RESOURCE SERIES 16
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GEOTHERMAL RESOURCE ASSESSMENT OF IDAHO SPRINGS, COLORADO

by
F.N. Repplier
T.G. Zacharakis
C.D. Ringrose

Colorado Geological Survey / Department of Natural Resources / Denver, Colorado / 1982

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Department of Natural Resources
State of Colorado
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1982

NOTICE

NOTICE

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TABLE OF CONTENTS

	PAGE
Acknowledgments	ii
Abstract	1
Introduction	1
History	3
Geology	3
Introduction	3
Stratigraphy	5
Tertiary Intrusives	5
Quaternary sedimentary deposits	5
Structural history	5
Precambrian	5
Laramide uplift	7
Hydrogeology of Idaho Springs Thermal Waters	8
Introduction	8
Origin of thermal waters	9
Conclusion	12
Electrical Geophysical Resistivity Surveys	13
Conclusion	13
Soil Mercury Surveys	19
Introduction	19
Strategy and methodology	20
Sampling methods	20
Analysis	21
Indian Springs soil mercury surveys	22
Introduction	22
Soil description	23
Sample Analysis	23
Summary	28
Conclusion	29
References	30
Appendix A. Physical Properties and Chemical Analysis of Indian Springs Thermal Waters	33
Appendix B. Factors Affecting Resistivity	34
Appendix C. Scintrex RAC-8 System Description	35
Appendix D. Resistivity Field Procedures	36
Appendix E. Resistivity Calculations	43
Appendix F. Geometric Factor Tables	50

FIGURES

PAGE

1. Index Map 2
2. Topographic map 2
3. Geologic map of Idaho Springs region.......................... 4
4. Structural geology map of Idaho Springs region................ 6
5. Thermal springs and wells 10
6. Geophysical resistivity surveys............................... 14
7. Dipole-dipole Line A.. 16
8. Dipole-dipole Line B.. 17
9. Dipole-dipole Line C.. 18
10. Soil Mercury surveys... 24
11. Cumulative frequency distribution of east slope of Soda Creek 27
12. Schematic diagram for resistivity 37
13. Wenner array 38
14. Schlumberger array 40
15. Dipole-dipole array 41
16. Data plotting scheme for dipole-dipole array 42
17. Typical dipole-dipole array 42

TABLES

1. Description of Indian Springs thermal waters 8
2. Mercury content (ppb) of samples collected on east side of Soda Creek 23
3. Mercury content of samples collected on west side of Soda Creek ... 26
4. Mercury content of samples collected 0.6 mi southeast of Indian Springs Lodge 26
5. Resistivity line A calculations 45
6. Resistivity line B calculations 47
7. Resistivity line C calculations 48
8. Schlumberger method geometric factor table 50
9. Dipole-dipole geometric factor table 50
10. Wenner method geometric factor table 50

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Frank V. Repplier, Ted G. Zacharakis, and Charles D. Ringrose

ABSTRACT

Located in the Front Range of the Rocky Mountains approximately 30 miles west of Denver, in the community of Idaho Springs, are a series of thermal springs and wells. The temperature of these waters ranges from a low of 68°F (20°C) to a high of 127°F (53°C).

To define the hydrothermal conditions of the Idaho Springs region in 1980, an investigation consisting of electrical geophysical surveys, soil mercury geochemical surveys, and reconnaissance geological and hydrogeological investigations was made. Due to topographic and cultural restrictions, the investigation was limited to the immediate area surrounding the thermal springs at the Indian Springs Resort.

The bedrock of the region is faulted and fractured metamorphosed Precambrian gneisses and schists, locally intruded by Tertiary age plutons and dikes.

The investigation showed that the thermal waters most likely are fault controlled and the thermal area does not have a large areal extent.

INTRODUCTION

In 1979 the Colorado Geological Survey, in cooperation with the U.S. Department of Energy, Division of Geothermal Energy, under Contract No. DE-AS07-77-28365, initiated a program to determine the nature of those geothermal resources in Colorado with high potential for near term development. This effort consisted of a literature search, reconnaissance geological and hydrogeological mapping, resistivity geophysical surveys, and soil mercury geochemical surveys.

One of the thermal areas investigated was Idaho Springs, Colorado, a community of 2,000 persons located along Clear Creek approximately 30 miles west of Denver on U.S. Interstate Highway 70 (Fig. 1). A group of thermal springs and wells are located at the Indian Springs resort on the south side of Idaho Springs several hundred yards (184 m) south of I-70 along Soda Creek (Fig. 2). These springs are used commercially for recreation and therapeutic purposes.

As reported by Coe and Zimmerman (1981), the Mayor of Idaho Springs in 1979 expressed an interest in having the geothermal resources of her community developed. Based on the city's interest, the Colorado Geothermal Resource Assessment Team in 1980 decided to make an appraisal of the thermal conditions in and adjacent to Idaho Springs.

Based on published information and reconnaissance field investigations, the only sources of thermal waters found in the Idaho Springs region are located at the Indian Spring Resort. This evaluation was further confirmed by miners who

Figure 1. Index Map.

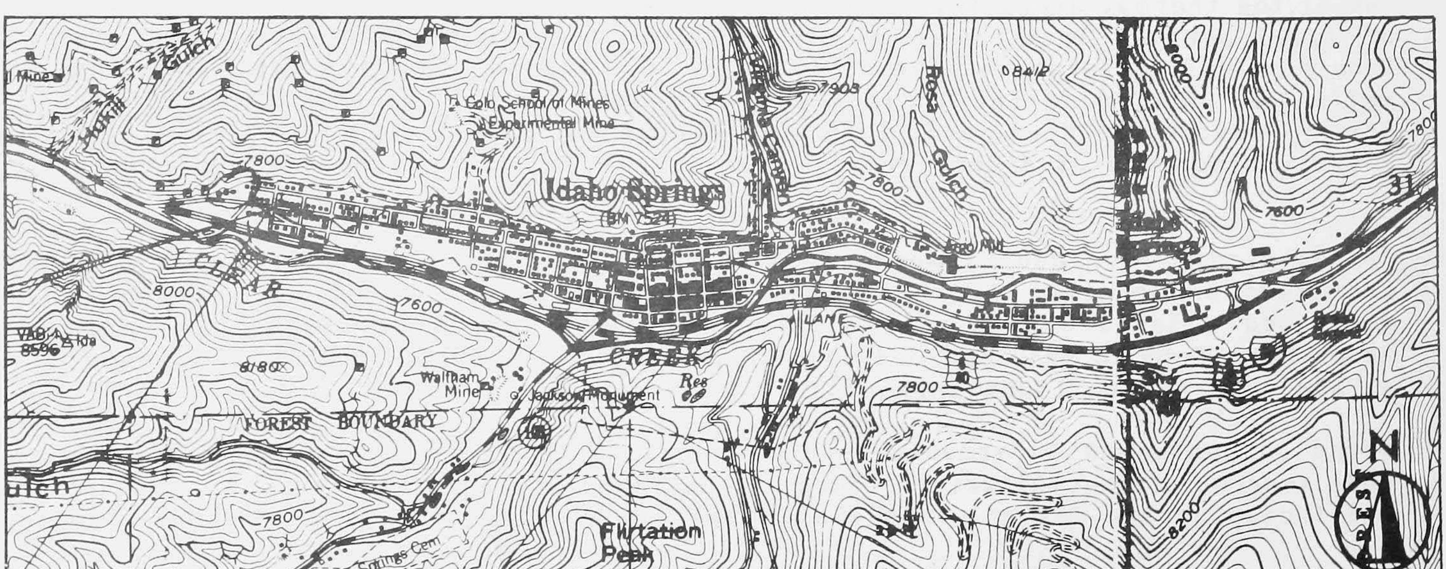


Figure 2. Topographic Map (U.S. Geol. Survey 7 1/2' Topo. Maps).

noted that the mine waters of the region are cold and they could not remember any warm mine drainage waters (J. Connors, pers. comm., 1982). Based on this information, the field investigations were restricted to the area immediately surrounding the Indian Springs Resort.

The field methods employed to delineate the areal extent of the Indian Springs geothermal reservoir consisted of electrical resistivity and soil mercury surveys. This report details the findings of these surveys.

HISTORY

The Ute and Arapahoe Indian tribes used the hot springs before they were discovered by prospectors in 1859 while searching for gold (Maxwell, oral comm., 1981). Kevin McCarthy (1982) has told the authors that at the time they were found by George A. Jackson, the springs had a temperature of 95°F (35°C). Placer gold was discovered to the west in Chicago Creek one week later by Jackson, and the area soon became the center for numerous mining camps. The community of Idaho Springs was eventually established by consolidation of various mining camps (K. McCarthy, oral comm., 1982).

Ownership of Indian Springs has changed many times since 1863, when the first commercial use of them was made by Dr. E. M. Cummings. The springs are presently owned by Jim Maxwell, who has expanded and improved both the facility and resource.

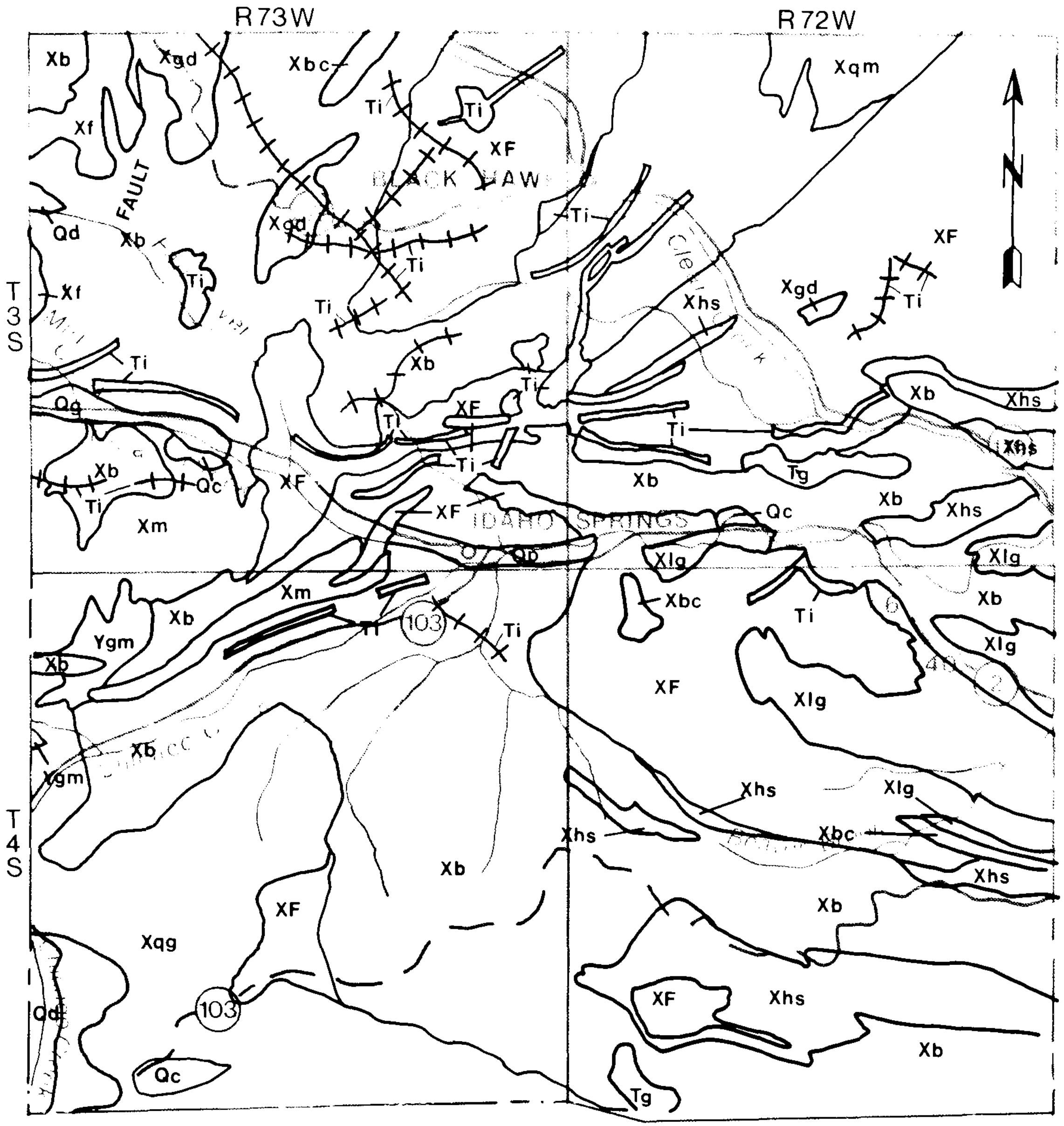
GEOLOGY

Introduction

The geological conditions of the Idaho Springs region have been described by Bryant and others (1981); Harrison and Wells (1959); Lovering and Goddard (1950); Moench (1964); Moench and Drake (1966); Moench and others (1962); Schwochow (1975); Sheridan and Marsh (1976); Sims (1960 and 1963); Spurr (1906); Spurr, Garrey and Ball (1908); Tooker (1963); Tweto (1975 and 1980); Tweto and Simms (1963); Warner (1980); and Wells (1960). The following description is taken from the above papers.

Idaho Springs is located in the Front Range of Colorado, a 30-35 mile wide mountainous uplift extending from Canon City on the south to the Wyoming border on the north, where it merges with the Laramie Range. The city also lies near the northeastern end of the Colorado Mineral Belt, a northeast trending zone of intrusive rocks extending from the Silverton area in the southwest part of Colorado to just north of Boulder. The bedrock of the area is a complex assemblage of Precambrian age rocks.

As shown on Fig. 3, the bedrock of the area in the vicinity of the Indian Hot Springs is a biotite gneiss of Precambrian age. According to Bryant and others (1981), this rock unit consists of the following: "Biotite-quartz plagioclase schist and gneiss, commonly contains abundant sillimanite and less abundant muscovite with some layers of cordierite-biotite gneiss and of garnet-biotite gneiss. Locally a few layers of hornblende gneiss and calc-silicate gneiss are evident. Lenses, pods, and thin layers of pegmatite are abundant. In some regions, layers of lenses of granodiorite and quartz monzonite are also abundant and the rock grades to migmatite, which probably is derived from shale, siltstone and sandstone."



EXPLANATION

Qp Alluvium (Holocene)
Qd Glacial drift (Plesitocene)
Qc Colluvium (Holocene and Pleistocene)
Tg Gravel (Pliocene? or Miocene)
Ti Intrusives: Bostonite, Monzonitic, and Granodioritic (Paleocene & upper Cretaceous)
Xgm Migmatitic quartz, monzonite and granodiorite (Precambrian)
Xqg Quartz monzonite and granodiorite (Precambrian)
Xgd Granodiorite (Precambrian
Xm Migmatite (Precambrian)
Xhs Amphibolite and calc-cilicate gneiss (Precambrian)
Xf Felsic gneiss (Precambrian)
Xls Layered gneiss (Precambrian)
Xb Biotite gneiss (Precambrian)
Xbc Cordiertite-bearing and garnet bearing sillimanite-biotite gneiss (Precambrian)

SCALE
1:125,000

Figure 3. Geologic Map of Idaho Springs Region.
(Adapted from Bryant, McGrew and Wobus, 1981)

The hot springs in Idaho Springs are located at the confluence of Soda Creek and Clear Creek, and at the intersection of a northeast trending, highly mineralized fault zone (Fig. 4). Hydrothermal alteration is readily apparent along I-70 where it traverses the Idaho Springs mining district.

Lovering and Goddard (1950) mapped a Tertiary (Eocene?) intrusive extending a short distance north from the hot springs. This unit was described as alkali syenite, diorite, monzonite, and sodic granite (Lovering and Goddard, 1950). This unit was not shown on Figure 3.

Stratigraphy

According to Tweto (1975), the Front Range has had postive tendencies since Precambrian time. The sedimentary rocks deposited over the Precambrian age rocks throughout geologic time, have been removed by the subsequent uplifts and erosion that occurred in Pennsylvanian, Cretaceous, and Tertiary time.

The biotite gneiss bedrock of the region consists of the Precambrian age rocks formerly called the Idaho Springs Formation (Tweto, 1977). With the exception of Tertiary intrusives, Quaternary and Recent sediments, no younger rocks are found in the region. The biotite gneiss bedrock locally is intruded by bodies of pegmatites and Boulder Creek and Silver Plume granites of Precambrian age.

Tertiary Intrusives

During early Tertiary time granitic type intrusive rocks were emplaced in the Idaho Springs region (Lovering and Goddard, 1950). The form of these intrusions varied from irregularly shaped plutons to radiating dikes (Fig. 4). The plutons generally have steep walls and range from several hundred feet to several thousand feet in diameter. The dikes may be miles in length but are generally only a few feet to a few tens of feet in width.

Quaternary Sedimentary Deposits

Quaternary age alluvial and colluvial deposits are found throughout the Clear Creek Valley and the Idaho Springs region (Fig. 3). Alluvial deposits occur along the various streams cutting the region. Colluvial deposits occur on the sides of the hills above the rivers. During Pleistocene time, glaciers were present to the west and extended to within approximately 3 to 4 miles (4.8 to 6.4 km) of Idaho Springs. Evidence of these glaciers appears in the Pleistocene terrace gravels found along Clear Creek about 3 miles (4.8 km) west of Idaho Springs.

STRUCTURAL HISTORY

Precambrian

After deep burial and subsequent high grade metamorphism of Precambrian sediments, two periods of deformation occurred. The first was a plastic deformation that produced large north-northeast trending, gently plunging folds

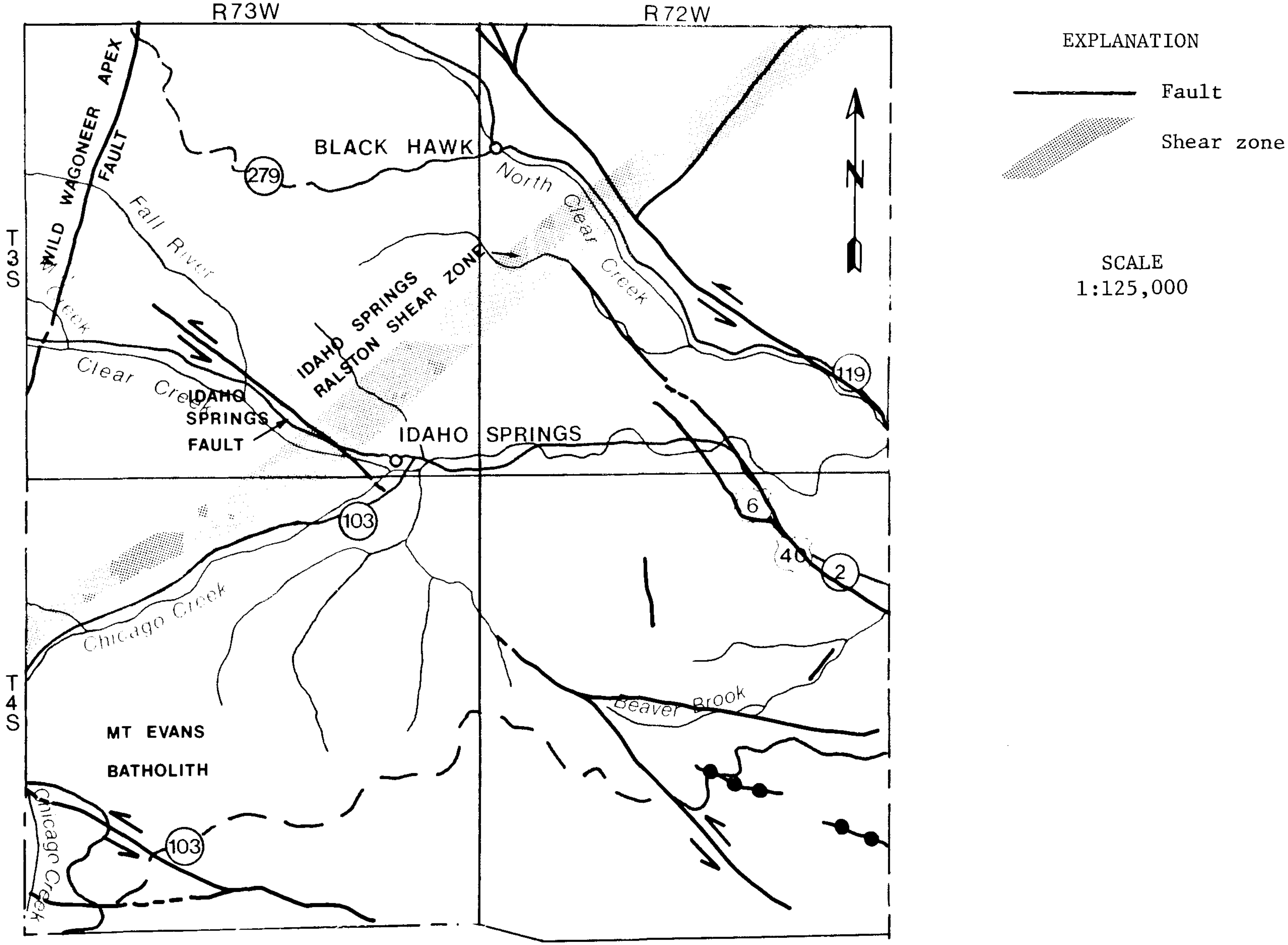


Figure 4. Structure geology map of Idaho Springs Region. (Adapted from Bryant, McGrew and Wobus, 1981)

that still dominate the structure in the area. This deformation included the intrusion of a series of granodiorite, quartz-diorite, hornblendite, and biotite-muscovite plutons.

The second deformation resulted from movement along the northeast trending Idaho Springs-Ralston shear zone of Tweto and Simms (1960) (Fig. 4). The competent rocks within this zone were sheared and granulated while the less competent beds were folded. These folds which formed along a uniform axis are only a few feet in width. The folds are all asymmetrical and have steep axial planes with the northwest side appearing to have been pushed up and over the southeast side.

Laramide Uplift

At the close of Cretaceous time and extending into early Tertiary time, the southern Rocky Mountains were uplifted piecemeal during the Laramide Orogeny (Tweto, 1975). As indicated by the sedimentary record, the Front Range area was uplifted and eroded from very late Cretaceous into early Tertiary time (Tweto, 1975).

The first part of this activity was marked by two periods of large scale faulting followed by intrusions of porphyritic rocks. The earlier faulting was along a northwesterly path while the more numerous younger faults trend east to northeast. Both sets of faults were steeply dipping. Orientation of intrusive rocks in the Colorado Mineral Belt was largely decided by the structural weaknesses caused by the faulting, particularly the second set of faults (Tweto, 1975).

Tweto (1975) noted that while the Mineral Belt contains intrusive rocks of three distinct ages, only one group has been mapped in the Front Range area. These rocks, 50-70 m.y. old, are largely concentrated in a sharply defined and narrow zone extending from the east side of the Front Range to the west side of the Swatch Range. In this zone they occur in thousands of dikes and sills and in numerous stocks, most of which are small, less than 3 mi (5 km) maximum dimension (Tweto, 1975).

Following igneous activity in early Tertiary time, a period of faulting and fracturing occurred. For the most part, the fracture and fissures strike east-northeast and dip steeply northwestward with strike-slip displacement averaging several inches to a few feet. An unusual exception is the Idaho Springs Fault, a major northwest trending fault, with strike-slip displacement of over 600 ft (183 m) (Moench, 1964). This fault terminates at the west edge of Idaho Springs and does not extend into the thermal area (Fig. 4).

Mineralization of the veins occurred concurrently with faulting. This activity produced a breccia surrounded by gangue and ore minerals. The gangue minerals are mostly quartz, calcite, barite, and fluorite. The ore minerals are sulphides and sulphosalts of iron, copper, lead, zinc, silver, and gold.

The various stocks and numerous other porphyry bodies of the Colorado Mineral Belt have been interpreted as expressions of an underlying batholith or string of batholiths. Geophysical data has shown that the Mineral Belt in the Front Range coincides with a gravity valley pocked by deep gravity lows (Tweto, 1975).

HYDROGEOLOGY OF IDAHO SPRINGS THERMAL WATERS

Introduction

As noted earlier, all the thermal waters in Idaho Springs are located at the Indian Springs Resort on the southside of the city. At the resort are three thermal springs and three thermal wells ranging in temperature from 68°F (20°C) to 127°F (53°C). Table 1, below, presents a brief description of these waters. The complete chemical analysis of the waters is presented in Appendix A.

Table 1. Description of Indian Springs Thermal Waters

Spring A: This spring, located in a tunnel 75 ft (23 m) south of the lodge, used to be the primary source of hot water for the lodge. The temperature and flow rate have decreased markedly over the years, thus necessitating the drilling of new wells. The temperature ranges from 104°F (40°C) to 113°F (45°C) Barrett and Pearl, 1976). The spring has a discharge of 21 gpm. Total dissolved solids range from 1,940 to 2,110 mg/1, with the waters being a sodium-bicarbonate type.

Spring B: The spring is located 50 ft (15 m) east of the southeast corner of the lodge in a tunnel along the cliff face. The spring has a temperature of 75°F (24°C), with a discharge of less than one gpm. Total dissolved solids are 1,070 mg/1 and the waters are a sodium-bicarbonate type.

Spring C: This spring is located in a tunnel 100 ft (30 m) south of the lodge. The spring has a temperature of 68°F (20°C), a discharge of one gpm, and 1,070 mg/1 of total dissolved solids. The waters are a sodium-bicarbonate type.

Well A: This flowing well is located just north of the swimming pool. It was drilled in 1979 to a depth of 140 ft (43 m) to replace the deteriorating 10 year old lodge well and yields about 36 gpm. The water is at a temperature of 127°F (53°C) and is now the primary source of hot water for the resort. During the course of drilling this well, it was reported that temperature and discharge increased with depth.

Well B: Located south of the hotel, flows 3 or 4 gpm and drains into Soda Creek. The well is 40 ft (12 m) deep. The waters have a surface temperature of 111°F (44°C) and are presently unused.

Well C: Located north of the hotel, surrounded by a concrete ring, this well was capped and is no longer used. Sloughing is thought to have occurred in the well bore.

The thermal conditions of these waters have been described by Barrett and Pearl (1976 and 1978); Coe and Zimmerman (1981); George and others (1920); Lewis (1966); Mallory and Barrett (1973); Pearl (1972 and 1979); and Spurr, Garrey, and Ball (1908). Several of the authors, Barrett and Pearl (1978); Coe and Zimmerman (1981); Pearl (1979); and Spurr, Garrey, and Ball (1908), have attempted to explain the origin of the springs and to estimate their subsurface temperatures. Barrett and Pearl (1978) made the most comprehensive estimate of the subsurface temperatures. Using the Silica Mixing Model, Na-K, and Na-K-Ca Geothermometer Models, they estimated that the subsurface temperatures could range from a low of 47°C (117°F) to a high of over 200°C (393°F). However, these estimates are unreliable due to the ambiguous geochemistry of the thermal waters (Barrett and Pearl, 1978). Using historical geochemical data with the silica geothermometer model, Barrett and Pearl (1978) showed that the estimated maximum reservoir temperature has increased from approximately 104°F (40°C) to a high of approximately 176°F (80°C). They (Barrett and Pearl, 1978) pointed out that many factors could be influencing these estimates, and as such the estimates should be used as guides only.

Lacking any wells and subsurface data in the area, Pearl (1979) made several general assumptions about the size, extent, and temperature of the resource. This analysis determined that the areal extent of the system could encompass approximately 1.52 sq mi and contain as much as $.1714 \times 10^{10}$ B.T.U.'s of energy, at an estimated maximum temperature of 80°C (176°F). The accuracy of these estimates cannot be verified until more detailed appraisal work is done, including the drilling of test wells.

Another thermal spring once discharged into Clear Creek, but was destroyed by construction of I-70 (R. Fargo and unidentified citizen, oral comm., 1980). Even though this spring does not exist today, its approximate location is noted on Figure 5.

Origin of Thermal Waters

The Indian Springs appear to be fault controlled, as Tweto (1979) has mapped a northwest trending fault system just south of the springs (Fig. 5). Interpretation of the electrical resistivity data collected during the course of this investigation suggests the presence of several east-west trending faults through the springs (see section on electrical resistivity surveys in this report). If the springs are fault controlled, evidence is lacking on the deep, subsurface conditions of the controlling fault system, or the extent of the thermal system. Pearl (1979) noted that the system appeared to be fault and fracture controlled and estimated that this system could encompass approximately 1.52 sq mi (2.45 sq km). As pointed out by Pearl (1979), lack of definitive subsurface data renders this estimate quite tentative.

Due to the lack of any deep water wells in the Idaho Springs region or isotope data, from which meaningful hydrogeological data could be collected, the authors were limited in their efforts to fully evaluate the conditions of the region and the preparation of a working model of the thermal conditions. Therefore, a number of assumptions are presented regarding the possible origin of the Indian Springs thermal waters.

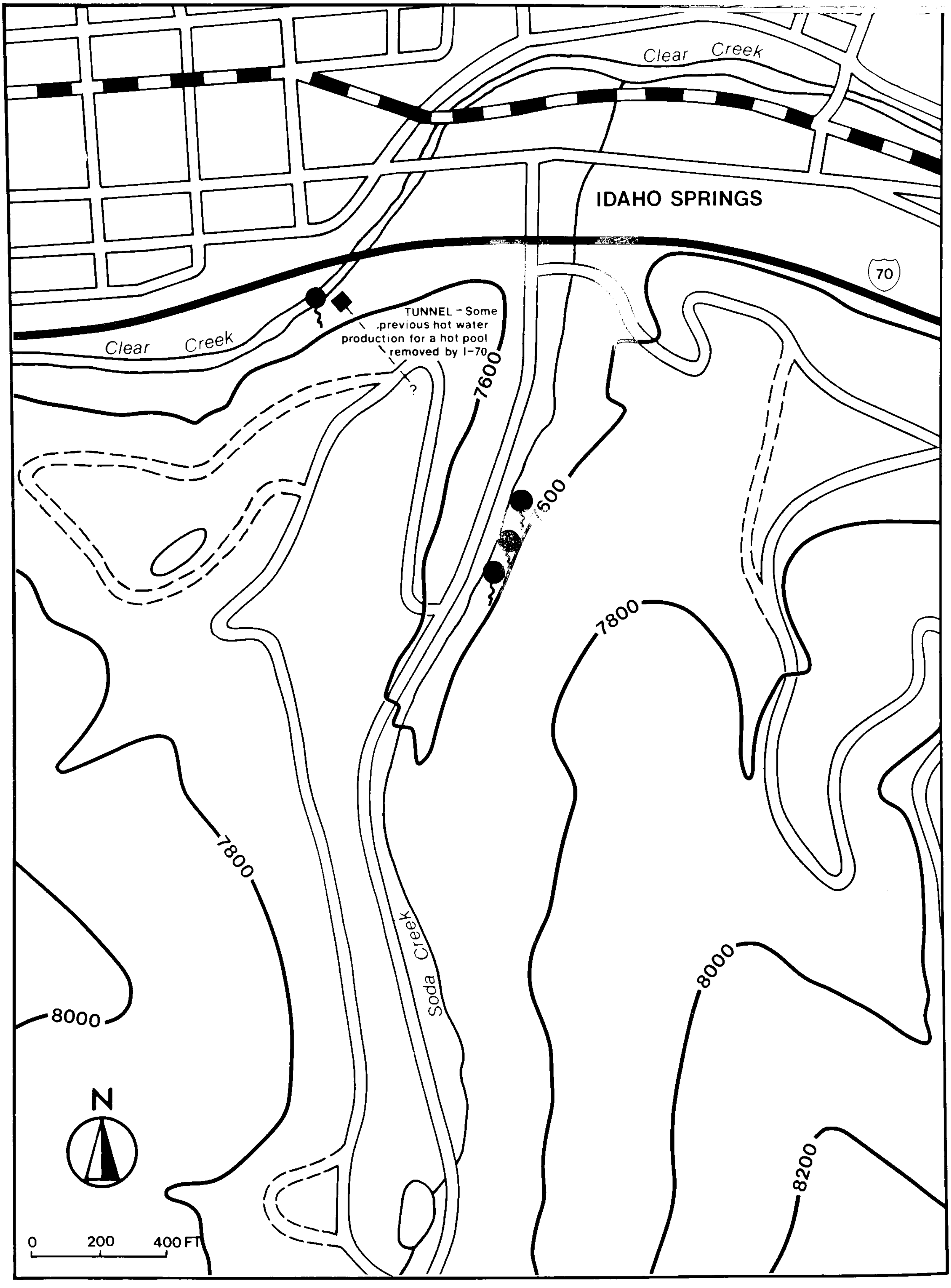


Figure 5. Thermal springs and wells.

Thermal waters can be of several origins, magmatic or meteoric. Magmatic waters are waters derived from a cooling igneous rock body, while meteoric waters are those which have fallen on the surface of the earth in the form of precipitation. Deeply migrating meteoric waters can become heated by several possible means: 1) Natural heat of the earth; 2) heat from decay of radioactive minerals; and 3) cooling magma bodies. Craig (1961) has shown that, under most conditions, thermal waters are of meteoric origin. Based on Craig's (1961) findings, it is the authors' opinion that the Indian Springs thermal waters are of meteoric origin.

One possible source of the heat is the disintegration of radioactive minerals. Wells (1960) has shown that the Tertiary age igneous rocks of the Colorado Mineral Belt in the Front Range are 15 to 25 times more radioactive than the average granitic rock. In the calculation of the natural flow of heat from the earth, the contribution from the decay of radioactive minerals is always considered. While the Tertiary intrusive rocks are highly radioactive, this radioactivity does not yield exceedingly high heat flow. Zacharakis (1981) has shown that this area has a heat flow of approximately 80 mW/m^2, slightly in excess of the heat flow of the Front Range. The geothermal gradient in this area has been measured at 25°C/km (U.S. Geol. Survey, 1974).

High mountain peaks, such as Mount Evans, Squaw Mountain, and the Continental Divide, are situated south and west of Idaho Springs. These mountains receive from 15 to over 40 inches of precipitation annually (Colorado State Planning Division, 1957). Some of this precipitation migrates downward along the numerous faults and fractures in the Front Range where they become heated by the natural heat of the earth. Using the reported geothermal gradient of this area, 80 mW/m^2 (Zacharakis, 1981), and allowing for cooling of the waters before they reach the surface, it can be calculated that if the waters reach a maximum temperature of 200°F (93°C), they must circulate to a depth of approximately 12,000 ft (3.65 km) below the recharge point. For the waters to reach a maximum temperature of 125°F (51.67°C), they would only have had to circulate to a depth of 6,562 ft (2 km) below the recharge point. This should not be misinterpreted to indicate that the thermal waters would be found at 2 km depth below Idaho Springs, but rather that the waters may circulate to such depths below the recharge point. If the recharge point lies at some higher elevation than Idaho Springs, then the difference in elevation between the recharge point and Idaho Springs has to be subtracted to determine the maximum depth at which the thermal waters may be found below Idaho Springs.

Another possible origin for the thermal waters is that they are, at least in part, of magmatic origin. As noted earlier, Idaho Springs is located in the Colorado Mineral Belt and extensive hydrothermal mineral deposits occur in the immediate vicinity. Tweto (1975) has noted that some authors have suggested the presence of buried batholiths beneath the Mineral Belt. Tooker (1963) believed that the hydrothermal (fluids warmer than 5°C of the enclosing environment [White, 1975]) mineral solutions of Idaho Springs are thought to be dilutions of magmatic water driven off from these batholiths, mixed with metamorphic and meteoric waters. He (Tooker, 1963) explained the origin of the Indian Springs as "representing the late stages of a long period of hydrothermal activity in the region, and are, as they issue at the surface, worked out, oxidized, diluted hydrothermal (mineral deposit forming) solution." Tooker (1963) did not estimate at what depth these fluids may have come from. This theory of Tooker's (1963) is within the guidelines of the authors' hypothesis of deep circulation of meteoric waters which become heated by the natural gradient of the earth.

Conclusion

From analysis of available geological and hydrogeological data, it appears that the Idaho Springs thermal system is very complex and not fully understood. Until more deep wells are drilled in the area which will help define the system, only a hypotheses can be presented regarding the origin and distribution of the thermal system.

ELECTRICAL GEOPHYSICAL RESISTIVITY SURVEYS

To define the thermal conditions of the Indian Springs area, electrical resistivity surveys were conducted to determine the location of low resistive zones in the Indian Springs area. Low resistivity is due to water saturation, higher than normal temperatures, and high clay matrix zones. For a complete description of the factors which may affect electrical resistivity measurements, see Appendix B.

Using a Scintrex RAC-8 Electrical Resistivity System (see Appendix C for description) measurements were made along three lines totalling 3900 ft (1189 m) in length in the vicinity of the Indian Springs. These resistivity measurements indicated a low resistive zone on a ridge immediately to the east and several hundred feet above the hot spring area (Fig. 6). Similarly, immediately to the west of the hot springs, another low resistive zone was observed which aligned itself with the first low zone to the east (Fig. 6).

Another low resistive zone was located on line A, 900 ft (274 m) south of the first zone between stations 11 and 13. On line C, which is located west of line A (Fig. 6), a low resistive zone was located between stations 4 and 6. An east-west fault may be projected through these two low resistive zones paralleling the fault to the north. Due to the lack of additional resistivity data, the two low resistive zones were not combined. No surface indication of faulting or rock alteration was found in these areas.

The only fault mapped in the area is located 1,600 ft (488 m) south of the hot springs area (Fig. 6), but no resistivity measurements were conducted in this area due to terrain obstacles. In the interpretation of the resistivity pseudo-sections of this area (Fig. 7, 8, & 9), the reader should be aware of the fact that values obtained along the line of traverse may be influenced by lateral variations of unknown features. This could be the case in the Idaho Springs area (see Appendix D for a description of field procedures pertaining to the various arrays employed.) Resistivity calculations for lines A, B & C are presented in Appendix E. Appendix F presents the geometric factors table used to calculate the resistivity values in Appendix E.

Conclusion

Due to cultural and topographic affects, the electrical resistivity surveys were limited to the immediate surroundings of the Indian Springs. Analysis of the data indicates two areas worthy of further consideration pertaining to a potential geothermal resource. The first area is adjacent to the hot springs and trends in a northwest-southeast direction. The second area is located about 900 ft south of the springs and has a similar strike. A mapped fault immediately to the south of this second low resistive zone could be the conduit for the thermal waters to the north. Fractures in the Precambrian bedrock may also serve as conduits at depth for the thermal water.

From the resistivity field work conducted, only the upper 250 to 350 ft of the geothermal system in the Indian Springs area has been delineated. In order to further determine the gradient and heat flow of this area, several geothermal gradient wells of 300 ft depth would have to be drilled. In addition, more resistivity geophysical surveys should be attempted where more control is required. This may be a difficult task due to cultural and terrain obstacles throughout the area.

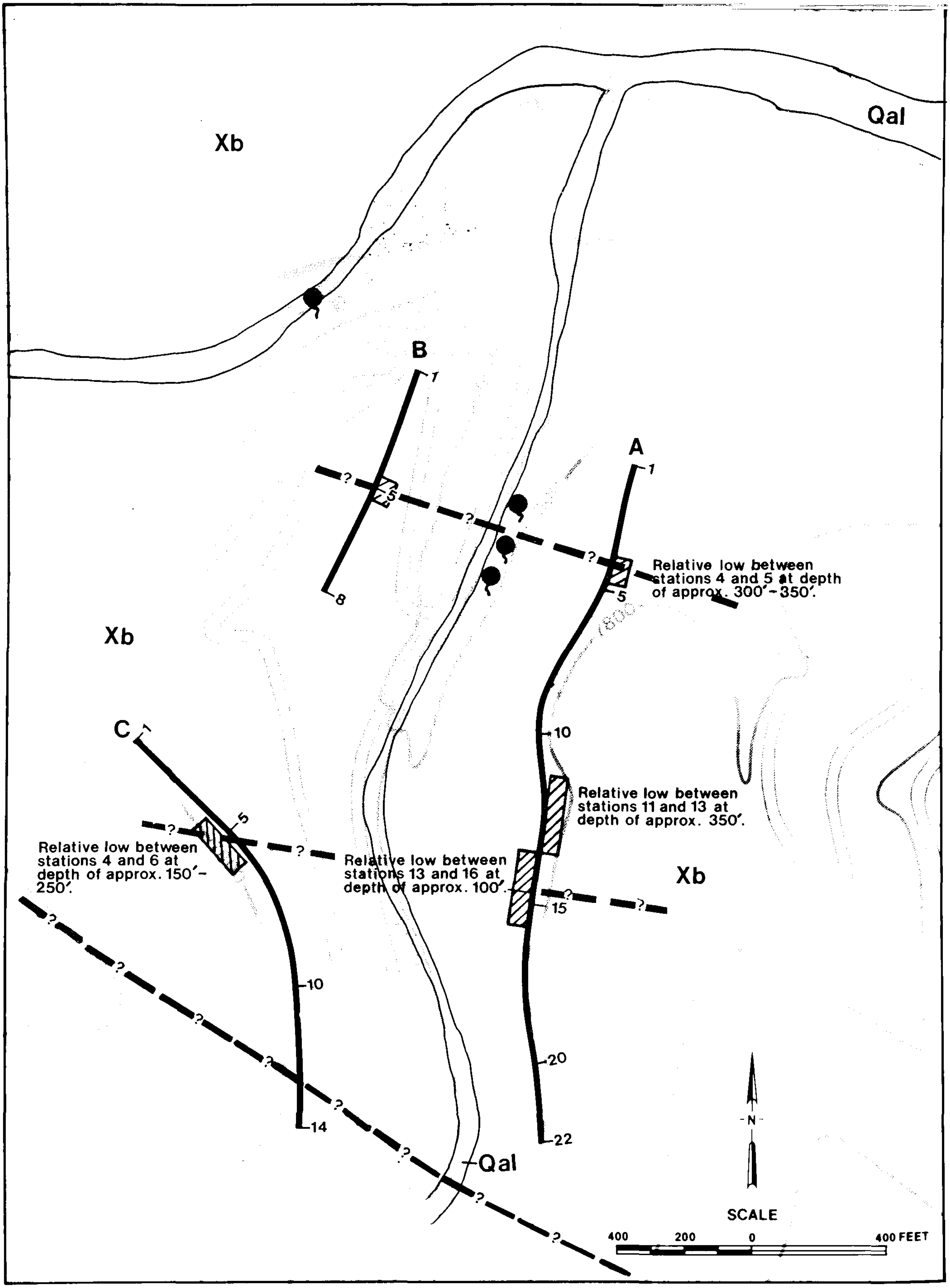


Figure 6. Geophysical resistivity survey.

EXPLANATION

Hot spring

Dipole-dipole resistivity line and station marker and identifying letter

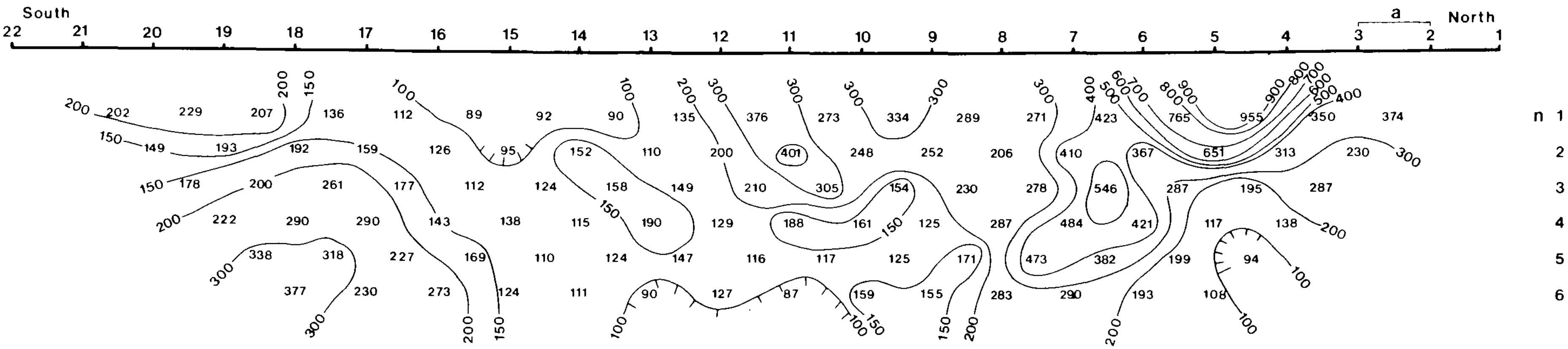
Relative resistive low

Estimated areal extent of low resistivity

Qal	Alluvium	Recent
Xb	Biotite Gneiss	} Precambrian

Contact

—?— Fault; dashes where approximate, queried where probable



Dipole-Dipole Line A: Located 400 ft east of the Indian Springs Resort and 300 ft above and parallel to Soda Creek (Fig. 6). A moderately low resistivity zone is indicated between stations 4 and 5 at an approximate depth of 300 ft to 350 ft (91 to 107 m). This low resistive zone is probably associated with fracture permeability in the Precambrian biotite gneiss bedrock. Additional low resistive zones are indicated between stations 11 and 13, at an approximate depth of 350 ft (107 m) and between stations 14 and 16 at a depth of 100 ft (30 m).

SEPARATION: n Value

DATE: June 13, 16, 1980

TYPE: Dipole Dipole

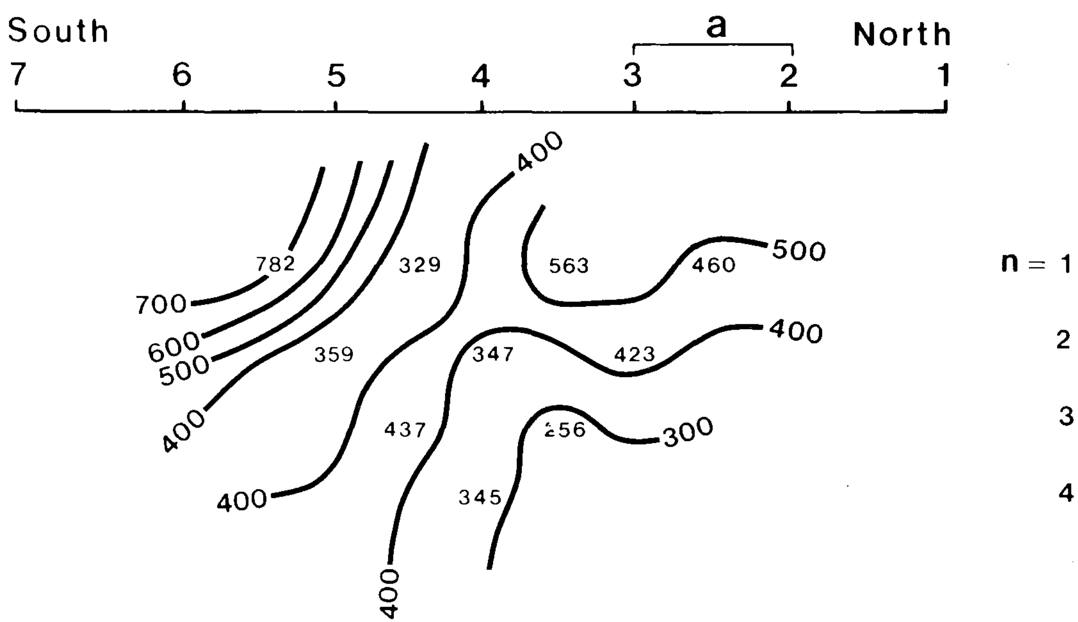
SPREAD: a 100 Feet

RESISTIVITY: In ohm meters

SCALE

0 50 100 FEET

Figure 7. Dipole line A.



Dipole-Dipole Line B: Located on a ridge approximately 500 ft northwest of the hot springs area (Fig. 6). Due to difficult terrain and accessibility conditions this line was only 700 ft (213 m) in length. A highly low resistive zone at stations 3 - 5 aligns itself on the surface with the low resistive zone noted on line A by station 5 and also with the thermal area. A fault could therefore be projected between these zones. However, surface expression of this condition was not evident. In general, the surface resistivity was much higher than the resistivity at depth.

SEPARATION: n Value

DATE: June 17, 1980

TYPE: Dipole-Dipole

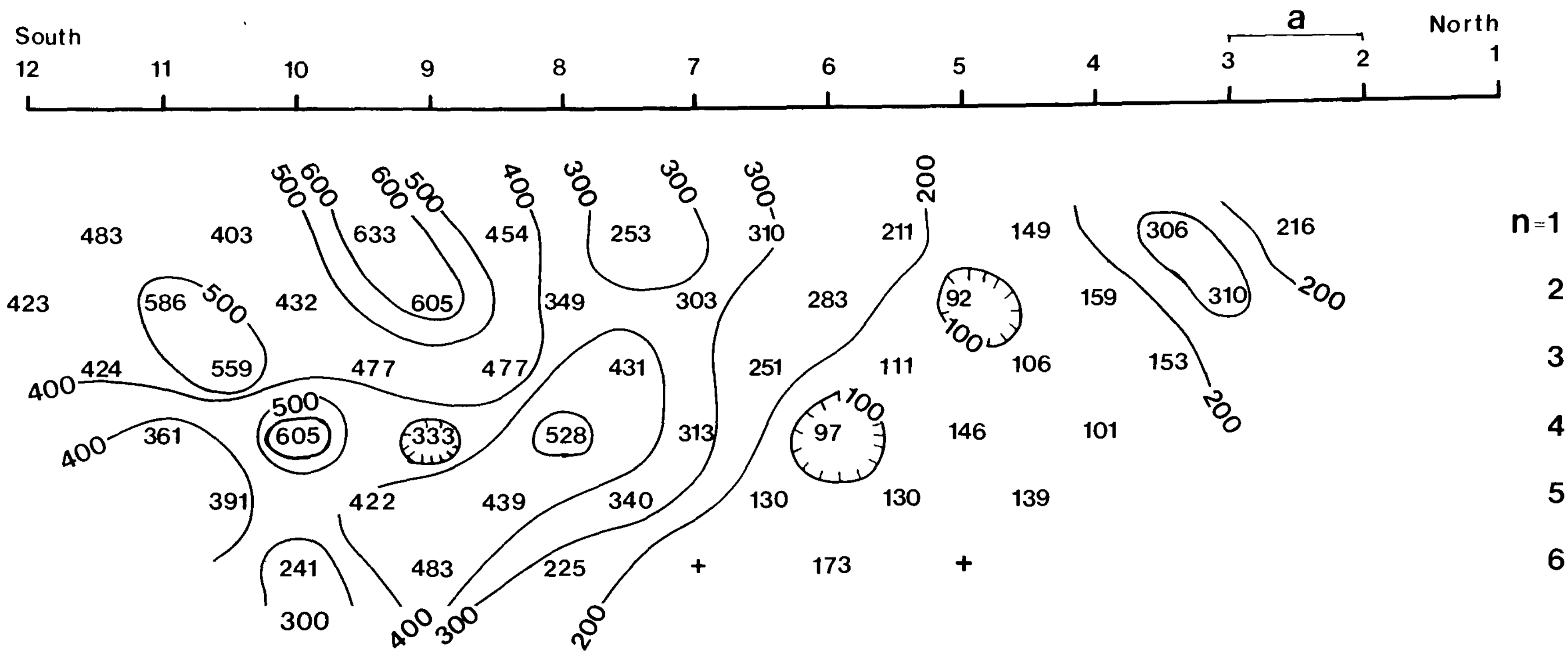
SPREAD: a=100 Feet

RESISTIVITY: In ohm-meters

SCALE

0 50 100 FEET

Figure 8. Dipole line B.



Dipole-Dipole Line C: Located 1100 ft (335 m) southwest of the Indian Springs Resort, along a promontory which parallels Soda Creek (Fig. 6). A low resistive zone is indicated between stations 4 and 6 at depths of approximately 150 ft to 250 ft (46 m to 76 m). This low resistive zone could be associated with a projected low zone noted at station 15 on line A (Fig 7). There is a major fault zone approximately 600 ft (183 m) south of this low resistive zone (Fig. 6) that may have contributed to the lower resistivities with depth. The surface resistivities by comparison are much higher.

SEPARATION: n Value

DATE: June 17, 1980

TYPE: Dipole-Dipole

SPREAD: a= 100 Feet

RESISTIVITY: In ohm-meters

SCALE

0 50 100 FEET

Figure 9. Dipole line C.

SOIL MERCURY SURVEYS

Introduction

The majority of exploration methods used in geothermal exploration are the more common ones such as geology, geophysics, and hydrogeological mapping; however, new methods are beginning to be used. One of these, soil mercury surveys, has proven successful in a number of instances. For example, Capuano and Bamford (1978); Cox and Cuff (1980); Klusman et al (1977); Klusman and Landress, (1979); and Matlick and Buseck (1976) have demonstrated the use of soil mercury surveying as a geothermal exploration tool. Both Matlick and Buseck (1976), and more recently, Cox et al (1980), have used soil mercury surveys on a regional scale. On a detailed scale, Klusman and Landress (1979) and Capuano and Bamford (1978) have shown how soil mercury surveys can delineate faults or permeable zones in geothermal areas. The association of mercury with geothermal deposits has been shown by White (1967). Matlick and Buseck (1976) stated that areas with known thermal activity, such as: Geysers in California; Wairakei, New Zealand; Geyser, Iceland; Larderello, Italy; and Kamchatka in Russia contain mercury deposits.

Matlick and Buseck (1976), in presenting the geochemical theory behind the associations of mercury with geothermal deposits, noted that mercury has great volatility, and the elevated temperatures of most geothermal systems tends to cause the element to migrate upward and away from the geothermal reservoir. In addition, they noted the work of White (1967), and White and others (1970), which showed that relatively high concentrations of mercury are found in thermal waters. Matlick and Buseck (1976) then pointed out that soils in thermal areas should be enriched in mercury, with the mercury being trapped on the surfaces of clays and organic and organometallic compounds.

Matlick and Buseck (1976) presented four case studies where they used soil mercury concentrations as an exploration tool. Three of the four areas tested, Long Valley, California, Summer Lake and Klamath Falls, Oregon indicated positive anomalies. At the fourth area, East Mesa in the Imperial Valley of California, no anomaly was observed, although isolated elevated values were recorded.

Klusman and others (1977) evaluated the soil mercury concentration at six geothermal areas in Colorado. These areas were Routt Hot Springs, Steamboat Hot Springs, Glenwood Springs, Cottonwood Hot Springs, Mt. Princeton Hot Springs, and Poncha Hot Springs. Their sampling and analysis procedures differ from Matlick and Buseck (1976) in that they first decomposed the soils using hydrogen peroxide and sulfuric acid; then a flameless atomic absorption procedure was used to determine the concentration of mercury. They presented the results for only one of six areas sampled, Glenwood Springs. Their survey indicated anomalous zones at Glenwood Springs.

Soil Mercury surveys were run by Capuano and Bamford (1978) at the Roosevelt Utah Hot Springs Known Geothermal Resource Area. They analyzed the soil samples with a Jerome Instrument Corp. gold film mercury detector. The results of their investigation showed that mercury surveys can be useful for identifying and mapping faults and other structures controlling the flow of thermal waters and for delineating areas overlying near-surface thermal activity.

Strategy and Methodology

The aim of the geochemical sampling program by the Colorado Geological Survey was to evaluate those thermal areas deemed to have high commercial development potential. As the time allotted for this program was limited, the soil mercury surveys had to be preliminary in nature. The geochemical sampling program started in 1979 and continued into 1980. The surveys conducted during the summer of 1979 were aimed at determining the structural conditions controlling the hot springs. This approach was strongly influenced by the work of Capuano and Bamford (1978). In 1980 a broader sampling target was selected. Rather than just sampling along traverses located over suspected faults, grid sampling patterns were used. If anomalous mercury concentrations were detected, then follow-up samples were collected at a more detailed level. For those thermal areas where grid sampling was not possible due to lack of access, soil disturbance, or urban development, traverses were chosen in a similar method to the procedure used in 1979.

During the course of the investigations the following restrictions became apparent: urban development; alluvial and colluvial deposits; and mining areas. In urban developments one cannot really be sure whether the surface deposits in the back streets and lawns are original or have been brought in. In sampling alluvial and colluvial surficial deposits such deposits because of their origin, age and mineral content tend to mask, dilute, and/or distort any anomalies. In old mining area the problem becomes whether the mercury concentrations found are caused by mineralization or by geothermal actitivty.

Sampling Methods

At selected sample sites, one to eight samples were taken at points within 15 to 20 ft of each other. The notation of sampling locality is explained in Miesch (1976). The interval between sampling sites depends on the target being considered. For areas investigated, the sample site interval was either 100 ft to 200 ft or 400 ft (30 m to 61 m or 122 m). When using a 400 ft (122 m) interval, the area in the immediate vicinity of the hot spring was considered the target rather than any particular fault. Sampling intervals of 200 ft (61 m) or less were used where attempts were made to delineate controlling faults. This spacing was used by Capuano and Bamford (1978). However, Klusman and Landress (1979) seem to think that the sample must be taken directly over the faulting for detection. Considering the empirical result of Capuano and Bamford (1978), it was believed that some anomalous mercury values should be encountered if a grid pattern encompassing the hot spring area was used. A definite structural pattern may be obvious, but if the study area is being influenced by geothermal activity, the trend should indicate that the hot springs area entirely or partially is high in mercury relative to surrounding area.

The sampling procedure used during 1979 consisted of laying out a series of sample lines across suspected faults in the thermal areas. Samples were collected at predetermined intervals (usually 100 ft) along the lines.

In most of the areas investigated during 1980, three or more samples were taken at random sample localities. This was done to get an estimate of how the variance between sample localities compared with the variance at a sample locality. If the comparison suggested that there is as much variance at a sample locality as there is between sample localites, then the data would be interpreted on a point to point basis. Contouring the data would more than likelv lead to false interpretation.

Two rationals have been used for determining the sampling depth. The method recommended by Capuano and Bamford (1978) is to determine the profile of mercury down to a depth of approximately 16 in (40 cm), the depth at which the profile peaks determines the sampling depth. The other method consistently samples a soil horizon, such as the A or B horizon. The problem with using the A horizon is that its normally high organic content has been shown to have strong secondary effects in controlling mercury in the soil. Also, the sampling depth in the A horizon may not be deep enough to avoid the "baking" effect of the sun.

The method used during 1979 consisted of using profiles to determine sampling depths. A sampling depth of approximately 6 in (15 cm), with an interval of about 0.4 in (1 cm), was used for most of the profiles. During 1980 each sample was taken over an interval of 5 to 7 in (13 to 18 cm). It was hoped that some of variance due to depth would be smoothed out by sampling over a wider interval. Also, at that depth it was hoped that the sun would not be affecting the soil's ability to retain mercury.

To collect a sample, the ground was broken with a shovel to a depth of 9 to 10 in (20 to 25 cm). Then a spatula and metal cup were used to collect approximately 100 grams of material. The contents of the cup were then put in a marked plastic bag. At the end of the day the material in each bag was laid out and allowed to dry overnight. Sometimes it would take more than one night to dry. Normally, the following morning the dried material would be sieved down to an 80 mesh size outside in a shaded area and stored in 4 ml glass vials with screw caps. Within a period of seven days later, the samples were analyzed for mercury using the Model 301 Jerome gold film mercury detector.

Analysis

For an accurate analysis of geochemical data, it is necessary to differentiate between background and anomalous values. There are various statistical ways of accomplishing this. For those areas where the statistical sample approaches 100 samples and a lognormal distribution can be assumed, a method which looks for a break in the cumulative frequency plot of the mercury data can be used. Hopefully, the break distinguishes the two populations -- the background and the geothermal induced population (Capuano and Bamford, 1978; Lepelitor, 1969; and Levinson, 1974).

For those instances where the data was analyzed using a cumulative frequency diagram, the following procedure was used.

1). Determine the number of class intervals by multiplying the logarithm of the sample by 10.

2). Determine the range of each class interval by dividing the maximum recorded value, determined above, by one less.

3). Determine logarithm of top end of each interval.

4). Determine class frequency by calculating the number of values in each class.

5). Determine relative frequency by dividing each class frequency value by total number of values.

6). Construct frequency distribution graph by plotting class frequency log values by cumulative frequency.

7). Note where break in slope of graph occurs.

For those cases where the data was sparce and the values were clustered near the lower detection limit of the instrument with a few high values at the opposite extreme, a more empirical method was used. This method called for arranging the data in ascending numerical order then inspecting the data for any gaps. The anomalous values are differentiated from background values. For the lack of a proper sampling design and computer facilities, the gap between background and the anomaly was chosen subjectively, rather than using a statistical test as recommended by Miesh (1976). When background was determined in this manner, sometimes the anomaly criteria of four times typical background was used to see how it compared with the anomalous results of the ranking method.

As a further aid in determining background mercury values, sample localities were chosen within a mile or two of the study area. Care was taken to try to sample on the same parent material as in the study area. It was assumed that there were no extreme regional trends.

INDIAN SPRINGS SOIL MERCURY SURVEYS

Introduction

To evaluate the Indian Springs area, and to determine if there were other geothermal manifestations present not having a surface expression, a series of profile lines were laid out and 138 samples were collected and analyzed during the summer of 1980 (Fig. 10). The sample lines were designed to cross all suspected controlling structural features in and adjacent to the hot springs.

The location of these lines and the analytical results are shown on Figure 10. During the course of this investigation 109 samples were collected and analyzed in the vicinity of Indian Springs. Nineteen "background" samples were collected 0.6 mi (1 Km) southeast of the springs along Soda Creek. The mercury content of the samples from the Indian Springs area ranged from a low of 0 ppb to a high of over 900 ppb (Tables 2 and 3). The mercury content of the samples collected 0.6 mi southeast of Indian Springs ranged from a low of 0 ppb to a high of 20 ppb (Table 4). The mean soil mercury content was 83 ppb, with a standard deviation of 137 ppb.

Table 2. Mercury content (ppb) of samples collected on west side of Soda Creek (Fig. 10).

0	14
0	16
0	16
0	16
0	20
8	20
9	39
9	40
10	50
12	

Soil Description

The soil in the area, which comes from weathering of the biotite gneiss bedrock, is usually less than 1 ft thick. Thus, the B horizon was very thin, rocky, and sandy; however, sometimes it contained clayey material. The slope of the terrain was quite steep, usually from 20° to 30°, and the vegetation consisted of fir and pine with a thin grass cover.

Sample Analysis

One of the problems apparent at the outset of this study was the extensive mineralization of the Idaho Springs region. For example an old mine dump was found in the study area and traces of alteration were sighted. Thus, a problem arises as to whether the measured mercury anomalies are caused by mineralization or by geothermal activity.

Enough data was collected in the study area so that its distribution could be statistically analyzed. Because of the high contrast in sampled results between the east and west slopes of Soda Creek, the areas were considered separately.

Using methods described earlier, the analytical data was analyzed statistically in order to construct a frequency distribution (Fig. 11). It was calculated that there were 19 class intervals having a range of 50 ppb each. When the logarithm of the mercury concentration for each interval was plotted against the cumulative frequency distribution, it was noted that a change occurred in the slope of the curve at a log mercury value of 2.3. This value corresponded to the 150 - 199 ppb class interval. Therefore, all soil mercury values greater than 150 ppb were considered as anomalous. This value of 150 ppb is much higher than the 1 ppb to 20 ppb for 21 localities sampled about .6 miles southeast of the lodge (Table 4). The eastern slope of Soda Creek above the lodge is definitely anomalous compared to the surrounding area.

The mercury values of the localities on the western side of Soda Creek (Table 3) range from less than 1 ppb to 50 ppb, much more in line with the values from the localities outside the study area. Though the probability for the localities marked as anomalies is not as great as those on the eastern side of Soda Creek, they should be considered until other surveys prove them false. The criteria for the anomalies is based upon the gas in rank ordered data for the western slope and the data collected outside of the study area.

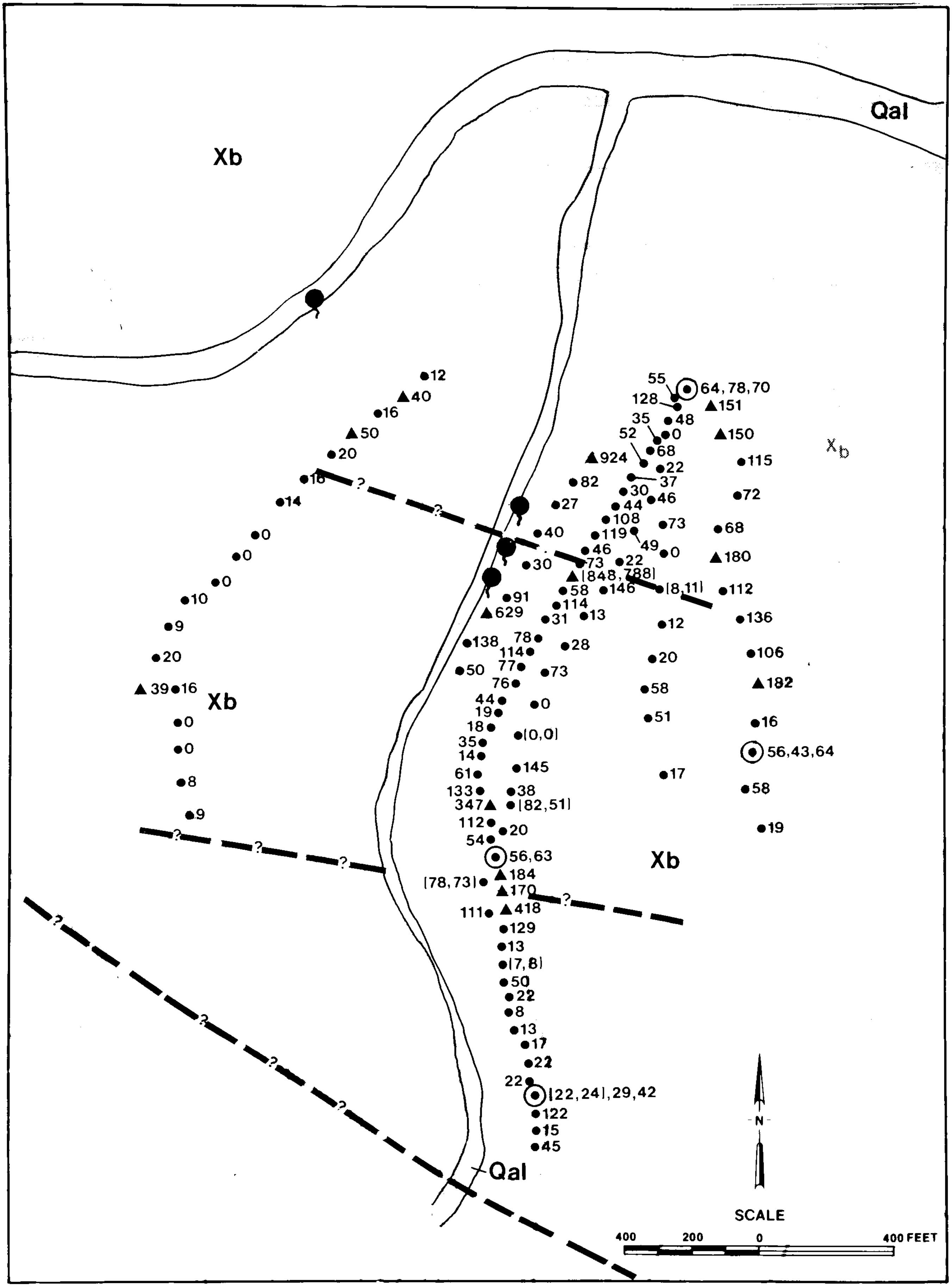


Figure 10. Soil Mercury survey

EXPLANATION

Hot spring

Sampling locality

Anomalous sample value at a locality

Values taken at a locality with two or more samples collected approximately 20 ft apart

At least one anomalous sample value at a locality with two or more samples collected approximately 20 ft apart

7,8,345
(3,9) Each value indicates the analysis of a single sample in ppb of mercury. Values in parentheses indicate more than one analyses of a single sample.

Qal	Alluvium	Recent
Xb	Biotite Gneiss	Precambrian

Contact

Fault; dashes where approximate, queried where probable

Table 3. Mercury content* (ppb) of samples collected on east side of Soda Creek (Fig. 10). Arranged in ascending rank order.

0	31	63	128
0	31	66	129
0	35	68	133
7	35	68	136
8	37	70	138
13	38	72	145
13	40	73	146
13	44	73	150
14	44	75	151
15	45	76	170
16	46	77	180
17	46	78	182
18	48	82	184
19	49	91	347
19	50	106	418
20	50	108	629
22	52	111	818
22	54	112	924
22	54	112	
22	55	114	
27	56	114	
28	58	115	
30	58	119	
30	61	122	

*Represents just the first value recorded at a sampling locality.

Table 4. Mercury content (ppb) of samples collected 0.6 mi southeast of Indian Spring Lodge.

0	11
0	11
0	12
5	13
6	14
6	14
7	15
8	20
8	

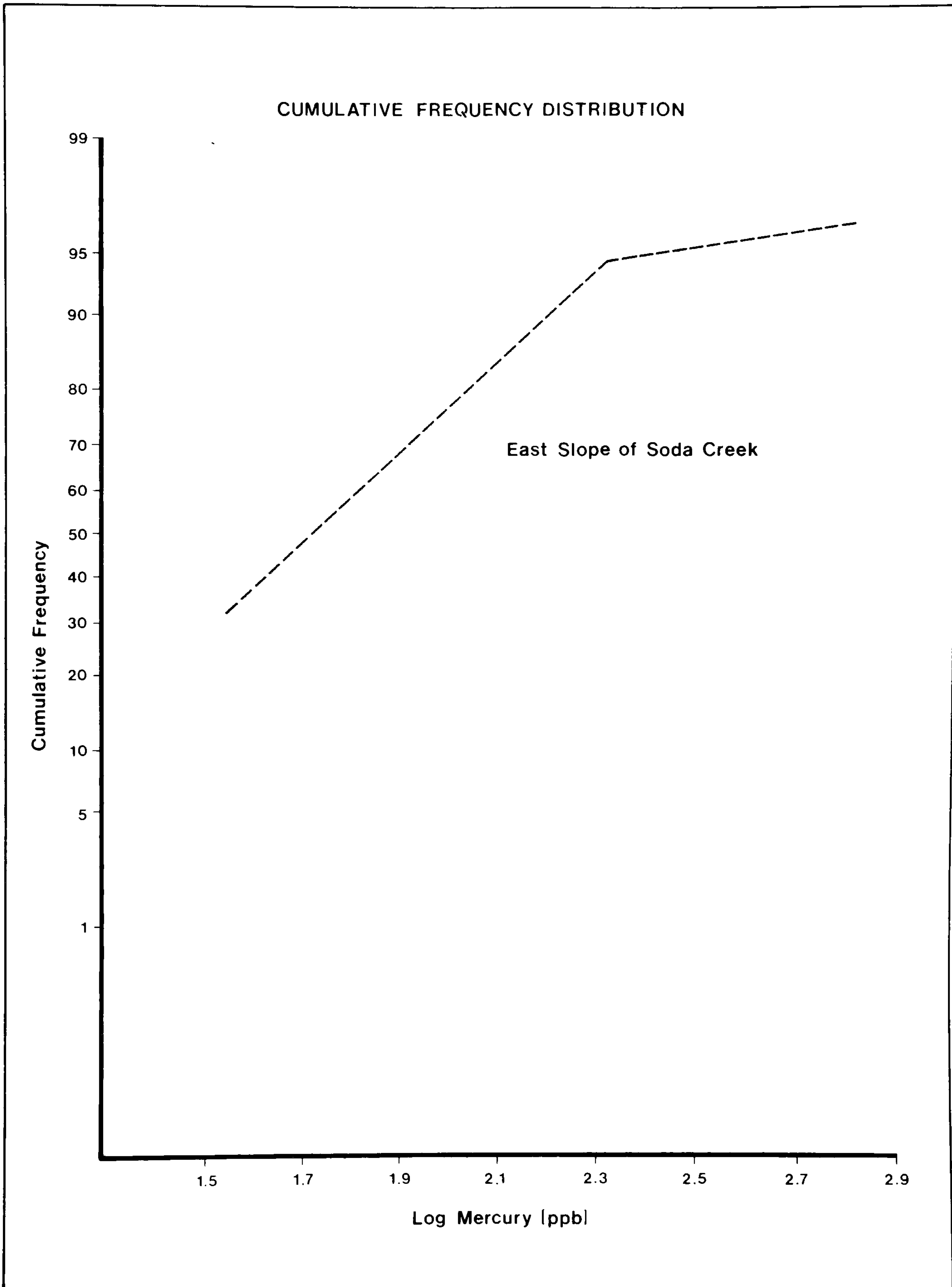


Figure 11. Soil Mercury cumulative frequency distribution.

Summary

The soil mercury survey for this area shows a definite anomalous zone on the lower slope east of Soda Creek above the lodge. There is also limited evidence of the anomalous trend existing on the west side. Further investigations may determine whether this anomalous zone is caused by mineralization or geothermal activity.

SUMMARY AND CONCLUSIONS

In 1980 the Colorado Geological Survey, with funding from the U.S. Department of Energy/Division of Geothermal Energy, initiated an investigation to evaluate the geothermal resources of the Idaho Springs region. Upon initiation of this investigation, it became apparent that due to cultural and topographic affects, this investigation would have to be limited to the area immediately surrounding the thermal Indian Springs on the southside of the city.

The investigation conducted consisted of the following facets: library research, field geological reconnaisance investigation, electrical resistivity geophysical surveys, soil mercury geochemical surveys, and hydrogeological modeling.

The geochemical and geophysical surveys conducted near the hot springs showed that the thermal waters most likely are fault controlled. As part of their preliminary evaluation of the Idaho Springs geothermal resources Barrett and Pearl in 1978 ran geothermometer model analyses. These models showed that the maximum reservoir temperatures may range between 178°F (81°C) and 446°F (230°C). They cautioned that these temperatures are unreliable due to the ambiguous geochemistry of the thermal waters and, until they are proven by deep drilling, should be used as guides only. An estimate of the size of the Indian Springs thermal reservoir by Pearl (1981) noted that it may be restricted to an area approximately .152 sq mi (.256 sq km) in extent.

Studies at the Mount Princeton Hot Springs in Colorado and elsewhere in the world have shown that most thermal waters are of meteoric origin. Hydrogeological models developed for the Idaho Springs region based on geological evidence indicate that the thermal waters are probably of meteoric origin. However, they also could be of magmatic origin or a mixture of the two. Thermal waters of meteoric origin originate as deep circulation of normal groundwaters along faults in an area of above normal heat flow. Recharge of the thermal system occurs from melting snows and precipitation falling on the surrounding highlands. Thermal magmatic waters would be waters driven of from the cooling of batholiths which have been postulated to underlie the Colorado Mineral Belt.

The geothermal resources of the Idaho Springs area do not appear to be of extremely high temperatures and the reservoir probably does not extend over a large geographic area. Due to the apparent low subsurface temperature of the resource, it most likely would be suited for direct uses such as space heating, recreation, or some light industry requiring low temperature heat.

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APPENDIX A. PHYSICAL PROPERTIES AND CHEMICAL ANALYSIS OF INDIAN SPRINGS THERMAL WATERS (From Barrett and Pearl, 1976)

	Spg. A	Spg. B	Spg. C	Well C
Arsenic, (UG/L)	20	12	2	46
Boron, (UG/L)	350	370	170	360
Cadium, (UG/L)	0	0	1	0
Calcium, (MG/L)	140	130	77	150
Chloride, (MG/L)	66	69	36	66
Fluoride, (MG/L)	4.8	4.8	2.9	3.5
Iron, (UG/L)	20	20	40	1,000
Lithium, (UG/L)	640	660	340	870
Magnesium, (MG/L)	36	50	23	38
Manganese, (UG/L)	40	20	40	70
Mercury, (UG/L)	0	0.1	0	0
Nitrogen, (MG/L)	0.13	0.08	0.13	0
Phosphate				
Ortho diss. as P, (MG/L)	0.11	0.06	0.01	0.05
Ortho, (MG/L)	0.34	0.18	0.03	0.15
Potassium, (MG/L)	80	82	44	82
Selenium, (UG/L)	0	0	0	0
Silica, (MG/L)	68	68	45	58
Sodium, (MG/L)	500	520	260	520
Sulfate, (MG/L)	380	400	210	420
Zinc, (UG/L)	10	10	20	10
Alkalinity				
As Calcium Carb., (MG/L)	1,240	1,250	623	1,220
As Bicarbonate, (MG/L)	1,510	1,520	759	1,490
Hardness				
Noncarbonate, (MG/L)	0	0	0	0
Total, (MG/L)	500	530	290	530
Specific Conductance (Micromohs)	3,400	2,900	1,620	2,920
Total Dissolved Solids (TDS), (MG/L)	2,020	2,070	1,070	2,070
ph, Field	6.9	-	-	6.9
Discharge (gpm)	21	-	1	30
Temperature (^{o}C)	45	24	20	46
Date Sampled	7/75	7/75	7/75	10/75
Remarks				Formerly called Lodge Well

APPENDIX B

FACTORS AFFECTING RESISTIVITY

Electrical resistivity geophysical methods used in geothermal exploration measure the electrical resistivity of rocks at various depths. Temperature, porosity, salinity of fluids, and the content of clays will normally be higher within the geothermal reservoir than in the surrounding subsurface rocks. Consequently, the electrical resistivity in thermal reservoirs is low compared to the surrounding rock. Basically, resistivity methods utilize manmade currents which enter the subsurface via two electrodes with the resultant potential measured at two other electrodes (Soil Test Inc., 1968).

The difficulty with interpretation stems from the fact that resistivity is a complicated function of the following parameters: temperature, porosity, salinity, and clay content. For example, a low temperature, highly saline ground water can provide the identical low resistivity anomaly as a high temperature, moderatately saline geothermal system. Therefore, to be most effective, this method should be used in conjuction with direct temperature gradient measurements and other types of data that are of value in determining the reason for the resistivity values obtained (Soil Test Inc., 1968).

Zones of low resistivity in a geothermal environment can be caused by a high dissolved solid content of thermal water versus ground water, higher clay content due to the hydrothermal alteration within the fault zones, and the higher temperature of the thermal fluids. Finally, the ability of the geophysicist to isolate any of the aforementioned factors and relate it to the objective of the resistivity exploration program rests upon a combination of elimination processes of constant or slowly varying factors from those that are most susceptible to change.

APPENDIX C

SCINTREX RAC-8 LOW FREQUENCY RESISTIVITY SYSTEM

The following description is taken from the Scintrex Manual (1971).

The Syntrex RAC-8 electrical resistivity equipment used by the Colorado Geological Survey is a very low frequency AC resistivity system with high sensitivity over a wide measuring range. The transmitter and receiver operate independent of each other, requiring no references wires between them. This allows a great deal of efficiency and flexibility in field procedures and eliminates any possibility of interference from current leakage or capacitive coupling within the system.

The transmitter produces a 5Hz square wave output at a preset electronically stabilized, constant current amplitude. The output current level is switch selectable at any one of five values ranging from 0.1 to 333 milliamps.

The receiver is a high sensitivity phase lock, synchronous detector which locks onto the transmitter signal to make the resistivity measurement. When set at the same current setting as the transmitter, the receiver gives a direct readout of V/I ratio.

The RAC-8, with a measuring range from .0001 to 10,000 ohms, high sensitivity to weight ratio, gives fast, accurate resistivity data. With the low AC operating frequency, good penetration may be obtained in excess of 1500 ft under favorable conditions. The system has an output voltage maximum 1000 V peak to peak. However, the actual output voltage depends on the current level and load resistance. The output power under optimum conditions approaches 80 watts.

In areas of very low resistive lithology, the penetration power was reduced by a sizeable amount. Realizing the aforementioned constraint, the intent was to delineate gross potential differences in resistivity. In some areas where the lithology reflected small differences in resistivity, the RAC-8 system appeared to average the penetrated lithologic sequences rather than picking up distinct breaks. Considering cost and time constraints, the system performed as indicated and performed best in areas of high resistivity.

APPENDIX D

RESISTIVITY FIELD PROCEDURES

One of the most widely used electrical processing techniques for geothermal resource exploration is the resistivity profiling and sounding method. The method utilizes various arrays, but the most common are the Wenner, the Schlumberger and the Dipole-Dipole schemes. The Colorado Geological Survey extensively employed the latter method primarily because of the ease of use and also being able to obtain horizontal and vertical sections.

Before discussing the various electrode methods used, it is necessary to consider what is actually measured by an array of current and potential electrodes (Fig. 12). By measuring (V) and current (I) and knowing the electrode configuration, a resistivity (p) is obtained. Over homogeneous isotropic ground this resistivity will be constant for any current and electrode arrangement. That is, if the current is maintained constant and the electrodes are moved around, the potential voltage (V) will adjust at each configuration to keep the ratio (V/I) constant (Sumner, 1976).

If the ground is nonhomogeneous, however, and the electrode spacing is varied, or the spacing remains fixed while the whole array is moved, then the ratio will in general change. This results in a different value of P for each measurement. Obviously, the magnitude is intimately involved with the arrangement of electrodes.

This measured quantity is known as the apparent resistivity, Pa. Although it is diagnostic of the actual resistivity of a zone in the vicinity of the electrode array, this apparent resistivity is definitely not an average value. Only in the case of homogeneous ground is the apparent value equivalent to the actual resistivity (Sumner, 1976).

The following formula is used by all methods to calculate the apparent resistivity at a site.

General Resistivity Formula

$$P_a = 2PIaV/I$$

a = Spread length
V/I = Voltage current ratio
Pa = apparent resistivity
2PI = 6.2

Wenner Array

In the Wenner Spread (Fig. 13) the electrodes are uniformly spaced in a line (Fig 13) (Sumner, 1976). In spite of the simple geometry, this arrangement is often quite inconvenient for field work and has some disadvantages from the theoretical point of view as well. For depth exploration using the Wenner Spread, the electrodes are expanded about a fixed center, increasing the spacing in steps. For lateral exploration or mapping the spacing remains constant and all four electrodes are moved along the line, then along another line, and so on. In mapping, the apparent resistivity for each array position is plotted against the center of the spread.

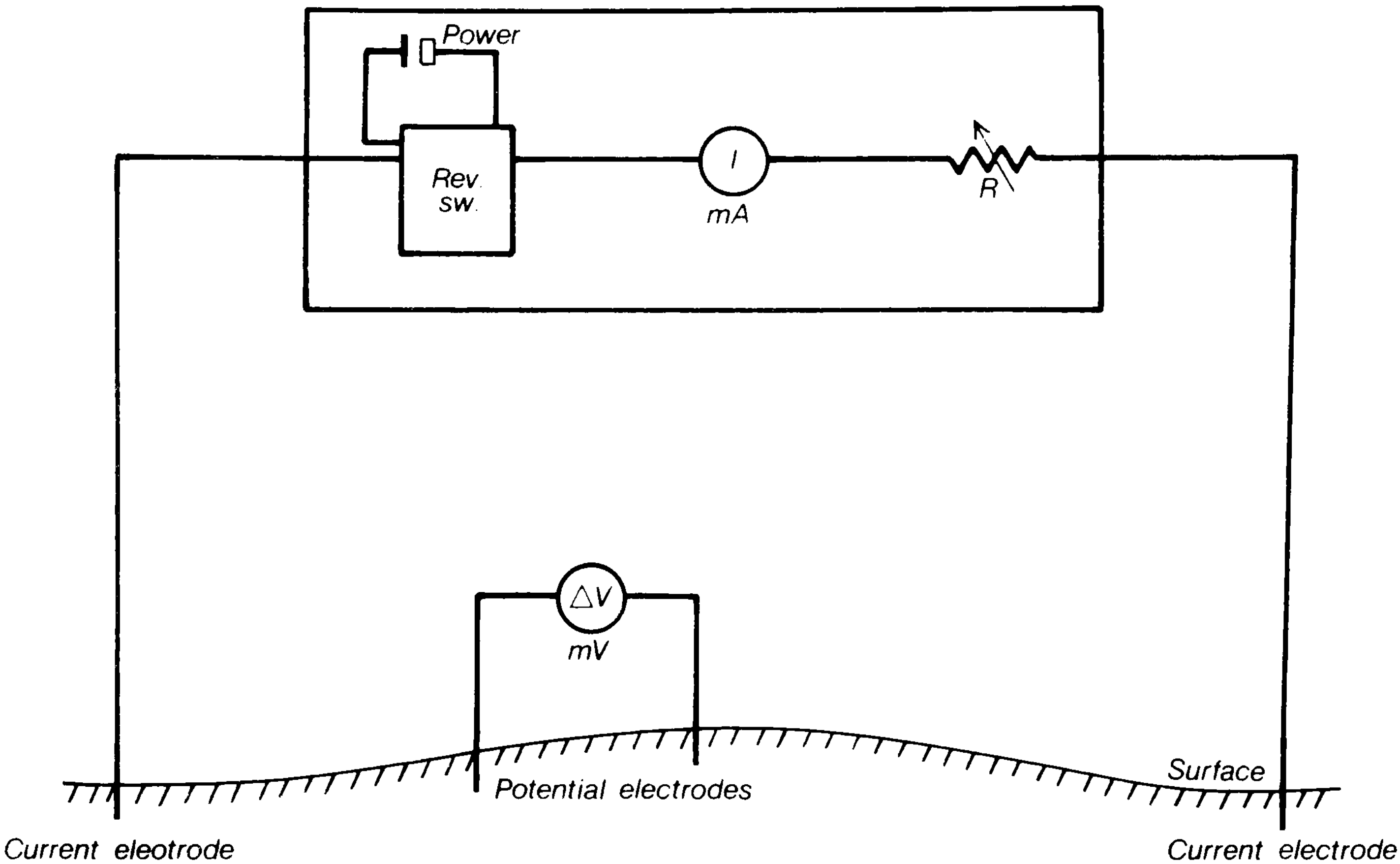
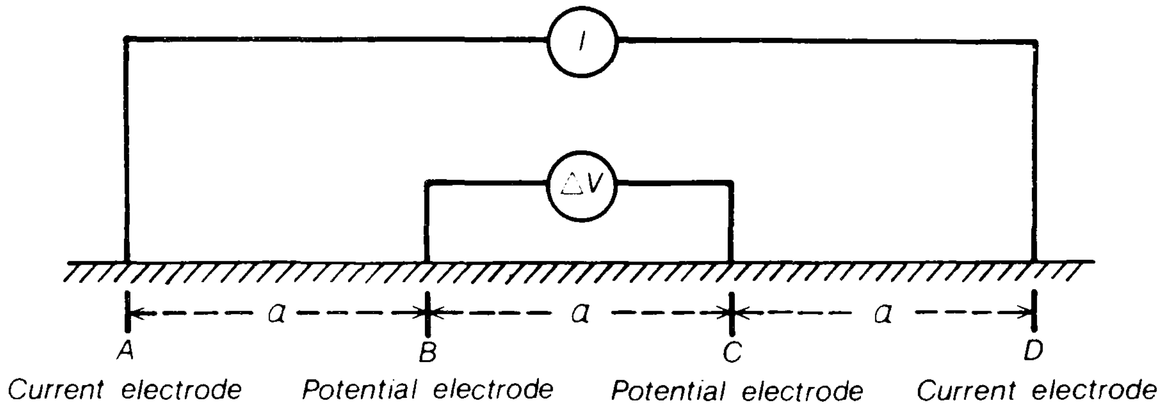


Figure 12. Schematic diagram for resistivity (From Combs, 1980).



$$\rho_a = 2\pi a(\Delta V/I)$$

Figure 13. Wenner array (From Combs, 1980).

This method was not used in the Idaho Springs area due to steep terrain and access problems.

Schlumberger Array

For the Schlumberger array, the current electrodes are spaced much further apart than the potential electrodes (Fig. 14).

In depth probing the potential electrode remains fixed while the current elecrode spacing is expanded symmetrically about the center of the spread. For large values of L it may be necessary to increase 2 1 also in order to maintain a measurable potential. This procedure is more convenient than the Wenner expanding spread because only two electrodes need move. In addition, the effect of shallow resistivity variations is constant with fixed potential spread (Sumner, 1976).

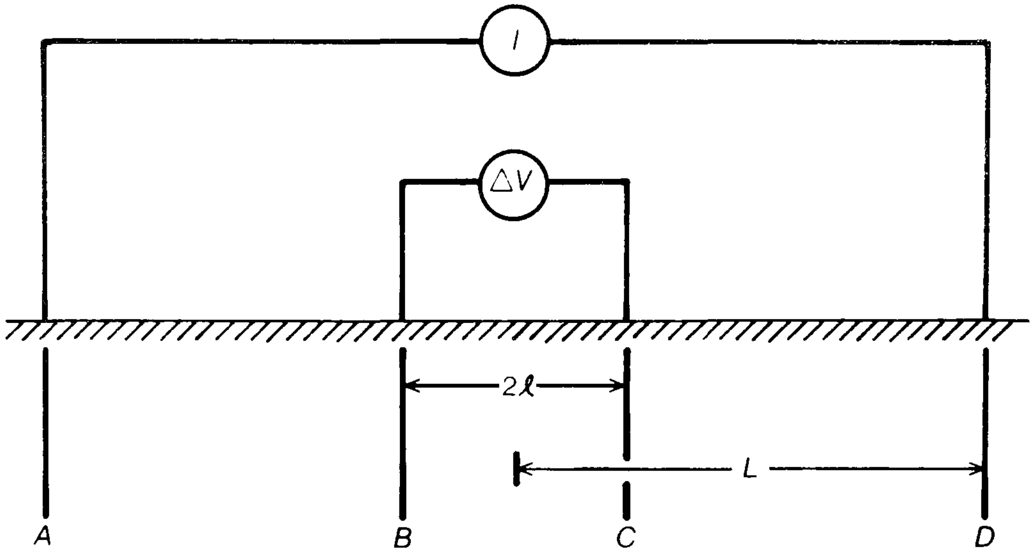
In summary, short spacing between the outer electrodes assumes shallow penetration of current flow and computed resistivity will reflect properties of shallow depth. As the electrode spacing is increased, more current penetrates to greater depth and conducted resistivity will reflect properties of each material at greater depth. This method was used on a few lines for sampling purposes in array.

Dipole-Dipole Array

The potential electrodes are closely spaced and remote from the current electrodes which are close together. There is a separation between C and P , usually 1 to 5 times the dipole lengths (Fig. 15).

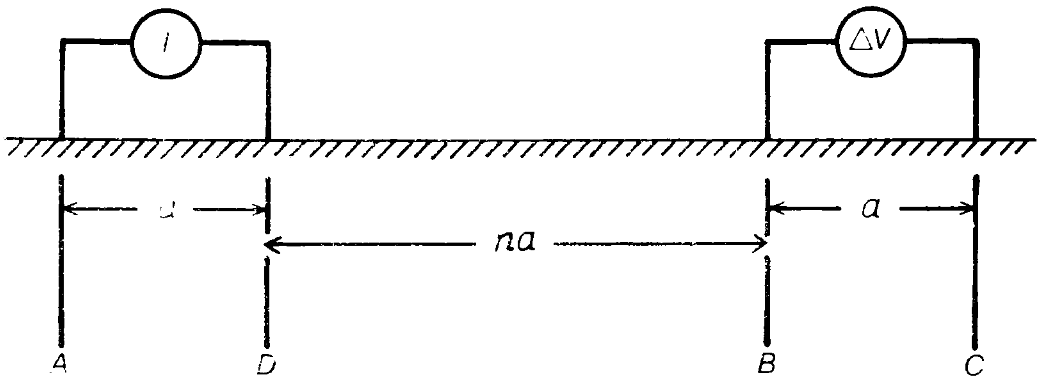
Inductive coupling between potential and current cables is reduced with this arrangement. This method was primarily used throughout all study areas because of reliability and ease of field operation. A diagram of this method is depicted in Figures 16 and Figure 17.

With reference to Figure 16 and 17, an in-line 100 foot dipole-dipole electrode geometry was used. Measurements were made at dipole separations of n = 1, 2, 3, 4, 5. The apparent resistivities have been plotted as pseudosections, with each data point being plotted at the intersections of two lines drawn at 45^{o} from the center of the transmitting and receiving dipoles. This type of survey provides both resolution of vertical and horizontal resistivity contrasts since the field procedures generate both vertical sounding and horizontal profile measurements. The principal advantage of this technique is that it produces better geologically interpretable results than the other two methods (Wenner, Schlumberger). In addition, the dipole-dipole array is easier to maneuver in rugged terrain than either of the other methods. Its main disadvantage compared to the Schlumberger array is that is usually requires more current, and therefore a heavier generator for the same penetration depth. However, this advantage is not sufficient compensation for the difficulties encountered in making geologic interpretation from the resulting data (Sumner, 1976).



$$\rho_a = \frac{\pi L^2}{2\ell}(\Delta V/I)$$

Figure 14. Schlumberger array (From Combs, 1980).



$$\rho_a = \pi n(n+1)(n+2)a(\Delta V/I)$$

Figure 15. Dipole-dipole array (From Combs, 1980).

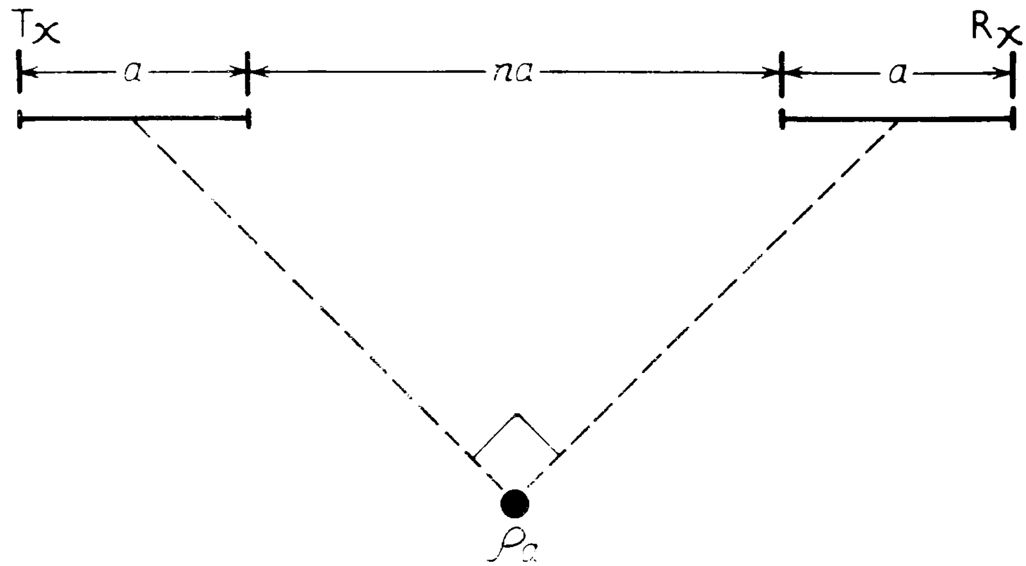


Figure 16. Data plotting scheme for dipole-dipole array (From Combs, 1980).

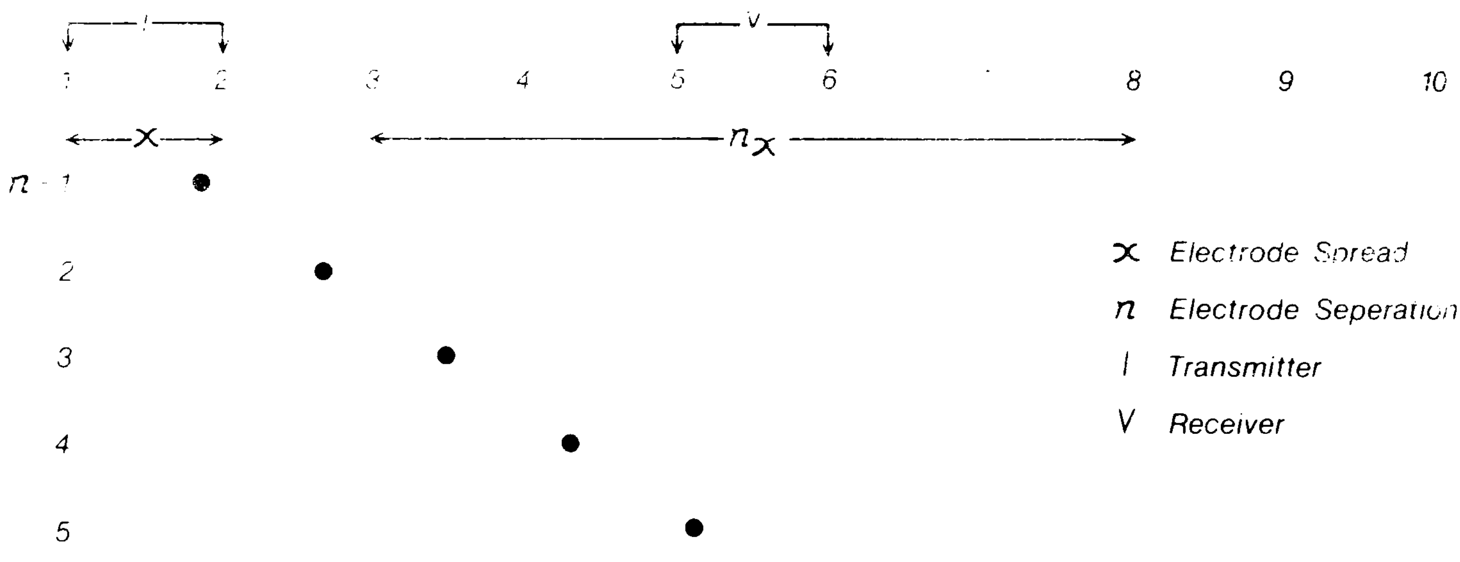


Figure 17. Typical dipole-dipole array (From Combs, 1980).

APPENDIX E. RESISTIVITY CALCULATIONS

TABLE 5. LINE A.

COLORADO GEOLOGICAL SURVEY
Geophysical Exploration
(Resistivity Survey)

LOCATION	PROJECT	DATE
Idaho Springs, Colo.	Line A	13 June 1980
CHIEF OPERATOR	ASSISTANTS	METHOD
Jay Jones	Fargo and Treska	Dipole-Dipole (Nx100')

Sta.	Range	MA	Voltage	V_P	DV/I	G.F.	P_a
1-2							
3-4	1	-1	100	.65	.065	575	374
4-5	1	-2	100	1.00	.100	2299	230
5-6	1	-2	100	0.35	.035	5747	287
6-7	1	-2	100	0.12	.062	11493	138
7-8	0	-2	100	.37	.0037	20113	74
8-9	0	-2	100	.335	.00335	32182	108
2-3							
4-5	1	-2	66	6.08	.608	575	350
5-6	1	-2	66	1.36	.136	2299	313
6-7	1	-2	66	0.34	0.34	5747	195
7-8	0	-2	66	1.02	.0102	11493	117
8-9	0	-2	100	0.99	.0099	20113	199
9-10	0	-3	250	6.08	.00608	32182	193
3-4							
5-6	2	-2	100	1.61	1.61	575	955
6-7	1	-2	100	2.83	.283	2299	651
7-8	0	-2	100	5.00	.0500	5747	287
8-9	0	-2	100	3.66	.0366	11483	421
9-10	0	-2	100	1.90	.0190	20113	382
10-11	1	-2	100	0.09	.009	32182	290
4-5							
6-7	2	-2	100	1.33	1.33	575	765
7-8	1	-2	100	1.67	.167	2299	367
8-9	1	-2	100	0.95	.095	5747	546
9-10	0	-2	100	4.21	.0421	11483	484
10-11	1	-3	400	2.35	.0235	20113	473
11-12	0	-3	433	8.80	.0088	32182	283

LEGEND: Range = Gain
MA = Dummy TX Current Switch
Vp = Balance Control to Null Meter
G.F. = Geometric Factor
Pa = Apparent Resistivity

TABLE 5. LINE A (CONT.)

COLORADO GEOLOGICAL SURVEY
Geophysical Exploration
(Resistivity Survey)

LOCATION	PROJECT	DATE
Idaho Springs, Colo.	Line A	13 June 1980
CHIEF OPERATOR	ASSISTANTS	METHOD
Jay Jones	Fargo and Treska	Dipole-Dipole (Nx100')

Sta.	Range	MA	Voltage	V_p	DV/I	G.F.	P_a
5-6							
7-8	1	-2	166	7.36	0.736	575	423
8-9	1	-2	66	1.79	0.179	2299	410
9-10	0	-2	66	4.83	0.0 83	5747	278
10-11	1	-2	66	0.25	0.025	11483	287
11-12	0	-3	275	8.50	.0085	20113	171
12-13	0	-3	275	4.80	.0048	32182	155
6-7							
8-9	1	-2	66	4.71	0.471	575	271
9-10	0	-2	66	8.97	.0897	2299	206
10-11	0	-2	66	4.00	.040	5747	230
11-12	1	-3	275	1.09	.0109	11483	125
12-13	1	-3	275	0.62	.0062	20113	125
13-14	0	-3	275	4.95	.00495	32182	159
7-8							
9-10	1	-2	66	4.16	.416	575	239
10-11	1	-2	66	1.10	.110	2299	252
11-12	1	-2	66	0.24	0.0268	5747	154
12-13	1	-3	100	1.40	0.0140	11483	161
13-14	1	-3	100	0.58	.0058	20113	117
14-15	1	-3		0.27	.0027	32182	87
8-9							
10-11	1	-2	66	5.81	0.581	575	334
11-12	1	-2	66	1.08	0.108	2299	248
12-13	1	-3	133	5.30	0.0530	5747	305
13-14	1	-3	133	1.64	0.0164	11483	188
14-15	0	-3	133	5.76	.00576	20113	116
15-16	0	-3	133	3.95	.00395	32182	127
9-10							
11-12	1	-2	66	4.74	0.474	575	273
12-13	1	-2	66	1.75	0.175	2299	401
13-14	1	-3	166	3.64	0.0364	5747	210
14-15	1	-3	166	1.12	0.0112	11483	129
15-16	0	-3	166	7.32	.00732	20113	147
16-17	0	-3	166	2.80	.0028	32182	90

TABLE 5. LINE A (CONT.)

COLORADO GEOLOGICAL SURVEY
Geophysical Exploration
(Resistivity Survey)

LOCATION	PROJECT	DATE
Idaho Springs, Colo.	Line A	13 June 1980
CHIEF OPERATOR	ASSISTANTS	METHOD
Jay Jones	Fargo and Treska	Dipole-Dipole (Nx100')

Sta.	Range	MA	Voltage	V_P	DV/I	G.F.	P_a
10-11							
12-13	1	-2	66	6.54	0.654	575	376
13-14	0	-2	66	8.68	0.0868	2299	200
14-15	1	-3	133	2.60	0.0260	5747	149
15-16	1	-3	133	1.65	0.0165	11493	190
16-17	0	-3	133	6.73	0.00613	20113	124
17-18	0	-3	133	3.46	0.00346	32182	111
11-12							
13-14	1	-2	66	2.34	0.234	575	135
14-15	1	-3	166	4.77	0.0477	2299	110
15-16	1	-3	166	2.75	0.0275	5747	158
16-17	1	-3	166	1.00	0.010	11493	115
17-18	0	-3	166	5.46	0.00546	20113	110
18-19	0	-3	166	3.86	0.00386	32182	124
12-13							
14-15	1	-2	66	1.57	0.157	575	90
15-16	0	-2	66	6.63	0.0663	2299	152
16-17	0	-2	66	2.15	0.0215	5747	124
17-18	1	-3	166	1.20	0.0120	11493	138
18-19	0	-3	166	8.40	0.0084	20113	169
19-20	0	-3	166	8.48	0.00848	32182	273
13-14							
15-16	2	-3	133	1.60	0.16	575	92
16-17	1	-3	133	4.15	0.0415	2299	95
17-18	1	-3	133	1.94	0.0194	5747	112
18-19	1	-3	133	1.24	0.0124	11493	143
19-20	1	-3	133	1.13	0.0113	20113	227
20-21	0	-3	133	7.17	0.00717	32182	230
14-15							
16-17	2	-3	100	1.54	0.154	575	89
17-18	1	-3	100	5.45	0.0545	2299	126
18-19	1	-3	100	3.08	0.0308	5747	177
19-20	1	-3	100	2.52	0.0252	11493	290
20-21	1	-3	100	1.58	0.0158	20113	318
21-22	1	-3	100	1.17	0.00117	32182	377

TABLE 5. LINE A (CONT.)

COLORADO GEOLOGICAL SURVEY
Geophysical Exploration
(Resistivity Survey)

LOCATION	PROJECT	DATE
Idaho Springs, Colo.	Line A	13 June 1980
CHIEF OPERATOR	ASSISTANTS	METHOD
Jay Jones	Fargo and Treska	Dipole-Dipole (Nx100')

Sta.	Range	MA	Voltage	V_p	DV/I	G.F.	P_a
15-16							
17-18	2	-3	100	1.94	0.194	575	112
18-19	1	-3	100	6.90	0.069	2299	159
19-20	1	-3	100	4.55	0.0455	5747	261
20-21	1	-3	100	2.52	0.0252	11493	290
21-22	1	-3	100	1.68	0.0168	20113	338
16-17							
18-19	1	-2	66	2.36	0.236	575	136
19-20	0	-2	66	8.33	0.0833	2299	192
20-21	0	-2	66	3.48	0.0348	5747	200
21-22	1	-3	133	1.93	0.0193	11493	222
17-18							
19-20	1	-2	66	3.60	0.360	575	207
20-21	1	-3	133	8.41	0.0841	2299	193
21-22	1	-3	133	3.10	0.031	5747	178
18-19							
20-21	1	-2	66	3.98	.398	575	229
21-22	0	-2	66	6.48	.0648	2299	149
19-20							
21-22	1	-2	66	3.52	.352	575	202

TABLE 6. LINE B.

COLORADO GEOLOGICAL SURVEY
Geophysical Exploration
(Resistivity Survey)

LOCATION	PROJECT	DATE
Idaho Springs, Colo.	Line A	13 June 1980
CHIEF OPERATOR	ASSISTANTS	METHOD
Jay Jones	Fargo and Treska	Dipole-Dipole (Nx100')

Sta.	Range	MA	Voltage	V_P	DV/I	G.F.	P_a
7-6							
5-4	2	-2	66	1.36	1.36	575	782
4-3	1	-2	66	1.56	0.156	2299	359
3-2	2	-3	333	0.76	0.076	5747	437
2-1	1	-3	333	3.00	0.030	11493	345
6-5							
4-3	2	-3	275	5.73	0.573	575	329
3-2	2	-3	275	1.51	0.151	2299	347
2-1	1	-3	275	4.46	0.0446	5747	256
5-4							
3-2	3	-3	250	0.98	0.9 0	575	563
2-1	2	-3	250	1.84	0.184	2299	423
4-3							
2-1	3	-3	133	0.80	0.800	575	460

LEGEND:
- Range = Gain
- MA = Dummy TX Current Switch
- Vp = Balance Control to Null Meter
- G.F. = Geometric Factor
- Pa = Apparent Resistivity

APPENDIX E. RESISTIVITY CALCULATIONS

TABLE 7. LINE C.

COLORADO GEOLOGICAL SURVEY
Geophysical Exploration
(Resistivity Survey)

LOCATION	PROJECT	DATE
Idaho Springs, Colo.	Line A	13 June 1980
CHIEF OPERATOR	ASSISTANTS	METHOD
Jay Jones	Fargo and Treska	Dipole-Dipole (Nx100')

Sta.	Range	MA	Voltage	V_p	DV/I	G.F.	Pa
1-2							
3-4	1	-2	66	3.75	0.375	575	216
4-5	1	-2	66	1.35	0.135	2299	310
5-6	1	-3	200	2.66	0.0266	5747	153
6-7	1	-3	200	0.85	0.0085	11493	101
7-8	0	-3	225	6.9	0.0069	20113	139
2-3							
4-5	1	-2	66	5.32	0.532	575	306
5-6	1	-3	166	6.90	0.0690	2299	159
6-7	1	-3	166	1.84	0.0184	5747	106
7-8	1	-3	166	1.27	0.0127	11483	146
8-9	1	-3	166	0.61	0.0061	20113	123
9-10	0	-3	166	5.38	0.00538	32182	173
3-4							
5-6	1	-2	66	2.59	0.259	575	149
6-7	1	-3	166	3.98	0.0398	2299	92
7-8	1	-3	166	1.98	0.0198	5747	114
8-9	1	-3	166	0.84	0.0084	11493	97
9-10	0	-3	166	6.46	0.00646	21113	130
10-11					N.R.	32182	
4-5							
6-7	1	-2	66	3.67	0.367	575	211
7-8	2	-3	225	1.23	0.123	2299	283
8-9	1	-3	225	4.37	0.0437	5747	251
9-10	1	-3	225	2.72	0.0272	11493	313
10-11	1	-3	225	1.61	0.0161	21113	340
11-12	0	-3	225	6.98	0.00698	32182	225

LEGEND: Range = Gain
MA = Dummy TX Current Switch
Vp = Balance Control to Null Meter
G.F. = Geometric Factor
Pa = Apparent Resistivity

TABLE 7. LINE C. (CONT.)

COLORADO GEOLOGICAL SURVEY
Geophysical Exploration
(Resistivity Survey)

LOCATION	PROJECT	DATE
Idaho Springs, Colo.	Line A	13 June 1980
CHIEF OPERATOR	ASSISTANTS	METHOD
Jay Jones	Fargo and Treska	Dipole-Dipole (Nx100')

Sta.	Range	MA	Voltage	V_p	DV/I	G.F.	P_a
5-6							
7-8	1	-2	66	5.38	0.538	575	310
8-9	1	-2	66	1.32	0.132	2299	303
9-10	2	-3	250	0.75	0.075	5747	431
10-11	1	-3	250	4.59	0.0459	11493	528
11-12	1	-3	250	2.08	0.0208	21113	439
12-13	1	-3	250	1.50	0.0150	32183	483
6-7							
8-9	1	-2	66	4.40	0.440	575	253
9-10	1	-2	66	1.52	0.152	2299	349
10-11	2	-3	200	0.83	0.083	5747	477
11-12	1	-3	200	2.90	0.0290	11493	333
12-13	1	-3	200	2.00	0.020	21113	422
13-14	1	-3	200	0.75	0.0075	32183	241
7-8							
9-10	2	-2	66	0.79	0.79	575	454
10-11	1	-2	66	2.63	0.263	2299	605
11-12	1	-2	66	0.83	0.083	5747	477
12-13	1	-3	166	5.26	0.0526	11493	605
13-14	1	-3	166	1.85	0.0185	21113	391
8-9							
10-11	2	-2	66	1.10	1.10	575	633
11-12	1	-2	66	1.88	0.188	2299	432
12-13	0	-2	66	9.72	0.0972	5747	559
13-14	0	-2	66	3.14	0.0314	11493	361
9-10							
11-12	2	-2	66	0.70	0.70	575	403
12-13	1	-2	66	2.55	0.255	2299	586
13-14	0	-2	66	7.37	0.0737	5747	424
10-11							
12-13	2	-2	66	0.84	0.840	575	483
13-14	1	-2	66	1.84	0.184	2299	423
11-12							
13-14	1	-2	66	5.54	0.554	575	319

APPENDIX F

TABLE 8
GEOMETRIC FACTOR TABLE
SCHLUMBERGER METHOD

2 (ft) L(ft)	25	50	75	100	200	300
50	95.78	47.89	31.93	23.94	11.97	7.98
75	215.5	107.75	71.83	53.87	26.94	17.96
100	383.11	191.55	127.70	95.78	47.89	31.93
200	1532.44	766.22	510.81	383.11	191.56	127.70
300	3447.99	1724	1149.33	862	431	287.33
400	6129.87	3064.89	2043.26	1532.44	766.22	510.81
500	9577.77	4788.89	3192.59	2394.44	1197.22	798.15
600	1391.99	6896	4597.33	3447.99	1724	1149.33
700	18772.43	9386.22	6257.48	4693.11	2346.55	1564.37
800	24519.1	12259.54	8173.03	6129.77	3064.89	2043.26
900	31031.99	15515.99	10344	7758	3879	2586
1000	38311.1	19155.55	12770.36	9577.77	4788.89	3192.59
1100	46356.42	23178.21	15452.14	11589.11	5794.55	3863.04
1200	55167.97	27583.99	18389.32	13791.99	6896	4597.33
1300	64745.74	32372.87	21581.91	16186.44	8093.22	5395.48
1400	75083.74	37544.87	25029.91	18772.44	9386.22	6257.48
1500	86199.96	43099.98	28733.32	21548.98	10774.99	7183.3

TABLE 9. DIPOLE-DIPOLE GEOMETRIC FACTOR TABLE

n a(ft)	25	50	100	150	200	300
1	143.67	287.33	574.67	862	1149.33	1724
2	574.67	1149.32	2298.67	3448	4597.32	6896
3	1436.7	2873.3	5746.7	8620	11493.3	17240
4	2873.4	5746.6	11493.4	17240	22986.6	3480
5	5028.45	1056.55	20113.45	30170	40226.55	60340
6	8045.52	16090.48	32181.52	48272	64362.48	96544
7	11924.61	23848.39	47697.61	71546	95394.39	143092
8	17240.4	34479.6	68960.4	103440	137913.6	206880
9	23705.55	47409.45	94820.55	14230	189639.45	284460
10	31607.4	63212.6	126429.4	189640	252852.6	379280

TABLE 10. WENNER GEOMETRIC FACTOR TABLE

2II a(ft)	25	50	100	200	300	400	500
6.2	157	314.16	628.32	1256.64	1884.64	2513.27	3141.6

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