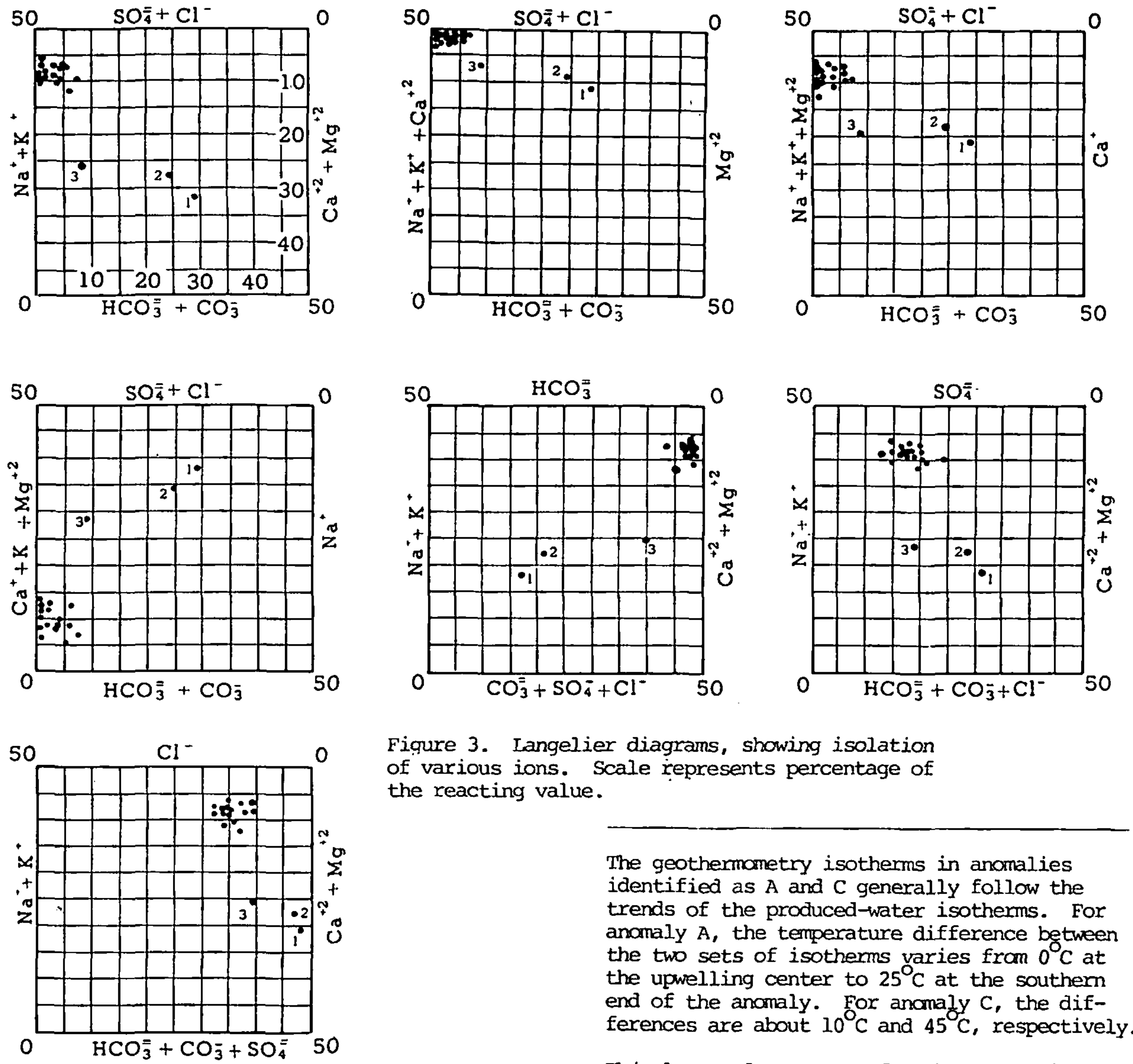

Figure 3. Langelier diagrams, showing isolation of various ions. Scale represents percentage of the reacting value.

COMBINED ISOTHERM MAP

Geothermometry isotherms were superimposed on the isotherms of produced water, yielding a map that is useful for determining areas of upwelling and directions of fluid flow (Fig. 2).

The areas of upwelling are assumed to be the areas where highest produced-water temperatures coincide with highest geothermometry temperatures. The initial direction of fluid flow away from the center of upwelling often is in the direction of the produced-water isotherms radiating from the center. In some areas, large temperature differences between the two sets of isotherms are an indication that geothermal water flows laterally through these areas and is cooled faster than the geothermometry indicators can equilibrate. This condition appears to arise downslope from the upwelling centers.

The geothermometry isotherms in anomalies identified as A and C generally follow the trends of the produced-water isotherms. For anomaly A, the temperature difference between the two sets of isotherms varies from 0°C at the upwelling center to 25°C at the southern end of the anomaly. For anomaly C, the differences are about 10°C and 45°C, respectively.

The elongated patterns of both sets of isotherms, combined with the large temperature difference to the southeast, indicate that the flow from both anomalies is primarily to the southeast -- a direction consistent with the regional hydraulic gradient.

The flow path for anomaly B is not as easy to interpret. The small area with the 80°C produced-water isotherm is assumed to be the center of upwelling. The pattern of the produced-water isotherms suggests that a large component of geothermal water initially flows northeasterly and southwesterly for about 1 km through channelways (fractures) that bisect the lobes of the isotherms in these directions.

The water flowing from anomaly B is undoubtedly mixed with water flowing southeast from anomaly A, as indicated by the change in direction of the 60°C and 70°C geothermometry isotherms. This mixing of geothermal waters makes it difficult

Table 1. Concentrations in ppm of selected ions and silica oxide from geothermal wells in the Desert Hot Springs GRA.

well number	Ca^{+2}	Mg^{+2}	Na^{+}	K^{+}	$SO_4^{=}$	Cl^{-}	HCO_3^{-}	SiO_2
1	60	19	42	8	130	14	295	19
2	53	16	52	8	150	17	257	17
3	101	22	133	10	394	101	164	21
4	33	1	239	4	400	144	71	15
5	55	1	300	4	606	153	32	21
6	74	4	393	9	660	306	72	15
7	29	1	221	6	385	93	48	19
8	58	3	298	5	460	167	82	26
9	34	1	276	6	460	159	40	45
10	52	5	285	9	502	216	47	32
11	69	1	349	8	445	215	59	23
12	41	1	301	7	450	147	55	28
13	49	1	332	8	465	180	35	34
14	68	2	302	9	465	191	51	45
15	64	11	291	10	500	156	114	21
16	47	0	310	4	536	136	43	-
17	64	2	345	12	510	201	76	-
18	51	5	275	11	564	180	45	56
19	46	5	235	7	428	150	69	18
20	55	2	328	12	565	212	34	59
21	48	6	223	7	464	138	36	13
22	54	1	333	9	548	185	15	47

Table 2. Selected geothermometry equations (Henley, et al., 1984) used in this study. Concentrations of Na, Ca, K, and SiO_2 are in parts per million.

Geothermometer	Equation	Restrictions
a. Quartz-no steam loss	$t^\circ C = \dfrac{1309}{5.19 - \log SiO_2} - 273.15$	t = 0-250°C
b. Quartz-maximum steam loss	$t^\circ C = \dfrac{1522}{5.75 - \log SiO_2} - 273.15$	t = 0-250°C
c. Chalcedony	$t^\circ C = \dfrac{1032}{4.69 - \log SiO_2} - 273.15$	t = 0-250°C
d. Na/K (Fournier)	$t^\circ C = \dfrac{1217}{\log (Na/K) + 1.483} - 273.15$	t > 150°C
e. Na/K (Truesdell)	$t^\circ C = \dfrac{855.6}{\log (Na/K) + 0.8573} - 273.15$	t > 150°C
f. Na-K-Ca	$t^\circ C = \dfrac{1647}{\log (Na/K) + \beta[\log(\sqrt{Ca}/Na) + 2.06] + 2.47} - 273.15$ where $\beta = 1/3$	
g. Na-K-Ca	$t^\circ C = \dfrac{1647}{\log (Na/K) + \beta[\log(\sqrt{Ca}/Na) + 2.06] + 2.47} - 273.15$ where $\beta = 4/3$	

Table 3. Produced-water temperatures measured at the wellhead and calculated subsurface temperatures for geothermometer equations in Table 2. The last column is the average of the two geologically most credible results, those derived from equations (a) and (g).

		Geothermometry Temperatures, °C							
Well number	Measured Produced-Water Temp., °C	SiO_2 (a)	(b)	(c)	Na-K (d)	(e)	Na-K-Ca (f) 1/3	(g) 4/3	Average Temperatures, °C, of (a) & (g)
1	21.7	62	67	29	279	269	180	59	61
2	26.7	57	64	25	257	239	174	64	61
3	30.5	65	71	33	194	159	148	67	66
4	42.2	53	60	21	100	52	102	68	61
5	52.2	65	71	33	89	40	93	60	63
6	43.3	53	60	21	117	70	115	82	68
7	43.3	62	67	29	126	80	120	82	72
8	33.3	74	78	42	100	52	101	65	67
9	43.3	97	98	67	114	67	113	81	89
10	49.0	82	85	51	137	93	128	87	85
11	47.8	69	74	37	117	70	114	78	74
12	60.0	77	81	45	117	70	116	83	80
13	65.5	85	88	54	119	73	118	85	85
14	77.7	97	98	67	131	86	123	80	89
15	30.0	65	71	33	140	96	129	85	75
16	43.3	-	-	-	91	42	96	65	65
17	43.3	-	-	-	141	96	131	93	93
18	38.0	107	107	78	149	106	136	93	100
19	34.0	60	65	27	130	84	121	77	69
20	70.0	110	109	80	144	100	134	96	103
21	35.0	48	55	15	135	58	124	77	63
22	77.0	99	100	69	126	80	122	87	93

to interpret the flow paths northwest and southeast of the anomaly. It is clear, though, that the combined geothermal water must flow primarily down the hydraulic gradient to the southeast since the temperature differences of the two sets of isotherms is greatest in that direction.

CONCLUSIONS

Rain runoff from nearby, high mountains could provide the source of recharge water for the Desert Hot Springs GRA. From the San Bernardino Mountains, the water may pass into a system of deep fractures, part of the North Branch of the San Andreas fault, and flow southeasterly toward the GRA. The water descends to depths of 2.7 to 3.0 km, where it attains a distinct sodium-sulfate chemistry and a maximum temperature of about 110°C (230°F) as suggested by the geothermometry data. The heated water continues to flow into the GRA along the Mission Creek fault and then ascends through the fractured rock near the intersections of the Blind Canyon and Long Canyon faults.

The combined isotherm map (Fig. 2) suggests that the ascending geothermal water crests in the areas labeled A, B, and C. The pattern of produced-water isotherms suggests that within anomaly B, some of the geothermal water flows to the northeast and some to the southwest. The water, however, soon becomes mixed with geothermal water flowing southeasterly from anomaly A. This combined flow is to the southeast, consistent with the regional hydraulic gradient. The geothermal water that upwells in anomaly C also flows to the southeast, consistent with the overall flow direction.

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CHEMISTRY OF LOW-TEMPERATURE GEOTHERMAL WATERS AT KLAMATH FALLS, OREGON

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ABSTRACT

Thermal water at Klamath Falls, Oregon, occurs in a heterogeneous aquifer at depths of 60 to 610 meters over an area of about 5 square kilometers. Waters measuring 70 to 100°C are utilized for district space heating. These thermal waters enter the shallow aquifer through a permeable fault zone on the northeast side of the city, and undergo mixing and cooling as they flow southwestward. Chemical and isotopic analyses of well discharges indicate that in the aquifer mixing occurs between shallow cold groundwater containing 2.0 TU tritium and a deeper tritium-free thermal groundwater at 100 to 120°C. This deeper water apparently results from the mixing of old, tritium-free cold groundwater and deep thermal groundwater at about 190°C and 120 mg/kg Cl. The temperature and chlorinity of the deep thermal water are based on SO_4-isotope and silica geothermometers and chloride and silica mixing models.

INTRODUCTION

The city of Klamath Falls is located east of the Cascade Range in south-central Oregon. This community of about 17,000 persons utilizes a shallow (<610 m depth), low-temperature (<140°C) geothermal resource for a variety of direct-use applications, the primary one being space heating. More than 450 wells tap heat from the thermal aquifer in an urban area of less than five square kilometers (Fig. 1).

For nearly 50 years, privately-owned wells have been supplying heat to individual houses or businesses. Development of a central district heating system began in 1979 with construction of two production wells, an injection well, a heat-exchange facility, pipelines, and heating units. In 1983, the Klamath Falls district initiated an effort to gather data on the geothermal resource and to monitor the aquifer. During the summer of 1983, investigators from Lawrence Berkeley Laboratory, Stanford University, the Oregon Institute of Technology, and the U.S. Geological Survey collaborated in an intensive study of the shallow thermal reservoir to obtain chemical and hydraulic data for evaluation of reservoir performance and geothermal development potential. Data from the aquifer tests are presented in Benson and others (1984) and interpretations of the test results are presented in Sammel (1984).

Thermal waters collected from the Klamath Falls geothermal aquifer the month preceding, and during the 1983 pumping tests were analyzed for chemical and isotopic constituents. In this report we interpret these and earlier analyses as indicators of the temperature and reservoir processes in the geothermal system.

GEOLOGIC SETTING AND RELATED HYDROLOGY

The geology of the Klamath Falls area has been mapped by Peterson and McIntyre (1970), and the geohydrology of the Klamath River drainage basin has been summarized by Sammel (1980). The Klamath Falls region is transitional between the Cascade volcanic chain to the west and the Basin and Range Province to the southeast. Rocks of the area are predominantly Pliocene and Pleistocene volcanics. Most of the thermal wells penetrate fine-grained volcanic sediments and diatomite that were deposited in a Pliocene lake. These sediments are interstratified with thin (1.5 to 6 m) layers of basaltic tuff, scoria, and breccia, and underlain by basaltic and andesitic flow rocks and tephra.

A major NW-trending fault roughly parallels the border of the hot-well area of the city and extends to the shore of Upper Klamath Lake (Fig. 1). This fault is one of several westward-dipping, high-angle normal faults that define the eastern margin of the Klamath graben which contains the Upper and Lower Klamath Lake basins.

The geothermal aquifer at Klamath Falls is not a clearly defined rock unit, but rather a series of stratified lithologic units having extensive vertical and areal variability. Lithologic logs and temperature measurements indicate that thermal water flows in permeable strata generally not more than a few meters thick. These strata include fractured, indurated lacustrine sediment, volcanic breccia, and fractured vesicular basalt flows. Temperature profiles show that hot water (>80°C) occurs less than 61 m below land surface under most of the hot-well area, and that the aquifer is at least 610 m thick.

The potentiometric surface of thermal water (based on data from wells >91 m in depth) dips toward the southwest as does the general topography. Reported maximum temperatures in the thermal aquifer are highest (95 to 120°C) near the vicinity of the major NW-trending fault, and decrease (to <80°C) toward the southwest (Fig. 2). These data suggest that hot water flows upward along the fault, feeds the thermal aquifer in the hot-well area, and then flows laterally toward the south and west. An area of artesian wells extends along the trend of the

Old Fort Road valley, suggesting that artesian pressures and high temperatures are transmitted along a permeable NE-trending fault zone that cuts across the main fault (Fig. 2).

CHEMICAL COMPOSITIONS

Analyses of Klamath Falls thermal and non-thermal waters from Benson and others (1984, Ch. 4), and Sammel (1980) along with new analyses are given in Table 1. Thermal waters from Klamath Falls wells contain (in order of decreasing concentration) SO_4, Na, SiO_2, Cl, HCO_3, Ca, and K with minor F, Li, Mg, and Al. Nonthermal well waters are more dilute and contain (in order of decreasing concentration) HCO_3, SiO_2, Na, Ca, Mg, Cl, K, and SO_4. Analyses of thermal and non-thermal Klamath Falls well waters are compared in Figure 3. Cold spring waters in the vicinity of Klamath Falls contain less Na and Cl than non-thermal well waters (Table 1). Constituents of thermal waters show limited ranges of concentration, with most variation in K, Ca, Mg and SiO_2 (Fig. 3).An increase in SiO_2, Na, K, and Cl concentrations is observed for samples collected during the pumping tests (Table 1). As discussed below, the variation in the chemistry of the thermal waters is apparently caused by mixing with cooler waters of different composition and by equilibration with rock minerals at different temperatures.

ISOTOPIC COMPOSITIONS

Water from Klamath Falls cold wells and springs is isotopically similar to rainwater but shows some effects of evaporation before infiltration. The oxygen-18 and deuterium contents of these waters fall along a local "meteoric water line" (MWL) similar to the global MWL (Craig, 1961), but with a "deuterium excess" of about +6 rather than +10 (Fig. 4). The thermal waters shown in Figure 4 are significantly lower in δD and higher in $\delta^{18}O$ than local cold waters. Concentrations of D and ^{18}O in precipitation worldwide have been observed to decrease with increase in elevation, latitude, and distance inland, and with decrease in temperature (Craig, 1961). Thus the lower deuterium content of the Klamath Falls thermal waters compared to that of the cold waters, suggests that the recharge to the geothermal aquifer occurs at greater elevations than the recharge to the cold aquifer or, much less probably, consists of old waters from a time of colder climate (Buchardt and Fritz, 1980). The higher ^{18}O concentrations in the thermal waters relative to waters on the MWL represents an "oxygen isotope shift" caused by long contact with ^{18}O-rich rock minerals at elevated temperatures. The isotopic (^{18}O, D) variation of the thermal waters results from mixing with local cold groundwater and from infiltration of evaporated surface water (Fig. 4).

The tritium content of a sample from the city's major cold-water supply (well 500) is very low at 0.14 Tritium units (TU). The residence time in the cold aquifer is at least 30 years because this water can have only a very small contribution from high-tritium precipitation (with 30 to 1000 TU) that postdates nuclear bomb testing in the mid-1950s. The tritium in this water may represent prebomb tritium (estimated at 10 TU originally), which has undergone radioactive decay during 6 half lives of 12.3 years indicating that the water is older than 70 years (Gat, 1980). Alternatively it could have orginated in a steady-state, well-mixed reservoir with the same fraction of inflow and outflow each year. In such a reservoir the average age of water with <0.2 TU would be greater than 10,000 years because of the effect of small additions of postbomb precipitation (Pearson and Truesdell, 1978). The second cold well sampled in 1983 (well 501) has higher tritium (0.71 TU), indicating a small addition of more recent precipitation.

The tritium contents of the thermal waters sampled in 1983 range from 0 to 1.0 TU; water from well 304 containing higher tritium is probably contaminated with irrigation water. This well also has lower temperature and chloride than most other wells. Most thermal waters sampled prior to the pumping test have tritium contents near zero (<0.3 TU), indicating greater than 60-year storage as discussed above. Thermal waters show increasing tritium with decreasing temperature, suggesting mixing with younger, more dilute waters (Fig. 5).

MIXING OF THERMAL AND NONTHERMAL WATER

The relations of temperature, chloride, tritium, and other constituents of the thermal waters indicate mixing. A reasonably linear chloride-temperature mixing relation is observed for samples collected prior to the pumping tests (Fig. 6), suggesting that the cold end member is a water at 20°C containing about 10.5 mg/kg Cl. The temperature of cold water at depths of any possible mixing is assumed to be 20°C because of heating by conduction. Recharge of this cold end-member water probably does not originate from modern Klamath Lake water because extrapolated tritium contents at 20°C from Figure 5 are only about 2.0 TU, whereas Klamath Lake had a tritium concentration of 25.7 TU when sampled (Sammel, 1980). Klamath Lake should have higher tritium concentrations than present precipitation because it contains stored older rainwater with higher tritium. (The tritium concentration of modern precipitation is decreasing faster than would be expected from radioactive decay because rainwater is being diluted with deep, tritium-free ocean water.) Klamath Lake also has a higher deuterium concentration (δD = -97.5; Sammel, 1980) than any hot waters, making it an unlikely source of recharge. Samples collected during the pumping tests have chloride concentrations that are, except for one sample, nearly independent of temperature (Fig. 6). These waters may have been out of thermal equilibrium because of more rapid flow in the aquifer. Higher concentrations of SiO_2 relative to Cl (Fig. 7) also indicate that non-equilibrium conditions occurred during the aquifer tests.

If there were a component of the cold end member that was absent from the high-temperature end member, then it would be possible to calculate the composition and temperature of the hot end member. Tritium and magnesium might seem suitable but do not work at Klamath Falls. The mixed thermal waters vary in temperature and chemistry but over a wide mixing range have essentially no tritium or Mg even though the cold waters have both (Table 1). Magnesium seems to be completely removed by mineral reactions at some temperature below 60°C. Well 304 water at 56°C is obviously mixed with cold surface water as indicated by relatively low temperature and Cl (40 mg/kg) and high tritium (9.5 TU) but has only 0.02 mg/kg Mg (Table 1). In the analyses reported here and by Sammel (1980) of waters over 60°C only one has more than 0.1 mg/kg Mg. Cooler spring and well waters below 38°C have increasing Mg with decreasing temperature (2 to >12 mg/kg, of the reliable analyses). There are no waters between 38 and 60°C.

The situation with tritium is no better, with most thermal well waters near the limit of detection (about 0.1 to 0.2 TU) and some cold waters also having little tritium (e.g., well 500 with 0.14 TU). Although a relation can be seen between tritium and temperature (Fig. 5) and chloride and temperature (Fig. 6), there is little trend between tritium and Cl. Probably two cold water sources are involved, one very shallow containing tritium and a deeper one that is tritium free. The shallow mixing that is observed does not define the deep thermal end member and there is no indication that the highest temperature reported (140°C in an unexploitable well; P.J. Lianau, written commun., 1982) is the maximum temperature of the system.

GEOTHERMOMETERS AND MIXING MODELS

Certain chemical and isotopic reactions re-equilibrate sufficiently slowly as fluids cool that evidence of higher temperature equilibria are preserved. These reactions may thus be used as geothermometers and have been calibrated experimentally or empirically to indicate probable maximum temperatures attained. Calculated geothermometer temperatures for Klamath Falls thermal waters are given in Table 2.

In dilute waters, cation geothermometers are likely to be affected by re-equilibration, and at Klamath Falls they show temperatures close to those measured at the sampling point. The average temperature from the Na-K-Ca geothermometer (Fournier and Truesdell, 1973) is 81 ±6°C. Cation geothermometer temperatures of samples taken before the pumping test agree closely with measured temperatures. Samples taken during pumping agree less well because waters chemically equilibrated at other temperatures were rapidly heated or cooled during passage to the wells.

Silica concentrations are greater than expected for saturation with silica minerals (other than amorphous silica) at sampling temperatures and suggest equilibration at higher temperatures, deeper in the reservoir (Fig. 8). Silica in the well waters cannot result from equilibrium with amorphous silica because the waters are undersaturated with this mineral. Direct use of silica geothermometers suggests temperatures of 100 to 150°C (Table 2) but silica concentrations are probably affected by mixing as discussed below.

The sulfate-water isotope geothermometer depends on fractionation of ^{18}O between SO_4 and H_2O, a process that is reasonably rapid at high temperatures but very slow at low temperatures (McKenzie and Truesdell, 1977). At Klamath Falls, this geothermometer is unlikely to be influenced by contamination because the thermal waters have higher SO_4 than cold waters and because no other sulfur-containing material (H_2S, sulfates) are reported in the system. The temperature indicated by using the observed water-^{18}O compositions is 189 ±4°C for thermal waters (Table 2).

At 189°C the half time of sulfate-water ^{18}O equilibration based on experimental rate studies is only 2.4 years, and 97 percent equilibration would be expected in 12.3 years (McKenzie and Truesdell, 1977). If the maximum temperature in the system were 140°C the water would be 97 percent equilibrated in 55 years. Lack of equilibrium from short residence at any temperature below 180°C is not consistent with the tritium concentrations. From the experimental data and experience with this geothermometer in other geothermal systems (Truesdell, 1976) we are confident that the waters have resided at about 180°C long enough to equilibrate and that they have not been at their present temperature (in the shallow thermal aquifer) more than 20 years.

Silica mixing calculations (Truesdell and Fournier, 1977) based on 1983 silica data indicate an average temperature of 185 ±18°C (1 standard deviation of 14 samples with 2 outlying values excluded). Using only data on samples collected during the pumping tests, the average calculated temperature is 192 ±11°C, but this temperature may be high because of lack of equilibrium. Silica concentrations previously reported from wells sampled in this study produced a wider range of mixing-model temperatures (148 to 180°C) and led to a lower estimate of reservoir temperatures (Sammel, 1980). Not all previous samples were properly treated to preserve silica and the recent analyses are probably more reliable. The estimate of 185°C is consistent with the sulfate isotope temperature of 189°C. Mixing temperatures based on chalcedony saturation are about 30°C lower. These are considered less likely because chalcedony is metastable and in the presence of water should convert completely to quartz given the minimum time and temperature indicated for the Klamath Falls thermal aquifer. Other sources of silica (feldspars, etc.) would also rapidly alter in part to quartz and would not control silica concentrations. Temperatures based on chalcedony or other silica sources do not agree with SO_4-isotope temperatures.

If equilibration with quartz is assumed at a temperature of 185°C (Fournier and Potter, 1982), and if the cold and mixed waters contain 45 and 120 mg/kg SiO_2 respectively, then the fraction of high-temperature water in the reservoir mixture is calculated to be about 44 percent. Using 185°C as the temperature of the hot-water end member in a chloride-temperature mixing model, and assuming the cold water to contain 10.5 mg/kg Cl at 20°C and the mixed water to contain 55 mg/kg Cl, on the basis of 1983 samples, the chloride concentration of the hot -water end member is calculated to be about 120 mg/kg. Applying the mixing fraction (44 percent) and assuming no oxygen shift for the cold end member, we calculate that the hot-water end member may have a $\delta^{18}O$ value about -13.5 and a δD value near -129 (Fig. 4).

SUMMARY

From considerations of mixing, from geothermometry, and from tritium analyses we can form a conceptual model of the geothermal system at Klamath Falls. Wells sampled appear to draw water from a shallow thermal aquifer at 70 to 100°C where hot water at 100 to 120°C with zero tritium mixes with a cold water at about 20°C with 10.5 mg/kg Cl and 2.0 TU tritium. Different mixing ratios in the aquifer result in well waters of different temperatures and compositions. The 100 to 120°C hot water may be derived by upflow from a deeper high temperature zone where mixing of older, tritium-free cold and hot waters occurs. Although the indicated high-temperature end-member water has not been encountered by wells drilled thus far, the geochemical relations indicate temperatures of 150 to 190°C somewhere in the system.

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Table 1. Chemical and isotopic analyses of water from wells and springs at Klamath Falls, Oregon. (Concentrations in mg/kg; stable isotope data in permil relative to SMOW.)

Sample	Date	T °C	SiO_2	Na	K	Li	Ca	Mg	Al	Cl	F	SO_4	HCO_3	δD	$\delta^{18}O$	Tritium TU	$\delta^{18}O$ SO_4
Hot water wells sampled the month prior to 1983 pumping tests[1]																	
25	6/04/83	78	98.0	206	5.08	--	31.8	0.02	0.027	51.7	--	398	45.2	-119.4	-14.78	0.23±0.08	-5.34
45	6/09/83	88	98.6	207	5.47	--	24.7	0.05	0.027	48.7	--	425	46.4	-119.4	-14.50	0.33±0.08	-5.18
45A	6/09/83	80	172.0	220	6.65	--	24.8	0.12	0.135	49.4	--	421	47.6	-120.2	-14.69	0.04±0.08	-5.37
65	6/14/83	66	83.0	194	4.30	--	29.8	0.05	0.027	47.7	--	382	50.0	-119.8	-14.54	1.03±0.09	-5.12
76	6/07/83	78	90.7	189	5.08	--	22.4	0.02	0.027	41.9	--	362	46.4	-119.4	-14.72	0.25±0.08	-5.44
110	6/21/83	91	98.6	213	5.47	--	26.3	0.02	0.027	52.8	--	423	46.4	-121.3	-14.45	0.52±0.09	-5.43
216	6/29/83	72	123.2	217	5.84	0.31	25.0	0.04	0.022	51.7	1.2	409	43.3	-121.3	-14.53	0.56±0.09	-5.36
259	6/08/83	56	89.7	205	5.08	--	28.9	0.05	0.027	51.7	--	--	45.8	-120.3	-14.61	0.21±0.08	-5.21
304	6/08/83	56	80.2	175	3.91	--	16.7	0.02	0.027	40.1	--	293	86.0	-116.1	-14.05	8.35±0.30	-4.56
Cold water wells[1]																	
500	6/20/83	21[b]	58.0	23.5	2.74	--	10.2	4.37	0.027	5.7	--	2.2	123	-112.3	-14.86	0.14±0.06	--
501	6/23/83	17	50.1	20.0	3.13	--	15.8	7.73	0.027	7.2	--	1.8	136	-108.8	-14.26	0.71±0.09[a]	--

Table 1. Chemical and isotopic analyses of water from wells and springs at Klamath Falls, Oregon (Continued)

Sample	Date	T °C	SiO_2	Na	K	Li	Ca	Mg	Al	Cl	F	SO_4	HCO_3	δD	$\delta^{18}O$	Tritium TU	$\delta^{18}O$ SO_4
Hot water wells sampled during 1983 pumping tests[1]																	
25	8/10/83	82	126.0	216	5.94	0.28	28.2	0.06	0.025	51.7	1.3	408	43.9	-120.2	-14.54	0.45±0.09	-5.40
45	8/15/83	84	120.7	225	6.52	0.31	28.5	0.09	0.030	51.1	1.3	428	43.9	-122.0	-14.53	0.55±0.10	-5.08
216	8/16/83	71	109.3	214	5.89	0.30	23.5	0.03	0.019	53.4	1.3	407	45.1	-119.0	-14.48	0.59±0.09	-5.49
272	7/08/83	95[c]	119.8	230	7.44	0.38	34.8	0.05	0.030	55.0	1.3	419	41.4	-121.4	-14.63	0.12±0.08	-5.07
272	8/09/83	98[c]	130.1	228	7.17	0.38	36.1	0.03	0.025	53.6	1.3	441	40.9	-122.1	-14.41	0.20±0.11	-5.62
272	8/24/83	--[c]	120.7	228	7.52	0.36	36.1	0.05	0.026	53.4	1.4	450	40.3	-120.8	-14.56	0.20±0.09	-5.14
304	8/17/83	47	97.3	177	4.59	0.28	20.5	0.05	0.010	38.4	1.2	294	90.3	-115.7	-13.96	9.55±0.33	-4.66
Earlier analyses of hot water wells[2]																	
OIT #6	8/24/72	88	--	195	3.9	--	24.2	<.1	--	58	1.45	400	44	--	--	--	--
"	3/31/75	79	90	--	--	0.12	--	--	--	--	--	--	--	-118.7	-14.4	1.6±0.5	-5.45
A. of God	8/27/76	87	110	213	4.6	--	28	0.1	--	54	--	393	61	-120.0	-14.75	--	--
Medo-Bell	1/24/55	81	81	213	4.2	--	23	0	--	54	1.2	403	48	-119.4	-14.6	--	-5.45
Friesen	2/19/55	73	87	221	4.4	--	25	0	--	56	1.6	431	48	--	--	--	--
Serruys	12/22/54	71	83	207	3.8	--	22	0	--	50	1.4	393	51	--	--	--	--
Earlier analyses of cold waters[2]																	
Ore. Water Corp Well	9/--/71	14	27	22	--	--	14.4	9.8	--	4.5	0.01	1.2	--	-109.4	-14.4	2.4±0.5	--
Sharp Sp	6/06/74	18	43	19	4.0	--	13	6.6	--	2.2	0	10	112	--	--	--	--
Shell Rk Sp	8/05/72	12	60[d]	5.9	1.5	--	8.1	3.9	--	<1	0.1	<2	50	-113.1	-15.35	--	--
Hummingbird	7/26/72	12	42[d]	7.0	1.2	--	10.6	8.2	--	1	0.1	<2	96	--	--	--	--
Barkley Sp	7/26/72	11	18	8.0	1.2	<0.02[d]	9.2	6.5	--	1	0.1	<2	84	-115.0	-15.11	--	--
Neubert Sp	8/05/72	17	48[d]	7.6	1.3	--	12.1	7.2	--	1	0.11	<2	94	--	--	--	--

[1]Benson and others (1984), Ch. 4, tables 4-2, 4-3. New Ca data from A. White and A. Yee, Lawrence Berkeley Laboratory, Berkeley, Calif.; new Cl data from J. Consul and D. Vivit, Analytical Chemistry Branch, U.S. Geological Survey, Menlo Park, Calif. Tritium analyzes under contract by the Univ. of Miami RSMAS Tritium Laboratory (H.G. Ostlund, director), Miami, Flor.

[a]Corrected transposition error of value previously reported.

[b]City of Klamath Falls reports a constant temperature of 18°C.

[c]Wellhead temperature reported to have remained constant at 100°C for entire pumping test.

[2]Sammel (1980), tables 5, 8.

[d]Analysis of water from same location collected Aug. 1975.

Table 2. Calculated reservoir temperatures (°C)

Sample	Date	Measured Temp	Qtz[1]	Chalc.[2]	SiO_2 Amorph[3]	Na/K[3]	Na-K-Ca[4]	Qtz mixing[5]	Chalc. mixing[5]	SO_4-H_2O[6] $\delta^{18}O$	X(hot)[7]
Hot water wells sampled the month prior to 1983 pumping tests											
25	6/04/83	78	136	108	16	74	78	175	138	186	0.38
45	6/09/83	88	137	108	16	78	82	165	156	189	0.48
45A	6/09/83	80	170	146	47	87	89	271	--	188	0.25
65	6/14/83	66	127	98	8	68	70	168	132	187	0.32
76	6/07/83	78	132	103	12	79	81	165	126	189	0.41
110	6/21/83	91	137	108	16	76	81	162	123	194	0.51
216	6/29/83	72	149	123	28	79	85	221	200	191	0.26
259	6/08/83	56	131	103	12	74	76	201	173	187	0.20
304	6/08/83	56	125	96	6	68	78	180	148	186	0.23
Hot water wells sampled during 1983 pumping tests											
25	8/10/83	82	151	124	29	81	83	206	178	192	0.34
45	8/15/83	84	148	121	27	84	86	196	165	186	0.37
216	8/16/83	71	142	115	21	81	86	203	173	194	0.28
272	7/08/83	95	148	121	26	91	86	182	147	184	0.47
272	8/09/83	98	153	126	31	89	84	189	156	198	0.47
272	8/24/83	*	148	121	27	93	86	178	142	187	0.52
304	8/17/83	47	136	107	16	77	79	256	--	189	0.12
Earlier analyses of hot water wells											
OIT #6	3/31/75	79	132	103	12	62	71	154	123	195	0.51
A. of God	8/27/76	87	143	115	22	66	74	180	145	--	0.43
Medo-Bell	1/24/55	81	126	96	7	61	75	148	109	191	0.48
Friesen	2/19/55	73	130	101	10	61	76	165	127	--	0.37
Serruys	12/22/54	71	127	98	8	57	73	161	123	--	0.37

*100°C assumed for calculations.

[1]Fournier and Potter (1982), with conductive cooling.

[2]Fournier and Rowe (1966), and Fournier (1977), with conductive cooling.

[3]White (1970).

[4]Fournier and Truesdell (1973), β = 4/3.

[5]Mixing model based on quartz or chalcedony equilibria and a cold-water component having a SiO_2 concentration of 45 mg/kg and a temperature of 20°C (Truesdell and Fournier, 1977; Fournier and Potter, 1982).

[6]McKenzie and Truesdell (1977), with conductive cooling.

[7]Fraction of hot end-member water in the SiO_2 mixing model.

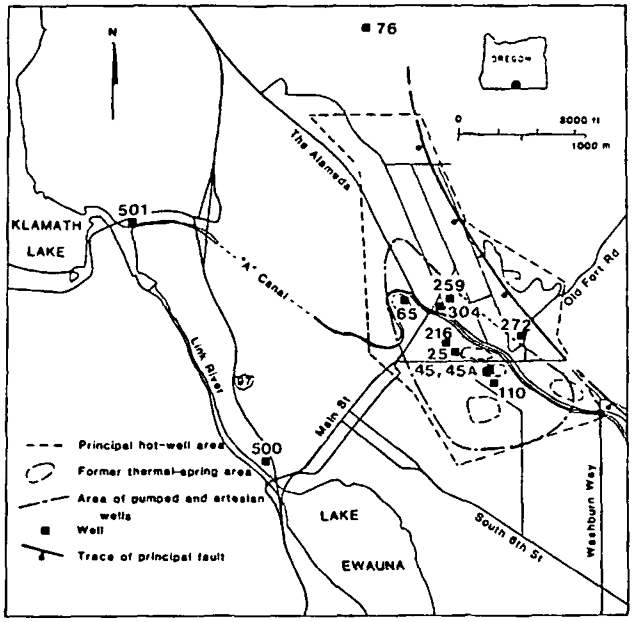

Figure 1. Map of Klamath Falls, Oregon, showing location of hot-well area, trace of the principal fault, and locations (numbered as in Tables 1 and 2) of wells sampled in 1983.

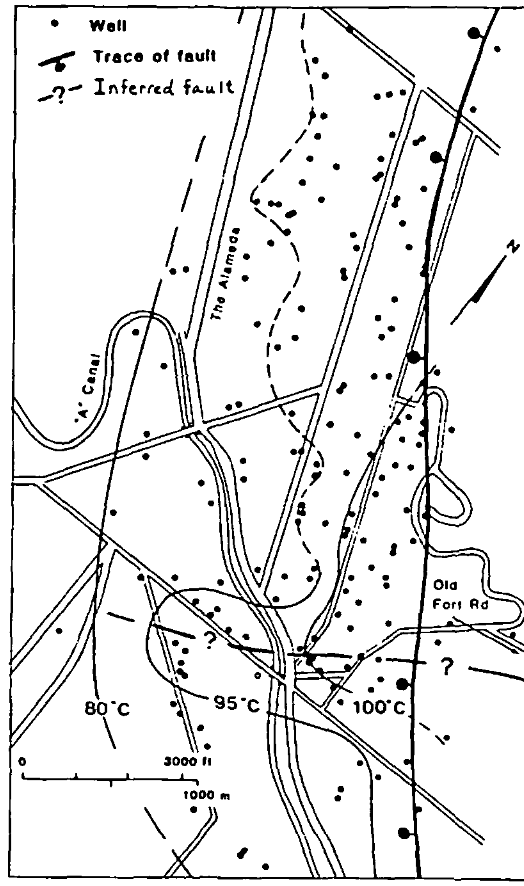

Figure 2. Isotherms based on reported maximum temperatures measured in wells or in well discharges. Most wells within each region have reported temperatures in the ranges indicated.

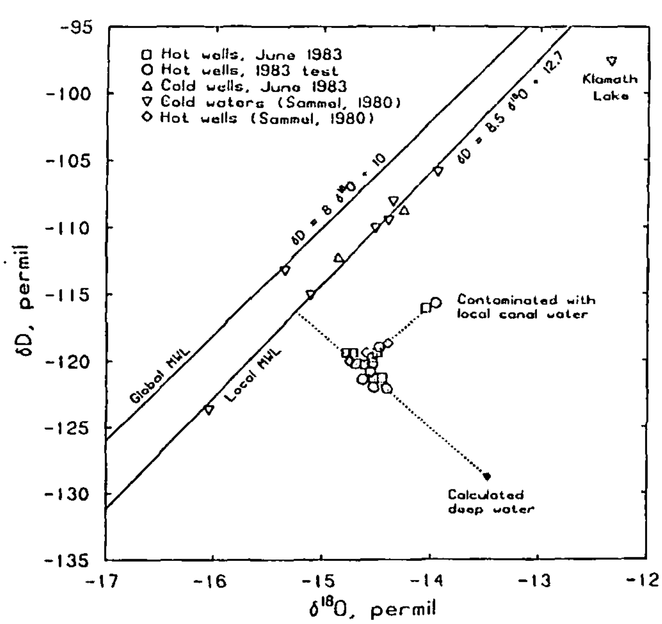

Figure 4. Hydrogen and oxygen isotope compositions of thermal and nonthermal waters, and calculated composition of deep reservoir water.

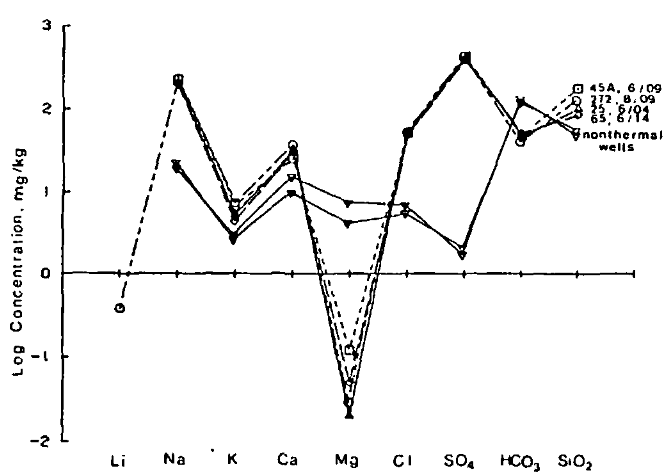

Figure 3. Schoeller plot showing comparison of representative hot and cold well-water compositions.

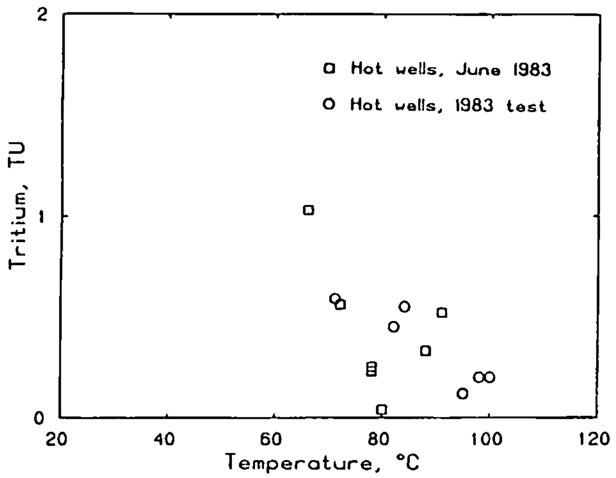

Figure 5. Tritium versus temperature of hot waters.

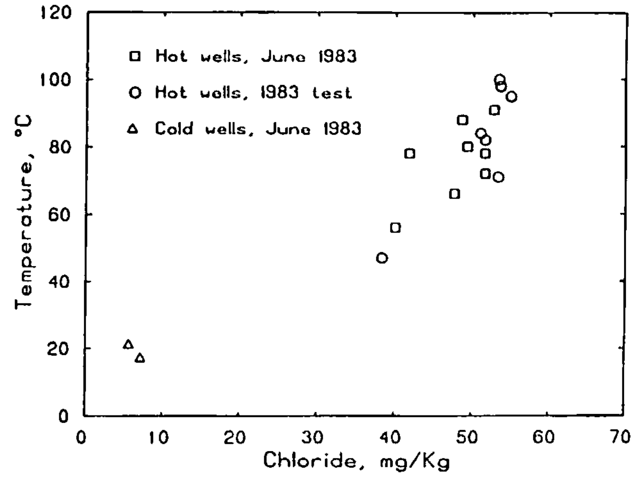

Figure 6. Temperature versus Cl concentrations of hot and cold well waters, showing a mixing trend.

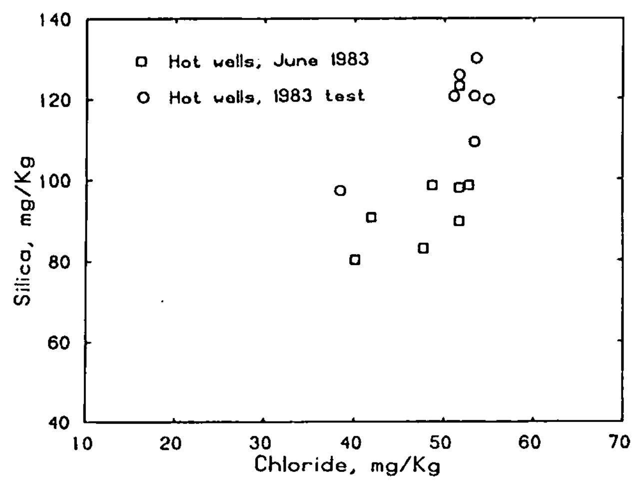

Figure 7. SiO_2 versus Cl concentration of hot waters.

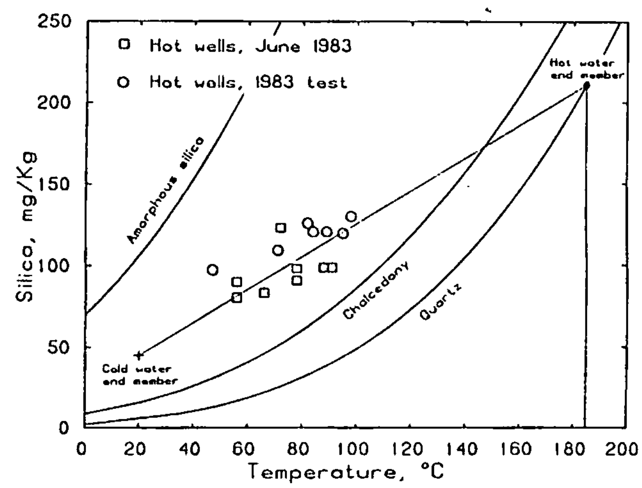

Figure 8. SiO_2 concentration versus temperature of thermal waters and their relation to SiO_2 solubilies

DIRECT HEAT
APPLICATION PROGRAM
SUMMARY

PRESENTED AT THE

GEOTHERMAL RESOURCES COUNCIL
ANNUAL MEETING

SEPTEMBER, 1979

PREPARED FOR THE
U.S. DEPARTMENT OF ENERGY
UNDER CONTRACT NO. DE-AC07-76ID01570

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ACKNOWLEDGMENTS

The project descriptions contained in this pamphlet were prepared by the Project Teams of each of the twenty-two direct heat application projects currently in progress throughout the United States. The Department of Energy gratefully acknowledges their assistance in providing this information which will assist other potential users in assessing the economic and technical viability of the direct use of geothermal energy. Additional copies of this pamphlet can be obtained through the Department of Energy Offices listed on page 5.

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SPECIAL SESSION - DOE-SPONSORED DIRECT HEAT

APPLICATIONS PROJECTS

September 25, 1979

Geothermal Resources Council 1979 Annual Meeting

Session Description:

This special open session on direct heat application project experience, sponsored by the Department of Energy, will feature panel discussions on geothermal:

- Space Conditioning Systems
- Applications for Agriculture/Aquaculture
- District Heating Systems
- Applications for Industry

Panel members are individuals with a wide variety of experience, who are currently involved in demonstration projects in the direct applications field. The panelists will present brief overviews of their projects, and respond to questions from the audience. Experience in resource exploration, well drilling, design, construction and permitting will be emphasized.

Agenda:

1:30 - 1:50 p.m. Direct Heat Applications Program Overview: Morris Skalka, Chief, Direct Heat Applications Section, DOE-Washington

1:50 - 2:00 p.m. Opening Remarks: Program Moderator, Bob Schultz, Manager, Hydrothermal Energy Commercialization, EG&G Idaho, Inc.

2:00 - 3:00 p.m. Panel Discussion: Geothermal Space Conditioning Systems

Moderator: Roland Marchand, Chief, Engineering Branch, Nevada Operations Office, DOE

Panelists	Project
Richard Berg, Project Engineer, Hengel, Berg & Associates	Haakon School, Philip, SD
Robert Sullivan, Project Engineer, Kirkham, Michael & Associates	St. Mary's Hospital, Pierre, SD
Gene McLeod, Project Manager, MERDI, Inc.	Warm Springs State Hospital, Mt

Panelists	Project
Marshall Conover, Project Manager, Radian Corporation	T-H-S Hospital, Marlin, TX and Navarro College, Corsicana, TX
Brian Fitzgerald, General Director, Klamath County YMCA	Klamath County, YMCA, OR
Jeff Burks, Research Analyst Utah Energy Office	Utah State Prison, UT
Sharon Province, Project Manager Westec Services, Inc.	El Centro, CA

3:00 - 3:45 p.m. Panel Discussion: Geothermal Application for Agriculture/Aquaculture

Moderator: Hilary Sullivan, Program Coordinator, San Francisco Operations Office, DOE

Panelists	Project
Ralph Wright, Chairman of the Board, Utah Roses, Inc.	Utah Roses, Inc., Sandy, UT
Dr. Stan Howard, Principal Investigator South Dakota School of Mines and Technology	Diamond Ring Ranch, SD
Frank Metcalfe, President, Geothermal Power Corporation	Kelley Hot Springs, CA
Becky Broughton, Hatchery Manager Aquafarms International, Inc.	Aquafarms International, Inc. Mecca, CA

3:45 - 4:00 p.m. BREAK

4:00 - 4:45 p.m. Panel Discussion: Geothermal District Heating Systems

Moderator: Eric Peterson, Program Manager, Division of Geothermal Resource Management, DOE-Washington

Panelists	Project
Roger Harrison, Project Engineer, Terra Tek, Inc.	Monroe City, UT
Dr. Glenn Coury, Project Manager Coury & Associates, Inc.	Pagosa Springs, CO
Phillip Hanson, Project Director, Boise Geothermal	Boise, ID

Special Session/Agenda (continued)

Panelists	Project
Harrold Derrah, Assistant City Manager, Klamath Falls, OR	Klamath Falls, OR
Phillip Edwardes, Principal Investigator Susanville, CA	Susanville, CA
David Atkinson, President, Hydrothermal Energy Corporation	Reno, NV

4:45 - 5:30 p.m. Panel Discussion: Geothermal Applications for Industry

Moderator: Robert Chappell, Project Manager Idaho Operations Office, DOE

Panelists	Project
Dr. Jay Kunze, Vice President & General Manager, Energy Services, Inc.	Madison County, ID
Robert Rolf, Director Technical Services, Ore-Ida, Inc.	Ore-Ida, Inc., Ontario, OR
Sheldon Gordon, Project Engineer, Chilton Engineering	Elko Heat Company, Elko, NV
Lee Leventhal, Project Engineer, TRW, Inc.	Holly Sugar, Brawley, CA

DIRECT HEAT APPLICATION PROJECTS

The use of geothermal energy for direct heat purposes by the private sector within the United States has been quite limited to date, yet there is a large potential market for thermal energy in such areas as industrial processing, agribusiness, and space/water heating of commercial and residential buildings. Technical and economic information is needed to assist in identifying prospective direct heat users and to match their energy needs to specific geothermal reservoirs. Technological uncertainties and associated economic risks can influence the user's perception of profitability to the point of limiting private investment in geothermal direct heat applications.

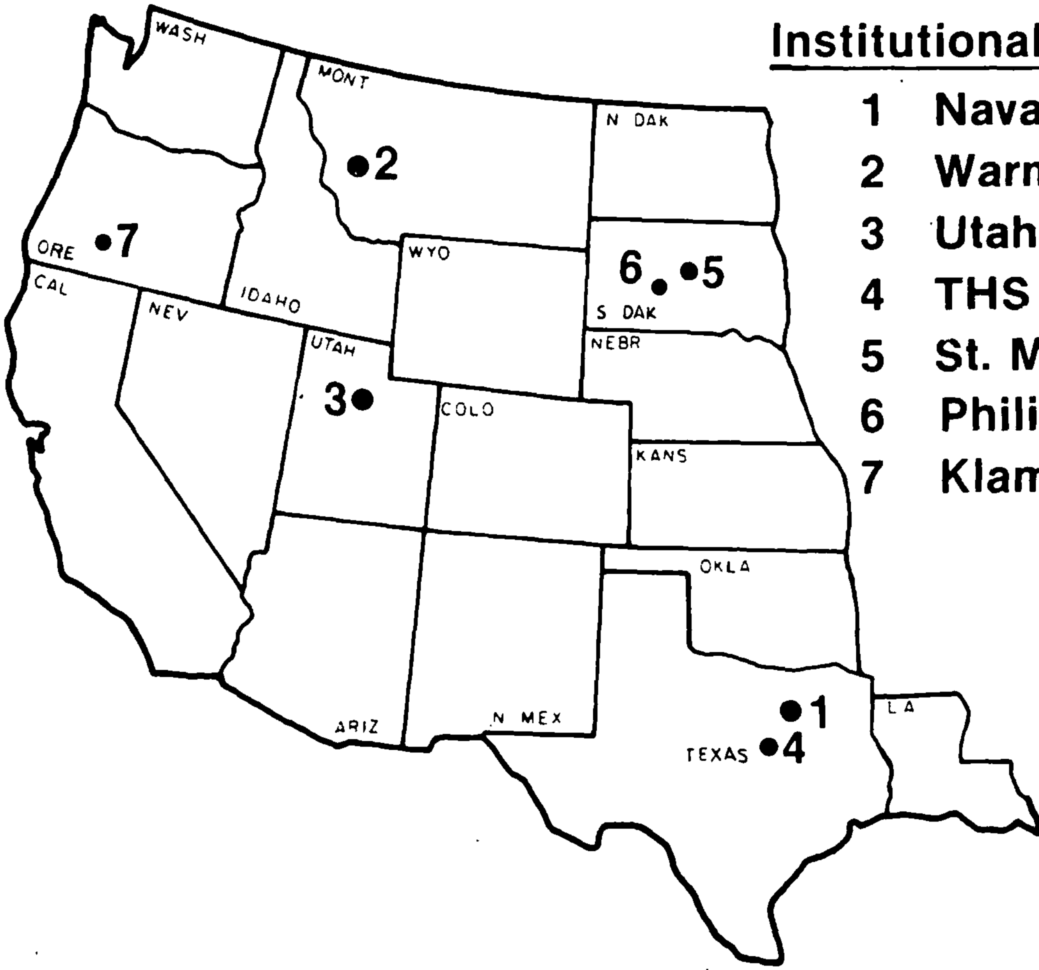
To stimulate development in the direct heat area, the Department of Energy, Division of Geothermal Energy, issued two Program Opportunity Notices. These solicitations are part of DOE's national geothermal energy program plan, which has as its goal the near-term commercialization by the private sector of hydrothermal resources. Encouragement is being given to the private sector by DOE cost sharing a portion of the front-end financial risk in a limited number of demonstration projects.

The twenty-two projects summarized in this pamphlet are a direct result of the Program Opportunity Notice solicitations. These projects will (1) provide visible evidence of the profitability of various direct heat applications in a number of geographical regions; (2) obtain technical, economic, institutional, and environmental data under field operating conditions that will facilitate decisions on the utilization of geothermal energy by prospective developers and users; and (3) demonstrate a variety of types of applications.

DOE PROJECT OFFICES

Three Department of Energy Operations Offices are responsible for the management of the direct heat application projects. The offices and their respective projects are:

	Projects
Idaho Operations Office 550 Second Street Idaho Falls, Idaho 83401 Contact: Robert Chappell Project Manager, DOE (208) 526-0085 Technical Support: EG&G Idaho, Inc. Idaho Falls, Idaho 83401	Boise Diamond Ring Ranch Elko Heat Company Haakon School Madison County Monroe City Ore-Ida, Inc. Pagosa Springs St. Mary's Hospital Utah Roses, Inc. Utah State Prison Warm Springs State Hospital
Nevada Operations Office P.O. Box 42100 Las Vegas, Nevada 89114 Contact: Roland Marchand Chief Engineering Branch, DOE (702) 734-3424	Navarro College T-H-S Hospital
San Francisco Operations Office 1333 Broadway Oakland, California 94612 Contact: Hilary Sullivan Program Coordinator, DOE (415) 273-7943 Technical Support: Energy Technology Engineering Center Canoga Park, California 91305	Aquafarms International Inc. El Centro Holly Sugar Kelley Hot Springs Klamath County YMCA Klamath Falls Reno Susanville



Institutional Heating Systems

1 Navarro College & Hospital Corsicana, Texas
2 Warm Springs Hospital, Montana
3 Utah State Prison, Utah
4 THS Hospital, Marlin, Texas
5 St. Mary's Hospital, Pierre, South Dakota
6 Philip School, South Dakota
7 Klamath Falls, Oregon, YMCA

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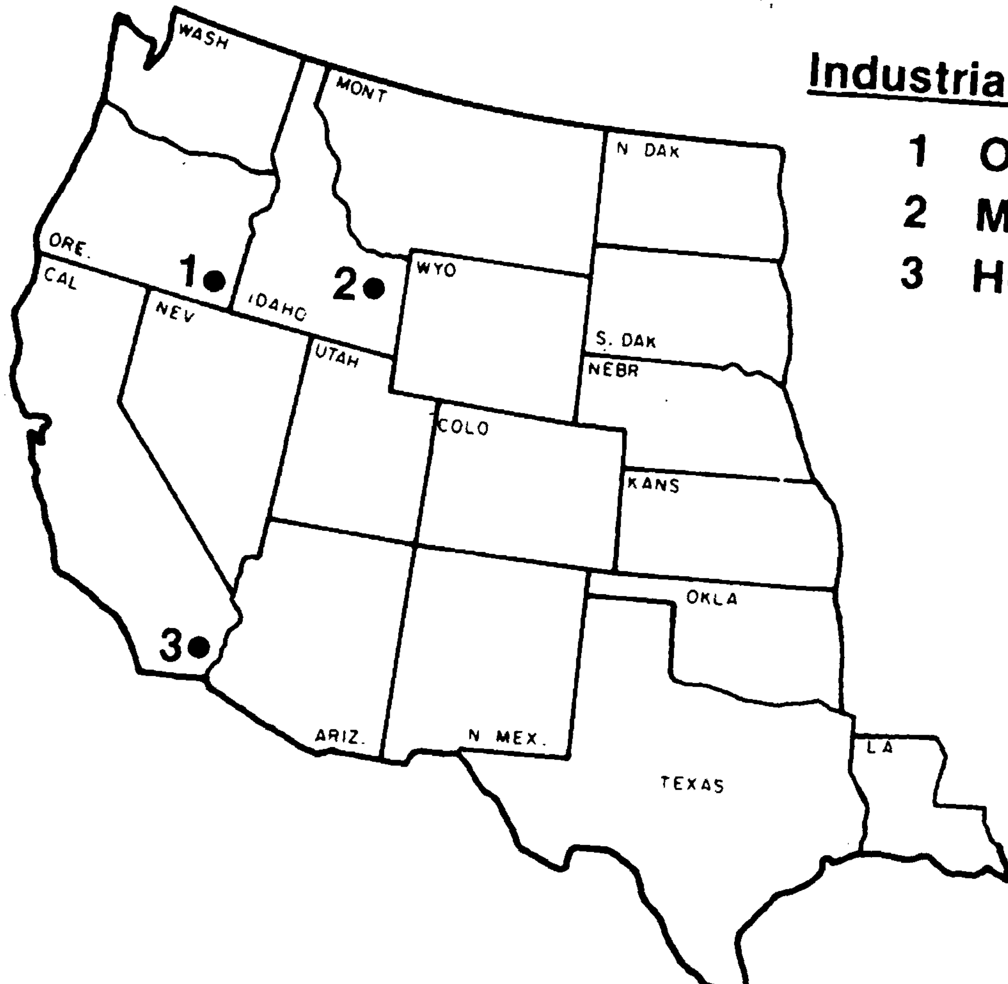

Industrial Process Sites

1 ORE-IDA — Ontario, Oregon
2 Madison County — Rexburg, Idaho
3 Holly Sugar — Brawley, California

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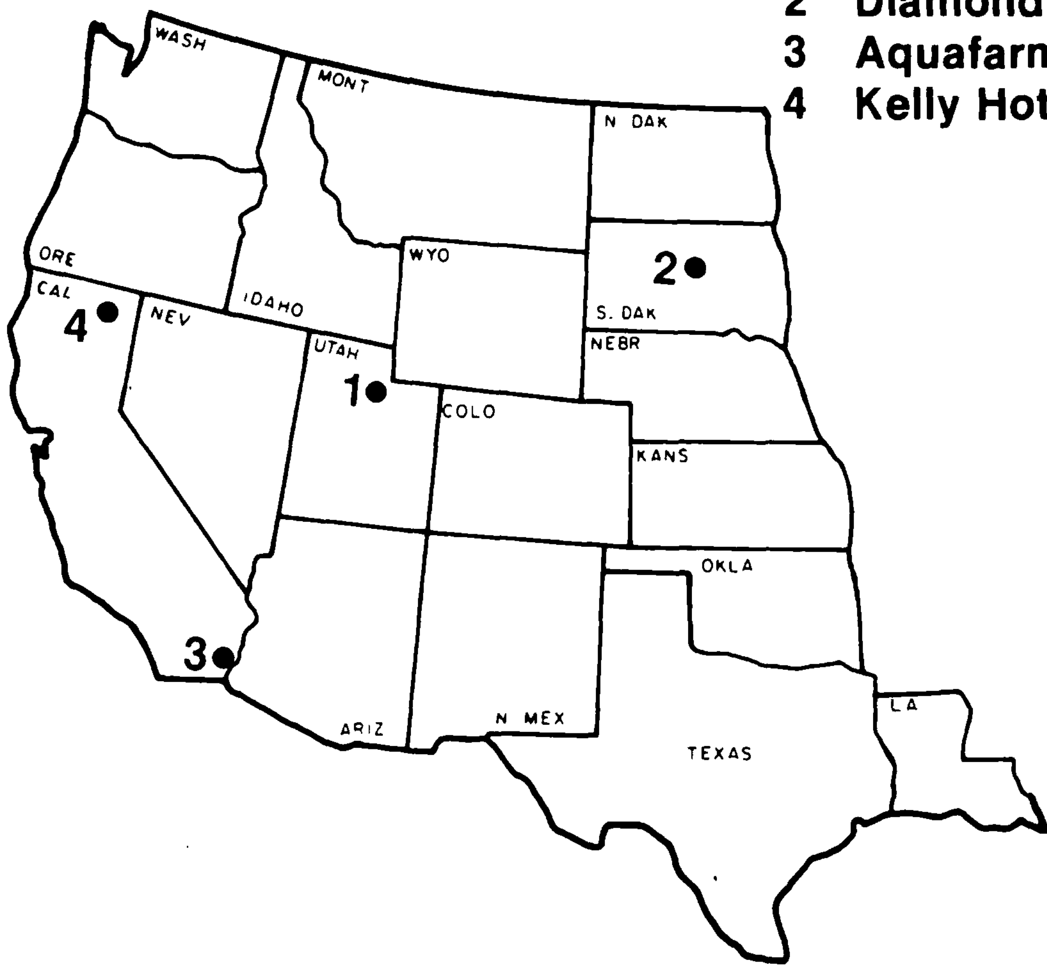
Agribusiness

1 Utah Roses — Sandy, Utah
2 Diamond Ring Ranch — South Dakota
3 Aquafarms International — Mecca, Calif.
4 Kelly Hot Springs — Novato, California



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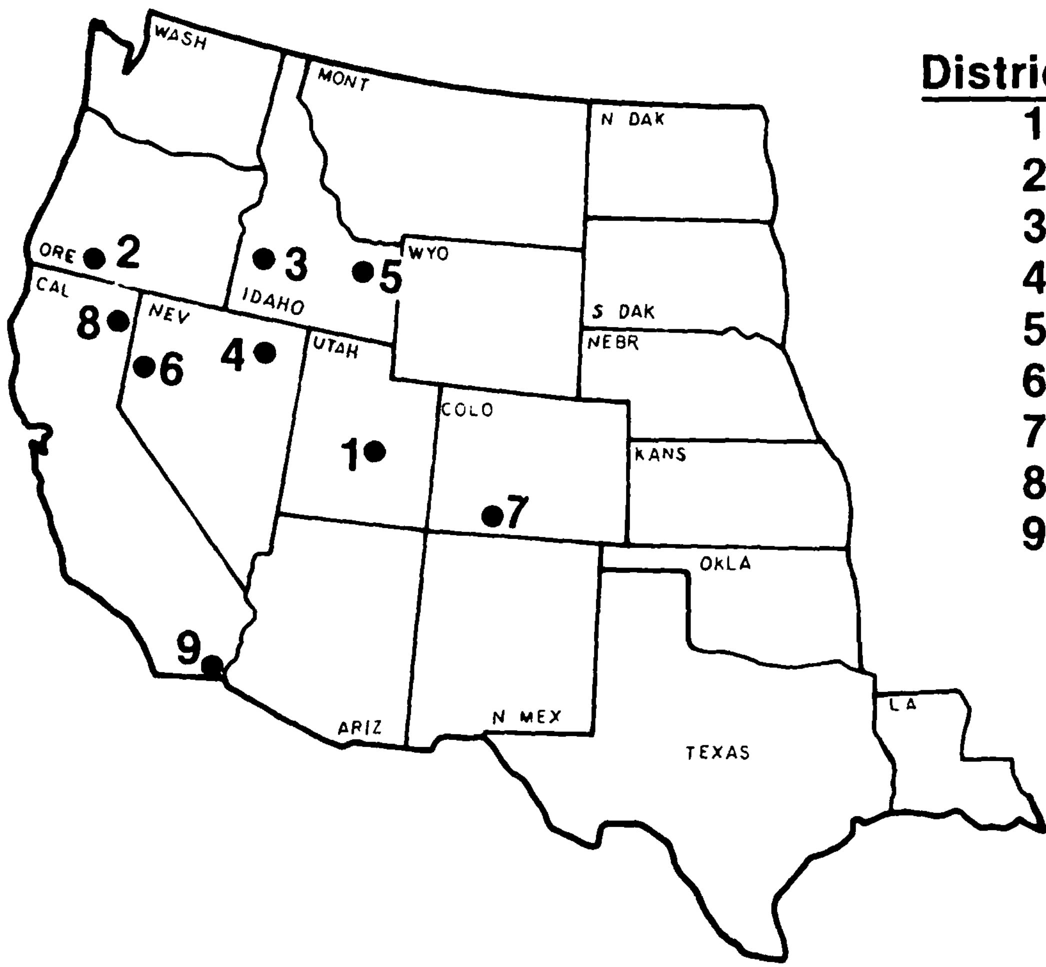


District Heating Systems

1 Monroe City, Utah
2 Klamath Falls, Oregon
3 Boise, Idaho
4 Elko, Nevada
5 Madison County, Idaho
6 Reno, Nevada
7 Pagosa Springs, Colorado
8 Susanville, California
9 El Centro, California

INEL-S-17 789

DIRECT HEAT

APPLICATIONS PROJECT

DESCRIPTIONS

Project Title: Commercial Culture of *Macrobrachium Rosenbergii* on Geothermal Water

Location: Mecca, California

Principal Investigator: Dr. Dov Grajcer, President, Aquafarms International, Inc.

Project Team:

- Aquafarms International, Inc.

Project Objective:

To develop a commercial-scale prawn farm in the Coachella Valley, utilizing geothermal water as the source of constant-temperature fluid. This would permit economical, year-round prawn farming.

Resource Data:

The project is located on 246 acres of desert land in one of the most desolate parts of the Salton Trough, the area between the Coachella Canal and the Salton Sea.

Subsequent drilling has proven successful. Three wells have been drilled to a depth of about 100 ft. They are all prolific producers of warm water under thermo-artesian pressure. The estimated artesian head is about 5 psig. The total flow rate is of the order of 300 gallons per minute per well, without pumping. The quality of the water is superb: the salinity of the water is less than 600 ppm TDS, making it less saline than the Coachella canal water flowing by. The salinity of the latter is on the order of 800 to 1,000 ppm TDS. The water issues out of the 10-inch (O.D.) wells, at a temperature range of 84 to 87°F, which is ideal for shrimp farming. Detailed chemical tests of water chemistry have established that the water is of an acceptable quality for giant shrimp (or prawn) farming.

System Design Features:

The equivalent energy demand for raising shrimp or prawn in artificially heated ponds is on the order of 170 billion Btu per year for Coachella Valley groundwater and ambient air conditions for a 50-acre pond farm. The equivalent energy saving would amount to about $560,000 per year (at $2/MMBtu and 60% boiler efficiency).

The three geothermal wells on the property provide water at the required pond temperatures. Geotechnical investigations will determine if slightly higher water temperatures would be expected at a slightly greater depth. In case of discovery of hotter water, it would be possible to control pond temperature in winter with greater ease, by proper mixing of water from wells of different temperatures.

It is estimated that well pumping would require a 10-kW generator to be installed on the deep well.

Project Description:

Aquafarms International, Inc. (AII), a small California corporation, is developing a 50 acre prawn farm on its property in the Dos Palmas area, east shore of the Salton Sea, utilizing geothermally heated water. Extensive genetic and field work have already been completed by AII to achieve adaptation of the giant Malaysian prawn, Macrobrachium rosenbergii, to local water, soil, and climate conditions.

The giant Malaysian shrimp enjoys many advantages over many other crustaceans. It adapts to a relatively wide temperature range, with the optimum temperature being in the 80 to 85°F range. The female prawn is highly productive and protects her eggs, resulting in a relatively high (30 to 50%) survival rate of the larvae. The larvae metamorphose to juvenile prawns in 22 to 35 days; depending upon temperature, the juveniles reach maturity in 7 to 8 months. And, finally, the meat has excellent taste and quality, and the product is much in demand worldwide.

The project will utilize geothermal water issuing from three existing shallow wells, plus one deeper well to be drilled as part of the project, to provide enough warm water for continuous, year-round, prawn farming operation. Appropriate geotechnical studies will be carried out to determine optimal location for the new well, to test hydrologic characteristics of all wells, and to determine best methods of disposal of the water after it has been used in the prawn ponds. Studies of optimum feed, flow-through, efficient water quality control methods, and harvesting methods will also be determined.

Status:

The environmental report has been submitted for approval.

Project Title: Boise City - A Field Experiment in Space Heating

Location: Boise, Idaho

Principal Investigator: Phil Hanson, Director, Boise Geothermal (208) 384-4013

Project Team:

- Boise City
- Boise Warm Springs Water District
- CH_2M Hill Engineers

Project Objective: To develop a geothermal space heating system to serve the largest possible market in and around the Boise central business district.

Resource Data: The resource area is commonly referred to as the Boise Front. This area appears to be fault controlled, with the source of water being the annual runoff in the mountains immediately behind Boise City. There is a long history of using this resource data, dating to 1892 when the first wells were drilled to a depth of 400 feet. These wells are still productive, with water temperatures relatively invariant at 170°F. Since that date, there has been fairly continuous development of hot water wells. Records available on some 70 wells show temperature ranges of 75 to 170°F, and depths ranging up to 1,200+ feet. Production varies over the range up to 800 gpm.

System Design Features: Boise Geothermal is a joint venture of Boise City and Boise Warm Springs Water District. This joint venture will develop a space heating system consisting of two major parts. The first part is based on the Warm Springs heating district, which, in one form or another, has been operating since the 1890's. This part of the system presently serves about 220 residences, based on two 400-ft wells with temperatures of 170°F. This part of the system will be improved to provide expanded service to the residential community.

The second part of the system will draw on a separate part of the resource area to supply heat to the central business district. It is planned that the system will serve, initially, approximately 11 major buildings. These buildings range from the 270,000 square foot Idaho First National Plaza, built in 1978, to a renovated 1930's hotel that is now an office building.

The types of heat exchangers used will vary. The system capacity is a nominal 5,000 gpm, designed to take advantage of a 40 to 50°F temperature drop (170 to 120°F) to heat residential and commercial buildings. Fuel savings are expected of 230,000 barrels of oil for a system serving 500 to 1,000 residences and 11 office buildings.

Project Description: The project is managed and operated through Boise Geothermal. A Board of Directors, made up of the Boise City Council and members of the Boise Warm Springs Water District Board, provides policy direction to Boise Geothermal. An Executive Committee maintains daily involvement in project work. Overall project management is the responsibility of a Project Director, who reports to the Project Board. CH_2M Hill provides project technical direction.

Funding for the project is being supplied by the Economic Development Administration, the Department of Energy, Boise City, and Boise Warm Springs Water District. As the project develops, funding is planned from other sources. These funds will be used to prove the extent of the resource. If the resource is large enough, the first segment of the system will serve 500 residences and 11 office buildings. Service to this segment will be evaluated, to determine the technical and economic feasibility of expanding the system.

Status: Contracts with EDA and DOE were signed in July 1979. Preliminary work on geological and environmental studies actually began in March 1979. An environmental report has been completed. Geology studies are continuing, with some concurrent drilling. All wells should have been drilled and service begun between 1980 and 1982.

Project Title: Diamond Ring Ranch Geothermal Demonstration Heating Project

Location: Mid-central South Dakota, 35 miles west of the state capitol

Principal Investigator: Dr. S. M. Howard, Professor of Metallurgical Engineering, (605) 394-2341

Project Team:

- South Dakota School of Mines and Technology
- Re/Spec, Inc.
- Diamond Ring Ranch

Project Objective:

Utilize existing Madison well to provide grain drying, cattle warming, and space heating for homes.

Resource Data:

The geothermal water is coming from the Madison Limestone, which is under most of western South Dakota, at depths from 2,500 to 7,500 feet and temperatures from 100 to 195°F. The Madison is a major aquifer of South Dakota, yielding good quality drinking water. The existing well is 4,100 feet deep, flowing at approximately 152°F and 180 gpm.

System Design Features:

The system is designed using PVC piping and plate-type isolation heat exchangers made of 316 stainless steel. The system is designed for a 20°F temperature drop across the isolation exchangers. The grain dryer was designed to use antifreeze in its "clean" water side, to avoid complications associated with freezing weather. The space heating provided to the homes will use existing duct work and installation of water-to-air exchangers in line with the existing heating system. This will permit the existing system to function as a backup unit, if necessary.

The system is designed for simplicity and minimal control systems, to permit economical installation and elementary operational problems.

Status:

Ground breaking ceremonies were conducted on July 19, 1979, with project completion slated for October 19, 1979. The main pipeline has been excavated and is nearly installed. The grain dryer is on-site and retrofit procedures are commencing. The retrofit for the space heating is also underway. The project will hopefully be completed to permit some grain drying operations to begin in late September.

Project Title: City of El Centro Geothermal Energy, Utility Core Field Experiment

Location: El Centro, California

Principal Investigator: Mr. G. L. Herz, Assistant City Manager of El Centro (714) 352-9440

Project Team:

- City of El Centro
- WESTEC Services, Incorporated
- Chevron Resources Company

Project Objective:

The overall objective of this field experiment is to demonstrate the engineering and economic feasibility of the utilization of moderate-temperature geothermal heat, on a pilot scale, in the City of El Centro, for space cooling, space heating and domestic hot water. This field experiment will provide visible evidence of the profitability of direct heat applications to residential/commercial space conditioning, particularly in the southwestern United States.

Resource Data:

The geothermal reservoir which is the energy source for this demonstration is embodied in a 13.5 square mile area known as the Heber Known Geothermal Resource Area in the Imperial Valley. The City of El Centro is 4-1/2 miles north of the center of the KGRA, in an area where well temperature gradients should be 2 to 4°F per 100 feet in depth.

Reservoir Characteristics (predicted) at the El Centro City Site

Total Dissolved Solids (TDS)	14,000 ppm
Brine Chemistry for Thermodynamic Calculations	14,000 ppm solution of NaCl
pH	6.2
CO_2 by weight of flashed steam	$\leq$ 0.3%
Methane and hydrogen sulfide by weight of flashed steam	trace
Maximum supply rate per well	365,000 lb/hr per well
Downhole Brine Condition at the City	Saturated 250°F at 8,500 feet
Brine return temperature at the reinjection well	$\geq$ 160°F

System Design Features:

The basic concept is to use the geothermal brine to heat clean City supply water and circulate this clean water to the Community Center for space and water heating purposes. Also, this clean, hot City supply water would be used in a lithium bromide/water absorption chiller to produce chilled water, which would be circulated to the Community Center for space cooling purposes. This design is based on the concept that the hot water/chilled water plant will be located at the proposed drill site, about 1/2 mile away from the Community Center. The reason this plant is not located at the Community Center is that the modular concept of district heating and cooling developed in the initial feasibility study will be evaluated in this demonstration. Under this concept, the area of a city to benefit from district heating and cooling would be divided into small districts in which one modular plant would serve a particular district. This demonstration plant is conceived as a modular plant serving not only the Community Center but, hopefully in the future, other consumers in the area--residential and industrial alike.

Key Design Features

Number of Production Wells	One
Number of Injection Wells	One
Type Absorption Chiller	Lithium bromide/Water
Cooling Capacity	101 tons nominal 65 tons available
Hot Water Temperature IN	235°F
Hot Water Temperature OUT	215°F
Type Heat Exchanger	Undetermined at this time
Capacity (max.)	1.2×10^6 Btu/hr
Brine Temperature IN	250°F
Brine Temperature OUT	$\geq$ 160°F
Hot Water Temperature IN	215°F cooling mode 175°F heating mode
Hot Water Temperature OUT	235°F cooling mode 195°F heating mode
Estimated Geothermal Fuel Cost/ Million Btu	$4.78 (based on fully developed district wide system, including industrial park use)
Annual Fuel Savings	4.6×10^5 cu ft/yr natural gas 1.7×10^5 kWh/yr electricity

Project Description:

The project plan calls for drilling a geothermal well within the city, building a pilot hot water/chilled water plant at the wellsite, and distributing the hot or chilled water to the El Centro Community Center (located about one-half mile away from the pilot plant). Heat from the brine will be transferred to the working fluid by way of heat exchangers located at the wellsite. City supply water has been selected as the working fluid because of its relatively low cost and availability.

The heated city water will be used in the winter to supply space heat and heat for domestic water for the Community Center. During the summer, the heated city water will be used in a lithium bromide/water absorption process to produce chilled water to be used for space cooling the Community Center. The Community Center will be retrofit with heating/cooling coils for the space conditioning requirements.

Status:

The prime contract for this project between the Department of Energy (DOE) and the City of El Centro was executed on July 11, 1979. The environmental impact report was certified by the El Centro City Council on July 5, 1979. The technical conceptual design was completed on August 3, 1979, and the detailed design phase is now in progress.

Project Title: Field Experiments for Direct Uses of Geothermal Energy: Elko Heat Company, Elko, Nevada

Location: City of Elko, Nevada

Principal Investigator: Mr. Ira S. Rackley, P.E., Project Manager Elko Heat Company, (702) 738-3108

Project Team:

- Elko Heat Company, Elko Nevada; Mr. Jim Meeks, President
- Chilton Engineering, Elko, Nevada; Mr. Ira S. Rackley, P.E., Project Manager, and Mr. Sheldon S. Gordon, P.E., Project Engineer

Project Objectives:

This project was selected to demonstrate the technical and economic feasibility of the direct use of geothermal brines from the Elko KGRA for the purpose of providing space, water, and process heat. In a more general sense, it is the aim of the project to develop information and approaches that will enable the proposers to develop the Elko resource as a viable alternative to the consumption of primary fuels for space, water, and process heating in Elko.

Objectives related to this overall goal are:

- Develop adequate resource information to allow for the design of the geothermal process system.
- Use this resource information to generate a plan for the continued development and use of this resource after the period of government support.
- Displace a significant portion of the primary fuel consumption in Elko for identified energy markets with geothermal energy.

Resource Data

Resource Area: Adjacent to the Elko KGRA, within the city limits of Elko.

Controlling Geologic Features: Fault zone trending north-northeast through city of Elko; hot water from depth ascending along the fracture zone.

Predicted Temperature: Geothermometry-based predictions (240-670°F) (actual unknown).

Predicted Flows: Unknown

Depth of Resource: 700 to 2,000 feet, based on cold water well drilling logs (actual unknown).

System Design Features:

Production/Injection Wells: One production well (700-2,000 feet in depth), one injection well (similar depth). (Actual use is dependent on water quality considerations.)

Heat Exchangers: Use is dependent on water quality and resource temperature considerations. Design at present allows for wellhead heat exchanger and closed-loop system circulation. Heat exchanger design anticipates 10-15°F approach.

System Capacity: Present extraction permits under assumed operating ΔT of 10°F on shell side of heat exchanger provide a net capacity of 6.74×10^6 Btu/hr (actual capacity unknown).

Unique Design Considerations:

- shallow resource
- clean resource (dilute samples at 550 ppm - TDS)
- variety of applications:
    - commercial laundry
    - 400-unit motor hotel
    - office building

Project Description:

The Elko project involves the location and drilling of a production well for the purpose of extracting hydrothermal fluids from the Elko KGRA. These fluids are to be used to displace primary fuel consumption for the operation of a commercial laundry, a motor hotel, and an office building.

The Vogue Laundry and Dry Cleaners requires energy for the operation of washing equipment, dryers, and ironers. The Stockmens Motor Hotel requires energy for domestic water, space, and swimming pool heating. The Stockmens Motor Hotel also has substantial cooling requirements that may be met if the geothermal source is of sufficient temperature. The Henderson Bank Building requires energy primarily for space heating, with a small domestic hot water requirement. Thus, several different applications of the direct utilization of the Elko geothermal resource will be demonstrated and tested in this program.

Status:

The Elko project is in the resource assessment phase, with Geothermal Surveys, Inc. just completing the temperature probe survey, geologic reconnaissance, electrical resistivity soundings, sling ram soundings, and some of the geochemical sampling of city wells and the Elko Hot Springs.

The Environmental Report has been completed for this project, with no significant environmental effects expected to be caused.

Project Title: Direct Utilization of Geothermal Energy for Philip Schools

Location: Philip, South Dakota

Principal Investigator: Charles A. Maxon, Superintendent of Schools, (605) 859-2679

Project Team:

- Haakon School District 27-1
- Hengel, Berg & Associates, Rapid City, South Dakota
- Francis-Meador-Gellhaus Inc., Rapid City, South Dakota

Project Objective: To obtain water at 155°F (66°C) from the Madison Formation that can be used for space heating and domestic water heating at the Philip School buildings of the Haakon School District 27-1.

Resource Data: The Madison Formation underlies most of western South Dakota. In the Philip area, the depth to the Madison Formation is approximately 4,000 feet. The temperature of the water from the Madison Formation in this area is 155°F (66°C). The flow rate of the well drilled by the school to a depth of 4,266 feet is 300 gallons per minute, at a temperature of 155°F (66°C).

System Design Features: A 4,266-foot well, with artesian flow, has been drilled. Two stainless steel plate-type heat exchangers will be provided. One will provide 1,800,000 Btu/hr for space heating of an Armory-High School building. The other will provide 1,130,000 Btu/hr for space heating of an elementary school building, a vocational education building, and two small music buildings. Temperature of geothermal water delivered to heat exchangers is 155°F (66°C). Leaving temperature from space heating heat exchanger is not less than 130°F (54°C). Water leaving the space heating heat exchanger is piped through a domestic water heat exchanger.

The heat energy remaining after the school space heating is satisfied will be piped through part of the Philip business district. The heating district plans to utilize the low-temperature water (approximately 125 to 130°F) in a direct heat application, i.e., fan-coil type heat exchangers.

The estimated annual fuel savings for the school is 36,200 gallons of fuel oil and 107,000 kWh of electricity.

Additional savings of fuel oil will be realized from the heating district.

Project Description: A 4,266-foot well was drilled to the Madison Formation. The well will produce a sustained flow of 300 gpm of water, at 155°F (66°C). However, to utilize the well pressure to circulate the water through the heat exchangers, the heating system was designed to use 250 gallons per minute.

The water will be piped from the well that is located near the school buildings, to the Armory-High School building and to the elementary school building.

Because of the corrosive action of the Madison Formation water, the recommended materials to be used in this system are 316 stainless steel and the plastics.

The pipe from the well house to the buildings will be high-density polyethylene pipe equal to Driscopipe. Supply piping inside the buildings will be chlorinated polyvinyl chloride.

A 316 stainless steel plate-type heat exchanger will be provided in each of the buildings. The existing low pressure steam heating system will be modified to hot water systems by replacing steam coils with hot water coils in the fan coil units, by adding additional fan coil units, and by using the existing baseboard radiation. One of the boilers will be replaced because of its condition and the other boiler will be retrimmed to a hot water boiler.

The Armory-High School building is approximately 30,000 square feet. The heat exchanger in the elementary shool building will be used to heat that building, the Vocational Education building, and two small buildings used for music classrooms. These four buildings have approximately 28,000 square feet.

In addition to the space heating, the domestic hot water will be provided for the Armory-High School building and the elementary school building. A 316 stainless steel plate-type heat exchanger will be located next to the space heating heat exchangers. These heat exchangers will use either the leaving water from the space heating heat exchangers or the geothermal water, depending upon the space heating demands.

The exit temperature from the heat exchangers will be approximately 130°F (54°C). The water will be piped through a part of the Philip business district. Several building owners have indicated an interest in utilizing the remaining heat energy. They propose to use fan coil type heating units. These units are estimated to have a life of ten years before they may have to be replaced.

Because of the presence of Radium 226 in excess of the EPA allowable for domestic water, the water will be treated with barium chloride. After removal of the Radium 226, the water will be discharged into the Bad River, which flows through Philip.

The proposed barium chloride treatment plant will have two 5,000-gallon mixing tanks that will have a 10 percent aqueous solution. The solution will be added to the water. The water will pass through a static mixer. From the static mixer, the water will be piped to two detention ponds. After three days, the water will be acceptable for discharge.

Status:

The well has been drilled, cased, and flow tested.

The design of the retrofit is approximately 85 percent complete. A design review was recently completed, with some changes in the plans.

A feasibility study, financed by nine businessmen, to use the leaving water from the school to heat their buildings, has been completed. Their final decision has been delayed because the location of the final discharge point has not been established.

The most economical method of removing the Radium 226 from the water appears to be the addition of barium chloride and the precipitation of the Radium 226.

Various locations for the treatment plant are being investigated.

Bids for the heat exchangers have been received by the school. A final decision on the award of the contract will be made in the near future.

The plans and specifications will be completed and construction contracts obtained, with construction starting April 1, 1980.

Project Title: Geothermal Energy for Sugar Beet Processing

Location: Brawley, California

Principal Investigator: J. J. Seidman, Program Manager, (213) 536-1955
J. M. Kennedy, Co-Investigator, Project Manager Geothermal Resource, (213) 535-1571
E. L. Leventhal, Co-Investigator, Project Manager System Design, (213) 536-1955

Project Team:

- TRW Energy Systems Group
- Holly Sugar Corporation

Project Objective:

The objective of this project is to implement a three-phase program to replace large quantities of fossil fuels with geothermal energy for sugar processing at Brawley, California, in a technical straightforward, economically sound and environmentally acceptable manner.

Resource Data:

The Imperial Valley of southern California is within a major rift zone. Tectonic stresses, acting on valley fill sediments, have caused faulting and opened fracture, permitting the formation of geothermal convective cells. Such a convective cell is thought to exist near the termination of the Imperial Fault, in section 30 T14S, R14E SBBM. The local fault system was first mapped on the surface and later verified by geophysical surveys (seismic, resistivity, and heat flow) performed as part of this project. The fault system, which consists of four tension faults splaying off the Imperial Fault, is thought to provide permeability and the conduit for the convective cell. The geologic structure, in concert with the thermocline of other wells in the region, suggests a source of 350°F (177°C) at about 8,000 feet (2440 m). If the geothermal system is as predicted, a flow of about 1,000 gallons per minute (63 liters/sec) may be achieved.

System Design Features:

The system will replace one boiler that is presently used to supply low pressure steam (≈ 25 psig) to the evaporators and juice heaters, and will supply heat to pulp dryers. There will be about 13 heat exchangers in the system. Three of these will be used to generate 25 psig steam, and the other 10 to preheat and heat the pulp-drying air. The capacity of the system will be 75 million Btu/hr for the steam generator and 160 million Btu/hr for the air heaters, for a total of about 235 million Btu/hr. This system will be used for about four months per year during the sugar campaign.

The total cost of the system, including the study phase and the design, is expected to be about $20 million. It is estimated that the heat supplied by the geothermal resource for the 4-month campaign will save about 100,000 barrels of oil/year. At an estimated cost of $5-6/MBtu, this is equivalent to $30/barrel of oil. Based on a full year's usage, the system could replace 300,000 barrels of oil and drop the cost to about $2/MBtu; this is equivalent to $10-15/barrel.

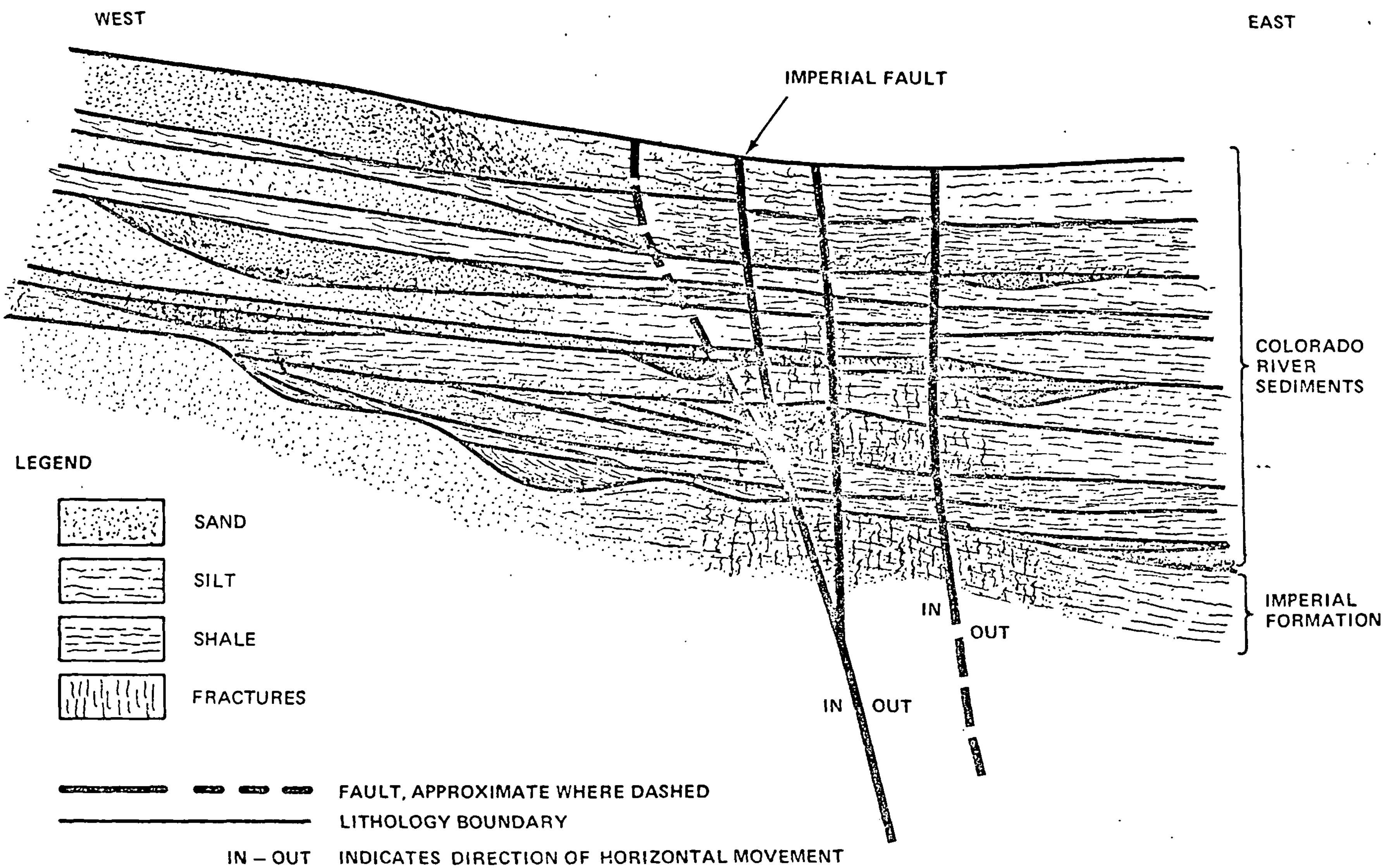
Project Description:

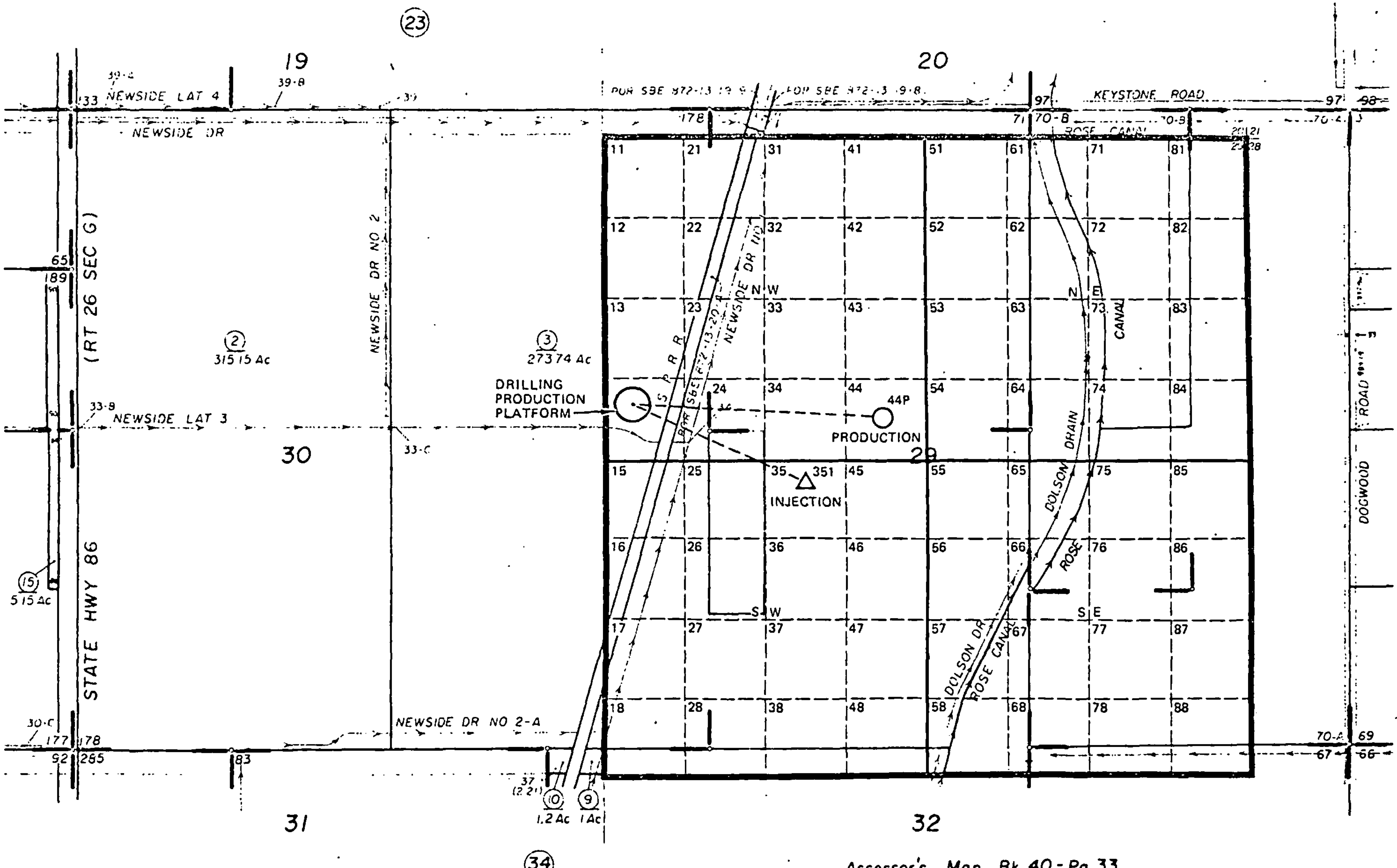
During the present phase of the contract (Phase I), the main users of geothermal heat in sugar processing have been investigated, an environmental report published, a drilling plan completed, and an application for a drilling permit submitted. An assumption was made that the well will produce water at 350°F, and, based on this assumption, low pressure steam generators and pulp dryers have undergone preliminary design. The accompanying equipment, piping, and required installation have also been identified. In Phase II, one production and one reinjection well will be drilled, and a pilot plant will be assembled. The equipment design will be based on data obtained during the drilling of the first production well.

Status:

The drilling of the first production well is scheduled to start in October 1979.

EAST - WEST DIAGRAMMATIC CROSS SECTION OF IMPERIAL VALLEY SEDIMENTARY COLUMN, SECTIONS 29 AND 30 T14S, R14E





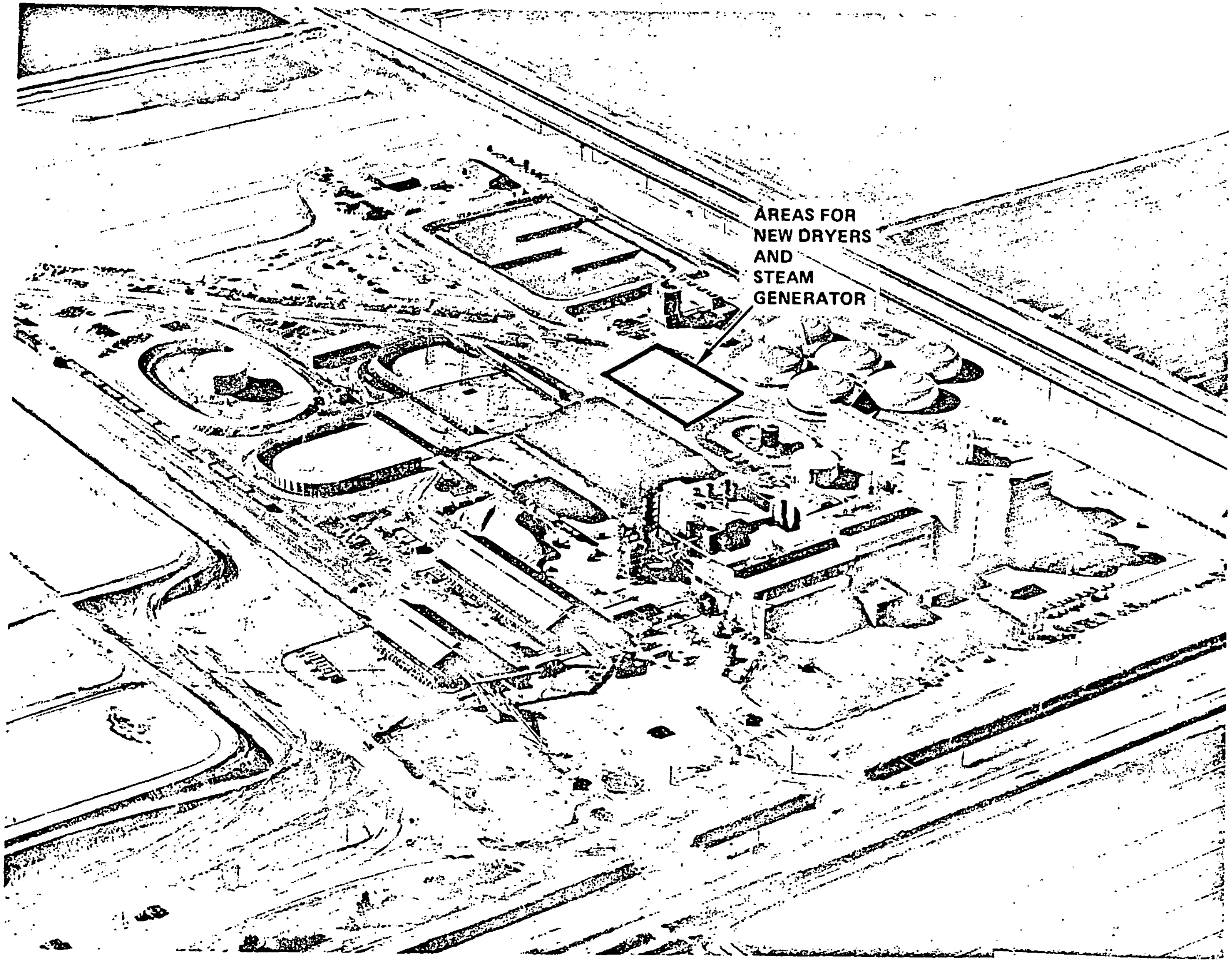
○ – PRODUCTION

△ – INJECTION

NOTE – Assessor's Block Numbers Shown in Ellipses
Assessor's Parcel Numbers Shown in Circles

Assessor's Map Bk. 40 - Pg. 33
County of Imperial, Calif.

SHOOT AT 65 %
FIG 3 PAGE 14
REMARKS


AREAS FOR
NEW DRYERS
AND
STEAM
GENERATOR

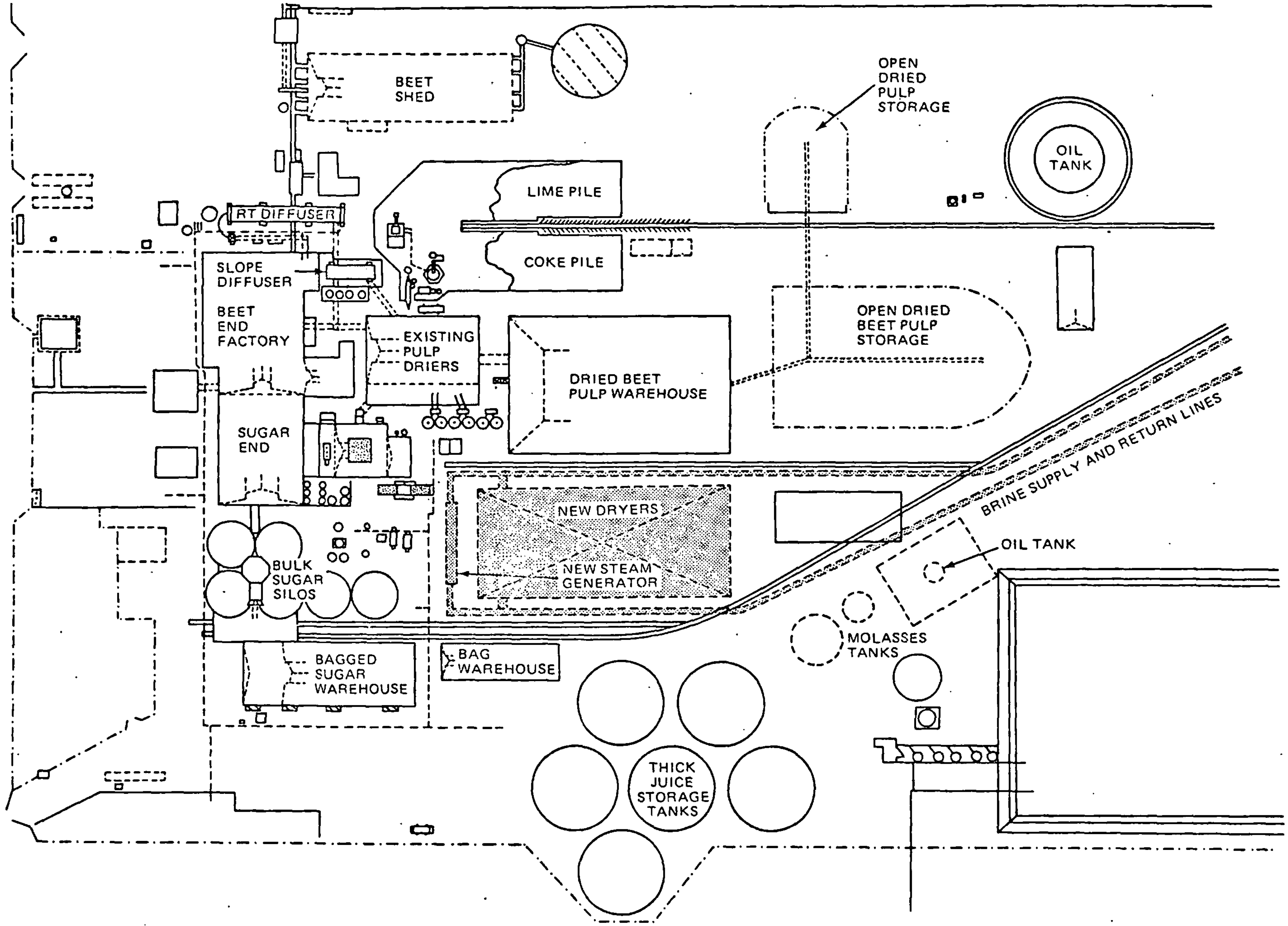


Figure 2. HOLLY SUGAR PLANT
With Geothermal Retrofit Equipment Areas

FIGURE NO. 2
PAGE NO. 13
REDUCE/~~ENLARGE~~
TO 80 %

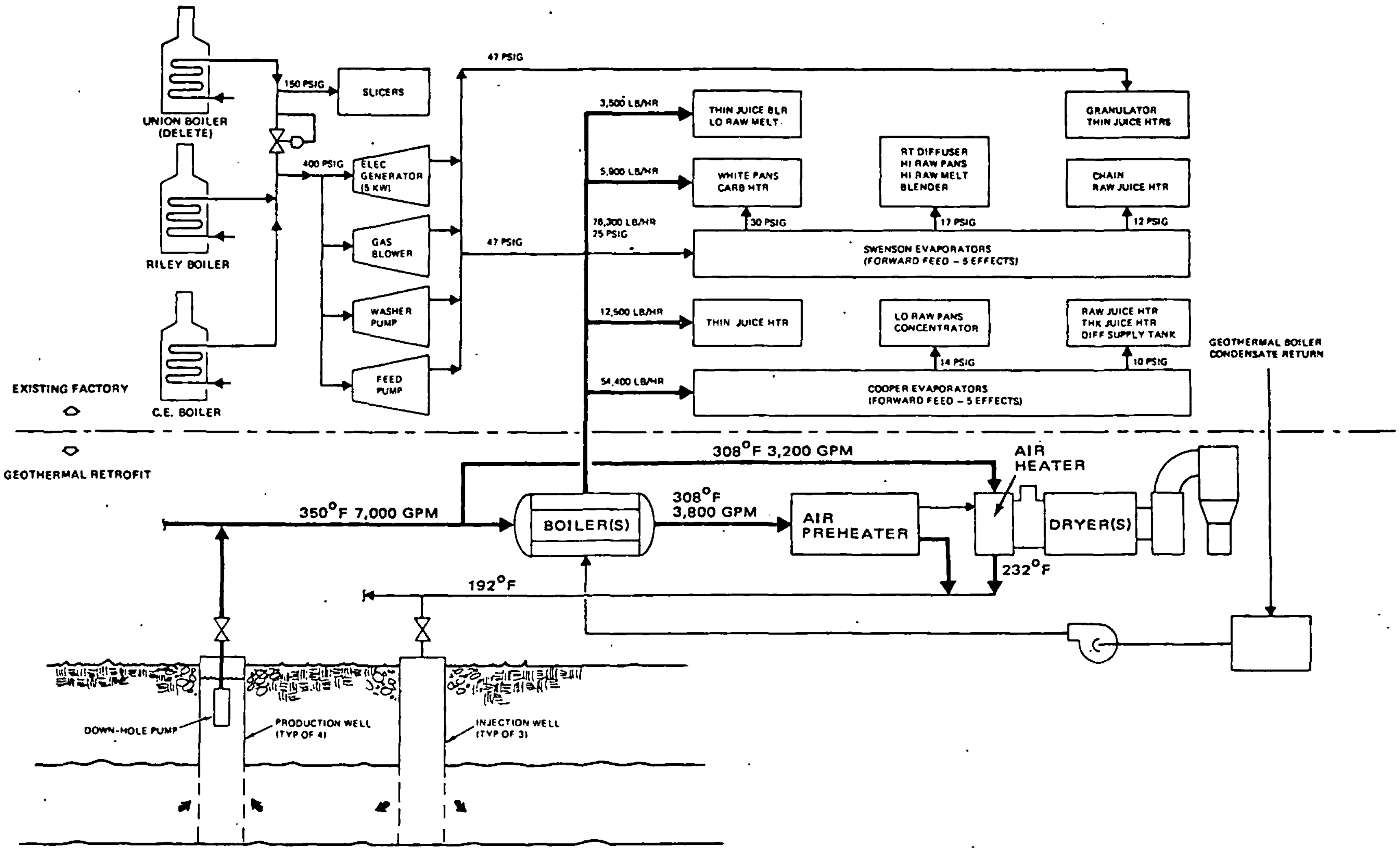


Figure 5. Geothermal Sugar Project Schematic

Project Title: Kelley Hot Springs Agricultural Center

Location: Kelley Hot Springs, Modoc County, California

Principal Investigator: Alfred B. Longyear, (916) 441-4510

Project Team:

- Geothermal Power Corp., Prime Contractor; Frank Metcalfe, Program Manager
- Agricultural Growth Industries, Inc.; Dr. Richard Matherson, Agricultural
- International Engineering Co.; Leonard A. Fisher, Civil Engineering
- Coopers & Lybrand; William R. Brink, Economics and Project Cost Management
- Ecoview; Dr. James Neilson, Environmental Report
- Carson Development Co.; Johan Otto, Construction Management

Project Objective: To demonstrate the economics and feasibility of using low-temperature geothermal energy for an integrated swine raising complex in Northern California.

Resource Data: The Warm Springs Valley of the Pit River is highlighted by Kelley Hot Springs, flowing at 96°C (205°F) at 320 gpm from a single orifice. The flow is at boiling for the elevation (4,360 feet). The region is part of the Modoc Plateau province. The Pit River Valley contains a thin veneer of stream-channel alluvium, flanked by terrace deposits and older and younger fan deposits. Beneath this are sedimentary and tuffacious beds of the Alturas Formation. Overlying these on higher hills are basalt flows of Pliocene and Pleistocene age. The principal fault of the region is the northwest-trending Likely fault, which passes about one mile west of Kelley Hot Springs, and which appears to be a significant regional boundary.

Extensive exploration data include: Reconnaissance-level geologic mapping and gravity surveys, an aeromagnetic survey, at least 30 sq mi of electrical resistivity surveys, a reconnaissance-type telluric survey, a ground-noise and micro-earthquake survey, geochemical analyses, extensive temperature gradient surveys over a 15 sq mi area with 2.5-3 HFU across the area and a high of 20 HFU in certain holes.

Two exploration wells have been drilled. In 1969, Geothermal Resources International drilled to 3,200 feet, 1/4 mile south of the spring with a maximum temperature of 110°C (230°F) at bottom. In 1974, Geothermal Power Corporation drilled to 3,396 feet, approximately 1-1/2 miles due east of the GRI #1 well. The maximum bottom hole temperature of 115°C (239°F) was measured in 1977 in KHS #1. The lithology of the two wells is similar.

The proven reserve in this project is a body of hot water at over 240°F in a porous reservoir between about 1,600 to 3,400 feet depth, covering an area of several square miles. A conservative estimate of the resource, assuming an areal extent of 4 sq mi, thickness of 2,000 ft, a reservoir temperature of 240°F, a reinjection temperature of 80°F, and a porosity of 20% (KHS #1 logs), will amount to heat in the fluid of 6.73 x 10^{16} calories. The reservoir within the drilled depth has sufficient reserve to supply the proposed plant, plus considerable additional development.

System Design Features: The GRI #1 well will be reentered and pump tested. An additional standby and reinjection well are planned. A primary tube and shell heat exchanger will be used. The geothermal fluid will be maintained in the fluid state from supply to reinjection. Fan coils, radiant floor and possible plate-type wall heaters will be the principal sources of space heating. Absorption refrigeration for sprouted grain raising and liquid cooling in the feed processes will be evaluated. Heating coils in the methane digesters and in the treatment ponds will also be evaluated. The system will operate with 240°F nominal supply and reinjection at 80°F. The peak capacity is on the order of 64 million Btu/hr. The annual fuel savings is estimated to be 2.5 million gallons of oil.

Project Description: Based upon the characteristics of the Kelley Hot Springs resource, regional markets and raw material supplies, and a recent related DOE PRDA[1], a totally confined swine raising complex was proposed for a field experiment.

A 1,200 sow swine raising complex, utilizing geothermal direct energy, will be designed, developed, constructed, and operated as a field experiment. During subsequent operations, it is planned that the complex will be expanded to 4,800 sow capacity. The field experiment will be composed of a feed production facility, a totally confined system of swine raising buildings, employee services and maintenance facility, and a waste management system.

All commercial hardware will be utilized. Commercial swine raising facilities will be evaluated. Engineering design effort will be directed to adapt the commercial hardware and systems to geothermal applications. Engineering and economic trade studies will be conducted to determine benefit/costs and optimum design. Some of the important options are:

- A reinjection well vs. a reinjection pipeline to KHS #1 well.
- Geothermal hydroponic-sprouted grain raising as a feed constituent vs. purchase of barley sprouts from malting processes vs. no sprout content in the feed.
- Geothermal absorption refrigeration for various cooling loads, i.e., sprout raising, sprout processing, space cooling.
- Geothermal wall heaters vs. added insulation, and similar evaluations for all other space heat loads.
- Geothermal methane generation vs. protein extraction vs. conventional commercial sewage treatment facilities.

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[1] Mountain Home Geothermal Project, DOE Contract DE-AC07-78ET28442, 1978.

Final design will be based upon economic analysis, including consideration of geothermal and other tax incentives and their impact on the rate of return to the investors. Emphasis will be placed upon simplicity and a straightforward commercial approach.

Status: The project was entering contract negotiation at the time of this writing.

Project Title: Klamath Falls YMCA 1-78
Geothermal Space/Water Heating

Location: Klamath County YMCA
Klamath Falls, Oregon

Principal Investigator: Brian C. FitzGerald, General Director, YMCA

Project Team:

- Klamath County YMCA Board of Directors
- E. E. Storey & Son Well Drilling
- Balhizer & Colvin, Engineers
- Alan Lee, Attorney at Law
- O.I.T., Geo-Heat Research Consultants
- Honeywell Control Systems

Project Objective:

To demonstrate the viability of geothermal energy used in a non-profit social service corporation. The project is a direct use retrofit for space and water heating.

Resource Data:

The YMCA is located over the Klamath KGRA, with a present well drilled to 2,016 feet. The production capacity is in excess of 500 gpm, at 110°F. The well is cased to 512 feet, leaving exposed 325 feet of shale material. Consequently, it is felt that considerable cold surface water is mixing with hotter water found at lower levels, which include 850, 950, 1,150, and 1,345 feet. Static water temperature at 1,350 is 146°F. Bottom rock temperature is 176°F.

Systems Design Features:

A second production well is being drilled, with greater production anticipated (casing will exceed 950 ft, closing off cold surface water). The first well will then be used for reinjection. Should we be unable to improve upon our 110°F resource with the #2 well, the following design specifications will be used:

1. 348 gpm peak load pump, with a Nelson variable drive (110 gpm).
2. Transmission line, 5-inch ID black iron bedded in insulation.
3. In building heat exchangers, hot water boosted by gas and several multi-zoned fan-coil units.
4. The current conventional system supplies approximately 64,000 therms per year. The geothermal system should replace 44,000 therms, primarily in heating the water in the pool, boosting domestic hot water, and general space heating.

5. Project life-cycle estimates indicate 50-year term savings net of $4,950,000.

Project Description:

This straightforward application seeks to space/water heat a private recreation facility with geothermal fluid. The long-term benefits can profoundly impact the quality of recreation services available. For example, at present our swimming pool costs $1.15 per hour to heat. Conservative estimates, for 10 years hence, indicate a cost of $7.50 an hour. Since we are the only indoor teaching facility available year-round for a population of 60,000 people, this resource is critical. Future projections indicate an expense which would require this facility to be shut down. To a lesser degree, many other facilities and services could become so expensive to maintain as to be prohibitive. An alternative energy source then becomes critical to the life of our organization.

Status:

We have completed our environmental study and predesign phases. Work on our second well is in progress. It is interesting to note that our second well, drilled 520 ft from our first well, is encountering virtually the same aquifer formations found in the first. We will improve our position through more accurate use of technical information. With Department of Energy support, i.e., teaming arrangements, management plan, technical assistance, and bid specification packages, we have grown as customers and general contractors. There does exist a learning curve which can bring the process within the capability of a small private social service agency.

Although technical advances are being made, problems exist in the fields of general contracting (a scarce resource in the geothermal field), accounting, legal, and engineering. Since the field is relatively new, general business support has no precedent and little in the way of "automatic" business procedures. As potential users become more sophisticated and knowledgeable as to their needs, we are able to turn a sellers market into a buyers market--a process which must occur if geothermal energy is to be put to widespread use.

Project Title: Klamath Falls Geothermal Heating District

Location: Klamath Falls, Oregon

Principal Investigator: Mr. Harold Derrah, Assistant City Manager, (503) 884-3161

Project Team:

- City of Klamath Falls
- LLC Geothermal Consultants
- Robert E. Meyer Consultants

Project Objective:

To provide for initial setup of the geothermal heating district. Initial project will provide heating to 14 city, county, state, and federal buildings.

Resource Data:

Project is within the Klamath Falls KGRA. The geothermal description of the KGRA is as follows:

> In general, the fractured basalts and cinders are highly porous, being capped by a nearly impervious zone of fine grained, lacustrine, palogonite tuff sediments and diatomite, referred to as the "Yonna formation" and locally as "chalk rock". This formation, Tst on the geologic map, is estimated to be 30 to 150 feet thick in the urban area. It is also interbedded with sandstone or siltstone and fine cinders.

The predicted temperatures for the project range from 200°F to 240°F, and the wells will be drilled to the approximate depth of 1,000 feet. Reported temperatures within the Klamath Falls area have been as high as 250°F, with flows being produced up to 700 gpm.

System Design Features:

The project will involve two production wells, each approximately 1,000 feet deep, with temperatures ranging up from 200 to 240°F. There will also be a reinjection well for injection of geothermal fluids after passing through a central heat exchange facility. The heat exchanger facility is designed for plate exchangers, with the following specifications:

Type - Single pass with 150 316 sst plates, EPDM gaskets

Size - 9'3" long x 1'7" wide x 5' high.

Geothermal side - 219°F, inlet
176°F, outlet
4.3 psig pressure drop

350 gpm flow
(1,000 gpm maximum flow)

Secondary side - 200°F, outlet
160°F, inlet
3.7 psig pressure drop

378 gpm flow
(1,000 gpm maximum flow)

The estimated ΔT is 40°F. The estimated heating peak requirements for the initial project is 15.3 x 10^6 Btu/hr. The system will involve the use of concrete conduits to allow for future expansion, longer life expectancy of the distribution system, and lower maintenance costs. Again, with the use of the conduit, expansion will be greatly facilitated. The estimated annual savings for the initial heating of the 14 buildings is $262,000, current dollar value.

Project Description:

The project is initially for the establishment of a heating district that will provide geothermal heating to 14 city, state, county, and federal buildings. The intent of the project is to be the initial stage for a total urban heating district. Included within the project is the development of a master plan for the distribution lines, production sources, storage requirements, and peaking facilities. The initial project will involve two geothermal wells, a distribution line appropriately over-sized for future development, central heat exchange facilities, and domestic reheated water distribution system to the initial buildings. The geothermal distribution line will be placed in concrete conduit, which will provide for future growth, increased life expectancy, easy access for future maintenance and repairs, and also provide better assurance that groundwater will not provide a deteriorating factor to the life expectancy of the pipe. The project also envisions that after the water circulates through the plate heat exchange facility and has transferred the energy through the plate heat exchanger, the geothermal water will be reinjected to the geothermal reservoir for reheating and reuse. The initial project is to generate a peak heating load delivery of 15.3 x 10^6 Btu, with 756 gpm of estimated temperature of 200°F. The total estimated cost of the project is $1.4 million, with approximately 75% financed by the Department of Energy and the remaining match generated by local sources. The estimated savings in relation to natural gas costs is $262,000 per year.

Status:

At the date of this paper, the project is currently in the drilling status, with the completion of the conceptual design report and also an acceptable environmental report. As of the date of this report, the well has been drilled to 250 ft, and temperatures are at 137°F. It is envisioned that by the time this paper is presented one well will have been completed to approximately 1,000 ft, and the results of that particular well can be made available at that time. To date, the temperature gradient received in constant monitoring of the well within the range of 150 to 250 ft was approximately 1°F per 7 to 8 feet of drilling depth. From all indications at this time, the well should prove out to at least the specifications drawn for the project.

Project Title: Madison County Geothermal Project

Location: Rexburg, Idaho

Principal Investigator: Dr. J. Kent Marlor, Chairman, Madison County Energy Commission, (208) 356-3431

Project Team:

- Madison County; Kent Marlor, Program Director
- American Potato Company; Eugene F. Berry, Deputy Program Director
- Energy Services, Inc.; Dr. Jay F. Kunze, Project Manager

Project Objective:

To demonstrate the economics and feasibility of using a low-temperature geothermal resource for food processing and space heating applications.

Resource Data:

Madison County and Rexburg are at the eastern edge of the Snake River Plain, a plain that has recently been characterized as a young volcanic rift. Northeast trending faults, concentrated along the plain boundaries, are the source of many hot springs. The Snake River Plain has been the site of intense bimodal basalt-and-rhyolite for the last ten million years. The youngest eruptions (Craters of the Moon and Cedar Butte) apparently occurred as recently as 1,625 years ago.

Extensive exploration data include: reconnaissance-level geologic mapping and gravity surveys, electrical resistivity surveys, groundwater investigations, geochemical analyses of area wells, and a heat flow of 4 HFU.

Further indications of warm water not far below the surface exist in the Newdale area (9 miles northeast of Rexburg), with shallow (< 500 foot) wells producing 105°F and 97°F water. In the immediate vicinity of Rexburg there are several shallow wells, with temperatures in the 60's, the most promising of which is a 460-foot well with a surface temperature of 70°F.

In consideration of the available data, a geothermal resource of 350 to 450°F is believed to exist at a depth of 8,000 to 9,000 feet below the Snake River Plain. The target depth of the production well will be between 5,000 and 7,000 feet, to encounter a resource of at least 250°F.

Project Description:

Two 1,500-ft hydrological test wells and production wells at 3,000 and 6,000 feet are planned at this time. The deep well will supply 250°F water to American Potato Company for use in food processing.

Two major heat exchangers, using fresh water, will discharge into blanching units used for boiler makeup water and supply heat to belt dryers, secondary dryers, and heat-filtered air entering the plant. Madison County will then receive the partially spent water from the processing plant, at 190°F, and will supplement it with water from the 3,000-ft well, if required. Madison County will use the water for space heating purposes, using conventional heat exchangers. The annual savings in gas and oil would amount to an equivalent of 470 billion Btu (approximately 140 million kWh) by changing to geothermal energy.

Final design will be based upon the temperature of resource encountered, flow rate, and economic analysis of these factors.

Status:

The two 1,500-ft exploratory wells are being drilled at the time of this writing.

Project Title: Direct Utilization of Geothermal Resources Field Experiments at Monroe, Utah

Location: Monroe City, Utah

Principal Investigator: Mr. Duane Nay, Mayor, (801) 527-3511

Project Team:

- Monroe City, Utah
- Terra Tek, Inc.

Project Objective:

Utilize geothermal fluids from source of local hot springs to heat high school, city hall, and fire station. Install nucleus of district heating system for private and commercial usage.

Resource Data:

Sevier fault on Monroe-Red Hill KGRA Hot Spring, discharge 230 gpm at 148 to 165°F. Aquifer temperature is 169°F at 500 feet, increasing to 179°F at 1,400 feet.

System Design Features:

One production well	=	1,471 feet
One injection well	=	800 feet (estimate)
Geothermal fluid temperature	=	167°F
Circulating water temperature	=	155°F
Well pump capacity	=	650 gpm
Circ. pump capacity	=	650 gpm
Load	=	17×10^6 Btu/hr (with some fossil peaking assistance)
Candidate pipeline materials	=	asbestos cement and fiberglass reinforced plastic

Project Description:

Monroe City is a community of 1,500 people, located 160 miles south of Salt Lake City, Utah. The local economy is based primarily on agriculture. The geothermal demonstration project currently underway in Monroe will explore the economic and technical viability of the application of a moderate-temperature resource for a district heating system. The project will entail the drilling of one production and one injection well. Geothermal fluid from the production well will be piped through a central heat exchanger and then to the injection well. The expected production temperature is 167°F, at 650 gpm. The initial buildings to be heated will be the South Sevier High School, the Monroe City Hall, the fire station, and a number of small stores and residences. The system will be capable of being expanded to include the major areas of Monroe, and it is estimated that the total system will be capable of a load of 12,000 kW.

The principal investigator is Mr. Duane Nay, Mayor of Monroe City. Terra Tek, Inc., Salt Lake City, Utah, has responsibility for implementing the project, under the direction of Mr. Roger Harrison.

Status:

A 1,471-ft production well has been drilled and flow rates up to 370 gpm and temperatures to 167°F have been obtained during preliminary pump testing. Preliminary system design and costing and planning for injection well drilling is underway. Further pump testing of the aquifer is also planned.

Project Title: Water and Space Heating for a College and Hospital by Utilizing Geothermal Energy at Corsicana, Texas

Location: Navarro College and Navarro Memorial Hospital Corsicana, Texas 75110

Principal Investigator: C. Paul Green, Institutional Development Director Navarro College, (214) 874-6501

Project Team:

- Navarro College, Corsicana, Texas - Prime Contractor and User Facility
- Navarro Memorial Hospital, Corsicana, Texas - Using Facility
- Radian Corporation, Austin, Texas - Geothermal Consulting Engineers
- H. H. Hardgrave, Corsicana, Texas - Drilling Consultant
- Ham-Mer Consulting Engrs, Austin, Texas - HVAC Engineers

Project Objective:

The objective of this project is to demonstrate the economic and technical feasibility of direct utilization of geothermal energy. To meet this objective, this project is designed to decrease the dependence of Navarro College and Navarro County Memorial Hospital on fossil fuel by making maximum use of the low-temperature geothermal resource for water and space heating.

Resource Data:

Well tests have produced sustained flow rates of 315 gpm of 125°F water, at about 5,300 ppm total dissolved solids. The producing zone is 2,400 to 2,600 feet below the surface. The source of the heat is faulting associated with the Ouchita fold belt, which outcrops in Arkansas and underlies much of central Texas. The Woodbine Formation is the groundwater reservoir that makes up the aquifer. Hydraulic interconnection of deeper and shallow formations provided by the Mexia-Talco fault system is the factor responsible for the area's low-temperature geothermal value.

System Design Features:

One 2,600-ft production well provides the required flow for this project. Flat-plate heat exchangers will be used to achieve maximum geoheat utilization and for ease of cleaning. Geothermal fluids will not be vented to the atmosphere so as to control corrosion and scaling phenomena. At peak winter heating periods, the geothermal heating system will deliver approximately one million Btu/hr to the college's Student Union Building (SUB), and about 3.5 million Btu/hr (peak) to the hospital water and space heating systems. This load is represented by a fluid temperature drop of 25°F at 315 gpm, and will reduce the college and hospital natural gas heating loads by 87 and 44 percent, respectively. The geothermal fluid will be disposed of by injection into a suitable horizon via a second well.

Project Description:

The purpose of this geothermal project is to retrofit a college SUB and county hospital space and water heating systems to use geothermal energy, thereby reducing their dependence on fossil fuels. The geothermal heating system will supply heat to the domestic water system, as well as the forced air heating and outside air preheating systems of the college SUB and hospital. At present, heat input to these systems is accomplished via steam provided by low-pressure, natural gas-fired boilers. These boilers will be maintained in place as backup and augmentation.

Readily available commercial piping, pumps, valves, controls, flat-plate heat exchangers, and insulation will be utilized. However, even though initial geochemistry has shown the Corsicana geothermal fluids to be relatively noncorrosive, a short series of field corrosion tests will reveal the most acceptable system materials.

The final phase is a one-year operational demonstration phase, during which potential geothermal users will be encouraged to visit and observe the geothermal heating system.

Status:

A submersible production pump has been set at 1,000 feet, and pumped at 315 gpm. Injection well drilling will commence in September 1979, and system preliminary design will begin in October 1979.

Project Title: Direct Utilization of Geothermal Energy for Food Processing at Ore-Ida Foods, Inc.

Location: Ore-Ida Foods Processing Plant, Ontario, Oregon

Principal Investigator: Mr. Robert W. Rolf, Director Technical Services, Ore-Idaho, Inc., (208) 336-6238

Project Team:

- Ore-Ida Foods, Inc.
- CH_2M Hill, Inc.
- GeothermEx, Inc.

Project Objective:

Locate and develop geothermal resource of 800 gpm at 320°F. Retrofit existing plant for potato processing, space heating, and hot potable water.

Resource Data:

Snake River Basin, (predicted) 320°F at 7,000 feet.

System Design Features:

Two Production Wells
One Injection Well
Central Heat Exchangers
Fluid Transmission Pipeline
Geothermal Fluid Temperature = 150°C (300°F)
Injection Fluid Temperature = 55°C (130°F)
Total Well Capacity = 800 gpm
Pipeline - Buried insulated steel
Maximum energy utilization via cascading
System Capacity $\sim 64 \times 10^6$ Btu/hr
Estimated Annual Fuel Savings - 97,200 MWh

Project Description:

Ontario, Oregon is located just across the Oregon-Idaho border, 57 miles northwest of Boise, Idaho. The existing Ore-Ida Foods, Inc. plant processes potatoes, corn, and onions. It is currently dependent on natural gas and oil for process heat. The plan for this demonstration program is to substitute geothermal energy for the potato processing heat and other heat loads of about 97,000 MWh annually (33.2×10^{10} Btu/yr).

Status:

An environmental report has been prepared which examines the impacts the project will have upon the environment and the Ontario area.

A seismic survey has been conducted to supplement existing geologic and geophysical data. Based upon all the data available, two production sites have been located on the Ore-Ida factory property. Drilling of the first production well commenced on August 19, 1979, and is expected to be at the target depth of 7,000 feet in 45 to 60 days.

The preliminary system design is underway. Equipment and material selections are being made and piping and heat exchanger locations are being laid out. Final design is expected to commence in late 1979.

Project Title: Pagosa Springs Geothermal Distribution and Heating System

Location: Pagosa Springs, Colorado

Principal Investigator: Fred A. Ebeling, Planning Administrator (303) 264-5851

Project Team:

- Town of Pagosa Springs
- Archuleta County
- School District 50 Joint
- Coury and Associates, Inc.

Project Objective: To provide the community with a means of using its natural hydrothermal resource for space heating at minimal cost to users and reduce local dependency upon fossil fuels. This project will determine the best methods of utilizing the hydrothermal resource, demonstrate the practicability of community space heating systems, and provide the basis for future expansion.

Resource Data: The geothermal resource in Pagosa Springs has been used on an individual basis since the early 1900's. Since then, nearly 30 wells have been drilled for heating and recreation purposes. These wells are drilled to depths of less than 500 feet and produce waters ranging in temperature from 130° to 170°F. The water quality of the resource is highly site specific. Some of the wells produce warm water which nearly meets the national drinking water standards. Others contain higher concentrations of dissolved solids similar to those of the production formation, the Mancos Shale.

System Design Features: Flow from several existing wells in the town can be used to supply the entire heating needs of the system. It is expected that with proper pretreatment the geothermal fluids can be pumped through the distribution system directly to the individual users. The geothermal fluid will then be collected and returned to a central location for either reinjection or surface discharge to the San Juan River, depending on the water chemistry. In the past, all geothermal fluids, including the natural hot springs, have been discharged to the San Juan River.

It is estimated that an average flow rate of 500 gpm is required for the system. A ΔT of 25° is anticipated at design conditions. The estimated annual cost for the fossil fuels to be replaced by the geothermal system is about $70,000.

Project Description: The Pagosa hot springs have been used for therapeutic purposes since prior to the coming of the white man. In this century, the underground reservoir has been tapped by wells and the hot water used for heating purposes in a relatively unsophisticated manner, which has presented corrosion and scaling problems. Characteristics of the resource have never been quantified--area, depth, source of heat, pressures, temperatures, water quantity, recharge mechanisms, specific geology, etc. The first phase of this project is to quantify characteristics of the geothermal reservoir. This provides a basis for determining its potential applications and the design of a system for practical utilization.

Actual construction and placing the system in operation is scheduled for completion by late 1980. The Town of Pagosa Springs will then operate and maintain the system. Operational data will be collected to allow ongoing evaluation of the system, to gain further knowledge concerning the resource characteristics and potential future capabilities.

At present, the project is in the early stages and specifics are not as yet determined. In general, all public buildings in the town (courthouse, Town Hall complex, schools, etc.) will be heated using geothermal energy. Location of these buildings will basically determine routing of the distribution piping. Other buildings, commercial or residential, which can logically be served from the distribution pipelines, may tap on. The piping will be located along easements, alleys, or streets provided by the town, county, or school district. User fee arrangements have yet to be determined.

Several options are available for the heat distribution medium. The hot geothermal water may be used directly from the underground reservoir or it may be chemically treated to counteract corrosive and/or scaling difficulties. Or, a closed, fresh water loop may go to the user facilities after heating by a heat exchanger, which isolates the geothermal water.

Options also exist for access to the subsurface geothermal reservoir. Existing, relatively shallow, wells, may be used. Or, new wells may be drilled to tap intermediate depths. Or, a combination of wells may be used. Final system design will involve consideration of several interdependent factors for optimum practicality.

The total cost of the project is estimated at $1,003,000, of which DOE will provide $779,000 and local community $224,000. The amount shared by the local community is comprised of in-kind contributions of wells, rights-of-way, easements, and work by local people. The Town of Pagosa Springs has been designated as the local lead entity by its partners, Archuleta County and the School District. Local control is, by agreement among the three entities, handled by an advisory committee consisting of interested and qualified citizens.

Status: The Environmental Report and the resource evaluation flow-test plan have been submitted to DOE-ID for review. Well monitoring equipment is being installed. A file of existing hydrological and geological data has been compiled. The project is coordinating with appropriate regulatory agencies and a survey of prospective users has been conducted. The conceptual design is ready for review.

Public meetings, news releases, and radio interviews have been used to keep the public informed. General public attendance at the Advisory Committee meetings is increasing.

Project Title: Multiple Use of Geothermal Energy at Moana KGRA

Location: Reno, Nevada

Principal Investigator: Dr. David J. Atkinson, President
Hydrothermal Energy Corporation
(702) 323-2306; (213) 654-6397

Project Team:

- Hydrothermal Energy Corporation, Developer and Heat Supplier
- Oak Grove Investors, Principal Heat User
- S.A.I. Engineers, Engineering Design and Construction
- W. L. McDonald & Sons, Drilling
- William E. Nork, Inc., Logging and Testing

Project Objective:

Thermal waters of the Moana KGRA in Reno have been used over several decades for heating buildings and swimming pools.

We shall use these waters for heating space and domestic hot water in the Sundance West apartment complex nearby.

To increase utilization of available heat and aid disposal of cooled geothermal fluids, we shall add whichever auxiliary uses prove most feasible after space and water heating is completed.

Resource Data:

The resource at Moana KGRA underlies part of southern Reno, though its exact limits have not been defined. Cool or cold water wells surround the general area of thermal water, but these wells are not spaced closely enough to map a boundary.

Geologic conditions are relatively simple. Valley fill in the area is generally 600 to 2,000 ft thick and consists of very young gravels, sands and clays. The hot water presently used at Moana comes mostly from shallow aquifers in this sequence, usually below a characteristic blue clay aquiclude.

Below this valley fill are Tertiary volcanics, principally andesite. Gravity data provide a straightforward indication of depth to this volcanic "basement", and, when combined with a detailed structural analysis, show that the shallow hot water reservoirs in the valley fill overlie part of a clearly defined upfaulted block.

Fault and fracture patterns show three main sets trending approximately N, N 40° E and N 35° W. The sense of relative displacement on these faults suggests they are conjugate shears (N 40° E and N 35° W), bisected by extension fracturing and normal faults that trend north.

Fracture zones and intersections in the volcanic basement may provide the best targets for high flow rates in our production wells.

Temperatures in some existing wells are close to boiling point, but more usually are in the range 140 to 190°F, with only about 1,100 ppm total dissolved solids.

System Design Features:

Two production wells about 1,000 ft deep will be drilled near the apartment complex where the heat will be used. The geothermal fluids will be piped underground to newly installed shell and tube heat exchangers in the existing boiler rooms.

The present heating system, using circulating hot water, was specifically intended by the creator and designer of the apartment complex, Mr. Larry Freels, to take advantage of the local geothermal energy. The task of retrofitting will accordingly be relatively straightforward. The existing natural gas boilers will be retained both as permanent backup and to handle peak loads.

Temperature drop in the geothermal water will be about 60°F. An average flow of about 70 gpm (180°F) will be needed to supply the major part of the heating load, which is 176,000 therms annually. Peak flow will probably be about 250 gpm.

The saving of fossil fuel energy (about 3.5×10^{11} Btu over twenty years) is quite significant, and will be increased by auxiliary uses of the heat remaining in the geothermal water after space and water heating.

Project Description:

The first stage of the project involves environmental clearances and obtaining the numerous permits that are required.

Then, by integrating data on geology, hydrogeology, geochemistry, geophysics, economics, and engineering, we shall select the first well site and design the well.

We shall drill a test production well, log it, and test selected intervals for flow rates and temperature. From the results, we shall design the well completion to maximize heat extraction.

From results of the pump tests, we shall finalize design of the production and distribution system, and the retrofit heat exchangers and related equipment.

Samples of the geothermal water will enable us to select, and obtain permits for, the most appropriate disposal method.

A second production well will next be drilled, utilizing the experience gained in the first.

Installation of the buried pipelines will follow, taking geothermal water into and out of the apartment complex boiler rooms. Heat exchangers will be installed in these, upstream of the present boilers where the cold return water enters after circulating through the buildings.

After testing and optimization and detailed analysis of the engineering results of the installation, the system will run on a routine commercial basis.

The best technically and economically feasible auxiliary applications will then be selected, and used to extract more heat from the geothermal water after the space and water heating load is handled.

An important aspect of the project is a program of public information to convey broadly how simple the concept of direct use of geothermal heat is, and exactly how this project was done, and what results we obtained.

Status:

At the time of writing, we are only four weeks into the project, and are working on the first stage. Drilling should begin before the end of this year.

Project Title: Geothermal Application of the Madison Aquifer for St. Mary's Hospital

Location: St. Mary's Hospital, Pierre, South Dakota

Principal Investigator: James Russell, St. Mary's Hospital Administrator (605) 224-5941

Project Team:

- St. Mary's Hospital
- Kirkham, Michael and Associates, Engineer
- Sherwin Artus, Reservoir Consultant
- Dr. J. P. Gries, Geologist

Project Objective:

Demonstrate that 106°F water can be used economically to heat buildings and also to preheat domestic hot water.

Resource Data:

The 2,100-ft well which taps the Madison aquifer is located on a vacant lot adjoining a residential neighborhood and across the street from the hospital complex. The site overlooks the Missouri River. Well test data indicate a static pressure of 480 psig maximum, and a flow of 375 gpm, with 27 psig, at 106°F.

System Design Features:

The system has been designed for 350 gpm flow at 105°F, producing 4,375,000 Btu/hr. The maximum supply water temperature out of the heat exchanger is expected to be 100°F. A corrosion and water quality report was completed by Dr. Howard and Dr. Carda of Rapid City, South Dakota. This report indicates that type 316 stainless steel is the recommended material for the thin wall plate fin-type heat exchangers.

Project Description:

(See attached well house and exchanger building schematic.) The system's three heat exchangers will provide heat for three existing hospital systems and will also serve the new hospital wing presently under construction. The existing hospital systems are: 1) space heat in existing fan coil units now used only for air conditioning; 2) space heat for the high volume of outside air (makeup air ventilation) that is required in some areas of a hospital; and 3) preheating of domestic hot water. The well is located across the street from the hospital. The heat exchangers will be located in a small building at the well site.

a) Heat Exchanger Design: Heat exchangers will be in accordance with the following design conditions:

No.	Function	Fluid		Flow gpm	Ent °F	Leav °F
1.	Building Heat	Geothermal	=	350	105	80
		Closed Loop Heating Water	=	350	75	100
2.	Preheat Dom HW utilizing geo-thermal discharge from exchanger #1	Geothermal	=	350	80	75
		Domestic Water	=	76	55	78
3.	Preheat dom HW (boost from #2 and full preheat when #1 is un-loaded)	Geothermal	=	97	105	70
		Domestic Water	=	76	55	100

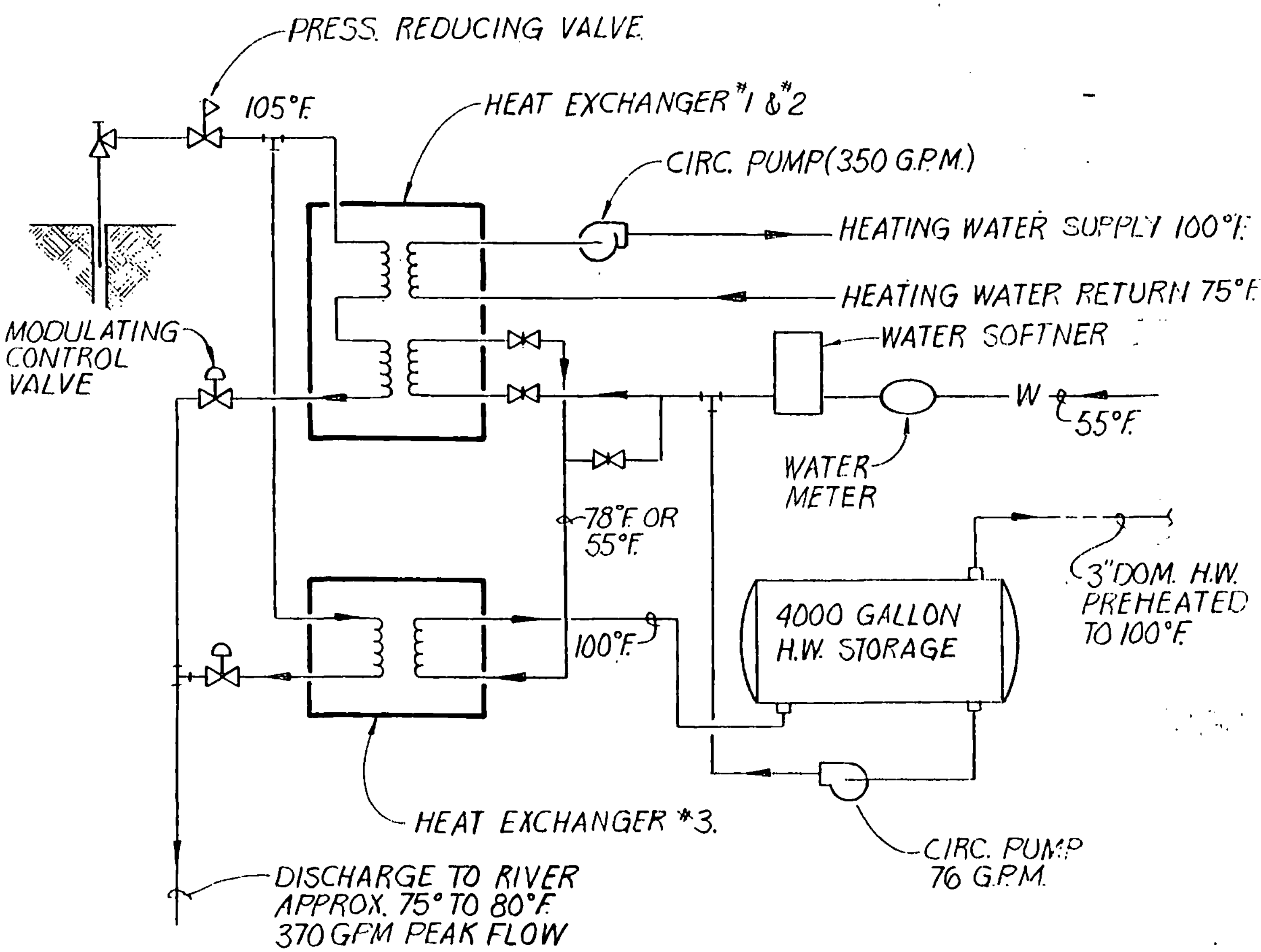
b) Makeup Air System Retrofit: A high volume of fresh air must be continuously introduced into certain areas of a hospital. This requires raising the outside air temperature to room temperature. Using the existing 6-row chilled water coil (15,650 CFM), the geothermal water supply flow would be 90 gpm at 100°F, and the leaving water temperature would be 64.5°F.

c) Fan Coil System Retrofit: The heating system in the existing hospital is basically steam perimeter radiation. The fan coil system was added to provide air conditioning. Chilled water at the average temperature of 50°F is circulated in the summer, to provide approximately 57 to 59°F supply air off the coils. In the winter time, 100°F water will be provided to these coils, to produce 87°F heated air, which is adequate to heat the spaces served during outside temperatures of approximately 2°F and above.

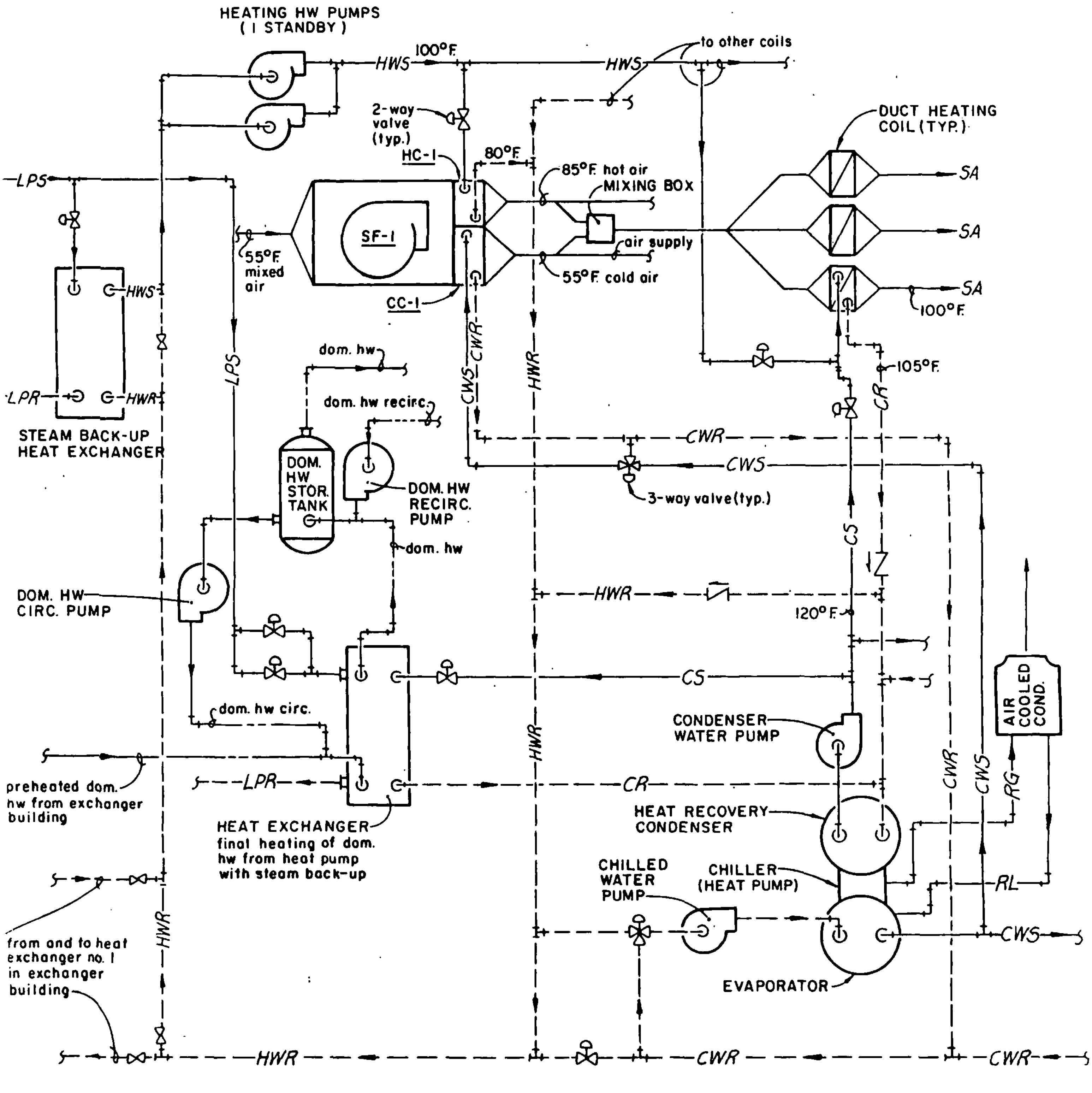
d) New Building Heating: 155 gpm from heat exchanger #1, representing 2,000,000 Btu/hr, will be available for use in the new hospital addition that is presently under construction. The new heating system is designed to utilize the geothermal heat source. (See new building heating schematic.)

Status:

The well is completed. Retrofit of the existing mechanical system will go out for bids at the end of August or early September.



3. Well House and Exchanger Building Schematic



NEW BUILDING HEATING SYSTEM SCHEMATIC

Project Title: Susanville Energy Project - Direct Utilization of Geothermal Energy

Location: North end of the Honey Lake Valley, Lassen County, California

Principal Investigator: Philip A. Edwardes
(916) 257-7259

Project Team:

- Aerojet Energy Conversion Company
- Donna Benner, Drury System Design
- Monte Koepf, Koepf and Lange, Engineering
- Fred Longyear, Lahontan, Inc., Technical Advisor
- Johan Otto, Carson Development, Management Information System/Construction Management
- Dr. Subir Sanyal, Energetics Marketing & Management Associates, Ltd., Reservoir Evaluation/Management

Project Objective:

To displace fossil fuels and create employment. Program will heat 17 public building complexes. Effluent fluids will be cascaded through a Park of Commerce. Ultimately the heating of all commercial buildings and private homes within the City of Susanville may be feasible.

Resource Data:

Most of the temperature gradient holes developed penetrated alternating layers of basalt and mud flow (ash flow) agglomerates. Some holes encountered alluvial conglomerates. Correlation of lithological strata from one hole to another indicates faulting. This confirms the surface evidence of extensive faulting in the area. Electrical logs through the basalt layers suggest fracturing at the upper and lower limits of the layers, indicating these basalt layers, as well as the agglomerates and conglomerates, may be potential reservoir units.

Ten holes were drilled within the city boundaries and its immediate surroundings, ranging in depth from 135 m to 640 m. Six existing private wells are also within the area. Temperatures varied between 35 and 75°C. Several holes, notably in the southwest portion of the reservoir, display marked temperature reversal, with depth of 100 to 150 m in the holes with the higher temperature. In the north part of the reservoir area, reversal takes place much deeper, and the temperature zone is also thicker.

Resource data to date suggest a temperature of 75°C possible, with individual well flow between 300 and 400 gpm; a flow rate of well over 2,000 gpm is considered feasible, pending final evaluation of BuRec resource work.

System Design Features:

It is projected that there will be 3 production wells, capable of 350 gpm, pumping from a depth of 150 m x 200 m at 72°C, producing 20,000,000 Btu/hr. Two reinjection wells are planned. Water quality data to date suggests total dissolved solids of less than 1,000 PPM and pH of 7-7.5. The heat requirement of the 17 building complex (320,000 sq ft) is 12,000,000 Btu/hr; the effluent fluid reaches the park at a temperature of 110°F. A heat pump will be incorporated within the system for peaking purposes, enabling further fossil fuel displacement, and also minimizing the necessity for further wells. A relatively limited use of heat exchangers is visualized in the retrofits. Heat exchangers will be utilized only where damage to the existing hardware could be caused by the geothermal fluids. In several cases, a direct hookup will be possible; in other buildings fan coils will be used.

The economic model allows for wells to be replaced at the rate of 25% every 7 years. Initial indications are that a price to the consumer of $2.75 per million Btu could be possible. This figure could dramatically change with full utilization of the effluent fluids by the Park of Commerce.

The main transmission lines are capable of an optimum flow of 2,000 gpm; the 12-inch transmission line will be insulated, and the 12-inch return line uninsulated.

Project Description:

The Susanville project envisions in its initial phases the development of a heating district to heat 17 public building complexes and to cascade the effluent heat through a Park of Commerce.

The City of Susanville, in 1974, recognized the necessity to hold down the escalating cost of heating to the local population and to create job opportunities (the local unemployment was reaching a peak of 20% in winter months). The existence of a resource had been identified and utilized in a limited manner from the 1920's; its extent and real potential was unknown. The city believed that it was beyond the capacity of private enterprise to establish and develop the resource, so by resolution of Council, expressed its intent to develop the resource potential on behalf of the maximum number of residents for their maximum benefit. It was because of this expressed intent that Public Law 94-156 was passed, and BuRec was authorized and funded by Congress to evaluate the resource potential on behalf of the city. This extensive program is currently ongoing and is proving to be successful in its objective.

It was deemed expedient that the initial development would address publically held buildings, thus spreading the cost savings benefits to the population in general. It was also anticipated that it would be easier to attract grant funds for this objective. The Park of Commerce would be developed concurrent with the heating district.

The potential for replicating the program in many western rural areas was identified, and this formed part of the basis of justification of the project.

In January 1978, a proposal was submitted to DOE which projected a DOE contribution of $2.4 million and a City share of $1.9 million. The program was expected to extend over a 33-month period. Phase I, the design and engineering effort, is currently under contract, with Phase II, the construction phase, hopefully under contract in time to develop the first production well in December 1979. The City and its team members believe they have the capacity to have fluid flow and utilization by December 1980.

The Park of Commerce is being promoted and developed independently of DOE, but, at the same time, the City is under contractural obligation to DOE to do so if deemed feasible. Currently the City is negotiating for land suitable for such a park (200 to 300 acres). It is the City's intent to secure options on behalf of its nominees (identified industry) but not to be involved in land purchases itself. The City will and has successfully identified potential long-term loan sources for the development of streets, utility reticulation, and sewerage system for the park; the repayment will come from the developers and operators within the park.

The City intends that the Park of Commerce will have an agricultural bias, feed mill, greenhouses, and confined animal raising units. Some heat augmentation of the residual effluent from the heating district will be necessary for refrigeration, air conditioning, and sterilization of wool, etc. Various alternate energy sources are being investigated (wood waste, city refuse, and methane from the animal fattening units) to accomplish this.

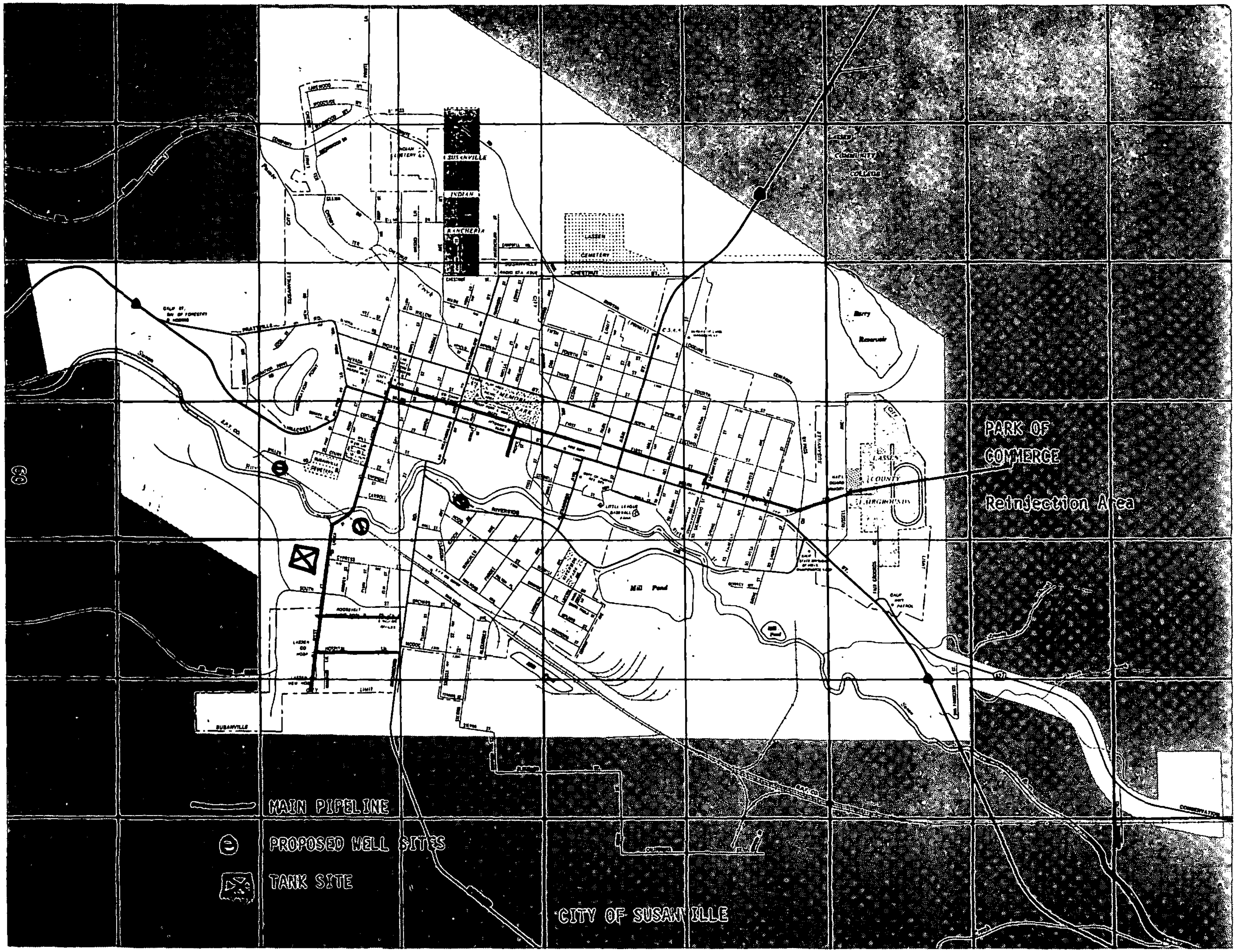
The intent of the City of Susanville is that eventually it will develop a heating district encompassing all buildings within the city. It is likely that a high percentage of Main Street commercial buildings could be heated by December 1981; progress to encompass single family homes could be considerably slower. It is recognized that the geothermal resource in itself may be insufficient for this ambitious program; however, the City believes that the cost effectiveness of utilizing wood waste from the nearby forest areas will allow it to continue in its objective of being self-sufficient in energy for heating purposes. Importantly, the City has unanimous support in its endeavor from the local population.

In recognition of the fact that a geothermal resource is finite by definition, the City introduced an ordinance to insure orderly and efficient utilization of the resource for the maximum benefit of the residents of the city. It is the declared intent of Council to make geothermal energy available to private enterprise at the lowest cost possible. The City's ownership of the total supply and distribution system will enhance this position.

Status:

Phase I: Under DOE contract since March 1, 1979 for design, engineering, and resource evaluation. Major permitting has been completed. Initial design criteria has been transmitted to engineers. Program to date on schedule.

Phase II: Construction period, hopefully, will commence in December 1979, with development of the first production well. Pipeline and storage tank construction is anticipated to commence in May 1980, concurrent with retrofit. First flow to part of the system is anticipated by December 1980. Completion and checkout is anticipated by June 1981.


PARK OF
COMMERCE
Reinjection Area
MAIN PIPELINE
PROPOSED WELL SITES
TANK SITE
CITY OF SUSANVILLE

Project Title: Direct Utilization of Geothermal Energy for Space and Water Heating at Marlin, Texas

Location: Torbett-Hutchings-Smith (THS) Memorial Hospital Marlin, Texas 76661

Principal Investigator: J. D. Norris, Jr., Administrator THS Memorial Hospital, (817) 883-3561

Project Team:

- THS Hospital, Marlin, Texas - Prime Contractor and User Facility
- Radian Corporation, Austin, Texas - Geothermal Consulting Engineer
- Ham-Mer Consulting Engineers, Austin, Texas - Economic Evaluation, Operation and Maintenance
- Layne Texas Company, Dallas, Texas - Well Drilling
- Spencer Associates, Austin, Texas - Architectural

Project Objective:

The objective of this project is to demonstrate the economic and technical feasibility of direct utilization of geothermal energy. To meet this objective, this project is to augment the space and water heating requirements of the THS Memorial Hospital in Marlin, Texas, with geothermal energy.

Resource Data:

Well tests have produced flow rates of over 300 gpm of 153°F water, at about 4,000 ppm total dissolved solids. The producing zone is 3,615 to 3,885 feet below the surface. The source of the heat is faulting associated with the Ouchita fold belt, which outcrops in Arkansas and underlies much of central Texas. The coarser-grained sandstones (especially the Houston member of the Travis Peak Foundation) are the groundwater reservoir that defines the aquifer. The factor which is responsible for the area's geothermal value is the hydraulic interconnection of deeper and shallow sandstones provided by the Mexia-Talco fault system.

System Design Features:

One 3,885-foot production well will provide more than the required flow for this project. Flat-plate heat exchangers will be used to achieve maximum geoheat utilization and for ease of cleaning. Geothermal fluids will not be vented to the atmosphere so as to control corrosion and scaling phenomena. At peak winter heating periods, the geothermal heating system will deliver approximately 2.5 million Btu/hr to the hospital heating load. This load is represented by a fluid temperature drop of 45°F at 110 gpm, and will reduce the THS Hospital natural gas consumption by 85 percent. The geothermal fluid disposal system design is yet to be defined.

Project Description:

The purpose of this geothermal project is to retrofit the 130-bed hospital space and water heating systems to use geothermal energy, thereby reducing its dependence on fossil fuels. The geothermal heating system will supply heat to the hospital domestic water system, as well as to the 130°F space heating and outside air preheating systems. At present, heat input to these systems is accomplished via steam provided by a low-pressure, natural gas-fired boiler. This boiler system will remain in place as backup and augmentation.

Readily available commercial piping, pumps, valves, controls, flat-plate heat exchangers, and insulation will be utilized. However, even though initial geochemistry has shown the Marlin geothermal fluids to be relatively noncorrosive, a short series of field corrosion tests will reveal the most acceptable system materials.

The final phase is a one-year operational demonstration phase, during which potential geothermal users will be encouraged to visit and observe the geothermal heating system.

Status:

A 3,885-foot deep production well was completed and tested in July 1979. Preliminary heating system design is underway, and the Preliminary Design Review is anticipated to be held in November 1979.

Project Title: Floral Greenhouse Industry Geothermal Energy Demonstration Project

Location: 567 West 90th South, Sandy, Utah (15 miles south of Salt Lake City center)

Principal Investigator: Ralph M. Wright, Chairman of the Board
Utah Roses, Inc. (801) 295-2023

Project Team:

- Utah Roses, Inc., Sandy, Utah
- Energy Services, Inc., Idaho Falls, Idaho

Project Objective:

To demonstrate to the public the potential offered by geothermal space heating in a highly populated area, by using geothermal heating in a commercial application.

Resource Data:

A large area of the southeast portion of the Salt Lake Valley appears to be underlaid by a source of warm water. Crystal Hot Springs, approximately 6 miles south, flows hot water at 180°. Several wells in the area of Utah Roses have shows of warm water, including one within 100 yards of the proposed site, which has 93°F water at 875 feet. There is evidence of a fault running east-west at our location. These indications, plus the normal temperature gradient, lead to an expectation of water at 150 to 180°F at 3,000 to 4,000 feet. Since no drilling below 1,000 feet has occurred in our area, flow rates and actual temperatures are difficult to project until actual drilling can take place.

System Design Features:

One well is projected to a 4,000-ft depth, with possibly a second well for reinjection. However, the primary plan is to discharge the water into a nearby irrigation canal, if water quality is high enough. Heat exchange will be dependent on water quality and temperature. Plans are to keep the water under pressure and run it through water/air heat exchangers in the greenhouse, with the air being distributed through polyethelene tubes located near ground level throughout the greenhouse. If sufficient flow and temperature are achieved, the entire heat load of the greenhouse will be taken over by geothermal, with an annual saving of $100,000.

Project Description:

A 4,000-ft well is to be drilled at the present site of Utah Roses, Inc., a 250,000 sq ft greenhouse which is producing cut roses for the national floral market. If water of sufficient temperature and quantity is developed, the water will be used to heat the greenhouse, replacing the current natural gas/oil usage. Since Utah Roses is well-known in the floral industry, with two of its officers serving as officers in national floral trade associations, a considerable amount of publicity has been and will be generated for geothermal energy in an industry that has a high potential for using geothermal energy.

Status:

The Environmental Report has been prepared and approved, the well design is completed, and the bid package has been sent to prospective drilling contractors. It is planned to begin drilling during September.

Project Title: Direct Utilization of Geothermal Resources, Field Experiment at the Utah State Prison

Location: Draper, Utah; approximately 22.5 km (14 miles) south of Salt Lake City.

Principal Investigator: Jack Lyman, Director, Utah Energy Office, (801) 533-5424

Project Team:

- Utah Energy Office
- Utah Department of Social Services
- Utah State Building Board
- Utah Geological and Mineral Survey
- Terra Tek, Inc.

Project Objective:

To demonstrate the economic and technical viability of using a low-temperature geothermal resource in a variety of direct applications at the Utah State Prison.

Resource Data:

The site of the Utah State Prison PON is located in the southern portion of Salt Lake County, near Draper, Utah. Located just west of the Wasatch Range, the resource is within the Basin and Range physiographic province. The surface expression of the resource is known as Crystal Hot Springs, and is located on the northern flank of the East Traverse Mountains; a horst that is intermediate in elevation, between the Wasatch range to the east and the valley grabens to the north and south. The northern flank of the Traverse Range is bound by a series of northeast striking normal range front faults, having a combined displacement of at least 900 m (3,000 ft). The thermal springs are located between two of the range front faults that are intersected by a north-northeast striking fault. Only 25 meters (80 ft) of basin alluvial material covers the bedrock surface in the immediate vicinity of the springs. The maximum measured temperature of the resource is 86°C, and total surface discharge is approximately 35 l/sec. (1.25 ft^3/sec). The total dissolved solids content of the spring water is on the order of 1,500 mg/l.

System Design Features:

The preliminary system design for the Utah State Prison minimum security block includes plans for a space heating system, with a design load of 750 kW and a culinary water heating system with a design load of 500 kW. The inlet temperature for both systems is 90°C; the outlet temperature is 75°C (T = 15°C) for the space heating system and 65°C (T = 25°C) for the water heating system. Together, these systems will require a design flow rate of 17 kilograms per second (270 gpm) and an average requirement of 5 kilograms per second (80 gpm).

One production well and one injection well (if needed) are anticipated. Siting of the production hole will be based on the results of a detail gravity survey and thermal test hole drilling program. The injection well will be drilled in the event that water quality parameters preclude surface disposal, in which case the water could be disposed of in near surface alluvial aquifers.

The conversion of the minimum security block to a geothermal heat source will result in a 10 to 25% reduction in the prison's use of natural gas and fuel oil.

Project Description:

The project is designed to provide geothermal space and water heating systems for the minimum security block of the Utah State Prison. Future expansion of the project may include the extension of these services to other buildings, as well as the use of the thermal water for a variety of other direct applications at the prison dairy and slaughterhouse. Where possible, the geothermal fluids may be used to heat greenhouses and irrigate crops.

Status:

The first phase of the project has just begun. The detailed gravity survey is in progress and plans are being made for a test hole drilling program.

Project Title: Geothermal Heating of Warm Springs State Hospital, Montana

Location: Warm Springs State Mental Hospital, Deer Lodge County, Montana

Principal Investigator: M. Eugene McLeod, Project Manager
(406) 494-6420; FTS-587-6402

Project Team:

- MERDI, Inc.
- Energy Services, Inc.
- CH_2M Hill, Inc.
- State of Montana

Project Objective:

The objective of this program is to develop the geothermal resource at Warm Springs for domestic water and space heating.

Resource Data:

The Deer Lodge Valley is within the Northern Rocky Mountains physiographic province and is bordered on the east by low (generally below 8,000 ft), rolling hills known locally as the Deer Lodge Mountains. The western boundary consists of the rugged, glaciated Flint Creek Range, with elevations up to 10,171 feet (Mount Powell). The Anaconda Range encloses the valley on the south and the Garnet Range is located to the north. The valley consists of high terraces that slope downward from the mountain peaks and terminate above low terraces that grade into the Clark Fork flood plain, which forms the valley floor. This basic topography has been modified by the formation of coalescent fans and glacial moraines at the mouths of the tributary valleys and canyons. This modification is especially evident on the west side of the valley.

The valley is predominately filled with Tertiary sedimentary strata derived from the surrounding mountains. This strata has a diverse lithology composed primarily of interbedded limestone, shale, sandstone, volcanic debris, and sand. It appears to be at least 1,600 feet thick northwest of Deer Lodge and is overlain by 300 feet of Pliocene channel sand and gravel. The strata also contains bentonitic clay beds, pebble conglomerate, cobbles, and granitic debris. The maximum thickness of the valley fill may be as much as 5,500 feet east of Anaconda. The Tertiary valley fill is estimated to be approximately 2,200 feet thick in the area of Warm Springs.

The valley is a closed structural basin produced by faulting along the boundaries. The sedimentary beds in the mountains surrounding the valley have been folded and faulted. Extensive thrusting has occurred within the Flint Creek Range and several northeast-southwest trending anticlines and synclines are evident. Several faults within the mountains to the south and west are traceable into the valley, although direct evidence such as faultline scarps are lacking. The spring currently discharges water at 171°F, with a dissolved solid content of 1,250 mg/l. The source of the geothermal water is attributed to deep circulation in fault zones with a probable limestone matrix.

System Design Features:

The present drilling plan calls for drilling one production well to a depth of 1,250 ± 250 feet. Disposal of the spent geothermal water will be utilized for the creation of a wetlands for waterfowl, eliminating the need for an injection well.

The engineering design will accomplish two well-defined tasks at Warm Springs, depending upon the well flow.

1. Heating of domestic hot water; and
2. Space heating of at least two buildings.

Both heating tasks will be accomplished independently by using plate-type counterflow heat exchangers, each task having its own exchanger. The domestic hot water heating requirements are estimated to be 100 gal/min of 170°F geothermal fluid, with a ΔT of 60°F; the space heating requirements are estimated to be 200 gal/min at the same ΔT.

Project Description:

The geothermal demonstration plan includes drilling one production well to a depth of approximately 1,250 feet. The expected production temperature is 170°F, at 300 gpm. The plan is to substitute geothermal energy for domestic hot water requirements and partial space heating of the Warm Springs facility, which is currently dependent upon natural gas. The water will be pumped through plate-type heat exchangers, with approximately 490 Btu per gallon of useful energy extracted in the process. The water will be discharged at 110°F into Montana Department of Fish, Wildlife, and Park's ponds adjacent to the hospital, for the creation of wetlands for migratory waterfowl.

Status:

The Environmental Report has been prepared and reviewed by DOE. The geophysical survey conducted by the Montana College of Mineral Science consisted of gravity and resistivity surveys. The geophysical survey was supplemented by a review and interpretation of existing geologic and geophysical literature by Roger Stoker. The well site has been determined and well drilling is scheduled in September 1979.

The legal review of state regulations for geothermal exploration and drilling has been completed. Applicable permits have been acquired. MERDI is presently working with various state and federal agencies for the creation of waterfowl wetlands, using the disposed geothermal water.

Appendix 1

Industrial Process Heat Requirements at Temperatures 300°F (149°C) and Below

Industry - SIC Group	Application Temperature Requirement °F	(°C)	Process Heat Used for Application 10^{12} BTU/Yr	(10^{12} KJ/Yr)
Group 10				
1. Copper Concentrate - 1021				
Drying	250*	(121)	1.7	(1.6)
Group 12				
2. Bituminous Coal - 1211				
Drying (including lignite)	150-250*	(66-104)	18.0	(19.0)
Group 14				
3. Potash - 1474				
Drying Filter Cake	250*	(121)	1.03	(1.09)
Group 20 - Food & Kindred Products				
4. Meat Packing - 2011				
Sausages and Prepared Meats - 2013				
Scalding, Carcass Wash and Cleanup	140	(60)	43.7	(46.1)
Edible Rendering	200	(93)	0.52	(0.55)
Smoking/Cooking	155	(68)	1.16	(1.22)
5. Poultry Dressing - 2016				
Scalding	140	(60)	3.16	(3.33)
6. Natural Cheese - 2022				
Pasteurization	170	(77)	1.28	(1.35)
Starter Vat	135	(57)	0.02	(0.02)
Make Vat	105	(41)	0.47	(0.50)
Finish Vat	100	(38)	0.02	(0.02)
Whey Condensing	160-200	(71-93)	10.2	(10.8)
Process Cheese Blending	165	(74)	0.07	(0.07)
7. Condensed and Evaporated Milk - 2023				
Stabilization	200-212	(93-100)	2.93	(3.09)
Evaporation	160	(71)	5.20	(5.48)
Sterilization	250	(121)	0.54	(0.57)
8. Fluid Milk - 2026				
Pasteurization	162-170	(72-77)	1.44	(1.52)

Appendix 1 (continued)

Industrial Process Heat Requirements at Temperatures 300°F (149°C) and Below

Industry - SIC Group	Application Temperature Requirement °F	(°C)	Process Heat Used for Application 10^{12}BTU/Yr	(10^{12} KJ/YR)
9. Canned Specialties - 2032				
Beans				
Precook (Blanch)	180-212	(82-100)	0.40	(0.42)
Simmer Blend	170-212	(77-100)	0.24	(0.25)
Sauce Heating	190	(88)	0.20	(0.21)
Processing	250	(121)	0.38	(0.40)
10. Canned Fruits and Vegetables - 2033				
Blanching/Peeling	180-212	(82-100)	1.88	(1.98)
Pasteurization	200	(93)	0.15	(0.16)
Brine Syrup Heating	200	(93)	1.02	(1.08)
Commercial Sterilization	212-250	(100-121)	1.67	(1.76)
Sauce Concentration	212	(100)	0.44	(0.46)
11. Dehydrated Fruits and Vegetables - 2034				
Fruit & Vegetable Drying	165-185	(74-85)	5.84	(6.16)
Potatoes				
Peeling	212	(100)	0.33	(0.35)
Precook	160	(71)	0.47	(0.50)
Cook	212	(100)	0.47	(0.50)
12. Frozen Fruits and Vegetables - 2037				
Citrus Juice Concentration	190	(88)	1.33	(1.40)
Juice Pasteurization	200	(93)	0.27	(0.28)
Blanching	180-212	(82-100)	2.26	(2.38)
Cooking	170-212	(77-100)	1.41	(1.49)
13. Wet Corn Milling - 2046				
Starch Dryer	120*	(49)	3.03	(3.20)
Steepwater Heater	120	(49)	0.77	(0.81)
Sugar Hydrolysis	270	(132)	1.89	(1.99)
Sugar Evaporator	250	(121)	2.74	(2.89)
Sugar Dryer	120*	(49)	0.16	(0.17)
14. Prepared Feeds - 2048				
Pellet Conditioning	180-190	(82-88)	2.28	(2.40)
15. Bread and Baked Goods - 2051				
Proofing	100	(38)	0.84	(0.89)

Appendix 1 (continued)

Industrial Process Heat Requirements at Temperatures 300°F (149°C) and Below

Industry - SIC Group	Application Temperature Requirement °F	(°C)	Process Heat Used for Application 10^{12} BTU/Yr	(10^{12} KJ/YR)
16. Cane Sugar - 2062				
Mingler	125-165	(52-74)	0.59	(0.62)
Melter	185-195	(85-91)	3.30	(3.48)
Defecation	160-185	(71-85)	0.44	(0.46)
Granulator	110-130	(43-54)	0.44	(0.46)
Evaporator	265	(129)	26.39	(27.84)
17. Beet Sugar - 2063				
Extraction	140-185	(60-85)	4.63	(4.88)
Thin Juice Heating	185	(85)	3.08	(3.25)
Thin Syrup Heating	212	(100)	6.68	(7.05)
Evaporation	270-280*	(132-138)	30.8	(32.5)
Granulator	150-200	(66-93)	0.15	(0.16)
Pulp Dryer	230-280*	(110-138)	16.5	(17.4)
18. Soybean Oil Mills - 2075				
Bean Drying	160	(71)	4.05	(4.27)
Toaster Desolventizer	215	(102)	6.08	(6.41)
Meal Dryer	300*	(149)	4.36	(4.60)
Evaporator	225	(107)	1.62	(1.71)
Stripper	212	(100)	0.30	(0.32)
19. Shortening & Cooking Oil - 2079				
Oil Heater	160-180	(71-82)	0.72	(0.76)
Wash Water	160-180	(71-82)	0.12	(0.13)
Dryer Preheat	200-270	(93-132)	0.60	(0.63)
Cooking Oil Reheat	200	(93)	0.32	(0.34)
Hydrogenation Preheat	300	(149)	0.37	(0.39)
20. Malt Beverages - 2082				
Cooker	212	(100)	1.53	(1.61)
Water Heater	180	(82)	0.53	(0.56)
Mash Tub	170	(77)	0.60	(0.63)
Grain Dryer	300*	(149)	9.18	(9.68)
Brew Kettle	212	(100)	3.98	(4.20)
21. Distilled Liquor - 2085				
Cooking (Whiskey)	212	(100)	3.16	(3.33)
Cooking (Spirits)	320	(160)	6.27	(6.61)
Evaporation	250-290*	(121-143)	2.32	(2.45)
Dryer (Grain)	300	(149)	1.94	(2.05)
Distillation	230-250	(110-121)	7.69	(8.11)

Appendix 1 (continued)

Industrial Process Heat Requirements at Temperatures 300°F (149°C) and Below

Industry - SIC Group	Application Temperature Requirement °F	(°C)	Process Heat Used for Application 10^{12} BTU/Yr	(10^{12} KJ/YR)
22. Soft Drinks - 2086				
Bulk Container Washing	170	(77)	0.21	(0.22)
Returnable Bottle Washing	170	(77)	1.27	(1.34)
Nonreturnable Bottle Warming	75-85	(24-29)	0.43	(0.45)
Can Warming	75-85	(24-29)	0.52	(0.55)
Group 21 - Tobacco				
23. Cigarettes - 2111				
Drying	220*	(104)	0.43	(0.45)
Rehumidification	220*	(104)	0.43	(0.45)
24. Tobacco Stemming & Redrying - 2141				
Drying	220*	(104)	0.50	(0.26)
Group 22 - Textile Mill Products				
25. Finishing Plants, Cotton - 2261				
Washing	200	(100)	15.4	(16.2)
Dyeing	200	(100)	4.5	(4.7)
Drying	275	(135)	22.2	(23.4)
26. Finishing Plants, Synthetic - 2262				
Washing	200	(93)	35.9	(37.9)
Dyeing	212	(100)	15.2	
Drying & Heat Setting	<275	(135)	23.2	(24.5)
Group 24 - Lumber				
27. Sawmills & Planing Mills - 2421				
Kiln Drying of Lumber	200*	(100)	63.4	(66.9)
28. Plywood - 2435				
Plywood Drying	250	(121)	50.6	(53.4)
29. Veneer - 2436				
Veneer Drying	212	(100)	57.8	(61.0)

Appendix 1 (continued)

Industrial Process Heat Requirements at Temperatures 300°F (149°C) and Below

Industry - SIC Group	Application Temperature Requirement °F	(°C)	Process Heat Used for Application 10^{12}BTU/Yr	(10^{12} KJ/YR)
Group 25 - Furniture				
30. Wooden Furniture - 2511				
Makeup Air & Ventilation	70	(21)	5.7	(6.0)
Kiln Dryer & Drying Oven	150	(66)	3.8	(4.0)
31. Upholstered Furniture - 2512				
Makeup Air & Ventilation	70	(21)	1.4	(1.5)
Kiln Dryer & Drying Oven	150	(66)	0.9	(0.9)
Group 26 - Paper				
32. Pulp Mills - 2611				
Paper Mills - 2621				
Paperboard Mills - 2631				
Building Paper - 2661				
Pulp Refining	150	(66)	175	(185)
Black Liquor Treatment	280	(138)	164	(173)
Pulp & Paper Drying	290	(143)	383	(404)
Group 28 - Chemical				
33. Cyclic Intermediates - 2865				
Styrene	250-300	(121-149)	35.0	(37.0)
Phenol	250	(121)	0.45	(0.47)
34. Alumina - 28195				
Digesting, Drying, Heating	280	(138)	113.2	(119.4)
35. Plastic Materials & Resins - 2821				
Polystyrene, suspension process				
Polymerizer Preheat	200-215	(93-102)	0.102	(0.107)
Heating Wash Water	190-200	(88-93)	0.067	(0.068)
36. Synthetic Rubber - 2822				
Cold SBR Latex Crumb				
Bulk Storage	80-100	(27-38)	0.179	(0.189)
Emulsification	80-100	(27-38)	0.086	(0.091)
Blowdown Vessels	130-145	(54-63)	0.865	(0.912)
Monomer Recovery by Flashing & Stripping	120-140	(49-60)	4.095	(4.319)

(continued on next page)

Appendix 1 (continued)

Industrial Process Heat Requirements at Temperatures 300°F (149°C) and Below

Industry - SIC Group	Application Temperature Requirement °F	(°C)	Process Heat Used for Application 10^{12}BTU/Yr	(10^{12} KJ/YR)
36. Synthetic Rubber - 2822 (continued)				
Dryer Air Temperature	150-200	(66-93)	3.663	(3.864)
Cold SBR, Oil-Carbon Black Masterbatch				
Dryer Air Temperature	150-200	(66-93)	0.506	(0.534)
Oil Emulsion Holding Tank	80-100	(27-38)	0.090	(0.095)
Cold SBR, Oil Masterbatch				
Dryer Air Temperature	150-200	(66-93)	1.09	(1.15)
Oil Emulsion Holding Tank	80-100	(27-38)	0.090	(0.095)
37. Cellulosic Man-made Fibers - 2823				
Acrylic	<250	(<121)	23.5	(24.8)
38. Noncellulosic Fibers - 2824				
Rayon	<212	(<100)	37.8	(39.9)
Acetate	<212	(<100)	37.6	(39.7)
39. Pharmaceutical Preparations - 2834				
Autoclaving & Cleanup	250	(121)	18.85	(19.88)
Tablet & Dry-Capsule Drying	250	(121)	1.00	(1.05)
Wet Capsule Formation	150	(66)	0.05	(0.05)
40. Soaps & Detergents - 2841				
Soaps				
Various Processes in Soap Manufacture	180	(82)	0.50	(0.53)
Detergents				
Various Low-Temperature Processes	180	(82)	0.36	(0.38)
41. Organic Chemicals, N.E.C. - 2869				
Ethanol	200-250	(93-121)	6.0	(6.0)
Isopropanol	200-300	(93-149)	11.0	(12.0)
Cumene	250	(121)	1.0	(1.0)
Vinyl Chloride Monomer	250-300	(121-149)	9.0	(9.0)
42. Urea - 2873215				
Low-Pressure Steam-Heated Stripper	290	(143)	0.89	(0.94)

Appendix 1 (continued)

Industrial Process Heat Requirements at Temperatures 300°F (149°C) and Below

Industry - SIC Group	Application Temperature Requirement °F	(°C)	Process Heat Used for Application 10^{12}BTU/Yr	(10^{12} KJ/YR)
43. Explosives - 2892				
Dope (Inert Ingredients)				
Drying	300	(149)	0.006	(0.006)
Wax Melting	200	(93)	0.118	(0.12)
Nitric Acid Concentrator	250	(121)	0.070	(0.07)
Sulfuric Acid Concentrator	200	(93)	0.027	(0.02)
Nitric Acid Plant	200	(93)	0.223	(0.23)
Blasting Cap Manufacture	200	(93)	0.016	(0.01)
Group 29 -- Petroleum				
44. Petroleum Regining - 2911				
Alkylation	45-300	(7-149)	59	(62)
Butadiene	250-300	(121-149)	60	(63)
45. Paving Mixtures - 2951				
Aggregate Drying	275-300*	(135-149)	88.1	(92.9)
Group 30 - Rubber				
46. Tires & Inner Tubes - 3011				
Vulcanization	250-300	(121-149)	6.18	(6.52)
Group 31 - Leather				
47. Leather Tanning & Finishing - 3111				
Bating	90	(32)	0.094	(0.099)
Chrome Tanning	85-130	(29-54)	0.060	(0.063)
Retan, Dyeing, Fat Liquor	120-140	(49-60)	0.15	(0.16)
Wash	120	(49)	0.034	(0.036)
Drying	110*	(43)	2.05	(2.16)
Finish Drying	110*	(43)	0.13	(0.14)
Group 32 - Stone, Clay, Glass & Concrete Products				
48. Hydraulic Cement - 3241				
Drying	275-300*	(135-149)	8.0	(8.0)
49. Concrete Block - 3271				
Low-Pressure Curing	165*	(74)	12.29	(12.96)

Appendix 1 (continued)

Industrial Process Heat Requirements at Temperatures 300°F (149°C) and Below

Industry - SIC Group	Application Temperature Requirement °F	(°C)	Process Heat Used for Application 10^{12} BTU/Yr	(10^{12} KJ/YR)
50. Ready-Mix Concrete - 3273				
Hot Water for Mixing Concrete	120-190	(49-88)	0.34	(0.36)
51. Gypsum - 3275				
Wallboard Drying	300	(149)	11.18	(11.79)
52. Treated Minerals - 3295				
Kaolin				
Drying	230*	(110)	12.7	(13.4)
Expanded Perlite				
Drying	160*	(71)	0.22	(0.23)
Barium				
Drying	230*	(110)	0.34	(0.36)
Group 33 - Primary Metals				
53. Ferrous Castings				
Gray Iron Foundries - 3321				
Malleable Iron Foundries - 3322				
Steel Foundries - 3323				
Pickling	100-212	(38-100)	151	(160)
Group 34 - Fabricated Metal Products				
54. Galvanizing - 3479				
Cleaning, Pickling	130-190	(54-88)	0.011	(0.012)
Group 36 - Electrical Machinery				
55. Motor & Generators - 3621				
Drying & Preheat	150	(66)	0.043	(0.045)
Baking	300	(149)	0.133	(0.140)
Group 37 - Transportation Equipment				
56. Motor Vehicles - 3711				
Baking-Prime & Paint Ovens	250-300	(121-149)	0.29	(0.31)

Note: SIC Groups 34, 35, 36, 37 utilize hot water for parts degreasing and washing in application temperatures of 80-180°F (27-82°C); total process heat used is not currently available.

*No special temperature required; requirement is simply to evaporate water or to dry the material.

FEDERAL ASSISTANCE PROGRAM

PROJECT STATUS REPORT

GEOTHERMAL TECHNOLOGY TRANSFER

GRANT NO. DE-FG-07-83ID 12478 M011

Reporting Period: June 1987

PAUL J. LIENAU AND GENE CULVER

Geo-Heat Center
Oregon Institute of Technology
Klamath Falls, Oregon 97601

1. PROJECT SUMMARY - JUNE 1987

1.1 Geothermal Information Services. Advising was provided to 13 inquiries and materials were sent to 23 requests, 6 hours of lectures and a tour of geothermal sites in the Klamath Falls area were presented to an Elderhostel at OIT, and one additional tour was given .

1.2 Geothermal Progress Monitor. Progress monitor activities are reported on: 1) New developments at Susanville, CA; 2) Binary power plant to be built at Amedee, CA; 3) Carson City, NV elementary school to use geothermal; and 4) Decision reached for test drilling near Crater Lake, OR.

1.3 Geothermal Direct Heat Applications Handbook. Draft status of work on chapters is indicated and delivery dates are given.

1.4 Geothermal Technology Support to End-Users reported under advising for June.

1.5 Geo-Heat Center staff that worked on the project in June include: Paul Lienau (1%), Gene Culver (72%), Cindy Nellipowitz (49%), Joyce Pryor (17%), Kevin Rafferty (84%), Charles Higbee (66%), and our student library assistant worked 27 hours.

2. GEOTHERMAL INFORMATION SERVICES

Transfer of technical information on geothermal resources and applications is provided to potential users, consulting engineers, industry groups, developers and the general public. This effort includes advising, distribution of published material, publishing a Quarterly Bulletin, maintaining a geothermal technology library, presentations and tours, and issuing geothermal technology development status reports for the Progress Monitor.

2.1 Advising. The following phone and letter inquiries, and personal visits were handled by the GHC during June 1987.

	Name	Date	Nature
a.	Jack McNamara 11752 San Vicento Bvd. Los Angeles, CA	6/1	Resource. Requested information on Raft River resources for direct use applications. Previous studies on direct uses for irrigation, fish farming, potato dehydration and soil warming were sent.

	Name	Date	Nature
b.	Darrel Seven D No. 7 Ranch Summer Lake, OR	6/8	Well Testing. Owns three wells, one producing 1000 gpm, and the other two at 200-300 gpm with an average temperature of 230°F. Provided referral to three pumping contractors in the Klamath area that can perform a pump test. Interested in bottling mineral water.
c.	Tom Drougas Guyer Springs Water Co. Ketchum, ID	6/10	Space Heating. Planning to provide geothermal for space heating of condominiums. Requested an economic analysis computer run on the project using RELCOST. A form for input data was sent.
d.	Alex Sifford Oregon DOE Salem, OR	6/10	Well Test. Requested cost of well test conducted at the Klamath College Industrial Park. The cost for a six day pump test was $8,250 excluding MOB and de-MOB and monitoring observation wells.
e.	Leo Glinkman 6th St. Steel Klamath Falls, OR	6/15	Equipment. Requested heat transfer rates from finned pipe for use in the Lakeview greenhouse. Calculated values were provided for given temperatures and pipe dimensions.
f.	Sandy Balsiger Caldwell Bankers Klamath Falls, OR	6/16	Wells. Provided data on two wells located in the Klamath Falls urban area.
g.	Rick Chitwood Chitwood Energy Mgt. Mt. Shasta, CA	6/18	Space Heating. Requested information on Cedarville Hospital and Fort Bidwell feasibility studies. Especially needed data for heat exchangers (copper coils and H2S reaction) and alternative pipe materials. He wanted to know how to sample for H2S. Provided materials selection guide, sampling techniques and information on corrosion in low temperature geothermal.
h.	Glen Townsend Ft. Bidwell Tribal Council Fort Bidwell, CA	6/19	Space Heating. Requested names of experienced geothermal contractors to install system for home heating regarding feasibility study completed by the GHC last year. Names and phone numbers of five contractors were supplied.

	Name	Date	Nature
i.	Gib Cooper Mendocino Jr. College Lakeport, CA	6/23	Greenhouses. Requested a review of plans for experimental geothermal greenhouse.
j.	Ray Rangila Confederate Tribes of Warm Springs Warm Springs, OR	6/25	Space Heating. Requested feasibility study for heating lodge, which is ½ mile from the springs.
k.	Klaus Findler Wheatley, NY	6/30	Power Generation. Requested names of U.S. manufacturers of modular binary power plants. Provided ORMAT and Barber-Nichols Engineering.
l.	Darrel Seven Paisley, OR	6/30	Swimming Pool. Planning to build RV park with hot tubs and swimming pool. Discharge temperature from the facility is estimated at 105°F. Disposal is by means of discharge into sewage ponds. Bacterial will not function above 80°F, with ideal temperatures between 60 to 65°F.
m.	John Hader 1108 W. Jon St. Pasco, WA	6/30	Greenhouses. Requested information on heating geothermal greenhouses with a 120°F hot spring at Lolo Pass, ID. Provided greenhouse heating guide.

2.2 Information Materials. The following requests were received by the Geo-Heat Center for published materials.

	Name	Date	Nature
a.	Sally Benson LBL Berkeley, CA	6/1	Letter requesting review of draft handbook chapter on DHE's.
b.	Mike Wright UURI, Earth Sci. Lab Salt Lake City, UT	6/1	Copy of same.
c.	Joe Kanta Caldwell, ID	6/1	Letter, feasibility study on Moana, papers on Eval & Design of DHE's, Natural Convection Promoter for Geo Wells and draft copy of DHE chapter for handbook.

	Name	Date	Nature
d.	Jack McNamara Los Angeles, CA	6/2	Letter regarding request for information on Raft River. Enclosed 4 technical publications on loan from GHC library.
e.	Alex Sifford ODOE Salem, OR	6/3	Copy of map from Oregon Site Data Base book.
f.	Jozef Csaba 2443 Szazhalombatta Pf 32 HUNGARY	6/4	Information on geothermal in HI, greenhouse heating, copy of draft chapter on greenhouses for handbook, back bulletin on greenhouses.
g.	S.S. Kittur Mgr. Planning Thermax Private Ltd. Thermax House 4 Bombay Pune Rd Shivajinagar Pune 411005 INDIA	6/8	Letter, Publication Request Form, information on geothermal in India, referral to Geol. Survey of India for more information.
h.	Robert Cherry Vice Pres., Engineering Layne & Bowler Inc. Memphis, TN	6/9	Letter regarding bearing failure at MWMC in Klamath Falls, OR (hospital). Enclosed slides and piece of bearing.
i.	Charles Sundquist, PE Richland, WA	6/11	Letter regarding his proposed absorption power generator. Suggested he write an article for the Bulletin. Enclosed back issues of the Bulletin.
j.	Gordon Bloomquist WSEO Olympia, WA	6/11	Letter regarding GHC's providing data to WSEO on DH systems in Klamath Falls area if WSEO's proposal for GEODIM is approved.
k.	Ben Lunis EG&G Idaho Idaho Falls, ID	6/11	Copy of same.
l.	Tom Drougas Guyer Spgs Water Co. Ketchum, ID	6/15	Letter, form to fill out for RELCOST life cycle analysis.

	Name	Date	Nature
m.	Kevin Fisher City Water Dept. San Bernardino, CA	6/15	Letter with slides of San Bernardino geothermal system.
n.	C. McGuire Centrilift Hughes Huntington Beach, CA	6/17	Letter requesting review of draft chapter on pumps for handbook.
o.	Jack Frost, Mgr. Geo. Services Div. Azusa, CA	6/17	Letter requesting review of draft chapter on pumps for handbook.
p.	Alex Sifford ODOE Salem, OR	6/18	Letter, enclosed slides of Oregon Trail Mushroom plant.
q.	Rick Chitwood Chitwood Energy Mgt. Mt. Shasta, CA	6/18	Copy of techniques for geothermal liquid sampling and analysis.
r.	R.C. Cherry Layne & Bowler Inc. Memphis, TN	6/18	Letter, draft chapter on pumps for handbook for his review.
s.	David Bomar Balzhiser, Hubbard & Assoc. Eugene, OR	6/22	Letter, copy of draft chapter on pumps for handbooks for his review.
t.	Cecil Kindle Battelle NW Richland, WA	6/22	Letter, copy of draft chapter for handbook taken primarily from Battelle publication. Asked for permission and review.
u.	Bob Helm PP&L, Columbia Div. Portland, OR	6/25	Letter explaining occupancy schedule used to determine heat load in study of Convention Center. Enclosed Publication Request Form and back Bulletin.
v.	B.J. Pfeffer Calgary, Alberta CANADA	6/25	General information on GHC, back Bulletins, Publication Request Form.
w.	George Wagner JUB Engineering Boise, ID	6/16	Letter with enclosures on piping and general geothermal.

2.3 Presentations and Tours.

2.3.1 Paul Lienau presented three lectures to 25 persons attended Elderhostel at OIT and provided a tour of Klamath geothermal sites.

2.3.2 A presentation was requested by the City of Klamath Falls to the Oregon Utilities Coord. Council for 21 Aug. 1987.

2.3.3 A tour and talk on Klamath geothermal sites was requested for 40 persons from Rogue Valley for September.

2.3.4 Gene Culver provided a tour for four people from Tigard, OR of the geothermal system at OIT on June 22nd.

3. GEOTHERMAL PROGRESS MONITOR

attached

4. GEOTHERMAL DIRECT HEAT APPLICATIONS HANDBOOK

4.1 Status of work on the "Geothermal Direct Heat Applications Handbook" is as follows:

Chapter		Status
1	Introduction & State of the Art	Not Started
2	Program Opportunities Notice (PON) Project Lessons Learned	In Progress
3	Nature & Distribution of Geothermal Resources	In Progress
4	Exploration for Geothermal Resources	In Progress
5	Water Sampling Techniques	Completed
6	Drilling & Well Construction	In Progress
7	Well Testing	In Progress
8	Materials Selection	Completed
9	Well Pumps	Completed
10	Piping	In Progress
11	Heat Exchangers	Completed
12	Space Heating Equipment	In Progress
13	Heat Pumps	Not Started
14	Absorption Refrigeration	Completed
15	Greenhouses	Completed
16	Aquaculture	Completed
17	Industrial Applications	Not Started
18	Engineering Cost Analyses	In Progress
19	Institutional, Legal & Permit Requirements by State	In Progress
20	Environmental Aspects	In Progress

3. GEOTHERMAL PROGRESS MONITOR

3.1 New Developments in Susanville. Two additional facilities will be connected to the Susanville geothermal district heating system and a new injection well will be drilled.

The Susanville California geothermal district heating system is designed to provide space heating to two separate geothermal loops. The first loop, in operation since 1982, circulates 170^{o}F fluid to nineteen public and commercial buildings, and a second to 23 homes.

Two additional facilities will be connected to the system by Oct. 1, 1987. The Roosevelt swimming pool will be connected to enable year around use. Geothermal will be used for space and pool heating. City shops will also be retrofitted to utilize geothermal for space heating from the district system.

Susanville has had difficulty finding a suitable disposal method for the geothermal effluent from the system. Originally, a well drilled by the Bureau of Reclamation, Richardson 1, was to be used. The Richardson well had a poor injectivity, 0.85 gpm/foot, which is only 1/10 the productivity of other wells in the area, and a high skin factor which indicated the well was damaged during drilling by mud invasion into permeable zones. This well was able to accept only 150 gpm of an estimated average 400 gpm required by the system. Temporary suface discharge permits had to be obtained from the State Department of Water REsources, which considers geothermal water to be a hazardous waste. To resolve this problem, the city has obtained funds from the California Energy Commission and HUD to drill a new 500 ft. injection well between Suzy's 5 and 7 production wells and near the Tsuji nursery. This well is expected to accept 500 gpm and about 600 ft. of pipeline will be required to connect to the existing system.

3.2 Binary Power Plant to be Built at Amedee. Amedee hot springs is located near Wendel and has been a site where several successful wells have been drilled. Transpacific Geothermal Corporation has received permits to build the Amedee Geothermal Power Plant Project. The 1.5 MW plant will utilize two existing wells, Norcal No. 1 and No. 2, to provide 3800 gpm of fluid at 226^{o}F to the plant. A 20 year contract has been signed with PG&E and CP National's transmission lines will be used to tie into the grid.

3.3 Carson City Elementary School to Use Geothermal. The Stanton Drive Elementary School of the Carson City School District will be heated by geothermal resources and will be a state energy pilot project. Design Concepts West of Carson City, Nevada, selected as the architect to design several new facilities for the district, has commenced the design of the Stanton Drive Elementary School.

3.4 Decision Reached for Test Drilling near Crater Lake. The Bureau of Land Management (BLM), Lakeview District, and the United States Forest Service (USFS), Winema National Forest, have reached a decision on a request to deepen temperature gradient wells and drill without circulation on the Winema National Forest by California Energy Company, Inc. The decision is to implement the proposed action, to drill on previously disturbed sites within the unitized areas to 5,500 feet and with fluid loss to the subsurface. The reasons given were:

a. A detailed computer model analysis of the possible impacts of fluid loss from drill holes in the vicinity of Crater Lake indicates that this fluid loss could pose no threat to Crater Lake or affect the hydrologic system in the immediate vicinity of the lake caldera.

b. Subsurface aquifers are adequately protected by the requirements to seal zones of inflow and to use non-toxic drilling fluids.

c. Surface related impacts are considered negligible.

d. With the exception of drilling depth and fluid loss, the special design features and mitigation measures developed in the original 1984 EA and the monitoring plan requirements remain unchanged.

e. BLM and USFS will conduct detailed on-site inspections for each proposed site location prior to any surface disturbance. Personnel from the NPS will be invited to attend this inspection.

f. Drilling to 5,500 feet will provide data that will improve the understanding of the hydrology and geology of the Crater Lake area. This information should prove useful for future decisions related to geothermal exploration and provide information to the National Park Service and the Forest Service to aid them in their management efforts.

g. If a new or amended Geothermal Drilling Permit Application is received from California Energy Company, Inc., the decision on the application will be based on the original and supplemental EAs and any newly generated information.

This decision will be implemented after a 30-day period has elapsed to allow for any appeal.

4.2 Schedule of Delivery Dates:

4.2.1 August 1, 1987 - Chapters to peers for review. Each author to choose peers.

4.2.2 September 1, 1987 - Chapters returned from peer review to allow time for updates and changes.

4.2.3 October 1, 1987 - Edit peer reviewed proof by Paul Lienau & Ben Lunis.

4.2.4 December 1987 - Printer ready.

5. GEOTHERMAL TECHNOLOGY SUPPORT TO END USERS

Technical assistance for June is reported under 2.1, Advising.