

GEOHERMAL RESERVOIR ASSESSMENT CASE
STUDY--NORTHERN DIXIE VALLEY, NEVADA

Final Report

November 1978--November 1980

by

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ABSTRACT

This report describes work performed by Southland Royalty as contractor and Mackay Minerals Research Institute, Haskins-Pfeiffer Inc., Thermal Power Company, Energy and Natural Resource Consultants Inc., and Christiansen Drilling Inc. as subcontractors for work on geothermal reservoir assessment in northern Dixie Valley, Nevada. (Figure 1, Page 2)

Existing data comprised of multilevel aeromagnetic surveys, a magnetotelluric survey, thermal gradient holes, a report by Geothermex entitled "Geothermal Potential in Dixie Valley, Nevada" which included seismicity, gravity, magnetic, and gradient hole information, and a report by Keplinger and Associates Inc. entitled "Preliminary Evaluation of Dixie Valley Geothermal Potential and Associated Economics" were delivered at the beginning of the contract term. During the contract two additional 1500 foot temperature gradient holes and two deep exploratory wells were drilled and tested. Hydrologic-hydrochemical, shallow temperature survey, structural-tectonic, petrologic alteration, and solid-sample geochemistry studies were completed by the Mackay Minerals Research Institute of the University of Nevada at Reno. In addition, eighteen miles of high resolution reflection seismic data were gathered over the area.

The study indicates that a geothermal regime with temperatures greater than 400°F may exist at a depth of approximately 7500' to 10,000' over an area more than ten miles in length. Further work will be required to evaluate the reservoir.

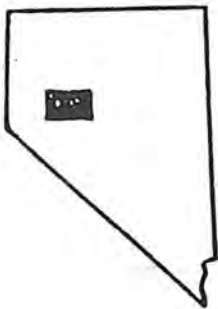
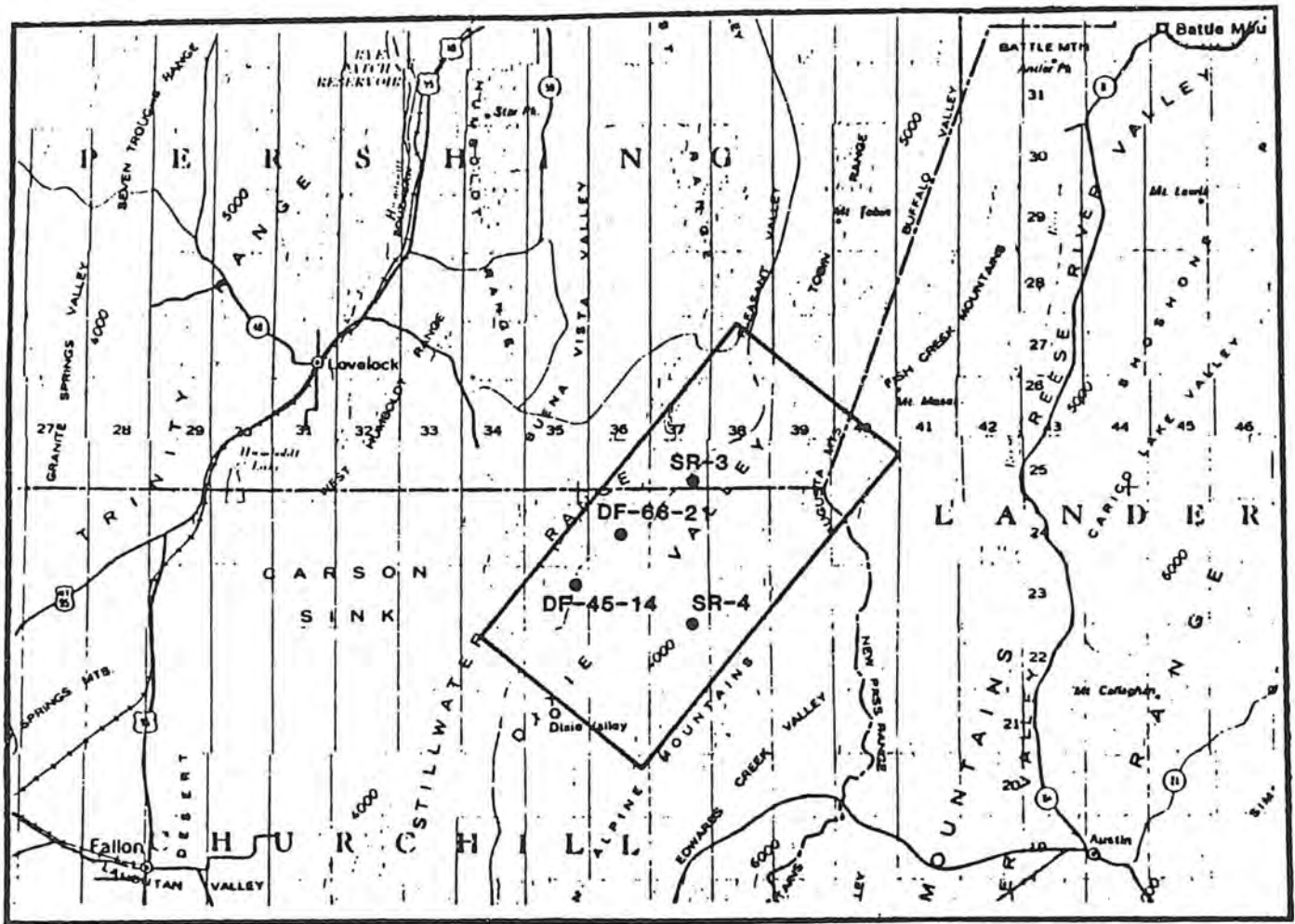


FIGURE 1. Case Study Area Map

SCOPE OF WORK

In May of 1978 Southland Royalty Company and the Mackay Minerals Research Institute of the University of Nevada at Reno submitted a proposal in response to RFP number ET-78-R-08-0003 which called for a joint industry/academic evaluation of the geothermal potential in the northern part of Dixie Valley, Nevada. The area to be covered extended from Dixie Hot Springs on the south to Sou Hot Springs on the north. The initial purpose of the study was to evaluate existing information regarding this area, including existing data delivered under this contract. It was then expected that a structural-tectonic review of the area would be commenced, additional intermediate depth temperature gradient holes would be drilled, existing wells and springs would be sampled, and a grid of one meter temperature holes would be installed. Coincident with submission of the proposal, eleven promising locations for deep well tests were permitted. It was assumed that preliminary evaluation would allow selection of the optimum two well sites by the time permits to drill were received. The proposal further anticipated that once the two deep wells were drilled petrologic alteration studies of the drill cuttings could be conducted, evaluation of fluids collected could be integrated with other hydrologic-hydrochemical information which had been collected, and structural and tectonic information could be updated if the wells intersected faults or thrust planes.

The contract was finally executed in November of 1978, at which time the review of existing data commenced and field work limited by adverse

weather conditions was commenced. Due to the nature of the academic calendar and the timing of the wells certain phases of the studies were not conducted in the sequences originally contemplated. In September of 1979 the scope of work was modified to add a high resolution reflection seismic survey of approximately eighteen miles designed to intersect the location of the Dixie Federal 66-21 wellbore and to provide for a geochemical analysis of the soil samples collected in the vicinity of the two deep wells and from 100' intervals in each of the wells.

ACKNOWLEDGEMENTS

The authors wish to thank the U. S. Department of Energy for its support of geothermal energy research in this country. Particular thanks go to Mr. Jack Salisbury who was instrumental in conceiving the program and to Mr. Joe Fiore for his cooperation and assistance in the administration of the program.

Thanks are due to Richard L. Jodry of Energy and Natural Resource Consultants, Inc. who supplied the original geologic concepts which convinced others that a high-temperature geothermal resource was possible in Dixie Valley despite the moderate geothermometers derived from the hot springs.

We wish to acknowledge once again the significant contribution of the Mackay Minerals Research Institute, the Mackay School of Mines, and the Desert Research Institute and the participating personnel and students of those institutions.

We wish to acknowledge the participation of Thermal Power Company in conducting the drilling operations on the two deep exploratory wells.

We wish to thank Mr. Howard P. Ross and the staff of the Earth Science Laboratory of the University of Utah Research Institute for their ideas and encouragement.

Thanks are due to Michael D. Campbell for his early work in Dixie Valley and his faith in a joint academic/industry approach to the project.

A special contribution was made by Mr. Bill Haskins and Mr. Rett Farrell of Haskins-Pfeiffer Inc. in overseeing the seismic program, processing and interpretation.

We wish to acknowledge the special talents of Mr. Jerry Christiansen in drilling the intermediate depth temperature gradient holes under adverse conditions. We thank Mary Jo Sweeney for her analysis of the drill cuttings from these gradient holes.

Finally, we wish to acknowledge the hospitality and cooperation of the people that live in Dixie Valley, Nevada.

RESULTS PREVIOUSLY DELIVERED

Data delivered under this contract is available from the University of Utah Research Institute, Earth Science Laboratory, 420 Chipeta Way, Suite 120, Salt Lake City, Utah 84108. Existing data delivered at the beginning of the contract is available under designations NV/DV/SR-1 through NV/DV/SR-10. The well history of Dixie Federal 45-14 (TD 9022') is available

under the designation NV/DV/SR-11. The well history for Dixie Federal 66-21 (TD 9780') is available under the designation NV/DV/SR-12. A graphic drilling history and a temperature profile for each well are presented in Figures 2 through 5. The Mackay Minerals Research Institute report on the first phase of the contract entitled "Geothermal Reservoir Assessment Case Study, Northern Basin and Range Province, Northern Dixie Valley, Nevada--Final Report" is available in two volumes designated NV/DV/SR-13. Because this report synthesizes most of the information obtained in the study, the narrative section (Volume 1) is included here as Appendix F-1. The final report of the Mackay Minerals Research Institute has the same title and is labeled Volume 3--Soil Geochemistry and Petrochemistry. It carries the designation NV/DV/SR-14. The text of this report is included as Appendix F-2.

NEW RESULTS

The interpretation of the seismic survey is included as Appendix F-3.

In addition, a final report on the isotope analysis which was performed by the Desert Research Institute on fluid samples from Dixie Valley is included as Appendix F-4. The isotope data were not available for earlier inclusion in the Mackay Minerals Research Institute Report. Results from two 1500 foot temperature gradient holes (SR-3 and SR-4) which were drilled in the latter part of the contract are included here as Appendix F-5.

THERMAL POWER COMPANY / SOUTHLAND ROYALTY COMPANY DIXIE VALLEY 45-14 CHURCHILL COUNTY, NEVADA

Spud date 4/25 5/4 5/14 5/24 6/3 6/13 6/23 7/3 7/13 7/23
 0 10 20 30 40 50 60 70 80 90 100 DAYS

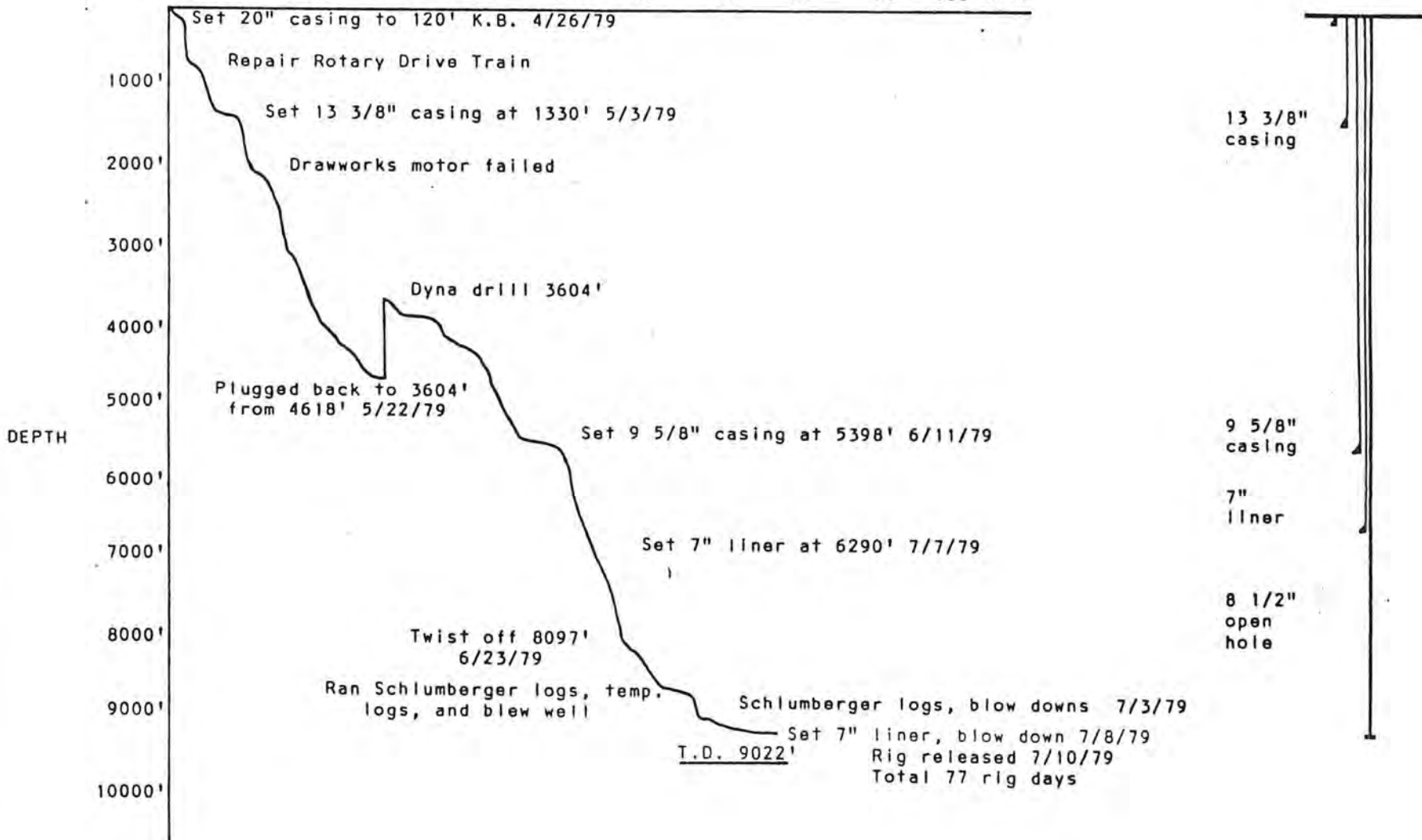


FIGURE 2. Drilling History & Casing Program for Dixie Federal 45-14

THERMAL POWER COMPANY SOUTHLAND ROYALTY COMPANY
 DIXIE FEDERAL 45-14 GEOTHERMAL WELL

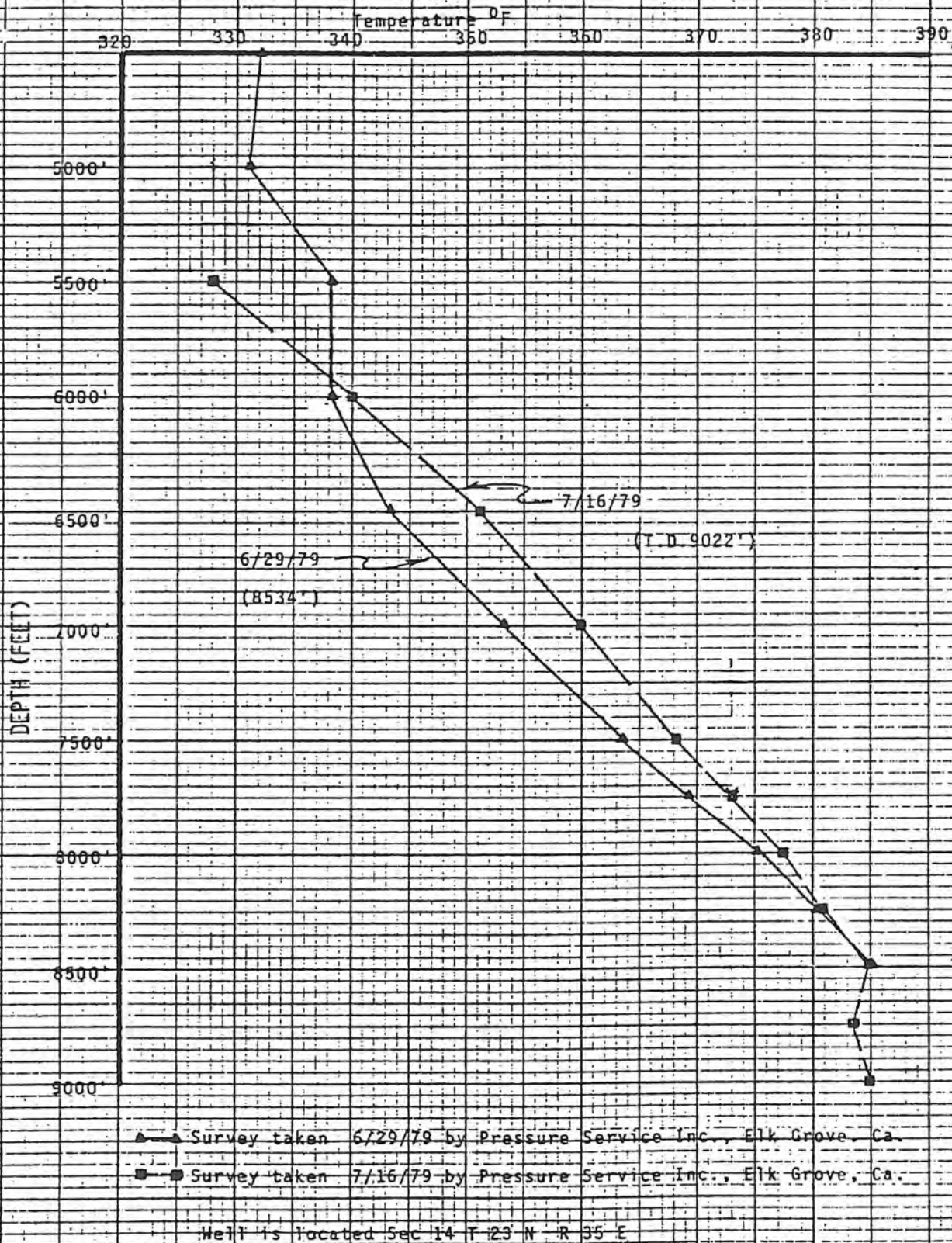


FIGURE 3. Downhole Temperature Profile for Dixie Federal 45-14

THERMAL POWER COMPANY / SOUTHLAND ROYALTY COMPANY DIXIE VALLEY 66-21 CHURCHILL COUNTY, NEVADA

7/15	7/25	8/4	8/14	8/24	9/3	9/13	9/23	10/3	1/4	1/14	1/24
0	10	20	30	40	50	60	70	80	0	10	20
DAYS											

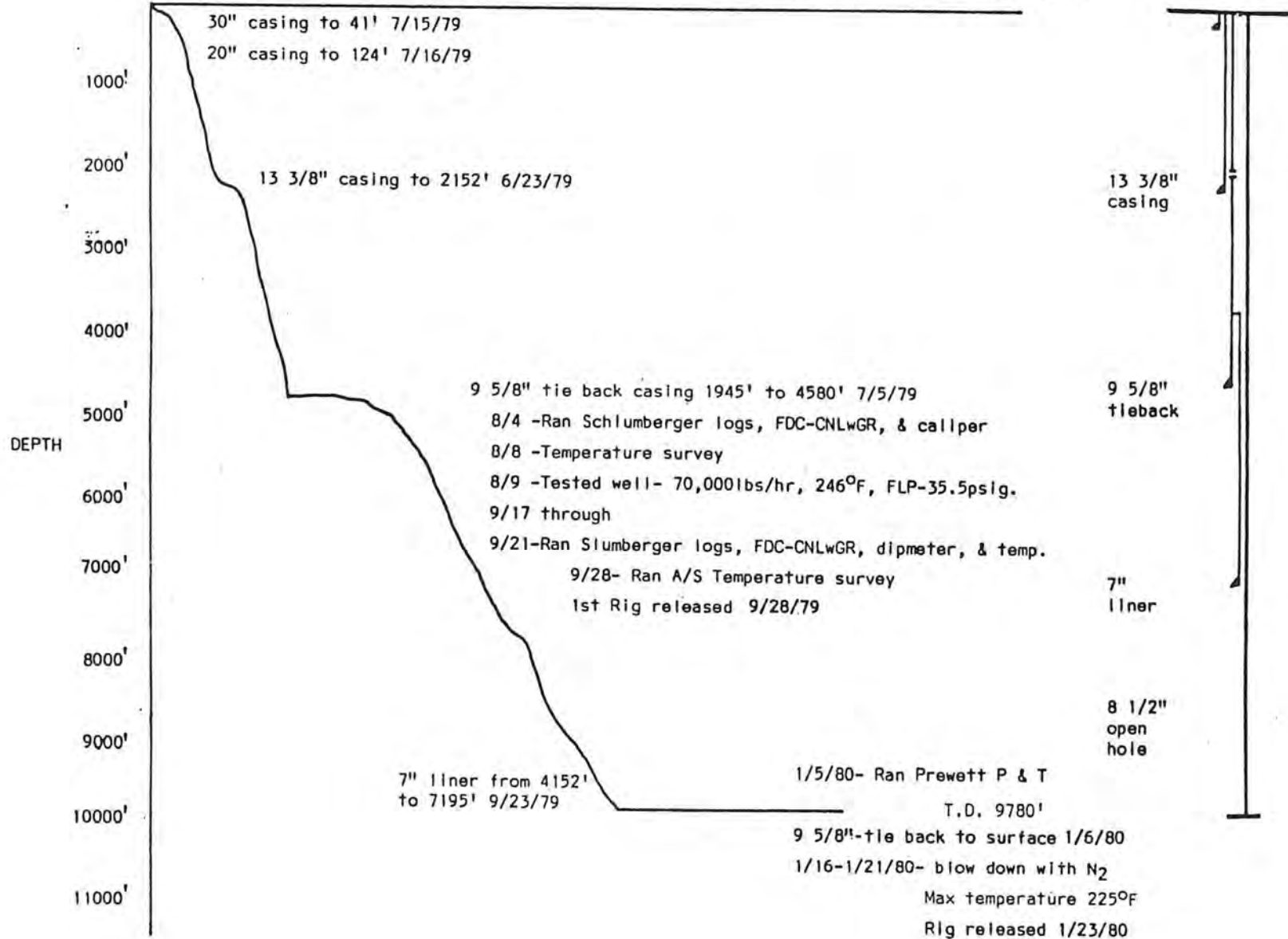


FIGURE 4. Drilling History & Casing Program for Dixie Federal 66-21

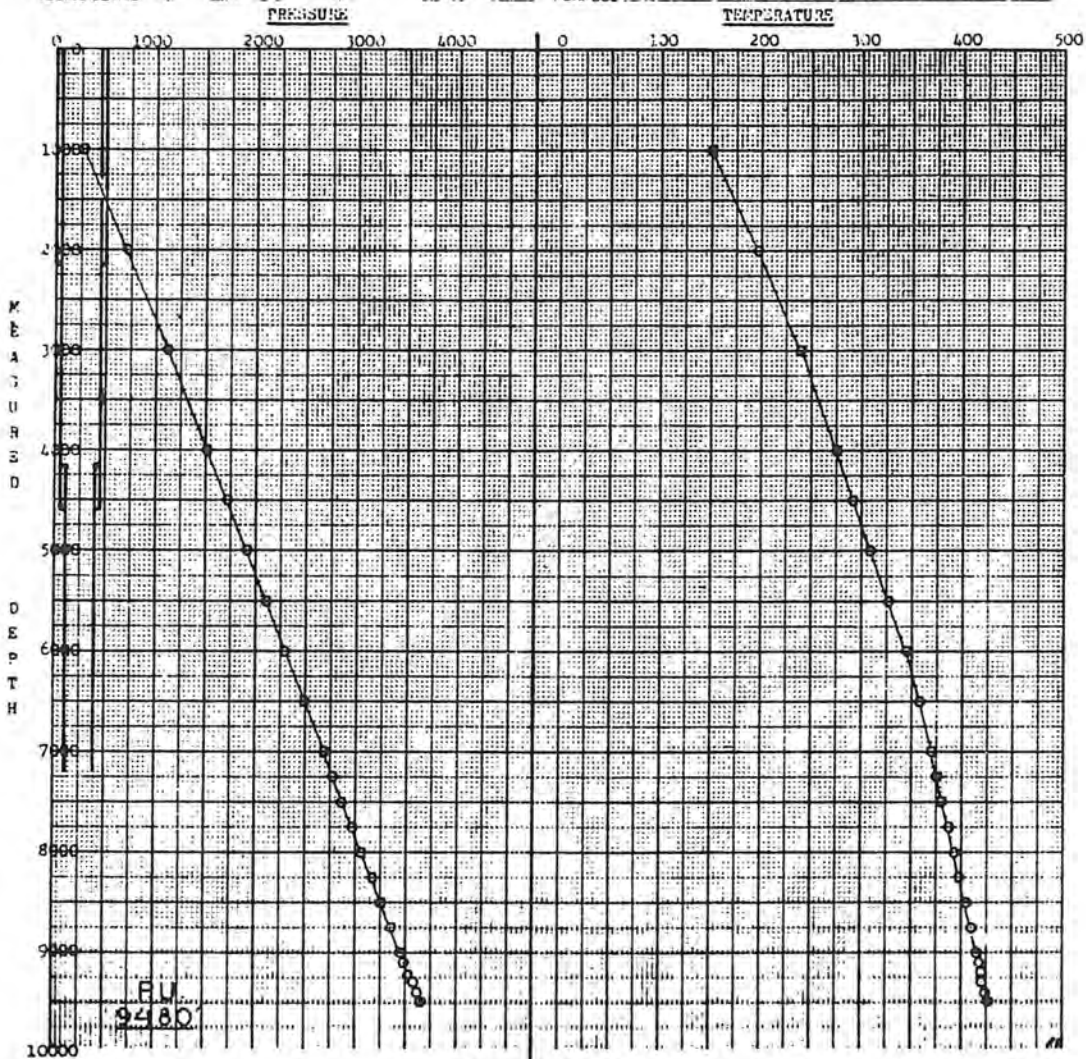
PRUETT WIRELINE SERVICE

8915 ROSEDALE HWY BAKERSFIELD, CA 93308
(805) 589-2788



SUB-SURFACE PRESSURE SURVEY

COMPANY Thermal Power FIELD Dixie Valley WELL NAME Dixie Federal 66-21
 TOTAL DEPTH 7750' WELL STATUS Static BOMB HUNG @ none DATE January 24, 1980
 CASING comments CASING PRESSURE --- TIME ON BOTTOM 4:27 am ELEMENT RANGE 6000 PSI / 36-499°
 LINER --- TUBING PRESSURE --- TIME OFF BOTTOM 4:37 am ZERO POINT ---
 PERFORATIONS --- ZONE none SHUT IN none ELEVATION none
 MPP --- MAX F 126.1° ON PRODUCTION none PICKUP 9480'
 TUBING DETAIL none PURPOSE Pressure/Gradient - Temperature Survey



SURVEY DATA

DEPTH	PRESS.	GRAD.	DEPTH	PRESS.	GRAD.	DEPTH	TEMPERATURE	DEPTH	TEMPERATURE
1000	275		7750	2970	.396	1300	151.2	7750	387.3
2000	706	.431	8000	3063	.372	2000	198.1	8000	393.0
3000	1118	.412	8250	3163	.400	3000	219.4	8250	397.6
4000	1514	.396	8500	3259	.384	4000	275.7	8500	404.8
4500	1716	.404	8750	3355	.384	4500	292.3	8750	410.1
5000	1912	.392	9000	3452	.388	5000	309.5	9000	415.2
5500	2108	.392	9100	3491	.390	5500	327.8	9100	418.2
6000	2299	.382	9200	3527	.360	6000	345.1	9200	419.8
6500	2491	.384	9300	3572	.450	6500	358.3	9300	421.4
7000	2686	.390	9400	3617	.450	7000	370.5	9400	423.7
7250	2772	.344	9480	3651	.425	7250	376.5	9480	426.7
7500	2871	.397				7500	380.3		

Casing: 13 3/8" @ 2152', 9 5/8" @ 0 to 4580', 7" @ 4152' to 7195'.

6000 PSI Instrument Serial No. 22883.

Instrument Serial No. 113, 3 Mosen
36-499.

FIGURE 5. Downhole Temperature Profile for Dixie Federal 66-21

EVALUATION OF RESULTS

Prior to the study it appeared that Dixie Valley would provide a good geothermal environment because it had numerous surface manifestations of heat, historic activity along major faults, and was the lowest valley in the northern part of Nevada with water movement into it from a very large contiguous hydrologic basin. Thus, the requisite components for a geothermal system - heat, active or open fractures and fluids - appeared to be available. In addition a postulated gabbroic complex appeared to provide a potential reservoir seal. The chief questions were whether commercial temperatures could be reached at an economic depth; whether a reservoir was present; and, if so, its exact nature.

With the results of this study it is possible to say that the depth to commercial temperatures may range from 7500 to 10,000 feet below the surface of the valley along its margins. It is postulated that this depth will increase as one approaches the center of the valley. The linear distance over which commercial temperatures can be expected to be found is probably greater than ten miles. The width of the temperature regime is uncertain, but there is at least some evidence (temperature gradient hole SR-4) to indicate that the heat regime lies on the eastern side of the valley as well as along the western margin.

The study also confirms the idea that active faulting in the area should result in open fracture patterns in many parts of the valley. However, in some areas, such as Dixie Federal 45-14, fractures may be so infrequent or so sealed that the potential for hot dry rock geothermal systems may be greater than for hydrothermal systems.

Fluids in the valley appear to be abundant although there is some evidence to indicate more than one hydrothermal regime exists in the valley. Structural or sedimentary controls may have an important bearing on which of these fluid regimes may have the greatest potential for economic use.

The part of the geothermal regime which is least well understood is the nature of the reservoir.

The nature of Dixie Valley geothermal resource is probably unique within the Basin and Range Province. The structural complexity of the area is related to apparent rifting of the valley and the presence of the Humboldt Lopolith. It is therefore unlikely that a continuous homogeneous reservoir will exist across the entire northern part of Dixie Valley. The area encompassed by mapped surface outcrops of the Humboldt Lopolith and the adjacent part of the valley containing extensive gabbroic sequences may be the area in which optimum conditions for geothermal energy production are most likely to be found. Areas in which faults are acting as thermal conduits may have potential for economic geothermal energy production in areas where no specific reservoir rocks are found. Long term development of the geothermal resource in Dixie Valley will require additional drilling to define areas of this reservoir that are characterized by both depth and temperature regime suitable for commercial application for electric power production. Although geothermal resources at lower temperatures than those presently considered economic for such power production exist in Dixie Valley, its remote location makes economic development of those resources unlikely in the near future.

APPENDIX F-1

GEOHERMAL RESERVOIR ASSESSMENT CASE STUDY, NORTHERN
BASIN AND RANGE PROVINCE, NORTHERN DIXIE VALLEY, NEVADA

Conducted for:

U. S. Department of Energy
Contract number DE-AC08-79ET27006

Work Performed under Subcontract to:

Southland Royalty Company
Fort Worth, Texas

Conducted by:

Mackay Minerals Research Institute
University of Nevada, Reno, Nevada

Submitted:

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Color composite Landsat scene of north-central Nevada on December 20, 1979. Dixie Valley appears as an island amid the snow-covered landscape.

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Chapter 8. EVALUATION OF THE INTEGRATED MODEL OF THE DIXIE VALLEY GEOTHERMAL SYSTEM

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*MyIars of plates are on file with the Nevada Bureau of Mines and Geology, University of Nevada, Reno, Nevada 89557. Copies are available for the cost of reproduction.

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Chapter 1. INTRODUCTION

By: Elaine J. Bell, Michael E. Campana, Roger L. Jacobson,
Lawrence T. Larson, and D. Burton Slemmons

1.0 INTRODUCTION

1.1 Foreward

This report was prepared for the U. S. Department of Energy (DOE) in compliance with conditions of the statement of work as part of Contract number DE-AC08-79ET27006 for Geothermal Reservoir Assessment in the northern Basin and Range Province. Work was performed by the Mackay Minerals Research Institute (MMRI) under subcontract to Southland Royalty Company (SRC), Fort Worth, Texas.

The MMRI, with the Mackay School of Mines as lead agency, is charged with performing research in the general field of non-renewable resources. Optimal use of the staff and facilities of the University results from the various components of the University cooperating in interdisciplinary research. The Geothermal Reservoir Assessment of Dixie Valley by the MMRI involved both the Mackay School of Mines (MSM) and the Desert Research Institute (DRI). MSM and DRI personnel performed the investigations and prepared the following final technical report on the Dixie Valley Geothermal System. Table 1-1 lists specific individuals involved in the project, their respective affiliation, title, investigation areas, and level of effort. Primary responsibility for accomplishing the tasks listed in the statement of work resided with the Graduate Research Fellows under the technical supervision of the Principal Investigators. The specific personnel and their individual responsibilities in completing the various investigations are indicated in Chapters 3 through 6 of this report.

The overall reservoir assessment dealt with the northern Dixie Valley area, Nevada (Figure 1-1; Plate I). Specific investigations conducted within the study area included: 1) Structural-Tectonic Analysis; 2) Petrologic Alteration Studies; 3) Hydrology and Hydrogeochemistry; and 4) Shallow Temperature Survey. The results of these investigations are presented in the following technical report in a written format supplemented by appropriate graphic data and appended information.

1.2 Purpose

The purpose of the MMRI program was to develop an integrated model of the Dixie Valley area with respect to a geothermal system based upon both pre-existing data and data derived from the scope of

Table 1-1. MMRI Personnel

<u>Name</u>	<u>Affiliation</u> *	<u>Title (Areas of Investigation)</u> **	<u>Level of Effort</u>
Bard, Thomas R.	MSM	GRF (PA; STS)	8 man-months
Bell, Elaine J.	MSM	PI/Administrator (S-T; PA)	6 1/2 man-months
Bohm, Burkhard W.	DRI	GRF (HG)	7 man-months
Campana, Michael E.	DRI	PI (H; HG; STS)	1/2 man-month
Dowden, John E.	DRI	GRF (H; HG; STS)	1/4 man-month
Henne, Mark S.	DRI	GRF (H; HG; STS)	1/4 man-month
Ingraham, Neil L.	DRI	GRF (H; STS)	2 1/2 man-months
Jacobson, Roger L.	DRI	PI (HG; H; STS)	1/2 man-month
Johnson, Cady	DRI	GRF (STS)	2 1/2 man-months
Juncal, Russell W.	MSM	GRF (PA; STS)	6 man-months
Kearl, Peter M.	DRI	GRF (H; HG)	1/2 man-month
Larson, Lawrence T.	MSM	Project Administrator	1 man-month
McKay, Alan	DRI	Technician (H; HG)	1 man-month
Nosker, Sue C.	MSM	GRF (PA)	1 1/2 man-months
Slemmons, D. Burton	MSM	PI (PA; S-T)	1 man-month
Wheatcraft, Steven	DRI	Geologist (H; HG)	1/2 man-month
Whitney, Robert A.	MSM	GRF (S-T; STS)	8 man-months

* MSM -- Mackay School of Mines; DRI -- Desert Research Institute.

** GRF -- Graduate Research Fellow; PI -- Principal Investigator; PA - Petrologic Alteration Studies; STS -- Shallow Temperature Survey; S-T -- Structural-Tectonic Analysis; HG -- Hydrogeochemistry; H -- Hydrology.

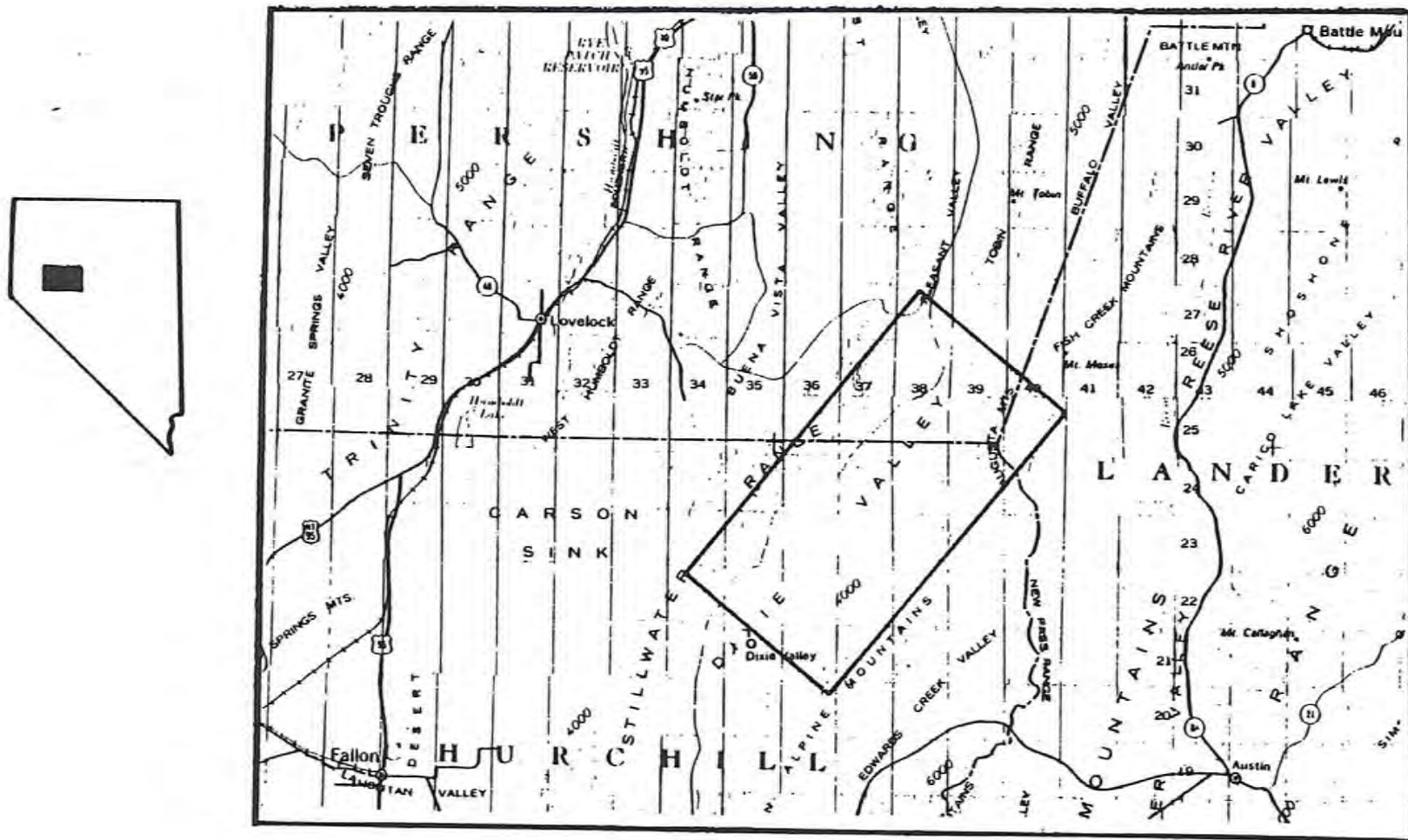


Figure 1-1. Index map of Dixie Valley study region. Shaded area of State shown enlarged. Specific study area outlined on enlarged map is shown in detail on Plate I.

work summarized below. This model will then serve the DOE as a case study of a Basin and Range geothermal system.

1.3 Scope

The scope of the study as outlined in Appendix A, Scope of Work, of both the Prime Contract and Subcontract (Contract number DE-AC08-79ET27006) included the following major tasks for each phase of investigation:

Structural-Tectonic Analysis

- (1) Review and evaluate available fault and lineament maps and high-altitude photography.
- (2) Conduct low-sun-angle and "snow lapse" photography studies.
- (3) Conduct field structural mapping.
- (4) Correlate and interpret data and develop a structural model.

Petrologic Alteration Studies

- (1) Analyze subsurface samples from thermal gradient hole drilling and representative surface samples.
- (2) Describe lithology and petrologic variation in drill holes.
- (3) Conduct detailed mapping of the primary stratigraphic units of the valley fill.
- (4) Review existing lithology and mineralogy data.
- (5) Integrate the above petrologic studies with the structural-tectonic studies.

Hydrology-Hydrogeochemistry

- (1) Review available hydrologic-hydrogeochemical data.
- (2) Sample selected wells, springs, and fumaroles.
- (3) Perform selected trace element and isotopic analyses.
- (4) Collect temperature data from existing wells and springs.
- (5) Determine recharge and groundwater flow rates and estimate reservoir geometry.
- (6) Construct an aquifer flow model.

Shallow Temperature Survey

- (1) Install and monitor a shallow depth temperature survey consisting of about 200 one-meter deep holes for a 6-month period.
- (2) Statistically evaluate the cumulative and individual readings of the survey.
- (3) Compare the near-surface hydrothermal discharge system with subsurface temperature gradient hole data.

- (4) Evaluate the validity of the shallow temperature survey as a geothermal exploration technique.

Model the Dixie Valley Geothermal System

- (1) Develop an integrated model of the Dixie Valley region with respect to a geothermal system based on pre-existing data and data derived from the above listed tasks.
- (2) Evaluate the Dixie Valley Geothermal System model.

Note:

During the course of investigation, Task 3 of the Petrologic Alteration Studies was transferred to the tasks to be accomplished under the Structural-Tectonic Analysis. This modification of the outlined program was based on 1) an increasing awareness of the integral relationship of the valley-fill stratigraphic units and their geographic distribution to the structural-tectonic history of the valley and 2) the extreme lateral and vertical variations (facies changes) inferred for the subsurface based upon limited drill hole information. More extensive subsurface stratigraphic data are needed upon which to base any subsurface-surface petrologic correlations.

1.4 Study Approach and Methods

Because the MMRI program involved several phases of study, the specific approach and methods of each investigation are presented in the respective chapters of this report. Models were developed based on the specific data of each of the various phases of investigation. These 'single-phase' models were then integrated to derive the model of the Dixie Valley Geothermal System.

1.5 Report Organization

This final technical report is presented in chapter format with each chapter authored by the person(s) primarily responsible for data development and accomplishing the tasks as outlined in the statement of work. The regional setting is presented in Chapter 2 to provide a basic framework within which to evaluate the data presented in the following chapters. Chapters 3 through 6 are devoted to each phase of the MMRI program of investigation. Chapters 2 through 6 serve as the basis for the integrated model of the Dixie Valley Geothermal Sys-

tem in Chapter 7. Chapter 8 evaluates the model in terms of completeness, conflicts in the existing data, and recommendations for further study that would verify and refine the model. Where necessary, graphic display of data in the form of tables, charts, photographs, figures, or maps is included to supplement and enhance the text presentation.

Preparation of the final technical report involved the following process:

- (1) Authorship of the respective chapters on the various phases of investigation.
- (2) Collation of Chapters 1 through 6 to form the Draft Final Report (DFR).
- (3) Review of DFR by all MMRI project personnel and by key SRC personnel.
- (4) Integration of data and 'single-phase' models to derive a model for the Dixie Valley Geothermal System and to evaluate the model. Cooperative effort by all MMRI personnel and key SRC personnel.
- (5) Preparation of DFR Chapters 7 and 8.
- (6) Review of DFR Chapters 7 and 8 by all MMRI personnel and key SRC personnel.
- (7) Preparation of Final Technical Report.

1.6 Acknowledgements

On behalf of the Graduate Research Fellows and ourselves, we express our sincere appreciation to the following representatives of Southland Royalty Company: Mr. Jere Denton, District Manager of Natural Resources, for initiating the joint venture between SRC and the MMRI, and for his continuing support, cooperation and guidance since the inception of this project; and Mr. Dennis S. McMurdie, Geothermal Geologist, for his probing questions and assistance during the course of our investigations. We also wish to acknowledge the stimulating discussions prompted by Mr. Richard L. Jodry, Energy Consultant to Southland Royalty Company.

We wish to acknowledge the participation of Mr. William D'Olier and Mr. Lou de Leon of Thermal Power Company in coordinating the deep exploratory drilling program in Dixie Valley with the MMRI investigations.

Dennis T. Trexler of the Nevada Bureau of Mines and Geology provided guidance and advice during the stages of proposal preparation

for this project and continued to be a source of administrative guidance throughout the course of the project; we especially thank him for his assistance. Purchasing agreements and bookkeeping were ably handled by Mrs. Betsy Peck, Ms. Sheryl Harding, Ms. Alta Sly, and Mrs. Louise Gibbs. Superb secretarial services provided by Mrs. Soni Cox, Mrs. Barbara Salmon, Mrs. Carol Hackney, Ms. Alta Sly, Mrs. Alma Smith and Ms. Mollie Stewart made the preparation of this report possible. Ms. Alice Kellames (MSM) is thanked not only for her patience but also for her capable assistance in handling personnel contracts and numerous unforeseen problems. Dale Schulke (DRI) was indispensable in dealing with the myriad of administrative, budget and other problems that arose during the course of the study. Contract expenditure records were maintained in a timely and efficient manner by Mr. James Murphy and Mr. Barry Myers of the UNR Controllers Office.

We wish to express our appreciation to the U. S. Department of Energy for their willingness to support the MMRI and its research in Dixie Valley. We would particularly like to thank Mr. Joe Fiore of the U. S. DOE for his cooperation and encouragement.

Special thanks are extended to Mr. and Mrs. Sheldon Lamb whose hospitality and cooperation were of inestimable value to the field effort of MMRI personnel in Dixie Valley.

We wish to extend our thanks to numerous other support personnel of the Mackay School of Mines and the Desert Research Institute who assisted our efforts toward completion of this project. Special thanks are due Ms. Patricia Harris and her staff at the DRI water analysis laboratories for providing excellent chemical analysis and related services, often under extremely difficult conditions. Jack Dowden, Cady Johnson, Robert Broadbent, W. Alan McKay, Rod Fricke, Mark Henne, David Bratberg, and Richard Goldfarb performed field work above and beyond the call of duty. Ron Sheen was invaluable in the portions of this project that involved computer analysis, and W. Alan McKay performed excellent drafting work.

And finally, we wish to thank Sun Oil Company for the grapes on the vine that aided us in understanding the complex nature of the Dixie Valley Geothermal System.

Chapter 2. REGIONAL SETTING

By: Elaine J. Bell

2.0 REGIONAL SETTING

2.1 Introduction

The geotectonic history of western Nevada, encompassing Dixie Valley, is unique and complex. Because the western Great Basin is a result of the dynamic interaction of the Pacific and North American Plates, the following discussion is focussed on the plate tectonic evolution of the California-Nevada region. While this interaction of lithospheric plates is not fully understood, the present state-of-the-art knowledge can be utilized for discerning a regional geotectonic framework for the Dixie Valley Geothermal System. This chapter is a summary; the reader is referred to Wilson (1971, 1976) for discussions of the basic concepts of plate tectonics, and to Atwater (1970) and Ernst (1979a) for more detailed evaluations of the Cenozoic plate interactions in western North America. Figure 2-1 illustrates selected key Cordilleran tectonic elements discussed in the following sections.

2.2 Geotectonic History

2.2.1 Precambrian

Three major divisions of the Precambrian are recognized in the Great Basin and adjoining regions: metasedimentary (schist and paragneiss) and plutonic rocks older than about 1400 million years (my); unmetamorphosed sedimentary rocks approximately comparable in age to the Belt Supergroup, 850 to 1250 my; and the unmetamorphosed uppermost Precambrian sedimentary rocks probably less than about 850 my old. The metasedimentary and plutonic rocks older than 1400 my are extensive in the surface and subsurface of Wyoming, Utah, Arizona and California east of the Wasatch Line. West of the Wasatch Line they are sparsely exposed in southeastern California, and in southern Idaho and northern Utah. Supposed older Precambrian rocks that crop out in southern Nevada (Ekren and others, 1971) may also be metamorphosed strata of latest Precambrian or Paleozoic age (Stewart and Poole, 1974). Unmetamorphosed to slightly metamorphosed pregeoclinal sedimentary rocks equivalent to the Belt Supergroup are recognized in northern Utah, southeastern California and in the Grand Canyon region of Arizona.

A distinctive diamictite unit (a rock composed of large clasts in a finer grained matrix) is considered to be of glacial origin (Critt-

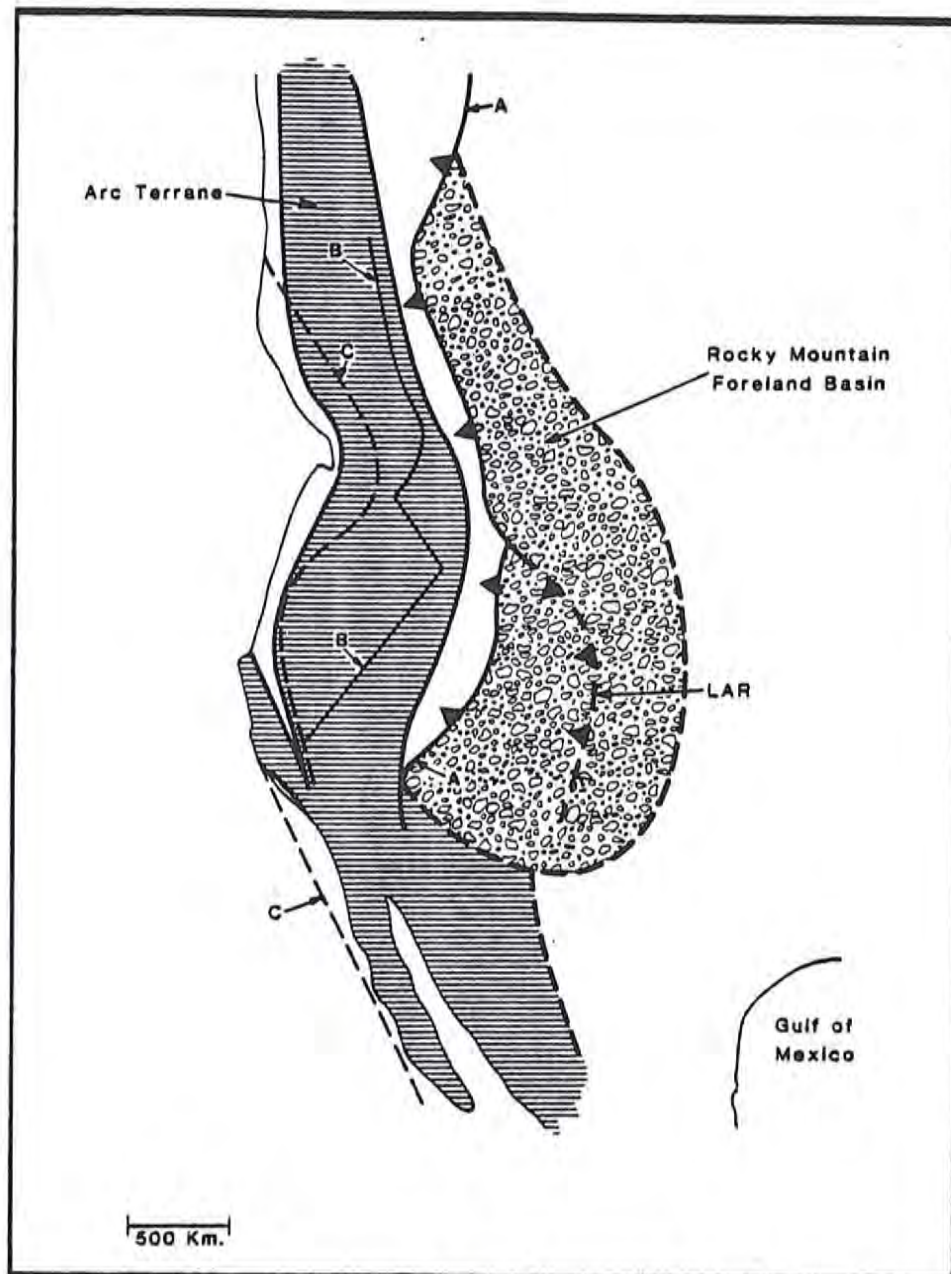


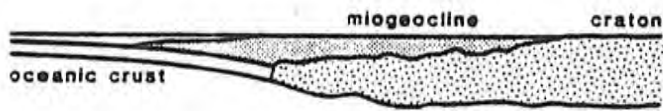
Figure 2-1. Diagrammatic sketch map of western North America showing selected key Cordilleran tectonic elements. Arc terrane is the main core belt of post-Triassic origin related to subduction of oceanic lithosphere beneath the continental margin. Line A denotes main backarc fold-thrust belt; Laramide front (LAR) is shown where it broke part of the foreland basin; line B is the edge of rifted Precambrian basement along the pre-Devonian continental margin; and Line C is edge of continental crust prior to initiation and evolution of the Cordilleran continental-margin arc-trench system. (from Dickinson, 1976)

enden and others, 1972; Stewart, 1972). A key component of this Windemere succession along the whole length of the Cordillera is a unit of tillites and glaciomarine turbidites. Part of a diamictite-volcanic sequence, this unit occurs at or near the base of the uppermost Precambrian rocks throughout westernmost North America from Alaska to California, with potassium-argon ages between 827 and 918 my on a greenstone member in northeastern Washington (Miller and others, 1973). This is the oldest unit that has a depositional pattern related to the Cordilleran geosyncline. A major change in the tectonic pattern of North America appears to have been initiated shortly before the deposition of the diamictite, and this change is inferred to mark the beginning of the geosyncline (Stewart, 1972).

The Cordilleran geosynclinal rocks described in this section range in age from latest Precambrian to late Devonian and consist of miogeoclinal carbonate and transitional assemblage rocks and of eugeoclinal siliceous assemblage rocks within the Great Basin. In general, these strata are well-sorted chemically mature sediments produced along an Atlantic-type continental margin (Figure 2-2a). At least half of the thickness of lower Paleozoic and uppermost Precambrian geosynclinal deposits are uppermost Precambrian and lower Cambrian quartzite and siltstone, and lesser carbonate rock and conglomerate. Because of the uncertainty in the Precambrian-Cambrian boundary (Stewart and Suczek, 1977), this terrigenous detrital sequence, which thickens westward, or northwestward from 300 m at about the Wasatch Line to locally as much as 7,500 m will be considered as a unit.

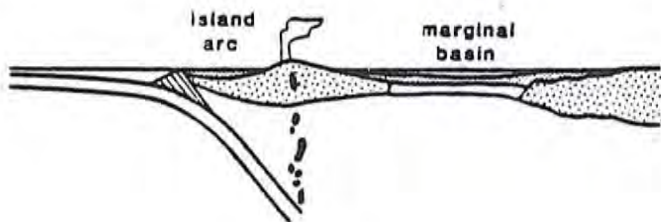
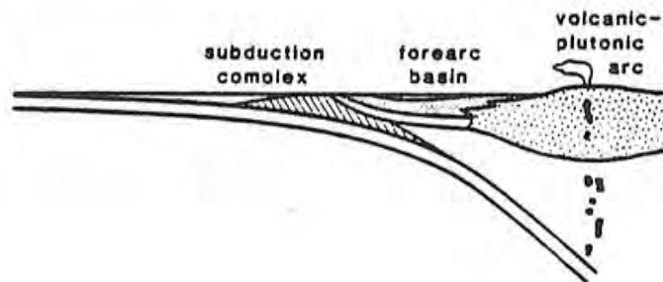
Three facies of the uppermost Precambrian and lower Cambrian rocks are generally recognized: (1) a quartzite and siltstone facies in the eastern Great Basin; (2) a siltstone, carbonate, and quartzite (transitional) facies in the central Great Basin; and (3) a siliceous and volcanic facies in the western Great Basin. The eastern and transitional facies comprise the terrigenous detrital sequence.

The eastern facies (Merriam and Anderson, 1942) or Millard belt (Kay, 1951) is miogeoclinal and consists mostly of thick cliff-forming fine- to medium-grained quartzite units ranging in composition from arkose to orthoquartzite (Stewart, 1970) separated by units of siltstone. Conglomerate, limestone and dolomite units are also present in thin layers (less than 30 m thick). Mafic volcanic flows and



A - Atlantic type

B - Andean type



C - Japanese type

D - Californian type



Figure 2-2. Diagrammatic sketch of the four major types of continental margins showing crustal topologies: (A) Atlantic type, divergent margin; (B) Andean type, convergent margin; (C) Japanese type, convergent margin; and (D) Californian type, transform margin. (from Ernst, 1979; after Dickinson, 1976)

breccias occur in this facies in the eastern Great Basin (Stewart, 1972, 1974). One mafic volcanic breccia in Utah is radiometrically dated as 570 my old (Crittenden and Wallace, 1973). The transitional siltstone, carbonate, and quartzite facies in central Nevada and southeastern California is thick and fossil-rich. It contains large amounts of siltstone (or phyllitic siltstone), thin to thick units of limestone and dolomite, and fine- to very fine-grained quartzite (Nelson, 1962; Stewart, 1970; Albers and Stewart, 1972). Fossils, including trilobites, archeocyathids, pelecypods, Hyalithes, Salterella, Skolithus, and algae indicate a Cambrian age from 650 (?) to 540 my (Palmer, 1971; Stewart, 1970; Nelson, 1976). The terrigenous detrital sequence is dominantly of shallow-water origin as indicated by an abundance of algae and archeocyathids and by the local occurrence of mudcracks, raindrop imprints, and runzel marks (Barnes and Klein, 1975; Klein, 1975) that indicate local exposure. Herringbone cross stratification, reactivation surfaces, superimposition of current ripples on larger current ripples, flaser and lenticular bedding, and other sedimentary structures as well as trace fossils indicate that sediments deposited by tidal currents occur in some units (Stewart, 1970; Barnes and Klein, 1975; Klein, 1975).

The siliceous and volcanic facies in the western Great Basin is identified in one locality near Battle Mountain and is represented by the Scott Canyon Formation (Roberts, 1964). It is allochthonous, known only in fault contact with both overlying and underlying units, and probably was transported eastward at least 80 km by the end of the Devonian-early Mississippian Antler Orogeny. It is composed of several thousand meters of chert, argillite and greenstones and minor amounts of sandstone, quartzite, and limestone (Roberts, 1964; Theodore and Roberts, 1971). The greenstone includes flows, which are in part massive and in part pillow lavas, and pyroclastic rocks ranging from fine tuffs to coarse breccias. Originally, these flows were probably mafic andesites or basalts, with subsequent alteration typical of volcanic rocks that have been extruded in a submarine environment (Roberts, 1964). The limestone occurs in lenses and contains algae, sponges, archeocyathids, and trilobites considered to be early or middle Cambrian in age. However, radiolaria in chert have been ten-

tatively assigned a Devonian age by Brian Holdsworth (Stewart and Suczek, 1977) suggesting that the Scott Canyon may consist of structurally interleaved rocks of at least two ages, a structural relation commonly mapped in other eugeosynclinal terranes in Nevada (Stanley and others, 1977). The exact age relations of the various parts of the Formation are not presently known. The Scott Canyon Formation may be a relatively deep-water oceanic deposit (Stewart and Poole, 1974). However, fossil material such as algae and archeocyathids is indicative of shallow-water origin. This anomaly may be explained as occurring in slump deposits or in structurally interleaved shallow-water layers in a predominantly deep-water deposit. Alternatively, the shallow-water deposits could have been on the flanks of oceanic volcanoes.

The type and distribution of Precambrian and lower Cambrian sediments are related to the tectonic setting of western North America during that time period. Two continental terrace wedge accumulations, the Belt-Purcell Supergroup and the Windemere Group and conformable Paleozoic strata (the Cordilleran miogeosyncline) may require two periods of continental rifting: one between 1400 to 1450 my ago and 1600 to 1700 my ago (Gabrielse, 1972) and the second approximately 850 my ago (Stewart, 1972, 1976). Whether initial Precambrian rifting was pre-Belt or pre-Windemere, by the beginning of Cambrian time the continental margin along western North America was reshaped and was the site of initial Cordilleran miogeoclinal deposition. Stewart and Suczek (1977) incorporate the concept that the Windemere deposits may have accumulated locally in rift-valley basins that formed prior to the main stage of rifting. The time of these events is poorly known, but rift valleys may have formed from 800 to 900 my ago and the main rifting event not until perhaps 650 my ago. In this setting, the terrigenous detrital sequence is related to erosion of the initial bulge or uplifted area created by thermal expansion during continental rifting (Figure 2-3). Rocks older than the rifting episode(s) would have been removed by rifting; the various detached fragments separated by suture belts may lie within the composite modern continent of Eurasia (including Alaska) (Dickinson, 1979).

The cessation of deposition of terrigenous detritus in the miogeocline by the end of early Cambrian time is related to the destruction of the bulge by thermal contraction due to cooling and by erosion.

800-900 (?) m.y.



650 (?) m.y.

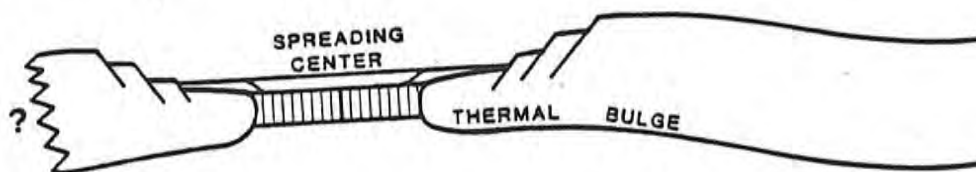


Figure 2-3. Diagrammatic sketch of late Precambrian and Cambrian development of western North America. Note the thermal bulge associated with the rift. (from Stewart and Suczek, 1977)

The ocean then transgressed eastward into cratonic areas. As a consequence, coarse detrital material was trapped in coastal areas on the craton and the miogeoclinal shelf was an area of carbonate deposition relatively free of detrital material. Figure 2-4 depicts the distal margin of the Precambrian basement (Dickinson, 1979). The location of the rifted continental margin transecting the Great Basin suggests that Precambrian age deposits are not present in western Nevada, and in particular, in Dixie Valley. The Scott Canyon Formation, as discussed above, may represent post-rifting Cambrian and younger Paleozoic deposits rather than a Precambrian eugeosynclinal sequence.

2.2.2 Paleozoic

2.2.2.1 Cambrian - Devonian (Antler Orogeny)

During the Cambrian, seas continued to transgress the continent and by late Cambrian they extended 600 to 1000 km east of the margin of the miogeocline, and much of the western United States was a shallow sea. Based upon unit thicknesses and sedimentary facies distributions, the ancient Pacific continental margin of North America extended from southern-most Idaho southward into southeastern California (Figure 2-5) prior to the end of the late Devonian-early Mississippian Antler orogeny (Burchfiel and Davis, 1972; Stewart and Poole, 1974; Stewart and Suczek, 1977). The overlapping sediments thicken from about 1 km on the southeast to almost 10 km on the north and northwest.

Development of the miogeocline culminated in the formation of a major oceanic basin by sea-floor spreading by Ordovician time. In late Ordovician or Silurian, westward subduction of this oceanic basin formed a volcanic arc -- the Klamath-Sierra arc. The continental margin had a Japanese-type configuration (Figure 2-2c), with a complicated offshore pattern of marginal seas and intra-oceanic island arcs (Figure 2-5) (Burchfiel and Davis, 1972; Monger and others, 1972; Silberling, 1973; Churkin, 1974). As subduction continued and the arc migrated eastward toward the continent, a lower Paleozoic accretionary wedge was formed.

During the middle to late Devonian, arc volcanism stepped eastward onto the accretionary wedge. The Antler orogeny marked the culmination of arc-continent collision as the leading edge of the accre-

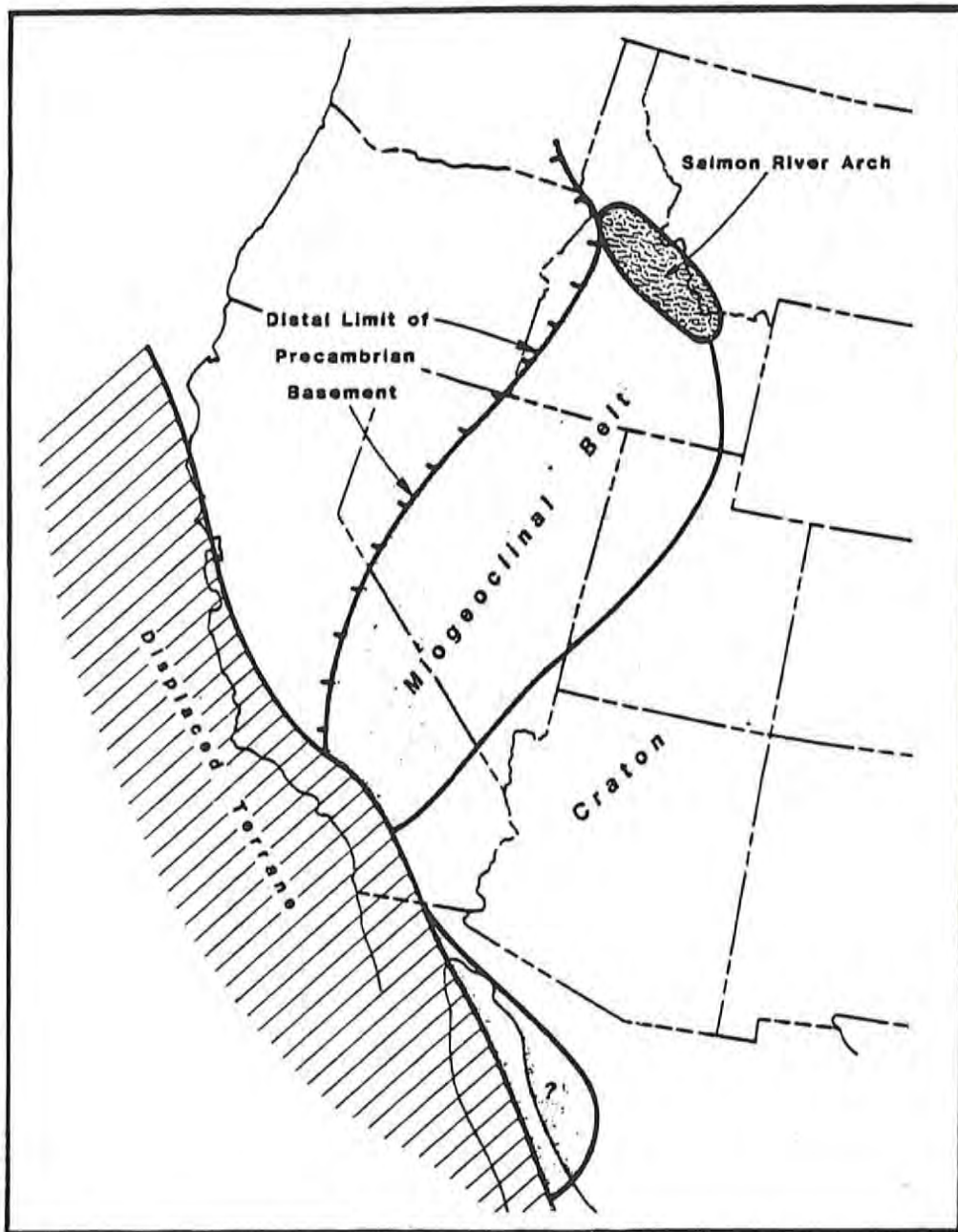


Figure 2-4. Diagrammatic sketch of early Paleozoic Cordilleran rifted continental margin. Note the location of the distal limit of Precambrian basement. (from Dickinson, 1979)

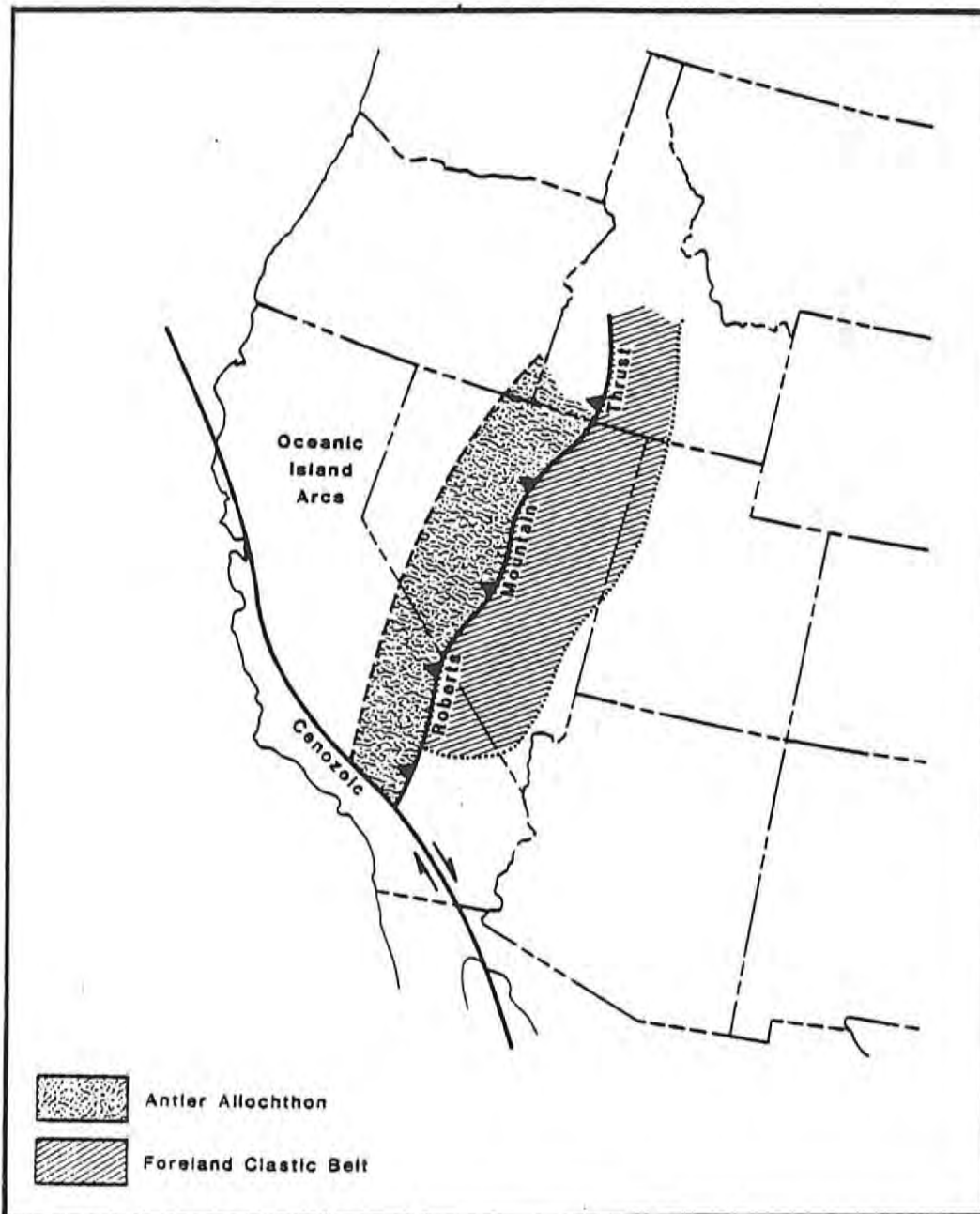


Figure 2-5. Generalized plate tectonic setting of California and Nevada during the Antler Orogeny. (from Ernst, 1979)

tionary wedge was thrust over the continent along the Roberts Mountains Thrust (the Harmony and Valmy-Vinini Formations in Nevada), bringing volcanogenic argillites of a subduction complex over an autochthonous miogeoclinal sequence (Stewart and Poole, 1974).

The upper Cambrian Harmony Formation originated as a sequence of turbidity flows in an extensive subsea fan complex on ocean floor west of the Cordilleran miogeocline (Suczek, 1977a,b). The sediments are believed to have been derived from the Salmon River arch, 400 km north (Stewart and Suczek, 1977). Underlain by upper Cambrian Paradise Valley Chert, both units were deposited as a fan complex on an abyssal plain floored by pelagic siliceous sediments.

The Valmy (Roberts, 1964; Gilluly, 1967; Gilluly and Gates, 1965) consists of conspicuous ledge-forming, poorly sorted quartzite and chert, with significant amounts of shale and greenstone. Numerous thrust fault slicing may account for the lack of lithologic continuity (Gilluly, 1967; Gilluly and Gates, 1965). Both the Valmy and the Vinini, a similar less quartzitic, more shaley unit to the east, include Ordovician and appreciable Devonian strata (Stanley and others, 1977; Jones and others, 1978). Schweikert and Snyder (1979) believe that the Valmy and Vinini should be regarded not as formations, but rather as tectonostratigraphic units that contain various lithologies that could range in age from Eocambrian to Devonian. Late Cambrian to Devonian deposition in the ocean basin produced pelagic and hemiterrestrial sediments, while carbonate and shale predominated in the miogeocline.

In order to close the extensive Eocambrian to Devonian ocean basin, the arc must have had reversed polarity (Figure 2-6) and faced southeastward, with the Roberts Mountains assemblage as part of an immense accretionary prism developed on the eastward side of the arc prior to collision (Dickinson, 1977, 1979; Schweikert and Snyder, 1979). Thus the stacking order of thrust sheets within the Roberts Mountains assemblage is related to the incorporation of successive packets of strata within the accretionary wedge during pre-collision subduction; higher structural units were accreted earlier and thus carried farther eastward than lower structural units. In the Shoshone, Galena and Sonoma Ranges, upper Cambrian Harmony generally occupied the highest post-Antler structural position, thrust over the Valmy-Vinini, which in turn was

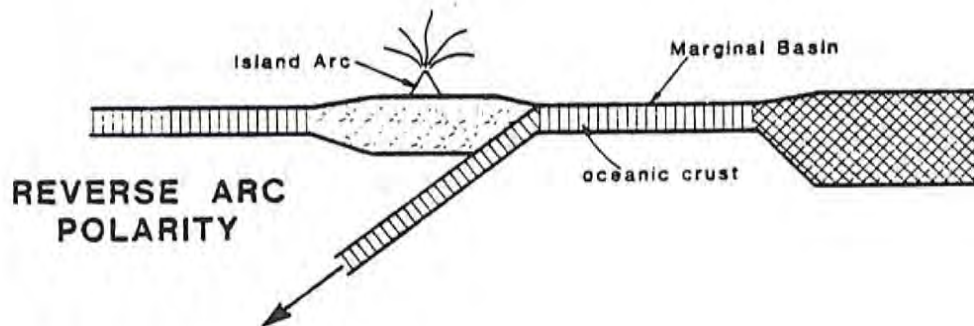
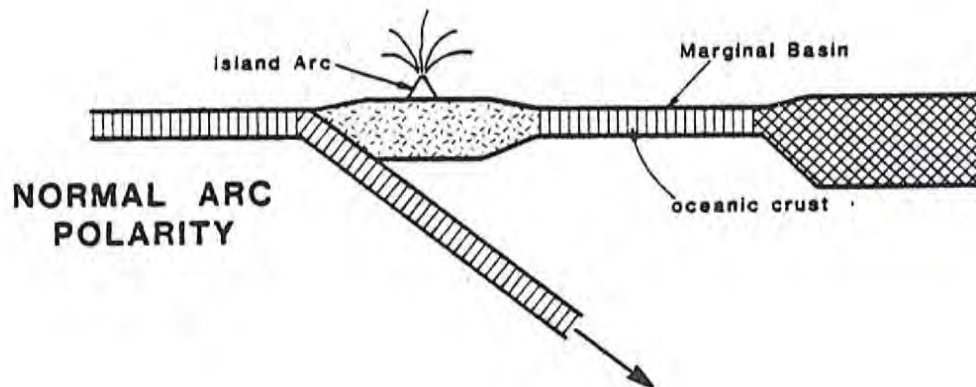


Figure 2-6. Diagrammatic sketch of normal arc polarity and reverse arc polarity relative to origin of oceanic marginal basins. (from Dickinson, 1977)

thrust over generally younger formations (Devonian Slaven Chert, Silurian Elder Sandstone) (Gilluly and Gates, 1965; Roberts, 1964; Silberling, 1975).

The location of pre-Ordovician oceanic crust remains a question. Since the Valmy-Vinini were probably deposited closer to the craton than the Harmony (i.e., farther from the rift), they presumably accumulated on older oceanic crust. The greenstone and quartzite within the Valmy-Vinini could be Cambrian or even Eocambrian in age, as they have not yet been dated (Schweikert and Snyder, 1979).

An island arc was active during Ordovician-Silurian time in order for westward subduction to occur. This arc and a related subduction sequence may be represented by the Alexander terrane. Paleomagnetic data of M. Jones and others (1977) very strongly supports the suggestion of D. L. Jones (1977) and Schweikert (1976, 1978) that the Alexander terrane in southeast Alaska, which contains abundant arc-like volcanic rocks of Ordovician, Silurian and Devonian ages (Churkin and Eberlein, 1977) was at or near the latitude of northern California during Ordovician and Silurian time.

Figure 2-7 depicts the early Paleozoic tectonic history for west-central North America. Subduction began beneath the Klamath-Sierra arc in late Ordovician or Silurian time (Figure 2-7A), with the incorporation of the Trinity ultramafic sheet into a nascent accretionary prism. During this interval, pelagic and hemiterrigenous sediment (in part the Valmy-Vinini) blanketed the ocean basin between the miogeocline and the arc. As subduction continued, the Harmony fan and the Valmy-Vinini were tectonically shuffled (thrust?) into a very large accretionary wedge (Figure 2-7B). This wedge formed the basement for arc volcanism which had stepped eastward by mid-Devonian. The Antler Orogeny marked the culmination of arc-continent collision. During the Antler Orogeny, the marginal basin closed, and the island arc was accreted to the continent, producing the Antler orogenic belt and foreland basin (Poole, 1974). Dickinson (1977) suggests westward consumption of the marginal basin, whereas Burchfiel and Davis (1975) explain a process of eastward plate descent involving uncoupling and overthrusting (obduction) of the island arc superstructure. The edge of the continent was partially subducted, causing the leading edge of the accretionary wedge to over-ride along the Roberts Mountains Thrust. Siliceous vol-

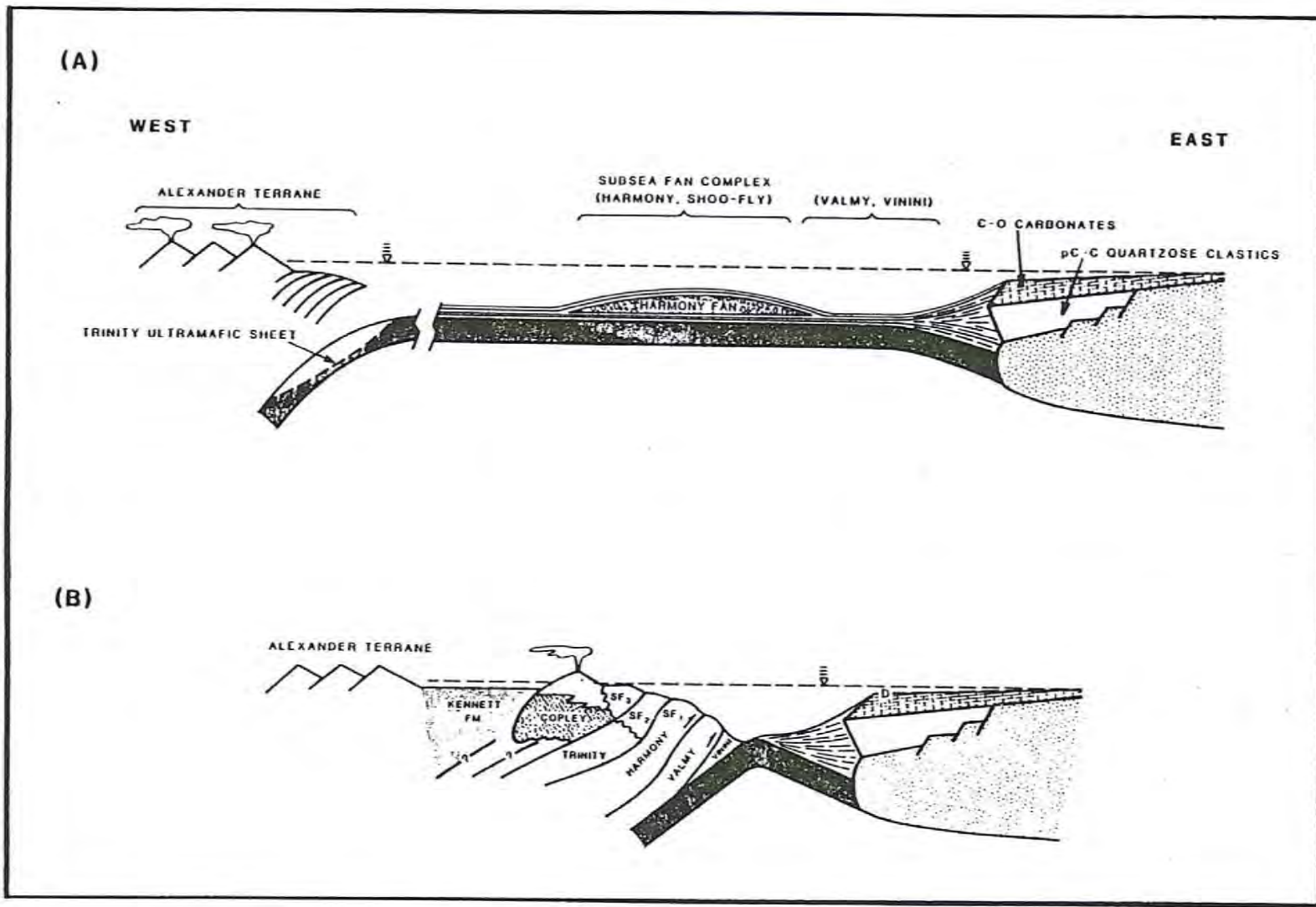


Figure 2-7. Schematic cross-section of early Paleozoic plate tectonics in late Ordovician - Silurian time (A) and in Devonian time (B). (from Schweikert and Snyder, 1979)

canism (Sierra Buttes Formation) occurred during this event. The Roberts Mountains Thrust was not a geologically instantaneous cataclysmic event, but the result of a slow, passive, accretionary process. The last 80 km of forward growth may have taken much of the Devonian and part of the Mississippian. In one locality, forward growth stopped at about the end of the Devonian (Smith and Ketner, 1968), but farther north, it continued in early Mississippian time (Ketner, 1970).

2.2.2.2 Mississippian - Pennsylvanian

Closely following the Antler Orogeny, the arc polarity reversed to a westward facing direction (Figure 2-6), and "normal" island arc andesitic subduction initiated back-arc spreading creating a marginal basin. Unlike the present-day Japanese-type arcs (Figure 2-2c) which face the subducting oceanic plate, these island arcs faced the North American Plate reflecting westward underflow of the marginal basin. Sediments of the Havallah sequence accumulated in the eastern portion of the basin and a thick volcanoclastic wedge (the base of the Calaveras Complex; Schweikert, 1979) formed on the arc-side. The Havallah sequence was deposited on structural remnants of the Roberts Mountains assemblage and on new oceanic crust formed between the rifted blocks. With continued back-arc spreading and marginal basin growth, olistostromes of chert and argillite, with rare lenses of greenstone and shallow-water limestone, were shed (from the Peale Formation?) into the deeper parts of the basin. Portions of the arc during Mississippian and Pennsylvanian time were apparently characterized by volcanic quiescence as the cherts of the Peale formation were deposited (Schweikert and Snyder, 1979). Early Permian faunas, one typical of the arc and one typical of the miogeocline (Stewart and others, 1977) occur in limestone olistoliths in the Havallah, suggesting that portions of the marginal basin, particularly in the north, were relatively narrow (Schweikert and Snyder, 1979). The marginal basin was thus composed of some rifted remnants of the Roberts Mountains assemblage of lower Paleozoic age in addition to upper Paleozoic oceanic crust, pelagic and clastic sediments.

By early Pennsylvanian time deposition of thick sequences of sand, gravel and mud that had filled the Antler foreland trough had nearly ceased (Roberts and others, 1958; Poole, 1974). In northern Nevada,

siliceous detritus continued to form local deposits until middle Pennsylvanian. On a regional scale, however, the early and middle Pennsylvanian are characterized by cherty limestone essentially free of detrital sediment. This limestone can be identified by various local names, but probably formed a nearly continuous sheet of uniform lithology over a very extensive area. Near the end of Pennsylvanian time, then, the western Basin and Range area was one of a tranquil sea broken by small, sparse insular remnants of the Antler Orogenic highland belt.

2.2.2.3 Permian

Facies and isopachous trends for miogeoclinal and cratonal rocks of late Precambrian and Paleozoic age can be projected southwestward across the California-Nevada border. However, the few small exposures of Paleozoic or probable Paleozoic rocks present in the western Mojave and southern Sierra Nevada regions are not counterparts of the units projected toward their locations from the northeast. A number of explanations have been offered to explain the distribution of these "out-of-place" Paleozoic units. Poole (1974), Poole and Sandberg (1977) and Poole and others (1977) attribute the irregular trend and distribution of the Antler orogenic belt (Roberts Mountains assemblage) between southern California and western Nevada to dextral oroclinal bending of Mesozoic and Cenozoic age. Dickinson (1979) concludes that the irregular trend of the Antler orogenic belt between the two regions may be a consequence of an initially irregular pattern of rifts and transform faults that outlined the western margin of the continent in late Precambrian time.

Three differing--but not totally inconsistent and exclusive--interpretations are provided by Dickinson (1977), Ketner (1977) and Burchfiel and Davis (1979) for the Permian in western North America that relate to the distribution of both structural trends and stratigraphic sequences.

Dickinson (1977) has proposed an episode of Carboniferous rifting as an explanation for the termination of the Antler foreland basin. He states that evolution of the foreland basin was anomalous in that the phase of flysch deposition was not succeeded by a phase of molasse deposition, but instead by shelf conditions. The presence of turbidites in the Death Valley region, the presence of Permo-Penn-

sylvanian oceanic assemblages in the Golconda allochthon, and mid-Carboniferous terrestrial and marine strata unconformably overlying the Roberts Mountain allochthon are given among the supporting data for this episode of rifting.

Ketner (1977) identifies the Humboldt Orogeny as a regional tectonic uplift beginning in late middle Pennsylvanian and persisting through Permian time that produced a north-trending belt of highlands (Figure 2-8). The Humboldt highland belt was a range of islands and shoals that extended northward from southern California through Nevada and Idaho and probably into western Canada. The distribution of orogenic sediments indicates that the belt was at least a thousand kilometers in length; the volume of orogenic sediments suggests the belt was comparable to a large mountain range, such as the present-day Sierra Nevada. Sediments shed eastward from this belt were deposited in a shallow epicontinental sea (Phosphoria sea) and are mingled with limestone, dolomite, evaporites, spicule chert, and phosphorite; those shed westward were deposited in deeper water as debris flows and turbidites, and mingled with coarse volcanic sediments, tuffs, lava flows, and radiolarian cherts. The volcanic sediments, tuffs, and lavas emanated from an island-arc that lay parallel to, and a short distance west of, the tectonic belt. Along the southern Nevada-California border, where Paleozoic facies boundaries and the Antler belt trended northeast, the Humboldt highland tectonic belt trends northwest, transecting the entire width of the Cordilleran sedimentation belt (Figure 2-8). Through central Nevada, the axis of this belt parallels the distal margin of the Precambrian North American continent, and succeeding Paleozoic trends.

Burchfiel and Davis (1979), however, propose that the distribution of outcrops of Precambrian through Paleozoic rock assemblages (including Roberts Mountain and overlying upper Paleozoic clastic wedge sediments) in southern California is a result of southward displacement along a major fault sliver or slivers. Attendant faulting is considered to represent the left-slip truncation (Davis and others, 1978) of a part of the North American continental margin during Permo-Triassic time (Hamilton, 1969; Burchfiel and Davis, 1972, 1975). The sinuous trend of this continental block and its marginal faults (Figure 2-9) is the product of later Mesozoic and Cenozoic deformations and rota-

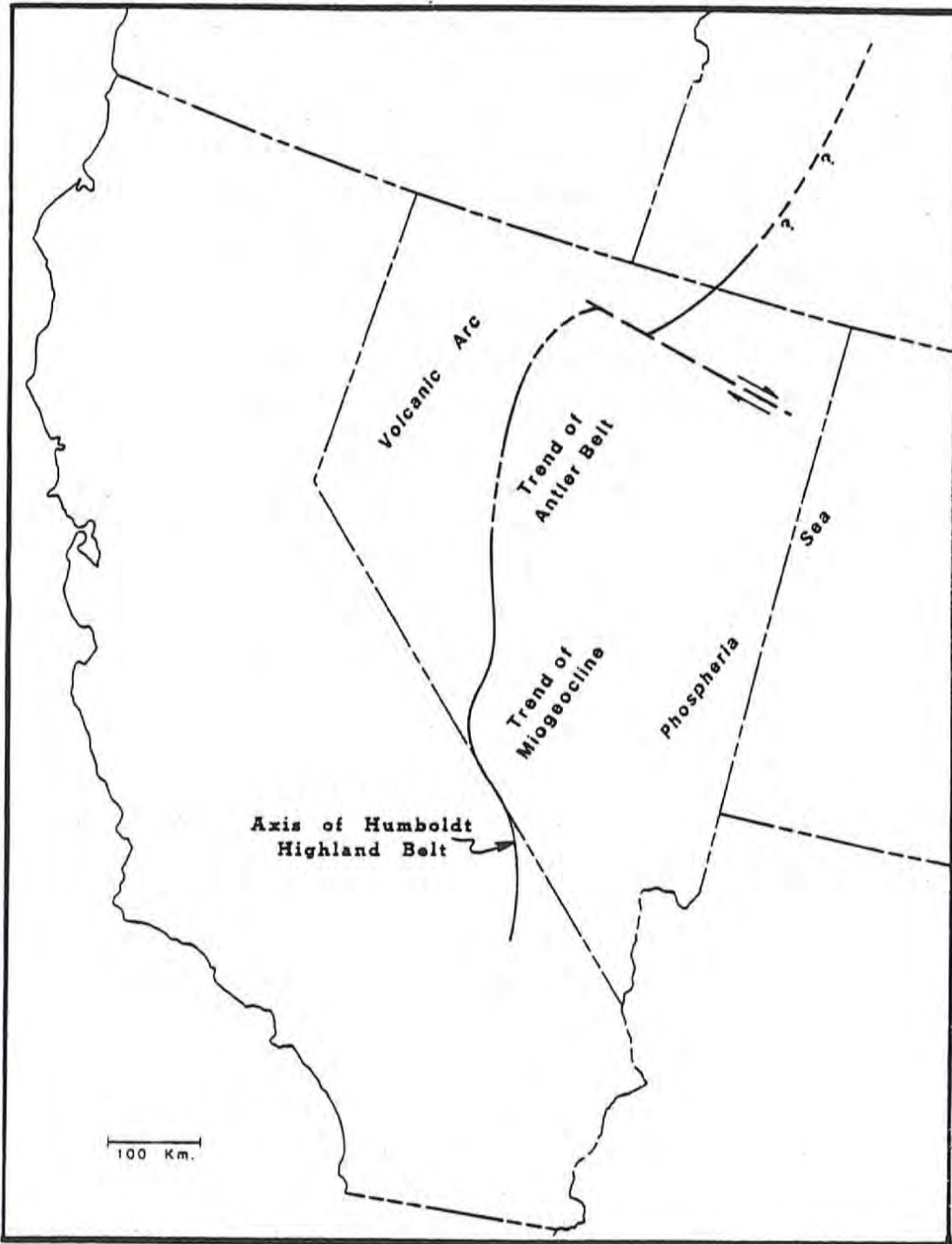


Figure 2-8. Diagrammatic sketch of Nevada showing position of the axis of the Humboldt highland belt in relation to the trend of the Antler Belt, the trend of the Cordilleran miogeocline, and the location of the Phosphoria sea. (from Ketner, 1977)

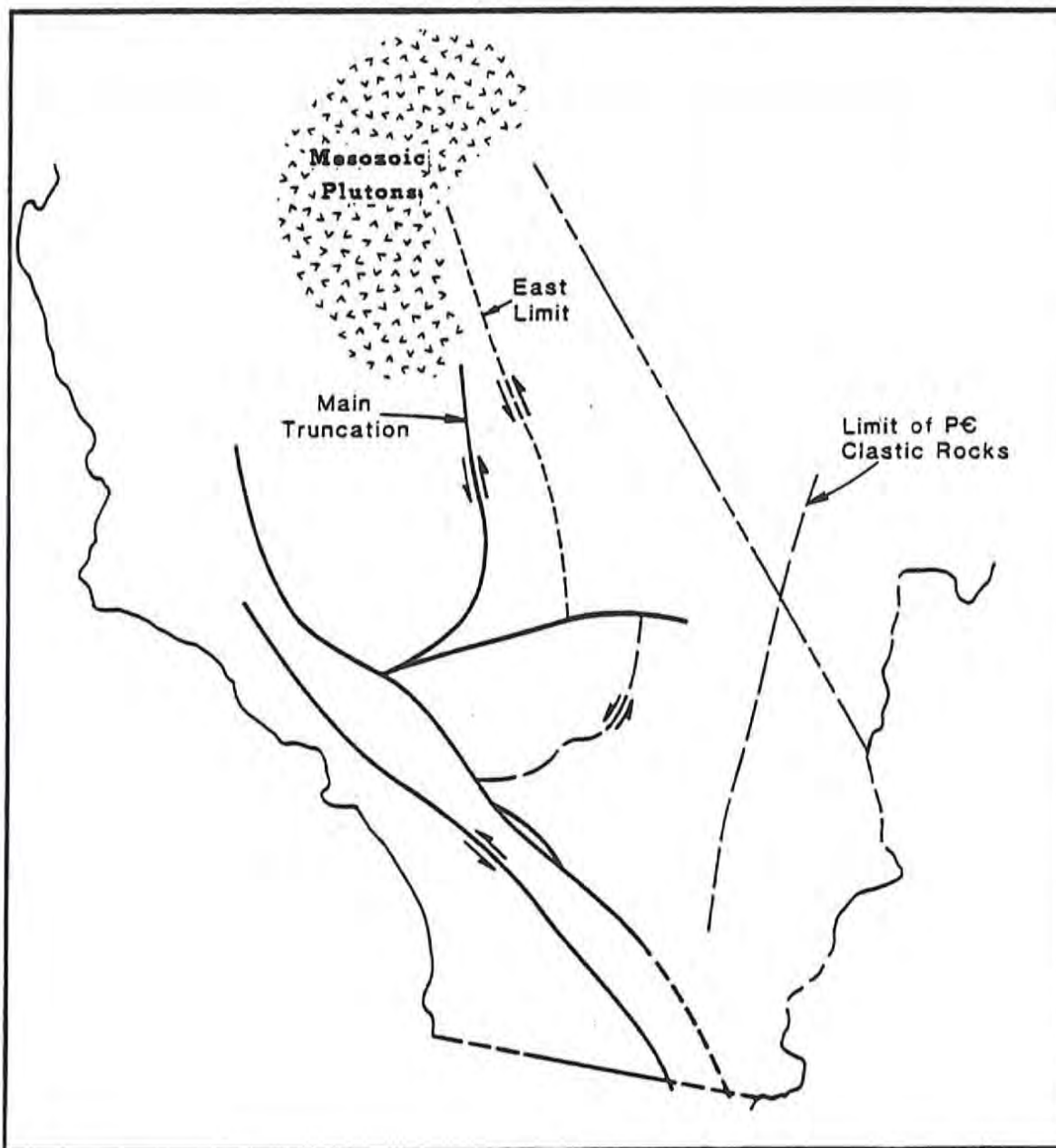


Figure 2-9. Diagrammatic sketch of southern California and Nevada showing location of the Permo-Triassic left-lateral truncation faults. Sinuous character of the faults is due to Cenozoic offsets. (from Burchfiel and Davis, 1979)

tions, with the northern extension of this zone obscured by younger Mesozoic plutonism. This truncation event must have occurred prior to the development of a Mesozoic marginal arc complex that cross-cuts Paleozoic geosynclinal and structural trends and must also pre-date the first phase of Mesozoic (Triassic) plutonic intrusion related to eastward subduction along a modified continental margin approximately 220 my ago (Evernden and Kistler, 1970; Silver, 1971). Lithologic and stratigraphic relationships, extensively discussed by Burchfiel and Davis (1972), in combination with the development of the Mesozoic arc and its associated volcanism thus place the time of continental truncation as middle to late Permian.

2.2.3 Mesozoic

2.2.3.1 Triassic (Sonoma Orogeny)

During the late Permian, the polarity of the island arc reversed again, and the arc began to subduct its own marginal basin. The Calaveras-Shoo Fly fault in California marks a deep-seated fault along which the island arc and its Paleozoic basement over-rode the marginal basin rocks of the Calaveras Complex during the initial phase of the Sonoma Orogeny. Eventually, the entire Calaveras-Havallah assemblage with numerous internal imbrications was obducted onto the continent as the Golconda allochthon. The Golconda thrust fault marks the maximum eastward extent of overthrusting of oceanic assemblages during the arc-continent collision of the Sonoma Orogeny (Figure 2-10).

In northwestern Nevada, the Sonoma orogen was sutured to the continent during mid-Triassic time (Silberling, 1973, 1975; Dickinson, 1977), being thrust across eroded remnants of the Antler allochthon (Roberts Mountains assemblage). However, no prominent foreland basin formed to the east of the Sonoma orogen. Instead, local rapidly subsiding basins along the trend of the suture belt received thick clastic accumulations, such as the Auld Lang Syne Group (Burke and Silberling, 1973), in upper Triassic and lower Jurassic time.

In the western Great Basin (north-central Nevada), the rocks in the upper plate of the Golconda thrust are referred to as the Pumpnickel and Havallah Formations (Ferguson and others, 1951; Muller and others, 1951; Roberts, 1951). As originally defined at the type section in the Tobin Range (Muller and others, 1951), the Pumpnickel Formation is an assemblage of primarily greenstone, bedded chert, and argillite; and the Havallah is the overlying, presumably younger, assemblage of primarily bedded chert, quartzite, limestone, and argillite, locally with conglomerate. Stewart and others (1977), however, suggest that lithologic, stratigraphic, structural and faunal evidence indicate that the Pumpnickel and Havallah are not lithostratigraphic units but are instead tectonic successions of upper Paleozoic rocks.

In general, the Calaveras complex exhibits a greater degree of deformation and is more strongly metamorphosed than the Havallah. Similarly, the Leach Formation (Ferguson and others, 1951) in the East Range, west of the Tobin Range, consists of structurally interleaved Valmy and Havallah (Whitebread, 1978; Stewart and Carlson, 1976) and

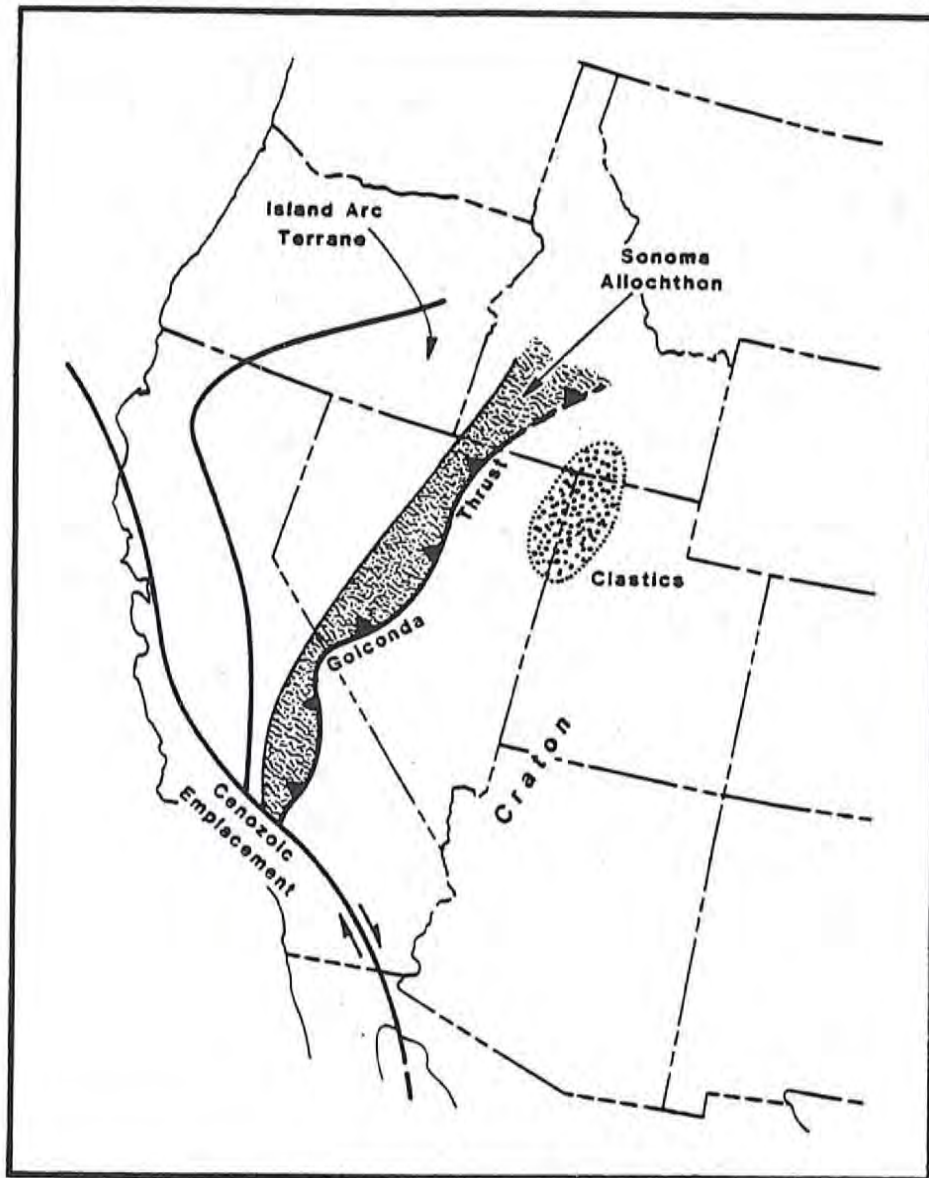


Figure 2-10. Generalized plate tectonic setting of California and Nevada during the Sonoma Orogeny. (from Ernst, 1979)

contains evidence of at least three episodes of deformation within the Sonoma Orogeny (Schweikert and Snyder, 1979). This distribution suggests that the grade of metamorphism and the degree of ductile deformation increased westward within the Golconda allochthon toward the volcanic arc.

Melting within the siliceous sediments of the Havallah sequence may have produced the silicic pyroclastic rocks of the Koipato sequence, analogous to the Sierra Buttes Formation of the Antler Orogeny (Schweikert and Snyder, 1979; Dickinson, 1977). Burchfiel and Davis (1972, 1975) and Silberling (1973, 1975), however, interpret the Koipato as the earliest vestige of a continental margin magmatic arc established on the continental edge by polarity reversals following arc accretion during the Sonoma Orogeny.

The episode of continental accretion during the Sonoma Orogeny abruptly shifted the continental margin several hundred kilometers from central Nevada to the western or outer side of the former Klamath-Sierra arc terrane (Burchfiel and Davis, 1979b).

By late Triassic time, an Andean-type (Figure 2-2b) continental margin had developed. The western margin of the North American continent northward from the southern Sierras had become the site of oblique (?) plate convergence and transcurrent motion (transpression; Saleeby, 1979), eastward subduction of Pacific Ocean lithosphere, and widespread arc magmatism (Hamilton, 1969; Monger and others, 1972; Burchfiel and Davis, 1979b). This continental-margin arc-trench system (Dickinson, 1977) would persist into the early Tertiary and included the following tectonic elements from west to east: 1) subduction complex composed of imbricated wedges and melange belts of dislocated ophiolitic slices and disrupted oceanic sediments with overprints of blueschist metamorphism; 2) forearc basin of arc-derived clastics in simple stratigraphic and structural order; 3) magmatic arc composed at high structural levels of generally andesitic volcanics and volcanoclastics, with overprints of greenschist metamorphism, and at depth of granitic plutons, with higher grade thermal aureoles; 4) infrastructural belt of deep-seated dynamothermal metamorphism with the potential for associated anatectic plutons; 5) backarc fold-thrust belt of uplifted imbricate thrust sheets with decollement; and 6) retroarc foreland basin with exogeosynclinal

clastic wedges shed from orogenic highlands in the fold-thrust belt along its orogenic flank.

2.2.3.2 Jurassic - Cretaceous (Nevada and Sevier Orogenies)

Until middle Jurassic, the Cordilleran magmatic arc was an offshore island arc (Stanley and others, 1971; Monger and others, 1972) near the edge of the continental block with shelf seas behind it (Figure 2-11). The main shoreline met the arc near the southern limit of the present Great Basin. Local thrusts (Burchfiel and Davis, 1972, 1975) and small plutons (Silberman and McKee, 1971; Armstrong and Suppe, 1973) in the Great Basin record the inception of a backarc fold-thrust belt, and possibly the initiation of an infra-structure belt. These structures propagated northward with time.

The subduction zone is well located only in California in the Sierra Nevada foothills (Schweikert and Cowan, 1975). The site of the subduction zone was stepped westward from the Sierra Nevada foothills (Figure 2-11) to a position in the Coast Ranges. Local evolution (westward migration) of the continental margin resulted from the tectonic accretion of island-arc structures to the continental-margin subduction complex following consumption of the intervening oceanic lithosphere (Schweikert and Cowan, 1975). By late Jurassic time, the magmatic arc in California had also stepped westward, with the shift of the subduction zone, to occupy the Sierra Nevada foothills.

During the Nevadan Orogeny in the western Great Basin, plutonic and volcanic activity generally parallels pre-Triassic trends. The eastern limit of Jurassic volcanic rocks (Burchfiel and Davis, 1972) roughly corresponds with the 118° line of longitude in western Nevada. Small, local thrusts also developed within Paleozoic rocks and early Mesozoic back-arc sedimentary and volcanic rocks. Hamilton (1969) and Dickinson (1969, 1970) related the plutonic rocks of this pulse of activity (160 to 210 my ago; Lanphere and Reed, 1973) to the underflow of the oceanic plate along the Pacific margin.

The partly contemporaneous middle Jurassic to Cretaceous Sevier Orogeny (Figure 2-12) is characterized by an eastward migrating zone of back-arc fold-thrusting principally in the eastern Great Basin affecting Cordilleran miogeoclinal Paleozoic rocks. A pulse of plutonic activity (132 to 158 my ago; Lanphere and Reed, 1973) corres-

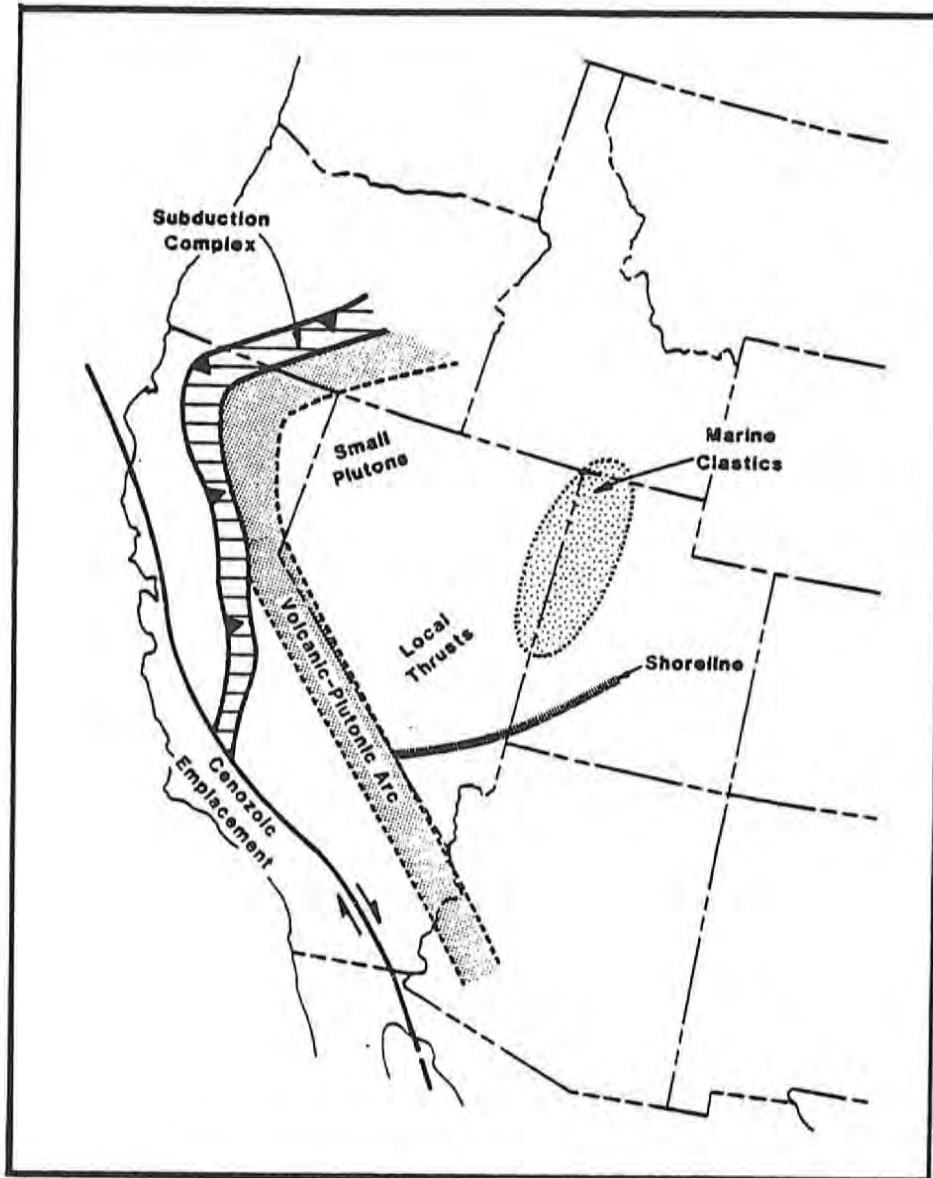


Figure 2-11. Generalized plate tectonic setting of California and Nevada during the Nevadan Orogeny. (after Dickinson, 1976; Ernst, 1979)

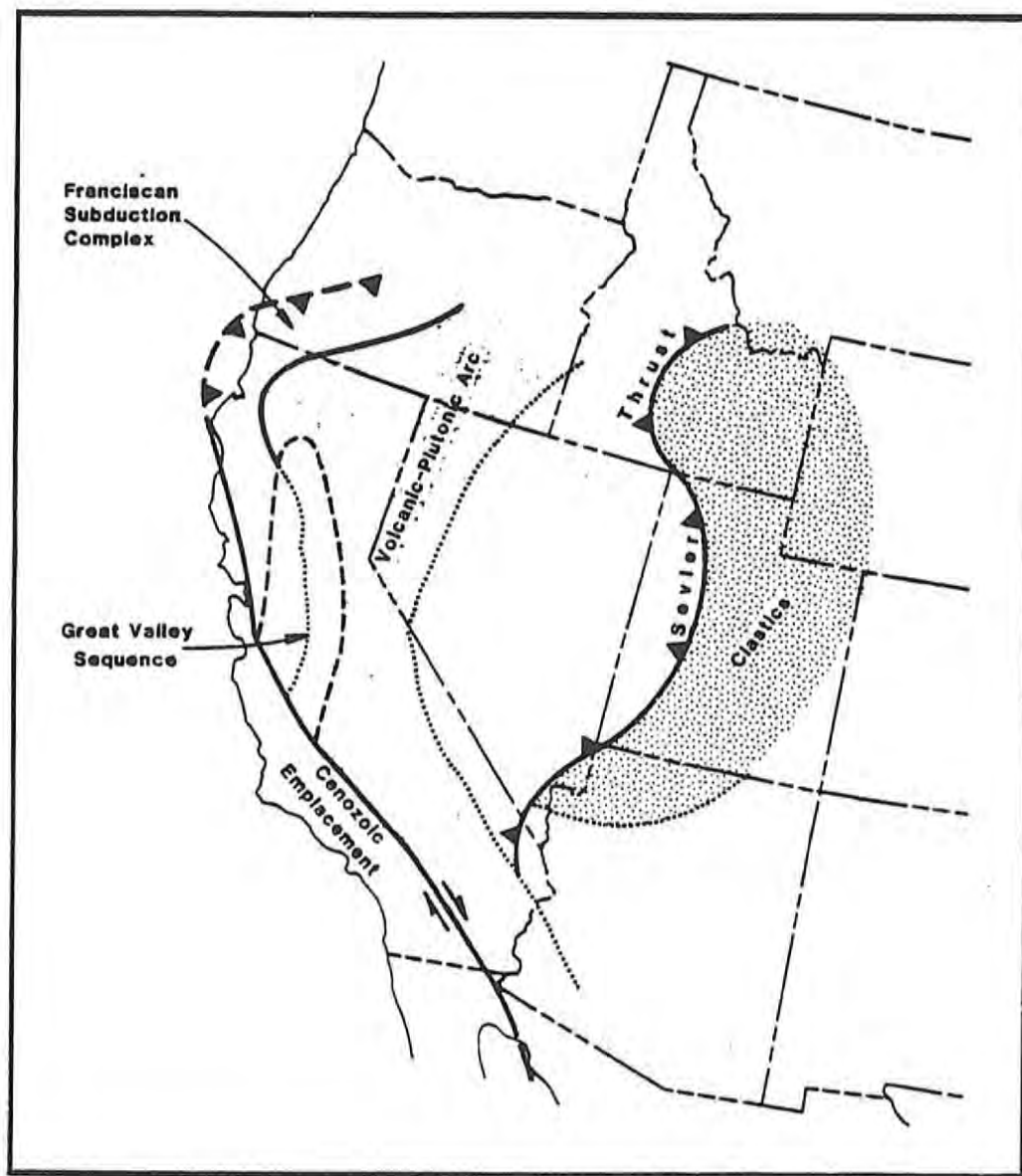


Figure 2-12. Generalized plate tectonic setting of California and Nevada during the Sevier Orogeny. (from Ernst, 1979)

ponds with the Sevier Orogeny. In general, this plutonism occurred farther eastward and northward of earlier Mesozoic intrusions.

By mid-Cretaceous time the magmatic arc had begun a retrograde migration (Dickinson, 1973) that continued throughout the Cretaceous and into the Tertiary. This trend is related to flattening of the Benioff Zone and entrainment of subducted slabs of Pacific lithosphere beneath the North American plate (Coney, 1973). Concurrent with retrograde movement of the arc, the subduction zone underwent prograde migration with accretionary growth of the Franciscan subduction complex. Thus, the arc-trench gap was progressively widened (Dickinson, 1973, 1977).

During the late Cretaceous both the retroarc foreland basin and forearc basin (Great Valley) achieved maximum extent. The arc-trench gap continued to increase. Localized plutonic activity continued in Nevada, with five major intrusive episodes for the entire late Triassic to late Cretaceous time interval recognized in the Sierra Nevada (Kistler and others, 1971). Episodes of regional metamorphism can generally be correlated with the Mesozoic orogenic intervals.

2.2.4 Cenozoic

2.2.4.1 Tertiary

2.2.4.1.1 Paleogene (Paleocene - Eocene - Oligocene)

The Laramide (Cordilleran) Orogeny in the eastern Cordillera included thrusting and folding along the same trend as the earlier Sevier Orogeny as well as an eastern belt of basement uplifts (Figure 2-13). Laramide deformation involved regional crustal buckling of the interior foreland to produce fault-bounded and basement-cored uplifts in the eastern Cordillera (Robinson, 1972; McDonald, 1972). During this same Paleocene time interval, a prominent hiatus in arc magmatism covered the Great Basin, while subduction continued along the coast (Lipman and others, 1972). Dickinson and Snyder (1978) attribute these conditions to an episode of shallow subduction beneath the Cordillera. Cross (1973) and Suppe (1970) however, suggest the lack of magmatic activity is the result of temporary cessation of subduction and plate consumption along the segment of the continental margin affected by development of the proto-San Andreas transform fault.

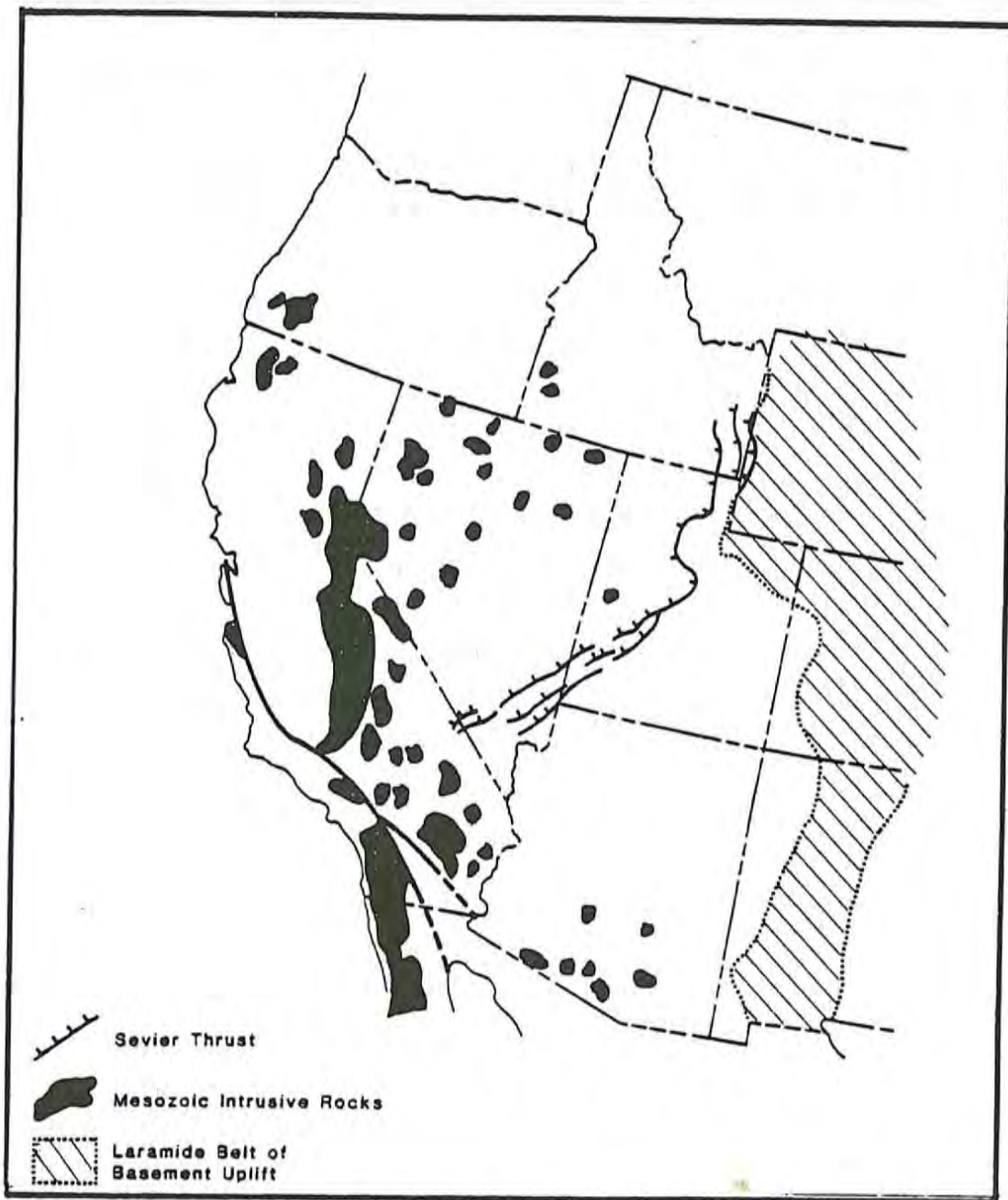


Figure 2-13. Generalized diagram of Mesozoic deformation belts in the western United States. (from Stewart, 1978)

Such a transform would prevent subduction of lithosphere beneath the continent to generate arc magmatism.

Armstrong (1974a) suggests that Laramide deformation reflects crustal contraction in a region where the lithosphere lacked sufficient ductility for the development of a backarc fold-thrust belt. Arc magmatism was dormant across the region between the Laramide uplifts and the proto-San Andreas. Burchfiel and Davis (1975) suggest that regional location of Laramide deformation reflects an eastward migration of the edge of the zone of lithosphere ductile enough to allow for crustal contraction. This is supported by the areal extent of arc magmatism when it resumed in the Oligocene extending to the Laramide uplifts in Colorado. These relationships must also be related to the subsurface geometry of subducted lithospheric slabs (Lipman and others, 1971) and the geometry of absolute plate motions (Coney, 1972, 1973). Coney (1971) has suggested a change in the direction of motion of the North American Plate corresponding with the change in pattern of igneous activity during the Laramide Orogeny.

Laramide deformation inland ended by about the end of the Eocene (Coney, 1971, 1972, 1973, 1976). The Laramide Orogeny was bracketed by major changes in global patterns of sea-floor spreading and motions of plates (Coney, 1976). This regionally extensive and profoundly compressional period of deformation may coincide with high convergence rates between the North American and Farallon plate to the west (Coney, 1978).

Contemporaneous with Laramide deformation in the Rocky Mountain, Colorado Plateau and eastern Great Basin region was an episode of tectonism in the Pacific trough--the Coast Range Orogeny. Deformation in the Coast Range belt was characterized by underthrusting along the continental margin and local intrusive activity (Dickinson, 1969, 1970, 1979; Page, 1970, 1979; Dewey and Bird, 1970). Both the Laramide and Coast Range belts developed in troughs that flanked the Mesozoic Nevadan and Sevier orogens and that were characterized by accumulations of thousands of feet of sediments derived from the orogens and local volcanic rocks of late Mesozoic age.

During the Paleocene and Eocene, the ancestral Cordillera extended unbroken from Mexico northward through western Nevada and into Canada.

The western Great Basin region (in central and eastern Nevada) received sediments from the Cordillera on the west and the Laramide uplifted region on the east in central Utah. Calc-alkaline volcanic activity was widespread in the Pacific Northwest and by the end of Eocene time had migrated through Idaho and perhaps into the northeastern corner of Nevada. Generally the Great Basin region was a low upland of subdued topography, westward external drainage, and extensive erosion. The earlier formed deformational structures and their accompanying terrains had been severely reduced to a surface probably no higher than a few hundred meters above sea level (Nelson, 1979).

The Oligocene was a time of progressive southwestward migration of volcanic activity across the Great Basin (Armstrong, 1975; Mckee and Silberman, 1975). By the end of the period volcanic activity had shifted from a diffuse belt in Utah and central Oregon to a northwest-trending belt in western Nevada. These volcanoes erupted calc-alkaline lavas and ash; the larger ash-flow eruptions resulted in the formation of collapse calderas in the Great Basin. The low topographic relief in the region offered little restriction to the ash-flow eruptions which spread over enormous areas as thin blankets of welded tuff (Figure 2-14).

The ancestral Cordillera, although now disrupted by ongoing rotation of the Klamath Mountains and Coast Range further north, became volcanically active in southeastern and east-central California and in northwestern Nevada (Nilsen and McKee, 1979). This relationship, in addition to the extensive regional volcanism in the western states, suggests that subduction must have been occurring along the entire western margin prior to collision with the Pacific Plate, which instituted a regime of right-slip transform faulting beginning about 29 my ago (Atwater, 1970; Atwater and Molnar, 1973).

The pattern of migrating calc-alkaline volcanism across the inland western United States during the Paleogene reflects, in part, the active subduction of the Farallon oceanic plate during this time. The lack of volcanic activity in the west-central part of the region during Paleocene and Eocene time (Lipman and others, 1972; Snyder and others, 1976) may be a result of strike-slip motion along a postulated proto-San Andreas fault in central and southern California (Cross,

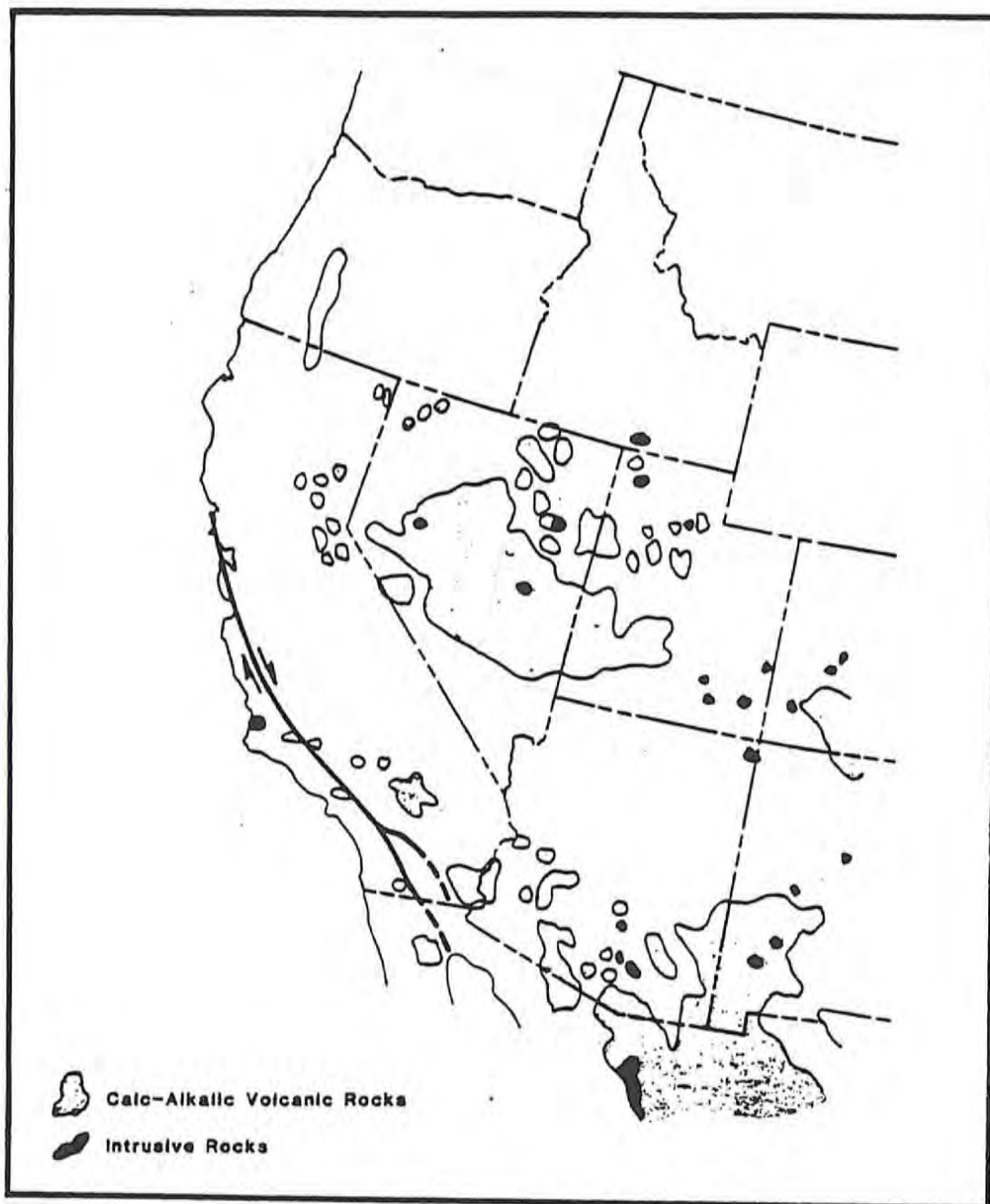


Figure 2-14. Generalized diagram of Oligocene volcanic rocks in the western United States. (from Stewart, 1978)

1973; Suppe, 1970; Nilsen, 1978); changing dips of the subducting Farallon plate (Coney and Reynolds, 1977; Keith, 1978; Burke and McKee, 1979); oblique subduction, rifting in southern California related to deposition, deformation, and eastward thrusting of the schists (Haxel and Dillon, 1978), or a combination of these factors (Nilsen and McKee, 1979).

Prior to 29 my ago, the Farallon Plate, separated from the Pacific Plate to the west by the East Pacific Rise spreading center, was being consumed at the subduction zone along western North America. The intersection of the spreading ridge with the subduction zone about 29 my ago changed the plate geometry and a transform fault system developed, marking the end of subduction along the southern California continental margin (Atwater and Molnar, 1973). As the transform boundary propagated northward, subduction was progressively terminated (Christiansen and Lipman, 1972; Lipman and others, 1972). The subducted plate continued to yield calc-alkalic volcanic rocks in the Basin and Range province, but by 18 my ago (early Miocene) this system was dead (Nilsen and McKee, 1979; Stewart, 1978). Coincident with the change in plate geometry, was the development of a new stress field that would control late Cenozoic tectonism and magmatic evolution in western North America, particularly the Great Basin.

2.2.4.1.2 Neogene (Miocene - Pliocene)

During early Miocene time the Sierra Nevada was an andesitic volcanic arc in which arc volcanism was progressively truncated from south to north as the continental margin transform fault system migrated northward. To the east, the Great Basin was dominated by intermediate volcanism. Rhyolitic volcanics and interbedded non-marine deposits were emplaced on an old erosion surface of low to moderate relief. This activity was not accompanied by any marked increase in tectonism with the exception of caldera collapse resulting from volcanic activity (McKee, 1971).

By mid-Miocene time, the remnant arc and subduction zone was about 2200 km in length and was bracketed by transform systems to the north and south (Christiansen and McKee, 1978). A subduction zone remains at the continental slope only along the continental edge segment adjacent to the Cascade arc (Silver, 1972). This configuration of relative plate

motions and interactions produced a stress field from the increasing Pacific-North American interaction that replaced the declining stress field from Farallon-North American plate interaction. Under these conditions, the regions of North America that had been heated and deformed during Oligocene arc magmatism behaved as ductile material and partial coupling developed. This forced extension across an axis essentially parallel to the continental margin arc about 300 to 400 km inland (Figure 2-15); extension roughly perpendicular to this axis opened a rift that was continuous northward from central Nevada (Christiansen and McKee, 1978). Stress relief at the base of the lithosphere along this rift produced basaltic magma generation in the upper mantle. In Nevada slight partial melting at considerable depth of a tectonically thickened Paleozoic sequence produced alkali-rich basalts. This change from intermediate-composition calc-alkaline to alkali-calcic suite to a bimodal basalt-rhyolite suite is correlated with the initiation of an episode of extensional faulting in the western United States by Christiansen and Lipman (1972), Lipman and others (1972), McKee and others (1970) and Armstrong and others (1969). Further north the plateau flood basalts (Figure 2-16) were produced by partial melting of Mesozoic oceanic crustal rocks.

During the next few million years, tectonic extension and uplift continued across the entire Great Basin region. Regional melting at the base of the lithosphere decreased lithospheric rigidity with subsequent deformation occurring at considerable depth by thinning of the crust and flowage of the mantle. Basaltic magma accumulation in the crust further increased regional heat flow, caused regional uplift and reduced rigidity. This resulted in progressive restriction of brittle deformation to upper-crustal levels away from the axial zone. Thus, faulting (seismicity) and basaltic volcanism have been progressively concentrated toward the Great Basin margins (Wasatch Line and Sierra front), and the axial zone may be hotter at depth than the margins (Christiansen and McKee, 1978).

The volcanic transition in most areas of the Great Basin was about 17 my ago (McKee and others, 1970; McKee, 1971). Christiansen and Lipman (1972), Proffett (1972), and Best and Hamblin (1978) have suggested a general northward migration of the time of inception of Basin and

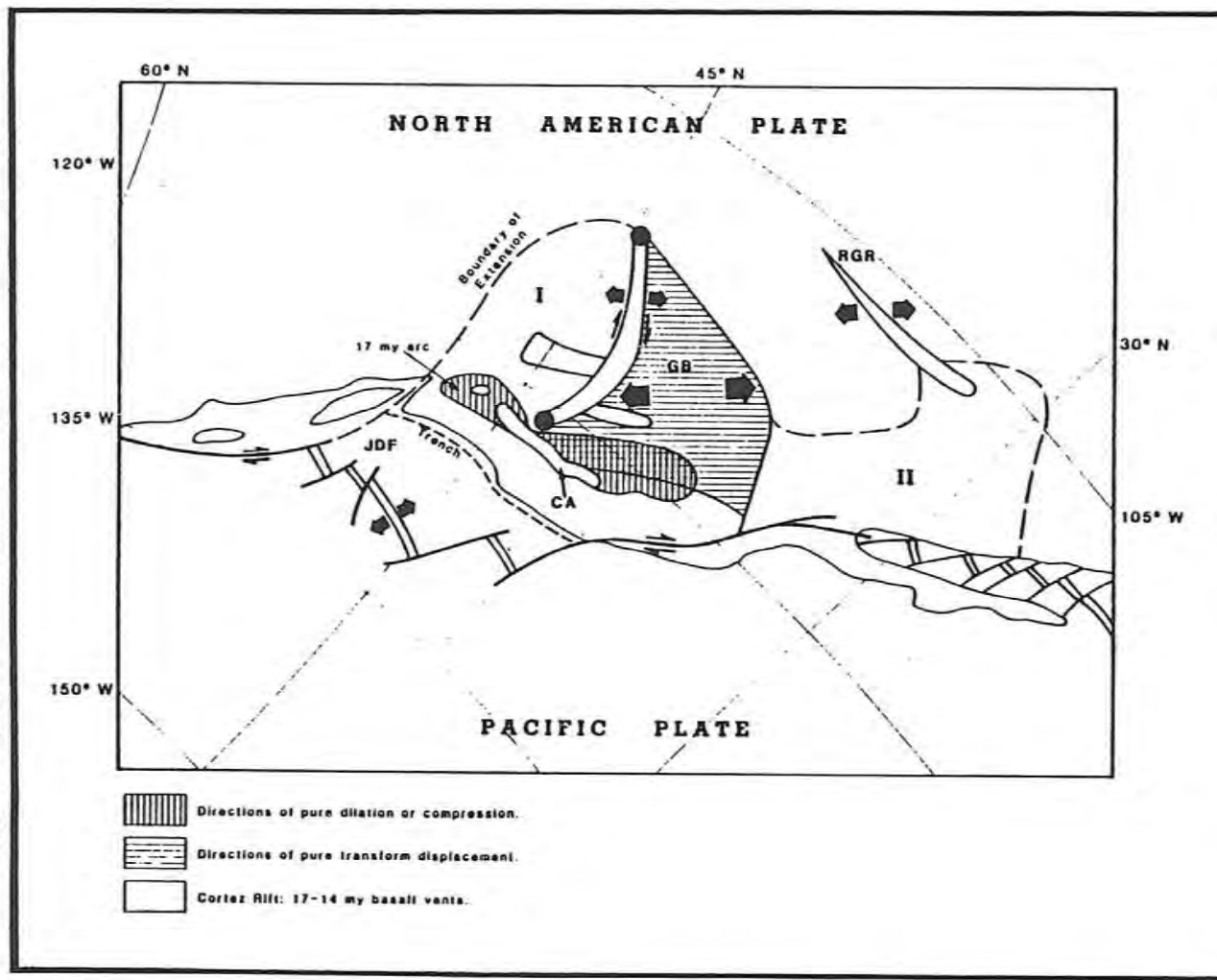


Figure 2-15. Generalized map of western United States showing late Cenozoic tectonic features and upper Cenozoic volcanic rocks in relation to Pacific-North American Plate interaction. Oblique Mercator projection of Atwater (1970); arrows indicate displacement vectors; volcanic centers shown as solid regions. CA - Active Cascade arc; GB - Great Basin, region of oblique extension; JDF - Juan de Fuca Plate, remnant of Farallon Plate; I - region of nearly inactive minor extension; II - region of nearly inactive major extension. (from Christiansen and McKee, 1978)

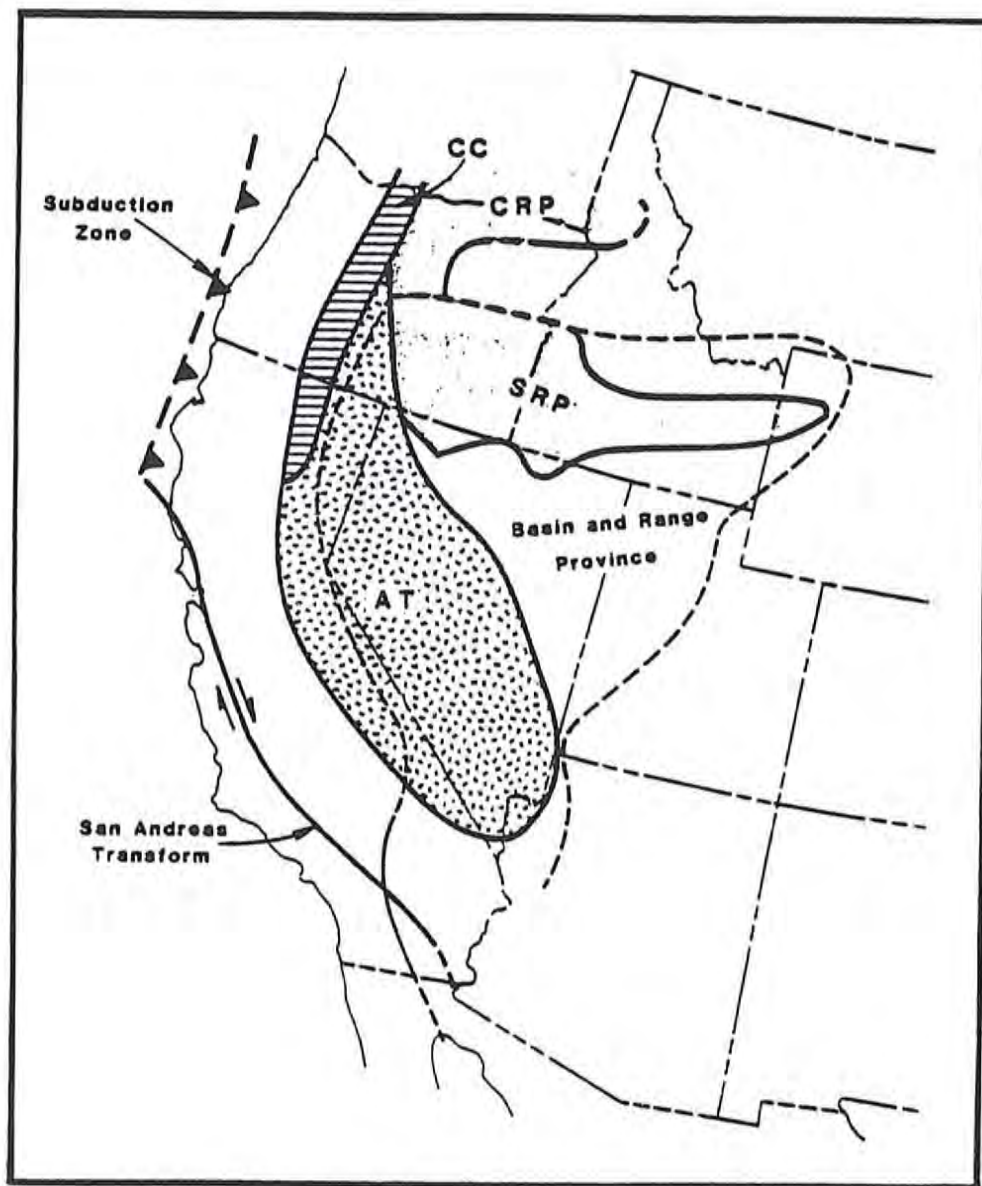


Figure 2-16. Generalized plate tectonic setting of the western United States, including the Great Basin, showing the Cordilleran arc-trench system in middle Miocene to Recent time. CC - Cascade Chain; CRP - Columbia River Plateau of mainly mid-Miocene flood basalts; SRP - Snake River Plain of mainly late Miocene and younger basalts. (from Dickinson, 1976)

Range faulting, with large sedimentary basins well defined by about 11 to 13 my ago (Axelrod, 1957; Robinson and others, 1968; Gilbert and Reynolds, 1973). Regional development of north-trending normal-fault bounded basins representative of present-day topography probably did not begin until mid-Pliocene time and continued into late Pliocene time (Stewart, 1978). During this time interval, approximately one km of uplift occurred in the Great Basin region and is documented by mid-Cenozoic floras (Proffett, 1972; Suppe and others, 1975).

Pliocene regional uplift, extension and graben formation in the Great Basin was coincident with an episode of uplift and westward migration of the Sierra Nevada. Tilting of the range was progressive with about one-third of the present tilt developed between 9.5 and 18 my ago and the latter two-thirds taking place in the last 9.5 my (Noble and Slemmons, 1975). Based on floral data, the Sierra Nevada was a broad ridge with elevations of about 3000 feet during late Miocene and early Pliocene time (Axelrod, 1962). By late Pliocene time the range stood at approximately its present elevation (Christensen, 1965).

The grabens produced by normal extensional faulting within the Great Basin became sites of interior drainage in which were deposited clastic wedges of fluvial, alluvial and lacustrine origin. The pattern of Basin and Range extension and non-marine sedimentation persisted through the Pliocene and into Pleistocene time. As the depositional basins filled, the drainage systems began to coalesce and exterior drainage developed.

2.2.4.2 Quaternary (Pleistocene - Recent)

The Pacific-North American plate interactions that developed during the Neogene have continued to produce regional extensional tectonics in the Great Basin-Sierra Nevada region through Quaternary time. The horsts and grabens within the Basin and Range Province have been enhanced by continued movements on normal tensional faults. Major historical earthquakes within the Province attest to the current activity along these zones (Ryall and others, 1966).

Depositional sequences indicate that the exterior drainage systems of late Pliocene-early Pleistocene time were replaced by interior basin drainage following the Pleistocene glacial cycles and coeval pluvial lakes, such as Lake Lahontan in western Nevada.

Quaternary basaltic activity has been concentrated near the margins of the Basin and Range Province (Best and Brimhall, 1974), with basalt erupted from vents along normal faults in only a few areas. Present volcanic arc activity in the western United States in the Sierra Nevada is limited to areas north of about 40 degrees latitude (Snyder and others, 1976).

The distribution and orientation of late Cenozoic extensional fault patterns in western North America clearly follow pre-existing patterns. Some similarity in the distribution of late Cenozoic extensional faulting and the position of the Paleozoic and latest Precambrian Cordilleran miogeosyncline can be noted, particularly the eastern margin of the Basin and Range with the Wasatch Line. A close correlation can be seen in the patterns of Mesozoic and earliest Cenozoic tectonic deformation, early and middle Cenozoic igneous activity, and late Cenozoic extensional faulting (review Figures 2-4 through 2-16).

2.3 Basin and Range Province Characteristics

The geotectonic history of western North America has resulted in the development of the Basin and Range Province that can be characterized by the parameters presented below.

2.3.1 Crustal Thickness

Seismic refraction data indicate a relatively thin crust, with average values ranging from 25 to 35 km (Pakiser, 1963; Pakiser and Zeitz, 1965; Prodehl, 1970). The region is characterized by a 25 km thick crust in the east and an 18 km thick crust on the west side (Dixie Valley region) of the Great Basin, with a 30 km thick central zone (Smith, 1978).

2.3.2 Upper Mantle Velocity

Low upper mantle (P_n) velocity values ranging from 7.7 km/sec to 7.9 km/sec (Herrin and Taggart, 1962; Pakiser, 1963; Koizumi and others, 1973) suggest a lithospheric plate of NE strike and SE dip within the upper mantle; such plates have been identified in areas of active subduction (Isacks and Molnar, 1971). However, the presence of tensional rather than compressive stresses normal to the zone and a maximum 15 km focal depth of earthquakes with no deeper seismic activity suggest that

the zone is inactive and that the present NW-SE extension may be at least partly due to a release of compressive stress behind a NE trending paleo-subduction zone (Koizumi and others, 1973). Termination of active subduction in the zone may have coincided with initiation of Basin and Range spreading approximately 17 my ago (Stewart, 1971).

2.3.3 Heat Flow

High average heat flow values of greater than 2.0 have been recorded for the Basin and Range (Lee and Uyeda, 1965; Roy and others, 1972; Morgan, 1974; Thompson and Burke, 1974) compared to normal continental values of about 1.5 HFU. Sass and others (1971) indicate consistently high values--mostly above 3 HFU--within the Battle Mountain High, which is located at the northern end of the Ventura-Winnemucca zone of historic seismicity and active faulting (Ryall and others, 1966). Blackwell (1978) indicates that transitions between thermal provinces are narrow (generally less than 20 km); therefore the sources directly responsible for the surface-measured variations in heat flow must be in the crust. Thermal boundaries also usually appear in areas of contemporary seismicity. The origin of the high heat flow of the Battle Mountain High is attributed by Sass and others (1971) to transient effects of a fairly recent crustal intrusion.

2.3.4 Fault Patterns

The pattern of fault rupture within the Basin and Range is generally described as high-angle normal faults trending north to northeast. However, Wright (1976) has shown that faulting within the Province is primarily related to the orientation of the structures with respect to the regional WNW tensional stress direction. Northwest trending faults have primarily right-lateral offsets similar to the San Andreas (e.g., Walker Lane) and NE trending faults have left-lateral offsets similar to that of the Garlock system of California (e.g., Carson Lineament). Extensional relationships generate horst and graben structures defined by predominantly normal-slip faults that parallel the N-S to N-NNE trends (e.g., Dixie Valley). Zones of extension are systematically spaced about 24 to 32 km apart. Considerable warping and tilting of the blocks accompany the faulting, particularly near the ends of elongate basins. The resultant fault pattern tends to be rhomboid or orthogonal and is

apparent on regional (e.g., Sierra Nevada front), local (e.g., Sand Springs Range, Churchill County, Nevada), and site-specific (e.g., Dixie Valley) scales (Trexler and others, 1978). The 15 km maximum focal depth common to the region suggests a lower limit to which the faults are propagated by brittle failure.

2.3.5 Strain Rates and Orientations

Focal mechanism solutions compiled by Scholz and others (1971), Suppe and others (1975), and Smith and Lindh (1978) are compatible with conjugate faulting on N-S normal-slip and NE-SW and NW-SE strike-slip faults in response to regional strain directions. Strain measurements (Priestley, 1974; Ryall and Priestley, 1975) suggest regional and temporal changes in rate and direction of strain with alternation in compression and extension on the NW-SE principal axis. The orientation of the strain axes is in agreement with the earthquake focal mechanisms of the region. Extension across individual grabens averages about 2.4 km and total extension across the Basin and Range is estimated at 50 to 100 km (Hamilton and Myers, 1966; Stewart, 1971; Thompson and Burke, 1974), with extension rates on the order of 0.5 to 1.5 cm per year across the Province (Stewart, 1971; Thompson and Burke, 1974; Slemmons and others, 1979). The region appears to be a topographically up-domed area, averaging 1-2 km above sea level, with an area in the west-central Great Basin elevated the greatest amount, about 2.2 km (Suppe and others, 1975).

2.3.6 Petrologic Relationships

Prior to Basin and Range faulting, lower and middle Cenozoic volcanic rocks were largely intermediate composition becoming more alkalic toward the continental interior (Lipman and others, 1972). A major change to basaltic volcanism, including bimodal mafic-silicic associations, took place during late Cenozoic time at about the inception of Basin and Range faulting (Christiansen and Lipman, 1972). The transition began in the southeastern part of the region and moved northwestward. Prior to Basin and Range faulting, the upper mantle to a depth of about 100 km contained about 75 percent peridotite and pyroxenite and 25 percent eclogite (McGetchin and Silver, 1972). Eclogite is capable of converting into gabbro with a volume expansion of about 10 percent in response to a

rise in temperature or a decrease in pressure. The expansion of eclogite in only 60 km of mantle could produce an uplift of 1.5 km. The former eclogite may now be represented by gabbro dispersed in the mantle low velocity zone, or by crustal additions of basic metaigneous rock, or by basaltic volcanics.

2.3.7 Magnetic and Electrical Anomalies

Zietz (1969) showed that magnetic anomalies are subdued in amplitude and that long-wavelength anomalies are absent. This fact suggests that the lower crust and mantle may be above the Curie temperature (578° for magnetite). Porath and Gough (1971) measured geomagnetic fluctuations to determine variations in mantle electrical conductivity. The anomalies are well represented by variations in depth to a half-space conductivity 0.2 ohm-m^{-1} . The top of this conductor is inferred to correspond approximately with the 1500° isotherm. Depths to the surface of the conductor are about 190 km. These data generally support a hot upper mantle in the Basin and Range Province as compared to the surrounding regions. The predominant magnetic highs in western Nevada generally correspond with north trending belts of Miocene intrusives, including the mid-Miocene Cortez Rift of Mabey and others (1978).

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Chapter 3. STRUCTURAL-TECTONIC ANALYSIS
By: Robert A. Whitney

3.0 STRUCTURAL-TECTONIC ANALYSIS

3.1 Introduction

3.1.1 Purpose and Scope

This study was undertaken to develop a geologic model for the Dixie Valley geothermal system with respect to structural and tectonic features. The study was conducted in two stages: Stage I included evaluation of available geophysical and geologic data, including fault and lineament maps and various imagery formats, such as Landsat, SKYLAB, NASA underflights, AMS photography and other conventional and low-sun-angle (LSA) photography. A bibliography search was initiated during Stage I. Stage II consisted of conducting a LSA photographic program of the study area, interpreting the resultant photography, and field checking the photogeologic data. Field work also included a preliminary fault scarp morphology study and compilation of a generalized surficial geologic map of the study area.

3.1.2 Methods and Analytical Techniques

3.1.2.1 Literature Search

The literature search was conducted primarily at the library of the University of Nevada-Reno; results of a limited computer search conducted by the Lawrence Livermore Laboratory, University of California-Berkeley, were incorporated into the reference bibliography. Additional pertinent geologic and geophysical data were provided by Southland Royalty Company. The literature search was initiated during Stage I of the investigation and continued throughout the program to incorporate pertinent new articles.

3.1.2.1 Low-Sun-Angle Photography

Six low-altitude aerial missions were flown under low-sun conditions from March to August, 1979, to generate LSA photographic coverage of the study area. A turbo-charged Cessna 207, designed for in-cabin mounting of the aerial camera and viewfinder, was chartered for each mission. The camera used for this program was a K17 military camera with 9x9 format and 6-inch focal length high speed lens actuated by a mechanical switch. The camera was manually leveled and timing of photo intervals was accomplished with an auto-reset stopwatch.

Navigation for the photography was by in-flight guidance from

1:62,500 scale topographic maps of the study area. Photos were flown along flight lines subparallel to the regional structure of N31E - S31W. Flight altitude approximated 6000 feet above mean terrain to yield photography with an approximate 1:12,000 scale. Flight lines were spaced at 1 5/16 miles to yield 20 percent side lap with photo intervals timed to yield 60 percent forelap for adequate stereoscopic images.

Kodak Tri-X (#2403) high speed black and white aerographic film was used for four missions, and Kodak Double-X (#2405) black and white aerographic films was used for two missions. The film was commercially processed and printed by America Aerial Surveys of Sacramento, California. A total of 1,324 images of good to excellent quality were obtained for the study region; an index of these images is incorporated on Plate II.

Flight times to obtain optimum results from LSA photography are defined by Slemmons (1969) as the two-hour intervals beginning one-half hour after sunrise and ending one-half hour before sunset. These optimum time intervals were used for the LSA program in Dixie Valley.

Seasonal variations in sun azimuth indicated an early spring program for optimum LSA imagery with respect to the regional trends in Dixie Valley (Slemmons, 1969; Glass and Slemmons, 1978; Slemmons, pers. commun., 1979). Approximately 90 percent of the imagery was obtained in early March under optimal conditions, with follow-up flights conducted in June and August.

Slemmons (1969) and Glass and Slemmons (1978) summarize the advantage of using highlighting techniques versus shadowing techniques for structural enhancement. The latter was used where possible in the program to detect very subtle structural features.

Interpretation of all stereoscopic images (LSA, NASA and AMS imagery) was accomplished with a Wild ST-4 stereoscope. Structural and tectonic photographic interpretations were delineated on mylar overlays on alternate images and then transferred to preliminary 7½-minute topographic quadrangle sheets (1:24,000 scale) with the aid of a Bausch and Lomb Zoom Stereo Transfer Scope. These map sheets were reduced by Photo-Mechanical Transfer (PMT) and assembled into a composite map (Plate II). Field verification of structural interpretations was conducted during the summer and fall of 1979.

3.1.2.3 Snow-lapse Photography

Snow-lapse photography planned for the winter months of late 1978 and 1979 was not accomplished. Major regional snowstorms occurred on December 19 and 20, 1978, and on February 22 and 23, 1979. Aerial reconnaissance flights immediately following the two storms (December 20, 1978, and February 23, 1979) verified the lack of adequate snow cover in Dixie Valley to accomplish a snow-lapse study.

The frontispiece, a Landsat composite color image of west-central Nevada, shows the regional extent of the storm of December 19 and 20, 1978. Dixie Valley, in the center portion of the image, is anomalously devoid of snow cover. This situation which recurred throughout the winter months, may be the result of local climatic influences, or a valley-wide 'hot spot' effect reflecting the underlying geothermal system.

3.1.2.4 Surficial Geology

Generalized surficial geology of northern Dixie Valley was delineated by interpretation of black and white 1:62,500 scale AMS photographs, with refinements based upon the 1:12,000 scale LSA photography. Because the various surficial geologic units correspond with specific geomorphic surfaces that also indicate genetic origin of the units, the surficial geology is presented as a generalized geomorphic surfaces map (Plate III). Surface and near-surface samples of the various units were collected for grain-size analysis; the results are presented in Appendix A.

3.1.2.5 Fault Scarp Morphology

Fault scarp morphology studies were conducted within the study area on scarps delineated on LSA photography. The method of Wallace (1977) was used to profile scarps by laying a stadia rod directly on the scarp and measuring its inclination. Scarps of moderate to high relief were chosen for profiling; graphic presentations of these profiles are presented in Appendix B.

3.1.3 Previous Work

Numerous authors have conducted studies within the Dixie Valley region. However, only those papers in the various critical disciplines that contributed most to the present study are listed here. Active faulting in Dixie Valley and nearby areas was studied by Larson (1957),

Slemmons (1957a,b), Burke (1967), and Ryal and Malone (1971). Regional studies were conducted by Stewart (1971), Thompson and Burke (1974), Wright (1976), and Speed (1976). Geologic maps of the Dixie Valley region were prepared by Page (1965), Willden and Speed (1974), and Johnson (1977). Geophysical data of Smith (1967, 1968), Meister (1967), Thompson and others (1967), Thompson and Burke (1974), Exploration Data Consultants (1976), Senturion Sciences (1977, 1978), and Keplinger and Associates, Inc. (1978) were integrated into this report.

3.1.4 Acknowledgements

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3.2 ANALYTICAL RESULTS

3.2.1 Geomorphic Setting and Surficial Geology

A generalized geomorphic map of the study area is presented on Plate III. The area is bounded by the Stillwater Range to the northwest, the Clan Alpine and Augusta Mountains to the southeast, Jersey and Pleasant Valleys to the northeast, the southernmost extent of the Humboldt Salt Marsh to the southeast.

Five distinct geomorphic surfaces were delineated by photogeologic interpretation (Plate III): (1) active alluvial fan surfaces, (2) inactive or highly dissected alluvial fan surfaces, (3) surfaces developed

by lacustrine processes, (4) composite or undifferentiated surfaces produced by aeolian, alluvial, and lacustrine processes, and (5) the surface of the presently active playa. These surfaces represent distinct geologic and geomorphic environments (including paleo-environments) and can be correlated with distinct surficial geologic units that are assumed to be present at depth but with lateral and vertical facies changes.

The active alluvial fans are eroded bedrock material derived from nearby mountain ranges and consist of clay-, silt-, sand-, and cobble- to boulder-size material of the varying lithologies present in the mountain ranges. Appendix A contains grain-size distribution curves for representative samples of the active fans and other surficial deposits. In general, coarser and heavier material would be preferentially deposited near the apex of a fan with finer and lighter materials being carried further down the apron (Ritter, 1979; Hooke, 1967, 1968).

The older alluvial fan deposits are similar to those of the active fans, but the older surfaces are primarily undergoing degradation. These surfaces are highly dissected and surface cobbles are coated with desert varnish.

The landforms produced by lacustrine deposition include deltas, strand lines, and longshore bars. The strand lines are conspicuous sandy deposits presently being subjected to subaerial erosion. The delta deposits and longshore bars are relatively stable and consist of well sorted cobbles, pebbles and silt with little or no primary clay or sand. The cobbles form large patches of desert pavement and are coated with desert varnish. Carbonate cobbles on the surface exhibit flutes from aeolian processes. Locally longshore bars have dammed pre-lacustrine drainages resulting in the ponding of fine-grained (clay- and silt-size) alluvium on the upslope side. Many of these dams have been breached by more recent fluvial activity.

The composite geomorphic surfaces include valley fill deposits derived from three distinct genetic processes. These undifferentiated deposits are predominantly silt with major amounts of clay and minor amounts of sand and gravel. This composition suggests two presently active processes with a third inactive. Active aeolian and alluvial processes are depositing predominantly finer particles. The scattered sand grains are possibly aeolian with the finer silt and clay being the outwash from active alluvial fans. The occasional cobbles are assumed

to be the result of flash flooding which has carried larger material beyond the active alluvial fan surfaces. Much of the silt is a result of inactive lacustrine processes and thus, may be Lahontan in age.

The active playa deposits are composed of fine sand, silt and clay-sized particles and salt crystals (predominantly sodium chloride). The center of the presently active playa is marshy with active deposition of sodium chloride. The water level fluctuates and, with the extreme low relief of the playa, the surface extent of the marsh varies greatly. Surrounding the marsh is a zone composed of clay and silt with salt crystals emplaced by littoral action as the water level fluctuates. Three sides of the high water area are bounded by conspicuous "beach berms" which may result from the capture of aeolian material by salt cementation as salt water is carried to these areas by capillary action from the marsh and evaporated. Locally sparse vegetation is present near the perimeter of the active playa with conspicuous mounds of aeolian material, mostly silt, built up around isolated bushes. Some areas of the playa are sparsely populated by salt grasses.

3.2.2 Structural-Tectonic Features

The structural-tectonic features map (Plate II) is a composite of features delineated by photogeologic interpretation of previously available imagery and the LSA imagery generated by this study. Fault zones, generalized as a single line, are shown on Plate IV. Interpretation of features critical to the structural analysis have been verified in the field. The structural-tectonic features delineated by surface expression correspond with subsurface structures delineated by gravity, aeromagnetic, and other geophysical surveys (Meister, 1967; Smith, 1968; Stewart, 1971; Senturion Sciences, 1977, 1978; Koenig and others, 1976; Keplinger and Associates, 1978) (Figure 3-1).

3.2.2.1 Old Stillwater Fault

This range-front fault zone bounds the Stillwater Range to the southeast and trends N36E from the southern edge of the study area near Dixie Meadows into Pleasant Valley on the northern edge of the study area. The zone is marked at the surface by very fresh appearing scarps in alluvium and bedrock, alignment of fumaroles, vegetation and tonal contrasts, back-facing grabens, breccia slumps, and slumps. Geomorphic

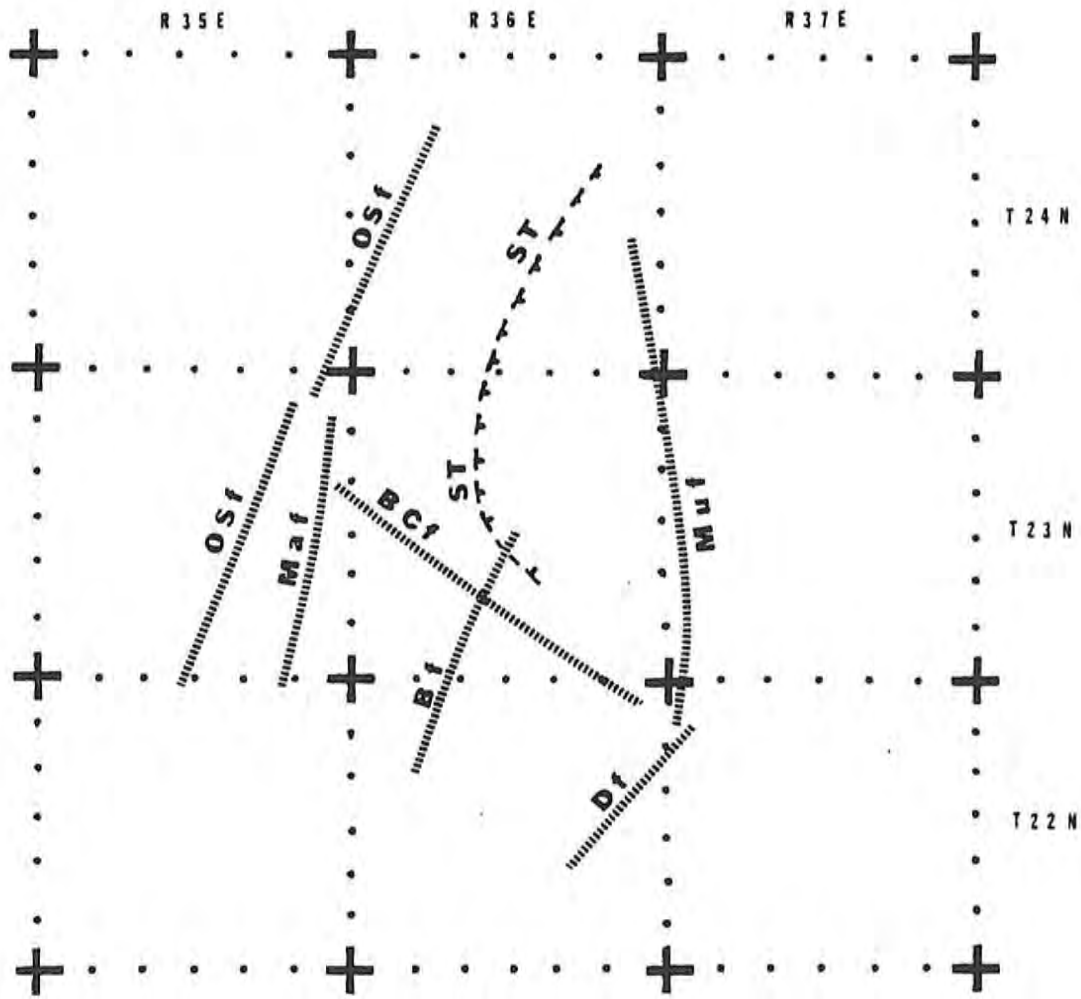


Figure 3-1. Diagrammatic sketch of major structural elements defined by aeromagnetic data interpretation by Senturion Sciences (1977, 1978). Bf - Buckbrush fault; Df - Dyer fault; Maf - Marsh fault; Muf - Mud fault; OSf - Old Stillwater fault; ST - Stillwater Thrust fault (?); BCf - Bernice Creek fault.

features indicate this zone is very active and include 'V' shaped valleys and steep faceted spurs with little or no dissection (Glass and Slemmons, 1978) and fault scarps in alluvium with a very high slope angle or even free face (Wallace, 1979). Slemmons (1966) indicates the southern portion of this zone ruptured in the 1954 Dixie Valley earthquake as far north as the Boyer Ranch. Mrs. Sheldon Lamb of the Boyer Ranch (verbal commun., 1979) reports that features near the Boyer Ranch were liquefaction phenomena and the actual fault surface rupture extended only as far north as Dixie Meadows.

Field measurements of a southeast dip of 50 to 60 degrees for this zone are consistent with the literature (Slemmons, 1957a; Larson, 1957; Page, 1965; Burke, 1967; Meister, 1967; Ryall and Malone, 1971; Stewart, 1971). Site-specific evidence for the southeast dip includes: (1) measurable fault planes on bedrock surfaces, (2) surficial geomorphic expression of normal faulting including back-facing grabens, and (3) fault plane solutions for a N55W tensional (Sigma 1) direction in northern Dixie Valley (Ryall and Malone, 1971). First order Riedel shears (Tchalenko, 1970) indicate the fault zone has a minor right-slip component. Regional studies also indicate extension of the study area in a general N55W direction (Thompson and Burke, 1973; Wright, 1976). This data is incompatible with the interpretation of aeromagnetic data by Senturion Sciences (1977, 1978) indicating a N55W dip for this zone.

The maximum elevation of the Stillwater Range to the northwest of the fault zone is greater than 7200 feet and alluvial fill in the valley to the southeast of the fault zone is nearly 2000 feet below the present valley elevation indicating greater than 6000 feet total vertical displacement along this zone.

3.2.2.2 Marsh Fault

The Marsh fault extends from the southern edge of the study area along a N30E trend to intersection with the Old Stillwater fault in Section 36, T24N, R35E. The surface expression, although somewhat subdued along the northwest edge of the Humboldt Salt Marsh, is marked by stream and vegetation alignments, subtle scarps and grabens. Scarps preserved in Sections 6 and 7, T23N, R35E, and Sections 10 and 16, T22N, R35E, indicate normal faulting with the southeast block downdropped. This is compatible with aeromagnetic interpretations (Senturion Sciences, 1977,

1978).

3.2.2.3 Buckbrush Fault

The Buckbrush fault is a semi-arcuate fault concave to the southwest trending N30E from the southern edge of the study to the Bernice fault when it splays into two branches. Both splays intersect the Old Stillwater fault: the western splay in Section 4, T24N, R36E, and the eastern splay near the Senator fumeroles in Sections 28 and 29, T25N, R37E. Surface expression is marked by scarps, spring and vegetation alignments, and vegetation and tonal contrasts. The northwest block is downdropped in the southern portion of the zone but relative movement is not discernible in the northern portion.

3.2.2.4 Mud Fault

The Mud fault trends N36E from the Dixie Settlement area to Section 4, T23N, R35E, then follows a N15E trend to Sou Hills. Surface expression of the southern portion indicates the northwest block is downdropped. Near Sou Hills several interpretations are possible: (1) the fault joins the frontal fault system on the east side of Pleasant Valley (Pleasant Valley fault), (2) the fault joins the Old Stillwater fault system and continues along the west side of Pleasant Valley, or (3) the fault is related to the frontal fault system on the east side of the Tobin Range (Tobin fault).

A southeast splay of this fault, mapped by Secturion Sciences (1977, 1978) as the Mud fault, shows surface expression as moderate scarps near the mouth of Deep Canyon and trends N25E to intersection with the main zone in Section 30, T23N, R37E. However, this splay could be related to the Shoshone fault with the Mud fault being a separate and distinct structure. In either interpretation, the block northwest of this splay is downdropped.

3.2.2.5 Dyer Fault

The Dyer fault transects the study area along a N36E trend from the mouth of Meadow Spring Canyon on the south and is the frontal fault of the Clan Alpine and Augusta Mountains. The northwest block is downdropped along this zone. A splay of this fault appears to arc across Dixie Valley just north of Hole-in-the-Wall and join the frontal fault

system on the west side of the Tobin Range in Pleasant Valley (Pleasant Valley fault).

The Clan Alpine Mountains reach a height in excess of 8,800 feet, with alluvial fill of unknown thickness below the 3600 feet mean elevation valley floor, which indicates greater than 5000 feet total vertical displacement along this zone.

3.2.2.6 Bernice Fault

The Bernice fault is a main crosscutting feature in northern Dixie Valley. Surface expression of vegetation and tonal contrasts and stream and spring alignments indicate the zone extends along an arcuate trace northwest from Bob Canyon toward the Dixie site in the Stillwater Range. A splay of this zone may arc to the north and join the Marsh fault near Section 17, T23N, R36E. Surficial geomorphic deposits indicate the southwest block is downdropped as the lowest area of the valley floor; the salt marsh is immediately southwest of the fault trace. However bedrock geology near the Dixie Site indicates the northeast block may be downdropped, as does aeromagnetic data (Senturion Sciences, 1977, 1978) Surface expression is subtle and relative movement on the zone is not discernible from surface evidence.

3.2.2.7 Stillwater Thrust

Senturion Sciences (1977, 1978) postulated a thrust fault in the center of the study area based on interpretation of aeromagnetic data. No surface expression of a thrust fault was discerned by photogeologic interpretation.

3.2.2.8 Mississippi Fault

Another fault zone (Mississippi fault) extends across Dixie Valley along a S40E trend from the mouth of Mississippi Canyon, crosscutting the valley. Surface expression includes low scarps, vegetation and tonal contrasts, and stream alignments. These features are subtle but indicate the southwest block is downdropped. Surficial geomorphic deposits indicate the opposite may be true as the Humboldt Salt Marsh lies immediately northeast of this zone.

3.2.2.9 Dixie Meadows Fault

A major fault zone (Dixie Meadows fault) trends N36E from the southern edge of the study area across Dixie Meadows, sub-parallel to and between the Old Stillwater and Marsh faults. This Dixie Meadows fault corresponds to the valley branch splay of the Stillwater Range frontal fault of Trexler and others (1978). The zone continues in an arcuate path concave to the northwest and merges with the Old Stillwater fault in Sections 35 and 36, T24N, R35E. The fault zone apparently re-emerges from the Old Stillwater fault zone just northeast of this area and again forms an arcuate zone which intersects the frontal fault system in Section 14, T24N, R36E. The southwest block is downdropped as evidenced by numerous slumps, asymmetric grabens, and southeast facing scarps. Spring and vegetation alignments are also present. Liquefaction phenomena, including slumps and sand boils, are present along the portions of this fault that lie near highly saturated playa deposits. Slemmons (1957a) has mapped surface rupture on this zone during the 1954 Dixie Valley - Fairview Peak earthquake in Dixie Meadows.

3.2.2.10 Shoshone Fault

A fault (the Shoshone fault) northwest of and sub-parallel to the Dyer fault trends N30E with maximum surface expression near the mouth of Shoshone Creek. In addition to vegetation contrasts, surface expression is marked by scarps which indicate the block to the northwest is downdropped. Hyder Hot Springs may be associated with this structure.

3.2.2.11 Pleasant Valley Fault

The Pleasant Valley fault is the frontal fault system on the west side of the Tobin Range. The fault zone is wide but activity in 1915 created a very sharp southwest facing scarp from surface rupture near the bedrock-alluvium contact (Slemmons, 1966). This zone crosses the Sou Hills to the southwest and apparently ends in the northern portion of Dixie Valley.

3.2.2.12 Tobin Fault

The frontal fault zone on the east side of the Tobin Range (Tobin fault) extends southwest into the study area and apparently dies out in

the north end of Dixie Valley. A splay of this fault may cross to the southeast of Sou Hills and merge with the Old Stillwater fault. Scarps along this zone indicate the southeast block is downdropped. Sou and Seven Devils Hot Springs may be associated with this structure.

3.2.2.13 White Rock Canyon Fault

A fault (White Rock Canyon fault) traces diagonally north-south across the Stillwater Range, following White Rock Canyon in the study area. This fault may join the Old Stillwater fault immediately southwest of the mouth of White Rock Canyon and trend south, splaying from the Old Stillwater fault in Section 23 of T23N, R35E. Surface expression for this zone includes vegetation and tonal contrasts and lineations in Dixie Valley, displaced bedrock in White Rock Canyon and, regionally, the apparent left-lateral displacement of the Humboldt gabbroic complex (Humboldt lopolith of Speed, 1976) (Figure 3-2).

The displacement shown on the Humboldt complex contrasts with that proposed by Smith (1968). From aeromagnetic data Smith (1968) interpreted about 1 mile of right lateral displacement of the eastern portion of the complex on what corresponds to the Buckbrush fault in this report. A review of this data indicates a noticeable anomaly which Smith interpreted as the southern edge of the complex beneath alluvium in Dixie Valley. This anomaly may instead correspond to a downdropped southwest block on the Mississippi fault. Another anomaly crossing the valley about 11 miles to the northeast may represent the southern boundary of the Humboldt complex in Dixie Valley (Figures 3-3 and 3-4). This interpretation is consistent with outcrops of the gabbroic complex in the Clan Alpine Mountains and Stillwater Range. Figures 3-5 and 3-6 depict the aeromagnetic data of Smith (1968) re-interpreted to include the left-lateral White Rock Canyon fault.

3.2.3 Fault Scarp Morphology

A preliminary fault scarp morphology study was initiated along the Old Stillwater and Tobin faults, primarily to discern number of movements and amount of displacement on scarps in alluvium near White Rock Canyon and Sou Hot Springs. The methods of Wallace (1977) were used. Scarp profiles for these traverses are presented in Appendix B.

Profiles 1 through 5 are on the Old Stillwater fault near White

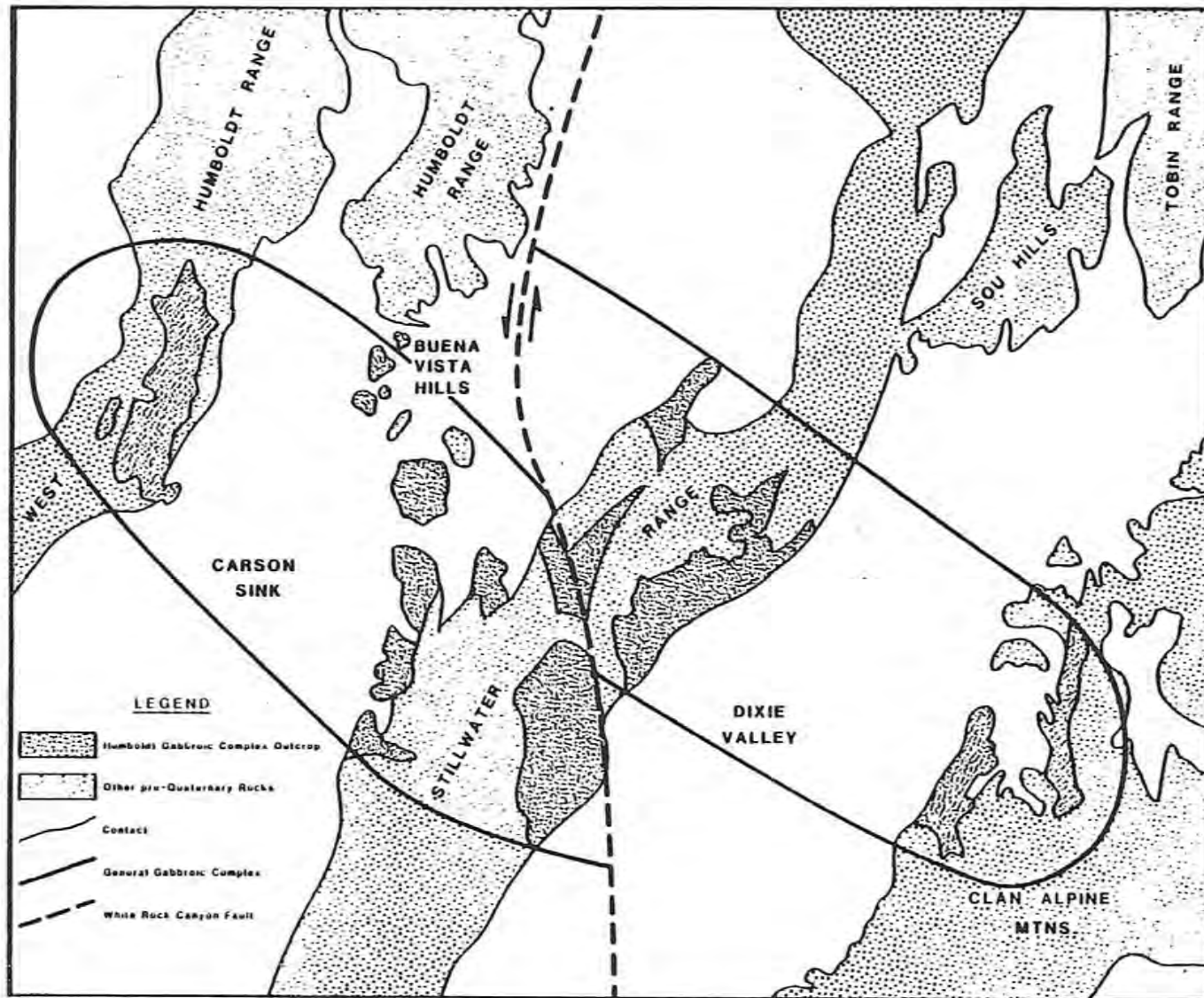


Figure 3-2. Generalized geologic map showing extent of Humboldt gabbroic complex (Humboldt Lopolith of Speed, 1976)

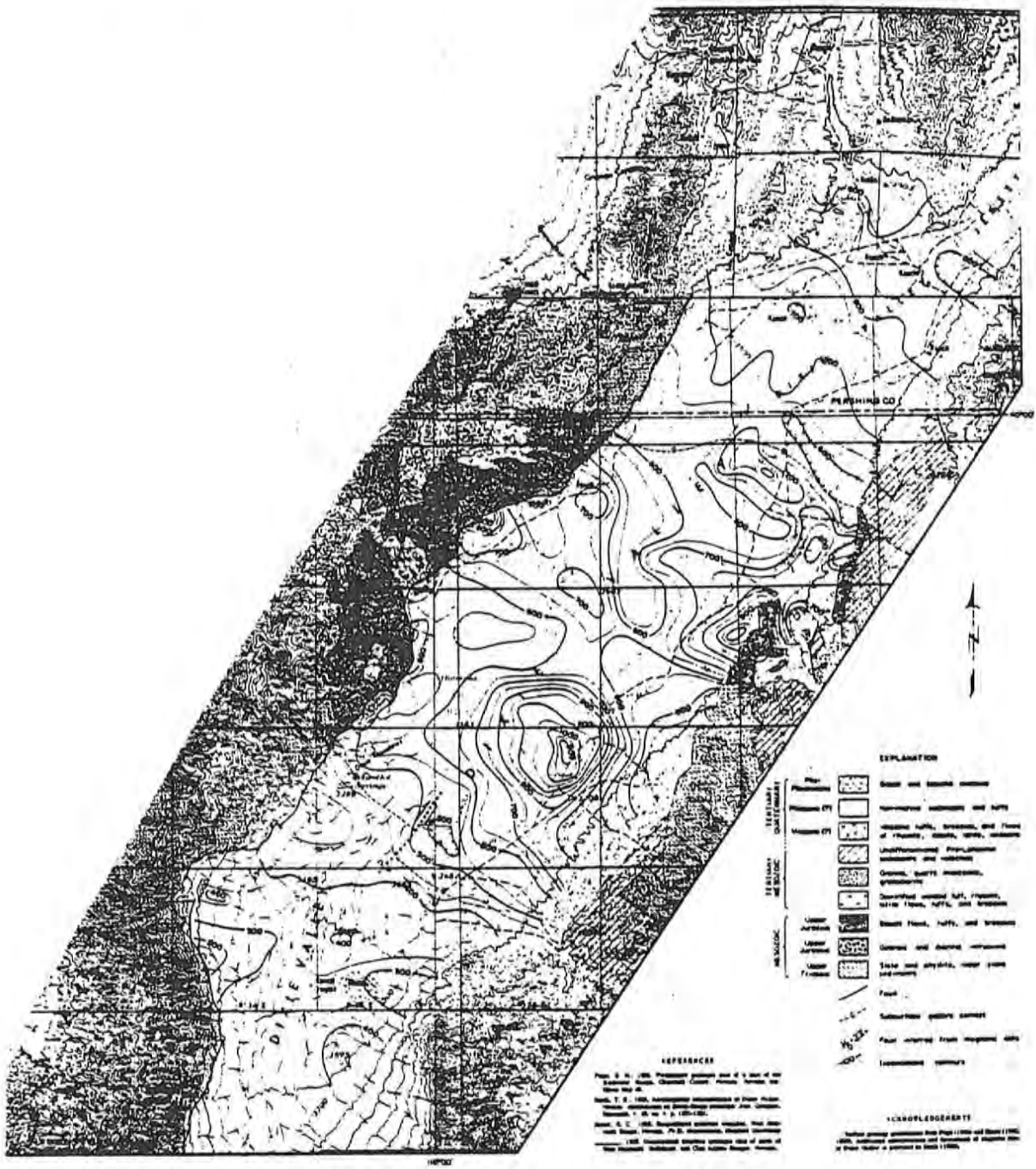


Figure 3-3. Aeromagnetic map of Dixie Valley. Approximately one to two miles of right-lateral offset is interpreted across the southern boundary of the gabbro by Smith (1971).



Figure 3-4. Second vertical derivative of aeromagnetic data for Dixie Valley. (from Smith, 1971)

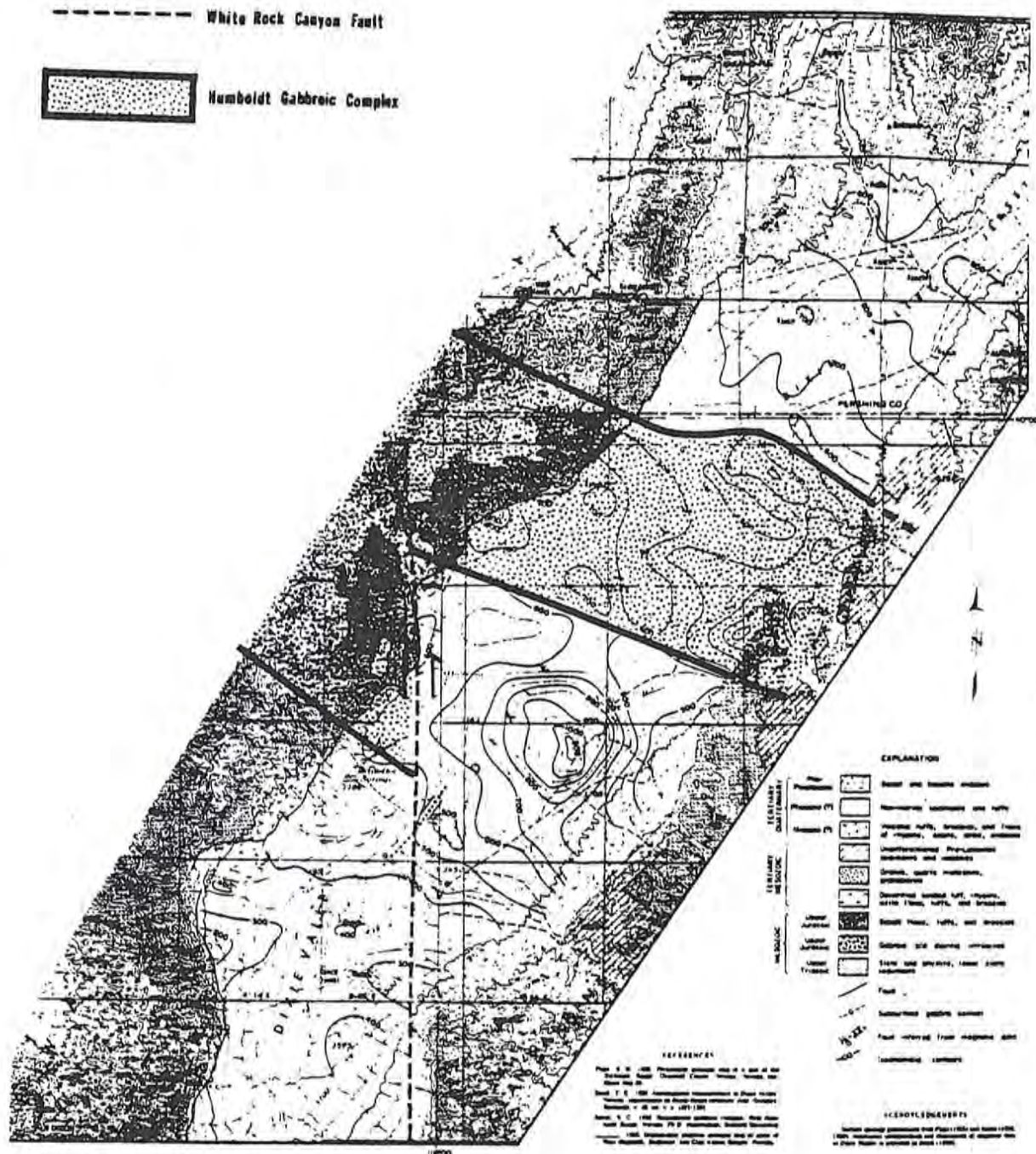


Figure 3-5. Re-interpretation of aeromagnetic map of Dixie Valley showing the left-lateral White Rock Canyon fault and the proposed boundaries of the Humboldt gabbroic complex. (aeromagnetic data from Smith, 1971)

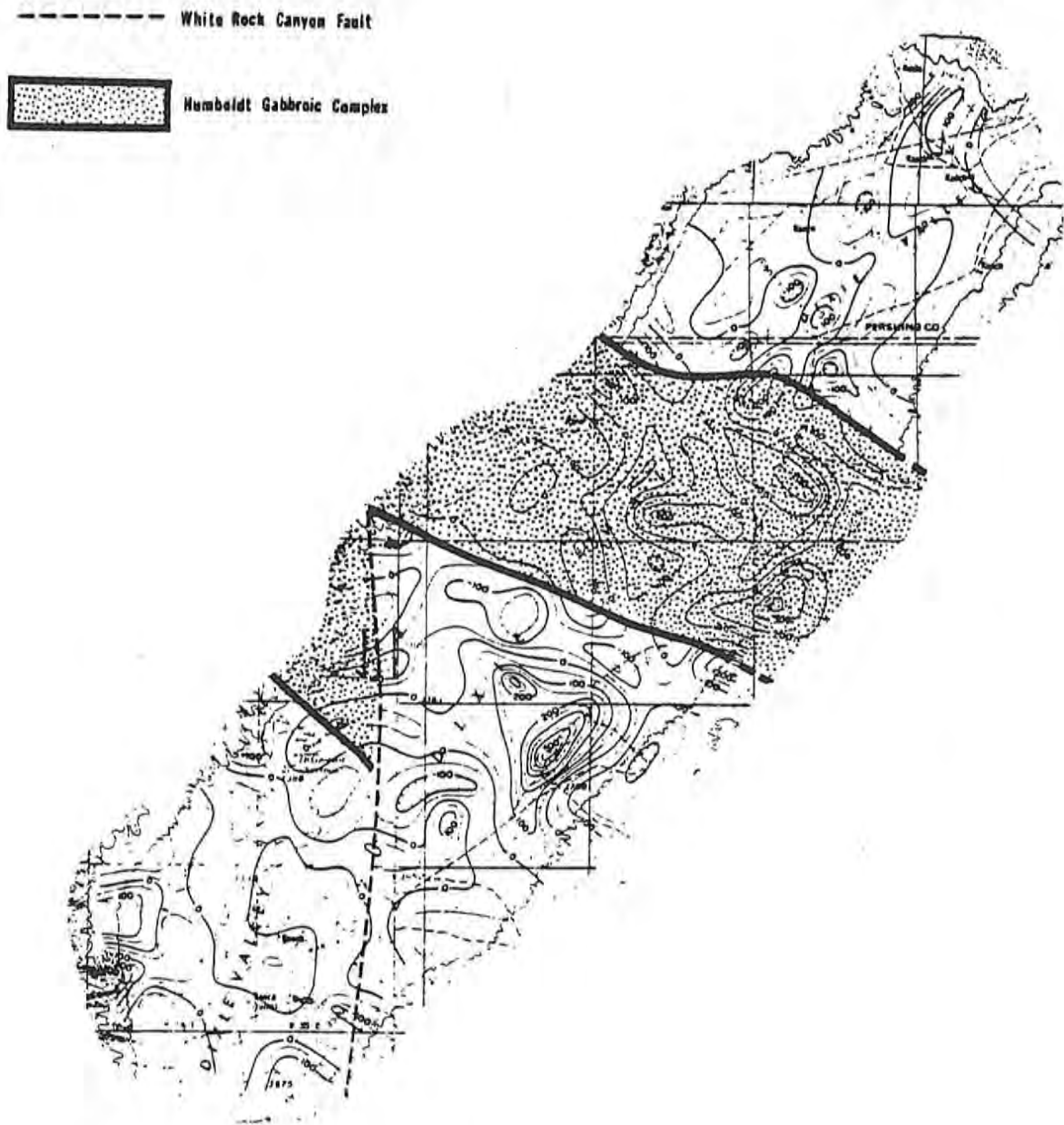


Figure 3-6. Interpretation of the second vertical derivative of the aeromagnetic data of Smith (1971) showing the left-lateral White Rock Canyon fault and the proposed boundaries of the Humboldt gabbroic complex.

Rock Canyon (Plate IV). The profiles show only one movement has occurred in this scarp with offset averaging 10.7 feet. This offset does not include a 15.5 foot offset on one traverse (#4) as this data may be invalid; the scarp profile at #4 may have been altered by lateral erosion at the scarp base or affected by liquefaction.

A scarp profile was measured on the Tobin fault (#6) to attempt to determine if this fault differs from the scarp observed near White Rock Canyon. The profile reveals three distinct movements of this fault, with displacement ranging from 3.5 to 10.9 feet.

This suggests that the Tobin and Stillwater faults represent different tectonic regimes with the Tobin fault more active in Quaternary and Recent time. The steeper faceted spurs on the Stillwater fault scarp indicate greater activity in the late Cenozoic however.

3.2.4 Bedrock Geology

The following discussion of the major geologic units and structure within the study area is synthesized from Page (1965), Willden and Speed (1974), Speed (1976) and Johnson (1977).

3.2.4.1 Triassic Rocks

The most important rock unit consists of upper Triassic metasediments, mostly phyllite and slate with about 10% metaquartzite and limestone. This complexly deformed unit is estimated to be 5000 to 10,000 feet thick, with the base unexposed. Locally this unit is overthrust by an upper Triassic massive limestone.

3.2.4.2 Jurassic Rocks

The Jurassic sequence of altered and locally schistic metavolcanics consist chiefly of fine-grained slaty andesite tuffs, breccias, andesite flows with minor quartzites and calcareous sandstone. This sequence is in excess of 5000 feet thick. The base is in thrust contact with the Triassic limestone.

The Humboldt gabbroic complex (Speed, 1976) outcrops in large areas of the Stillwater Range and Clan Alpine Mountains. The intrusive phase of this complex is gabbroic to dioritic in composition. Extrusive units include basalt flows, tuffs and breccias.

Jurassic and older rocks are metamorphosed to the greenschist fac-

ies.

The Boyer thrust (Fencemaker thrust of Speed, 1976) has transported quartzites of Jurassic age over lower Jurassic units with emplacement of the Humboldt complex as the propelling mechanism. Age dating of the gabbroic complex (Speed and Armstrong, 1971) at 151 to 156 my also dates the thrusting.

3.2.4.3 Jurassic to Miocene Rocks

Four sequences of volcanic units from 2000 to 10,000 feet thick include acidic tuffs and breccias and basaltic andesite flows. The lower units are intruded by granitic plutons of late Cretaceous age. The episode of acidic intrusion accentuated and coarsened the metamorphic fabric of the invaded units, inducing chiastolite growth and the formation of chloritoid porphyroblasts. Locally marble and skarn were produced. Most intrusions produced contact aureoles of black hornfels or chiastolite bearing phyllite.

3.2.4.4 Miocene Rocks

A widespread Miocene volcanic sequence, up to 4000 feet thick, includes tuffs, breccias and flows which vary in composition from latite through rhyolite. Riehle and others (1972) have located this Miocene volcanic center in the southern Clan Alpine Mountains.

3.2.4.5 Late Cenozoic Deposits

Plio-Pleistocene to Recent deposits include sediments of lacustrine and fluvial origin, locally interbedded with acidic ashes, tuffs, and flows. Pliocene basalt flows, aggregated up to 1600 feet thick, are present in the lower portions of the sequence. Thicknesses of greater than 5000 feet are postulated in Dixie Valley based on aeromagnetic data (Smith, 1968).

3.3 Interpretations and Conclusions

3.3.1 Geomorphic Surfaces

A distinct relationship is seen between the geographic position of geomorphic surfaces and the structural-tectonic features in Dixie Valley. In most obvious of these relationships, the active playa surface and the

Humboldt Salt Marsh are bounded on all four sides by structural features (Plate IV). On the northwest side of the playa the Marsh fault marks the boundary zone while to the southeast the boundary is marked by the Buckbrush fault. The two crosscutting features of Dixie Valley --the Bernice and Mississippi faults--lie close to the northeast and southeast boundaries of the Salt Marsh. The boundary of the active playa on the northeast extends over the Bernice fault and corresponds generally with a splay of the Buckbrush fault where it transects the valley. These studies place the active inner-graben to the northwest of the longitudinal axis of Dixie Valley and the salt marsh is displaced in the same manner with respect to the active playa surface.

The alluvial, lacustrine and aeolian deposits included in the undifferentiated valley fill are located northeast and southwest of the active playa surface. These sediments extend northward into both Pleasant and Jersey valleys, and southward beyond the study area. Like the playa deposits, these sediments are shifted northwest of the longitudinal valley axis.

Active alluvial fan surfaces extend from bedrock on both sides of the valley to the playa or valley fill deposits. These deposits are asymmetrically distributed as a narrow band (up to 4 miles wide) of steeply sloping fan surfaces extending southeast from the Stillwater Range, and a wide band (3 to 7 miles wide) of gently sloping surfaces extending northwest from the Clan Alpine and Augusta Mountains. A third narrow band encircles the southern end of the Tobin Range. The wider band of coalesced fan surfaces (bajada) on the east side of the valley are embayed deeply into the source mountain ranges but those on the west side extend only a short distance into canyons of the Stillwater Range.

The older alluvial deposits are nearly absent on the western, more active side of the valley but are more common to the north and east where they generally adjoin bedrock.

The lacustrine deposits, including deltas, strand lines, and long-shore bars, are predominantly near the playa and some distance from bedrock on the eastern and northern portions of the valley where they overlie alluvium and valley fill deposits. The scattered small deltas on the west side of the valley overlie fan surfaces only and generally adjoin bedrock.

The valley fill, older alluvial fans, and lacustrine deposits are

all transected by active alluvial surfaces where major drainages have entrenched through these areas.

Boundaries between the geomorphic surface units are generally sharp with the exceptions of the facies changes between the valley fill and playa and the valley fill and active alluvial fans.

3.3.2 Structural-Tectonic Features

A three dimensional model of the northern portion of Dixie Valley depicts the structural relationships among the various tectonic elements, with the alluvium removed and bedrock surfaces restored (Figure 3-7). This modified interpretation incorporates all fault zones described in Section 3.2.2 with the exception of the postulated Stillwater thrust fault (Senturion Sciences, 1977, 1978). Although the relative movement on the Bernice and Mississippi faults is uncertain, a downthrown southwestern block is depicted in the model.

A mechanism for the formation of graben valleys similar to Dixie Valley is reviewed by Stewart (1971). Tchalenko (1970) and Cloos (1955, 1968) have assumed laboratory clay models of fault and/or deformation systems mimic geological systems. A model developed by Cloos (1968) for Basin and Range horst and graben formation (Figure 3-8) closely resembles the three dimensional model presented in Figure 3-7 for Dixie Valley.

The longitudinal asymmetry of Dixie Valley also fits well with laboratory models developed by Cloos (1968). The downdropping of one side of the graben in the models corresponds with structural interpretation along the Stillwater margin of the Valley. Slump block faulting of the model closely resembles interpretation along the Clan Alpine and Augusta Mountains. A distinct arcuate crosscutting feature developed in the model may correlate with the arcuate crosscutting splays of the Buckbrush and Dyer faults.

The graben model postulated above would suggest more active (i.e., greater amounts of) subsidence in a zone along the western side of the valley. Alluvial fans (and other deposits) in this zone would tend to be covered more rapidly by playa deposition on this downdropping block. Conversely, older alluvial fans, older delta deposits, and a much wider zone of active alluvial deposition would be expected on the surface of the more brecciated side of the graben where the total displacement is distributed across a series of step faults. This is consistent with the

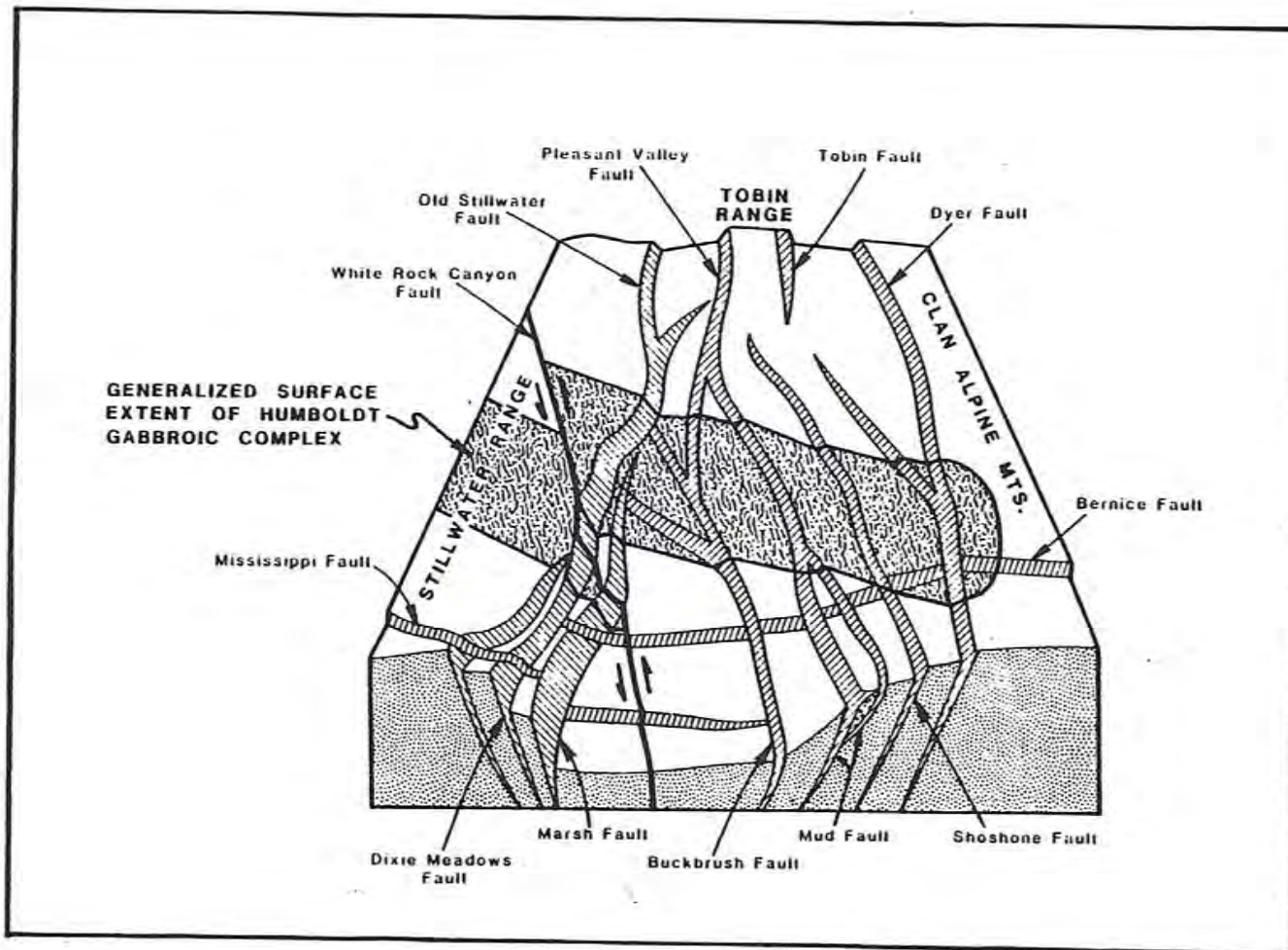


Figure 3-7. Three dimensional model of the northern portion of Dixie Valley. The structural relationships among the various tectonic elements are depicted, with alluvium removed and bedrock surfaces restored.



Figure 3-8. Clay model of graben formation showing assymetry of the bounding faults. Note arcuate cross-cutting structure. Right block moved right; left block remained stationary. Reprinted from Cloos (1968) with the permission of the American Association of Petroleum Geologists.

distribution of the various geomorphic surfaces within northern Dixie Valley (Plates III and IV).

The above model does not explain crosscutting features which extend into bedrock such as the Mississippi, Bernice, and White Rock Canyon faults. These are inferred to be older faults which predate the inception of Basin and Range faulting (15 to 17 m.y. ago; Stewart, 1971) but are still active. Geomorphic features along these fault zones include subtle scarps, spring alignments and vegetation contrasts indicating late Quaternary, possibly Recent, movement.

Gravity data (Stewart, 1971; Erwin and Bittleston, 1977; Erwin and Berg, 1977) are consistent with the model of a complex asymmetric graben. Structural relationships defined by aeromagnetic surveys (Smith, 1971; Nevada Bureau of Mines and Geology, 1977; Senturion Sciences, 1977, 1978) are in close agreement with clay model experimental data (Cloos, 1955, 1968) and the structural-tectonic features delineated by surficial expression in Dixie Valley (Plate II).

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Chapter 4. PETROLOGIC ALTERATION STUDIES

By: Thomas R. Bard

terized by high clay content, fracturing, brecciation, veining, very fine drill cuttings, visibly altered minerals, or other anomalous features.

4.1.2.1 Petrographic Analysis

The second step consists of making epoxy impregnated grain-mount thin sections representing both specific ten-foot intervals in anomalous zones and intervals of one hundred feet to obtain overall distribution patterns. One-hundred foot composite samples are prepared by combining five original samples. Analyzing composite samples helps minimize potential random sampling biases of the original samples and reduces the number of samples which must be prepared and analyzed to a practical total while maintaining adequate spatial resolution for defining most types of mineralogical distribution patterns. While the examination of one-hundred foot composite samples defines the overall distribution of mineral species, the ten-foot interval thin sections are useful in evaluating specific anomalous intervals.

A factor which may affect the validity of any interpretation regarding the relative mineral abundances and their distribution is the ability to obtain a sample representative of a one-hundred foot interval in a small thin section. Additionally, obtaining a representative sample of a particular depth is hindered by sample contamination and mixing caused by variations in circulation rates, settling velocities, and by caving. A representative sample is also a function of the technique used by the well logger to provide a split of the bulk sample. Moreover, during thin section preparation, certain minerals or rock types may be preferentially "plucked out", thereby decreasing their relative abundance.

Each thin section is examined under a polarizing microscope to identify and visually estimate the relative abundance of individual mineral species and to determine their inter-relationships. Plates V and VI show the distribution of selected minerals as a function of various parameters, including depth, lithology, temperature, and percent plagioclase alteration for DF 45-14 and DF 66-21, respectively.

4.1.2.3 X-ray Diffraction Analysis

The third step is the x-ray diffraction analysis identification

and interpretation of clay minerals present in the drill cuttings. This is the most useful method in identifying the distribution of alteration effects. Studies of the occurrence of various types of clay minerals are widely applied in the evaluation of hydrothermal alteration associated with geothermal systems (Steiner, 1968; Hoagland, 1976).

One-hundred foot composite samples are examined to provide an overall distribution of the clay minerals while specific ten-foot samples are taken every fifty feet to delineate and characterize specific anomalous intervals. Moreover, the data from the thin section analysis are more meaningful when interpreted in light of the x-ray diffraction data. Plates V and VI also show the relative abundance and distribution of the clay minerals for DF 45-14 and DF 66-21, respectively.

The samples selected for x-ray diffraction analysis were pulverized, slurried in distilled water and then allowed to settle for two hours. This process isolated the clay fraction of 8 phi (\emptyset) and smaller size (Folk, 1974). This suspended fine fraction is then pipeted onto glass slides and allowed to dry. This procedure tends to preferentially orient the platy phyllosilicate minerals which enhances their basal reflections and facilitates their identification. The samples are placed in a dessicator for at least 24 hours before analysis. Data was obtained on a Philips x-ray diffractometer using CuK α radiation at an operating voltage of 20 KV, a current of 15 MA, and a scanning speed of 2 $^{\circ}$ two-theta per minute. Each sample was analyzed from 2 $^{\circ}$ to 30 $^{\circ}$. Selected samples are heated to 550 $^{\circ}$ C for one-half hour and/or placed in an ethylene glycol saturated atmosphere at 75 $^{\circ}$ C for 24 hours and rerun for confirmation of certain clay minerals.

Clay minerals are identified from an examination of the basal x-ray reflections. Table 4-1 summarizes the parameters used for identification of the various clay minerals. Discrete illite was identified by its asymmetric d(001) peak at 10.0A $^{\circ}$ to 10.16A $^{\circ}$ which is unaffected by glycolation or heat treatment. Discrete montmorillonite was identified by d(100) = 12.25A $^{\circ}$ to 12.5A $^{\circ}$ or 14.2A $^{\circ}$ to 14.3A $^{\circ}$ in unglycolated samples, shifting to near 17A $^{\circ}$ upon glycolation. Na-montmorillonite is assumed at a d(001) of 12.25A $^{\circ}$ to 12.5A $^{\circ}$ and Ca or Mg-montmorillonite at a d(001) of 14.25A $^{\circ}$ (K. Papke, pers. commun., 1979). Randomly interstratified mixed-layer illite-montmorillonite was identified by the following properties:

Table 4-1. X-Ray Diffraction Data

<u>Mineral</u>	<u>d(001) A⁰</u>	<u>Glycolated d(001) A⁰</u>	<u>Heated for ½ Hour d(001) A⁰</u>
Illite	9.9 - 10.16	No Change	No Change
Na-Montmorillonite	12.25 - 12.4	16.8 - 17.2	9.8 - 10.1
Ca/Mg Montmorillonite	14.5 - 15.0	16.7 - 17.2	9.8 - 10.1
Randomly Interstratified Illite/Montmorillonite	10.16 - 14.5	Varies Depending on % of Expandable Layers	Collapses to 9.8 - 10.0
Regular Mixed-Layer Chlorite/Vermiculite	27.7 - 29.0	No Change	Collapses to Lower Value
Kaolinite	7.1 - 7.2	No Change	Peaks Disappear
Chlorite	13.9 - 14.2	No Change	Increase in Intensity

(1) The appearance of an asymmetrical, sometimes broad peak intermediate between 10.16\AA and 12.5\AA on diffraction patterns of unglycolated samples. This peak represents an average reflection from the (001) reflections of both illite and montmorillonite, and will be referred to as Ill (001)/Mo (001) (Brown and MacEwan, 1950; Weaver, 1956). The precise position of this peak will depend upon the relative amounts of the two layers.

(2) In glycolated samples the Ill (001)/Mo (001) reflection between 10.16\AA and 12.5\AA is shifted to a higher d-spacing value up to 17\AA .

The approximated percentage of expandable layers in the randomly interstratified illite-montmorillonite was estimated from the position of the glycolated Ill (001)/Mo (001) reflection, using the curve of Weaver (1956) (Figure 4-1) and the work of Steiner (1968) (Table 4-2).

Regular mixed-layer clays differ from random mixed-layer clays in that the (001) series is an integral sequence. That is, the (001) value of a regular mixed-layer clay is equal to the total thickness of the two or more types of layers which are present; for example, the $d(001)$ value of a regular mixed-layer chlorite (14.1\AA) and vermiculite (14.2\AA) would be 28.3\AA . Thus the observed appearance of a reflection in the vicinity of 28\AA in some samples must be due to two 14\AA clay minerals. As there is no observable shift in this 28\AA peak upon glycolation it is assumed to be a chlorite-vermiculite. Some of the diffraction patterns of these regular mixed-layer clays also indicate the presence of a randomly interstratified mixed-layer mineral (K. Papke, pers. commun., 1979).

Chlorite is identified by the presence of (001), (002), (003), and (004) reflections of 14.1\AA , 7.07\AA , 4.72\AA and 3.54\AA , respectively; all unaffected by glycolation. Kaolinite is distinguished by the appearance of (001) and (002) reflections at 7.16\AA and 3.57\AA , respectively. Heating kaolinite to 550°C for one-half hour destroys the internal structure and the reflections disappear. Difficulty was encountered in distinguishing chlorite from kaolinite due to an uncertain response of chlorite to heating that is mainly a function of grain size and crystallinity. The (001) reflection at 14.1\AA for chlorite will increase slightly after heat treatment, indicating an Fe-rich variety (Steiner, 1968). In addition, chlorite with strong (002) and

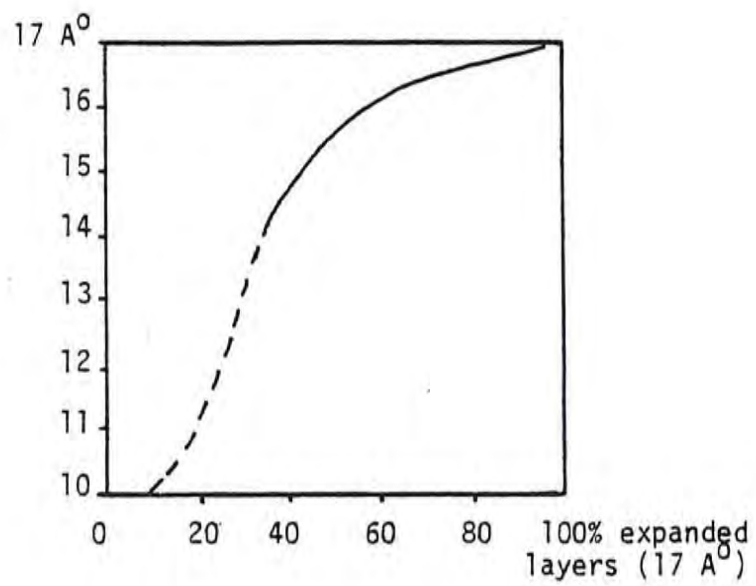


Figure 4-1. Curve showing migration of (001)/(001) peak of randomly interstratified 10 Å and 17 Å layers (Weaver, 1956).

Table 4-2. Montmorillonite Content of Micaceous Clay Minerals

(001)	(A ⁰)	Amount of Interstratified Montmorillonite Layers
Air-dried	Glycolated	(%)
12.45	14.97	> 40
11.78	13.60	< 35
11.63	13.39	> 30
11.33	13.19	> 30
11.05	12.98	30
11.19	12.77	< 30
10.91	11.94	25
10.78	11.33	> 20
10.52	11.19	20
10.40	10.91	< 20
10.40	10.65	> 15
10.28	10.16	< 10
10.16	10.16	0 or < 10
10.16	10.04	0 or < 10
10.16	9.93	0
10.04	10.04	0
9.93	9.93	0

Estimated proportions of interstratified montmorillonite in commonly observed micaceous clay minerals ranging from pure illite to randomly interstratified illite-montmorillonite with $d(001) = 12.45 \text{ \AA}$ (Steiner, 1968).

(004) reflections and relatively weak (001) and (003) reflections are assumed to be Fe-rich chlorites. The disappearance of certain reflections then is not conclusive evidence that only kaolinite is present. The appearance of the (002) kaolinite (004) chlorite double maxima at 3.57\AA , respectively, indicated the presence of both kaolinite and chlorite.

The above discussion shows that the identification of clay minerals is a difficult task. Certain factors inherent in x-ray analysis tend to bias results, including the structural characteristics of the mineral being analyzed, its degree of crystallinity, preferred orientation, and the ability to produce an internally consistent sample. All these factors will influence the observed peak intensity which is the parameter used to estimate relative abundances. Hence, the validity of the interpretations derived from these results must be viewed with discretion.

4.1.3 Previous Work

No previous petrologic and/or geochemical studies of the Dixie Valley Geothermal System are known to exist. It is believed that this study is the first of its kind to be conducted in this area. However, similar studies performed in other geothermal areas have provided useful guidelines and insights into the study of hydrothermal alteration in Dixie Valley. In particular, much use has been made of work done by Steiner (1968) in Wairakei, New Zealand, by Browne and Ellis (1970) at Ohaki-Broadlands, New Zealand, and by Hoagland (1976) in the Imperial Valley, California.

4.1.4 Acknowledgements

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4.2 Analytical Results

4.2.1 Shallow Thermal Gradient Holes

The results of the analysis of the six shallow thermal gradient holes (SR2, SR2A, S-8, DD-9, H-1, H-2) are presented in the November 1, 1979, report submitted to Southland Royalty Company and the U.S. DOE. The significant conclusions derived from the study are as follows:

(1) The propylitic mineral assemblage consisting of albite, calcite, quartz and clay is dominant in each of the drill holes.

(2) With the exception of S-8 and possibly H-1, none of the holes encountered bedrock material.

(3) The unconsolidated heterogeneous alluvial material is derived from a source area that was subjected to at least one episode of hydrothermal alteration prior to deposition.

(4) Zones of lost circulation encountered during drilling in some of the holes correlate with both more intense alteration and with the calculated depth of intersection with the range front fault.

(5) The distribution of clay minerals is the most useful indicator of alteration intensity.

(6) Alteration effects derived from the present hydrothermal system are only weakly developed and difficult to distinguish from previous regional alteration effects.

4.2.2 General Stratigraphy

Willden and Speed (1974) mapped and described seven rock groups within the Dixie Valley region. These include: Upper Triassic metasediments; Middle Jurassic quartz arenite, mafic volcanic rocks, and gabbroic rocks; Miocene rhyolites; and Pliocene and younger sedimentary rocks, basalt, and andesite. A few of the lithologies encountered, such as granodiorite and aplite, are distinctly atypical of the local Stillwater Range source area.

The dominant rock types encountered in the wells drilled in Dixie Valley are, in order of decreasing abundance: Upper Triassic metasedi-

ments, gabbroic to granodioritic intrusive rocks and silicic to intermediate volcanic rocks. The volcanic rocks are the dominant constituents of the alluvial sediments because these rocks are well exposed along the front of the Stillwater Range in the immediate catchment areas where the wells have been drilled. Although outcrops of the metasediments are present in the Stillwater Range, they are conspicuously rare or absent in the alluvial sediments.

The rock types encountered in the two wells are similar, however, the relative amounts and stratigraphic position of the units are quite different. The complex structural-tectonic and depositional history of Dixie Valley, as well as the effects of the hydrothermal system underlying the region, make stratigraphic correlations between the two deep exploratory wells tentative. A more thorough discussion of the correlation between DF 45-14 and DF 66-21 is presented in Section 4.2.7. Plate IV shows the location of the six shallow thermal gradient holes and the two deep exploratory wells in relation to the Stillwater Range front and the structural-tectonic features in Dixie Valley.

4.2.3 Lithology of Deep Exploratory Wells

4.2.3.1 Well DF 45-14

Rocks penetrated in the deep exploratory well DF 45-14, drilled to a total depth of 9020 feet, include 1100 feet of unconsolidated heterogeneous alluvial sediments, 1500 feet of silicic to intermediate volcanic rocks, 6500 feet of Upper Triassic metasediments and minor intrusions of diorite/gabbro. The different rock types exhibit varying degrees of hydrothermal alteration locally controlled by fracture permeability; however, the type of alteration is fairly consistent in the drill hole.

DF 45-14 penetrated approximately 6500 feet of a possible 10,000 feet (Page, 1965) of Upper Triassic metasediments. The unit as a whole is quite homogeneous; however, local facies variations are present, and their differential response to hydrothermal alteration is striking (compare Figures 4-2 and 4-3). The metasedimentary sequence consists chiefly of metasiltstone/metashale, quartz arenite/metaquartz arenite, and metalithic wacke.

The metasediments are here divided into two facies based mainly on grain size, as the mineralogy is quite consistent for the entire



Figure 4-2. (a) Photomicrograph of metasiltstone/meta-shale facies under polarized light. Note the quartz vein. (scale 13 mm : 0.22 mm)

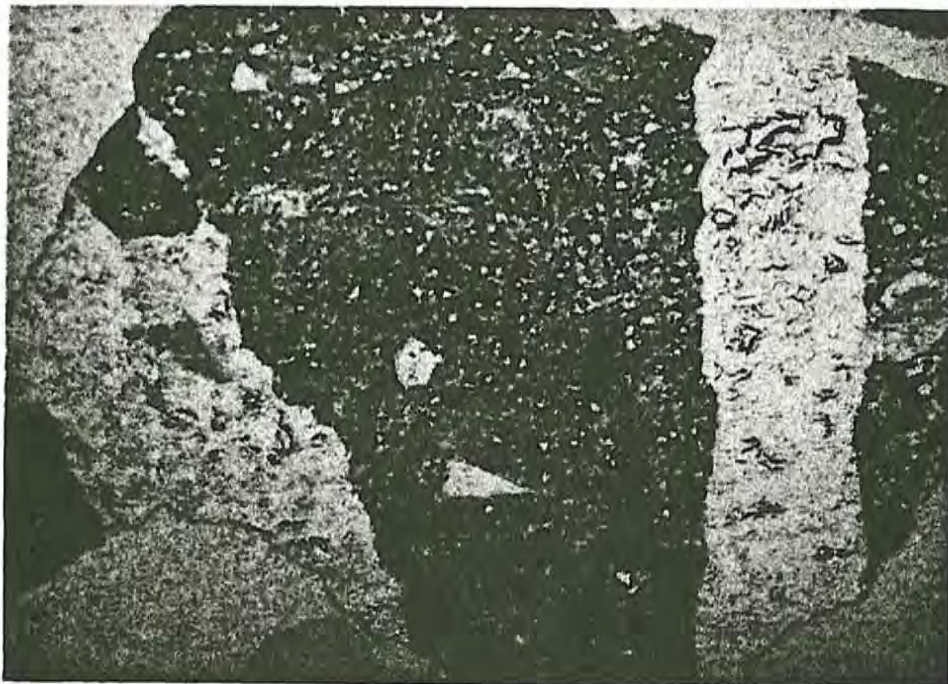


Figure 4-2. (b) Photomicrograph of metasiltstone/meta-shale facies. Note 'dirty' altered appearance. (scale 13 mm : 0.22 mm)



Figure 4-3. (a) Photomicrograph of metarenite facies under polarized light. (scale 13 mm : 0.22 mm)

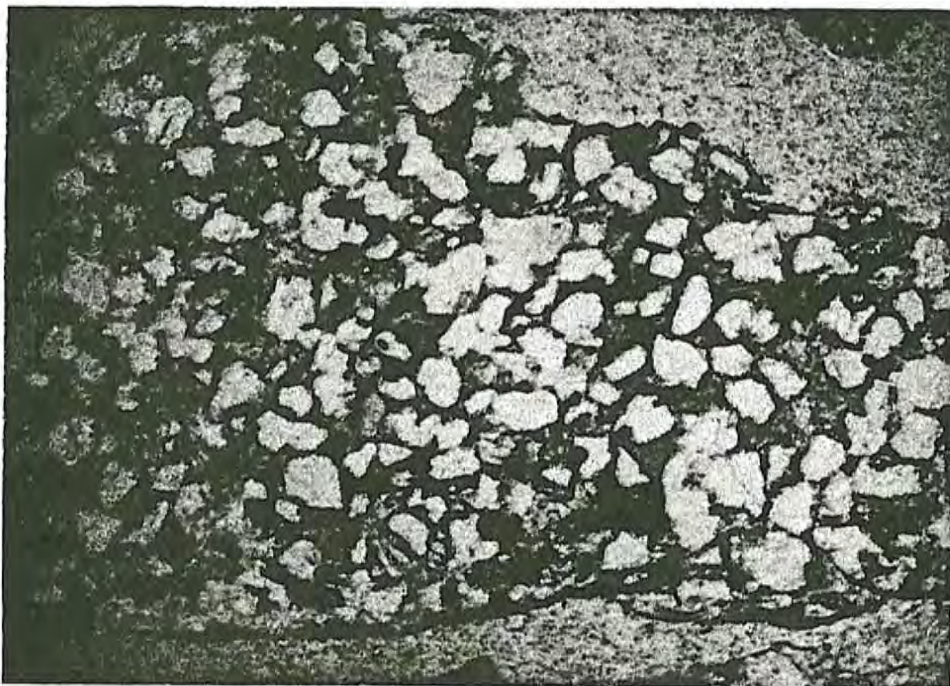


Figure 4-3. (b) Photomicrograph of metarenite facies. Note the lack of visible alteration. (scale 13 mm : 0.22 mm)

sequence. The majority of this metasedimentary sequence consists of a very fine-grained, black to medium-dark gray, well sorted, very hard, silicic, in places recrystallized and well foliated metasilstone/metashale, often containing abundant carbonaceous matter and varying amounts of disseminated and crystalline pyrite. The foliation is a result of regional low grade metamorphism during the late Jurassic (Willden and Speed, 1974). The unit is quite dense, possesses limited porosity and shows only local signs of fracturing. Where it is intruded by diorite/gabbro dikes (sills?) contact metamorphism has resulted in the formation of andalusite/chiaistolite porphyroblasts (Figure 4-4) and incipient spherulitic cordierite (?) (Figure 4-5).

The overall effects of the regional low grade metamorphic event on the metasediments include recrystallization, chemical reconstitution and textural changes. Figure 4-5a shows the well developed foliation observed in the organic-rich fine-grained metasilstone/metashale (compare with Figure 4-3). The present mineral assemblage observed in the metasedimentary sequence (quartz, K-micas, chlorite, clay minerals \pm calcite \pm laumontite) is the result of the composite effects of the regional and contact metamorphic events and the hydrothermal activity. Figure 4-5b is an example of the chiaistolite/cordierite bearing metasilstone from an area approximately 45 miles south in the Stillwater Range showing a relatively unaltered appearance when compared with Figure 4-5a. The difference in appearance may be largely a function of the hydrothermal activity in the northern portion of Dixie Valley.

The andalusite/chiaistolite crystals and the incipient spherulitic crystals of cordierite (?) were observed to occur only in the fine-grained facies of the metasedimentary sequence. Here, the foliation exhibited by orientation of the phyllosilicate minerals and carbonaceous matter, invariably wraps around the 'spots' (Figure 4-5a). The andalusite/chiaistolite crystals, however, generally grow across the foliation. Spry (1976) used the term 'maculose' structure to describe porphyroblasts of andalusite/chiaistolite in a fine-grained matrix. The organic 'cross' or inclusion characteristic of chiaistolite (Figure 4-4) is believed due to preferred adsorption of carbonaceous particles against the prism faces, particularly at the prism edges (Spry, 1976). Chiaistolite and its organic inclusion in particular are highly susceptible to sericitic alteration. This feature has



Figure 4-4. (a) Photomicrograph of andalusite/chiastolite porphyroblasts in organic-rich metasilstone/metashale from DF 45-14. (scale 13 mm : 0.34 mm)



Figure 4-4. (b) Photomicrograph of andalusite/chiastolite crystals in very organic-rich metasilstone/metashale. (scale 13 mm : 0.34 mm)

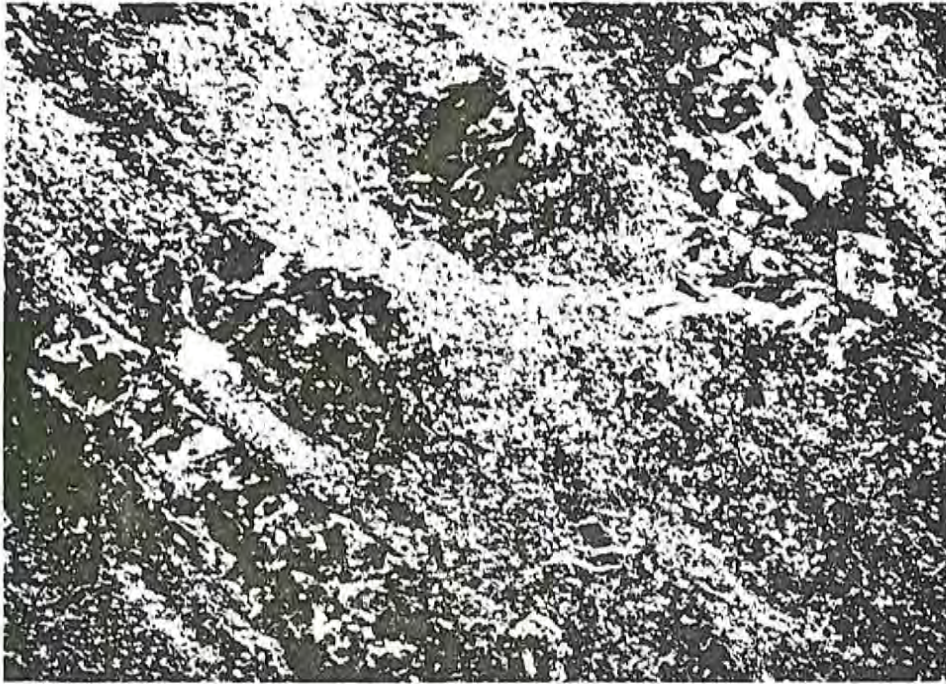


Figure 4-5. (a) Photomicrograph of andalusite/chiasmolite and incipient spherulitic cordierite porphyroblasts. Note foliation wrapping around the spherules. (scale 13 mm : 0.34 mm)

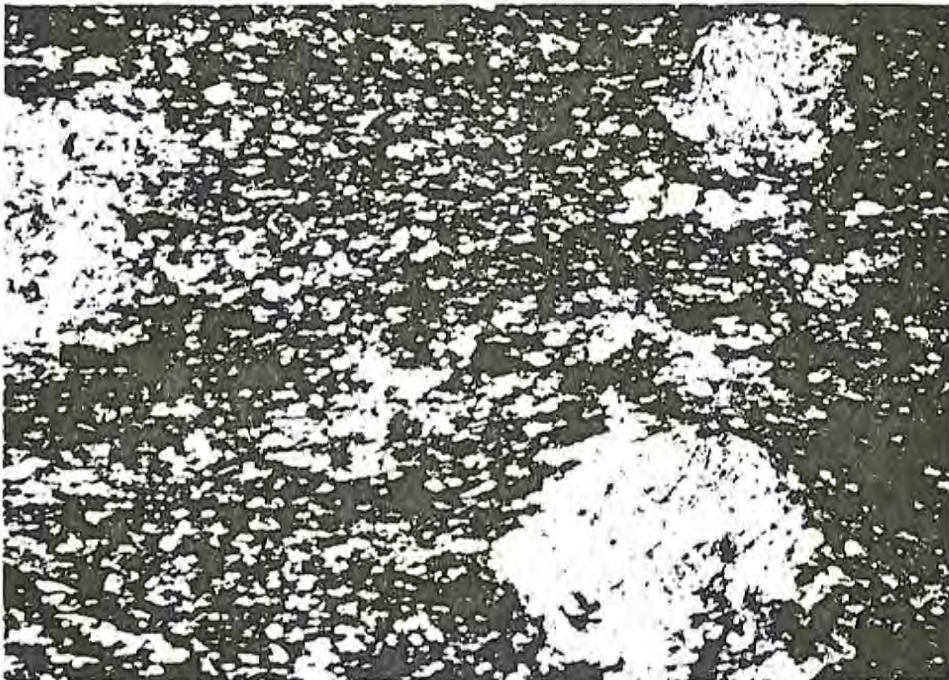


Figure 4-5. (b) Photomicrograph of relatively fresh andalusite/chiasmolite and cordierite-bearing meta-siltstone from the southern portion of the Stillwater Range. (scale 13 mm : 0.34 mm)

been widely observed in thin section as seen in Figure 4-6.

The second facies of the Triassic metasediments is a coarser-grained, light to dark gray to brown, very hard, silicic, recrystallized quartz arenite or meta-quartz arenite (Figure 4-3). It consists predominantly of round to subround optically clear and partially recrystallized quartz grains with varying amounts of micaceous minerals (chlorite, biotite, K-mica), calcite and clay occurring as intergranular phases. Trace amounts of magnetite, pyrite, garnet, epidote, tourmaline and carbonaceous matter are also present. Recrystallization of quartz in this unit is evidenced by grain boundary triple points and by overgrowths. This recrystallization may be responsible for the conspicuous lack of a well defined foliation as seen in the metasiltstone/metashale facies. This unit appears to have been largely unaffected by the metamorphism. The mineralogy of the finer-grained facies is essentially the same as that described above except for the occurrence of andalusite/chiasmolite and cordierite porphyroblasts and the abundance of carbonaceous matter and pyrite.

Several intervals of the metasedimentary sequence in DF 45-14 (3700-3800, 4100-4200, 4600-4700, 7600-7700, 8000-8100 feet) (Plate V) have been intruded by Middle Jurassic (?) diorite/gabbro dikes. The average thickness of these intrusions is about 30 feet as described in driller's logs. They consist predominantly of a medium to coarse-grained crystalline network of plagioclase and clinopyroxene with little or no potassium feldspar or quartz. Secondary minerals include hornblende, chlorite, magnetite, epidote, albite, calcite, and pyrite. An x-ray fluorescence analysis of the diorite/gabbro indicated the presence of significant quantities of titanium suggesting the presence of ilmenite. Similarly the presence of leucosene also indicates the existence of ilmenite. Figure 4-7 shows the mineralogy and texture of diorite/gabbro although it commonly takes on a more altered, 'dirty' appearance.

A total of 1500 feet of volcanic rocks were encountered in DF 45-14 in the intervals 700 to 800 feet and 1100 to 2500 feet. These consist primarily of andesite and basalt and lesser amounts of silicic tuff. The tuffs are gray to green and brown and consist of fine-grained interlocking plagioclase lathes in a very fine-grained, glassy, often devitrified matrix that is altering to clay. These vitric and



Figure 4-6. Photomicrograph of highly sericitized andalusite/cordierite porphyroblasts in metasilstone/metashale from DF 45-14 under polarized light. (scale 13 mm : 0.34 mm)



Figure 4-7. Photomicrograph illustrating the texture and mineralogy of diorite/gabbro under polarized light. (scale 13 mm : 0.34 mm)

lithic tuffs contain varying amounts of disseminated and crystalline pyrite. The andesites and basalts are predominantly medium to dark brown, dark gray to green, medium to fine-grained rocks consisting of interlocking plagioclase lathes in a groundmass of very fine-grained plagioclase and minor quartz. Pyrite, hornblende, epidote, and traces of original pyroxene are also observed. The age of this sequence of volcanic rocks is unknown, however they are believed to be of Tertiary age.

The approximately 1100 feet of alluvial sediments penetrated in DF 45-14 are an unconsolidated heterogeneous mixture of all the rock types described above except for the conspicuous rarity of the meta-sediments. The majority of the alluvium consists of volcanic fragments with lesser amounts of diorite/gabbro fragments. The volcanic material is intensely altered (Figure 4-8) as it has largely been derived from the erosion of the nearby highly altered, vari-colored volcanic rocks in the Stillwater Range (Figure 4-9). This alteration zone represents a previous period of more intense and possibly more acidic hydrothermal alteration. This zone of altered volcanics is probably indicative of the alteration of the volcanic sequence at depth. Delineating the weak overprint produced by the present hydrothermal activity from the effects of the previous alteration is difficult in the alluvial material.

4.2.3.2 Well DF 66-21

The rock types encountered in DF 66-21 are essentially the same as those observed in DF 45-14 although the proportions of each are rather different as indicated by a comparison of the lithologic logs (Plates V and VI). The thicker sequence of alluvial sediments in DF 66-21 is in part a function of increased distance from the range front and at least one additional increment of downdropping along normal faults (Plate IV). DF 66-21 also contains a much greater amount of intrusive material (granodiorite and diorite/gabbro). The mineralogy of these intrusive rocks are quite similar and it is uncertain if they originated from the same magma body. The relative ages of these two types of intrusives are unknown. The most obvious difference between the two wells is the much smaller amount and type of metasediments encountered in DF 66-21. The detailed description of the rock types in DF 45-14

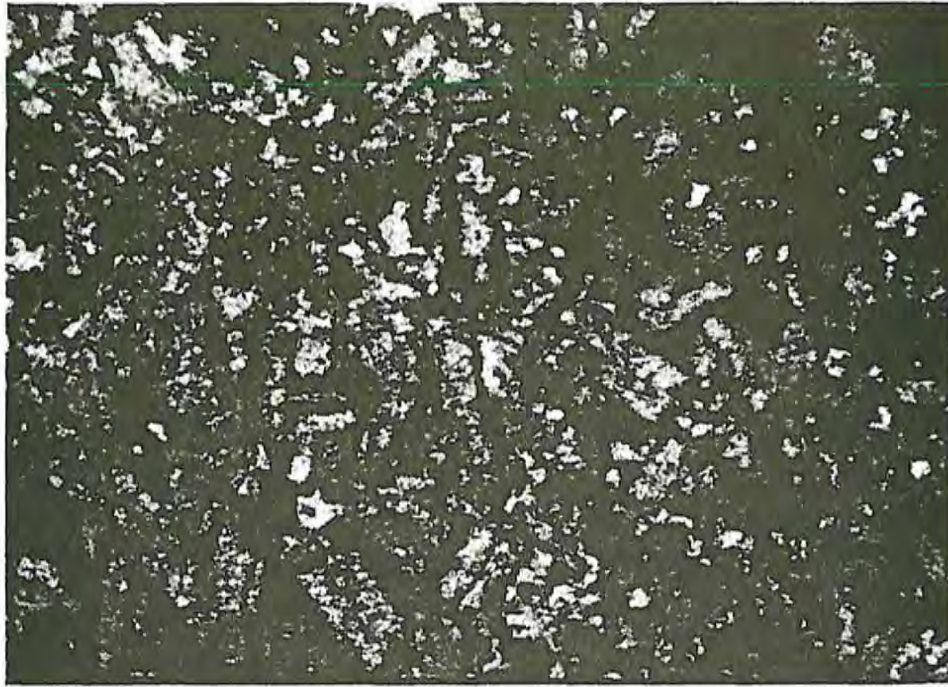


Figure 4-8. Photomicrograph of altered volcanic rock from alluvial material under polarized light. Note relict sericitized plagioclase lathes. (scale 13 mm : 0.27 mm)



Figure 4-9. Photograph of vari-colored zone of alteration of volcanic rocks in the Stillwater Range.

given above can also apply to the rock types in DF 66-21. Only obvious departures from these descriptions will be noted below.

The sequence of alluvial sediments encountered in DF 66-21 consists of essentially the same rock types as found in the alluvium in DF 45-14. In many intervals, however, the proportion of volcanic fragments is much greater in DF 66-21. Additionally, the presence of abundant reddish-brown clay in the alluvium in the lower 2000 feet and extending into the upper portions of the volcanic sequence, is unique to DF 66-21. This clay is essentially a montmorillonite which is fairly uniform in composition throughout the interval. This clay constituted up to 75 percent of the sample taken by the well logger. The origin of this layer and its significance as a possible cap rock will be discussed later.

The lithology and thickness of the sequence of volcanic rocks in DF 66-21 (1300 feet) is nearly the same as in DF 45-14 (1500 feet). One difference was the relative abundance of pyroxene blebs and greater amounts of disseminated magnetite in the andesitic rocks of DF 66-21. The significance of this observation, however, is uncertain. Drilling into these volcanic rocks encountered hot water and steam under high pressure. Presumably, all or part of this volcanic sequence has been highly fractured and brecciated, evidenced by increased porosity on logs and by calcite deposition.

A relatively homogeneous granodiorite/diorite (?) unit was encountered from 5300 to 6600 feet in DF 66-21. The mineralogy of this unit is not significantly different from the intrusive rocks encountered in DF 45-14, although the proportions of the various minerals are different. The 'granodiorite' intrusive in DF 66-21 does not contain as much magnetite; biotite and hornblende, rather than pyroxene and hornblende, are the dominant ferromagnesian minerals. This observation may be a secondary effect. Additionally, the 'granodiorite' in DF 66-21 contains more plagioclase and minor amounts of K-spar. The question is whether these two rock types represent parts of the same major intrusive body or not. The structural-stratigraphic relationships of this intrusive unit(s) is complex, and the lack of any granodiorite in DF 45-14, however, does suggest that it is of limited areal extent. It is possible that the granodiorite intrusion pinches out to the south toward DF 45-14, or that its absence is fault-controlled. This intrusive body

is considered to be part of the Humboldt gabbroic complex (Speed, 1976).

The metasediments in DF 66-21 are poorly represented, as only about 1100 feet of them were encountered. They consist primarily of quartz arenite, meta-quartz arenite (?) and metasiltstone/metashale. The mineralogy and structures are similar to those for similar rocks in DF 45-14. The difference is in the greater amount of coarse-grained quartz arenite (meta-quartz arenite) relative to the metasiltstone/metashale in DF 66-21. As in DF 45-14 the fine-grained facies is more intensely altered, whereas the coarse-grained facies is relatively unaltered in appearance. Another difference is the lack of andalusite/chiasolite porphyroblasts in the metasiltstone/metashale facies in DF 66-21.

Several intervals within the drill hole were reported to contain mineralized zones. Petrographic examination and a rapid binocular scan of the cuttings, however, could not verify the presence of any significant quantities of metal sulfides or oxides with the exception of pyrite, magnetite, and possible traces of chalcopyrite/bornite.

4.2.4 Alteration Effects in Deep Exploratory Wells

The varied rock types in DF 45-14 and DF 66-21 with their differential responses to the processes of alteration have given rise to a complicated assemblage of alteration minerals. The alteration in the two wells, however, is quite similar, the difference being one chiefly of degree rather than type.

4.2.4.1 Well DF 45-14

The alteration described below is not solely due to the effect of the hydrothermal activity that exists in Dixie Valley, as many of the rocks encountered have experienced regional low-grade metamorphism and/or contact metamorphism. Delineating the effects of these metamorphic events from those derived from the hydrothermal activity is at times difficult, as only a very fine line separates the two processes.

The sequence of volcanic rocks and the alluvial materials, which consist primarily of altered volcanic fragments, exhibit a very similar alteration mineral assemblage. This assemblage consists of albite, chlorite, sericite (illite), epidote, calcite, and clay minerals. A similar alteration assemblage for altered volcanic rocks of the Comstock Lode is described by Coats (1940) and is termed propylitic alteration.

Coats attributes the origin of propylitic alteration to hydrothermal alteration at relatively high temperatures. The same mineral assemblage has been documented as being indicative of low grade regional metamorphism (Winkler, 1976) or the equivalent greenschist facies of Eskola (1915).

The basic mineral assemblage described above has been noted in other geothermal areas. Work by Brown and Ellis (1970) at Ohaki-Broadlands, New Zealand, found hornblende and biotite replaced by chlorite, illite, calcite, quartz, or pyrite, with plagioclase replaced by albite, epidote, illite, calcite, adularia, wairakite, and quartz in rhyolitic and tuffaceous volcanic rocks. Steiner (1968) at Wairakei, New Zealand, described ferromagnesian minerals altering to chlorite, micaceous clay minerals (illite, illite/montmorillonite), epidote, calcite, quartz, and pyrite with plagioclase altering to montmorillonite, micaceous clays, epidote, calcite, quartz, and K-feldspar in tuff, andesite, and ignimbrite. Work by Schoen and White (1965) at Steamboat Springs, Nevada, identifies hornblende altering to epidote, calcite, mixed-layer illite/montmorillonite clays, and K-feldspar in acidic to intermediate volcanic rocks.

Propylitic alteration in DF 45-14 is best developed in the acid to intermediate volcanic rocks where it imparts a greenish-gray color to the rocks. It renders the albite phenocrysts less translucent and the twinning becomes diffuse or is totally destroyed. Widespread albitization of the plagioclase in both the volcanic rocks and the intrusive rocks is supported by x-ray diffraction data of random whole rock samples of the alluvial material which revealed an abundance of albite. Albitization is the process whereby the more reactive Ca-rich plagioclase is converted to the Na-rich end member of the plagioclase series, albite ($An_0 - An_{10}$). Presumably, Na-rich solutions attack the Ca-plagioclase through a cation exchange process, thereby releasing Ca ions which are probably incorporated into the formation of epidote, calcite, Ca-montmorillonite and/or possibly laumontite. This process is most likely a function of the low-grade metamorphism although albitization due to hydrothermal activity cannot be completely ruled out.

Sericitization of plagioclase, or its replacement by illite, is pervasive and in some instances complete as it consumes entire crystals, leaving only a relict form of the crystal (see Figure 4-10a). This



Figure 4-10. (a) Photomicrograph of a plagioclase crystal completely consumed by sericite. View under polarized light; note retention of crystal form. (scale 13 mm : 0.27 mm)



Figure 4-10. (b) Photomicrograph of sericitization confined to the core of a zoned plagioclase crystal. View under polarized light. (scale 13 mm : 0.27 mm)

process is most commonly developed in the cores of plagioclase crystals (Figure 4-10b) as the calcic cores of zoned plagioclase are more susceptible to alteration. Somewhat of a conflict exists in the literature regarding the physical and chemical conditions necessary for sericitic alteration. Turner and Verhoogen (1960) state that the process of sericitization is indicative of alteration by alkaline solutions of higher concentrations of potash. Lovering (1950), on the other hand, says that sericitic alteration takes place in mildly acidic solutions at higher temperatures. The preponderance of sericite (illite) in thin section and in the x-ray diffraction patterns support the ongoing process of sericitization.

The plagioclase is also altered to an epidote (clinozoisite (?)) mineral. This process is generally confined to the calcic cores of the plagioclase crystals (Figure 4-11) as the Ca ions necessary for the formation of epidote are available. Some of the intrusive rocks are saussuritized, although the epidote (clinozoisite (?)) is commonly coarsely crystalline. White and Sigvaldason (1962) suggested there are upper limits to the depth of formation of epidote in hydrothermal environments. In this regard, epidote may be useful in defining limits for depth of original cover in areas of altered volcanic rocks and epithermal ore deposits. If epidote occurs at the present land surface, for example, erosion of at least 400m (1300 feet) of cover is indicated (White and Sigvaldason, 1962). However, a hydrothermal origin for the epidote is very difficult to ascertain in many instances. When the epidote occurs in veins it may be assumed to be hydrothermal.

Replacement of plagioclase by clay and lesser amounts of calcite is commonly developed in the groundmass of the volcanic rocks and has largely masked much of the original texture. The presence of montmorillonite and mixed-layer illite/montmorillonite in the upper 2500 feet of DF 45-14 is evidence for this process as the volcanic rocks contain abundant plagioclase. Montmorillonite is characteristic of relatively low temperatures and its formation by alkaline solutions appears to be compatible with the near-surface water chemistry as will be discussed later.

The alteration of the primary and, in some instances, the secondary ferromagnesian minerals and ilmenite in the rocks of DF 45-14 is pervasive and commonly complete. The end product is generally either bio-



Figure 4-11. (a) Photomicrograph of epidote alteration in cores of plagioclase crystals. View under polarized light. (scale 13 mm : 0.34 mm)



Figure 4-11. (b) Photomicrograph of epidote alteration in cores of plagioclase crystals. (scale 13 mm : 0.34 mm)

tite, chlorite, magnetite, clay minerals and/or epidote or combinations of these minerals. Leucoxene is believed to represent the alteration of ilmenite under hydrothermal conditions. Original clinopyroxenes of the intrusive and volcanic rocks are replaced by hornblende, biotite, chlorite, epidote and magnetite. Primary and secondary hornblende is replaced by biotite, chlorite, magnetite, epidote and calcite; biotite is replaced by chlorite, vermiculite and magnetite. The iron derived from the alteration of these minerals as well as pyrite is probably responsible for the pervasive limonite and hematite staining of the alluvial material and possibly the abundance of magnetite. The precipitation of magnetite from solution is evidenced by its deposition in some veins and in intergranular positions as seen in Figure 4-12. X-ray diffraction patterns for the chlorites show that the second and fourth orders of the basal reflection are consistently stronger than the first, indicating an iron-rich type (Grim, 1968).

The propylitic alteration developed in the alluvium and the volcanic sequence in DF 45-14 is also recognized, to a lesser extent, in the intrusive rocks. For instance secondary hornblende is replaced by chlorite and magnetite with plagioclase altered to albite, sericite (illite), epidote, clay and calcite. Propylitic alteration of a diorite is described in the Tongonan Geothermal Field, Leyte, Philippines (Kingston, 1979). Despite these phase transitions, the diorite/gabbro maintains a relatively fresh unaltered appearance, except in localized areas where fracturing may be a factor.

The alteration products of primary minerals in the alluvial material, in the sequence of volcanic rocks, and in the intrusive rocks are all similar, differing only in the degree of alteration. This observation along with a similar mineralogy documented in other geothermal areas indicates a hydrothermal origin for much of the observed alteration assemblage in DF 45-14. The degree of alteration is mainly a function of rock texture. The texture will determine both the porosity and permeability and hence the fluid migration rates and ultimately the reaction times. Grain size is also an important consideration. The finer-grained material, as a rule, is more altered because of higher susceptibility to hydrothermal solutions and closer contact to flow channels as a result of a greater amount of surface area (compare Figures 4-2 and 4-3). The porosity-permeability relationships in DF 45-14 are very important as



Figure 4-12. (b) Photomicrograph of magnetite completely surrounding a vitric clast in a fragment of volcanic rock. (scale 13 mm : 0.1 mm)



Figure 4-12. (a) Photomicrograph of magnetite completely surrounding a vitric clast in a fragment of volcanic rock. View under polarized light. (scale 13 mm : 0.1 mm)

they are the dominant control in the distribution of the various hydrothermal alteration effects.

The hydrothermal alteration observed in the metasediments is rather limited except in localized regions associated with inferred zones of fracturing and/or faulting (Plate V). The post-metamorphic mineral assemblage of quartz, K-mica, biotite and chlorite has not been significantly changed. This is supported by x-ray diffraction data and thin section analysis.

Sericitization (illitization) has been an active process, particularly in the fine-grained metasiltstone/metashale, where it occurs predominantly as an intergranular phase alteration product of the primary K-micas. Quartz is essentially nonreactive, except for recrystallization which most probably was a result of the regional metamorphic events. Alteration of biotite is observed predominantly in the coarse-grained quartz-arenite and/or meta-quartz arenite facies where it is altered to chlorite and vermiculite (?). Petrographic evidence for biotite altering to vermiculite includes: (1) lack of the "bird's-eye" extinction characteristic of biotite, (2) a decrease in birefringence, (3) a decrease in pleochroism and a bleached appearance, and (4) pseudomorphs after biotite. One of the major occurrences of vermiculite is as an alteration product of biotite by weathering or by hydrothermal action (Deer and others, 1966). Vermiculite has also been documented in the Roosevelt Hot Springs Thermal Area, Utah (Ballantyne, 1978) where it has replaced biotite.

Andalusite/chiasmolite porphyroblasts in the foliated, organic-rich metasiltstone/metashale are invariably partially or wholly sericitized, particularly the included organic material. The stability of chiasmolite is such that even under normal atmospheric conditions its alteration to sericite is entirely possible (Spry, 1976). A large proportion of the 'spots' or incipient cordierite (?) porphyroblasts are altered to clay. Apart from the above observations, the metasediments are not highly altered; rather, they appear to be relatively impervious to the altering solutions. Only in localized zones, where the various well logging techniques (gamma-ray and neutron density logs) indicate anomalous porosity, is there any variation in the observed mineralogy, in particular the clay mineralogy. The clay mineralogy is discussed in Section 4.2.5.

Calcite and quartz are the dominant vein and fracture fill material along with the Ca-zeolite laumontite and lesser amounts of adularia and gypsum (?) in the metasediments of DF 45-14. The distribution and relative abundance of calcite and laumontite in relation to lithology and postulated fault and fracture zones are depicted on Plate V. These minerals exhibit a variety of structural and temporal relationships. Laumontite and adularia invariably occur in post-metamorphic veins whereas calcite and quartz are less predictable. Both laumontite and adularia have been described as minerals typically found in geothermal systems (Steiner, 1968; Browne and Ellis, 1970; Ellis and Mahon, 1977).

The presence of the Ca-zeolite laumontite ($\text{CaAl}_2\text{Si}_4\text{O}_{12}\cdot 4\text{H}_2\text{O}$) was confirmed by x-ray diffraction data as well as in thin section analysis (Figure 4-13). The main diffraction peaks at 9.1A° to 9.3A° and 6.7A° to 6.9A° disappear after heating to 550°C for one-half hour. Laumontite reportedly dehydrates with increasing temperature to wairakite (Winkler, 1976); however, this was not observed to occur. Leonhardite is a slightly dehydrated form of laumontite. The stability range of laumontite ranges from about 200°C to 300°C at relatively low pressures of about 3.0 Kb (Winkler, 1976). The first appearance of laumontite is often cited as indicating the beginning of metamorphism.

In thin section, laumontite is coarsely crystalline with low birefringence, low relief and exhibits two directions of observable cleavage, one distinct and the other obscure. Laumontite occurs with quartz and calcite in veins and fracture fillings (Figure 4-13b). The assemblage quartz-laumontite is typical of the zeolite facies (Deer and others, 1966; Winkler, 1976). This same assemblage is described in Onikabe, Japan (Ellis and Mahon, 1977), where it was indicative of a temperature of 170°C . The occurrence of laumontite has been reported elsewhere in other geothermal areas, including: Reykjavik and Jvergerdi, Iceland (Sigvaldason and White, 1962); Wairakei, New Zealand (Steiner, 1968); and in Matsukawa, Japan (Ellis and Mahon, 1977).

Lovering (1950) stresses the fact the zeolites have been synthesized only in alkaline aqueous solutions and that the temperature in some cases was as low as 60°C . According to Lovering, the zeolites develop near the end of the reaction series of hydrothermal solutions.

The distribution and relative abundance of laumontite (Plate V)



Figure 4-13 (a) Photomicrograph of a fragment of laumontite vein material under polarized light.
(scale 13 mm : 0.22 mm)



Figure 4-13. (b) Photomicrograph of laumontite vein in metasilstone/metashale under polarized light.
(scale 13 mm : 0.22 mm)

is spatially related to the known or inferred areas of increased porosity and permeability, i.e. fractures or faults. The laumontite occurrences in the intervals between 4000 and 6000 feet correlate not only with anomalous clay occurrences, but also with increases in porosity deduced from the various well logs. Temperatures in this interval range from 110°C to approximately 140°C. Large concentrations of illite (sericite) occur in this interval, which as noted earlier may be indicative of alteration by alkaline solution (Turner and Verhoogen, 1960). Randomly interstratified illite/montmorillonite is also concentrated in this interval (4000 to 6000 feet) further supporting the idea that slightly alkaline solutions may exist or have existed at depth. This idea will be discussed in more detail in Section 4.2.6 Water Chemistry.

4.2.4.2 Well DF 66-21

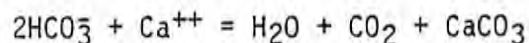
The propylitic alteration assemblage characteristic of the alluvial sediments, volcanic rocks and, to a lesser extent, the intrusive rocks in DF 45-14 is likewise well developed in their counterparts in DF 66-21. Detailed descriptions of the effects of propylitic alteration on the rocks are given above; only significant differences in the alteration mineralogy of DF 66-21 are discussed.

Examination of Plate VI shows that the distribution of various mineral species is quite localized, particularly for the clay minerals. This evidence, in conjunction with the interpretation of available electric logs, indicated the following intervals are anomalous: 3900 to 4700, 5000 to 5200, 5300 to 6100, 6300 to 6900, 7100 to 7400, 7600 to 8800, and 9100 to 9200 feet. The distribution of calcite, epidote and, to a lesser degree, chlorite and the percentage of plagioclase alteration as well as the various clay minerals is evidence that these zones represent regions of more intense hydrothermal alteration. Although lithology cannot be ruled out as a controlling factor, fracture porosity and permeability appear to be the dominating influences in the distribution of many of the various mineral species.

The intervals mentioned above are interpreted as representing zones of increased porosity, i.e. fault or fracture zones, which provide permeable avenues for migrating thermal fluids and hence the greater degree of alteration. The percent plagioclase alteration, although mainly a function of rock type, does show a weak correlation with structure

especially in the interval 4000 to 4500 feet. The distribution of chlorite is likewise somewhat indistinct, however certain correlations are discernible, i.e. 4000 to 5500 feet and 6300 to 6700 feet. Here again the distribution may be primarily controlled by lithology; however the structural correlations in the above instances are rather strong. Epidote is fairly localized as well, although in this instance the dominating control is the occurrence of the granodiorite intrusive rock. This relationship is clearly visible on Plate VI. In thin section, the intrusive rock was observed to be highly saussuritized although this is likely a function of the metamorphism.

The relative amount and distribution of calcite exhibit strong correlation with the position of known or inferred zones of fracturing and/or faulting. This is particularly evident in the intervals 4000 to 5500, 6400 to 6700 and 7600 to 9400 feet. It may be that these intervals represent zones of boiling where CO_2 is lost and calcite is precipitated according to the following equation:



As CO_2 is lost, the equilibrium constraints cause the equation to move to the right resulting in calcite precipitation. Figure 4-14 shows the intense amount of fracturing in the zone around 6500 feet and the filling of these fractures with calcite. Calcite as vein material is much more abundant in DF 66-21 than in DF 45-14. Minor amounts of adularia, epidote and quartz were also observed in veins in DF 66-21. Most of the epidote veins were altered to calcite and clay.

Limonite and hematite are abundant in the upper 4000 feet of alluvium and then drop off dramatically below 5100 feet. Sporadic increases occur at 6200 to 6700 feet and 8800 to 9100 feet, corresponding closely to zones of inferred fracturing or faulting. These occurrences may indicate the presence of oxygenated waters at great depths suggesting possible communication with the surface via faults and fractures.

The alluvial sediments in DF 66-21 not only differ in amount from those in DF 45-14 but also in their preponderance of volcanic fragments. Many intervals in DF 66-21 are comprised almost solely of volcanic fragments or more appropriately, 'volcanic gravels'. This is probably a function of the location of the well in relation to the zone of altered volcanic rocks directly to the west in the Stillwater Range. This all-

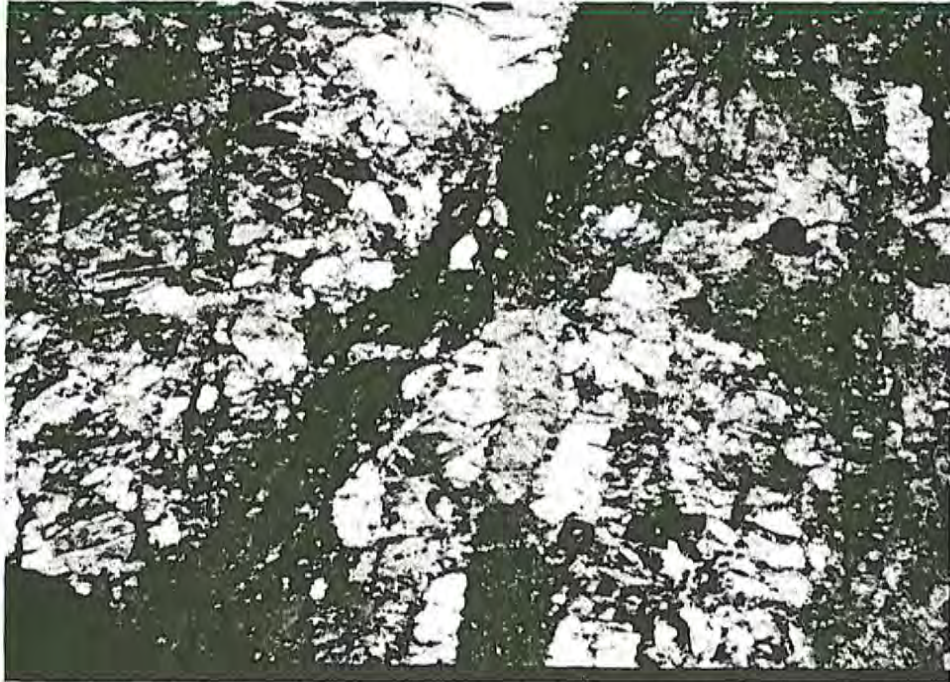


Figure 4-14. (a) Photomicrograph of an intensely fractured zone at 6500 feet in DF 66-21 with abundant calcite deposition. View under polarized light. (scale 13 mm : 0.34 mm)

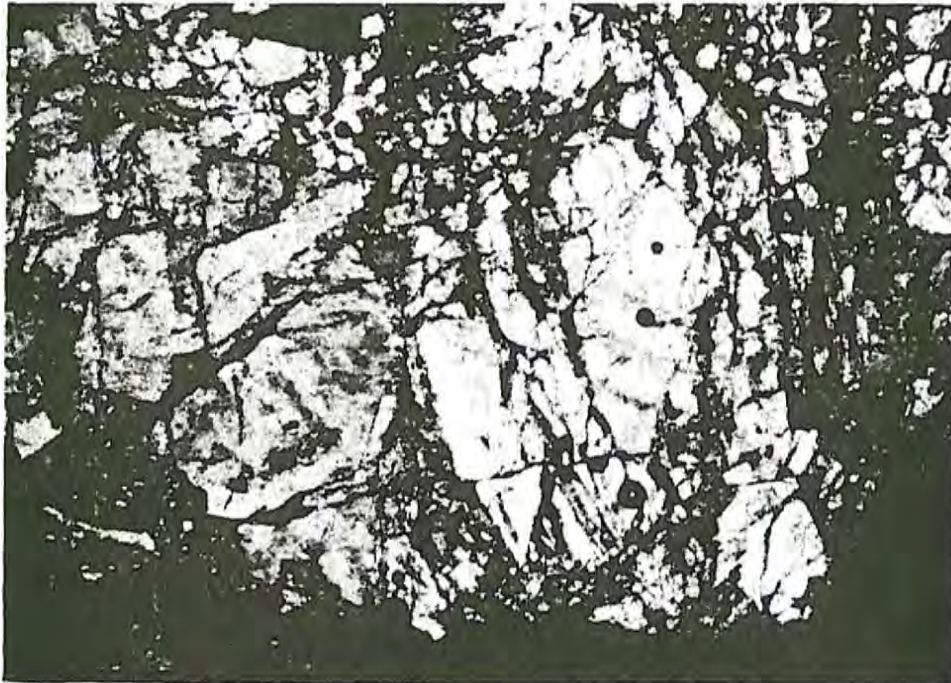


Figure 4-14. (b) Photomicrograph of an intensely fractured zone at 6500 feet in DF 66-21 with abundant calcite deposition. View under polarized light. (scale 13 mm : 0.34 mm)

uvial material also exhibits a greater degree of hydrothermal alteration, apart from the observed propylitic alteration, that is a result of the ongoing hydrothermal activity in the area.

4.2.5 Clay Mineralogy of Deep Exploratory Wells

4.2.5.1 Well DF 45-14

Plate V shows the relative abundances and distribution of the clay minerals identified by x-ray diffraction techniques and their relationships to fracture zones, lithology, temperature and other mineral species in DF 45-14. The clay minerals are the most useful indicators of the distribution of alteration effects. The clay minerals exhibit a consistent and highly localized pattern of distribution, particularly the mixed-layer and micaceous clays (illite). In Wairakei, the distribution of the montmorillonite zone and the micaceous clay mineral (illite, mixed-layer illite-montmorillonite) zones are related to temperature and fissures carrying geothermal fluid (Steiner, 1968). This seems to be the case in both DF 45-14 and DF 66-21.

The x-ray diffraction analysis of the fine fraction from selected intervals in DF 45-14 reveals the presence of the following clay minerals, in order of increasing temperature and depth: Ca/Mg-montmorillonite, Na-montmorillonite, mixed-layer illite-montmorillonite, and illite. This relationship very closely parallels the clay mineral distribution in Wairakei, New Zealand (Steiner, 1968). It is also documented in other geothermal areas, including Ohaki-Broadlands (Browne and Ellis, 1970), Reykjanes, Iceland (Tomasson and Kristmannsdottir, 1972), and in the Imperial Valley, (Hoagland, 1976).

The temperature/depth relationships of chlorite and kaolinite in DF 45-14 are less certain. Both chlorite and kaolinite have fairly wide stability fields. The difficulties associated with distinguishing kaolinite from chlorite using x-ray diffraction techniques make these minerals tenable indicators of varying conditions of hydrothermal alteration. However the presence of kaolinite is an important indicator of the pH conditions. It occurs in close proximity to fracture zones.

Montmorillonite is indicative of temperatures of 100^o to 200^oC in altered rhyolitic tuffs in Wairakei (Steiner, 1968). In the Salton Sea Geothermal Area, a temperature of about 100^oC is observed for the termination of mixed-layer illite-montmorillonite, while the conversion to

illite is complete at water temperatures of about 210°C (Ellis and Mahon, 1977). At Wairakei, Steiner (1968) found that low temperatures between about 120°C and 230°C prevail at shallow depths (i.e. low lithostatic and water vapor pressures) where mixed-layer minerals are forming. He found that high pressures and temperatures above 230°C are required for the formation of illite and mixed-layer clays deficient in montmorillonite. In Reykjanes, Iceland, montmorillonite is the dominant sheet silicate up to temperatures of about 200°C; for temperatures of 200° to 300°C, the most common sheet silicates are random mixed-layer minerals of chlorite and montmorillonite. Chlorite in Reykjanes characterizes a higher temperature range of 230° to 280°C (Tomasson and Kristmannsdottir, 1972). Thus a fairly wide range of temperatures have been reported for clay minerals. In DF 45-14, montmorillonite is found in zones at water temperatures of 40° to 80°C. Random mixed-layer clays deficient in montmorillonite exist at water temperatures of 80° to 140°C. Illite occurs in a wide temperature range, but it generally indicates higher temperatures (100°C) and pressures. This is evidenced by its concentration in the interval 8000 to 9000 feet in DF 45-14. The available temperature logs were taken only down to a depth of about 5500 feet in DF 45-14. Chemical geothermometers however indicate a downhole temperature of about 215°C.

The zone of montmorillonite clays in the interval 300 to 1200 feet occurs directly above the sequence of volcanic rocks. The change from Ca/Mg-montmorillonite to Na-montmorillonite corresponds to the transition from alluvium into a tuff unit at 700 to 800 feet. Thus in this interval, the abundance and distribution of montmorillonite is related either to rock composition or to the lithologic contacts which may provide permeable avenues for migrating thermal fluids. Another possible interpretation of the occurrence of montmorillonite clays in this zone is that they may represent vestiges of playa lake sediments, formed when the playa covered a much larger area. A very similar occurrence was observed in DF 66-21 but to a much larger extent. This clay layer appears to be very important as it may serve as a 'cap rock' for the geothermal system. Below this impermeable clay layer in DF 66-21 (below 4700 feet) large quantities of hot water under high pressure were encountered during drilling. From 1100 to 2400 feet, in DF 45-14, the tuff/andesite sequence does not exhibit any appreciable degree of

clay mineral formation. The interval 2300 to 2400 feet, corresponding to the contact between the metasediments and the overlying volcanics, is characterized by a regular mixed-layer chlorite/vermiculite clay. The genetic significance of this type of clay occurrence is uncertain.

The interval 4500 to 5500 feet in DF 45-14 is also anomalous with respect to clay mineral distribution. Electric logs indicate that these intervals have anomalous porosity, suggesting these intervals may be fracture or fault zones. This interval is characterized by high concentrations of random mixed-layer illite-montmorillonite deficient in montmorillonite. Steiner (1968) found that with an increase in temperature the relative amount of expandable layers decreased. Steiner also noted that a temperature inversion occurred below a fracture, characterized by a slight increase in the amount of expandable layers. The presence of illite is indicative of higher temperatures as thermal fluids may be migrating along these postulated fractures.

The interval 5100 to 5500 feet corresponds to the intersection of the range front fault zone projected at depth at an angle of 55 to 60 degrees. This zone is marked by a reduction in the amount of cuttings and an abundance of 'trash' in the cuttings indicating a zone of lost circulation. Thus in this instance, the distribution and type of clay minerals is directly related to temperature and proximity to permeable structures. The random mixed-layer illite-montmorillonite clay minerals give way at greater depths to illite in the interval 8000 to 9000 feet. This is compatible with observations in other geothermal systems where illite is indicative of higher temperatures as noted above.

The limited and highly localized nature of the distribution of the various clay minerals is strong evidence in support of the interpretation that hydrothermal alteration in DF 45-14 is structurally controlled and that the majority of the rocks encountered have not been significantly affected by the hydrothermal activity.

4.2.5.2 Well DF 66-21

Referring to Plate VI, the first 4500 feet in DF 66-21 is characterized by a gradual increase in the amount of random mixed-layer clay. This clay is abundant in the interval 2000 to 4500 feet where it appears to act as an impermeable cap rock. Below 4500 feet the amount of this clay gradually decreases. This clay has been identified by x-ray diff-

reaction as a randomly interstratified mixed-layer illite-montmorillonite with abundant montmorillonite. The $d(001)$ of unglycolated samples ranges from 12.0\AA to 12.45\AA and shifts to a value of 13.8\AA to 14.7\AA suggesting the presence of 35 to 40 percent or more of expandable layers (refer to Figure 4-1 and Table 4-2). The composition of this clay is fairly uniform. It persists throughout the alluvium and into the sequence of volcanic rocks (?). At 5000 feet, corresponding to a change from intercalated tuff-andesite to a homogeneous tuff unit, the type of clay changes to a randomly interstratified mixed-layer illite-montmorillonite with illite dominating. The $d(001)$ of unglycolated samples ranges from 11.8\AA to 12.1\AA and shifts to an average $d(001)$ of 14.0\AA upon glycolation. This same relationship was observed in Wairakei (Steiner, 1968) where the $d(001)$ of mixed layer illite-montmorillonite gradually decreased with an increase in depth and temperature. The random mixed-layer illite-montmorillonite (with illite dominating) in DF 66-21 persists to a depth of 6100 feet where it gives way to a small amount of randomly interstratified mixed-layer illite-montmorillonite with abundant montmorillonite (35 percent) possibly as a result of a temperature inversion below a fracture zone. This clay is found in small amounts in the 6100 to 6300 foot interval.

In the intervals 6400 to 6900 feet and 7100 to 7400 feet and in sporadic occurrences throughout the remainder of the well is a clay mineral which has been identified as a regular mixed-layer chlorite-vermiculite occasionally mixed with a randomly interstratified chlorite-vermiculite or illite-montmorillonite (K. Papke, pers. commun., 1979). The $d(001)$ of this mineral averages about 28.25\AA in unglycolated samples. Successive orders of the basal reflection are an integral sequence of the $d(001)$ reflection. To get a $d(001)$ of 28.2\AA , two 14.1\AA clay minerals must be combined (i.e., chlorite $d(001) = 14.1\text{\AA}$ and vermiculite $d(001) = 14.1\text{\AA}$). Glycolation resulted in a variable shift of only some of these samples indicating the presence of expandable layers probably in a mixed-layer relationship with illite or chlorite. The significance of the regular mixed-layer chlorite-vermiculite is uncertain, although it most likely represents conditions of higher temperature and pressure in the well. Its occurrence also coincides with fracture zones. The relative amount of illite increases dramatically at approximately 6500 feet and predominates from this depth

down to 8800 feet where it begins to gradually decrease. It increases again in the interval 9100 to 9200 feet and is fairly abundant to the total depth of the well at 9780 feet. Thus illite again appears to represent the alteration taking place at higher temperatures and pressures, in association with postulated fracture zones.

The general relationship of the illite and the mixed-layer clay minerals can be expressed as follows, in order of increasing temperature and depth: (1) randomly interstratified mixed-layer illite montmorillonite (with more than 30 percent montmorillonite), (2) randomly interstratified mixed-layer illite-montmorillonite (with less than 30 percent montmorillonite), (3) regular mixed-layer chlorite/vermiculite mixed with randomly interstratified illite-montmorillonite, and (4) discrete illite. Steiner (1968) states that mixed-layer clays with abundant interstratified montmorillonite (1 above) form at low lithostatic and water vapor pressures, whereas comparatively high pressures are required for the formation of illite and mixed-layer clays deficient in montmorillonite (2 and 4 above). This same overall temperature/depth relationship is well documented in other geothermal regions.

Weaver (1956) states that mixed-layer clays in most cases are derived from the degradation or aggradation of pre-existing clay minerals; for example, during weathering, K^+ can be removed from some biotite layers, Mg^{++} and H_2O deposited and a mixed-layer biotite-vermiculite formed; or in sea water the Ca^{++} and H_2O between some montmorillonite layers can be replaced by K^+ and a mixed-layer illite-montmorillonite formed.

The concentration of kaolinite in the interval 4200 to 5400 feet is directly correlated with a zone of increased porosity where abundant water and steam were encountered in the highly fractured and brecciated volcanic rocks (the identification of kaolinite in this interval is certain). It is also associated with the red mixed-layer clay in this interval. The formation of kaolinite at high temperatures (above $150^{\circ}C$) requires relatively acidic conditions whereas montmorillonite is formed under alkaline conditions (Krauskopf, 1967). Thus the occurrence of these two minerals in the interval 4200 to 5400 feet seems contradictory. Kaolinite and montmorillonite may coexist in a very narrow range of temperature-pressure conditions (refer to Figures 4-16, 4-24). This observation may represent the superposition of the alteration effects of the present system on the alteration derived from an earlier system char-

acterized by different chemical conditions. However it is uncertain which of the phases, kaolinite or montmorillonite, is indicative of the present hydrothermal activity.

Evidence from DF 45-14 in conjunction with the observations from DF 66-21 appears to favor the formation of montmorillonite as representing the present system. It is conceivable then that, at some time in the past, more acidic conditions attended by the formation of kaolinite in localized fault zones existed in DF 66-21. It is quite likely that these fracture or fault zones have had a communication with more acidic waters at depth. It is also possible that these acidic waters at depth still exist but that there is no communication with them at the present time. Rather, the avenues of communication have been closed off possibly by deposition of material and/or by tectonic adjustment within the rocks. One other likely alternative is that kaolinite is presently forming from acidic waters in the immediate vicinity of fracture zones but that further away from these zones these acidic waters mix with more alkaline waters thereby making kaolinite unstable.

The interpretation of the alteration mineralogy with regard to kaolinite and chlorite is a problem as a result of the difficulty in distinguishing the two in x-ray diffraction patterns. The distribution pattern of chlorite depicted on Plate V does not exhibit any strong correlations with known or inferred structure or zones of increased porosity as do the other clay minerals. Chlorite does appear though to represent higher temperatures and pressures as it is concentrated in the lower portions of the drill hole. Because of its fairly wide range of stability, chlorite would not seem to be a good indicator of changes in temperature or pressure or alteration conditions.

4.2.6 Water Chemistry

The relationship between the fluid chemistry and the alteration mineralogy in geothermal systems is typically complex and Dixie Valley is no exception. It is evident from an examination of the alteration mineralogy that the chemical flow rate conditions of the hydrothermal system in Dixie Valley have varied through time. The variation of these conditions has resulted in complex mineralogical relationships. The water samples collected from DF 45-14 are considered to be reliable as indicating the chemical conditions at depth. On the other hand, the

reliability of the samples from DF 66-21 must be viewed with discretion as there are holes in the near-surface portions of the casing and the samples were very 'dirty' (D. McMurdie and M. Campana, pers. commun., 1979). Thus the chemical conditions (i.e. pH) of the water samples collected from DF 66-21 are only rough approximations of the conditions at the sampling depth (4700 feet).

In studying the fluid compositions of the two deep wells, 4 samples will be considered; DV-90 and DV-30 from DF 45-14 and DV-80 and DV-93 from DF 66-21. DV-90 is assumed to be representative of deep water conditions in DF 45-14 while DV-30 represents shallow water conditions. Both samples from DF 66-21, DV-80 and DV-93 were collected from a depth of about 4700 feet however the analytical results of these samples are questionable (M. Campana, pers. commun., 1979). The water samples were collected and analyzed by Desert Research Institute personnel. Tables 4-3 through 4-6 list the chemical analyses for these samples. These samples have low to moderate total dissolved solids (TDS) (1900 to 5300 ppb) when compared to geothermal waters in other areas. This suggests that the waters are largely of meteoric origin or that they have been greatly diluted by meteoric water during their migration upward. The waters have apparently achieved their chemical nature largely through interaction with the host rock. The low TDS value may also indicate a relatively short residence time.

The presence of constituents such as lithium, boron, fluorine, arsenic, carbon dioxide and hydrogen sulfide have been suggested in the past as indicating at least a partial magmatic origin of the fluids. More recently though experimental work has shown that concentrations of these elements can occur through normal water-rock interaction at moderate temperatures (Ellis and Mahon, 1977). The high boron content in the Dixie Valley water is not particularly uncommon for geothermal waters, especially within high organic content rocks as those encountered in DF 45-14 and to a lesser extent DF 66-21.

White (1957) has pointed out that within very dense formations sufficient rock/water contact could not occur to maintain the chemical output of a system over its lifetime. For the Wairakei system, Grindley, (1956) showed that over its lifetime (500,000 years) its water would have to have been completely replaced 100 times. Fluid flow through formations of limited permeability (tight, fracture-porosity formations

Table 4-3. Chemical Analysis of Sample DV-90

I	SPECIES		PPM	MOLALITY	LOG MOL	ACTIVITY
1	CA	2	.48496E+01	.12120E-03	-3.9165	.53220E-04
2	MG	2	.84886E-02	.34973E-06	-6.4563	.15628E-06
3	NA	1	.35839+03	.15615E-01	-1.8065	.12621E-01
4	K	1	.39987E+02	.10243E-02	-2.9896	.82217E-03
64	H	1	.19079E-03	.18959E-06	-6.7222	.15849E-06
5	CL	-1	.49287E+03	.13925E-01	-1.8562	.11177E-01
6	SO4	-2	.40033E+00	.41743E-05	-5.3794	.18185E-05
7	HCO3	-1	.44848E+02	.73623E-03	-3.1330	.60024E-03
18	CO3	-2	.50945E-02	.85036E-07	-7.0704	.37571E-07
86	H2CO3	0	.64823E+03	.10468E-01	-1.9801	.10531E-01
27	OH	-1	.74611E+01	.43942E-03	-3.3571	.35248E-03
62	F	-1	.75984E+01	.40061E-03	-3.3973	.32135E-03
19	MGOH	1	.43867E-02	.10634E-06	-6.9733	.87426E-07
23	MGSO4 AQ	0	.15420E-04	.12831E-09	-9.8917	.12884E-09
22	MGHCO3	1	.21796E-02	.25586E-07	-7.5920	.20617E-07
21	MGC03 AQ	0	.43908E-02	.52159E-07	-7.2827	.52372E-07
20	MGF	1	.36344E-02	.84055E-07	-7.0754	.68025E-07
29	CAOH	1	.70743E-01	.12413E-05	-5.9061	.10167E-05
32	CASO4 AQ	0	.68908E-02	.50699E-07	-7.2950	.50906E-07
30	CAHCO3	1	.11779E+02	.11671E-03	-3.9329	.95593E-04
31	CAC03 AQ	0	.36282E+02	.36310E-03	-3.4400	.36458E-03
44	NASO4	-1	.26616E+03	.22394E-02	-2.6499	.18257E-02
69	NA2SO4	0	.12514E-03	.88249E-09	-9.0543	.88609E-09
43	NAHCO3	0	.33201E+00	.39595E-05	-5.4024	.39757E-05
42	NAC03	-1	.15273E+00	.18432E-05	-5.7344	.15027E-05
49	NA2CO3	0	.29631E-05	.28003E-10	-10.5528	.28118E-10
94	NACL	0	.20495E+00	.35126E-05	-5.4544	.35270E-05
46	KS04	-1	.12745E-01	.94451E-07	-7.0248	.77004E-07
95	KCL	0	.17712E-01	.23796E-06	-6.6235	.23893E-06
63	HSO4	-1	.15942E-02	.16450E-07	-7.7833	.13313E-07
96	H2SO4	0	.44544E-15	.45492E-20	-20.3421	.45678E-20
93	HCL	0	.86357E-05	.23724E-09	-9.6248	.23821E-09
24	H4SI04AQ	0	.50928E+03	.53074E-02	-2.2751	.53291E-02
25	H2SI04	-1	.10447E+02	.11002E-03	-3.9585	.88655E-04
26	H2SI04	-2	.50997E-01	.54285E-06	-6.2653	.23984E-06
8	FE	2	.16100E+01	.28877E-04	-4.5395	.12997E-04
51	AL	3	.12522E-17	.46486E-22	-22.3327	.92667E-23
52	ALOH	2	.11694E-11	.26627E-16	-16.5747	.11765E-16
53	AL(OH)2	1	.14773E-06	.24260E-11	-11.6151	.19779E-11
54	AL(OH)4	-1	.35213E+00	.37124E-05	-5.4303	.30044E-05
55	ALF	2	.31660E-14	.68970E-19	-19.1613	.30472E-19
56	ALF2	1	.17570E-06	.27085E-11	-11.5673	.22082E-11
57	ALF3	0	.13526E-10	.16134E-15	-15.7923	.16200E-15
58	ALF4	-1	.65881E-12	.64083E-17	-17.1933	.51862E-17
59	ALSO4	1	.17806E-19	.14496E-24	-24.8388	.11731E-24
60	AL(SO4)2	-1	.75821E-23	.34663E-28	-28.4601	.28052E-28
36	H3BO3 AQ	0	.48620E+02	.78761E-03	-3.1037	.79083E-03
81	LI	1	.99965E+00	.14430E-03	-3.8407	.11819E-03
82	LIOH	0	.35696E-01	.14931E-05	-5.8259	.14992E-05
83	LIS04	-1	.11872E-03	.11545E-08	-8.9376	.93823E-09
88	SR	2	.10572E+01	.12086E-04	-4.9177	.52715E-05
89	SROH	1	.33177E-02	.31763E-07	-7.4981	.25813E-07

TDS=2493

Field pH=6.8

Temperature=210°C

Table 4-4. Chemical Analysis of Sample DV-30

I	SPECIES		PPM	MOLALITY	LOG MOL	ACTIVITY
1	CA	2	.12096E+03	.30238E-02	-2.5194	.15347E-02
2	MG	2	.22001E+02	.90666E-03	-3.0426	.46847E-03
3	NA	1	.36560E+03	.15933E-01	-1.7977	.13370E-01
4	K	1	.18356E+02	.47034E-03	-3.3276	.39125E-03
64	H	1	.14996E-03	.14906E-06	-6.8266	.12882E-06
5	CL	-1	.57484E+03	.16245E-01	-1.7893	.13514E-01
6	SO4	-2	.19579E+03	.20420E-02	-2.6899	.10216E-02
7	HCO3	-1	.19462E+03	.31957E-02	-2.4954	.26977E-02
18	CO3	-2	.18287E+00	.30532E-05	-5.5152	.15505E-05
86	H2CO3	0	.43660E+02	.70525E-03	-3.1517	.71001E-03
27	OH	-1	.22192E-01	.13073E-05	-5.8836	.10863E-05
62	F	-1	.40840E+01	.21538E-03	-3.6668	.17898E-03
19	MGOH	1	.14983E-01	.36330E-06	-6.4397	.30913E-06
23	MGSO4 AQ	0	.14637E+02	.12183E-03	-3.9143	.12270E-03
22	MGHCO3	1	.20723E+01	.24333E-04	-4.6138	.20309E-04
21	MGCO3 AQ	0	.13993E+00	.16626E-05	-5.7792	.16746E-05
20	MGF	1	.71891E+00	.16631E-04	-4.7791	.13940E-04
29	CAOH	1	.35595E-02	.62471E-07	-7.2043	.52970E-07
32	CASO4 AQ	0	.56750E+02	.41765E-03	-3.3792	.42065E-03
30	CAHCO3	1	.17118E+02	.16965E-03	-3.7705	.14385E-03
31	CACO3 AQ	0	.13494E+01	.13508E-04	-4.8694	.13605E-04
44	NASO4	-1	.20044E+03	.16868E-02	-2.7729	.14240E-02
69	NA2SO4	0	.40590E+00	.28631E-05	-5.5432	.28837E-05
43	NAHCO3	0	.15755E+01	.18794E-04	-4.7260	.18929E-04
42	NACO3	-1	.11817E+00	.14264E-05	-5.8457	.12041E-05
49	NA2CO3	0	.13678E-03	.12930E-08	-8.8884	.13023E-08
94	NACL	0	.26163E+00	.44853E-05	-5.3482	.45175E-05
46	KSO4	-1	.82496E+00	.61151E-05	-5.2136	.51621E-05
95	KCL	0	.10157E-01	.13650E-06	-6.8649	.13748E-06
63	HSO4	-1	.47326E-02	.48848E-07	-7.3112	.40943E-07
96	H2SO4	0	.16478E-12	.16833E-17	-17.7738	.16954E-17
93	HCL	0	.19774E-08	.54338E-13	-13.2649	.54729E-13
24	H4SI04AQ	0	.16706E+03	.17414E-02	-2.7591	.17539E-02
25	H3SI04	-1	.89482E+00	.94265E-05	-5.0257	.78678E-05
26	H2SI04	-2	.82756E-03	.88114E-08	-8.0550	.44745E-08
8	FE	2	.40000E-01	.71761E-06	-6.1441	.37093E-06
51	AL	3	.16146E-09	.59957E-14	-14.2222	.16130E-14
52	ALOH	2	.22338E-06	.50877E-11	-11.2935	.25836E-11
53	AL(OH)2	1	.23584E-03	.38738E-08	-8.4119	.32701E-08
54	AL(OH)4	-1	.34991E+00	.36899E-05	-5.4330	.30928E-05
55	ALF	2	.26696E-06	.58172E-11	-11.2353	.29541E-11
56	ALF2	1	.11578E-02	.17853E-07	-7.7483	.15071E-07
57	ALF3	0	.13201E-03	.15749E-08	-8.8027	.15862E-08
58	ALF4	-1	.10650E-04	.10363E-09	-9.9845	.86856E-10
59	ALSO4	1	.60138E-09	.48969E-14	-14.3101	.41045E-14
60	AL(SO)2	-1	.10137E-09	.46353E-15	-15.3339	.38852E-15
36	H3BO3 AQ	0	.26723E+02	.43301E-03	-3.3635	.43612E-03
	H2BO3	-1	.15788E+00	.26006E-05	-5.5849	.21405E-05
81	LI	1	.97064E+00	.14015E-03	-3.8534	.11883E-03
82	LI0H	0	.12605E-04	.52738E-09	-9.2779	.53117E-09
83	LISO4	-1	.64736E-01	.62970E-06	-6.2009	.52993E-06
88	SR	2	.32000E+01	.36591E-04	-4.4366	.18353E-04
89	SROH	1	.20513E-04	.19643E-09	-9.7068	.16531E-09

Field pH=6.89

TDS=1903

Temperature=64.5°C

Table 4-5. Chemical Analysis of Sample DV-93

I	SPECIES		PPM	MOLALITY	LOG MOL	ACTIVITY
1	CA	2	.72161E+-0	.18192E-04	-4.7423	.55761E-05
2	MG	2	.74068E+00	.30632E-04	-4.5138	.98537E-05
3	NA	1	.15834E+04	.69249E-01	-1.1596	.50809E-01
4	K	1	.43966E+02	.11305E-02	-2.9467	.81199E-03
64	H	1	.12737E-04	.12705E-07	-7.8960	.10000E-07
5	CL	-1	.17184E+04	.48734E-01	-1.3122	.35004E-01
6	SO4	-2	.11898E+01	.12453E-04	-4.9047	.37114E-05
7	HCO3	-1	.69581E+03	.11465E-01	-1.9406	.8548E-02
18	CO3	-2	.46544E+01	.77982E-04	-4.1080	.24101E-04
86	H2CO3	0	.10141E+03	.16439E-02	-2.7841	.16764E-02
27	DH	-1	.31589E+02	.18675E-02	-2.7288	.13380E-02
62	F	-1	.78380E+01	.41480E-03	-3.3822	.19720E-03
19	MGOH	1	.90011E+00	.21903E-04	-4.6595	.16644E-04
23	MGSO4 AQ	0	.17047E-02	.14239E-07	-7.8465	.14474E-07
22	MGHCO3	1	.75992E+00	.89542E-05	-5.0480	.64898E-05
21	MGCO3 AQ	0	.84398E+01	.10064E-03	-3.9972	.10230E-03
20	MGF	1	.14148E+00	.32844E-05	-5.4835	.24061E-05
29	CAOH	1	.26825E-01	.47245E-06	-6.3256	.35604E-06
32	CASO4 AQ	0	.12351E-02	.91216E-08	-8.0399	.92720E-08
30	CAHCO3	1	.47974E+01	.47711E-04	-4.3214	.35955E-04
31	CACO2 AQ	0	.53334E+02	.53577E-03	-3.2710	.54460E-03
44	NASO4	-1	.48974E+03	.41361E-02	-2.3834	.30839E-02
69	NA2SO4	0	.60239E-02	.42640E-07	-7.3702	.43344E-07
43	NAHCO3	0	.18733E+02	.22425E-03	-3.6493	.22795E-03
42	NACO3	-1	.16564E+03	.20065E-02	-2.6976	.14961E-02
49	NA2CO3	0	.30318E-01	.28760E-06	-6.5412	.29235E-06
94	NACL	0	.25429E+01	.43748E-04	-4.3590	.44469E-04
46	KSO4	-1	.20125E-01	.14971E-06	-6.8248	.11162E-06
95	KCL	0	.53912E-01	.72705E-06	-6.1384	.73904E-06
63	HSO4	-1	.44368E-04	.45956E-09	-9.3577	.33667E-09
96	H2SO4	0	.35616E-17	.36512E-22	-22.4375	.37114E-22
93	HCL	0	.22891E-06	.63124E-11	-11.1098	.64165E-11
24	H4SI04AQ	0	.20691E+03	.21644E-02	-2.6647	.22001E-02
25	H2SI04	-1	.77758E+02	.82202E-03	-3.0651	.59578E-03
26	H2SI04	-2	.16486E+02	.17625E-03	-3.7539	.54472E-04
51	AL	3	.34627-20	.12904E-24	-24.8893	.14965E-25
52	ALOH	2	.82517E-14	.18841E-18	-16.7233	.58290E-19
53	AL(OH)2	1	.37448E-08	.61728E-13	-13.2095	.46025E-13
54	AL(OH)4	-1	.35213E+00	.37264E-05	-5.4287	.27299E-05
55	ALF	2	.67344E-17	.14726E-21	-21.8319	.45512E-22
56	ALF2	1	.31126E-10	.48163E-15	-15.3173	.35911E-15
57	ALF3	0	.13015E-13	.15583E-16	-16.6074	.15840E-18
58	ALF4	-1	.85666E-15	.83643E-20	-20.0776	.61276E-20
59	ALSO4	1	.50555E-22	.41311E-27	-27.3639	.30264E-27
60	AL(SO4)2	-1	.40418E-25	.18547E-30	-30.7317	.13588E-30
36	H3BO3 AQ	0	.16682E+02	.27126E-03	-3.5666	.27573E-03
37	H2BO3	-1	.55050E+02	.90997E-03	-3.0410	.63549E-03

TDS=5310

Field pH=8.0

Temperature=165°C

Table 4-6. Chemical Analysis of Sample DV-80

I	SPECIES		PPM	MOLALITY	LOG MOL	ACTIVITY
1	CA	2	.58474E+00	.14653E-04	-4.8341	.44836E-05
2	MG	2	.44489E+00	.18379E-04	-4.7357	.64531E-05
3	NA	1	.12083E+04	.52785E-01	-1.2775	.39576E-01
4	K	1	.30984E+02	.79583E-03	-3.0992	.58651E-03
64	H	1	.12586E-04	.12541E-07	-7.9017	.10000E-07
5	CL	-1	.12071E+04	.34196E-01	-1.4660	.25202E-01
6	SO4	-2	.37303E+00	.39002E-05	-5.4039	.12671E-05
7	HCO3	-1	.82894E+03	.13644E-01	-1.8650	.10380E-01
18	CO3	-2	.46174E+01	.77279E-04	-4.1119	.25887E-04
86	H2CO3	0	.15291E+03	.24759E-02	-2.6063	.25145E-02
27	OH	-1	.37906E+02	.22385E-02	-2.6500	.16466E-02
19	MGOH	1	.72631E+00	.17654E-04	-4.7531	.13652E-04
23	MGSO4 AQ	0	.38458E-03	.32088E-08	-8.4937	.32494E-08
22	MGHCO3	1	.66015E+00	.77702E-05	-5.1096	.57710E-05
21	MGC03 AQ	0	.86347E+01	.10255E-03	-3.9878	.10415E-03
29	CAOH	1	.28939E-01	.50912E-06	-6.2932	.39089E-06
32	CASO4 AQ	0	.37985E-03	.28023E-08	-8.5525	.28377E-08
30	CAHCO3	1	.60242E+01	.59848E-04	-4.2230	.45949E-04
31	CAC03 AQ	0	.81801E+02	.82083E-02	-3.0857	.83122E-03
44	NASO4	-1	.16074E+03	.13560E-02	-2.8677	.10316E-02
69	NA2SO4	0	.11848E-02	.83772E-08	-8.0769	.84832E-08
43	NAHCO3	0	.17804E+02	.21290E-03	-3.6718	.21560E-03
42	NACO3	-1	.15612E+03	.18892E-02	-2.7237	.14372E-02
49	NA2CO3	0	.19354E-01	.18814E-06	-6.7255	.19052E-06
94	NACL	0	.14330E+01	.24627E-04	-4.6086	.24938E-04
46	KSO4	-1	.51080E-02	.37955E-07	-7.4207	.28875E-07
95	KCL	0	.28173E-01	.37953E-06	-6.4208	.38433E-06
63	HSO4	-1	.18357E-04	.18994E-09	-9.7214	.14235E-09
96	H2SO4	0	.12219E-17	.12513E-22	-22.9026	.12671E-22
93	HCL	0	.22112E-06	.60910E-11	-11.2153	.61681E-11
24	H2SIO3AQ	0	.23207E+03	.24250E-02	-2.6153	.24557E-02
25	H3SIO4	-1	.86151E+02	.90977E-03	-3.0411	.67570E-03
26	H2SIO4	-2	.16441E+02	.17548E-03	-3.7558	.58784E-04
36	H3BO3 AQ	0	.62920E+02	.10220E-02	-2.9905	.10349E-02
81	LI	1	.12538E+01	.18147E-03	-3.7412	.13933E-03
82	LIOH	0	.12495E+00	.52408E-05	-5.2806	.53071E-05
83	LISO4	-1	.10456E-03	.10196E-08	-8.9918	.77-66E-09

TDS=4305

Field pH=8.0

Temperature=171°C

as encountered in DF 45-14 and DF 66-21) would be expected to rather rapidly (geologically) deplete the available constituents from the exposed rock. It might be postulated then that for the waters in Dixie Valley, some magmatic additions are responsible for the relatively high boron, fluorine, lithium and arsenic levels reported. However it must be considered that the water flow and composition have not been constant over time. Periods of appreciable inflow and outflow may occur over relatively short periods of time when tectonic activity creates new flow channels. Flow may be drastically reduced or cease altogether when these channels become sealed again by deposition of minerals or by deformations within the rocks. Waxing and waning of the geothermal flow systems at Dixie Valley is supported by evidence of increased flows of quite different character at many of the springs, such as the tufa domes at Sou and Hyder Hot Springs. Evidence for the chemical variation of these geothermal fluids through time is also reflected in the hydrothermal alteration. Even though the origin of chemicals in the Dixie Valley water is not completely clear, local water-rock interactions appear to be the dominant control in determining well, spring, and fumarole fluid compositions.

The hydrothermal alteration in Dixie Valley involves a complex interaction of processes, including: devitrification, recrystallization, solution, and deposition. The end product of these processes is dependent upon a number of factors such as temperature, pressure, fluid compositions, original rock compositions, time of reaction, rates of fluid flow, permeability of the rocks and the type of permeability (fissure or interstitial porosity). The concentration of carbon dioxide and hydrogen sulfide in the waters also has an important control on the type of secondary mineralogy. However their concentrations in the water samples are uncertain. In Dixie Valley, the fluid compositions, flow rates and rock compositions and permeabilities exhibit considerable variation which contributes to the range of observed alteration. Because of these factors and the very limited extent of subsurface sampling (2 deep wells), the vertical zoning, characteristic of many geothermal fields, is somewhat obscure in Dixie Valley. Permeability is a very important factor since many mineralogical changes are not isochemical and any fluid present must be free to move, allowing an influx and outflux of chemical components. The water may impose a

new composition on the rock if a compositional or temperature gradient exists along the direction of water flow.

In New Zealand it was found that dense impermeable rocks such as fine-grained sediments and ignimbrites were little altered by near-neutral pH solutions, even at high temperatures. Breccias and pumice, however, were altered to an equilibrium assemblage (Ellis and Mahon, 1977). This type of effect is seen in the drill cuttings from both DF 45-14 and DF 66-21. Much of the metasiltstone/metashale, quartz arenite and basic intrusive rocks are only weakly altered, while the fractured and brecciated volcanics in the upper portions of both wells have been extensively altered. Significantly, fracture zones (zones of increased permeability indicated by lost circulation during drilling, by brecciation of the drill cuttings, and by the formation-density, neutron, and gamma-ray logs) within the various lithologies show extensive alteration, as would be expected for areas of high permeability.

In considering the alteration mineralogy and its relationship to fluid chemistry, the question of equilibrium is important. Given stable chemical conditions and temperature, it could be expected that the rock-water system would adjust to a new set of equilibrium conditions and minerals. From experimental silicate reaction rates, zones of slow water movement at temperatures above 200°C would probably produce an equilibrium mineral assemblage. However, in other cases equilibrium between rock and water may not be attained for reasons such as low rock permeability, rapid steam or water flow, boiling of water at particular depths, or condensation of steam into a water phase. The solubilities of the common rock-forming constituents (Si, Al, Na, K, Ca, Mg, Fe, Mn) are limited by particular mineral-water equilibria, which in all cases favor the retention of the elements in the mineral phases (Ellis and Mahon, 1977). In the two deep exploratory wells the flow rates are unknown and the downhole temperatures estimated from SiO₂ and Na/K geothermometry are approximately 190°C to 215°C for DF 45-14 and 165°C to 175°C for DF 66-21.

In correlating the alteration mineralogy to water chemistry, the computer program WATEQ (Truesdell and Jones, 1974) has been used to provide activities and saturation index data as determined from the laboratory water analyses. The following discussion is based upon the method of Helgeson (1969) for deriving phase diagrams expressing mineral stab-

ilities by ion concentration ratios in solution.

In the presence of quartz, with aluminum assumed fixed, the formation of various silicates depends on the activity of metal oxides in solution. For example the activity of CaO may be expressed as follows:

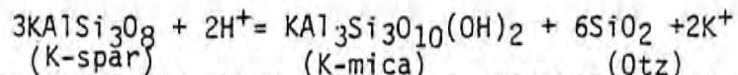


$$K_a = a_{\text{Ca}^{++}} / (a_{\text{H}^+})^2 \cdot a_{\text{CaO}} \quad ; \quad \text{with } a_{\text{H}_2\text{O}} \text{ assumed} = 1$$

$$\text{Thus, } \log a_{\text{CaO}} = \log (a_{\text{Ca}^{++}} / (a_{\text{H}^+})^2) - \log K_{\text{Ca}}$$

Since $\log K_{\text{Ca}}$ is a constant, the variation in CaO activity may be expressed in terms of the variation in the ratio $a_{\text{Ca}^{++}}/(a_{\text{H}^+})^2$ and similar expressions may be derived for other oxides.

Correlations can now be made between solution composition and mineral equilibrium using ion activities:



The equilibrium constant for the above reaction is directly related to the ratio of potassium ion activity to hydrogen activity.

The diagrams which follow (Figures 4-16 through 4-24) represent the thermodynamically stable phase for a given temperature and pressure, assuming quartz saturation but not amorphous silica saturation. The points plotted illustrate the migration of the fluid composition and stability fields with changing pH and temperature conditions. It should be noted that a fluid composition within a stability field for a given mineral phase does not imply its presence or indicate any quantity which may be present, merely that it should be a stable phase for the specified conditions. Other factors such as element availability, ability of a phase to nucleate, and the kinetics of grain growth must also be considered.

One of the major problems associated with the interpretation of these stability diagrams is the uncertainty of the pH at depth. Because of the combined effects of temperature, fluid mixing and degassing (especially CO₂) and/or boiling, the pH at the production depth may be considerably lower than the measured field pH. An increase in temperature with depth, such as that found in both wells, may lower the pH by one and as much as two pH units (R. Jacobson, pers. commun., 1979). As a generalization, the pH is controlled by the salinity of the waters and aluminosilicate equilibria involving hydrogen and alkali metal ions

(Ellis and Mahon, 1977). Figure 4-15 was used to provide an estimation of the pH in each of the wells using temperatures derived from chemical geothermometers and calculated salinities of the water samples. Temperatures of 170°C and 200°C for DF 66-21 and DF 45-14 yield pH values of 7.5-7.6 and 7.6-7.7, respectively (Figure 4-15). These values compare with measured field pH values of 8.0 for DV-93 (DF 66-21) and 6.8 for DV-90 (DF 45-14). The discrepancies between these pH values are likely an indication of fluid mixing wherein the salinities used in Figure 4-14 would not be truly representative. A water sample recently taken from DF 66-21, considered to be fairly reliable, yielded a pH of 5.8 (M.E. Campana, pers. commun., 1979).

Another mechanism which explains the apparent discordance between the measured pH and that determined from Figure 4-15 for DF 45-14 is the rise in pH associated with the release of CO₂ during boiling. Examination of the equation below illustrates this process.



As CO₂ is driven off (vaporizes), the equilibrium constraints force the reaction to proceed to the right, thereby consuming H⁺ (acid) in the process. CO₂ may also have been lost during sampling (R. Jacobson, pers. commun., 1979). Because of the uncertainty in the true down-hole pH in both wells, it is not totally clear which mineral phases represent an equilibrium assemblage. However it does appear that the downhole pH is slightly lower than the measured field pH in both wells.

At temperatures of 150°C and 200°C and the measured field pH, none of the samples are near equilibrium with kaolinite (Figures 4-16 through 4-24). This is particularly significant since kaolinite was prominent in localized intervals associated with postulated fracture zones in both wells (Plates V and IV). Considerably lower than field pH conditions (5.0 or less) or lower temperature conditions (100°C or less) would be required for kaolinite to exist as a stable phase. The presence of kaolinite together with montmorillonite is only possible in a very narrow range of physical and chemical conditions. It may be postulated that the presence of kaolinite represents a relict of a previous period of more acidic conditions and/or lower temperature conditions. The occurrence of kaolinite in the immediate vicinity of fault zones may indicate that the faults are in or have been in communication with

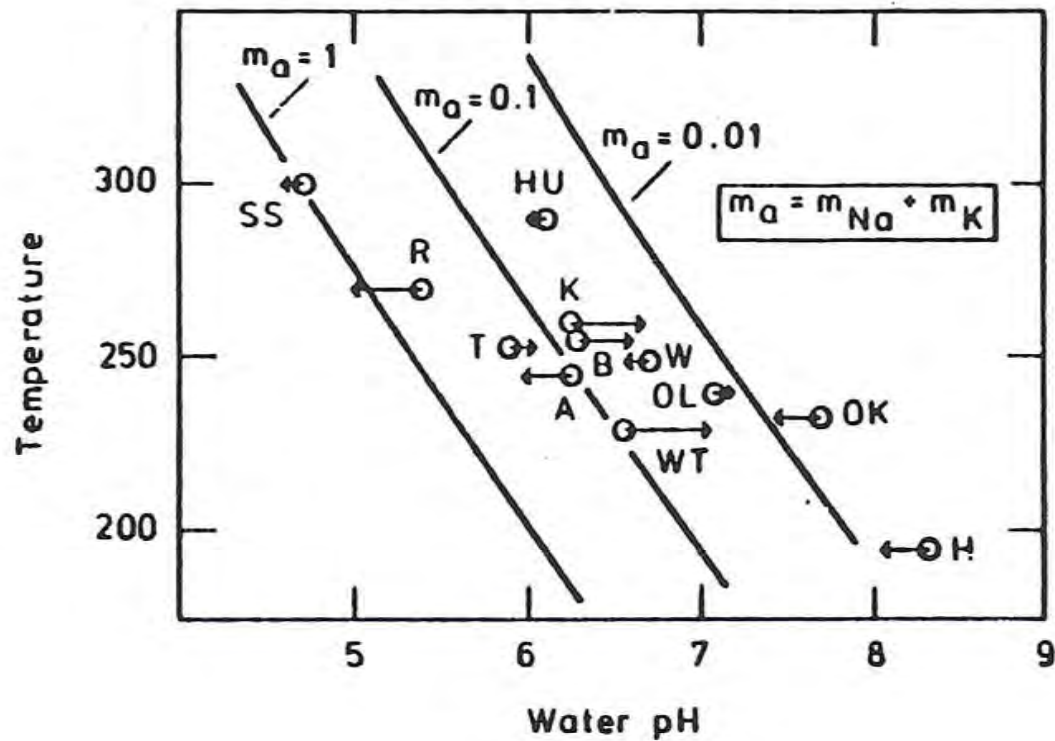
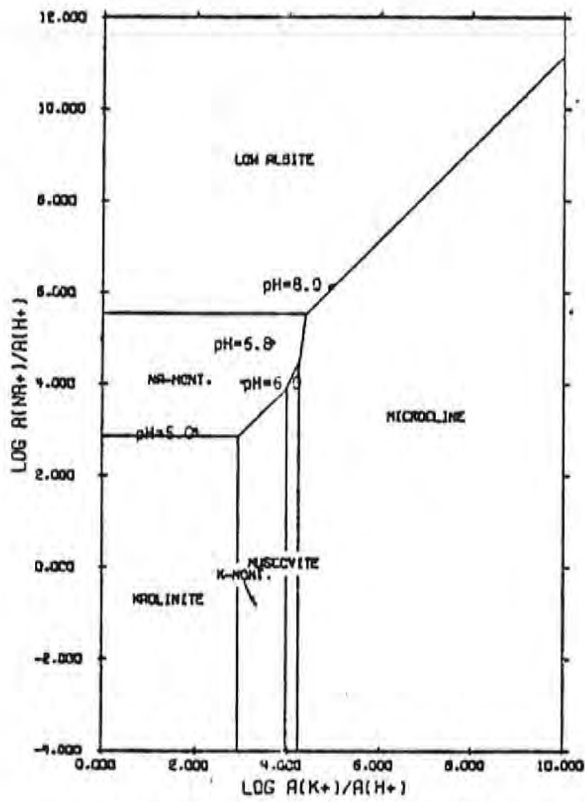
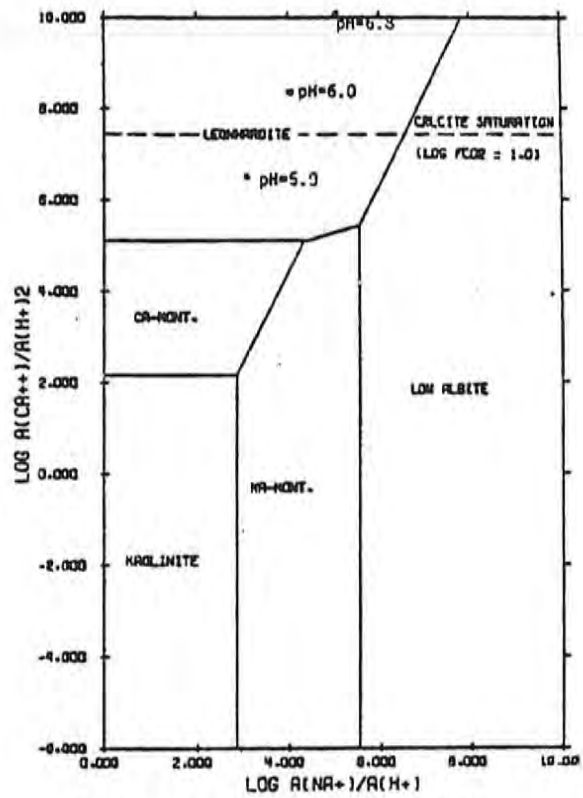


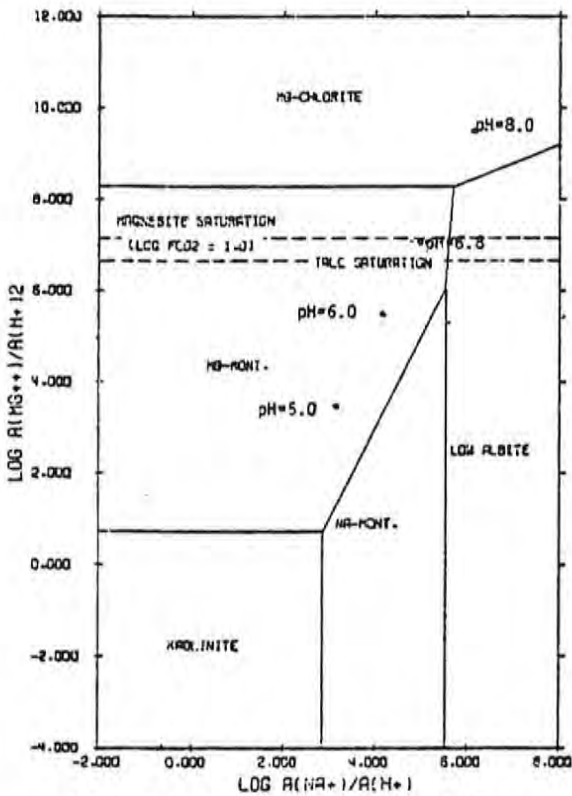
Figure 4-15. The calculated trends of pH with temperature for waters of three salinities. Open circles show pH of waters in geothermal systems; connected arrowheads indicate pH calculated for the salinity of the water. SS - Salton Sea; R - Reykjanes; H - Hveragerdi; HU - Hatchobaru; T - El Tatio; OL - Olkaria; A - Ahuachapan; W - Wairakei; K - Kawerau; B - Broadlands; WT - Waiotapu; OK - Orakeikorako. (from Ellis and Mahon, 1977)



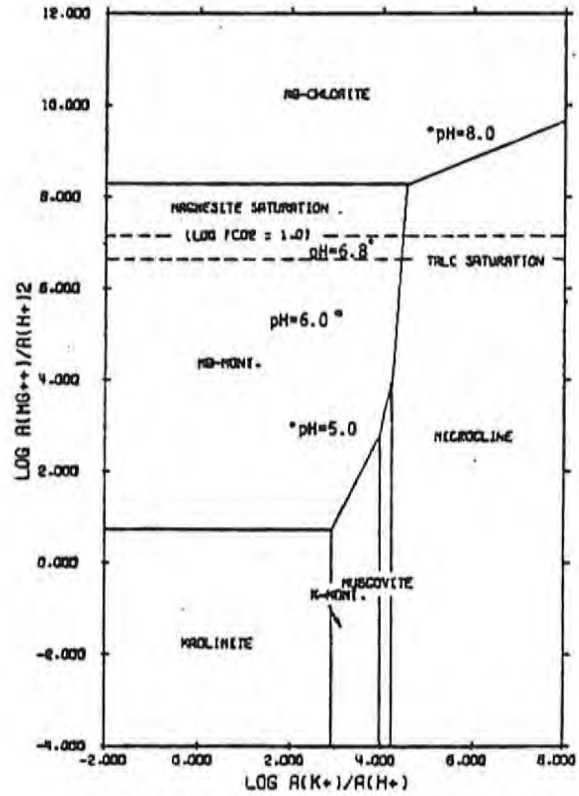
The System $HCl-H_2O-Al_2O_3-K_2O-Na_2O-SiO_2$ at $200^\circ C$: $\log a_{quartz} = -2.35 = \text{quartz saturation}$.



The System $HCl-H_2O-Al_2O_3-CaO-CO_2-Na_2O-SiO_2$ at $200^\circ C$: $\log a_{quartz} = -2.35 = \text{quartz saturation}$.

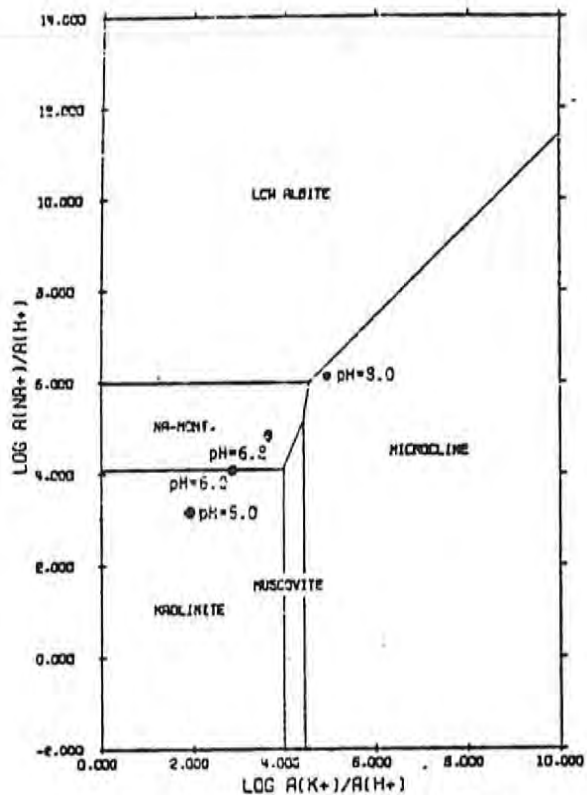


The System $HCl-H_2O-Al_2O_3-CO_2-MgO-Na_2O-SiO_2$ at $200^\circ C$: $\log a_{quartz} = -2.35 = \text{quartz saturation}$.

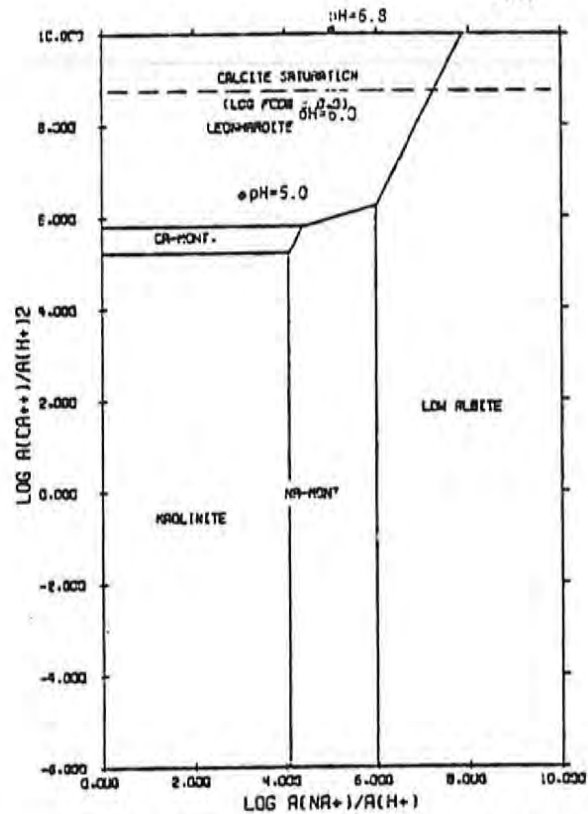


The System $HCl-H_2O-Al_2O_3-CO_2-K_2O-MgO-SiO_2$ at $200^\circ C$: $\log a_{quartz} = -2.35 = \text{quartz saturation}$.

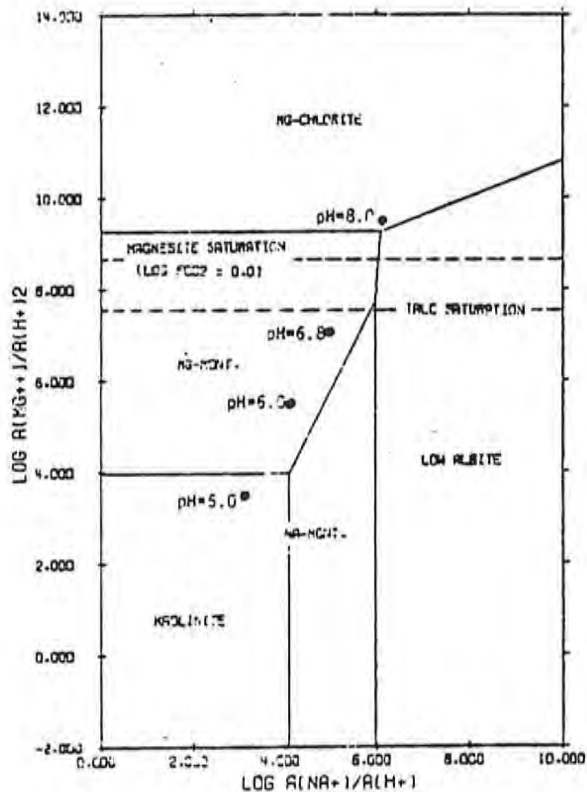
Figure 4-16. Phase stability diagrams for Sample DV-90 at $200^\circ C$.



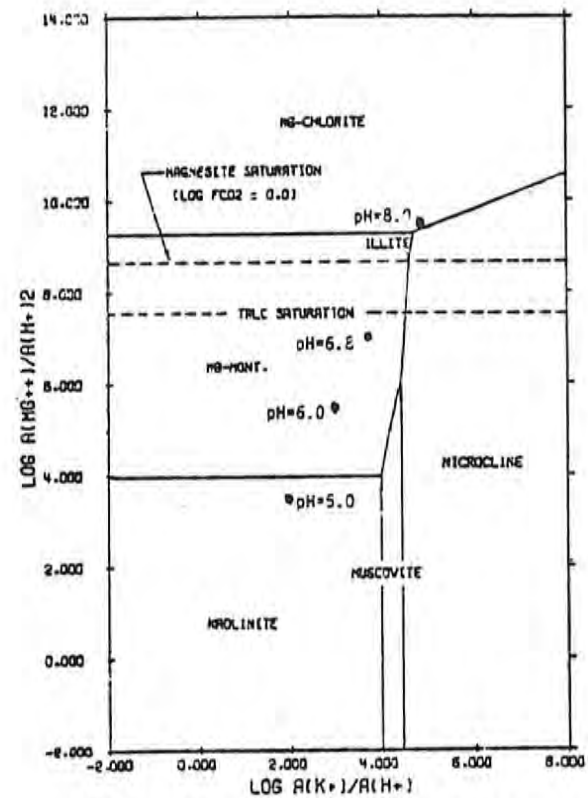
The System $HCl-H_2O-Al_2O_3-K_2O-Na_2O-SiO_2$ at $150^\circ C$; $\log a_{H_2SiO_3} = -2.67 = \text{quartz saturation}$.



The System $HCl-H_2O-Al_2O_3-CaO-CO_2-Na_2O-SiO_2$ at $150^\circ C$; $\log a_{H_2SiO_3} = -2.67 = \text{quartz saturation}$.

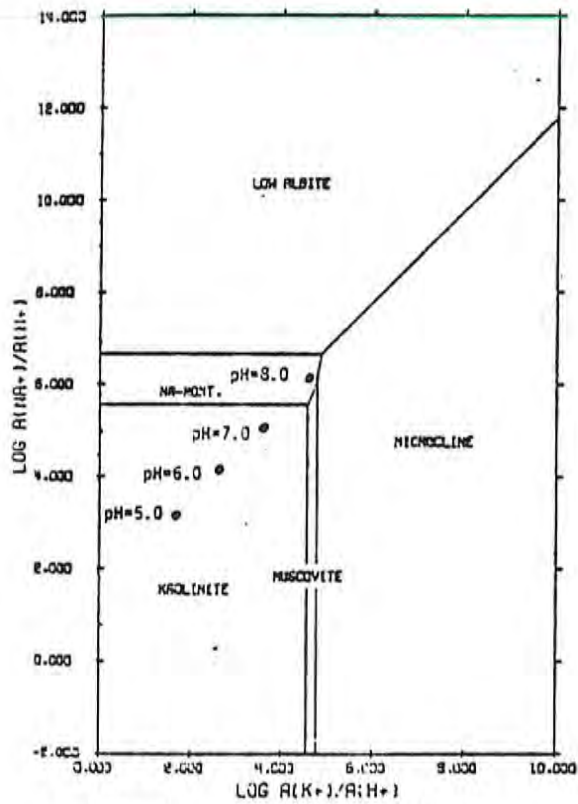


The System $HCl-H_2O-Al_2O_3-CO_2-MgO-Na_2O-SiO_2$ at $150^\circ C$; $\log a_{H_2SiO_3} = -2.67 = \text{quartz saturation}$.

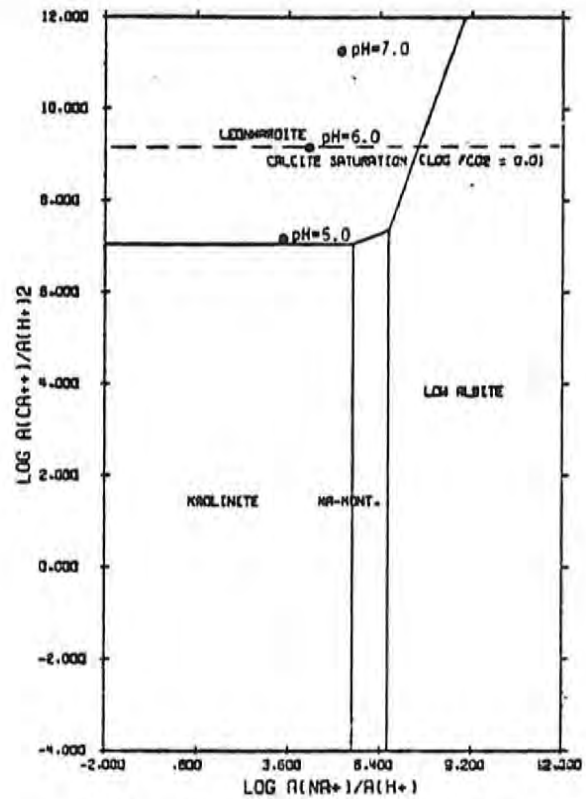


The System $HCl-H_2O-Al_2O_3-CO_2-K_2O-MgO-SiO_2$ at $150^\circ C$; $\log a_{H_2SiO_3} = -2.67 = \text{quartz saturation}$.

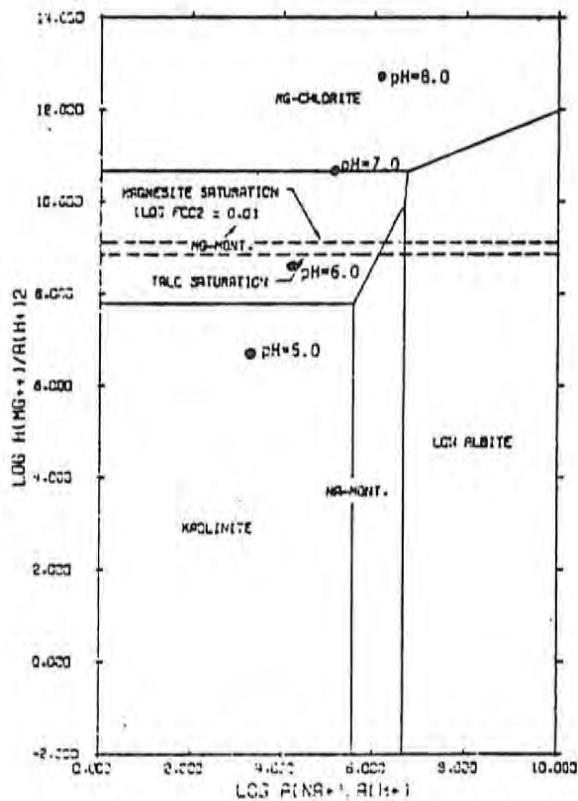
Figure 4-17. Phase stability diagrams for Sample DV-90 at $150^\circ C$.



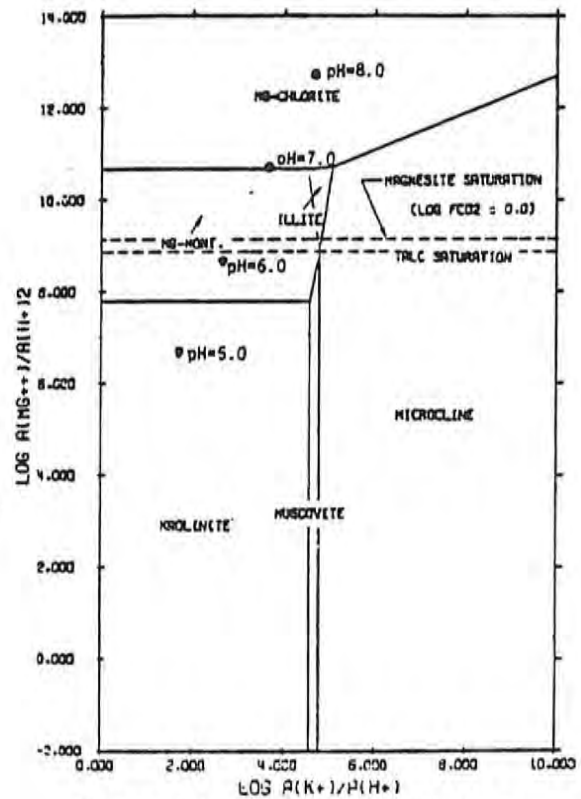
The System HCl-H₂O-Al₂O₃-K₂O-Na₂O-SiO₂ at 100°C; log a_{HCl} = -3.08 = quartz saturation.



The System HCl-H₂O-Al₂O₃-CaO-CO₂-Na₂O-SiO₂ at 100°C; log a_{HCl} = -3.08 = quartz saturation.

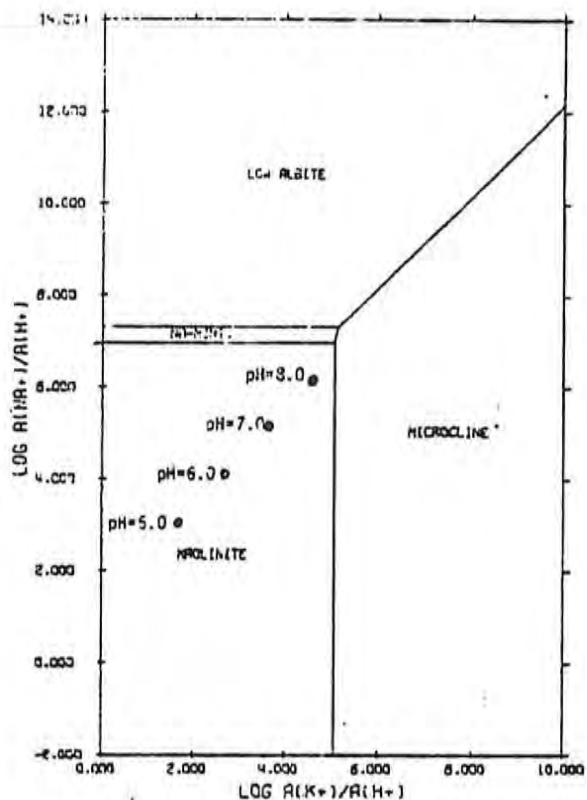


The System HCl-H₂O-Al₂O₃-CO₂-MgO-Na₂O-SiO₂ at 100°C; log a_{HCl} = -3.08 = quartz saturation.

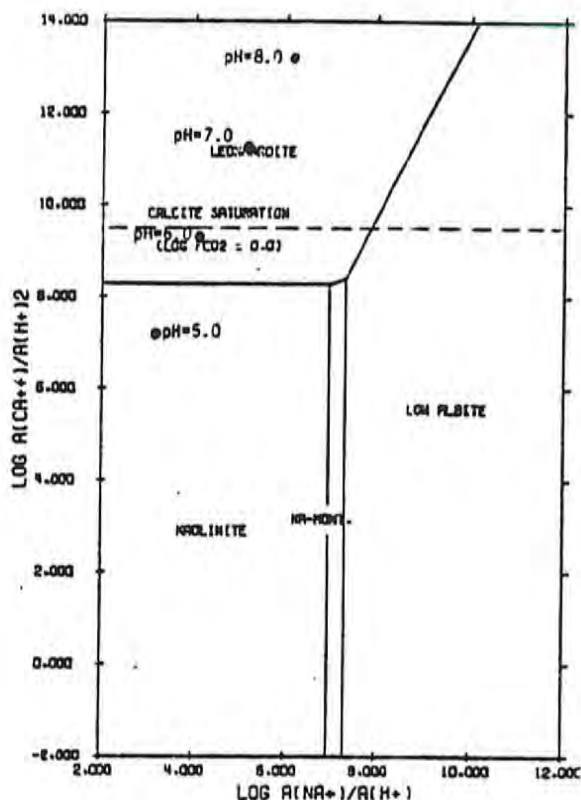


The System HCl-H₂O-Al₂O₃-CO₂-K₂O-MgO-SiO₂ at 100°C; log a_{HCl} = -3.08 = quartz saturation.

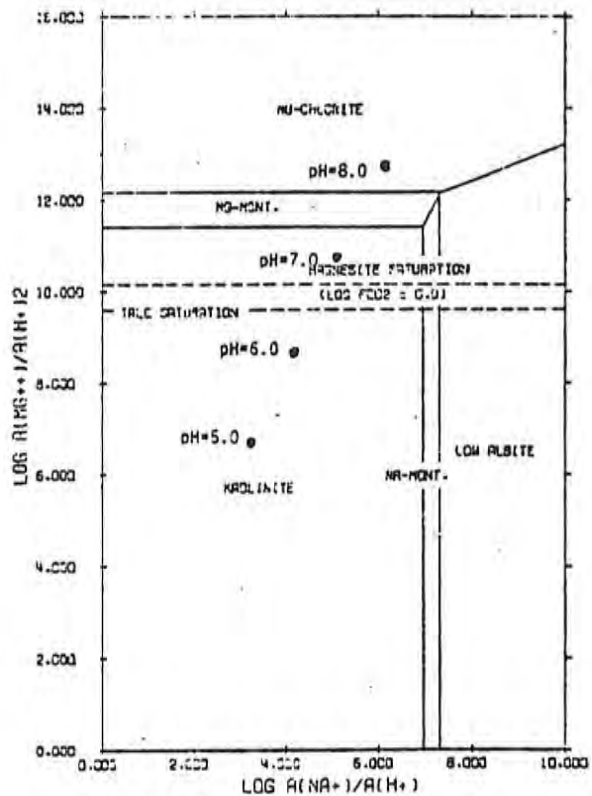
Figure 4-18. Phase stability diagrams for Sample DV-30 at 100°C.



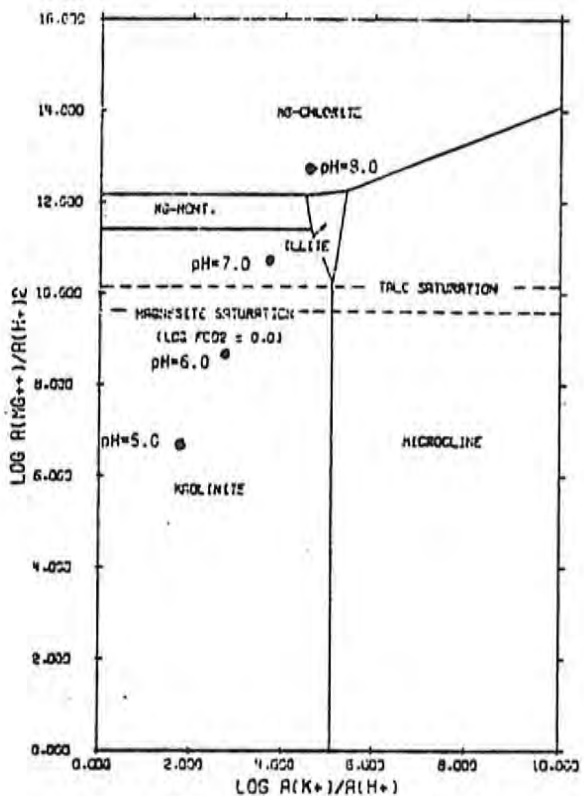
The System HCl-H₂O-Al₂O₃-K₂O-Na₂O-SiO₂ at 60°C; log a_{H₂O} = -3.52 = quartz saturation.



The System HCl-H₂O-Al₂O₃-CaO-CO₂-Na₂O-SiO₂ at 60°C; log a_{H₂O} = -3.52 = quartz saturation.

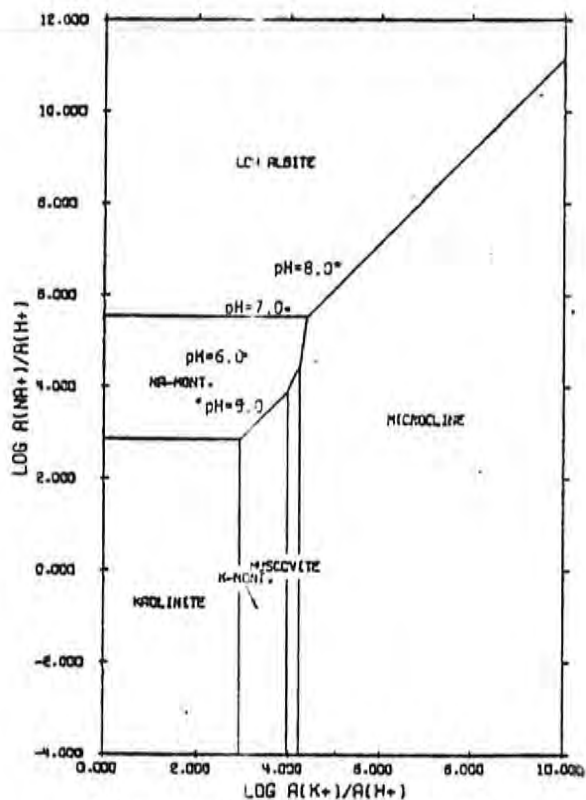


The System HCl-H₂O-Al₂O₃-CO₂-MgO-Na₂O-SiO₂ at 60°C; log a_{H₂O} = -3.52 = quartz saturation.

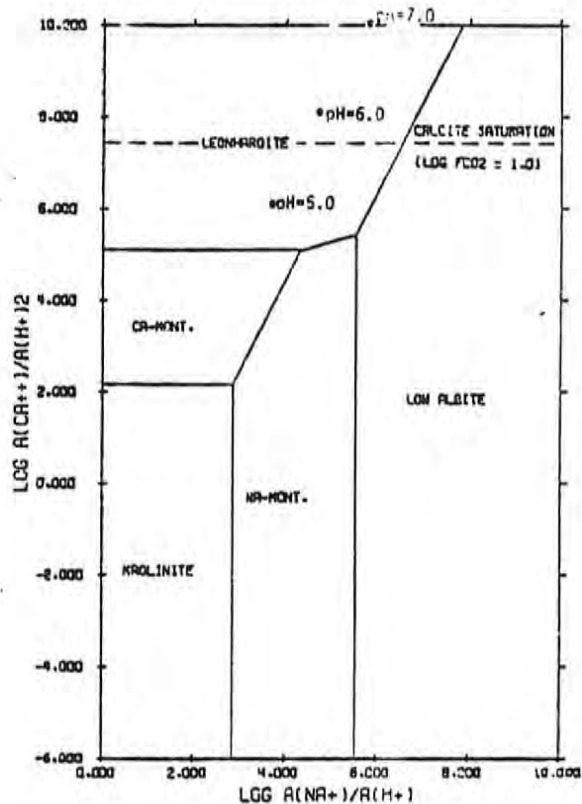


The System HCl-H₂O-Al₂O₃-CO₂-K₂O-MgO-SiO₂ at 60°C; log a_{H₂O} = -3.52 = quartz saturation.

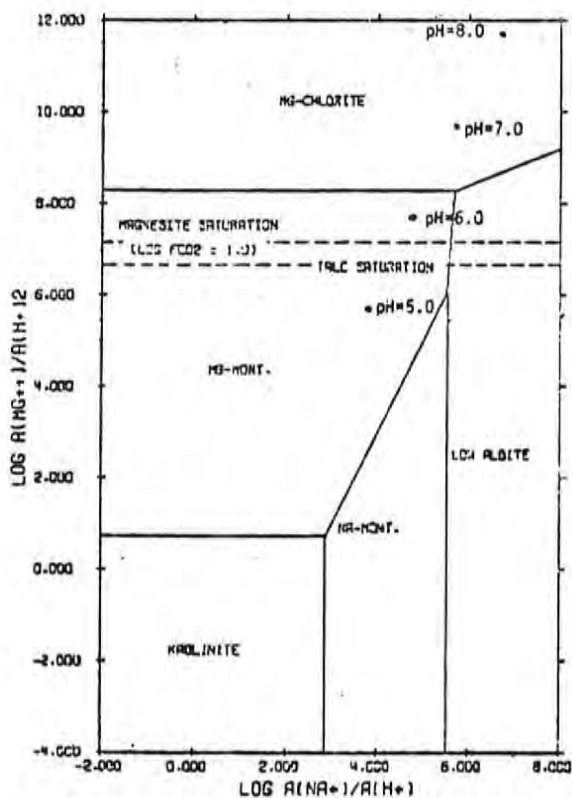
Figure 4-19. Phase stability diagrams for Sample DV-30 at 60°C.



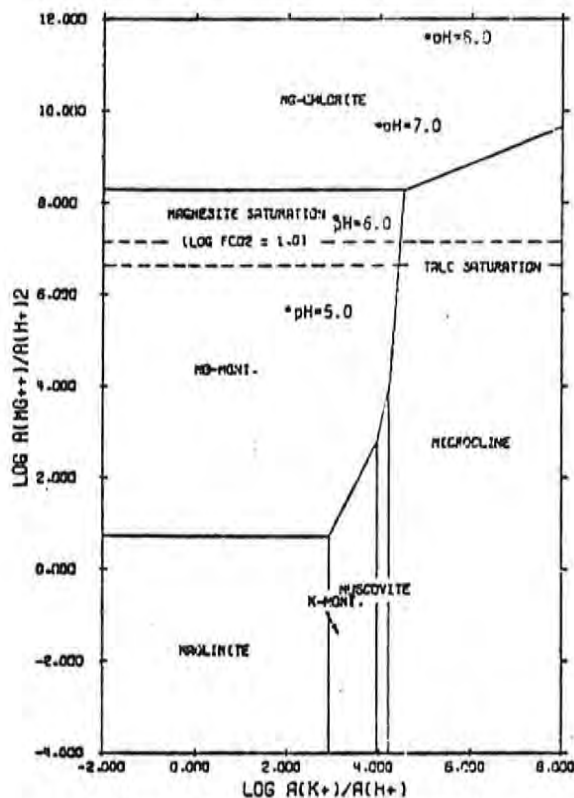
The System $HCl-H_2O-Al_2O_3-K_2O-Na_2O-SiO_2$ at $399^\circ C$;
 $\log a_{H_2O} = -2.59 = \text{quartz saturation}$.



The System $HCl-H_2O-Al_2O_3-CaO-CO_2-Na_2O-SiO_2$ at $200^\circ C$;
 $\log a_{H_2O} = -2.35 = \text{quartz saturation}$.

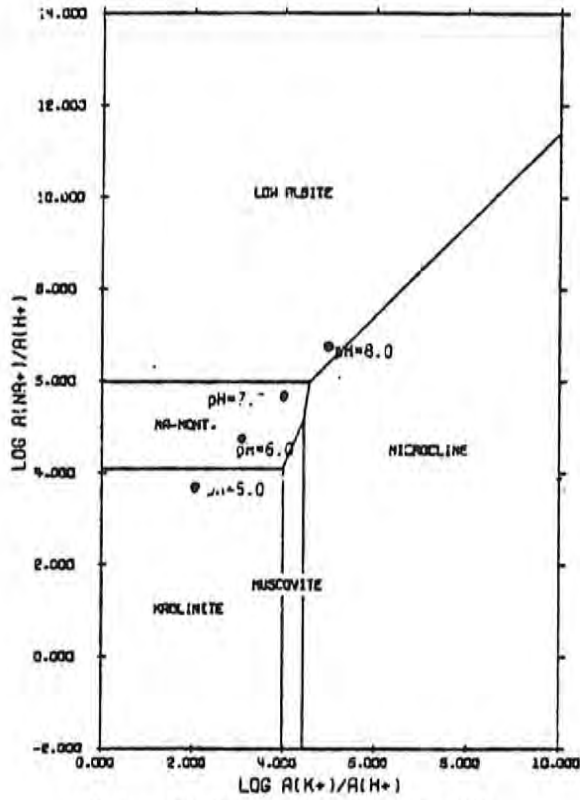


The System $HCl-H_2O-Al_2O_3-CO_2-MgO-Na_2O-SiO_2$ at $260^\circ C$;
 $\log a_{H_2O} = -2.35 = \text{quartz saturation}$.

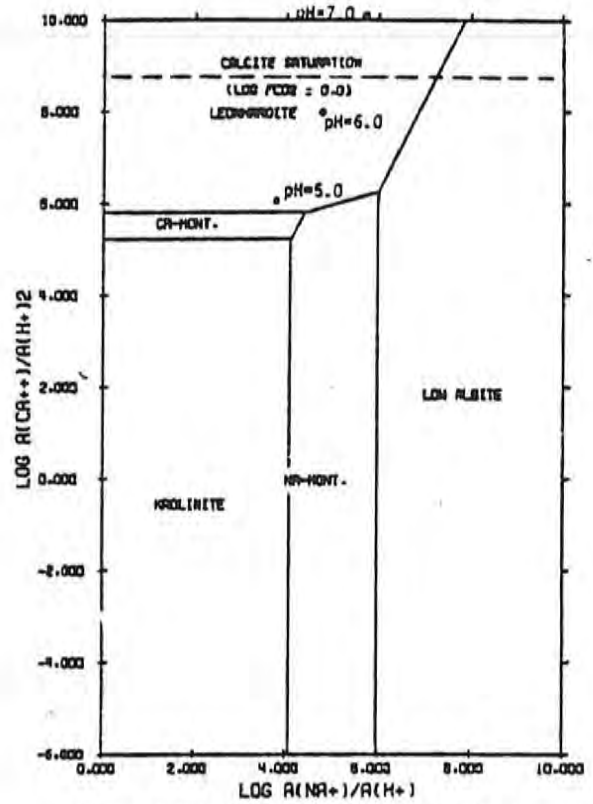


The System $HCl-H_2O-Al_2O_3-CO_2-K_2O-MgO-SiO_2$ at $200^\circ C$;
 $\log a_{H_2O} = -2.35 = \text{quartz saturation}$.

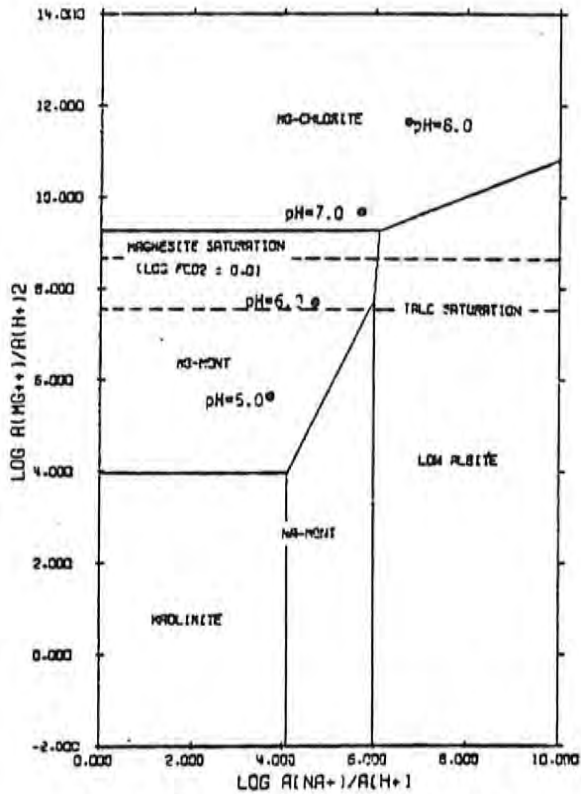
Figure 4-20. Phase stability diagrams for Sample DV-93 at $200^\circ C$.



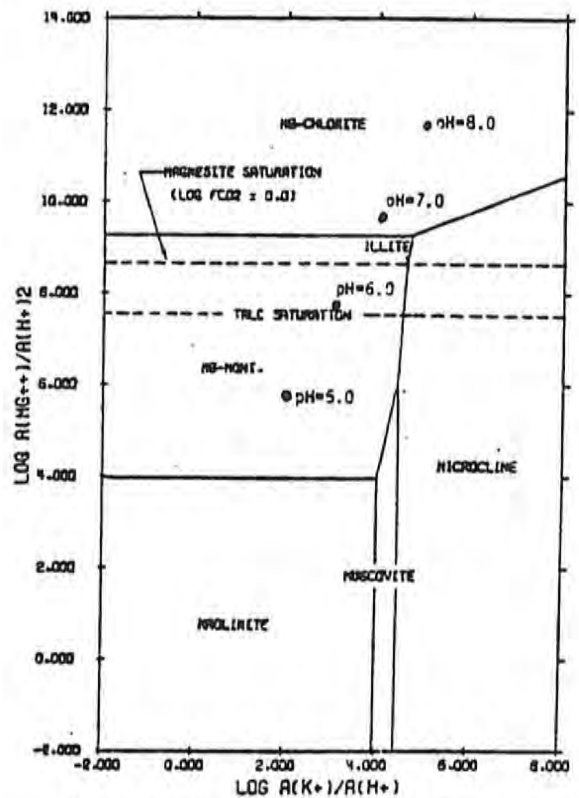
The System $HCl-H_2O-Al_2O_3-K_2O-Na_2O-SiO_2$ at $150^\circ C$; $\log a_{H_2O} = -2.67 = \text{quartz saturation}$.



The System $HCl-H_2O-Al_2O_3-CaO-CO_2-Na_2O-SiO_2$ at $150^\circ C$; $\log a_{H_2O} = -2.67 = \text{quartz saturation}$.

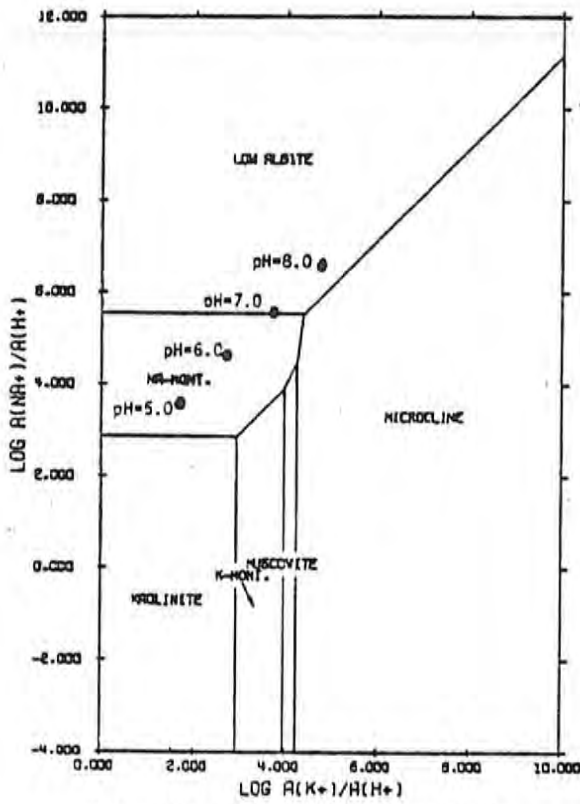


The System $HCl-H_2O-Na_2O-CO_2-MgO-Na_2O-SiO_2$ at $150^\circ C$; $\log a_{H_2O} = -2.67 = \text{quartz saturation}$.

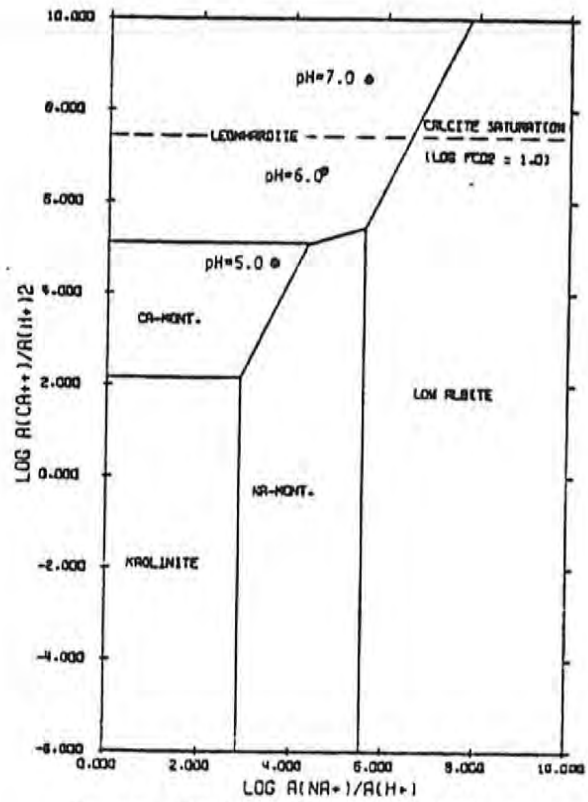


The System $HCl-H_2O-Al_2O_3-CO_2-K_2O-MgO-SiO_2$ at $150^\circ C$; $\log a_{H_2O} = -2.67 = \text{quartz saturation}$.

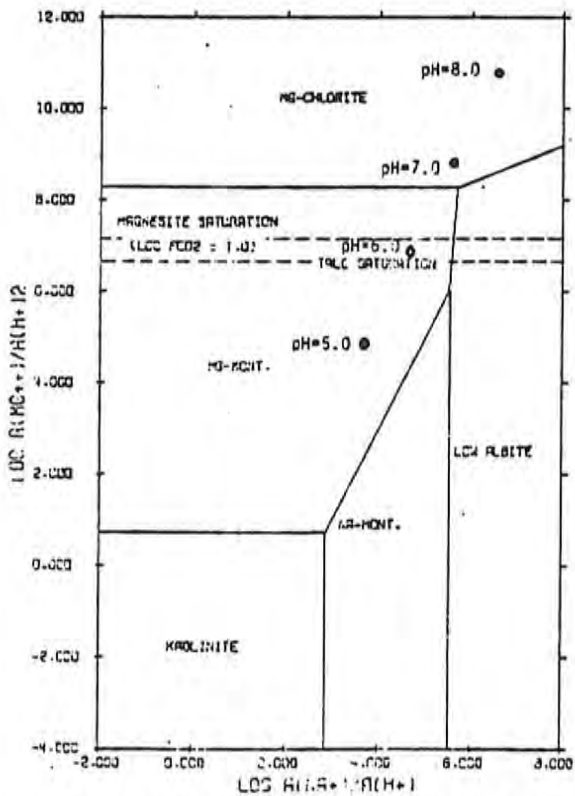
Figure 4-21. Phase stability diagrams for Sample DV-93 at $150^\circ C$.



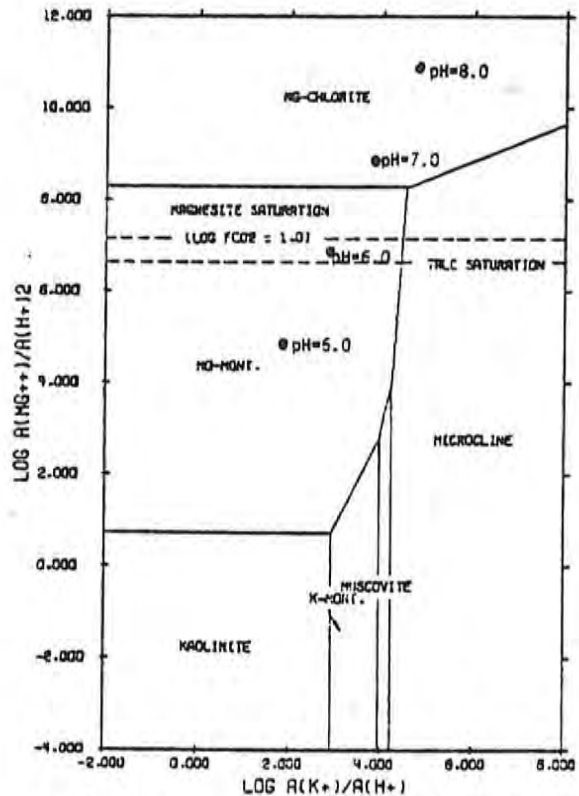
The System HCl-H₂O-Al₂O₃-K₂O-Na₂O-SiO₂ at 200°C;
log a_{H₂O} = -2.35 = quartz saturation.



The System HCl-H₂O-Al₂O₃-CaO-CO₂-Na₂O-SiO₂ at 200°C; log a_{H₂O} = -2.35 = quartz saturation.

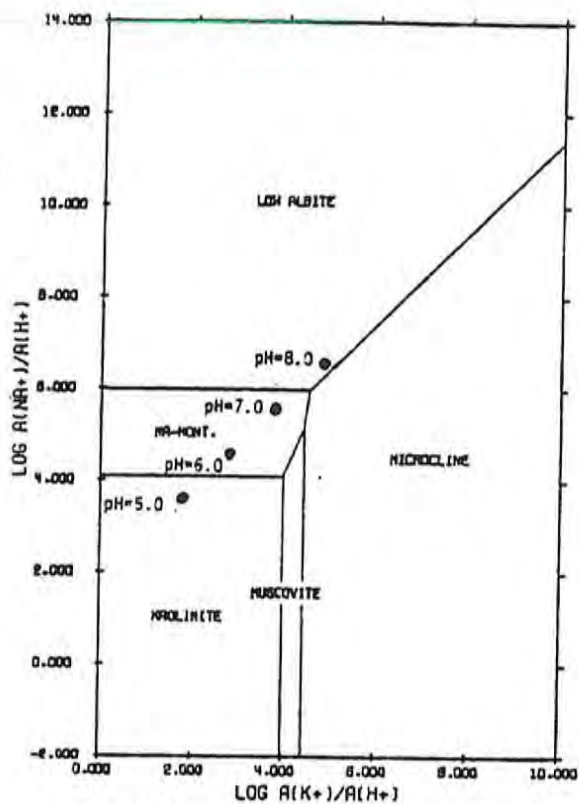


The System HCl-H₂O-Al₂O₃-CO₂-MgO-Na₂O-SiO₂ at 200°C; log a_{H₂O} = -2.35 = quartz saturation.

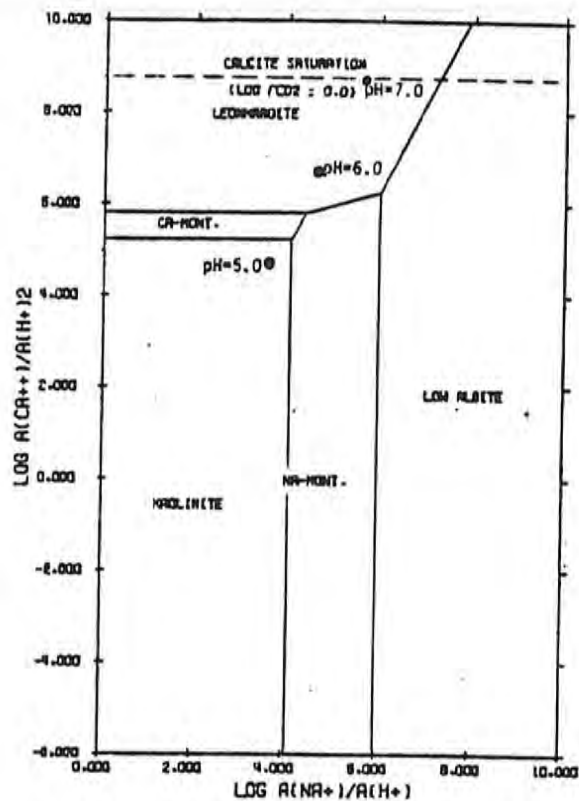


The System HCl-H₂O-Al₂O₃-CO₂-K₂O-MgO-SiO₂ at 200°C; log a_{H₂O} = -2.35 = quartz saturation.

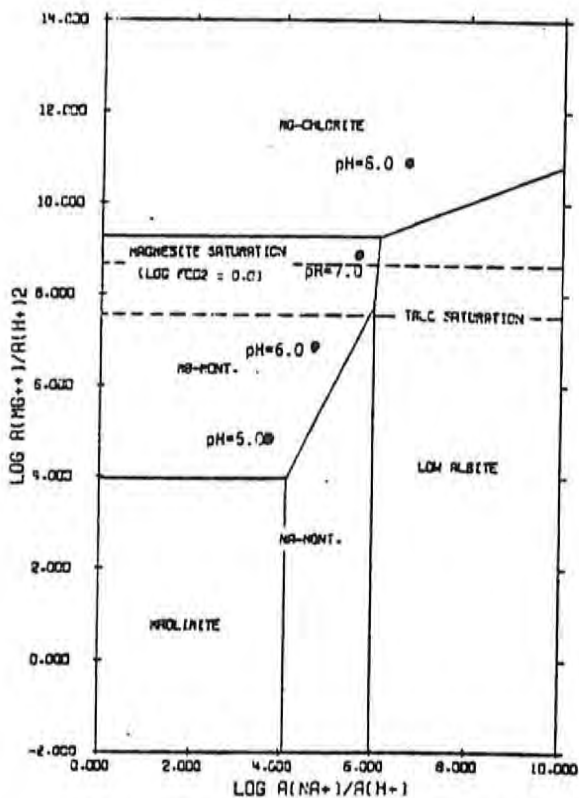
Figure 4-22. Phase stability diagrams for Sample DV-80 at 200°C.



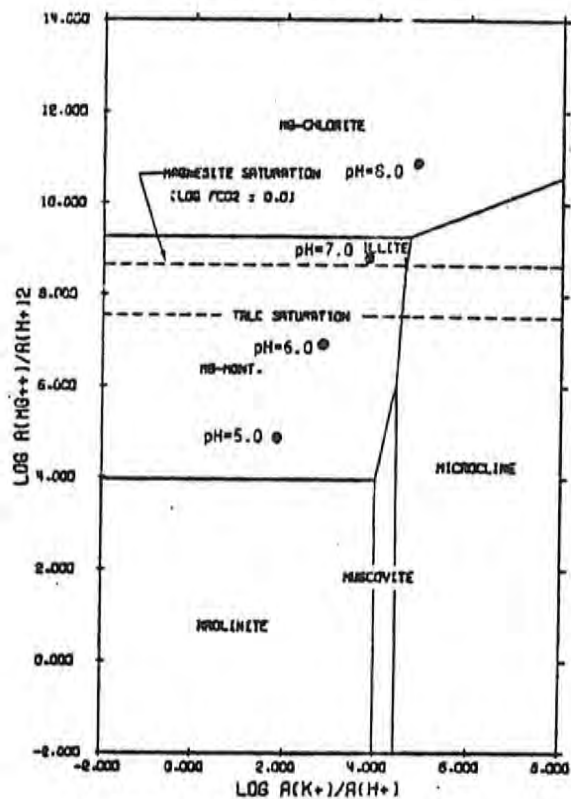
The System HCl-H₂O-Al₂O₃-K₂O-Na₂O-SiO₂ at 150°C; log $a_{\text{H}_2\text{O}}$ = -2.67 = quartz saturation.



The System HCl-H₂O-Al₂O₃-CaO-CO₂-Na₂O-SiO₂ at 150°C; log $a_{\text{H}_2\text{O}}$ = -2.67 = quartz saturation.

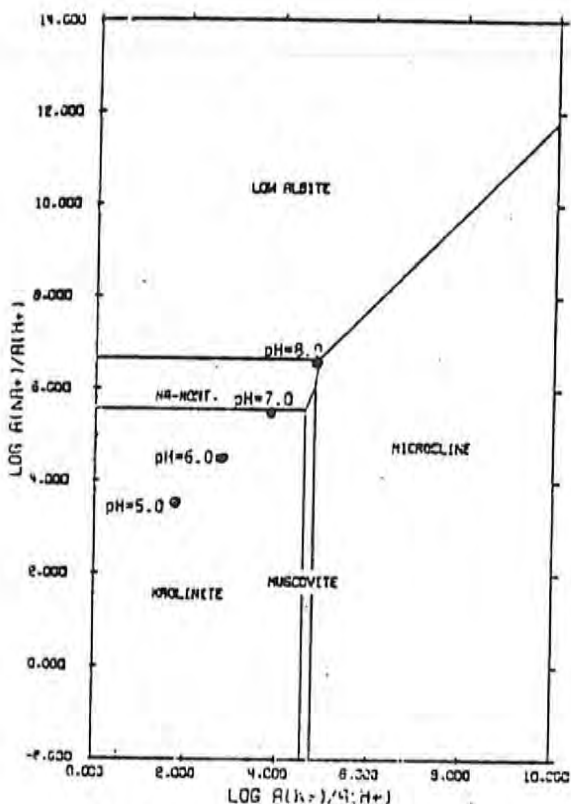


The System HCl-H₂O-Al₂O₃-CO₂-MgO-Na₂O-SiO₂ at 150°C; log $a_{\text{H}_2\text{O}}$ = -2.67 = quartz saturation.

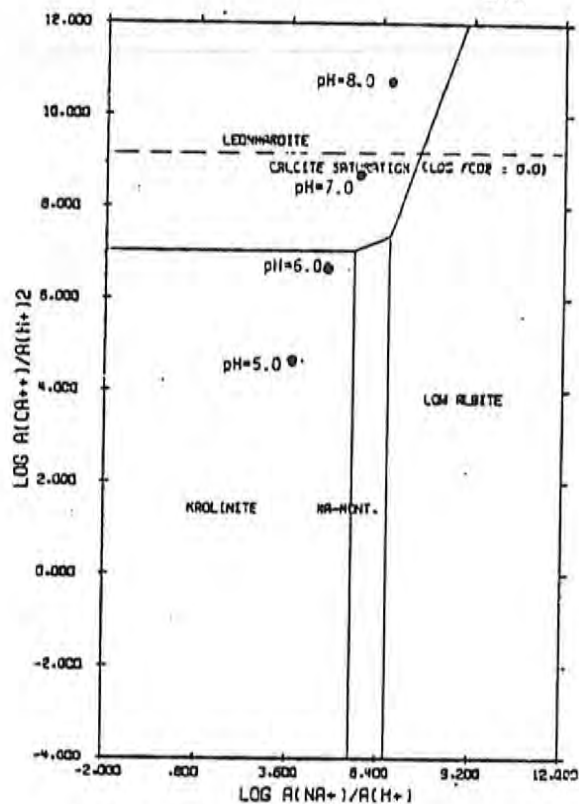


The System HCl-H₂O-Al₂O₃-CO₂-K₂O-MgO-SiO₂ at 150°C; log $a_{\text{H}_2\text{O}}$ = -2.67 = quartz saturation.

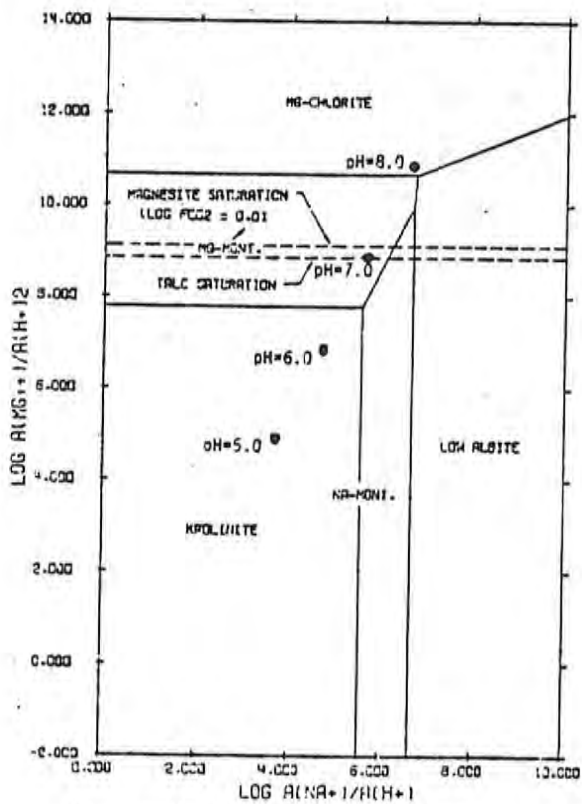
Figure 4-23. Phase stability diagrams for Sample DV-80 at 150°C.



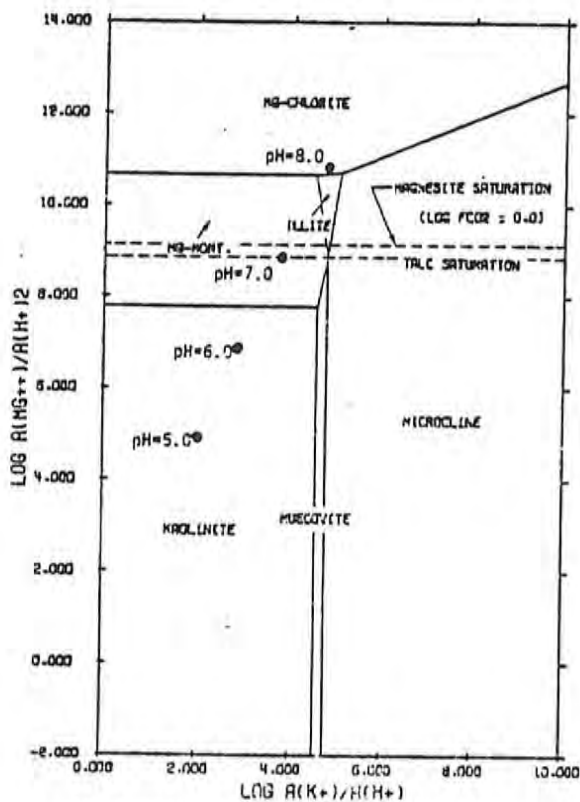
The System HCl-H₂O-Al₂O₃-K₂O-Na₂O-SiO₂ at 100°C; log a_{quartz} = -3.5 = quartz saturation.



The System HCl-H₂O-Al₂O₃-CaO-CO₂-Na₂O-SiO₂ at 100°C; log a_{quartz} = -3.68 = quartz saturation.



The System HCl-H₂O-Al₂O₃-CO₂-MgO-Na₂O-SiO₂ at 100°C; log a_{quartz} = -3.08 = quartz saturation.



The System HCl-H₂O-Al₂O₃-CO₂-K₂O-MgO-SiO₂ at 100°C; log a_{quartz} = -3.08 = quartz saturation.

Figure 4-24. Phase stability diagrams for Sample DV-80 at 100°C.

more acidic waters at depth which affect the kaolinitic alteration as they migrate upward. On the other hand the occurrence of montmorillonite, which is favored under more alkaline conditions, may reflect the action of downward percolating alkaline surface waters. These same waters probably mix with the more acidic waters issuing from the fractures thereby confining kaolinite to the fracture zones.

The observed mineralogy in DF 45-14 below 6000 feet (approximate depth of casing) is chiefly K-mica, quartz, chlorite, calcite, andalusite/chiaastolite, clay minerals \pm laumontite. Potassium feldspar as adularia occurs in trace amounts as vein material, and albite and epidote are abundant within the basic intrusions. If sample DV-90 (pH = 6.8) truly represents the deep water conditions in DF 45-14 and assuming a temperature of 200°C, the assemblage quartz, Na/Mg-montmorillonite-leonhardite (partially dehydrated laumontite) and/or calcite, \pm albite \pm K-spar would represent an equilibrium assemblage (Figure 4-16).

Only when the temperature is lowered to 100°C or the pH is reduced to 5.0 or below will kaolinite become stable. Figure 4-17 shows the migration of the stability fields as the temperature is lowered to 150°C for DV-90 and as the pH changes. Another significant observation of the activity diagrams for DV-90 is the apparent disequilibrium of K-mica with the present physical and chemical conditions. This mineral occurred pervasively in the cuttings from DF 45-14. The activity diagrams indicate only a small stability field for K-mica. As in the case of kaolinite, lower pH values than the measured field pH of 6.8, as well as a higher potassium ion activity, are required to bring K-mica into equilibrium. Illite (K-mica) is also spatially related to fracture zones supporting the hypothesis that more acidic conditions exist in these intervals. The occurrence of laumontite, mentioned earlier, may reflect alkaline conditions and may also be due to the downward migration of near surface alkaline waters along fracture or fault zones. This may come about when deep communication with more acidic water is temporarily cut off due to mineral deposition or tectonic closures of the fractures at depth.

Two major zones of alteration appear in the lower 5000 feet of DF 45-14. Highly altered intervals (4700 to 5300 and 8300 to 8600 feet) are characterized by an assemblage of kaolinite, illite, and mixed-

layer illite-montmorillonite that is spatially associated with presumed fracture zones. Larger generally less altered zones are negatively correlated with fracture zones (Plate V). It is evident then that fracture porosity and permeability is the dominant control in influencing the intensity and distribution of the various alteration effects observed in DF 45-14.

The alteration mineralogy above 1500 feet in DF 45-14 is dominated by Na-montmorillonite, illite (K-mica) and lesser chlorite. In the deeper portions of the well mixed-layer illite-montmorillonite is the predominant clay mineral. Examination of the activity diagrams (Figures 4-18 and 4-19) for sample DV-30, believed to be representative of the conditions above 1500 feet, show this assemblage to be at or very near equilibrium. An important shift occurs in the water chemistry between the deep (DV-90) and shallow (DV-30) water samples from DF 45-14. Calcium and magnesium increase significantly while sodium and potassium decrease. This may indicate alteration of Mg and Ca-silicates to form K and Na clays in the upper portion of the well. At greater depths Ca and Mg are less abundant due in part to rock type and also because they may be held in chlorite and Ca-zeolites or calcite.

The apparent disequilibrium in the deeper portions of DF 45-14 may be due in part to mixing of downward moving meteoric water via faults and fractures with upward migrating thermal water that has been heated at depth. The amount and depth of fluid mixing is unknown but the process may be very important in influencing the physical and chemical conditions at depth. The deposition of quartz veins must occur from solutions moving upward from presumably hotter areas where silica is dissolving. This does not appear to be taking place in the SiO_2 -saturated DV-90 water nor in the waters from DF 66-21. Also, more acidic, CO_2 -charged waters may be rising from deeper in the system along fractures and faults affecting kaolinite alteration and depositing calcite as CO_2 is lost.

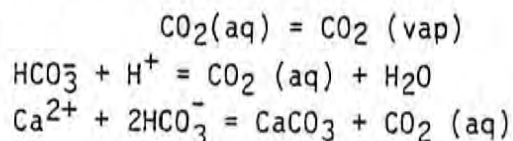
In considering the alteration effects by equilibrium activity plots it should be clear that this is not the whole picture. It is important to realize that only selected components have been considered, hence only selected phases. We have considered in part the 7-component system $\text{MgO-CaO-K}_2\text{O-Na}_2\text{O-Al}_2\text{O}_3\text{-SiO}_2\text{-CO}_2\text{-H}_2\text{O}$. This implies that for a given temperature and pressure seven stable phases should exist.

Muffler and White (1969) report for the Salton Sea Geothermal field the above components produce seven phases: quartz, albite, potassium feldspar, chlorite, K-mica, epidote and calcite. This approach, however, ignores other important components such as iron and sulfur. Iron is particularly important in Dixie Valley where biotite and/or vermiculite is encountered in much of the altered metasediments. The x-ray diffraction data indicate that the majority of the chlorites are iron-rich varieties. Additionally, the presence of abundant ilmenite (FeTiO_3) and its alteration product leucoxene have been established by x-ray fluorescence and optical data, respectively.

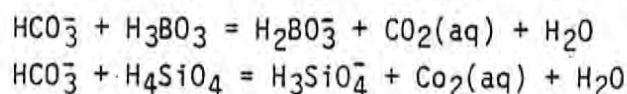
The relationship between the alteration mineralogy and water chemistry in DF 66-21 is likewise complex and, in some cases, obscure. The diversity and structural complexity of the rock units, in conjunction with the physical and chemical variations of the geothermal fluids, are the dominating factors contributing to this complexity. Two water samples, DV-93 and DV-80, are used to represent the chemical conditions at a depth of about 4700 feet in DF 66-21. The casing in DF 66-21 extends to a depth of 7300 feet. Tables 4-5 and 4-6 list the chemical analyses for DV-93 and DV-80, respectively. The analytical data for samples DV-93 and DV-80 were calculated at a pH of 8.0 rather than the recently acquired pH of 5.8.

For samples DV-93 and DV-80 the activity diagrams (Figures 4-20 through 4-24) depicting mineral equilibria at 150°C and to a lesser extent 200°C are believed to be most nearly representative of the conditions existing in DF 66-21 (170°C , pH = 5.8). The diagrams indicate an equilibrium assemblage of quartz, calcite, montmorillonite, illite, kaolinite, and albite. This essentially is what is observed in DF 66-21 in the upper 5000 feet where the water samples are applicable. Thus, excluding the abundance of kaolinite in localized zones, the above mineral assemblage appears to be in approximate equilibrium with the present physical and chemical conditions in DF 66-21. As described above for DF 45-14, it is possible that the pH in DF 66-21 is lower than the measured field pH (5.8) due to temperature effects, CO_2 loss upon boiling and fluid mixing. The decrease in pH results in a progressively more clay dominated mineral assemblage (Figures 4-21 through 4-24). The diagrams also indicate that the acidity required for kaolinite stability becomes less at lower temperatures.

Abundant kaolinite in the interval 4300 to 5300 feet corresponds to a zone of highly fractured volcanic rocks where large quantities of hot water (100°C to 150°C) under high pressure (320 psi) were encountered during drilling. This phenomenon may indicate communication at depth with more acidic waters. This interval from 4300 to 5300 feet, as well as 6300 to 6700 feet, and a broad zone, from 7500 to 9500 feet are also marked by increased calcite abundance (Plate VI). These intervals are interpreted as representing fault zones where boiling is occurring. Water which was originally close to saturation with calcite rapidly becomes supersaturated following boiling, loss of carbon dioxide and rising pH. The changes in chemistry which occur when steam boils from high-temperature water can be summarized in the following equations (Ellis and Mahon, 1977):



Samples DV-93 and DV-80 also reveal relatively high concentrations of boric and silicic acid (Tables 4-5 and 4-6). During the later stages of CO₂ loss, the following reactions may be occurring:



Bicarbonate ions are destroyed and the degree of supersaturation with calcite may be reduced possibly to the extent that the water becomes undersaturated again. The end effect of boiling and gas evolution is for the water to cool and for its pH to rise. The water is then no longer in equilibrium with the mineral assemblage with which it was in contact. This process would explain the localization of kaolinite near fracture zones. The effects produced depend mainly on the fluid flow rates. Calcite veining and fracture filling was much more abundant in DF 66-21 than in DF 45-14.

The distribution of chemical constituents between the reactant minerals and the product minerals listed in Table 4-7 suggests a relative chemical mass balance. This chemical balance is observed in both DF 66-21 and DF 45-14. One exception though is the addition of CO₃²⁻. The source of the carbon dioxide to bring about the CO₃²⁻ for calcite is

Table 4-7. Reactant and Product Minerals

Reactant Mineral	Chemical Formula	Product Mineral	Chemical Formula
Biotite	$K_2(Mg,Fe)_2(AlSi_3O_{10})(OH)_2$	Biotite	$K_2(Mg,Fe)_2(AlSi_3O_{10})(OH)_2$
Hornblende	$Ca_2(Mg,Fe,Al)_5(Si,Al)_8O_{22}(OH)_2$	Hornblende	$Ca_2(Mg,Fe,Al)_5(Si,Al)_8O_{22}(OH)_2$
Epidote	$Ca_2(Al,Fe)_3(SiO_4)_3OH$	Epidote	$Ca_2(Al,Fe)_3(SiO_4)(OH)$
Muscovite	$KAl_2(AlSi_3O_{10})(OH)_2$	K-Mica	$KAl_2(AlSi_3(O,OH)_{10})(OH)_2$
Augite	$Ca(Mg,Fe)(Si,Al)_2O_6$	Chlorite	$(Fe,Al)_6(Si,Al)_4O_{10}(OH)_8$
Plagioclase	$NaAlSi_3O_8$ -- $CaAl_2Si_2O_8$	Magnetite	Fe_3O_4
Pyrite	FeS_2	Vermiculite	$(Ca,Mg)(Mg,Fe,Al)_6(Al,Si)_8O_{22}(OH)_4 \cdot 2H_2O$
Magnetite	Fe_3O_4	Kaolinite	$Al_4(Si_4O_{10})(OH)_8$
		Limonite/Hematite	$FeO(OH)/Fe_2O_3$
		Montmorillonite	$(Na,Ca,Mg)O \cdot Al_2O_3 \cdot 5SiO_2 \cdot nH_2O$

uncertain; however, four possibilities can be considered. The first is that the carbon dioxide is derived from the organic matter identified in thin sections of the Triassic metasilstone/metashale. The second possibility is that the CO₂ is derived from magmatic additions. A third possibility is that the CO₂ is derived from carbonate rocks at depth although much more evidence is needed to substantiate their existence. The fourth and most likely source is the release of CO₂ upon boiling and/or degassing. This rough chemical mass balance can be interpreted as representing relatively stagnant flow conditions at depth, meaning that a significant influx of chemical constituents is not necessary to account for the observed alteration mineralogy. It also emphasizes the idea that significant flow of fluids is confined to fracture zones. Thus the mechanics of alteration in the two drill-holes consists mainly of the reconstitution of the chemical elements in the original rocks through the processes of solution, deposition and recrystallization according to the various equilibrium constraints.

There is a great deal more to learn and understand about the relationship between the alteration mineralogy and the water chemistry in the Dixie Valley region. However, a more thorough and detailed sampling of the deep wells is needed before this can be accomplished.

4.2.7 Well Correlations

4.2.7.1 Lithologic Correlation

The various lithologic units encountered in each of the two deep exploratory wells are compositionally very similar, differing mainly in the proportion of each unit present. The alluvium in both wells consists predominantly of altered volcanic fragments with lesser amounts of granodioritic to gabbroic igneous rocks. Fragments of the metasediments and quartz arenite are rare in the alluvial materials of both wells. The sequence of silicic to intermediate volcanic rocks in each well is very similar compositionally, texturally and structurally, although the degree of alteration is more intense in DF 66-21. The lithologies and textures of the metasediments in each well are essentially identical, although the association of andalusite/chiasolite porphyroblasts with igneous intrusion in these rocks appears to be lacking in DF 66-21.

A slight compositional variation exists between the intrusive

rocks observed in each well. The intrusions in DF 45-14 consist solely of rocks ranging in composition from diorite to gabbro. The predominant ferromagnesian minerals of these rocks are clinopyroxenes and hornblende. Equivalent intrusive material is encountered in DF 66-21. However, in DF 66-21, the majority of the intrusions are granodioritic to dioritic in composition. These rocks are characterized by biotite, hornblende, and lesser pyroxene as the dominant ferromagnesian minerals. Both compositional varieties contain magnetite and/or ilmenite that may in part be derived from the alteration of the ferromagnesian minerals. The textures of the two types are also similar. These slightly different intrusive rocks may represent two phases in the differentiation of a magma with the more basic intrusions representing earlier phases and the granodiorite/diorite representing the bulk composition. Detailed field relationships and age dates are required to evaluate this hypothesis.

4.3 Conclusions

The various lithologic units encountered in each of the two deep exploratory wells and, to the extent known, the Sun Oil Company well (SW Lamb #1) are compositionally and texturally similar, differing mainly in the proportion of each unit present. Plate VII shows the stratigraphic correlation between DF 45-14, DF 66-21 and SW Lamb #1. Figure 4-25 shows a generalized stratigraphic section of the Dixie Valley region. It can be seen that there is a strong correlation between all three wells in the occurrence of the sequence of volcanic rocks. This volcanic sequence can serve as a common datum plane.

In both DF 66-21 and SW Lamb #1, this volcanic sequence is overlain by an impervious red clay layer. Beneath this clay, both wells produced hot water and steam under high pressures from fractured volcanic rock (and underlying intrusive rock?). The general absence of an impermeable layer of this clay in DF 45-14 may explain the lack of a similar 'production zone'. The absence of this production zone may also be a function of the location of the well beyond the limits of the Humboldt gabbroic complex.

The distribution of the clay layer is very significant as it provides a capping mechanism for the geothermal system. Regardless of the origin, the clay represents a favorable cap rock and its association

GENERALIZED STRATIGRAPHIC SECTION OF DIXIE VALLEY

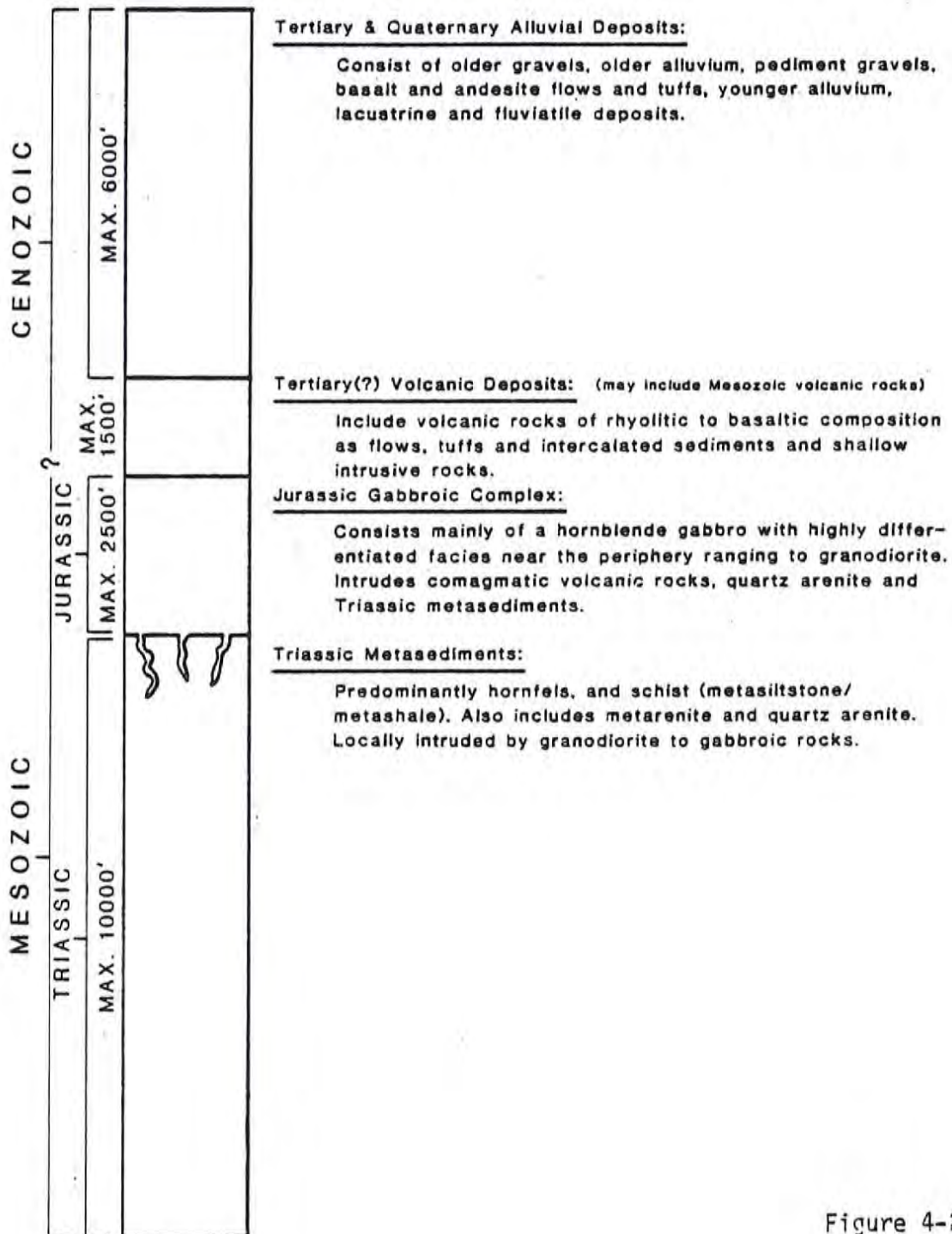


Figure 4-25

with the volcanic sequence overlying the Humboldt gabbroic complex may represent the key to exploiting the geothermal resource in Dixie Valley. The observations in the three wells support this hypothesis.

The absence of a sizeable intrusive body in DF 45-14 (Humboldt gabbroic complex) is significant for two reasons: first, if the boundary of the Humboldt gabbro represents the boundary of the 'producing' portion of the reservoir, this would indicate that DF 45-14 is located at the periphery of the unit. The distribution and stratigraphic relationships of the gabbroic complex in the three wells indicates that it pinches out toward the south. The gabbro however is too impermeable to act as a reservoir and so communication with the reservoir at depth is via fractures. Conductive heating in this body of rock may be an important factor. Another significant observation is that there are extensive exposures of the Humboldt gabbroic complex in the Stillwater Range in the immediate vicinity of DF 45-14 (Willden and Speed, 1974) yet very little was encountered in the well. Additionally, the nearest major outcrop of the Triassic metasediments is located about 10-12 miles to the south. This evidence suggests that post-Jurassic left-lateral displacement (up to 12 miles) has occurred in Dixie Valley along the White Rock Canyon fault (Bob Whitney, pers. commun., 1979).

Willden and Speed (1974) report that the only rocks intruded by the gabbro are quartz arenite and comagmatic basic volcanic rocks. However, minor gabbroic intrusions in the Triassic metasiltstone/metashale may be related to the complex. Willden and Speed (1974) also state that the quartz arenite forms a sort of ring surrounding the gabbro. This idea may be used in drilling to determine the relative position of the well in relation to the gabbroic complex.

The role of the Triassic metasedimentary sequence in the reservoir model is uncertain although it most likely functions as a cap rock as it is relatively impermeable. This unit is known to be fractured at depth but the extent of this fracturing is unknown. It is also known that these fractures are capable of conducting thermal fluids which may originate largely at depth. Willden and Speed (1974) postulated that Triassic carbonate rocks exist at depth below the Humboldt gabbroic complex. These carbonate rocks could conceivably serve as a reservoir for convecting thermal fluids. They could also supply the necessary CO₂ and CO₃ for the abundant calcite precipitation that was observed in DF 66-21.

The heat source for the geothermal system in Dixie Valley is elusive; although one surely exists. A couple of options are possible. The first is that heat is derived from the upper mantle which is relatively close to the surface in this region. The heat source may also be related to the gabbro which may conduct heat upwards from great depths. A shallow molten heat source is unlikely.

The vast discrepancy in the amount of alluvial material in DF 45-14 and DF 66-21 (1100 feet vs. 4100 feet) together with the distribution of alteration effects associated with areas of increased porosity and permeability provides evidence that DF 66-21 has been downdropped by at least one increment of normal faulting. Some of the difference may be due to its increased distance from the range front. Plate IV shows the location of the wells in relation to the structural-tectonic features. Surface expression of the faulting described above is to the west of DF 66-21. Assuming a 50 to 60 degree dip on these normal faults, they intersect the wells at depths of 4700 to 4800 feet and at about 6500 feet. These depths also correspond to zones of increased alteration indicating rocks with higher porosity and permeability. It is apparent that faulting has played a major role in influencing the distribution not only of rock types but also the hydrothermal alteration effects.

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Chapter 5. HYDROLOGY AND HYDROGEOCHEMISTRY

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5.0 HYDROLOGY AND HYDROGEOCHEMISTRY

5.1 Introduction

5.1.1 Purpose and Scope

The purpose of this portion of the study is to provide hydrologic and hydrogeochemical input to the construction of the Dixie Valley geothermal model. Since some of the most important aspects of hydrothermal reservoir assessment are the fluid characteristics and flow parameters, the hydrology and hydrogeochemistry of a particular reservoir must be considered in the formulation of any hydrothermal reservoir model, even a conceptual or qualitative one. Our primary efforts were devoted to extensive isotopic and hydrochemical sampling and interpretation of these data as direct and indirect indicators of the nature of the reservoir. Most of the work focused on the area of greatest interest to Southland Royalty Company, the northern half of Dixie Valley (between Dixie Meadows and Sou Hot Springs). The study also involved sampling outside this particular region.

The major tasks of the hydrologic-hydrogeochemical study are as follows:

- (1) Review the available hydrologic and hydrogeochemical data.
- (2) Obtain water samples from selected wells, hot springs and cold springs.
- (3) Analyze water samples for major, minor and selected trace chemical constituents and environmental isotopes.
- (4) Collect temperature and other data from selected wells and springs.
- (5) Estimate recharge rates and source areas and groundwater flow rates.
- (6) Estimate reservoir geometry.
- (7) Provide estimates of reservoir water chemistry and, using geothermometry, reservoir temperatures.
- (8) Provide hydrologic and hydrogeochemical input to the formulation of a conceptual model of the Dixie Valley hydrothermal system.

The interpretation of the environmental isotope data was intended to supplement the hydrogeochemical data and provide additional

hydrologic information, particularly with respect to recharge to the reservoir. However, the complete suite of isotope data has not yet been received. The interpretations and conclusions presented herein have been made without the benefit of any isotope data, and are based almost solely upon the hydrogeochemistry. This is not an ideal approach, but was dictated by circumstance. Therefore, this chapter is incomplete since it is devoted primarily to the hydrogeochemistry of the study area with little discussion of the hydrology. An addendum to this chapter will be written as soon as the isotope data have been received, interpreted and integrated with the rest of the information.

5.1.2 Methods and Analytical Techniques

Approximately 100 samples of the thermal and non-thermal waters were collected. The collection generally involved filtration through a 0.45 μm filter for a gross sample and an acidified sample. Nitric acid was added to lower the pH to approximately 2. Samples were collected for silica using a 1:10 field dilution of the thermal waters with distilled water. Isotope samples were collected by completely filling a 125 ml glass bottle and sealing the bottle with Parafilm and electrical tape.

Field measurements were made of temperature, electrical conductivity (EC), pH and dissolved oxygen (DO). Laboratory determinations were made for Ca, Mg, Na, K, HCO_3 , Cl, SO_4 , NO_3 , SiO_2 , F, B, Li, As, Cs, Al, Hg, Fe, Mn, Sr and Ba in the Water Resources Center Laboratories.

^3H analyses (both enriched and unenriched) were performed in Water Resources Center Laboratories. ^{18}O and ^2H samples were sent to the Laboratory of Isotope Geochemistry at the University of Arizona for analysis.

5.1.3 Previous Work

Very little previous work on the hydrology and hydrogeochemistry of Dixie Valley exists. A paper by Zones (1957) describes some of the hydrologic effects of the 1954 Dixie Valley earthquake. A reconnaissance study by Cohen and Everett (1963) gives an overview of the groundwater hydrology of Dixie and Fairview Valleys. This report also

includes a brief description of the groundwater chemistry of the Dixie-Fairview area. A total of 13 water chemistry samples were collected in an area of 2360 square miles, a very low sampling density. The data presented in this report were insufficient to draw any substantive conclusions concerning the groundwater system in Dixie Valley. Additional work of a limited nature in the valley was conducted by Keplinger and Associates (1977 and 1978) and by GeothermEx, Inc. (1976).

5.2 Analytical Results

5.2.1 Chemical Characteristics of Dixie Valley Waters

Table 5-1 shows the chemical analyses of all the samples collected during the study. Variations in gross chemical properties of Dixie Valley groundwater and surface water are evident on a trilinear diagram (Figure 5-1). Percentages of equivalents were plotted for major anions and cations (Cl , SO_4 , HCO_3 and CO_3 , Ca , Mg and $\text{Na} + \text{K}$). The three hot spring systems of the valley plot as separate groups. Dixie Hot Springs (D) show significant variation, but generally they are sulfate-chloride-potassium-sodium waters. Hyder Hot Springs (H) show little variation, and are bicarbonate-potassium-sodium type waters. Sou Hot Springs (S) is an intermediate type, having roughly similar equivalent percentages of the major ions. The hot springs of McCoy Ranch (M), Lower Ranch and southern Jersey Valley seem to be related to Hyder Hot Springs. Buckbrush Seeps (B) also seem to be closely related to Hyder, perhaps due to structural relationships. However, Buckbrush Seeps might be affected by evaporation. Surface runoff from the Stillwater Range, irrigation wells from the northern part of the valley and waters from the eastern ranges (Clan Alpine and Augusta Mountains) show wide variations. There is, however, a rough grouping of the water quality samples according to their areal distribution in the valley. For example, the samples from the deep wells (DF 45-14 and DF 66-21) and well SR2-A seem to be related to Dixie Hot Springs. It should be noted that samples from DF 45-14 and DF 66-21 may not be representative of the geothermal reservoir fluids because of contamination from drilling fluids and additives and/or shallow groundwater.

Table 5-1. Chemical Analyses of Dixie Valley Waters.

The letter or number in the first column identifies the group sample according to the following code:

- A - Artesian wells in the vicinity of Dixie Settlement.
- B - Buckbrush Seeps.
- C - Carson Sink (western slopes of the Stillwater Range).
- D - Dixie Hot Springs.
- E - Spring or stream from the eastern mountain ranges (Clan Alpine and Augusta Mountains).
- H - Hyder Hot Springs.
- I - Irrigation wells in the northern part of Dixie Valley.
- L - Wells in the vicinity of Dixie Settlement.
- M - Intermediate temperature springs around McCoy Ranch.
- S - Sou Hot Springs.
- W - Spring or stream from the Stillwater Range.
- 1 - DF 45-14 (probably contaminated).
- 2 - DF 66-21 (probably contaminated).
- 3 - DF 45-14.
- 4 - DF 66-21 (probably contaminated).

Table 5-1. Chemical Analyses of Dixie Valley Waters.

	DATE	TEMP	PH	CA	HG	NA	K	CL	SO4	HC03	SI02	C03	T0S
W	LL047	90878	15.00	8.00	105.600	43.300	133.000						
W	LL109	92878	16.00	6.90	270.400	31.300	211.000						
W	DV21	32079	13.00	7.40	46.500	26.600	464.000						
W	DV22	32079	11.00	8.50	56.000	32.000	116.500						
W	DV031	51579	17.00	8.35	82.500	33.500	108.000						
W	DV032	51579	26.00	8.77	54.500	52.500	176.000						
W	DV045	51579	14.00	7.69	68.000	35.000	100.000						
W	DV046	51579	15.75	8.35	44.500	21.500	70.000						
W	DV047	51579	18.00	8.50	42.500	24.000	80.000						
W	DV048	51579	13.50	7.78	48.000	30.700	78.000						
W	DV049	51579	15.75	8.21	58.000	37.000	123.000						
W	DV050	51579	16.00	8.45	39.500	26.000	94.000						
W	DV051	51579	17.00	8.17	41.500	22.500	114.000						
W	DV052	51579	17.00	8.54	53.000	24.000	165.000						
W	DV053	62579	18.00	7.20	175.000	78.000	118.000						
W	DV055	62579	15.00	7.86	66.000	34.500	120.000						
W	DV56	62579	15.00	7.66	65.000	37.800	117.000						
W	DV057	62579	15.00	8.45	44.000	40.000	138.000						
W	DV058	62579	15.00	8.60	41.500	27.800	105.000						
W	DV059	62579	28.00	8.70	38.000	31.300	123.000						
W	DV067	62579	16.00	7.33	262.000	33.300	215.000						
W	DV078	72379	22.00	7.40	95.700	16.900	167.000						
W	DV092	091279	43.00	8.48	32.000	65.000	197.000						
W	DV102	62579	14.00	7.71	104.000	36.000	138.000						
W	DV103	62579	14.00	7.79	62.500	26.500	82.500						
W	DV104	62579	13.00	8.13	74.000	31.000	112.000						
W	DV105	62579	28.00	8.57	39.500	22.000	102.000						
W	DV106	62579	26.80	8.32	42.000	26.500	118.000						
W	DV107	62579	15.50	7.55	65.000	36.500	130.000						
W	DV108	62579	16.50	7.62	70.500	39.800	140.000						
W	DV109	62579	22.50	8.35	65.000	46.500	182.000						
W	DV110	62579	15.50	7.87	60.000	34.500	65.000						
W	DV111	62579	14.00	7.79	100.000	41.300	97.500						
C	CS01	72379	15.00	7.93	60.200	33.200	108.000						
C	CS02	72379	16.00	7.60	59.900	34.900	110.000						
C	CS04	72379	16.00	7.58	97.800	35.100	48.000						
C	CS06	72379	21.00	7.63	483.000	36.500	108.000						
C	CS07	72379	22.00	7.93	267.000	32.300	266.000						
C	CS08	72379	24.00	8.73	135.000	31.800	266.000						
A	DV041	51579	28.00	9.59	13.700	2.100	1150.000						
A	DV042	51579	15.00	9.87	1.000	.600	352.000						
F	LL102	92878	10.00	7.60	14.200	2.300	18.000						
F	DV03	32079	11.00	7.70	37.500	4.500	154.000						
F	DV10	32079	15.00	7.33	165.000	66.000	55.500						
F	DV12	32079	11.00	7.27	54.500	14.000	37.000						
F	DV060	62579	9.00	7.85	14.500	2.720	21.000						
F	DV061	62579	10.00	8.33	42.300	7.600	43.500						
F	DV062	62579	23.00	7.92	34.000	7.100	41.600						
F	DV063	62579	18.00	7.40	19.000	6.700	25.000						
F	DV091	091279	32.00	8.58	3.700	.060	105.000						
L	LL103	92878	19.00	7.70	20.900	2.070	62.800						
L	LL104	92878	16.00	7.90	28.300	2.200	56.000						

Table 5-1. Chemical Analyses of Dixie Valley Waters. (cont'd.)

	DATE	TEMP	PH	CA	MG	NA	K	CL	SO4	HCO3	SI02	CO3	TDS
M DV05	32079	39.00	7.10	39.000	13.700	144.000	12.000	27.000	65.600	454.000	39.000	-0	799.385
M DV06	32079	29.00	6.95	68.000	19.100	164.000	12.300	28.000	124.000	554.000	44.000	-0	1019.315
M DV11	32079	43.00	6.90	95.000	37.500	213.000	10.300	27.000	209.000	310.000	36.800	.100	1191.515
M DV115	62579	50.00	7.31	94.000	36.500	218.000	10.000	28.000	216.000	312.000	32.500	-0	1206.370
M DV116	62579	40.00	6.89	67.000	18.500	240.000	23.800	39.000	88.000	803.000	26.300	-0	1312.200
I DV069	070979	68.5	9.5	20.500	.050	685.000	65.000	815.000	512.000	32.400	6.500	18.700	2163.205
I DV70	070979	75.5	8.9	12.500	.010	605.000	53.000	740.000	352.000	5.100	31.000	52.200	1470.145
I DV72	070979	76.5	9.1	22.500	.010	618.000	65.000	700.000	352.000	6.100	300.000	117.000	2201.726
Z DV080	81479	114.0	8.00	35.750	3.550	1288.000	31.000	1208.000	130.000	1410.000	-0	-0	4118.590
Z DV081	81479	51.00	-0	42.500	4.300	1238.000	27.000	-0	-0	-0	210.000	-0	1523.070
Z DV082	81479	86.00	7.90	27.000	3.900	1275.000	30.000	1150.000	127.500	1391.000	232.000	-0	4248.290
Z DV090	091279	94.	6.8	24.100	.015	410.000	40.000	493.000	215.000	130.500	325.000	-0	1657.545
4 DV093	091379	83.	8.	24.000	4.000	1730.000	44.000	1720.000	396.000	1283.000	189.000	-0	5410.600
4 DV094	091379	95.	8.5	23.600	9.900	2010.000	37.000	2315.000	280.000	1197.000	160.000	-0	6041.300
T DV15	042679	71.	7.63	156.000	30.000	400.000	30.000	535.000	448.000	200.000	98.000	-0	1905.930
T DV16	71.	7.42	145.000	28.800	398.000	18.000	550.000	445.000	203.000	100.000	-0	-0	1898.660
T DV030	51579	64.50	6.89	145.000	26.000	405.000	18.600	575.000	410.000	211.000	105.000	-0	1909.045

Table 5-1. Chemical Analyses of Dixie Valley Waters. (cont'd.)

	F	B	AL	FE	MN	LI	SR	NO3	AS	HG	CS
L LL105	-0	-0	-0	-0	-0	-0	-0	1.200	-0	-0	-0
L LL107	8.600	-0	-0	-0	-0	-0	-0	.310	-0	-0	-0
L LL108	.750	-0	-0	-0	-0	-0	-0	4.380	-0	-0	-0
L LL110	-0	-0	-0	-0	-0	-0	-0	2.260	-0	-0	-0
L LL111	-0	-0	-0	-0	-0	-0	-0	.090	-0	-0	-0
A LL106	-0	-0	-0	-0	-0	-0	-0	.220	-0	-0	-0
A DV13	6.000	.310	.100	.010	.015	.050	.110	-0	-0	-0	-0
A DV043	6.400	.320	.100	.010	.005	.052	.130	-0	-0	-0	-0
A DV065	6.500	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
A DV066	5.800	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
I DV034	.490	1.000	.100	.020	.017	.260	1.730	-0	.010	-0	-0
I DV037	.700	1.100	.100	.040	.071	.430	2.800	-0	.010	-0	-0
I DV038	1.000	1.300	.100	.010	.008	.350	1.870	-0	-0	.006	.050
I DV039	1.300	1.000	.100	.010	.005	.250	1.280	-0	-0	-0	-0
I DV040	.800	.900	.100	.010	.016	.300	2.400	-0	.010	-0	.050
I DV100	1.150	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
I DV101	.900	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
I DV113	.740	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
I DV114	.450	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
P DV20	.330	-0	-0	-0	-0	-0	-0	-0	-0	-0	.050
D LL098	-0	-0	-0	-0	-0	-0	-0	1.150	-0	-0	-0
D LL099	8.400	-0	-0	-0	-0	-0	-0	.090	-0	-0	-0
D LL100	-0	-0	-0	-0	-0	-0	-0	.090	-0	-0	-0
D LL101	-0	-0	-0	-0	-0	-0	-0	.400	-0	-0	-0
D DV23	12.600	.860	.100	.030	.015	.380	.020	-0	.010	.000	.100
D DV24	7.700	.400	.100	.010	.012	.550	4.350	-0	.010	.000	.170
D DV054	13.000	.400	.100	.010	.005	.382	.020	0	.010	.000	-0
D DH16	5.300	1.600	.100	.010	.005	.315	.380	0	.010	.000	-0
D DH15	12.500	1.000	.100	.010	.005	.425	.030	0	.010	.000	-0
D DH09	9.700	1.700	.100	.010	.008	.580	.950	0	.010	.000	-0
D DH06	11.200	1.800	.100	.010	.005	.470	.050	0	.010	.000	-0
D DH08	-0	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
D DH10	-0	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
H DV04	7.600	4.000	.100	.020	.022	1.600	1.200	-0	.030	.000	.280
H DV033	7.500	4.200	.100	.040	.007	1.680	1.260	-0	.030	.000	.260
H HHS01	8.100	4.900	.100	.020	.028	1.590	1.070	0	.020	.005	-0
H HHS03	8.200	6.100	.100	.070	.033	1.610	1.150	0	.010	.003	-0
H HHS05	8.100	5.000	.100	.010	.040	1.590	1.090	0	.010	.004	-0
H HHS06	-0	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
H HHS12	8.200	4.400	.100	.010	.026	1.560	1.100	0	.010	.000	-0
H HHS09	8.200	5.100	.100	.010	.021	1.630	1.060	0	.020	.000	-0
S DV01	5.100	1.360	.100	.010	.054	.650	10.800	-0	.010	.000	.120
S DV02	5.000	1.370	.100	.050	.045	.650	11.100	-0	-0	-0	-0
S DV035	5.100	1.200	.100	.040	.009	.710	6.200	-0	.010	.000	.110
S DV036	4.900	1.300	.100	.410	.006	.680	10.500	-0	-0	-0	-0
S DV075	5.300	2.000	.100	.380	.063	.670	11.400	-0	.010	.000	-0
S DV076	-0	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
S DV077	-0	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
S DV112	4.900	-0	-0	-0	-0	-0	-0	0	-0	-0	-0

Table 5-1. Chemical Analyses of Dixie Valley Waters (cont'd)

	DATE	TEMP	PH	CA	MG	NA	K	CL	SO4	HCO3	SI02	CO3	TDS
L LL105	92878	20.00	8.20	13.100	.300	65.000	2.900	19.500	60.000	106.000	73.900	-0	341.900
L LL107	92878	17.00	8.40	7.000	.200	67.000	1.900	17.500	50.200	80.500	70.800	7.100	311.110
L LL108	92878	16.00	7.60	37.800	3.720	43.000	4.300	29.500	55.000	124.000	62.400	-0	364.850
L LL110	92878	24.00	7.20	18.300	2.060	69.000	3.200	26.000	88.000	93.000	58.900	-0	360.720
L LL111	92878	14.00	7.20	45.800	7.300	38.800	1.300	32.800	38.700	180.000	28.600	-0	373.390
A LL106	92878	19.00	8.10	22.500	1.130	65.700	4.800	23.500	80.000	111.000	74.300	-0	383.150
A DV13	32079	19.00	8.30	18.800	.950	68.000	4.500	23.000	72.800	98.000	74.500	-0	367.145
A DV043	51579	18.30	8.26	18.000	.900	67.000	3.900	27.000	72.400	103.000	69.000	-0	368.217
A DV065	62579	20.00	8.35	17.500	.900	68.000	4.100	23.000	72.000	101.000	70.000	-0	363.000
A DV066	62579	20.00	8.18	19.000	1.150	65.000	3.700	23.500	69.000	104.000	62.500	-0	353.650
I DV034	51579	19.50	7.44	182.000	60.500	206.000	8.850	515.000	194.000	265.000	52.000	-0	1486.967
I DV037	51579	16.00	7.13	359.000	112.000	252.000	16.100	905.000	366.000	321.000	47.000	-0	2383.281
I DV038	51579	18.50	7.25	147.500	44.500	191.000	13.800	330.000	234.000	361.000	63.000	-0	1389.445
I DV039	51579	22.50	7.83	64.000	25.500	159.000	19.300	125.000	134.000	399.000	77.000	-0	1006.754
I DV040	51579	15.50	7.50	305.000	99.000	195.000	14.900	715.000	332.000	297.000	49.500	-0	2011.976
I DV100	62579	19.00	7.55	55.000	24.800	158.000	16.000	166.000	138.000	291.000	65.000	-0	914.950
I DV101	62579	20.00	7.60	57.000	22.000	108.000	14.000	122.000	128.000	242.000	70.800	-0	764.700
I DV113	62579	24.50	7.52	95.000	34.000	120.000	17.900	196.000	166.000	255.000	85.000	-0	969.640
I DV114	62579	22.50	7.52	112.000	41.300	128.000	14.600	262.000	161.000	262.000	80.000	-0	1061.350
P DV20	32079	21.00	7.60	50.000	19.700	102.000	6.800	134.000	78.800	200.000	40.800	-0	632.430
D LL098	92878	53.50	7.80	77.200	.800	446.000	10.500	675.000	164.000	64.900	74.000	-0	1513.550
D LL099	92878	73.50	8.70	3.900	.200	180.000	5.600	125.000	116.000	54.600	122.000	28.300	644.090
D LL100	92878	70.00	8.40	7.600	.200	195.000	5.200	156.000	129.000	87.400	117.000	9.470	706.960
D LL101	92878	25.00	7.40	51.800	2.900	232.000	2.600	250.000	160.000	163.000	52.700	-0	915.400
D DV23	32079	73.00	8.65	4.300	.200	181.000	6.000	133.000	108.000	56.000	123.000	28.000	653.505
D DV24	32079	55.00	7.80	67.500	.800	430.000	10.200	650.000	161.000	62.000	75.000	-0	1470.122
D DV054	62579	73.00	8.77	4.000	.200	175.000	5.500	130.000	112.000	46.500	118.800	31.100	637.597
D DH16	72379	26.00	7.65	55.000	2.950	228.000	2.270	250.000	150.000	166.000	44.000	-0	905.930
D DH15	72379	52.00	8.25	9.600	.050	210.000	4.330	208.000	102.000	76.600	112.000	7.200	743.850
D DH09	72379	73.00	7.80	31.000	.210	311.000	8.720	362.000	220.000	75.100	85.000	-0	1106.078
D DH06	72379	73.00	8.20	11.500	.120	216.000	5.420	160.000	202.000	76.600	109.000	9.600	803.875
D DH08	72379	57.00	-0	-0	-0	-0	-0	-0	-0	-0	102.000	-0	102.000
D DH10	72379	76.00	-0	-0	-0	-0	-0	-0	-0	-0	66.000	-0	66.000
H DV04	32079	63.00	6.61	46.500	10.300	362.000	22.000	49.000	122.000	936.000	68.500	-0	1630.842
H DV033	51579	65.50	6.45	49.000	10.800	342.000	21.200	47.000	116.000	919.000	67.000	-0	1586.787
H HHS01	72379	75.00	7.30	44.000	10.100	324.000	20.600	50.000	108.000	880.000	66.000	-0	1518.508
H HHS03	72379	67.00	6.50	47.000	10.000	335.000	21.300	47.000	112.000	911.000	69.000	-0	1569.563
H HHS05	72379	72.00	7.22	43.000	10.000	334.000	20.200	47.500	111.000	869.000	69.000	-0	1519.630
H HHS06	72379	39.00	-0	-0	-0	-0	-0	-0	-0	-0	83.000	-0	83.000
H HHS12	72379	63.00	6.90	47.000	9.900	322.000	20.400	50.000	110.000	844.000	66.000	-0	1525.196
H HHS09	72379	58.00	7.50	38.000	10.200	334.000	21.000	47.000	113.000	897.000	67.000	-0	1543.321
S DV01	32079	55.00	7.30	110.000	20.400	160.000	28.000	77.000	370.000	303.000	66.800	-0	1153.274
S DV02	32079	51.00	7.39	103.000	21.200	163.000	28.000	77.000	370.000	292.000	71.000	-0	1143.515
S DV035	51579	53.50	7.80	112.000	22.000	162.000	26.200	88.000	374.000	301.000	63.800	-0	1162.359
S DV036	51579	73.00	6.50	105.000	21.500	155.000	26.000	78.000	366.000	321.000	-0	-0	1090.306
S DV075	72379	73.00	6.05	105.000	20.800	150.000	22.000	76.000	348.000	309.000	53.000	-0	1103.713
S DV076	72379	55.00	-0	-0	-0	-0	-0	-0	-0	-0	58.000	-0	58.000
S DV077	72379	30.00	-0	-0	-0	-0	-0	-0	-0	-0	54.000	-0	54.000
S DV112	62579	73.50	6.80	105.000	20.500	162.000	28.000	78.000	352.000	313.000	60.000	-0	1123.400

Table 5-1. Chemical Analyses of Dixie Valley Waters. (cont'd.)

	F	H	AL	FE	MN	LI	SR	NO3	AS	HG	CS
W LL097	-0	-0	-0	-0	-0	-0	-0	.090	-0	-0	-0
W LL109	-0	-0	-0	-0	-0	-0	-0	.090	-0	-0	-0
W DV21	.800	.810	.100	.080	.017	.050	.860	-0	-0	-0	-0
W DV22	.370	.550	.100	.010	.015	.020	.780	-0	.010	.000	.050
W DV031	.190	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
W DV032	.320	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
W DV045	.300	-0	-0	-0	-0	-0	-0	-0	-0	-0	.050
W DV046	.290	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
W DV047	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
W DV048	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
W DV049	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
W DV050	.310	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
W DV051	.340	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
W DV052	.490	-0	-0	-0	-0	-0	-0	-0	-0	-0	.050
W DV053	.460	.550	.100	.010	.005	.039	1.210	0	.010	.000	.050
W DV055	.380	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV56	.380	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV057	.370	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV058	.320	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV059	.340	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV067	1.330	-0	-0	-0	-0	-0	.350	0	-0	-0	-0
W DV078	.570	.950	.100	.010	.005	.090	1.760	0	.010	.000	-0
W DV092	.300	1.100	.100	.010	.005	.030	.650	-0	.010	.000	.020
W DV102	.330	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV103	.220	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV104	.310	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV105	.260	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV106	.300	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV107	.360	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV108	.400	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV109	.620	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV110	.530	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
W DV111	1.370	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
C CS01	-0	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
C CS02	-0	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
C CS04	-0	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
C CS06	2.500	1.000	.100	.010	.240	.045	4.600	0	.010	.000	-0
C CS07	1.900	1.900	.100	.010	.032	.075	3.350	0	.010	.000	-0
C CS08	.330	1.700	.100	.010	.005	.050	1.630	0	.010	.000	-0
H DV041	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
R DV042	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
F LL102	-0	-0	-0	-0	-0	-0	-0	.710	-0	-0	-0
F DV03	.600	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
F DV10	.160	.240	.100	.020	.015	.020	4.250	-0	-0	-0	-0
F DV12	.300	.230	.100	.010	.015	.010	.690	-0	-0	-0	-0
F DV060	.360	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
F DV061	.290	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
F DV062	.300	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
F DV063	.180	.220	.100	.010	.005	.010	-0	0	-0	-0	-0
F DV091	.560	1.200	.100	1.610	.050	.050	.080	-0	.010	.000	.020
L LL103	-0	-0	-0	-0	-0	-0	-0	1.510	-0	-0	-0
L LL104	-0	-0	-0	-0	-0	-0	-0	1.240	-0	-0	-0

Table 5-1. Chemical Analyses of Dixie Valley Waters. (cont'd.)

	F	R	AL	FE	MN	LI	SR	NO3	AS	HG	CS
M DV05	3.100	.900	.100	.010	.015	.270	.690	-0	-0	-0	-0
M DV06	2.800	1.260	.100	.010	.015	.350	1.380	-0	-0	-0	-0
M DV11	1.300	.800	.100	.010	.005	.150	2.450	-0	-0	-0	-0
M DV115	1.370	-0	-0	-0	-0	-0	-0	-0	-0	-0	-0
M DV116	6.600	-0	-0	-0	-0	-0	-0	0	-0	-0	-0
1 DV069	1.550	5.200	-0	.030	.005	1.060	.160	-0	-0	-0	.130
1 DV70	8.500	9.200	-0	.100	.005	1.290	.190	-0	-0	-0	.290
1 DV72	9.500	9.600	-0	.100	.006	1.540	.320	-0	-0	-0	-0
2 DV080	-0	11.000	-0	-0	-0	1.290	-0	0	1.400	-0	.210
2 DV081	-0	-0	-0	-0	-0	1.270	-0	0	2.100	-0	.280
2 DV082	-0	10.600	-0	-0	-0	1.290	-0	0	1.800	-0	.200
3 DV090	7.600	8.500	.100	1.610	.050	1.010	1.060	-0	.590	-0	.325
4 DV093	7.900	12.700	-0	-0	-0	-0	-0	-0	.160	-0	-0
4 DV094	8.800	-0	-0	-0	-0	-0	-0	-0	.160	-0	-0
T DV15	4.390	-0	-0	-0	-0	.970	3.440	-0	-0	-0	-0
T DV16	4.400	5.500	-0	-0	-0	.960	-0	-0	-0	-0	-0
T DV030	4.400	4.700	.100	.040	.030	.975	3.200	-0	.040	.000	.140

All concentrations are in parts per million (ppm) and temperatures are in °C.

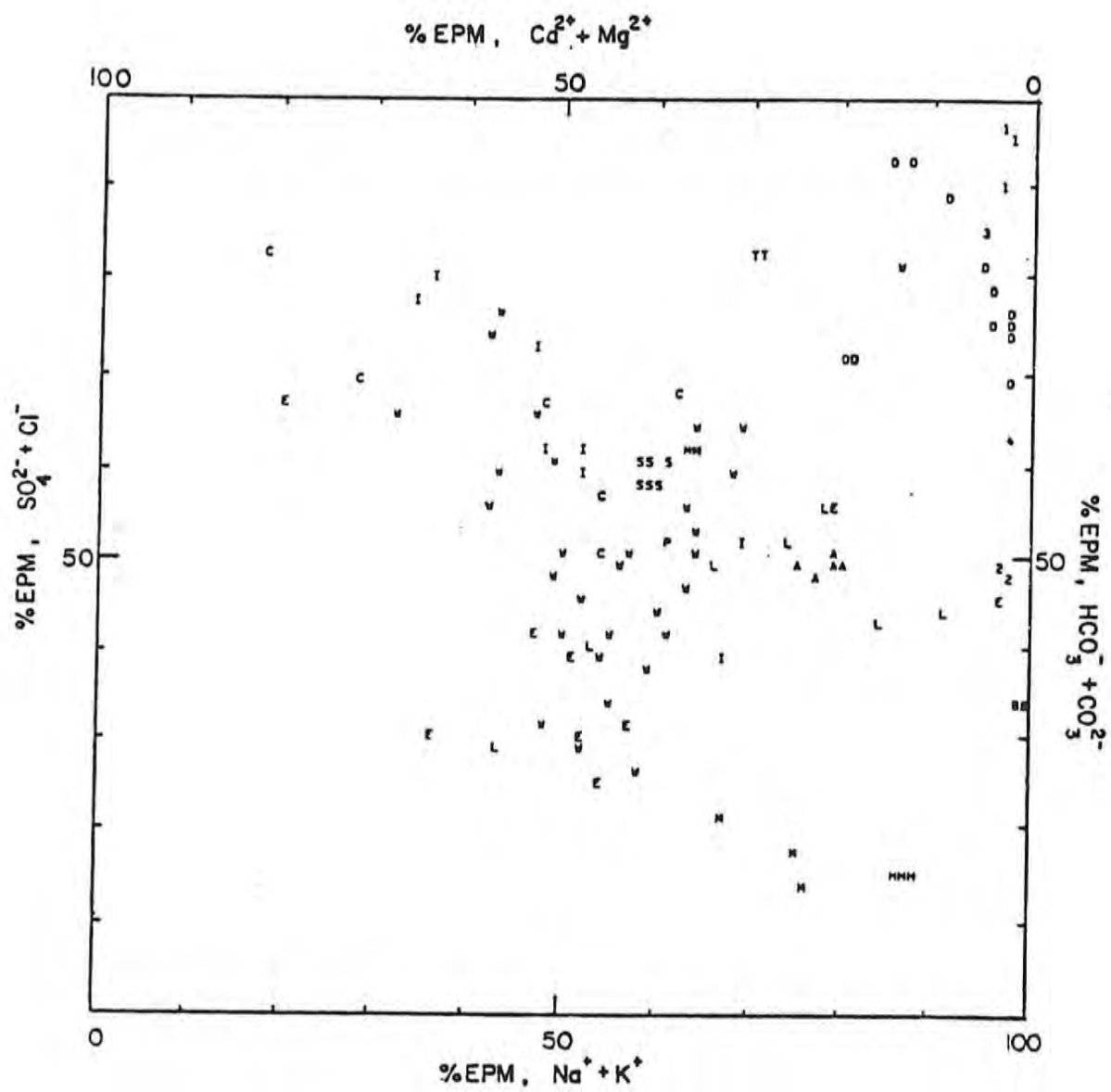


Figure 5-1. Trilinear Plot of Dixie Valley Waters.
 (See legend opposite Table 5-1 for letter
 and number code.)

All samples have comparatively low Ca and Mg levels. The hot spring systems, as well as the deep wells, seem to experience a substantial shift towards increased Na + K levels. The anions seem to indicate more clearly the differences among the three hot spring systems.

If one considers total dissolved solids (TDS), the variations in water chemistry become more apparent (Table 5-2).

Table 5-2 clearly illustrates the variations both within a given hot spring system and among the three hot spring systems. The Dixie Hot Springs system has the greatest TDS range, possessing both the lowest and highest levels. Sou and Hyder Hot Springs have very narrow TDS ranges, with Sou Hot Springs having the lower values of the two systems.

The artesian wells in the vicinity of Dixie Settlement have low values, similar to the surface water and spring water derived from the Clan Alpine Mountains. The hydrogeochemical evidence would tend to indicate that the Clan Alpine Mountains are a source of recharge for groundwater in the Dixie Settlement area. This situation is more likely than specifying the Stillwater Range waters as a source of groundwater recharge. These latter waters have high TDS levels for mountain spring water and surface runoff. These high TDS values definitely anomalous for "normal" mountain runoff will be discussed further in a later section.

5.2.2 Waters from the Clan Alpine Mountains

Most of the waters from the Clan Alpine Mountains are chemically similar to the waters from the wells at Dixie Settlement. Both groups of waters are low in TDS, but the Dixie Settlement waters exhibit relatively high temperatures for groundwater. Two Dixie Settlement samples, DV 13 and DV 66, have unusually high pH values (8.30 and 8.18) and fluoride levels (6.00 and 5.80 ppm). One anomaly in the Clan Alpine Mountains is Shoshone Spring (DV 91), which has a surprisingly low dissolved oxygen content for a mountain spring, a high pH (8.58) and temperature (32°C). The boron level in this spring, 1.2 ppm, is a significant amount.

Table 5-2. Total Dissolved Solids (TDS) Contents
of Dixie Valley Waters.

<u>Sample Location</u>	<u>TDS (ppm)</u>
Dixie Hot Springs	640-3000
Sou Hot Springs	1100-1160
Hyder Hot Springs	1580-1630
McCoy and Lower Ranch Hot Springs	800-1190
Eastern Mountain Ranges (Clan Alpine and Augusta Mountains)	170-435
Stillwater Range	600-1770
Artesian Wells in Southern Dixie Valley (Dixie Settlement)	350-370
Irrigation Wells in Northern Dixie Valley	760-2380
DF 45-14	1657-2201
DF 66-21	4100-5410
SR2-A	1900

Another important aspect is that the calculated CO_2 pressures are oversaturated with respect to atmospheric CO_2 pressure in many Clan Alpine Mountain waters (Table 5-3). These elevated CO_2 pressures are caused by various dissolution reactions occurring in the subsurface, and the CO_2 pressures are indicators of the extents of these various reactions. At the present time the sources of CO_2 in geothermal systems are poorly understood.

5.2.3 Waters from the Stillwater Range

Even in the early stages of field work it was recognized that the waters from the eastern slopes of the Stillwater Range, which is considered one of the major recharge areas for the study area, are surprisingly high in TDS. One normally expects low TDS waters in a recharge area. Springs on the Carson Sink (west) side of the Stillwater Range also exhibited high TDS levels. Although some thermal springs are believed to exist in the mountains, none was found. All sampled springs have low discharges and most streams deposit travertine (CaCO_3). All computed CO_2 pressures are greater than atmospheric CO_2 (see Table 5-3). This tends to imply the existence of a source of CO_2 other than the atmosphere.

Samples from the eastern slopes have Cl levels as high as 260 ppm, SO_4 as high as 520 ppm, HCO_3 up to 380 ppm, SiO_2 up to 46 ppm, Mg up to 80 ppm, Na + K up to 190 ppm and Ca as high as 180 ppm. These waters apparently comprise most of the shallow groundwater north of Dixie Hot Springs. The ratios Cl/SO_4 , $(\text{Na} + \text{K})/\text{SO}_4$, Ca/SO_4 and Mg/SO_4 are relatively constant in these samples. The constancy of these ratios is apparently characteristic of volcanic rock weathering, since volcanic rocks are abundant in the Stillwater Range.

The question remains as to why the Stillwater Range waters have such high levels of TDS, a phenomenon which is usually not anticipated from waters derived from precipitation. One reason could be that these waters have a relatively long residence time in the subsurface, a hypothesis weakly supported by the few available tritium data. However, the waters from the Clan Alpine Mountains are from a similar rock environment, but have much lower TDS levels.

A second explanation might be the rapid dissolution of hydrothermal minerals, by either old or young waters. Another possibility that could also explain the high calculated CO_2 pressures (Table 5-3)

Table 5-3. Calculated $\log P_{\text{CO}_2}$ Values in Dixie Valley Waters.

<u>Sample Locations</u>	<u>$\log P_{\text{CO}_2}$</u>
Stillwater Range	-3.4 to -1.5
Clan Alpine Mountains	-3.5 to -2.0
Artesian Wells (Dixie Settlement)	-3.5 to -3.7
Irrigation Wells	-2.5 to -1.8
Dixie Hot Springs	-4.0 to -2.6
Hyder Hot Springs	-1.4 to 0.0
Sou Hot Springs	-2.3 to 0.0
DF 45-14	-1.2
DF 66-21	-1.5
SR2-A	-2.2 to -2.0

is the ascension of thermal waters along deep reaching fault zones. These ascending thermal waters would then be slightly diluted by infiltrated precipitation, and the mixture would surface at springs.

Further work in the Stillwater Range itself will be required to verify the origin of the high TDS waters. It is possible that the isotope data will shed some light on this problem.

5.2.4 Thermal Waters in Dixie Valley

Thermal groundwaters commonly have chemical constituents that can serve as specific indicators of geothermal reservoirs. Elevated levels of F, Cl, B, SiO_2 , H_2S , Na and TDS can indicate geothermal activity. SO_4 can be high if sufficient sulfide and free oxygen are available at depth. HCO_3 is commonly high in thermal waters if an appropriate source of CO_2 exists at depth. However, it should be emphasized that F, B and other trace elements are low if the reservoir rocks lack these constituents. Ca and Mg levels are usually low in thermal waters due to cation exchange with clay minerals and other similar reactions.

In the following discussion the aforementioned indicators are utilized to determine relative circulation depths of waters and reservoir temperatures.

5.2.4.1 Dixie Hot Springs

The Dixie Hot Springs system is comprised of about 35 springs and seeps. The springs show wide variations in temperature and electrical conductivity. All springs emerge from alluvium, which is about 1000 feet thick in this area. Springs and seeps discharge over an area of about four square miles.

The Dixie Hot Springs system is typical for its relatively low HCO_3 (60-90 ppm). The pH ranges between 7.4 and 8.77, which is considered rather high. Electrical conductivities change significantly between adjacent spring orifices. Temperatures are generally highest for the lowest TDS springs (Figure 5-2). Cl and Na are highest for the low temperature springs. SO_4 and Cl correlate very well, and reach their highest levels in the coldest springs. SiO_2 decreases with increasing Ca, which indicates an increasing influx of cold

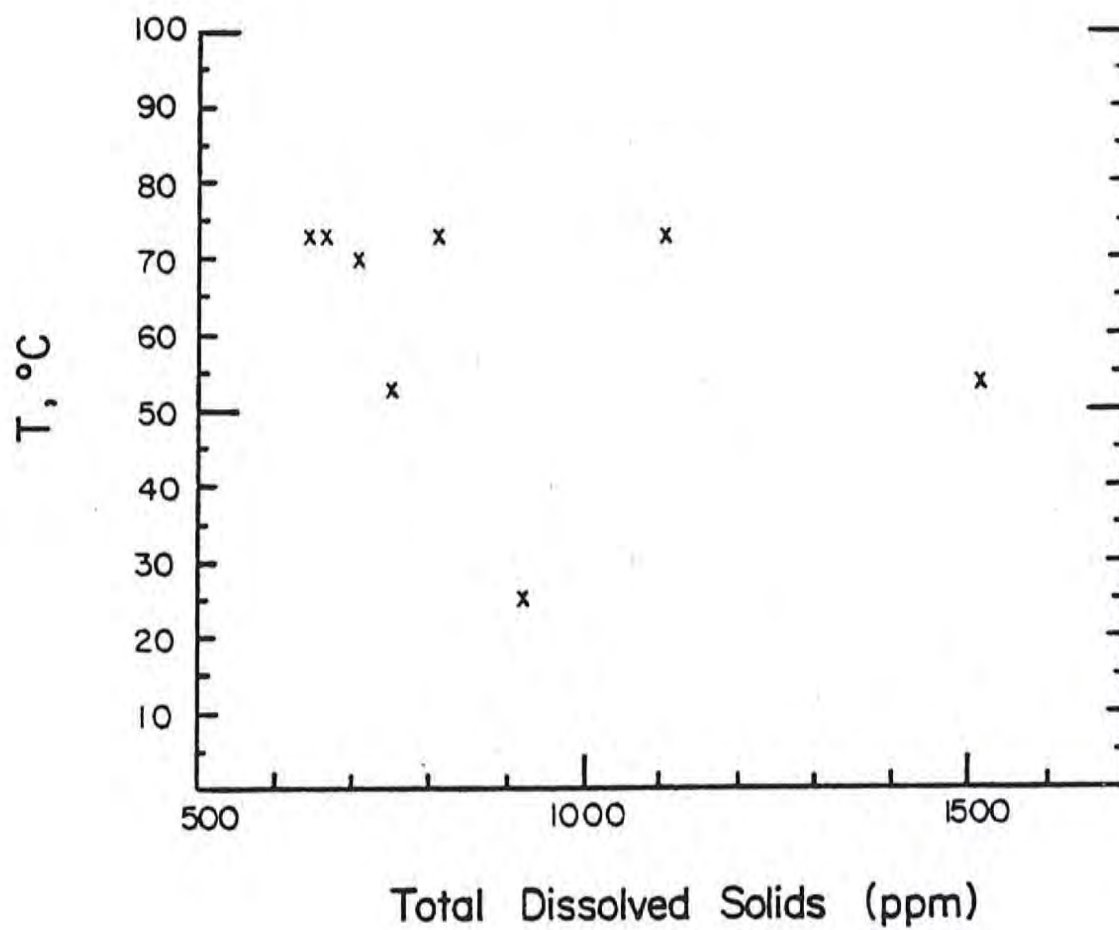


Figure 5-2. Temperature versus Total Dissolved Solids (TDS) for Dixie Hot Springs Waters.

water into the system. High Cl and Na levels indicate dissolution of shallow evaporites by cold water. F levels are the highest in the valley, whereas B, Li, As and Cs have comparatively low concentrations.

Table 5-4 shows estimated reservoir temperatures, calculated using chemical geothermometry. Conductive cooling was assumed for the silica method. Different temperatures between the two methods suggest that chemical equilibrium has not yet been attained.

The silica mixing model, based on a recharge water with 30 ppm SiO_2 , yields completely different results. For DV 23 a temperature between 225 and 235°C was computed. However, the mixing model is considered to be unreliable in this case, since the results are inconsistent.

In conclusion, the Dixie Hot Springs system probably shows various degrees of mixing between thermal and cold waters, which accounts for the varying temperatures and electrical conductivities. Based upon the structural analysis (Chapter 3), it is likely that the springs are fault-controlled. If fault control does exist then the colder waters with higher TDS could result from a less direct flow path that allows more cooling and more mixing within the alluvium which permits more dissolution of soluble salts. It is also possible that the low TDS waters are linked to the artesian wells of Dixie Settlement, since these are the only low TDS waters in the immediate vicinity. The increasing SO_4 with Cl (Figure 5-3) could also be attributed to cold waters from the Stillwater Range, since only there is the correlation between these ions good. Since the HCO_3 levels are low (less than 150 ppm), Dixie is probably not linked to any CO_2 source at depth. This accounts for the absence of any significant spring deposits. Low levels of most trace elements indicate that the Dixie Hot Springs system may be a relatively shallow system, whereas the high F could be due to the dissolution of ancient CaF_2 deposits. It is possible that the Dixie system is heated by a more extensive sealed geothermal reservoir at depth.

Table 5-4. Chemical Geothermometers Applied to Dixie Hot Springs Waters.

Sample No.	Quartz (°C)	Na-K-Ca (°C)	Field Temperature (°C)
DV23	149	139	73
DV54	147	137	73
DH6	142	97	73
DH15	143	95	52
DH9	128	100	73
DV24	122	88	55

5.2.4.2 Hyder Hot Springs

The Hyder Hot Springs system emerges in the middle of the valley, where bedrock is apparently covered by a thick cover of alluvium. However, the chemical homogeneity of the springs seems to require a well-defined, fault-controlled flow system. The springs have deposited a vast amount of travertine up to 100 feet thick. Some of the springs are surrounded by delicate travertine structures.

A detailed survey of the system showed that all springs have electrical conductivities ranging from 1600 to 1850 micromhos (temperature compensated). Gross chemistry of the springs does not vary much, although temperatures differ considerably, ranging from 39°C to 75°C. Thus the system is probably relatively homogeneous, issuing from one reservoir.

Hyder waters are of the bicarbonate-sodium-potassium type. The bicarbonate levels range between 870 and 936 ppm and are the highest among all the hot springs. They are about six times as high as those from DF 45-14 but somewhat lower than those from DF 66-21. The high HCO_3 accounts for the extensive travertine deposits. The pH is as low as 6.5 in springs with high HCO_3 and as high as 7.5 in those with low HCO_3 . Some evidence of degassing (probably CO_2) was found, which probably accounts for the inverse correlation of pH with HCO_3 .

The SO_4 and Cl levels are among the lowest of the hot springs in the valley (108-116 ppm SO_4 , and 47-50 ppm Cl). Sulfate is about as low as the first sample from DF 66-21. The relatively high bicarbonate concentrations indicate that Hyder is connected directly to a geothermal reservoir. In addition the low SO_4 indicates a relatively deep water with limited capability to oxidize sulfides (Figure 5-3). However, no evidence of H_2S was found in Hyder, and the dissolved oxygen content was as high as 2.9 ppm in some of the springs.

Since electrical conductivities and silica levels are relatively constant, the silica geothermometer, assuming conductive cooling and no mixing, was applied. Calculated temperatures range from 115°C to 127°C. The cation geothermometer yielded consistent temperatures between 159°C and 162°C. Slight calcite precipitation is possible for those springs with highest discharge temperature, but the increase in calculated temperature would be negligible in this case.

In conclusion, the Hyder Hot Springs system is of one water type and thus emerges from a well-defined flow system, with little mixing with shallow groundwater. The high HCO_3 values indicate connection to a CO_2 source at depth. However, the water chemistry is inconclusive; therefore the relationship of the spring water to the deep system is unknown.

5.2.4.3 Sou Hot Springs

Vast amounts of spring deposits are present at Sou Hot Springs. The chemistry of the hot springs do not vary considerably, and are of an intermediate type. The temperatures range between 30°C and 73°C . The pH is as low as 6.5; a pH of 6.05 was measured but it is believed to be too low due to a faulty meter. Among the three major hot spring systems, Sou has the highest Ca, Mg and SO_4 levels. Cl levels are comparatively low (less than 100 ppm) as are F and B (DV36 in Figure 5-4). Sr is the highest among all the hot springs. HCO_3 levels are lower (about 300 ppm) than they are at Hyder Hot Springs.

The silica geothermometer, assuming no mixing and conductive cooling, yields temperatures between 105°C and 118°C . The cation geothermometer yields temperatures between 93°C and 194°C . The drastic changes in the calculated reservoir temperatures are probably associated with seasonal fluctuations.

Evidence of extensive CO_2 degassing was found in a flowing well. This probably indicates that Sou Hot Springs is connected in some way to a CO_2 -producing reservoir at depth.

In conclusion, Sou is probably a relatively shallow hot spring system. The low SiO_2 temperature, low F, B and Cl and high Ca and Mg are indicative of relatively shallow circulating meteoric water that is heated to low temperatures.

5.2.4.4 Well SR2-A

The measured temperature of the discharge water was 65°C (DV30).

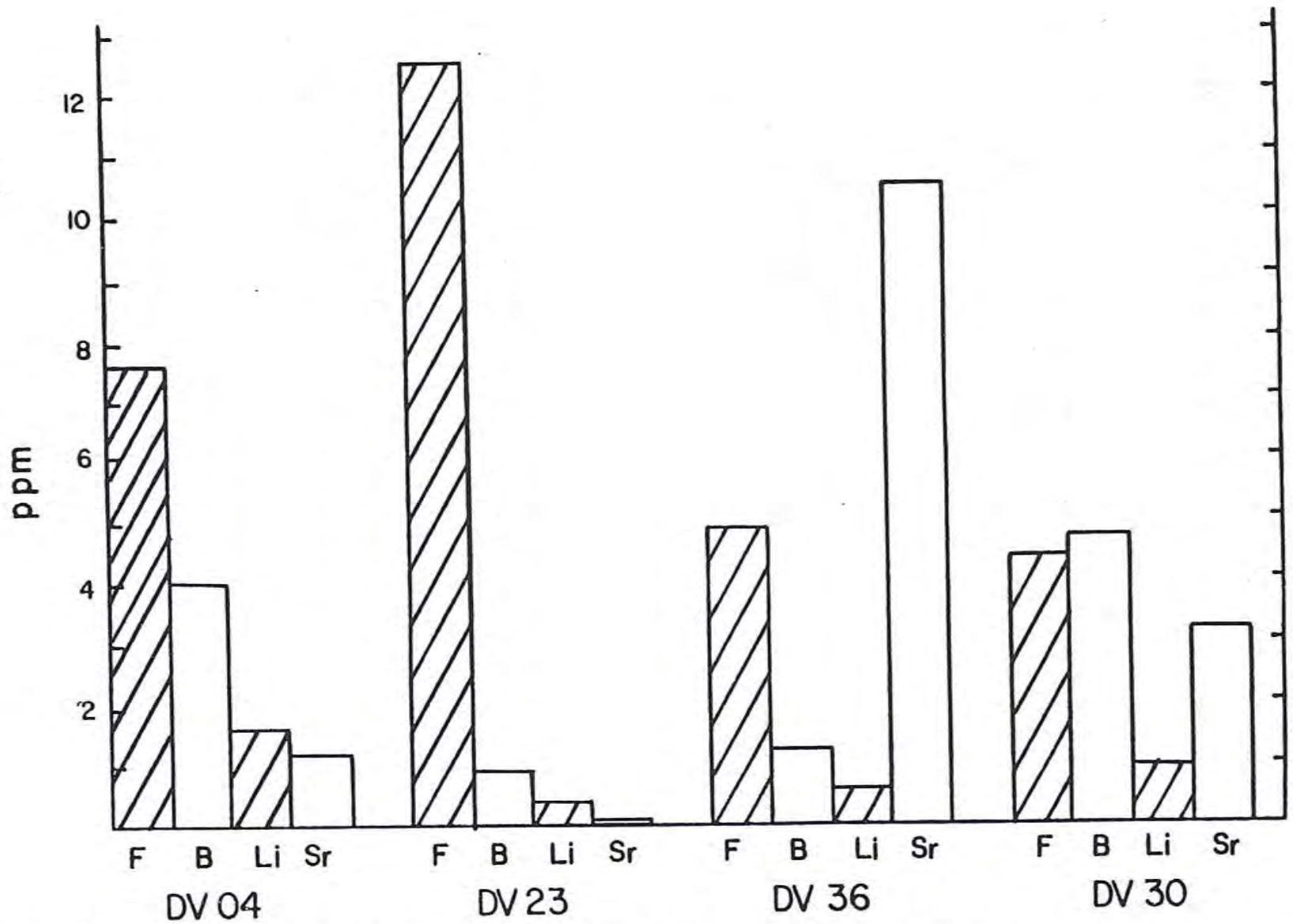


Figure 5-4. Fluoride, Boron, Lithium and Strontium in Selected Dixie Valley Waters.

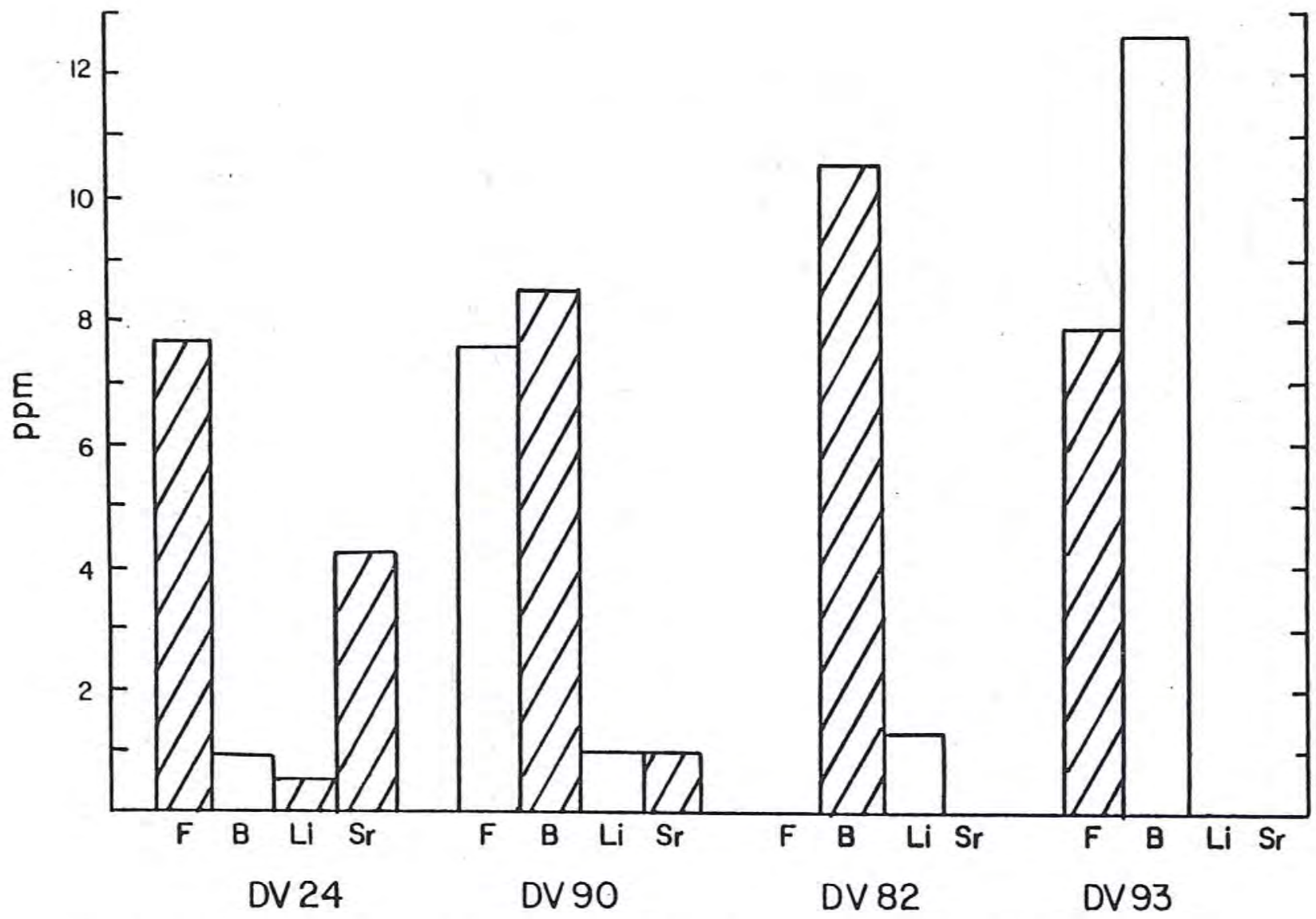


Figure 5-4 (con't.) Fluoride, Boron, Lithium and Strontium in Selected Dixie Valley Waters.

Ca and Mg levels are significant, indicating the influence of cold water. The Cl/SO₄ is very similar to that of the water from White Rock Canyon, indicating the influx of water from the Stillwater Range. F, B and Sr are also significant. Application of chemical geothermometers yields 91°C for the cation method and 139°C for the silica method. The different results indicate inadequate chemical equilibration of warm water at depth. However, the silica method seems to be more reliable. SR2-A is probably fed by a mixture of hot water from depth and cold recharge water from the Stillwater Range.

5.2.4.5 Deep Wells DF 45-14 and DF 66-21

A discussion of the water chemistry with special reference to the alteration mineralogy was given in Chapter 4. In this section the deep wells will be compared to the hot spring systems and the recharge waters. It must be indicated that more than 95% of our data is related to shallow groundwater systems and hot springs. The deep well samples are the only ones that represent deep thermal waters. However, the samples from DF 45-14 and DF 66-21 are not the most reliable because of suspected contamination from drilling operations and/or shallow groundwater. This is especially true of the samples from DF 66-21. Nevertheless, certain gross chemical features of the deep waters can be inferred from the existing samples.

DF 45-14 (DV90) has relatively low levels of Mg and Ca and high Cl and SO₄, which clearly indicate thermal waters. However, HCO₃ is low, much lower than in Sou Hot Springs. F is comparable to Hyder Hot Springs but much lower than in Dixie Hot Springs, whereas B is much higher than in any of the hot springs.

Application of the silica geothermometer (assuming no steam loss) indicates a reservoir temperature of 216°C for DF 45-14. Since the silica content at the sampling temperature is at the saturation level of amorphous silica, precipitation of amorphous silica might be expected in the ascending hot water. Thus, the silica temperature might be even higher. The cation geothermometer yields 193°C.

In DF 66-21 (DV82, DV93) the Cl and Na levels are about three times as high as in DF 45-14, although samples DV82 and DV93 are probably contaminated to a large extent. The SO₄ levels are comparable

to the hot spring systems. HCO_3 levels are extremely high. B and F are comparable to those of DF 45-14. Generally the fluids from DF 66-21 can be considered as Na-Cl waters, high in HCO_3 .

The silica geothermometer yields a temperature of 190°C (assuming no steam loss) and the cation geothermometer gives 139°C for DV82. For DV93 176°C was calculated with the silica method and 148°C with the cation method. The silica method is considered to be more reliable. DV82 is from a depth of about 4700 feet and DV93 from about 9500 feet in DF 66-21.

5.3 Conclusions

The preceding discussion about the chemistry of the three hot spring systems indicates that it is unlikely all three systems are linked to a common source. The Dixie Hot Springs system seems to be isolated from a deep CO_2 -supplying reservoir, and its waters probably originate from the Stillwater Range and/or other sources. The Sou and Hyder systems are different from each other as well as from the Dixie system as was seen in Figure 5-1. The comparatively low Cl levels in all three systems make a connection to a deep geothermal reservoir that supplies the deep wells unlikely. High HCO_3 levels could be produced by the upward diffusion of CO_2 from a source at depth. The fact that the maximum temperatures in all three systems are about the same (75°C) is conspicuous. F increases with SiO_2 if F is greater than 1 ppm, a relationship that applies to all the hot springs but not to the deep wells (Figure 5-5).

It is clear that the two deep wells tap waters that are different from all the other thermal waters in the valley. Additionally, DF 45-14 is apparently different from DF 66-21. The most significant differences are the higher HCO_3 in DF 66-21 and a TDS level almost double that of DF 45-14. However, since samples from DF 66-21 are undoubtedly contaminated to an unknown extent, these differences may be more apparent than real.

The origin of HCO_3 is unclear. One source could be the oxidation of organic materials at depth, evidence of which was found in some of the sedimentary rocks at depth (T. Bard, personal communication). Another source could be the dissolution of limestones under

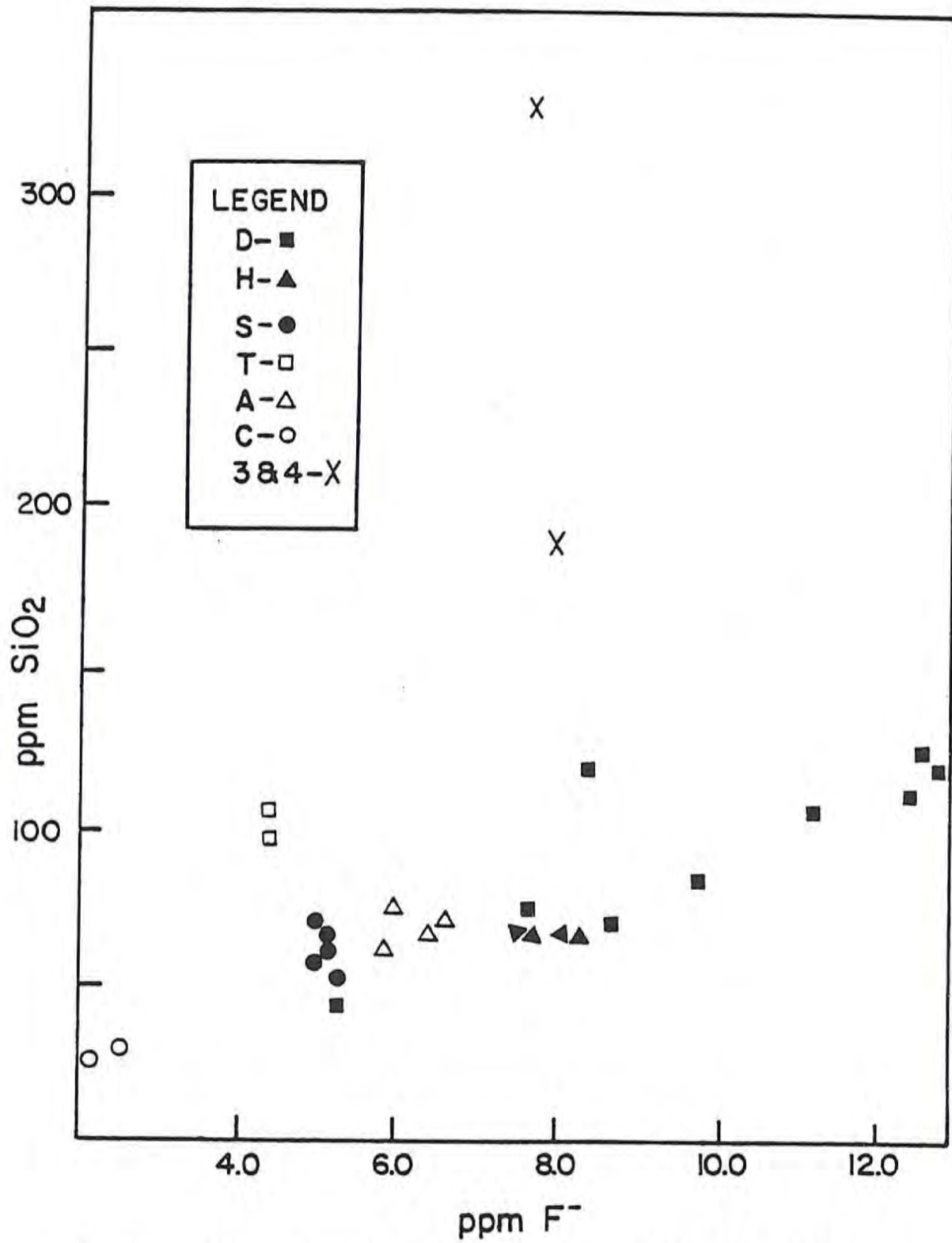


Figure 5-5. Silica versus Fluoride in Selected Dixie Valley Waters.
 (See legend opposite Table 5-1 for letter and number code.)

acidic conditions. Fracture zones could allow the CO_2 to escape to shallow aquifers, thus accounting for the high P_{CO_2} in most shallow groundwaters.

SO_4 probably originates from the oxidation of sulfides as well as the solution of SO_4 minerals. However, the level of SO_4 is probably limited at depth by the limited supply of free oxygen due to lack of deep-reaching circulation. This constraint might be active in both DF 45-14 and DF 66-21, thus accounting for the comparable levels of SO_4 .

It is still not clear how the three hot springs are related to the deep reservoir(s) tapped by DF 45-14 and DF 66-21. The collection and analysis of valid samples from the deep wells might provide more information, as might the isotope data. Age determinations using ^{14}C would prove useful in determining circulation depths. Sulfur isotopes would be helpful in determining the sources of SO_4 .

In summary, there appear to be different thermal systems in Dixie Valley. Their interrelationships are not clear, but it seems that there is little interaction among them other than occasional CO_2 transfer and conductive heat transfer.

5.4 References

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- Chapter 6. SHALLOW TEMPERATURE SURVEY

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6.0 SHALLOW TEMPERATURE SURVEY

6.1 Introduction

6.1.1 Purpose and Scope

The purposes for utilizing a shallow (1 meter) temperature survey in the Dixie Valley geothermal area are to: 1) delineate the near-surface hydrothermal discharge system; 2) compare the temperatures obtained from the survey to the data from the thermal gradient holes; and 3) test the usefulness of the shallow temperature survey as a rapid and inexpensive geothermal exploration technique in Dixie Valley. The shallow temperature survey did not cover the entire valley, but focused on the area of greatest interest to Southland Royalty Company, the northern half of Dixie Valley between Dixie Hot Springs and Sou Hot Springs. Plate VIII shows the locations of both the 1 meter holes and the thermal gradient holes.

6.1.2 Methods and Analytical Techniques

The shallow temperature holes were installed by augering to a depth of approximately 1 meter in a predetermined location. A length of 1 inch (ID) PVC pipe, sealed at the bottom, was then emplaced in the hole, and the hole was then backfilled. A 25%-75% mixture of anti-freeze and water was then poured into the pipe, and a removable cap placed on top of the pipe. Two hundred and forty-seven holes were installed in this manner. One hundred and seventy-five holes were emplaced by the end of March 1979; the remaining 72 were installed during May and June of 1979.

The network of holes was sampled once a month for six consecutive months (April through September). At each site, the ambient air temperature was recorded as was the temperature at the bottom of the fluid-filled hole. If necessary, fluid was added to the hole after measuring the temperature. During the early sampling runs, temperatures were recorded at the top, middle, and bottom of the hole. However, this practice was discontinued for two reasons: 1) it proved to be time-consuming; and 2) the bottom hole temperature was less affected by fluctuations in air temperature. At the end of each sampling run, the temperatures at a depth of 1 meter were plotted on a map similar to Plate VIII.

The analysis and interpretation of the data were performed by visual examination of the maps and by calculating and interpreting various statistical parameters (mean, standard deviation and range) for both the air temperatures and the 1 meter temperatures. The 1 meter temperatures were also compared to the data from the six thermal gradient holes.

6.1.3 Previous Work

No previous shallow temperature surveys have been conducted in Dixie Valley. Thermal gradient holes, some over 450 meters deep, have been drilled (see Plate VIII) in the valley. Data from these holes were provided to the authors by the Southland Royalty Company.

Shallow temperature surveys have been used in a variety of geological, geothermal and hydrogeological investigations. The first reported survey was by van den Bouwhijsen (1934), who used a thermocouple to measure temperatures at a depth of 1.5 meters to detect a buried fault zone in Holland. Poley and van Steveninck (1970) used temperatures at a depth of 2 meters in an attempt to delineate salt domes and faults near a natural gas field.

Shallow temperature surveys have shown to be very useful in groundwater investigations. The fact that subsurface temperature data can yield information on groundwater flow regimes has been known for some time. The first significant advance in applying subsurface heat measurements to groundwater flow problems was made by Stallman (1960) who derived a general partial differential equation relating conductive and convective heat flow through a saturated porous medium. He suggested that heat flow measurements could be useful in the study of groundwater flow systems. Stallman (1965) and Bredehoeft and Papadopoulos (1965) developed solutions of Stallman's (1960) equation for vertical water flow under certain boundary conditions. Cartwright (1970) applied the solution of Bredehoeft and Papadopoulos to groundwater discharge from the Illinois basin.

Groundwater flow in shallow aquifers has been shown to affect subsurface temperatures. Cartwright (1968) used soil temperatures at a depth of 50 centimeters to prospect for shallow glacial and alluvial aquifers. He based his method on the assumption that an aquifer acts as a heat sink. Birman (1969) studied shallow soil temperatures in

Johnson Valley, California, and developed an empirical relationship between shallow subsurface temperatures and the depth to the water table. Parsons (1970) studied the relationship between groundwater flow and subsurface temperatures in a glacial complex in Canada. He made comparisons between measured temperature profiles and those obtained by digital computer simulation. Cartwright (1971) developed a model to demonstrate the effect of a shallow confined aquifer on soil temperature. He concluded that the quantity of geothermal heat redistributed by moving groundwater in a shallow aquifer depends upon the velocity of the fluid and the thermal properties of the rock-fluid complex. Supkow (1971), using field data in conjunction with computer simulation developed a "valley mapping function" to delineate zones of maximum groundwater flow. His interpretations of subsurface temperature data in the Tucson basin were supported by independent studies of the subsurface properties of the Tucson basin groundwater reservoir. Cartwright (1974) measured soil temperatures in parts of Illinois and found that the greater the groundwater flow, the greater the temperature difference between recharge and discharge zones. He also discovered that the effect of fluid movement on soil temperature decreases with increasing horizontal distance between recharge and discharge zones. Smith (1974) used a shallow temperature survey to prospect for a buried glacial valley aquifer. He was unable to distinguish a summer low-temperature anomaly trend, which is indicative of a buried valley. He attributed his lack of success to the depth of the aquifer (90-150 feet), which was too deep to contact the summer portion of the annual temperature wave. Because of this, a near-surface low-temperature anomaly was not produced.

Geothermal investigations have also made use of shallow temperature surveys. Kintzinger (1956) conducted a 1 meter survey to outline the thermal effects in an area near Lordsburg, New Mexico where newly drilled wells discharged steam and boiling water. Kappelmeyer (1957) measured temperatures at a depth of 1.5 meters to locate natural steam occurrences in Italy. Banwell (1964) prospected the Wairakei hydrothermal region in New Zealand using a 1 meter temperature survey. He reported "some measure of correlation" between the spatial distribution of temperatures and the locations of some major faults. Combs (1975) stated that shallow temperature surveys have limited value for selecting deep drilling sites because of their ineffective depth of penetration

and the masking effects of shallow groundwater. Olmsted (1977) conducted 1 meter temperature surveys in Nevada and reported varying degrees of success in delineating temperature and heat flow anomalies at greater depths. Miller (1978), using a 1 meter survey in the Allen Springs - Lee Hot Springs area of Churchill County, Nevada, found evidence suggesting a structural control on the discharge portion of the hydrothermal system.

6.2 Analytical Results

6.2.1 Shallow Temperature Survey Data

Table 6-1 contains the entire assemblage of data collected during the survey. Table 6-2 shows the statistical data for the bottom hole temperatures, and Table 6-3 contains the statistical information for the air temperatures collected during the sampling runs.

6.2.2 Visual Delineation of Temperature Trends

A visual examination of the data reveals that, to no surprise, the strongest thermal activity lies in the vicinity of Dixie Hot Springs, where a maximum temperature of 42.1°C was measured in August. This was the highest temperature measured during the entire survey.

The single line of holes running across the valley from Shoshone Creek to the west side of the valley also showed an interesting trend. The easternmost end of this line (hole 6) showed higher temperatures than the far western end (hole 5) of the line. This difference was not apparent during the first survey (April), but became evident during later surveys. Temperature differences between these two holes ranged from a low of less than 1°C in April to a high of 6.9°C in July. Between these two holes there was no clear trend.

Interesting temperature trends were also noted in the isolated grid of holes located in the Bernice Creek-Dyer Flat area. The April survey revealed a temperature difference of just 2.8°C between the hottest (hole 98) and coldest (hole 91) locations. In May, June and July the temperature differences between these two holes were greater than 5.5°C but by September the difference dropped down to about 3.3°C . In fact, during April and September the temperatures in this area showed

Table 6-1. Monthly Shallow (1 meter) Temperatures in Dixie Valley, Nevada. (All temperatures in °C; I indicates no measurement taken.)

MOILE	APRIL	MAY	JUNE	JULY	AUGUST	SEPT	OCTOBER
1	10.7	14.9	22.7	26.0	28.0	25.9	
2	10.1	14.8	23.0	26.6	27.7	25.2	
3	10.6	15.3	24.2	27.2	29.3	26.7	
4	I	I	I	I	I	I	
5	9.1	10.8	I	18.1	21.0	21.5	
6	9.7	13.5	22.1	25.0	26.0	25.0	
7	9.9	13.2	20.7	23.1	25.5	23.9	
8	10.1	13.5	21.2	24.0	25.9	24.3	
9	9.5	13.1	20.5	23.0	25.1	23.5	
10	9.8	13.1	19.5	23.5	25.0	23.5	
11	10.1	13.2	20.5	23.5	25.8	23.8	
12	9.5	12.1	19.2	22.0	24.0	23.1	
13	9.2	13.3	21.7	23.3	25.5	24.0	
14	9.6	13.3	20.8	23.2	25.4	24.1	
15	8.7	13.3	21.7	23.8	25.7	24.3	
16	8.7	12.9	19.1	22.9	24.5	24.3	
17	9.9	12.8	19.0	22.5	23.2	22.3	
18	9.1	12.4	18.9	22.1	24.5	23.3	
19	8.9	11.8	19.0	22.0	24.4	23.6	
20	8.6	11.7	18.0	20.0	23.0	22.3	
21	11.0	14.9	21.8	24.9	26.5	25.9	
22	11.1	15.4	23.2	26.8	26.8	26.4	
23	12.0	15.4	23.0	26.7	27.8	26.5	
24	12.3	16.5	23.9	27.4	28.5	27.2	
25	13.8	18.5	25.2	28.4	29.3	28.2	
26	11.2	16.1	24.1	I	I	I	
27	11.0	15.8	23.4	27.1	28.2	I	
28	10.9	14.7	22.4	30.1	27.0	26.0	
29	10.2	13.5	20.1	23.5	I	I	
30	11.7	14.7	22.3	26.2	27.3	27.0	
31	10.6	14.5	22.4	26.3	27.2	25.9	
32	11.0	15.0	22.6	26.6	30.6	26.2	
33	I	I	I	I	I	I	
34	10.2	14.3	21.2	24.6	26.0	24.4	
35	10.5	14.5	21.9	25.6	26.6	24.9	
36	8.9	15.2	23.6	27.2	26.2	25.9	
37	10.4	15.6	24.0	27.3	28.8	26.2	
38	19.9	25.6	32.5	36.2	38.1	36.1	
39	24.4	30.8	I	40.1	42.1	39.1	
40	11.6	17.2	25.1	28.2	29.9	28.2	
41	10.3	13.1	19.0	22.0	23.8	22.7	
42	10.4	15.0	22.7	26.2	27.8	26.8	
43	10.1	14.3	21.2	24.7	26.0	24.5	

Table 6-1 (cont'd.)

Hole	April	May	June	July	August	Sept
44	10.5	16.0	21.9	25.5	27.6	26.7
45	8.7	13.7	21.4	25.6	26.9	24.0
46	9.3	14.5	21.5	25.4	26.9	24.8
47	10.5	15.3	22.8	26.6	27.8	25.8
48	10.5	15.7	23.2	26.4	27.8	25.7
49	I	-0	I	I	I	I
50	10.8	17.1	22.8	26.3	27.2	25.9
51	10.5	14.3	21.0	24.5	I	I
52	9.7	15.0	23.0	26.8	28.0	25.1
53	I	13.8	19.2	22.8	24.4	22.9
54	I	12.1	14.0	18.0	20.6	20.8
55	I	16.4	23.9	27.3	29.6	27.5
56	I	14.6	20.1	23.7	26.1	24.2
57	8.2	10.8	15.4	18.4	20.6	22.0
58	9.3	14.0	22.2	25.4	27.0	25.1
59	10.0	14.1	21.9	25.3	26.6	24.8
60	10.4	13.8	I	23.3	24.2	23.9
61	-0	21.5	30.1	33.8	34.3	31.5
62	9.9	14.9	23.2	25.3	28.1	26.1
63	9.4	12.2	17.6	19.0	22.7	21.3
64	9.1	12.1	20.3	22.9	24.5	22.7
65	9.4	I	17.5	21.0	22.0	21.8
66	9.8	13.6	I	24.9	I	I
67	7.8	9.9	17.0	18.0	21.8	21.2
68	8.4	10.6	16.4	18.7	21.9	21.2
69	9.7	14.7	22.4	26.0	I	25.9
70	9.4	14.1	21.8	25.3	I	24.7
71	10.9	13.9	21.1	24.9	I	25.0
72	10.6	14.9	22.5	25.9	27.2	26.6
73	11.2	13.3	21.0	23.8	25.9	25.2
74	11.3	14.0	24.4	24.8	27.0	26.2
75	10.9	14.9	23.4	26.8	28.4	27.0
76	10.4	12.8	27.2	23.9	25.7	24.8
77	8.9	11.1	17.6	21.2	22.0	21.9
78	8.8	11.5	18.2	20.3	22.2	21.2
79	9.2	11.2	18.8	22.0	23.5	22.9
80	8.5	11.1	17.4	19.8	21.9	20.9
81	8.4	12.5	20.4	22.8	24.6	23.5
82	9.9	11.8	I	23.1	25.3	23.9
83	9.7	12.4	20.5	23.7	25.0	24.0
84	-0	12.4	21.3	24.8	25.8	24.7
85	10.0	13.9	22.2	25.2	26.8	25.8
86	10.1	14.5	22.7	26.0	27.7	26.1
87	15.2	18.4	24.1	26.5	27.3	26.8

Table 6-1 (cont'd.)

Hole	April	May	June	July	August	Sept
88	I	13.3	21.7	25.3	27.0	25.5
89	10.1	15.2	24.4	26.8	29.3	26.0
90	9.6	14.0	22.3	I	27.7	24.7
91	8.3	11.0	17.8	19.5	23.5	22.9
92	9.2	13.0	21.3	22.5	25.4	23.3
93	10.3	16.0	24.6	27.1	24.4	26.5
94	9.8	14.1	21.3	23.8	27.1	24.3
95	9.7	18.2	24.5	26.5	29.4	24.9
96	11.0	17.8	23.6	26.1	I	25.8
97	10.2	15.4	21.1	22.8	I	23.6
98	11.2	17.5	23.9	25.5	I	26.1
99	9.3	15.1	26.1	23.1	I	23.5
100	9.4	15.8	21.0	24.0	I	23.7
101	9.2	12.6	18.5	22.0	23.2	22.8
102	9.4	11.7	19.8	22.2	26.1	24.3
103	10.7	14.3	23.1	25.0	28.8	26.5
104	10.7	13.2	20.5	23.0	26.4	23.7
105	11.0	13.1	20.5	23.6	26.3	24.7
106	9.9	11.7	16.8	18.3	21.8	20.7
107	9.2	I	16.0	17.8	21.6	20.4
108	10.4	14.2	22.6	25.7	28.0	25.2
109	9.2	11.6	19.0	22.5	25.6	23.4
110	10.1	I	18.9	22.4	25.1	23.0
111	9.2	I	16.9	18.8	23.3	22.5
112	8.9	11.4	19.3	23.0	25.5	22.8
113	9.1	10.6	17.9	21.0	24.2	21.8
114	10.8	14.1	I	25.8	I	I
115	9.9	14.4	21.2	23.9	26.8	24.9
116	10.7	13.4	19.9	24.0	25.8	24.3
117	9.1	11.9	18.4	20.0	22.6	20.8
118	10.1	13.6	21.2	23.7	26.2	23.8
119	10.7	13.4	21.6	23.8	26.1	24.1
120	9.9	12.7	17.0	19.0	22.8	22.8
121	11.7	14.7	22.8	25.7	26.7	25.8
122	10.6	13.7	21.2	23.8	27.1	24.8
123	10.7	I	23.2	25.9	26.3	26.4
124	9.6	15.8	22.1	25.4	26.9	25.0
125	I	14.2	I	I	I	25.8
126	I	16.1	22.5	26.2	24.0	21.5
127	I	15.4	22.8	26.0	28.7	26.2
128	I	13.3	I	20.7	22.1	I
129	I	10.0	14.3	16.8	20.6	I
130	I	14.2	21.1	24.8	I	I
131	I	14.6	21.1	23.5	25.7	23.8

Table 6-1 (cont'd.)

Hole	April	May	June	July	August	Sept
132	I	15.1	20.9	23.6	I	25.0
133	I	I	I	I	I	I
134	10.1	15.2	22.6	26.7	28.3	25.7
135	9.4	14.8	I	I	25.4	23.5
136	I	I	I	I	I	I
137	8.9	12.8	18.9	21.9	24.5	23.7
138	I	15.4	I	23.5	I	24.7
139	12.2	18.3	26.0	29.5	30.2	28.5
140	I	I	21.7	I	I	24.9
141	11.5	-0	I	I	I	I
142	10.5	15.8	21.9	25.3	26.1	24.7
143	10.2	16.2	24.4	28.3	29.4	27.1
144	9.8	17.9	25.9	29.8	30.3	26.9
145	10.0	16.7	I	26.5	27.7	26.2
146	10.3	14.9	24.8	28.4	28.8	26.3
147	10.5	14.9	22.8	25.6	27.5	25.2
148	13.3	18.9	27.5	31.2	32.9	29.3
149	16.7	21.5	29.2	33.2	34.8	32.0
150	8.7	14.4	21.5	25.1	26.0	24.0
151	10.4	14.6	22.3	26.1	27.7	26.0
152	9.7	14.2	22.8	25.6	27.8	25.1
153	10.6	14.3	21.4	25.2	26.1	24.9
154	10.0	16.7	23.1	26.7	28.5	26.5
155	10.1	16.8	23.4	26.7	28.2	26.2
156	9.2	16.1	I	27.0	28.7	26.5
157	9.5	14.8	21.9	25.5	26.8	I
158	9.2	14.4	22.4	26.1	28.1	I
159	11.0	14.7	I	24.9	I	I
160	I	15.4	I	27.0	I	I
161	10.0	15.1	I	I	I	I
162	12.1	17.9	I	I	I	I
163	10.5	I	26.7	30.4	I	29.4
164	I	I	I	I	I	I
165	10.7	16.1	23.8	27.1	I	I
166	11.6	17.8	25.7	29.3	30.6	28.8
167	10.4	-0	I	I	I	I
168	9.4	14.2	21.8	24.8	26.0	24.6
169	9.8	14.2	22.2	29.0	26.9	25.2
170	11.1	15.4	24.6	28.3	29.1	27.5
171	11.8	16.1	24.6	27.9	28.6	27.1
172	10.9	14.5	I	26.1	27.1	26.0
173	11.1	15.9	I	28.2	29.7	27.2
174	10.9	15.1	I	I	28.9	26.7
175	10.6	16.0	I	I	30.2	27.2

Table 6-1 (cont'd.)

Hole	April	May	June	July	August	Sept
176	10.0	13.5	I	25.3	I	26.8
177	9.6	13.7	21.4	I	26.6	23.6
178	9.1	13.9	21.3	I	26.1	23.8
179	I	I	I	I	I	I
180	I	I	21.8	25.5	27.2	25.0
181	I	I	I	28.8	29.6	I
182	I	I	I	I	I	I
183	I	I	I	I	I	I
184	I	I	21.0	25.1	26.7	25.5
185	I	I	22.3	24.6	26.2	24.0
186	I	I	24.2	27.8	29.5	27.9
187	I	I	24.5	27.9	29.8	28.2
188	I	I	22.4	27.1	28.0	26.4
189	I	I	23.0	26.3	28.0	26.5
190	I	I	22.9	26.6	28.0	25.9
191	I	I	22.8	26.3	27.9	25.9
192	I	I	22.4	I	28.1	26.2
193	I	I	22.2	I	27.5	24.7
194	I	I	23.5	I	29.0	26.1
195	I	I	22.0	I	I	I
196	I	I	23.1	26.3	I	I
197	I	I	I	I	I	I
198	I	I	I	I	I	I
199	I	I	I	I	I	I
200	I	I	21.2	22.7	25.2	23.9
201	I	I	18.0	21.0	23.2	23.4
202	I	I	16.7	I	I	I
203	I	I	19.8	23.0	24.4	23.4
204	I	I	19.8	23.0	24.0	23.1
205	I	I	17.2	I	I	I
206	I	I	15.3	18.2	I	I
207	I	I	19.9	23.6	24.8	23.8
208	I	I	23.9	26.7	27.6	25.3
209	I	I	22.7	25.5	27.1	23.8
210	I	I	22.8	25.8	27.3	24.9
211	I	I	22.9	25.8	I	I
212	I	I	23.1	27.2	28.1	26.0
213	I	I	I	I	I	I
214	I	I	22.3	24.4	25.5	24.0
215	I	I	22.2	25.4	26.5	24.4
216	I	I	23.7	26.1	27.5	24.8
217	I	I	I	I	I	I
218	I	I	23.1	I	28.6	I
219	I	I	22.7	26.0	28.5	27.2

Table 6-1 (cont'd.)

Hole	April	May	June	July	August	Sept
220	I	I	18.3	21.8	24.2	25.6
221	I	I	18.1	21.0	23.0	23.8
222	I	I	17.4	21.0	24.5	25.2
223	I	I	18.3	21.5	24.4	24.9
224	I	I	22.2	25.2	26.1	25.3
225	I	I	18.3	21.0	23.0	22.8
226	I	I	I	I	I	I
227	I	I	I	I	I	I
228	I	I	23.8	I	I	I
229	I	I	24.0	I	I	I
230	I	I	24.3	27.8	29.9	28.5
231	I	I	31.9	I	I	33.8
232	I	I	31.9	35.4	I	34.4
233	I	I	30.1	I	I	I
234	I	I	42.0	I	I	I
235	I	I	27.1	31.0	32.0	28.9
236	I	I	18.9	22.0	23.4	22.0
237	I	I	23.2	I	28.8	26.2
238	I	I	24.9	28.5	29.9	27.1
239	I	I	22.4	25.7	27.1	25.6
240	I	I	20.8	23.9	25.0	23.6
241	I	I	20.3	23.2	25.3	23.4
242	I	I	21.8	I	I	25.5
243	I	I	22.6	26.9	27.9	25.8
244	I	I	21.7	24.8	I	I
245	I	I	I	24.6	I	I
246	I	I	I	26.2	28.0	I
247	I	I	I	27.4	28.4	26.9
248	I	I	20.6	24.3	25.5	24.3
249	I	I	22.9	26.4	27.5	24.8
250	I	I	22.0	25.3	26.1	25.5
251	I	I	24.4	27.0	29.5	26.5
252	I	I	24.3	26.5	29.4	26.6
253	I	I	21.8	24.5	27.7	24.2

Table 6-2. Statistical Parameters for Dixie Valley Shallow Temperature Data.

Month	Number of Sites	Mean (°C)	Standard Deviation (°C)	Maximum Temperature	Hole Number	Minimum Temperature	Hole Number	Range (°C)
April	156	10.4	2.0	19.8	38	7.8	4	12.0
May	163	14.5	2.5	30.8	39	9.9	67	20.9
June	207	21.9	3.2	42	234	14.3	129	27.7
July	219	25.0	3.2	40.1	39	16.8	129	23.3
August	197	26.7	2.8	42.1	39	20.6	54,57,129	21.5
September	206	25.2	2.4	39.1	39	20.4	107	18.7

Table 6-3. Means and Standard Deviations of Monthly Air Temperatures at Dixie Valley Shallow Temperature Sites.

Month	Number of Sites	Mean (°C)	Standard Deviation (°C)
April	158	11.7	2.4
May	152	22.9	2.7
June	68	31.1	3.9
July	213	33.3	4.4
August	193	33.1	3.6
September	204	30.9	3.9

less variation than during May, June, July and August.

The grid encircling Hyder Hot Springs also produced some interesting results. The holes closest to the springs generally had lower temperatures than those farther away. These differences were pronounced in June and July (maximum differences greater than 8.8°C) than in August and September (maximum differences on the order of 3.9°C). It is possible that the lower temperatures are caused by evaporative cooling on the hot spring mound. Other than these variations in maximum temperature differences, the temperatures in the Hyder Hot Springs area were not particularly high. The same statement can be made for the Sou Hot Springs-Seven Devils Springs area.

A region of high temperature is hole 61, located in Section 16 about 2.5 miles west of the Boyer Ranch. A maximum temperature of 34.3°C was measured in August. A few holes to the east of hole 61 showed relatively high temperatures, but those to the south and west did not exhibit such high temperatures.

One trend that seems apparent even from a cursory visual examination is the change in the "uniformity" of the temperatures. The April temperatures are the most uniform. The sole exceptions are a few locations in the vicinity of Dixie Hot Springs. This apparent uniformity becomes less obvious as the surveys progress into the summer. However, the same general patterns occur each month.

It is apparent (and has been all along) that the shallow temperatures are strongly influenced by the local hydrologic system. We were somewhat surprised not to see the effects of the July 1979 flash floods on some of the temperatures. These floods occurred along portions of the western side of the valley. One would expect infiltrating flood waters to affect the temperatures in certain areas. This expectation is not strongly supported by visual examination of the August and September temperature data. It is possible that the effects of these floods had not manifested themselves by the last (September) survey or that the effects were not very substantial. Since no quantitative data exist for these floods, it is virtually impossible to quantify their effects on the shallow temperatures. However, temperatures are lower on the alluvial fans directly below the areas where the mountain canyons enter the valley. These lower temperatures could be manifestations of groundwater flow, possibly mountain-front recharge. The

lower temperatures could also be the result of previous, undocumented infiltration events.

Another observation is that the temperatures generally tend to increase as one approaches the Stillwater Range. This could possibly indicate stronger thermal activity around or beneath these mountains.

A study of Tables 6-2 and 6-3 shows a strong correlation between the mean monthly air temperatures and mean monthly 1 meter temperatures. The fact that the air temperature standard deviation for a given month is greater than that for the shallow temperatures indicates that the 1 meter temperatures are less susceptible to variations caused by topography, vegetation and other factors.

6.2.3 Relationship of Shallow Temperatures to Thermal Gradients

Table 6-4 gives the relevant data on the six thermal gradient holes; the locations of these holes are shown on Plate VIII. Table 6-5 shows temperatures in each gradient hole at various depths as well as the June temperatures of the 1 meter hole nearest each hole. The June temperatures were used because they showed the greatest variability (highest standard deviation). Table 6-5 also shows the correlation coefficients from a linear regression analysis which relates the 1 meter temperatures (dependent variable) to temperatures at various depths in the thermal gradient holes (independent variables).

There is a fair correlation between the 1 meter temperatures and the temperatures in the gradient holes. The correlation appears to be stronger at a depth of 1000 feet in the gradient hole, although the amount of data used is insufficient to make a definitive statement. Although not presented in Table 6-5, a correlation coefficient of 0.841 was obtained for a depth of 1500 feet. This number was obtained by assuming that the temperatures at the bottoms of both DD-9 and SR2-A (total depth = 1420 feet each) could be extrapolated to a total depth of 1500 feet. The poorer correlation at 250 and 500 feet (especially the latter) could be the result of quenching by groundwater flow. Groundwater quenching is probably the cause of the low thermal gradients in H-2 and S-8. The latter gradient hole is probably strongly affected by groundwater flow from White Rock Canyon Creek. DD-9 has a relatively cool 1 meter hole nearby but a high thermal gradient.

Table 6-4. Data from Dixie Valley Thermal Gradient Holes.

Hole No.	Depth (feet)	Maximum Temperature (°C)	Thermal Gradient (°C/100')
S-8	500	24	0.2
H-1	1500	97	4.7
H-2	1500	51	2.3
DD-9	1420	84	4.8
SR2	500	63	5.6
SR2-A	1420	91	3.6

Table 6-5. Correlations Between 1 Meter Temperatures and Temperatures at Various Depths in Thermal Gradient Holes.

Gradient Hole	Nearest 1-Meter Hole		Gradient Hole Temperatures (°C) at Various Depths		
	Number	June T (°C)	250 feet	500 feet	1000 feet
S-8	155	24	19	24	--
H-1	61	26	61	68	82
H-2	34	21	25	30	41
DD-9	141	22	38	51	72
SR2	49	26	60	62	--
SR2-A	49	26	58	63	80
Correlation Coefficient			0.782	0.700	0.839

This could indicate a lack of groundwater quenching in DD-9. The other three gradient holes have both relatively high thermal gradients and warm 1 meter temperature holes nearby.

6.3 Conclusions

Shallow temperature surveys have been useful in geothermal investigations. In Dixie Valley, where there is abundant surface evidence of the hydrothermal discharge system, the survey was perhaps not as useful as it would have been in an area without such surface manifestations. With few exceptions, hot spots delineated by the survey corresponded to areas with surface evidence of thermal activity (hot springs and fumaroles). It was therefore difficult to delineate the near-surface hydrothermal discharge system solely on the basis of the shallow temperature survey.

The survey results did reveal that the 1 meter temperatures generally tend to increase toward the Stillwater Range. It would probably be worthwhile to extend the survey into these mountains, although this could be a difficult task. Similarly, it might prove useful to extend the single line of holes running approximately east-west across the valley farther into the Clan Alpine Mountains. Despite the generally warmer temperatures closer to the Stillwaters, temperatures on those portions of the alluvial fans closest to the canyon outlets generally had cooler temperatures. These depressed temperatures were the result of streamflow infiltration and/or groundwater flow. In addition, evidence for quenching at depth was found in thermal gradient holes H-2 and S-8 by comparing their gradients to the nearest 1 meter hole.

Some success was obtained correlating the shallow temperature distribution to faults in the valley. In a few isolated instances the location of faults could be confirmed, but success was not widespread.

Comparison of the six monthly surveys indicated that with the exception of the April survey, the same trends appeared each month. This implies that single monthly survey data are sufficient, and that six consecutive monthly surveys are unnecessary. A better approach than the one used herein would be to take one survey in the summer (perhaps July) and one in the winter (perhaps January) in order to

account for differing soil thermal conductivities (Olmsted, 1977). In fact, if only a single survey is to be done, it is probably best to conduct it during a winter month. A winter-month survey should show the maximum temperature contrast at a depth of 1 meter. The reason for this is the depletion of stored heat in the near surface which would otherwise equalize temperatures at a shallow depth (Miller, 1978).

Although no measurements were made of the soil thermal diffusivities at the 1 meter locations, Miller (1978) states that for most types of soil, temperatures at a depth of 1 meter scarcely react to diurnal temperature fluctuations. However, the soil temperature per se was not being measured, but the temperature of a water-antifreeze mixture at the bottom of a PVC pipe. The one meter depth should provide sufficient protection from the diurnal changes, although we did not substantiate this belief. Two meter or 1.5 meter holes would probably have been better. No corrections were made for the annual temperature wave, which penetrated to depths much greater than 1 meter.

In conclusion, the shallow temperature survey in Dixie Valley did yield useful information, although not as much as would have been obtained in an area without surface evidence of faulting and hydrothermal activity. The local hydrology, particularly the shallow groundwater flow system, undoubtedly affected the shallow temperatures to an unknown degree by masking and redistributing geothermal heat.

6.4 References

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Chapter 7. MODELS OF THE DIXIE VALLEY GEOTHERMAL SYSTEM

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7.0 MODELS OF THE DIXIE VALLEY GEOTHERMAL SYSTEM

7.1 Introduction

The purpose of the MMRI program of investigation was to develop an integrated model of the Dixie Valley area for a geothermal system that is based on both available data and new data derived from the specific investigations in four general disciplines or fields of study (structure-tectonics, petrology, hydrology-hydrogeochemistry, and shallow temperature surveys). The results of these individual investigations are presented in the preceding chapters of this report. While developing the program of study, an initial model based upon a knowledge of the general regional relationships and limited site-specific data was prepared; this initial model will be referred to as the "proposal model". During the MMRI program this proposal model was evaluated and modified as new data were generated. The second stage of developing a new model was to re-evaluate the data as a series of specific models for each of the disciplines. Finally, these models are now consolidated and integrated to form a new model, to be referred to as the "integrated model" of the Dixie Valley Geothermal System.

7.2 Proposal Model

The initial model for the proposal was based on an evaluation of limited site-specific data (Keplinger and Associates, 1978; Senturion Sciences, 1977, 1978a, b; Thompson and others, 1967; Micro Geophysics, 1976; Exploration Data Consultants, 1976). Figure 7-1 illustrates, in map view, the major elements of the model, including faults and areas of aeromagnetic and gravity anomalies. Figures 7-2 and 7-3 present generalized east-west cross-sections of the model indicating the inter-relationships among the various tectonic elements, inferred heat sources, and distribution of the Humboldt Lopolith as a capping mechanism for the reservoir.

There are three major problems with this initial proposal model. First, the Stillwater fault is shown as a high-angle reverse fault. This is inconsistent with the extensional tectonic regime of the Basin and Range Province and with both the field expression of the fault and the focal plane solutions of the 1954 Dixie Valley earthquake (Ryall and Malone, 1971). Second, a thrust fault, the Stillwater thrust, is shown in

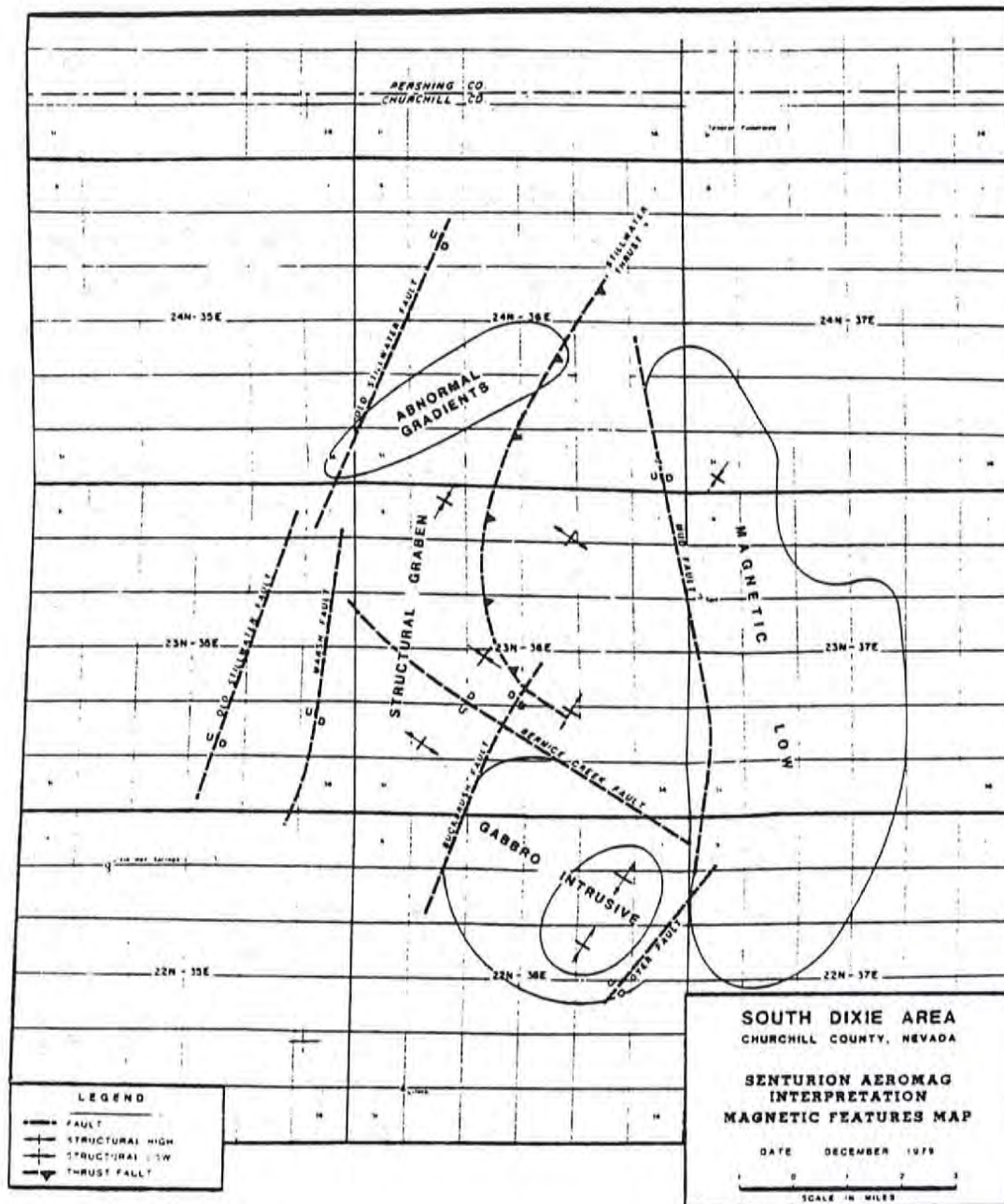


Figure 7-1. Map view of proposal model of the Dixie Valley Geothermal System. Faults and anomalous regions are based on an interpretation of aeromagnetic data by Senturion Sciences (1977, 1978).

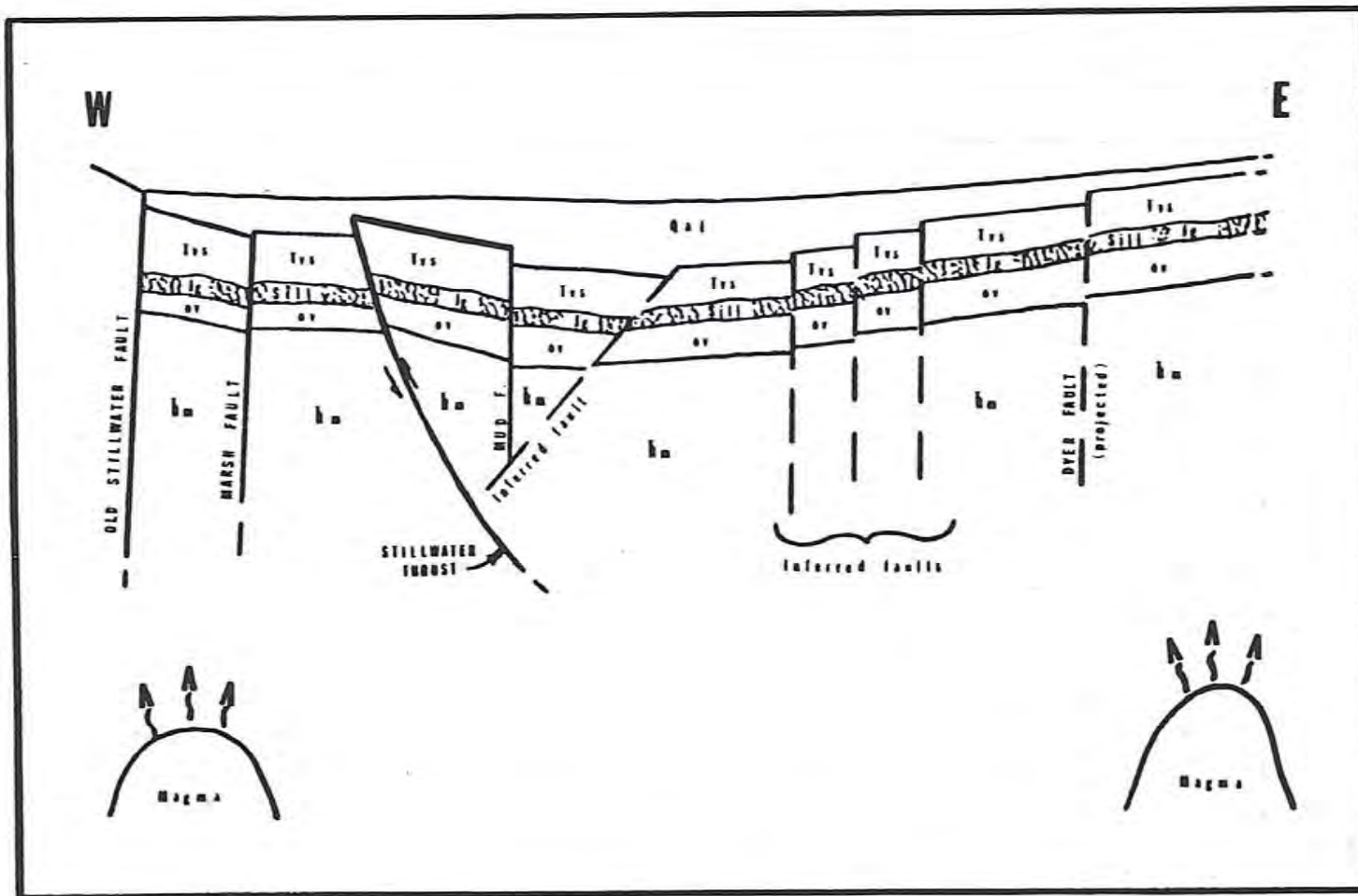


Figure 7-2. Generalized east-west cross-section of proposal model showing inter-relationships of the elements of the model. Qal - Quaternary alluvium; Tvs - Tertiary volcanics and sediments; Jg - Jurassic Humboldt Lopolith gabbro; ov - older volcanics; tm - Triassic metasediments, highly fractured phyllite and slate; arrows indicate displacement vectors on faults.

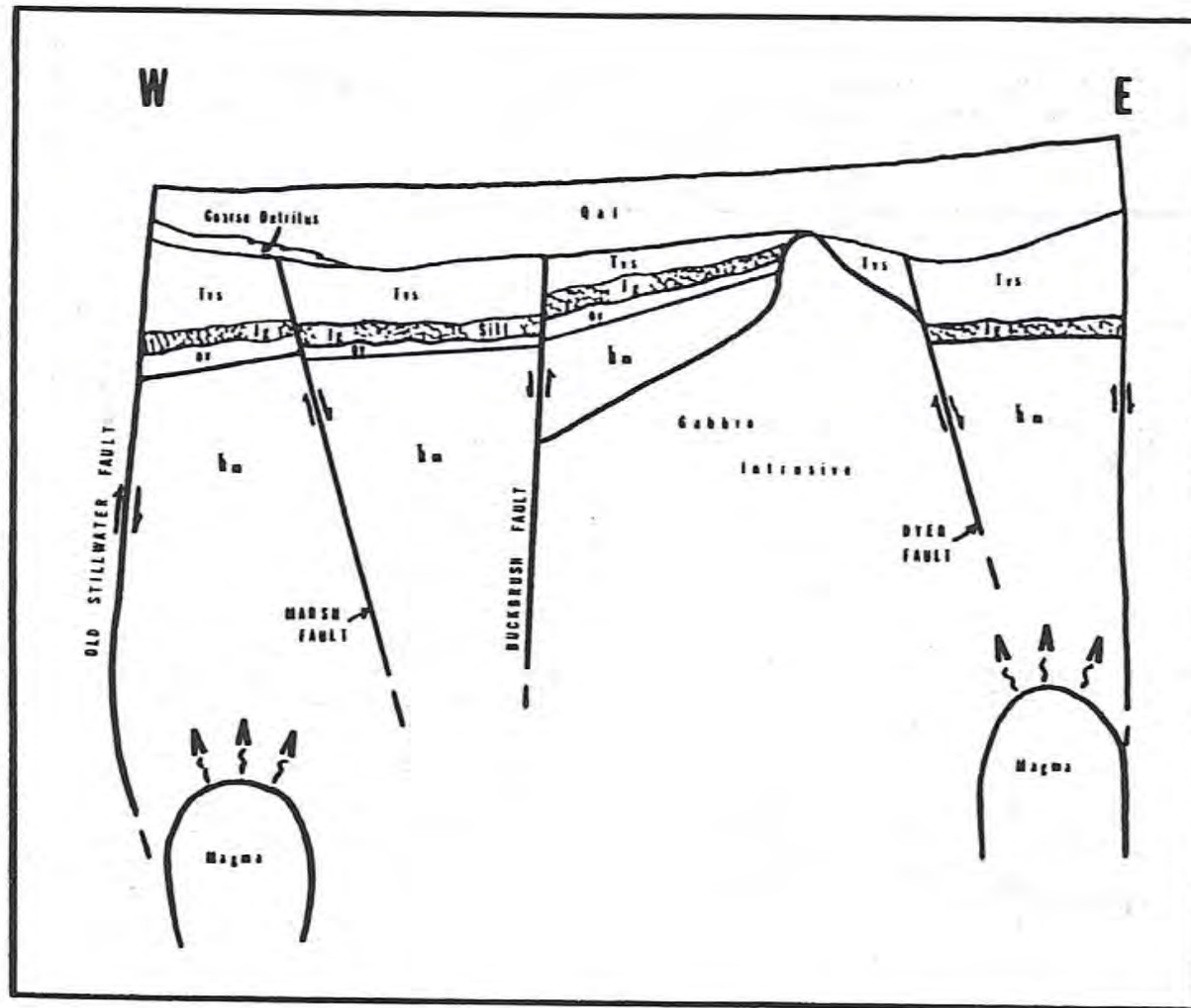


Figure 7-3. Generalized east-west cross-section of proposal model showing inter-relationships among the elements of the model. Qal - Quaternary alluvium; Tvs - Tertiary volcanics and sediments; Jg - Jurassic Humboldt Lopolith, gabbro sill; ov - older volcanics; lm - Triassic metasediments, highly fractured slate and phyllite.

the central portion of the Valley; this too is inconsistent with the extensional tectonic regime. And third, the subsurface stratigraphy is projected as consistent throughout the Valley and corresponding to the range front stratigraphy, even though the subsurface data and comparison of the lithologies exposed in the Stillwater Range and Clan Alpine Mountains are not adequate to substantiate either the stratigraphic sequence and physical nature of the units, or the geographic distribution and continuity of the units as shown. A "cap" of Humboldt Lopolith gabbro overlying very highly fractured Tertiary metasediments was assumed to be an important characteristic of the model.

Thus, this initial model provided the basis for defining the MMRI program of investigation and served as a working model that required refinement and verification to allow development of an integrated model of the Dixie Valley Geothermal System.

7.3 Integrated Model

The integration of the single-discipline models based on structural-tectonic analysis, petrologic alteration studies, hydrology-hydrogeochemistry, and shallow temperature surveys resulted in a new model of the Dixie Valley Geothermal System as presented in Figures 7-4 and 7-5.

7.3.1 Structural Setting

The structural setting of the integrated model is presented in Figure 7-4. Key elements of this setting include: 1) Basin and Range extensional faults; 2) Humboldt gabbroic complex; and 3) White Rock Canyon fault.

7.3.1.1 Basin and Range Extensional Faults

The structure of the northern portion of Dixie Valley is a complex, asymmetric graben, with portions of the inner graben complex defined by splays of the bounding faults. These structural-tectonic features are delineated by various geophysical data, as well as by their surface expression. These faults, their geographic distribution within the Valley, inter-relationships, and observable displacements are all consistent with the regional extensional tectonics of Dixie Valley and the northern Basin and Range Province. Many of these faults serve as preferential conduits for fluid migration, as evidenced by the spatial correlation of surface

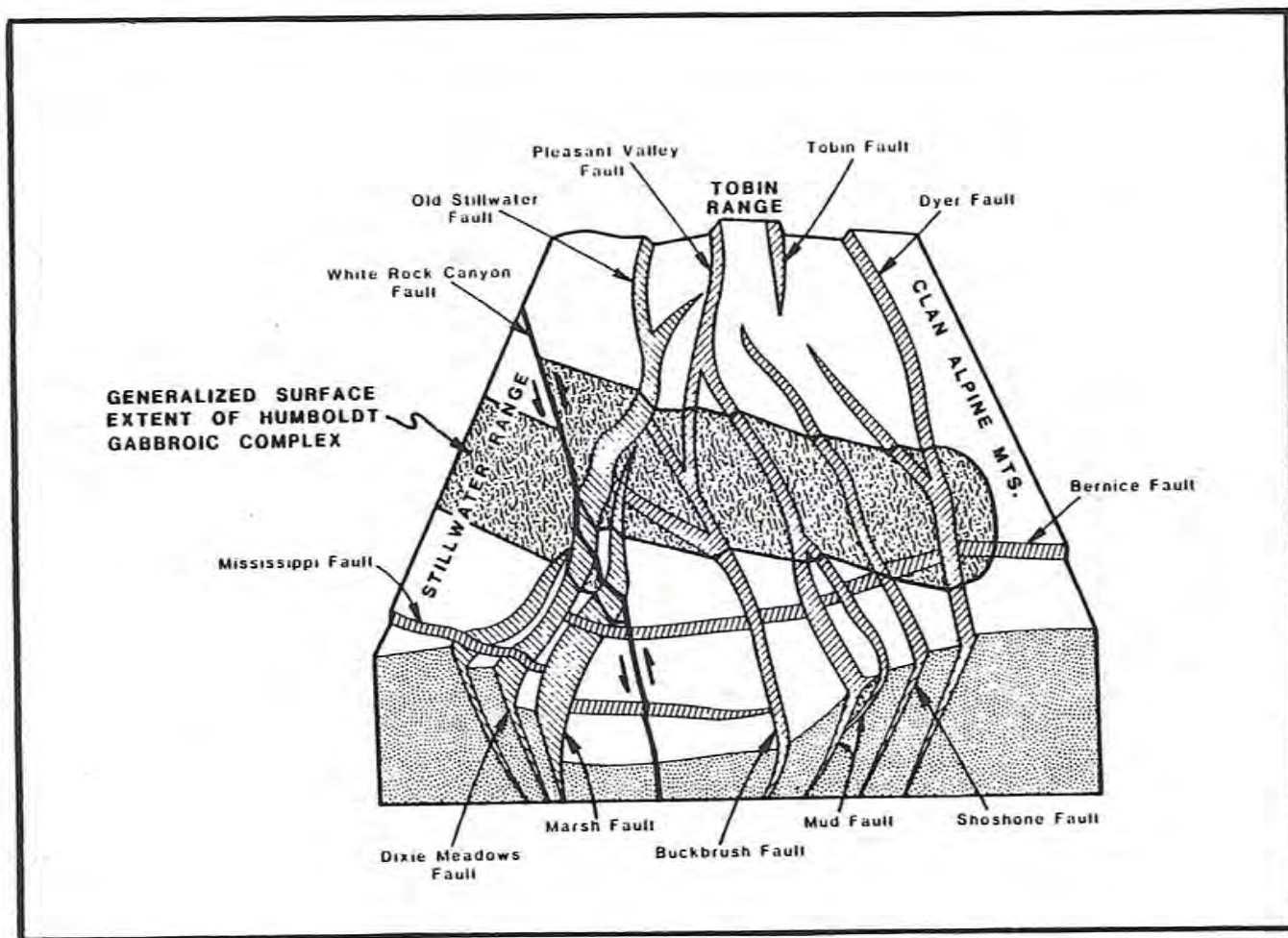


Figure 7-4. Three dimensional view of integrated model of the Dixie Valley Geothermal System. Note that the generalized surface extent of the Humboldt gabbroic complex merely defines an area within which the complex may be encountered, and in no way implies continuity of the unit within those boundaries.

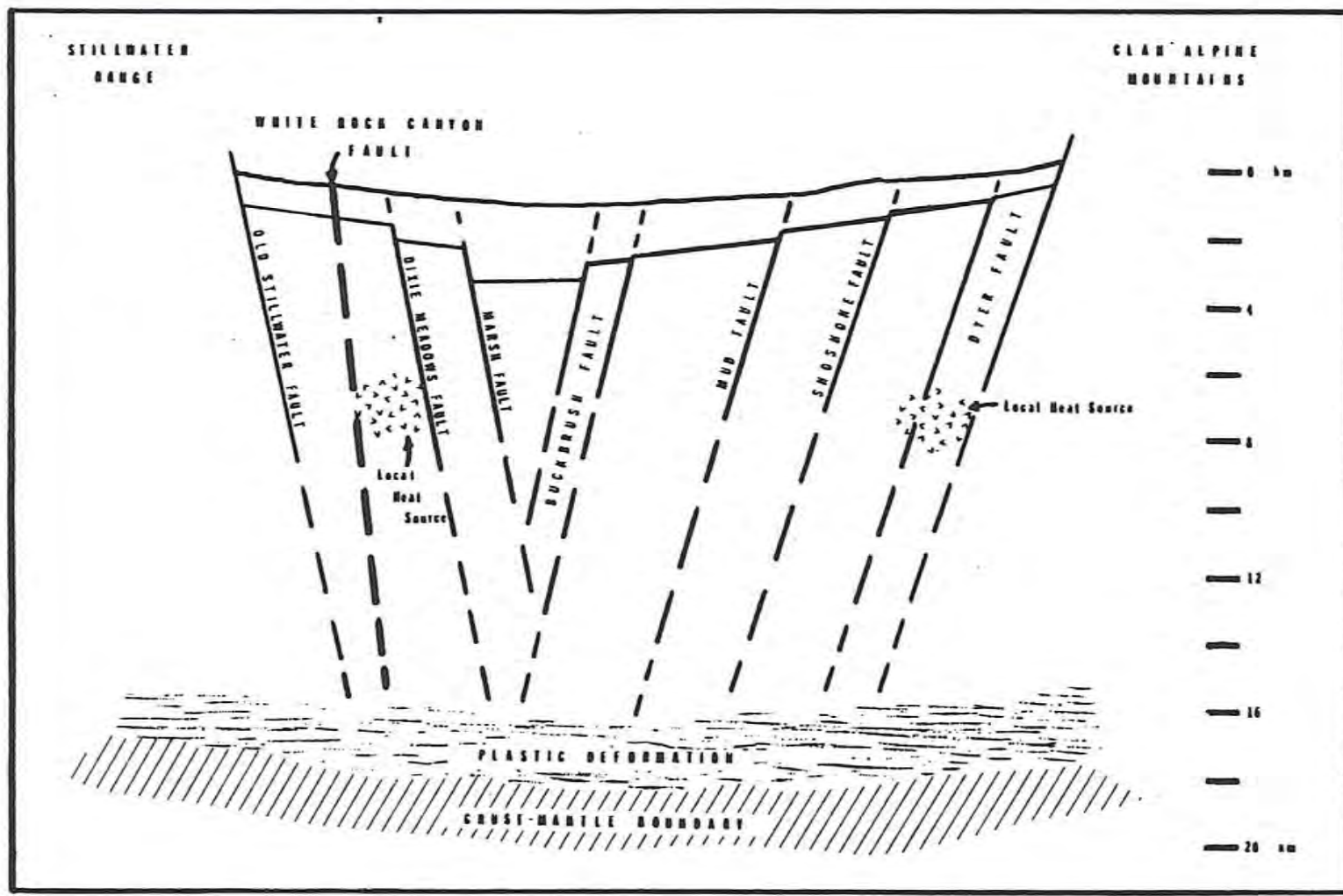


Figure 7-5. Generalized east-west cross-section of the integrated model of the Dixie Valley Geothermal System. Structural elements (faults) are propagated downward from the surface through the valley fill (alluvium with intercalated volcanics) and into the underlying undifferentiated bedrock and then projected through the zone of brittle failure.

springs, seeps and fumaroles, as well as the surface and subsurface concentration of intense hydrothermal alteration along these fracture zones, particularly along the major range front faults.

7.3.1.2 Humboldt Gabbroic Complex

The geographic distribution of surficial outcrops of the Humboldt gabbroic complex in the West Humboldt Range, Buena Vista Hills, Stillwater Range and Clan Alpine Mountains is accepted from Speed (1976). However, it appears that the Humboldt complex is not a continuous unit either on the surface or within the subsurface.

Two intrusive 'phases' of the complex were identified in the subsurface. The first, diorite/gabbro, is medium to coarse-grained primarily plagioclase and pyroxene, with secondary minerals including hornblende, chlorite, magnetite, epidote, albite, calcite, pyrite and titanium minerals (ilmenite and leucoxene). The second, granodiorite, has biotite and hornblende as the principal ferromagnesian minerals, with plagioclase and K-feldspar. These two phases have been reported in the Sun Oil Company wells SW Lamb #1 and #3, and SW Lamb #2, respectively. Only the diorite/gabbro phase was present in DF 45-14 as thin (average 30 foot thick) dikes intruding the Triassic metasediments; both phases were identified in DF 66-21 with the granodiorite phase much more dominant. The diorite/gabbro and the granodiorite phases of the Humboldt gabbroic complex exhibit evidence of deuteric alteration; however, obvious hydrothermal alteration is most intense and concentrated along fracture zones.

The intrusive phases of the Humboldt gabbroic complex are subjacent to an extrusive phase of a volcanic sequence and associated intercalated red clay ((?) hematitic mudstone of Speed, 1976). This relationship was identified in DF 66-21 and reported for the Sun Oil Company wells SW Lamb #1 and #3.

The subsurface stratigraphy encountered by the various deep exploratory wells, the surface outcrop pattern and a re-evaluation of the geophysical data (particularly the aeromagnetic data) suggest that the Humboldt gabbroic complex is not a continuous lopolithic unit, but rather is a discontinuous intrusive-extrusive complex of dikes, sills, plugs, and volcanics. Furthermore, distribution of the complex is controlled by faults associated with post-emplacement graben formation within the extensional Basin and Range tectonic regime and also with major basement

faults, including the White Rock Canyon fault.

The concentration of intense hydrothermal alteration along fracture or fault zones in the Humboldt gabbroic complex is also characteristic of other lithologic units encountered in the surface, such as the Triassic metasediments and younger volcanics and alluvium. Where units are not fractured, the alteration is much less intense and the water chemistry suggests that fluids within the non-fractured rock are essentially stagnant or have very long residence times.

7.3.1.3 White Rock Canyon Fault

The north-south trending White Rock Canyon fault exhibits left-lateral displacement suggested by geophysical data and by surficial expression of the fault. The 10 to 12 miles of apparent left-lateral offset of the Humboldt gabbroic complex is post-middle Jurassic. The location and orientation of the fault zone appears to have been determined by pre-middle Jurassic structural control. Major left-lateral displacement on this fault zone may be related either to development of the mid-Miocene Cortez Rift, or to a conjugate structure of the apparent post-mid-Miocene right-lateral displacement of the Cortez Rift across the high lava plains at the northern boundary of the Great Basin (Figure 2-15). These relationships may also be due to compressional tectonics as the Farallon Plate was being subducted along the west coast margin. Surficial expression as vegetation and tonal contrasts and lineations within the young alluvial sediments of Dixie Valley indicates Quaternary fault activity, including Holocene displacement along this zone.

Regardless of the time of initiation and major deformation on the White Rock Canyon fault, it appears to be a significant regional structural element critical to the model of the Dixie Valley Geothermal System. The White Rock Canyon fault, as a major basement fault, may act as a western boundary for a region of fluid (i.e., heat) migration from depth to the surface. The overall geochemical difference between Dixie Meadows hot springs to the south and Hyder and Sou Hot Springs to the north implies that a major structural barrier may exist northeast of Dixie Meadows -- i.e., the White Rock Canyon fault. Major tectonic activity, including the 1954 Dixie Valley earthquake, along the Old Stillwater fault south of the intersection with the White Rock Canyon fault may have effectively terminated direct communication of the Dixie Meadows hot springs

with a deep heat source. It should be noted that there are distinct geochemical differences between the Hyder-Buckbrush system (linked by the Buckbrush fault) and Sou Hot Springs, both east of the White Rock Canyon fault, implying a lack of general communication among fluids within the Dixie Valley Geothermal System.

7.3.2 Structural Relationships

The structural relationships of the integrated model are depicted in a generalized east-west cross-section in Figure 7-5. The structural-tectonic features (faults) delineated by surficial expression and by geophysical surveys are propagated downward from the surface through a zone of brittle failure to approximately 12 to 15 km depth. From 16 to 18 km depth, is a zone where plastic deformation dominates. The crust-mantle boundary in the Dixie Valley region is between 18 and 20 km depth with the underlying mantle acting as a regional heat source. The White Rock Canyon fault is shown as a major western boundary for a region of upward migration of thermal fluids.

Localized heat sources are depicted within the crust at depths of 6 to 8 km. Based upon magnetotelluric data (Keplinger and Associates, 1978), shallow heat sources were defined along the eastern edge of the Stillwater Range in the vicinity of the Dixie Comstock mine and northward. Similar shallow crustal heat sources are depicted on the diagram; these may represent localized intrusions of magma, with locations and extent that are structurally controlled. For the Dixie Comstock area, the Old Stillwater and Marsh faults, and possibly the Bernice Creek fault, are inferred to be the conduits. Based upon an interpretation of the distribution of the shallow temperature survey data in which anomalies on the west side of the Valley correlate with the locations of these shallow crustal heat sources, a localized heat source is also inferred to exist under the east side of Dixie Valley associated with the Dyer and Bernice Creek faults.

7.3.3 Summary

The Dixie Valley Geothermal System thus appears to be a result of regionally high mantle heat flow through a thin crust, with preferential conduits for fluid migration structurally controlled by the White Rock

Canyon and other major basement faults, and by Basin and Range extensional faults within the Valley. The geothermal system apparently does not have a specific cap rock. Rather, it appears that several lithologies within the Valley such as the Humboldt gabbroic complex, the 'red clay', or the Triassic metasediments are impervious to general fluid flow and production zones may be found in association with any of these units if there is communication by fracture or fault zones with the reservoir at depth and sufficient fluid to transport the heat.

7.4 References

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Chapter 8. EVALUATION OF THE INTEGRATED MODEL OF THE
DIXIE VALLEY GEOTHERMAL SYSTEM

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8.0 EVALUATION OF THE INTEGRATED MODEL OF THE DIXIE VALLEY GEOTHERMAL SYSTEM

8.1 Introduction

The integrated model of the Dixie Valley Geothermal System should be considered as a second phase of model development for the Dixie Valley region. Therefore, the following recommendations are presented as the types of additional investigations considered necessary to refine and verify the integrated model. These recommendations are focussed on obtaining additional data from previous sources (e.g., the deep exploratory wells) and on developing new sources of data, particularly for the subsurface. These recommendations indicate where insufficient data are available to completely verify inter-relationships of elements of the model or to define the specific role that an element has in the model, or where additional data could define additional elements critical to understanding the Dixie Valley Geothermal System.

8.2 Recommendations

The following recommendations were generated by the present MMRI program of investigation:

1. Theoretical analysis of the thermal gradient data from existing shallow gradient holes, existing deep exploratory wells, and any additional wells that may be drilled in Dixie Valley. In theory, it is often possible to obtain transmissive, mixing and other properties of the porous medium from these data. It is sometimes possible to distinguish (separate) conductive and convective heat transfer. Analysis of this type is not routine, i.e., a fair amount of theoretical work would be required. However, the methods developed would be applicable to evaluating any hydrothermal system.
2. The hydrogeochemical and isotopic investigations should be expanded to include all of Dixie Valley as well as surrounding valleys. Extensive work of this nature should also be conducted in the Stillwater Range and Clan Alpine Mountains. Additional isotopes should be used; these might include ^{14}C , ^{13}C , ^{34}S and ^3He , among others. The purpose of this investigation would be to substantiate the regional nature of the Dixie Valley hydrologic system(s), to better delineate the heat

sources, and to understand the origin and circulation patterns of both thermal and non-thermal waters.

3. Fluid samples should be collected with time from both DF 45-14 and DF 66-21. Both chemical and isotopic analyses should be performed on these samples. Useful information would also be obtained by conducting drill-stem and/or other tests on these wells or on any future wells drilled in the Valley. Drill-stem testing would yield pressure and permeability data, as well as fluid samples from depth. Regardless of the method used, obtaining fluid samples at depth (in situ and uncontaminated) would provide important information. Time series samples should also be obtained from selected springs; vapor samples should also be collected. Stable isotope geothermometry should be utilized to provide additional information on reservoir temperatures.
4. The construction of a quantitative (numerical or mathematical) model should be investigated. Data obtained from the testing of DF 45-14 and DF 66-21 would be invaluable in formulating a model of this type -- it should be noted that the testing of these wells would provide a wealth of information even if the model were not constructed. The proposed model would be more interpretive than predictive; its prime purpose would be to guide exploration and exploitation by enhancing our comprehension of the physico-chemical processes occurring in the reservoir.
5. The shallow temperature survey should be expanded into selected areas, mainly into the Stillwater Range and the Clan Alpine Mountains to verify the presence of high heat flow. A logical place for this extension would be along the cross-valley profile line. Other locations in the mountains and the Valley itself should be surveyed. Two-meter holes should be installed.
6. Efforts should be made to obtain data from between the shallow groundwater zone (i.e., a few hundred feet deep) and the geothermal reservoir zone. Both pressure and fluid samples should be obtained from this intermediate zone, either by using the existing exploratory

wells or by drilling holes specifically for this purpose. In any case, efforts should be made during future drilling programs to obtain reliable hydrologic, isotopic and hydrogeochemical data from the wells. Collection of these data will add to the expense of drilling, but will provide invaluable data. It might be useful to conduct a drilling program especially for this purpose. These holes would be on the order of a few thousand feet deep, and would be located in areas designed to maximize the amount and types of information obtained, as well as the geographic distribution of such data.

7. Shallow gradient holes should be drilled both at selected hot spring areas and at selected intervals along the cross-valley line. The piezometric surface and water chemistry should be evaluated with respect to depth, and lateral variation in the piezometric surface should be noted. This data along with aquifer properties, such as permeability, will allow calculation of fluid flow rates under present conditions and will be important in understanding how the geothermal reservoir may react to development.
8. Petrographic geothermometers based on the concentrations of trace elements in single minerals or mineral species may indicate past history of the reservoir or present reservoir temperatures. This type of geothermometer has an established theoretical and practical base; however, it has not been utilized to any extent in geothermal investigations but could prove useful. As various minerals are heated, larger concentrations of trace elements are able to be incorporated into the crystal structure. With carefully selected sampling, it is possible to determine which areas have undergone the highest temperatures in the past. This information does not duplicate any presently collected data and could prove to be a very useful technique for exploration and reservoir analysis.
9. A detailed bedrock geology study of the Stillwater Range and the Clan Alpine Mountains should be conducted to determine specific stratigraphic and structural relationships. This would delineate the limits of exposed key stratigraphic units in the mountains and would provide additional frame of reference for evaluating the stratigraphy en-

countered in the existing deep exploratory wells and future wells drilled in the Valley. In addition, critical units should be age-dated for correlation with the subsurface stratigraphy.

10. Core samples should be taken from critical units encountered in any future wells drilled in the Valley and side-core samples should be obtained from the existing deep exploratory wells. These samples should be age-dated to provide a basis for correlations between wells and with the regional stratigraphic sequence. Thin-sections of these cores would provide uncontaminated whole rock samples for petrologic and petrographic analysis of the hydrothermal alteration effects, as well as allowing for identification of the total mineral assemblage present at a specific depth. This data when correlated with the water chemistry for that depth would allow an evaluation of the equilibrium status of the system. These core samples could also be used for the petrographic geothermometry.
11. A regional analysis of the White Rock Canyon fault zone should be conducted. If the White Rock Canyon fault is indeed a significant regional structure that may have communication with a deep heat source or control the migration of heat from that source, then a regional analysis of the White Rock Canyon fault, beyond Dixie Valley, is necessary and could provide an understanding of the geothermal resources of western Nevada. Regionally, the fault is defined by a strong topographic lineament, by contrasts in lithologies, by the distribution of known surface manifestations of geothermal systems, and by a close spatial correlation with the Ventura-Winnemucca seismic zone. Multi-level aeromagnetic and gravity profiling along the White Rock Canyon fault zone would aid in delineating subsurface evidence of associated geothermal activity. Structural-tectonic analysis of the zone based on low-sun-angle aerial photographic analysis would delineate the fault zone and would provide a basis for generalized surficial mapping and identification of specific surface manifestations of any underlying geothermal systems along the zone.

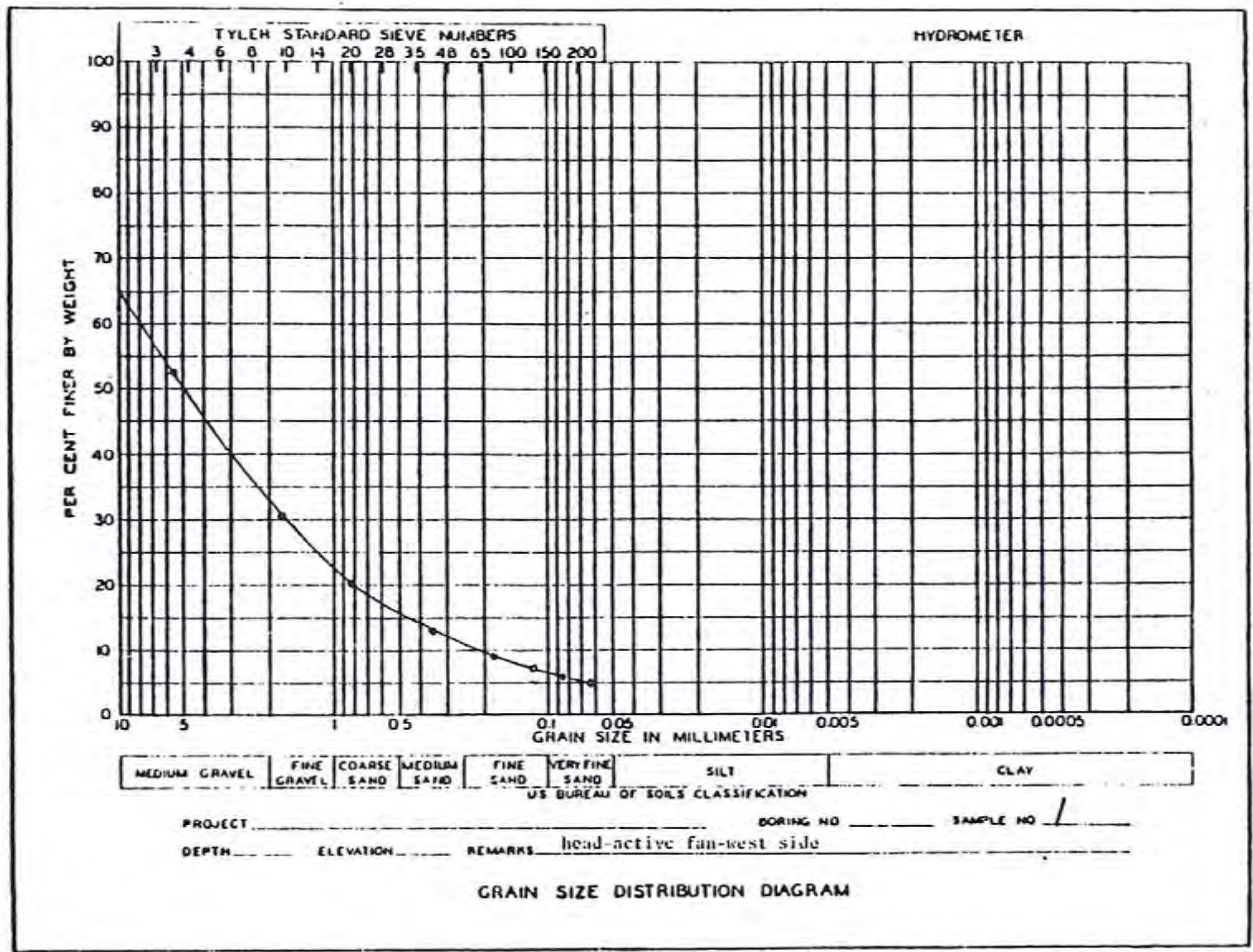
12. Detailed tectonic analysis of selected structural-tectonic features within Dixie Valley would provide a basis for understanding the mechanics of faulting within the Valley as it relates to the regional tectonic regime and provides a structural framework for the geothermal system. Such analysis might include fault scarp profiling, trenching, age-dating of appropriate materials, detailed surficial geologic mapping and a re-evaluation of available low-sun-angle aerial photography.
13. A minimum of three cross-valley reflection seismic profiles should be conducted. The interpretation of these profiles would be invaluable in verifying the subsurface structure of the Valley and the general geographic and stratigraphic relationships and distribution of the critical lithologies. One profile should transect the northern end of the Valley between Hyder Hot Springs and Sou Hills; a second profile should transect the Valley intersecting DF 66-21 and extend across the Shoshone Creek area; and a third profile should intersect DF 45-14 and extend across the Valley to the vicinity of Hoyt Canyon. These profiles should extend as far as possible into the mountains on both sides of Dixie Valley.
14. The multi-level aeromagnetic survey of Dixie Valley should be extended to include a larger area to the north and east. This could define anomalies of significant geothermal potential for exploration and development.
15. A detailed gravity survey should be conducted within Dixie Valley. This technique has proven useful in other regions where the local geologic and tectonic parameters control 'hot spots' that may be associated with either gravity highs or gravity lows depending on the specific setting. The anomalies could define areas of significant geothermal potential for exploration and development.
16. Additional deep exploratory wells should be drilled in Dixie Valley. A well should be drilled on the west side of the White Rock Canyon fault in the vicinity of Dixie Meadows to determine the extent to

which this structure acts as a barrier to migrating fluids or heat, and to identify the nature of the possible shallow crustal heat source and the conduits associated with Dixie Meadows. A well should be drilled in the center of the Valley, preferably along a cross-valley seismic reflection profile, to provide verification of the subsurface structural and stratigraphic configuration. A well should be drilled to the north of DF 66-21 and toward the center of the Valley to determine the subsurface configuration and extent of the reservoir encountered by DF 66-21 and by the Sun Oil Company wells. Ideally, during drilling of these and any other wells both uncontaminated water and core samples will be collected and the appropriate well testing will be conducted to obtain the maximum amount of data pertaining to the hydrologic system(s) encountered, such as flow rates, permeability, and porosity. This data base could then serve as the basis for model development and verification, as well as provide a baseline for evaluating the effects of development on the Dixie Valley Geothermal System.

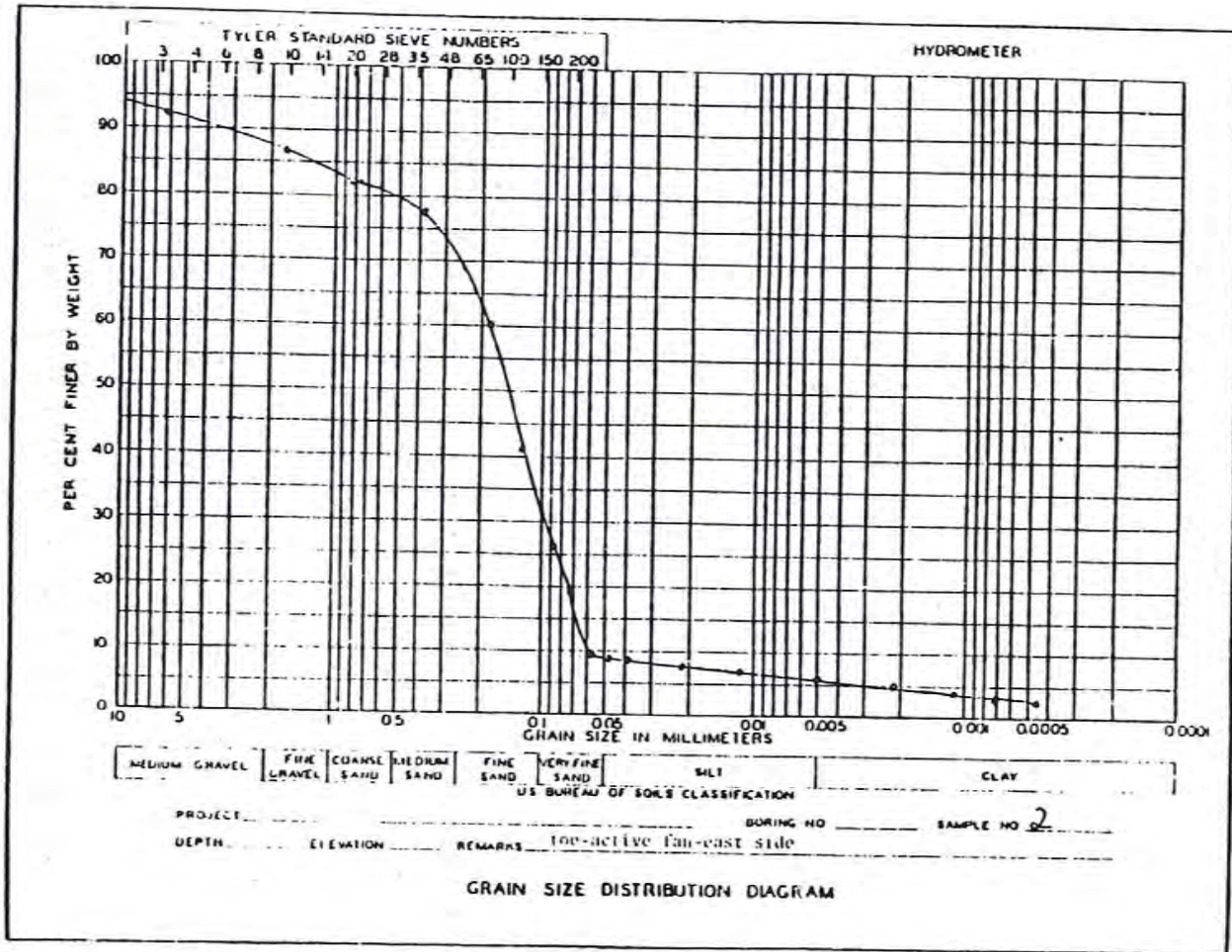
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Grain-Size Distribution Diagrams for
Selected Geomorphic Surface Samples

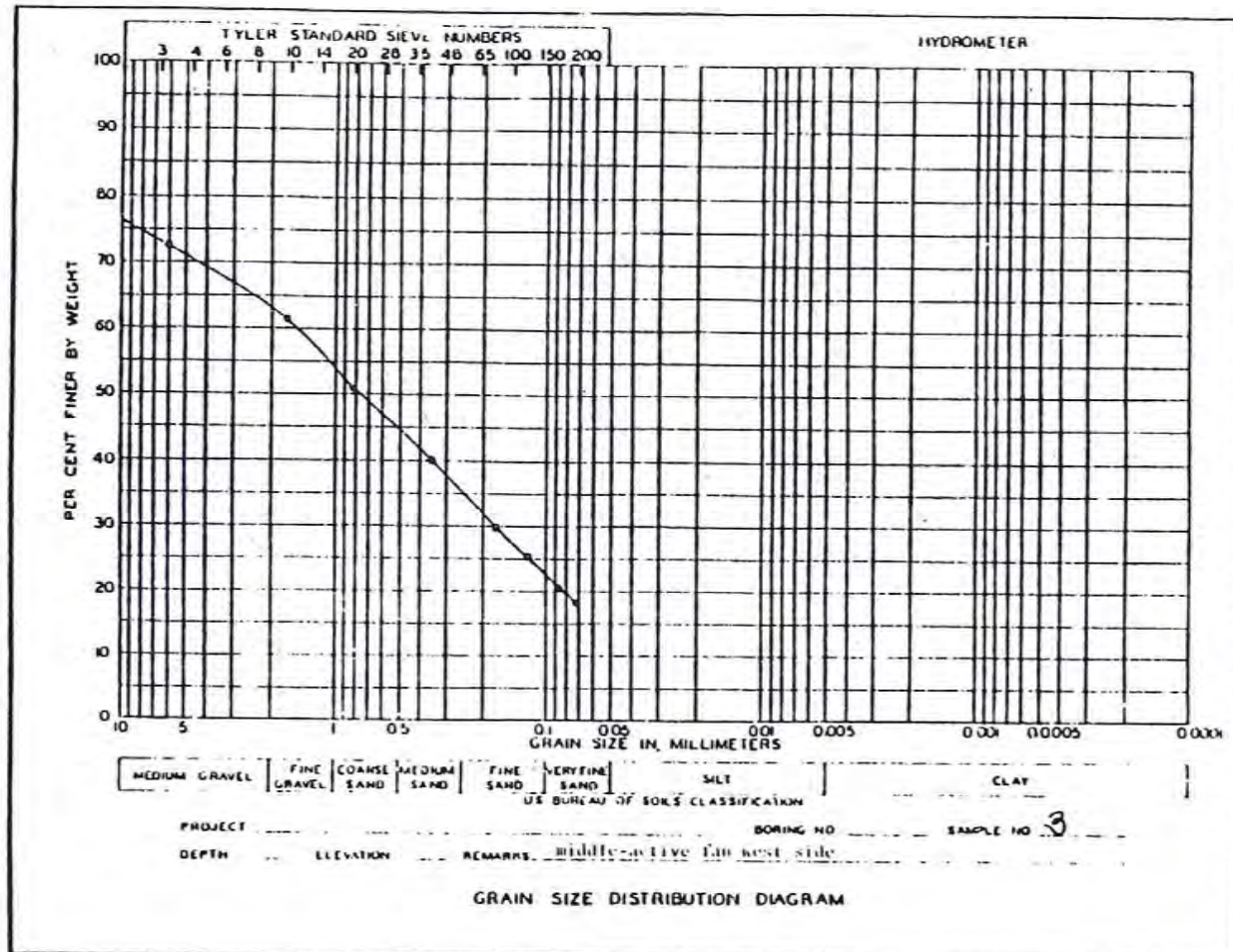
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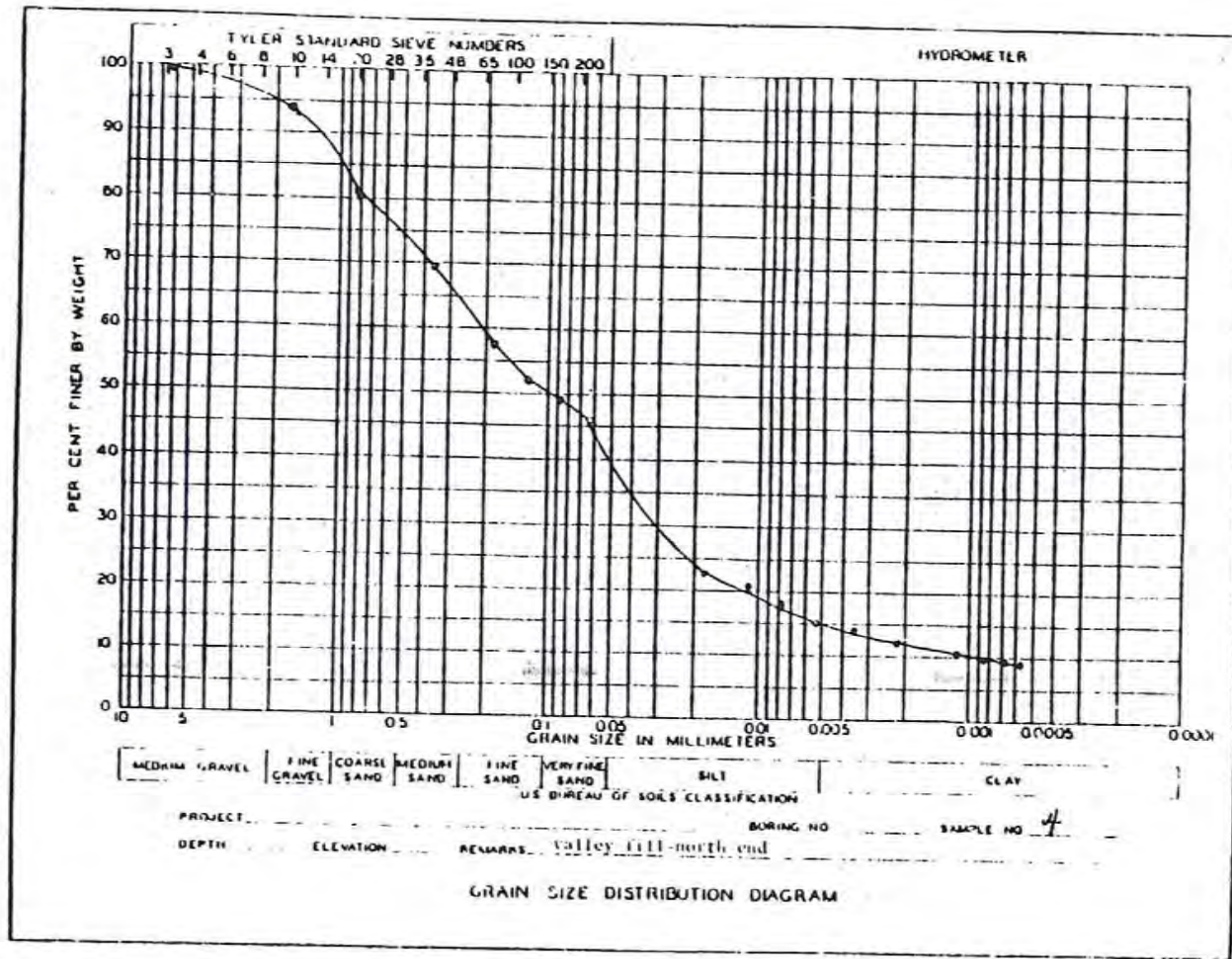
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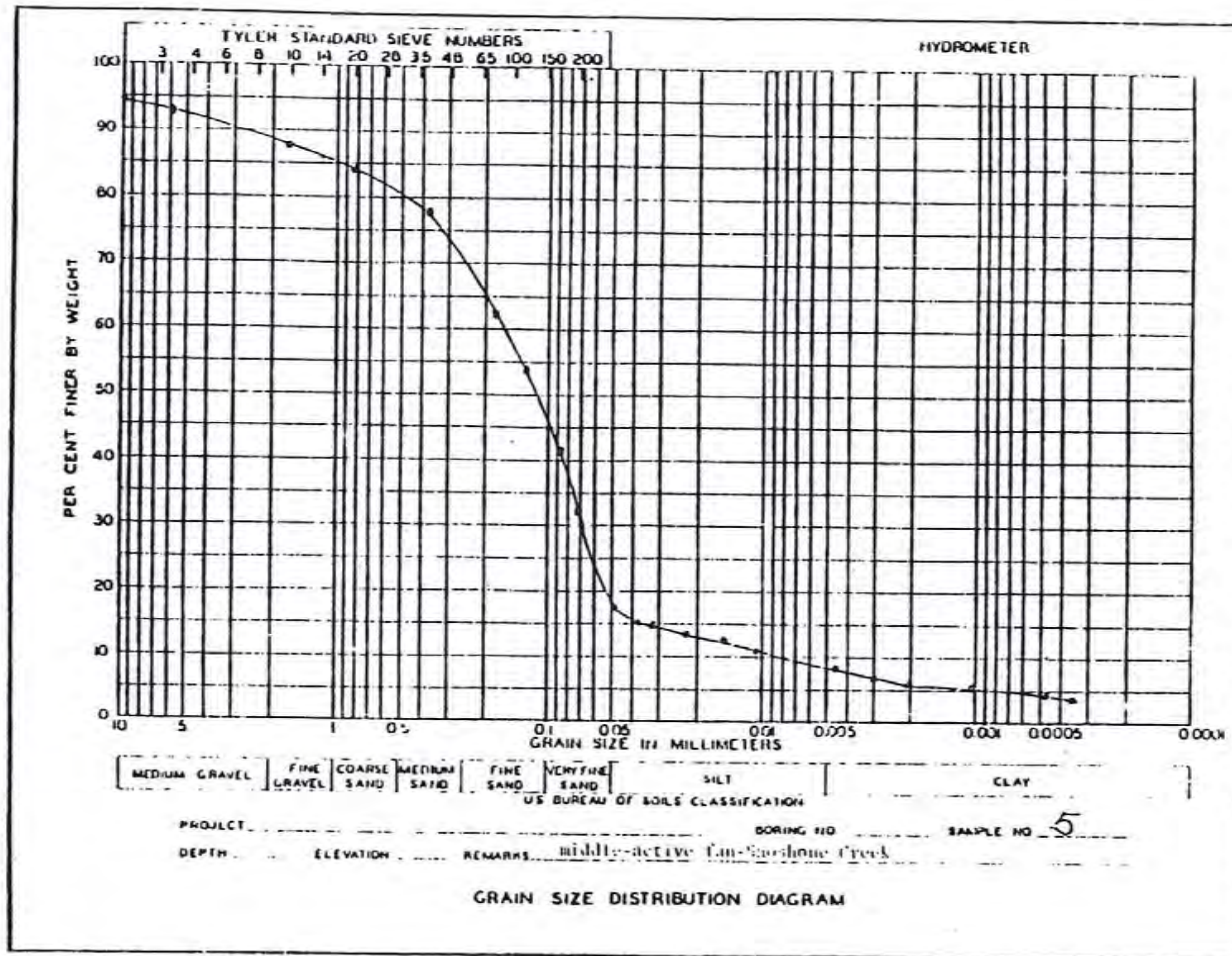
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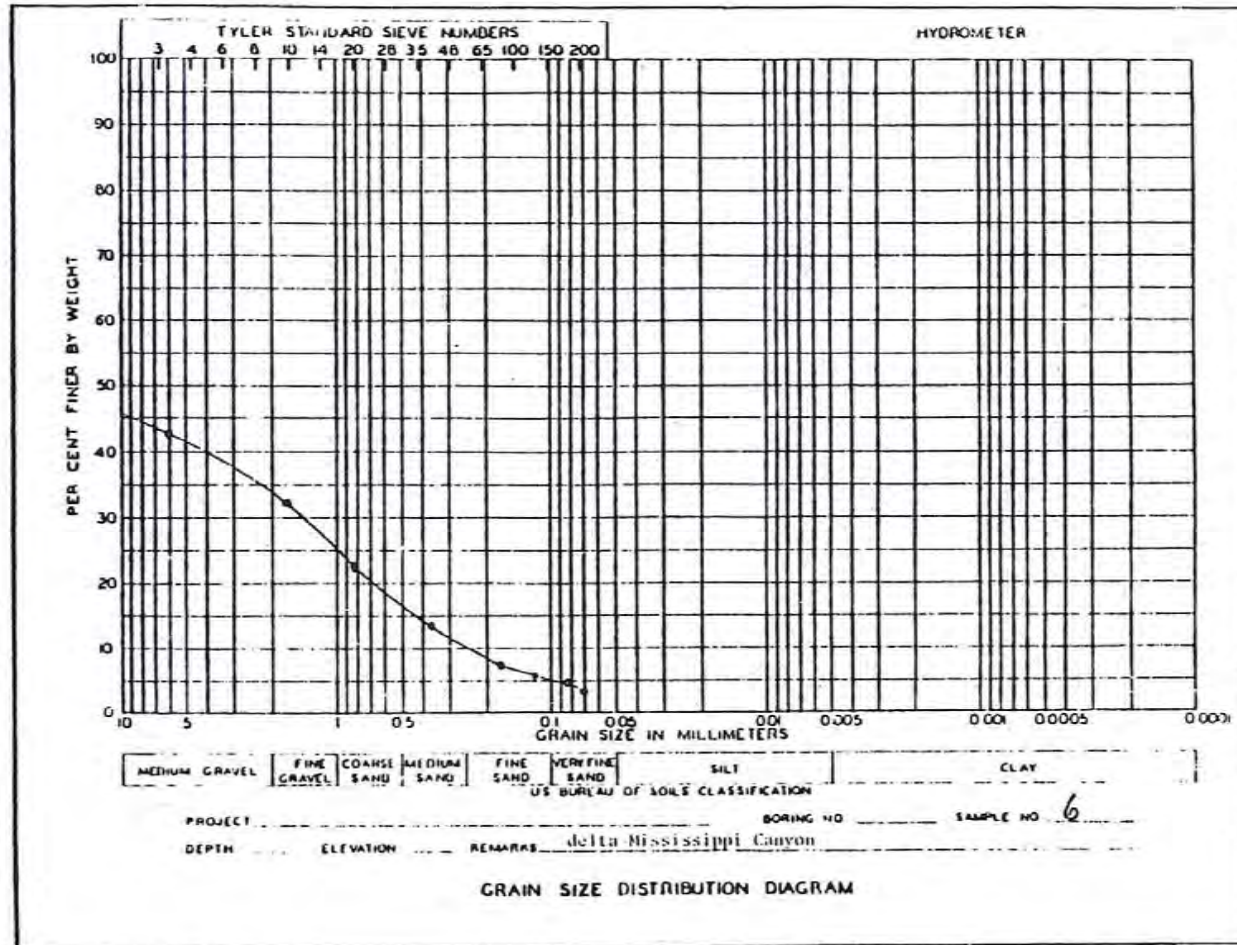
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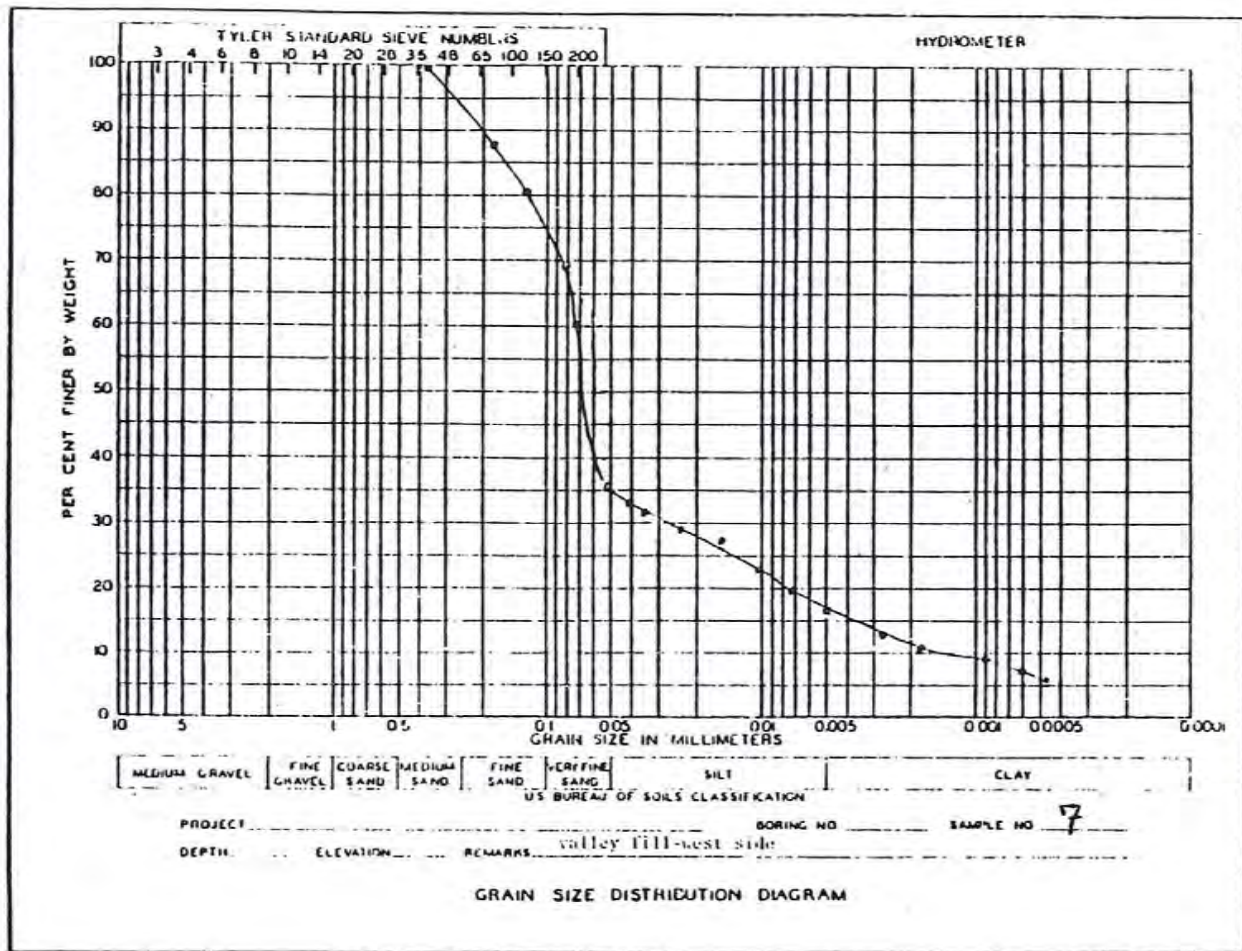
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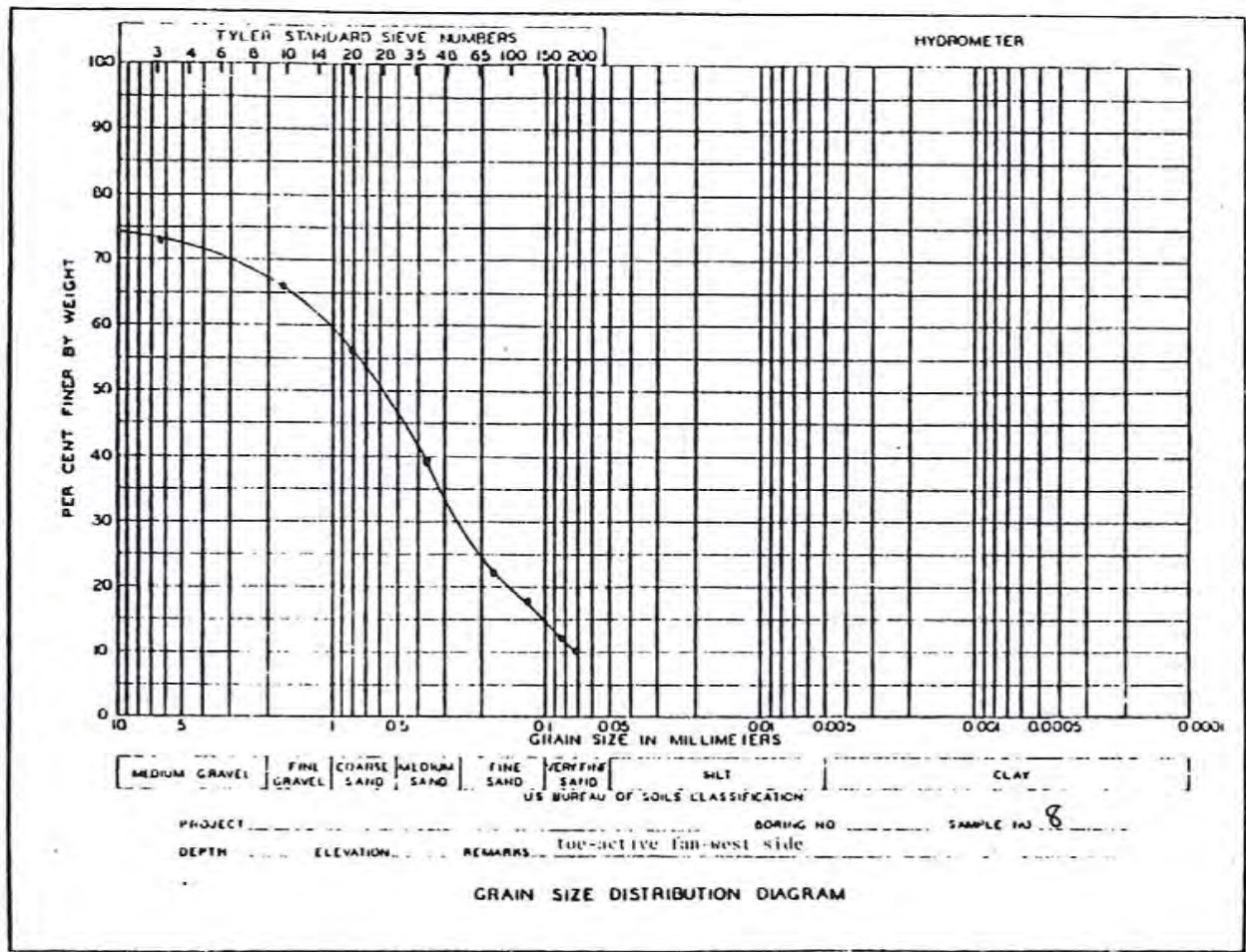
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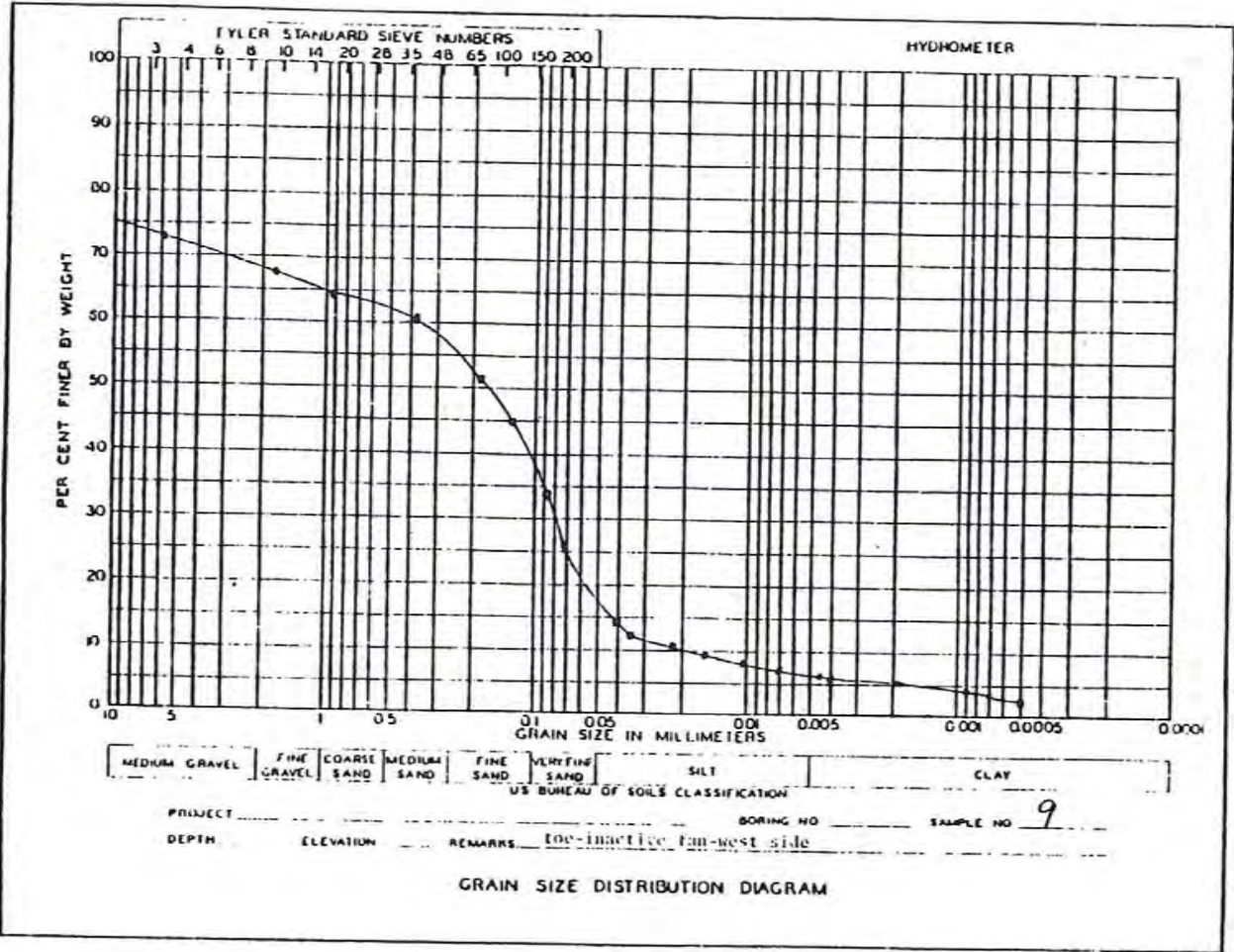


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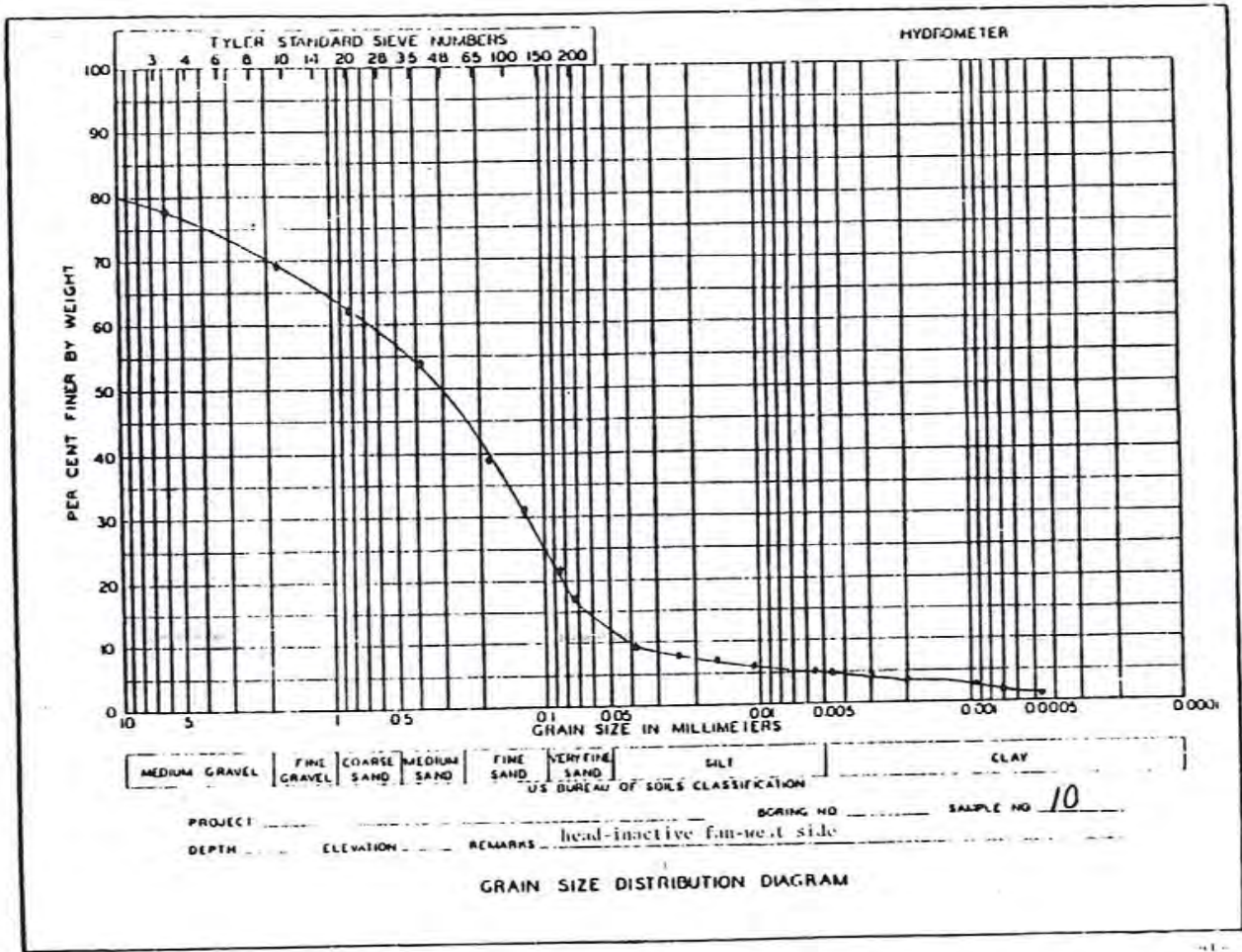


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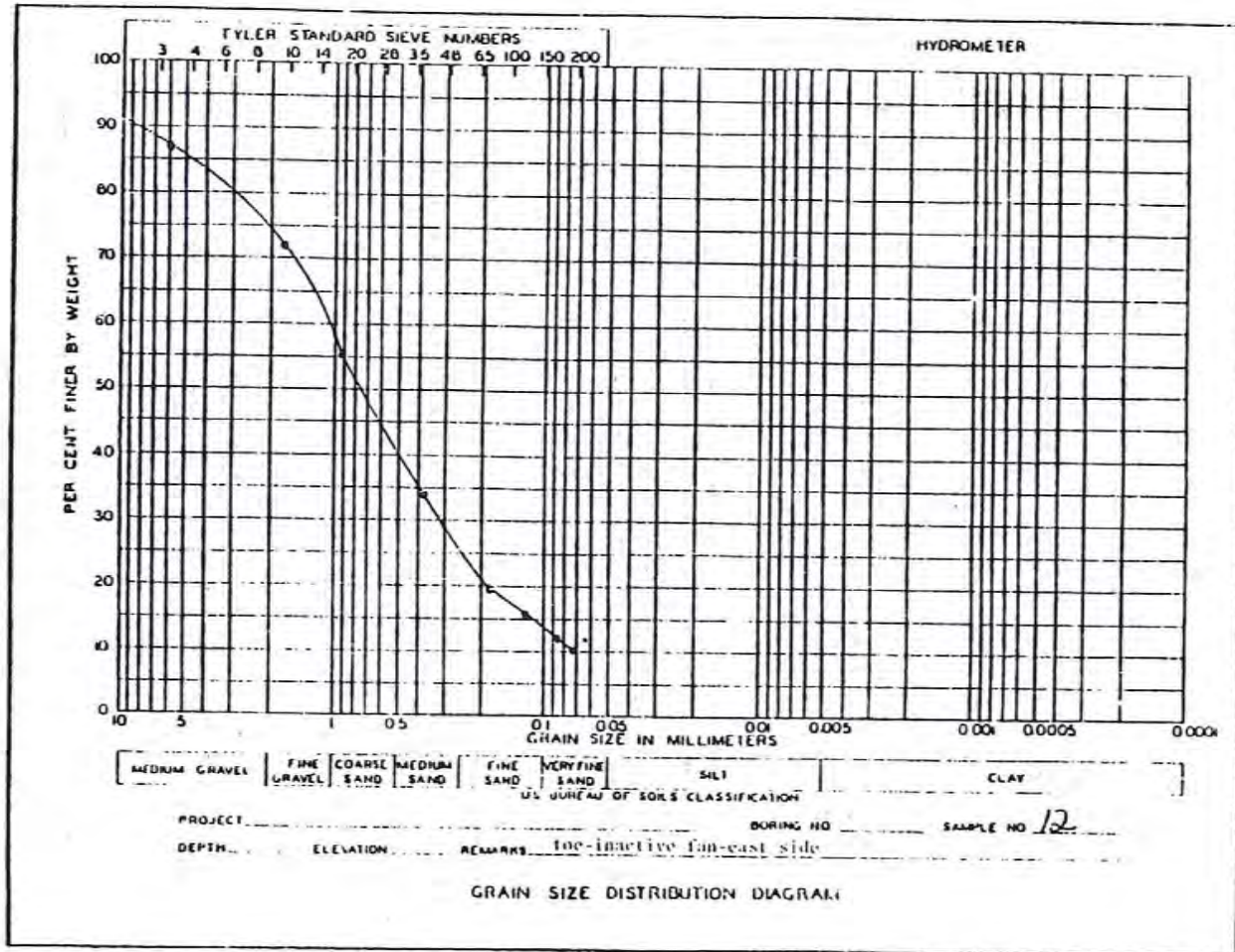




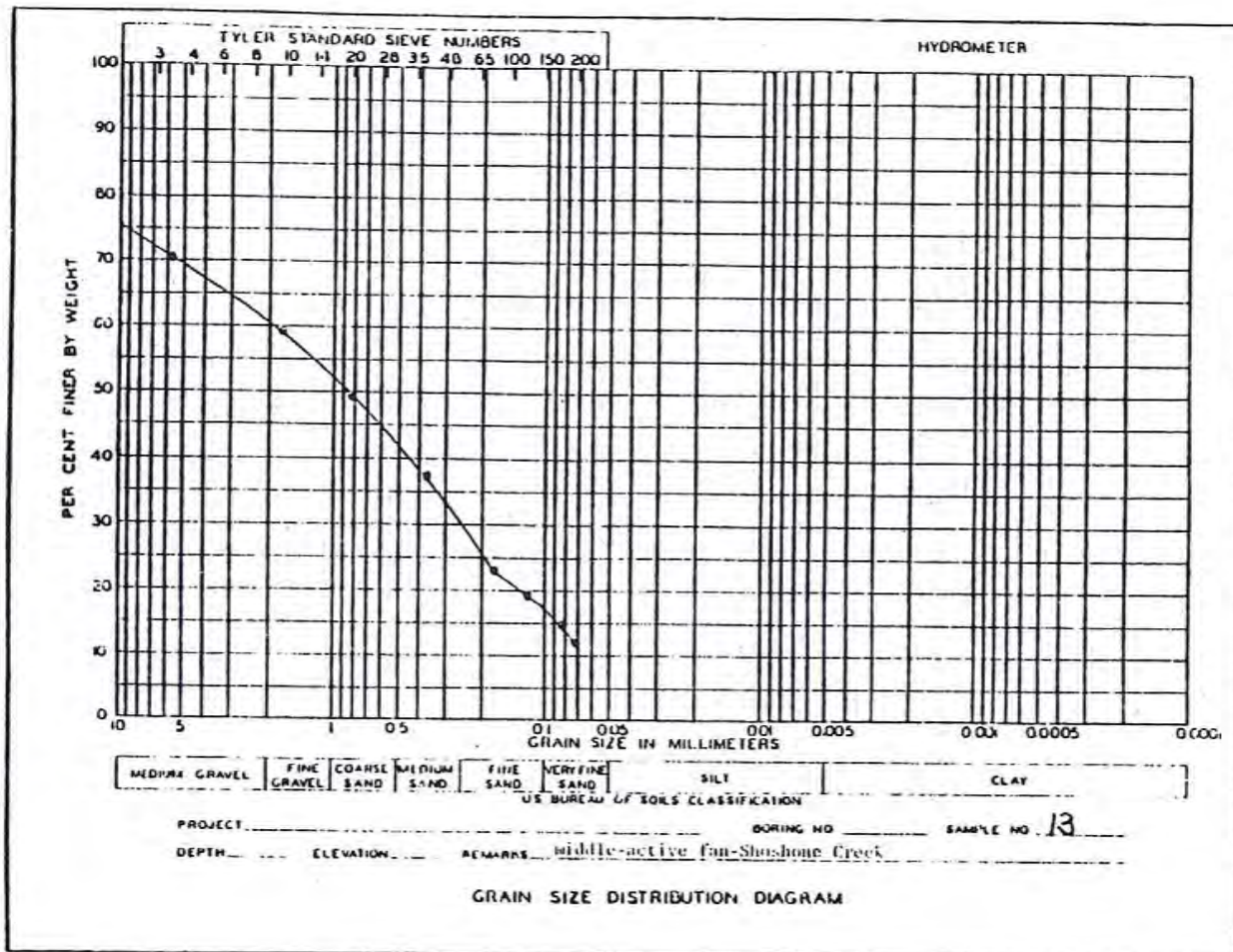
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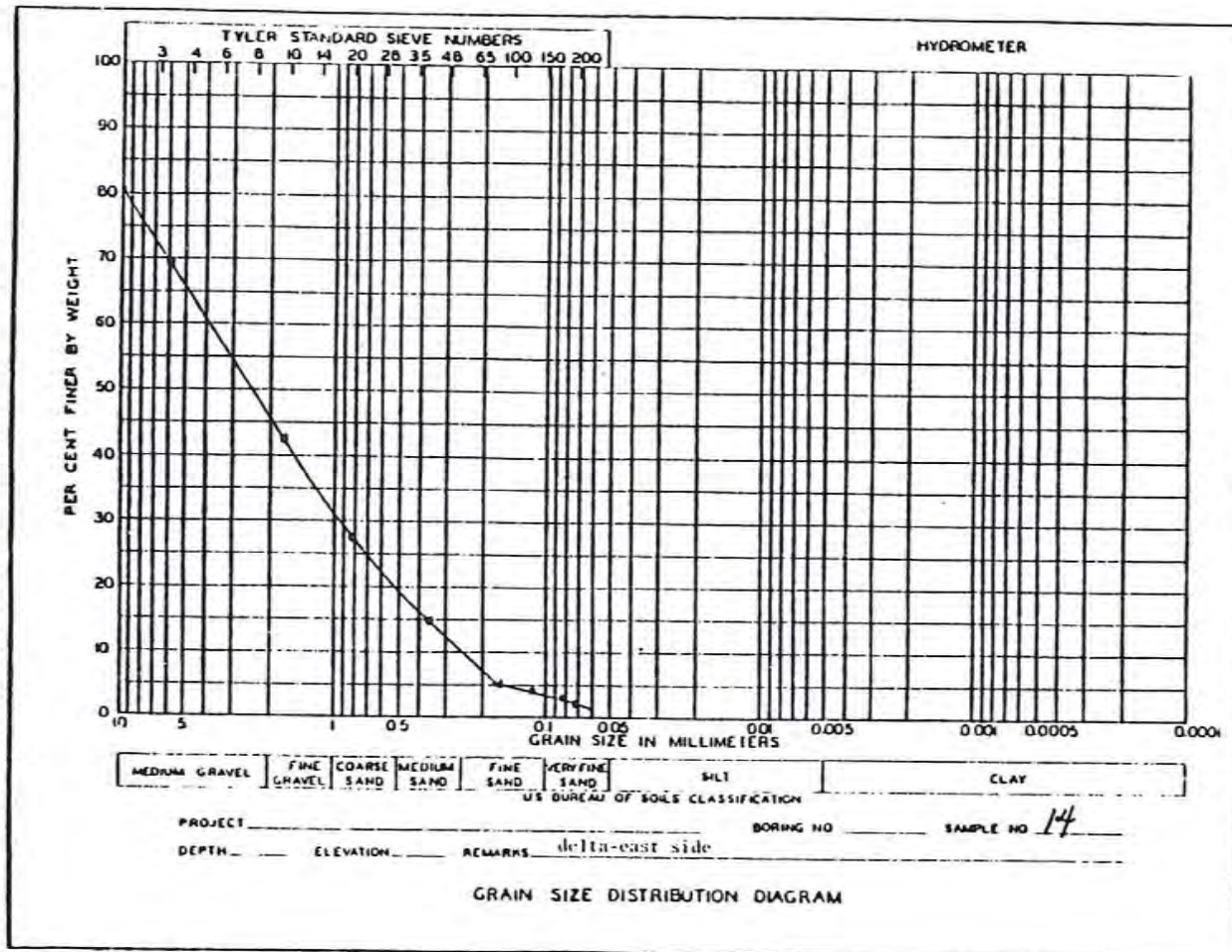
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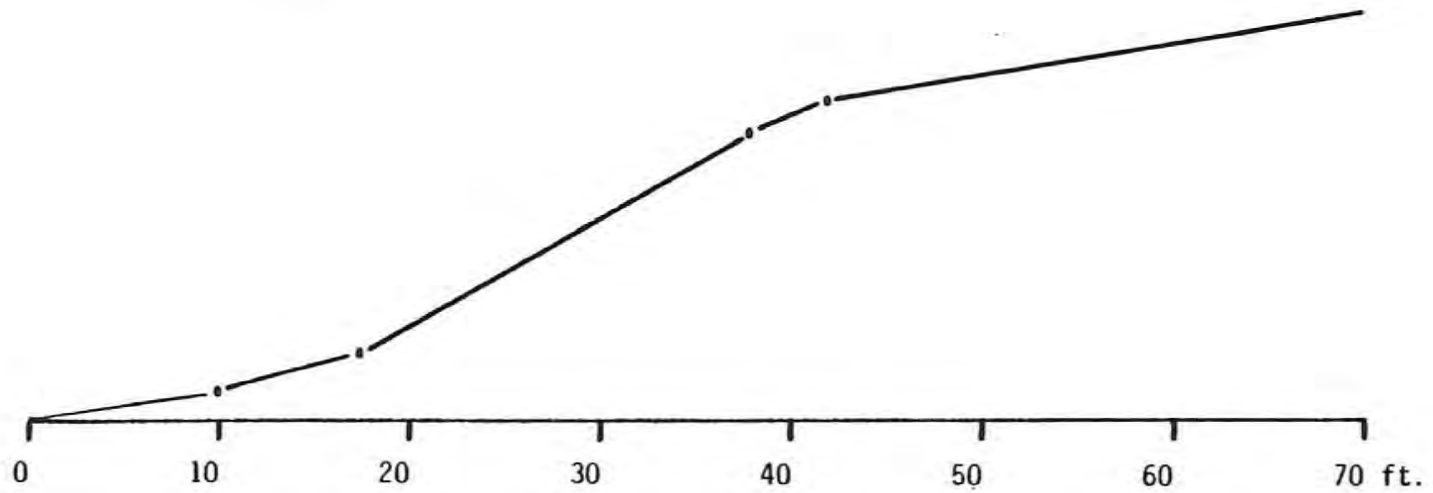
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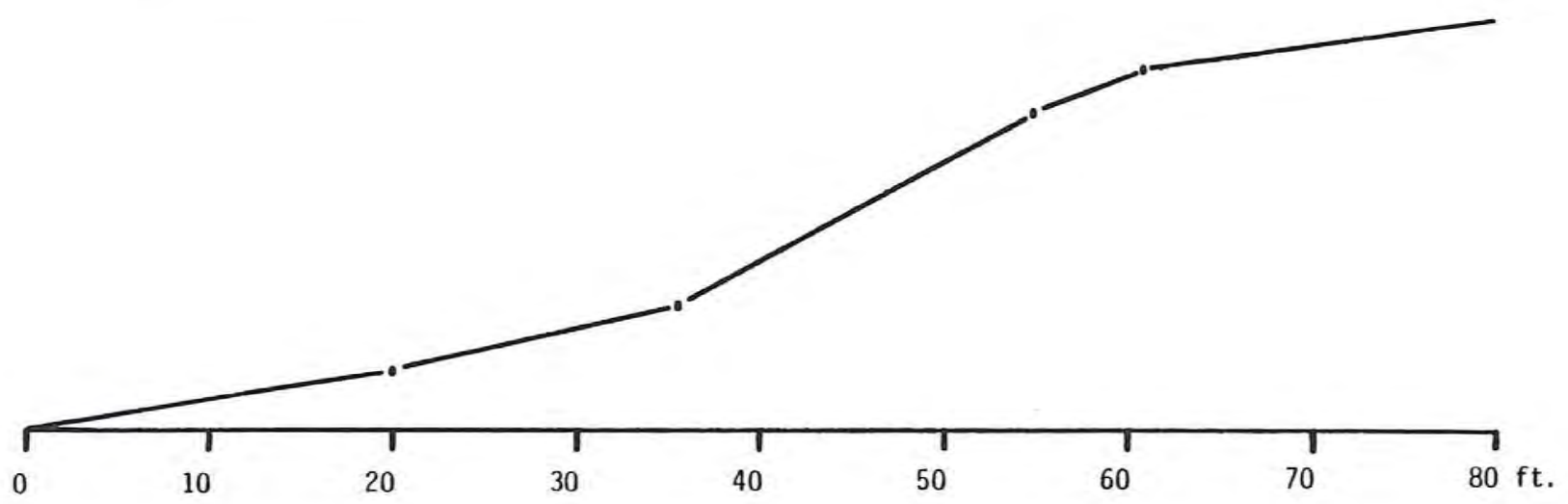
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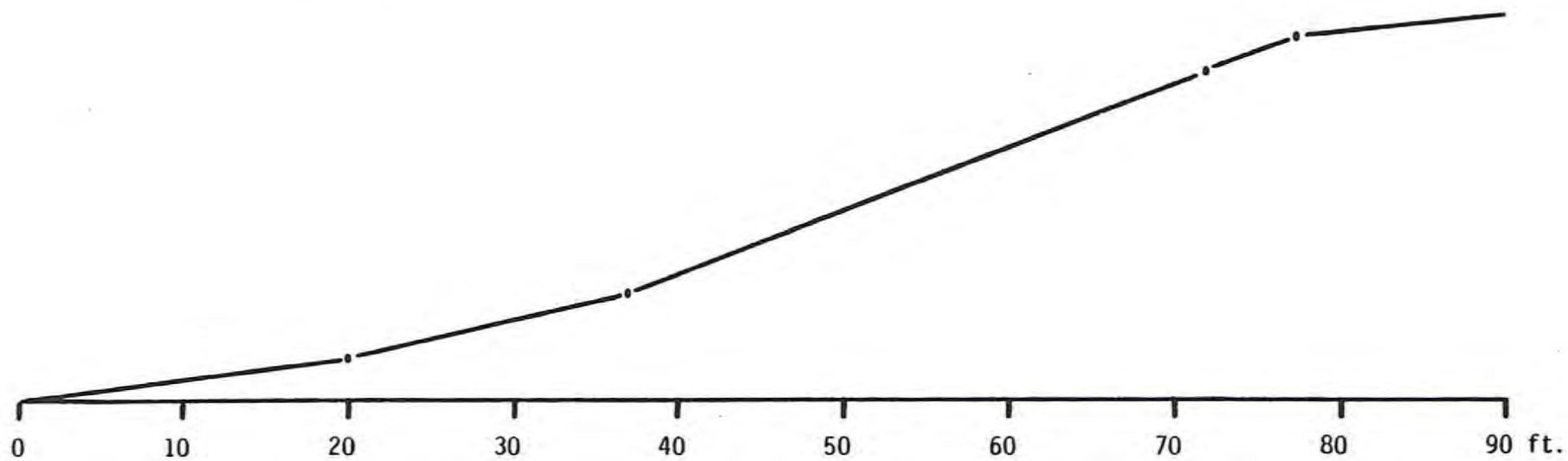
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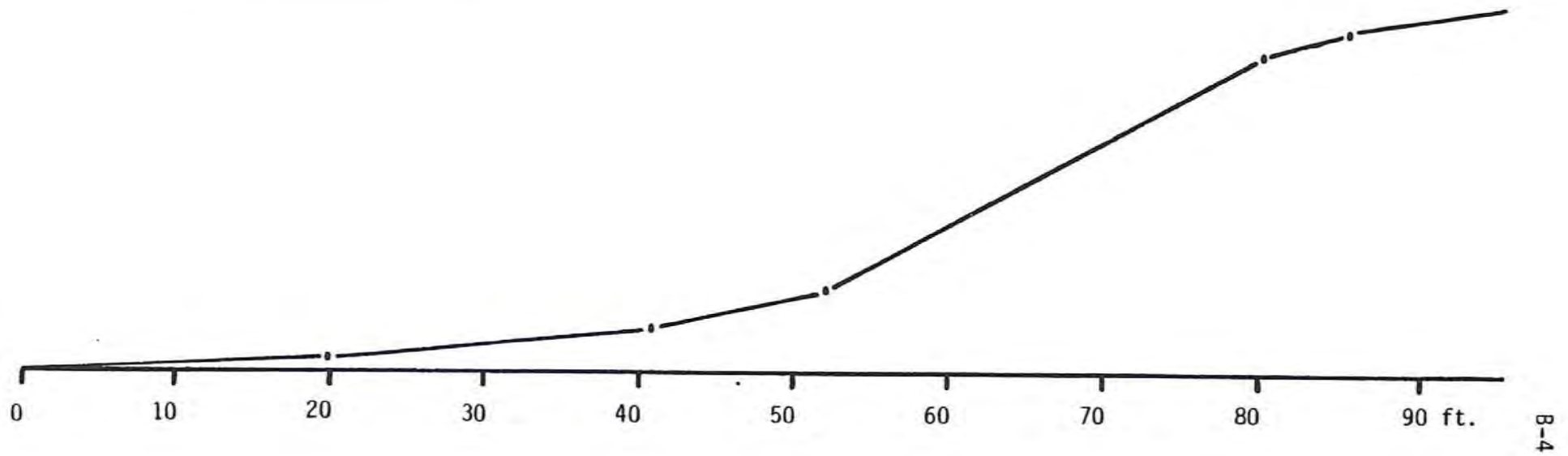
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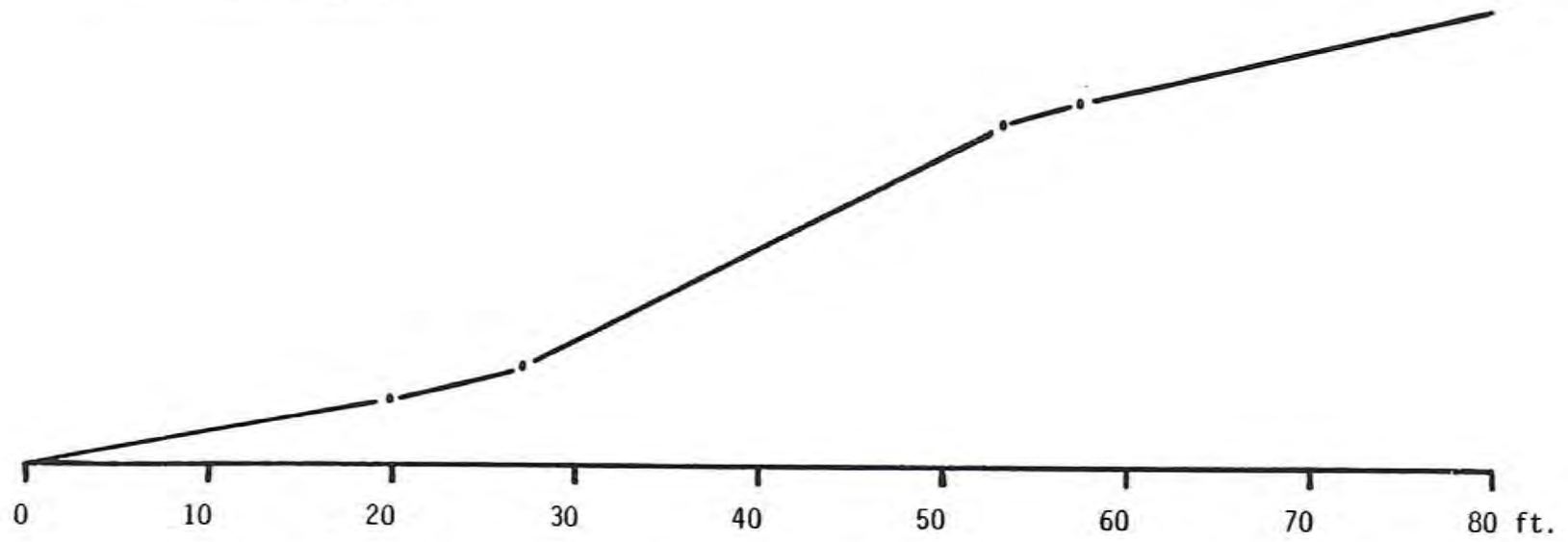
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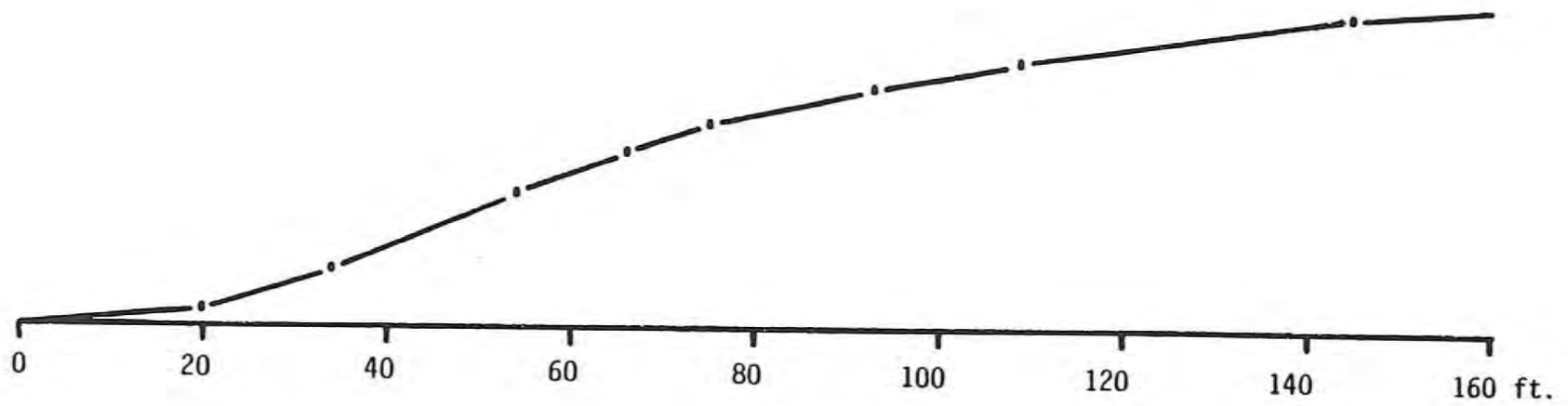
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o = inflection point



Fault Scarp Profile # 6

o = inflection point



APPENDIX F-2

GEOHERMAL RESERVOIR ASSESSMENT CASE STUDY, NORTHERN
BASIN AND RANGE PROVINCE, NORTHERN DIXIE VALLEY, NEVADA

Volume III: Soil Geochemistry and Petrochemistry

Conducted for:

U. S. Department of Energy
Contract number DE-AC08-79ET27006

Work Performed under Subcontract to:

Southland Royalty Company
Fort Worth, Texas

Conducted by:

Mackay Minerals Research Institute
University of Nevada, Reno, Nevada

Submitted:

July 31, 1980

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*MyIars of plates are on file with the Nevada Bureau of Mines and Geology, University of Nevada, Reno, Nevada 89557. Copies are available for the cost of reproduction.

Chapter 1. INTRODUCTION

By: Elaine J. Bell and Lawrence T. Larson

1.0 INTRODUCTION

1.1 Foreword

This report was prepared for the U. S. Department of Energy (DOE) in compliance with conditions of the statement of work as part of Contract number DE-AC08-79ET27006 for Geothermal Reservoir Assessment in the northern Basin and Range Province. Work was performed by the Mackay Minerals Research Institute (MMRI) under subcontract to Southland Royalty Company (SRC), Fort Worth, Texas.

The MMRI, with the Mackay School of Mines as lead agency, is charged with performing research in the general field of non-renewable resources. Optimal use of the staff and facilities of the University results from the various components of the University cooperating in interdisciplinary research. Table 1-1 lists specific individuals involved in this aspect of the project, their respective affiliation, title, investigation areas, and level of effort. The specific personnel and their individual responsibilities in completing the various investigations are indicated in the following chapters of this report.

The overall reservoir assessment dealt with the northern Dixie Valley area, Nevada (Figure 1-1), and included specific investigations conducted within the study area: 1) Structural-Tectonic Analysis; 2) Petrologic Alteration Studies; 3) Hydrology and Hydrogeochemistry; and 4) Shallow Temperature Survey. The results of these studies were presented in the report submitted January 31, 1980 (Mackay Minerals Research Institute, 1980). Follow-on studies specifically involving 1) Mercury-Arsenic Soil Geochemistry and 2) Petrochemistry are presented in the following technical report in written format supplemented by appropriate graphic data and appended information. This technical report should be viewed as part of the necessary and on-going process of investigation leading to an understanding of the Dixie Valley Geothermal System.

1.2 Purpose

The purpose of this phase of the MMRI program was to provide a large-scale surface geochemical survey for mercury (Hg) and arsenic (As) within and peripheral to an area of exploration for high temperature geothermal resources and to conduct petrochemical analysis of subsurface drill chip samples from two deep exploratory wells (DF 45-14

Table 1-1. MMRI Personnel

<u>Name</u>	<u>Affiliation*</u>	<u>Title (Area of Investigation)**</u>	<u>Level of Effort</u>
Bell, Elaine J.	MSM	PI/Project Administrator	4 man-months
Junca1, Russell W.	MSM	GRF (SG)	4 man-months
Larson, Lawrence T.	MSM	Principal Investigator	1 man-month
Bard, Thomas R.	MSM	GRF (P)	1 man-month
Nosker, Richard E.	MSM	GRF (P)	½ man-month
Nosker, Sue A.	MSM	GRF (P)	½ man-month

* MSM -- Mackay School of Mines

** GRF -- Graduate Research Fellow; P -- Petrochemistry; SG -- Soil Geochemistry;

PI -- Principal Investigator

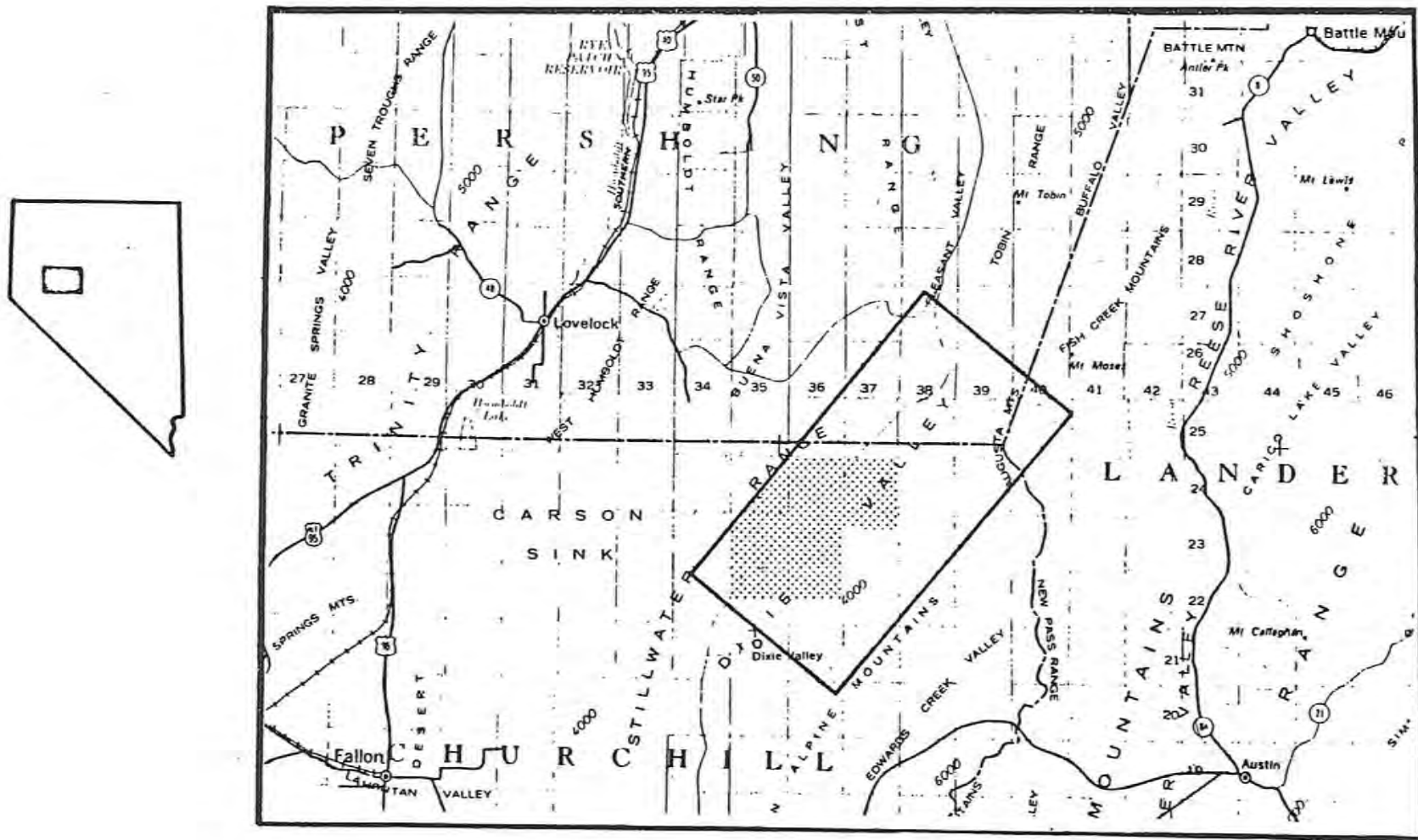


Figure 1-1. Index map of Dixie Valley study region. Shaded area of State shown enlarged. Area encompassing soil geochemistry and petrochemistry studies is shown (in pattern) within the Dixie Valley prospect area; refer to Plate I for detail.

and DF 66-21) to identify steam or hot water entries and to delineate the associated geochemical zonation.

1.3 Scope

The scope of the study as outlined in Appendix A, Scope of Work, of the Prime Contract and Subcontract (Contract number DE-AC08-79ET27006, contract modification A001) included the following major tasks for each investigation:

Soil Geochemistry

- (1) Establish broad scale and detailed grid networks and traverse lines within the approximately 30 square mile study area designated along the front of the Stillwater Range.
- (2) Collect and analyze an estimated 400 soil samples for mercury and arsenic content.
- (3) Evaluate the data.

Petrochemistry

- (1) Prepare whole rock and heavy mineral fraction samples for representative intervals within each of the wells: DF 45-14 and DF 66-21.
- (2) Analyze the samples, optimally 100-foot composite samples, for the following elements: lead, zinc, arsenic, antimony and mercury.
- (3) Evaluate the data.

1.4 Study Approach and Methods

The specific approach and methods of each investigation are presented in the respective chapters of this report.

1.5 Report Organization

This technical report is presented in chapter format with each chapter authored by the person primarily responsible for data development and accomplishing the tasks outlined in the statement of work. Chapter 2 presents a general geologic and geographic setting within which to evaluate the results of the specific soil geochemistry (Chapter 3) and petrochemistry (Chapter 4) investigations. Chapter 5 provides a brief re-evaluation of the integrated model of the Dixie Valley Geothermal System developed during the first phase of the Case Study in terms of the data contained in this report. Where necessary, graphic display of data in the form of tables, charts, photographs,

figures or maps is included to supplement and enhance the text presentation.

Preparation of this technical report involved the following process:

- (1) Authorship of the respective chapters on each investigation.
- (2) Collation of Chapters 1 through 4 to form the Draft Final Report (DFR).
- (3) Review of the DFR by MMRI project personnel and by key SRC personnel.
- (4) Preparation of the Final Technical Report.

1.6 Acknowledgements

On behalf of the Graduate Research Fellows and ourselves, we express our sincere appreciation to the following representatives of Southland Royalty Company: Mr. Jere Denton, District Manager of Natural Resources, for his continuing support, cooperation and guidance since the inception of this project as a joint venture between the MMRI and SRC; and Mr. Dennis S. McMurdie, Geothermal Geologist, for his patience and invaluable assistance during the course of our investigations.

We wish to express our appreciation to the U. S. Department of Energy for their willingness to support the MMRI and its research in Dixie Valley. We would particularly like to thank Mr. Joe Fiore for his cooperation and encouragement.

We also wish to extend our thanks to the numerous support personnel of the Mackay School of Mines and the University of Nevada who assisted our efforts toward completion of this project. Purchasing agreements and bookkeeping were ably handled by Mrs. Betsy Peck and Mrs. Louise Gibbs. Superb secretarial services provided by Ms. Mollie Stewart made the preparation of this report possible. Ms. Alice Kellames is thanked for her patience and capable assistance in handling personnel contracts and the various unforeseen problems that arose during the course of the study. Contract expenditure records were maintained in timely and efficient manner by Mr. James Murphy and Mr. Barry Myers of the UNR Controllers Office.

And finally, a special thanks to those persons who have patiently and continually supported our efforts and investigations throughout the

entire Dixie Valley project and who we may have inadvertently forgotten to acknowledge.

1.7 References

Mackay Minerals Research Institute, 1980, Geothermal reservoir assessment case study, northern Basin and Range Province, northern Dixie Valley, Nevada: Rept. prepared for U. S. Dept. of Energy, contract no. DE-AC08-79ET27006, v. 1, 223 p. plus appendices, v. II, map plates; also Univ. Utah Res. Inst. Rept. NV/DV/SR-13.

Chapter 2. GEOGRAPHIC AND GEOLOGIC SETTING

By: Russell W. Junca1 and Elaine J. Bell

2.0 GEOGRAPHIC AND GEOLOGIC SETTING

2.1 Introduction

The following discussion provides a framework within which to evaluate the soil geochemistry and petrochemistry investigations. General geographic and geologic settings are presented along with specific discussions of pertinent aspects of the study area, such as the mineralized zones. A brief summary is presented here, and the reader is referred to the Case Study (MMRI, 1980) for greater detail.

2.2 Geographic Setting

2.2.1 General

This study focusses on the southwestern portion of the Dixie Valley prospect area (Figure 1-1). Bounded by the Stillwater Range and the Humboldt Salt Marsh on the west and east, respectively, and extending from Dixie Meadows on the south to the Boyer Ranch on the north (Plates I through III), the study area encompasses the two deep exploratory wells (DF 45-14 and DF 66-21; Plate III).

Large alluvial fans slope from the mountain canyons and coalesce to form continuous alluvial plains. The valley floor is dominated by the nearly level playa, with the Humboldt Salt Marsh located within the southern portion of the playa. The Marsh is the sink for waters from Dixie, Jersey and Buffalo Valleys, as well as the Eastgate drainage from the south.

During late Wisconsinan time (approximately 12,000 years before present) a lake contemporaneous with but separated from Pleistocene Lake Lahonton reached an estimated depth of 73 meters (m) (220 feet) in the valley. Geomorphic remnants of the lake remain as old shorelines, bars and delta deposits on some of the fan surfaces.

2.2.2 Climate

The study area is arid to semi-arid with high evapotranspiration and abundant sunshine. Daytime temperatures are hot in summer and mild in winter; nights are cool or cold. Topographic position is an important factor in the local climate. The valley floor at an elevation of approximately 1037m (3400 feet) receives less than 20 centimeters (cm) (8 inches) of precipitation annually, while the upper portions of the

ranges with median elevations of 1830 to 2135 m (6000 to 7000 ft) average 30 cm (12 inches) or more. The elevation moderates the daytime temperatures in the mountains, although at night temperature inversions commonly occur. Recorded temperature extremes on the valley floor range from -24°C to 43°C (-11°F to 109°F).

2.2.3 Soils

The moderately to poorly developed arid soils of the area largely reflect their geographic position. Soils within and adjacent to the playa are poorly drained with groundwater within 1 m (39 inches) of the surface. Composed mainly of clay and silt they show little or no horizon differentiation (entisols). A surface salt crust or a salt horizon is sometimes present; vegetation is lacking.

Further from the playa are poorly and somewhat poorly drained soils of the basin-fill plains. The water table is commonly less than 1 m (39 inches) below the surface. The soils are clays, silty clay loams or silty loams underlain by silt and clay basin-fill deposits. These soils show little horizon development, although they may have a salic horizon. All are saline. Greasewood (*Sarcobatus vermiculitis*), salt-bush (*Atriplex lentiformis*), iodine bush (*Allenrophia accidentalis*) and salt grass (*Distichlis apicata stricts*) are the most common plants.

The third important soil environment is on the alluvial fans. Two important soil groups within the study area are found in this environment. The toes of the fans (2-4% slope) are covered largely by loamy, well drained soils showing a moderately fine textured subsurface horizon where clay has accumulated (argillic or cambic horizon). Shadscale (*Atriplex confertifolia*), and greasewood are the predominant shrubs. Rabbit brush (*Chrysothamus*, species unsure) inhabits the drainage courses and localized areas where the water table may be close to the surface. Further up the fan slopes (4-15% slope) are well drained loamy skeletal (35% gravel, cobble or stone) soils which generally show a weak argillic or cambic horizon. Typical vegetation is similar to the lower fans with the addition of bud sagebrush (*Artemesia spinecens*) and halogeton (*Halogeton glomeratus*); several species of grass are sparsely distributed. For a more detailed discussion of the soils, the reader is referred to Alexander and Peterson (1974).

2.3 Geologic Setting

2.3.1 General

The recent geologic history of the area is most important to this study and includes considerable seismic and geothermal activity. Large earthquakes accompanied by displacement and rupture of the ground surface occurred in 1903, 1915 and 1954. The waxing and waning of surface thermal activity may be related to the seismic activity in the area. Surface evidence of far greater thermal manifestations in the past is shown by areas of highly altered rock along the range front and large tufa mounds such as those at Sou Hot Springs in the north end of the valley. Numerous hot springs and fumaroles are still active, including a group of approximately ten fumaroles along the range front within the study area (Plate III).

The schematic model of northern Dixie Valley shown in Figure 2-1 depicts the valley structure as a complex asymmetric graben. The innermost portion of this stepped graben structure may contain as much as 3000 m (10,000 feet) of alluvial fill (Thompson and others, 1967). It can be seen that the general trends are northeasterly. However, cross-cutting structures such as the White Rock Canyon fault may be of particular significance to the geothermal system. Generalized structural trends within the specific study area are shown on Plate III.

The mountain ranges are largely composed of folded and faulted Triassic and Jurassic metasilstone and limestone overlain by upper Jurassic to Tertiary rhyolitic to dacitic tuffs, welded tuffs and flows, andesite and basalt flows and tuffaceous sediments (Figure 2-2). Intrusions of gabbroic and dioritic rocks of probable Mesozoic age are also volumetrically significant.

2.3.2 Dixie Valley Geothermal System

The complex structural setting of Dixie Valley has made the geothermal system difficult to characterize. Many of the structures serve or have served as preferential conduits for fluid migration as evidenced by the alignments of springs, seeps and fumaroles and the subsurface and surface concentration of intense hydrothermal alteration along these features. This is best seen along the front of the Stillwater Range where very intense localized alteration is observed, along with fumaroles

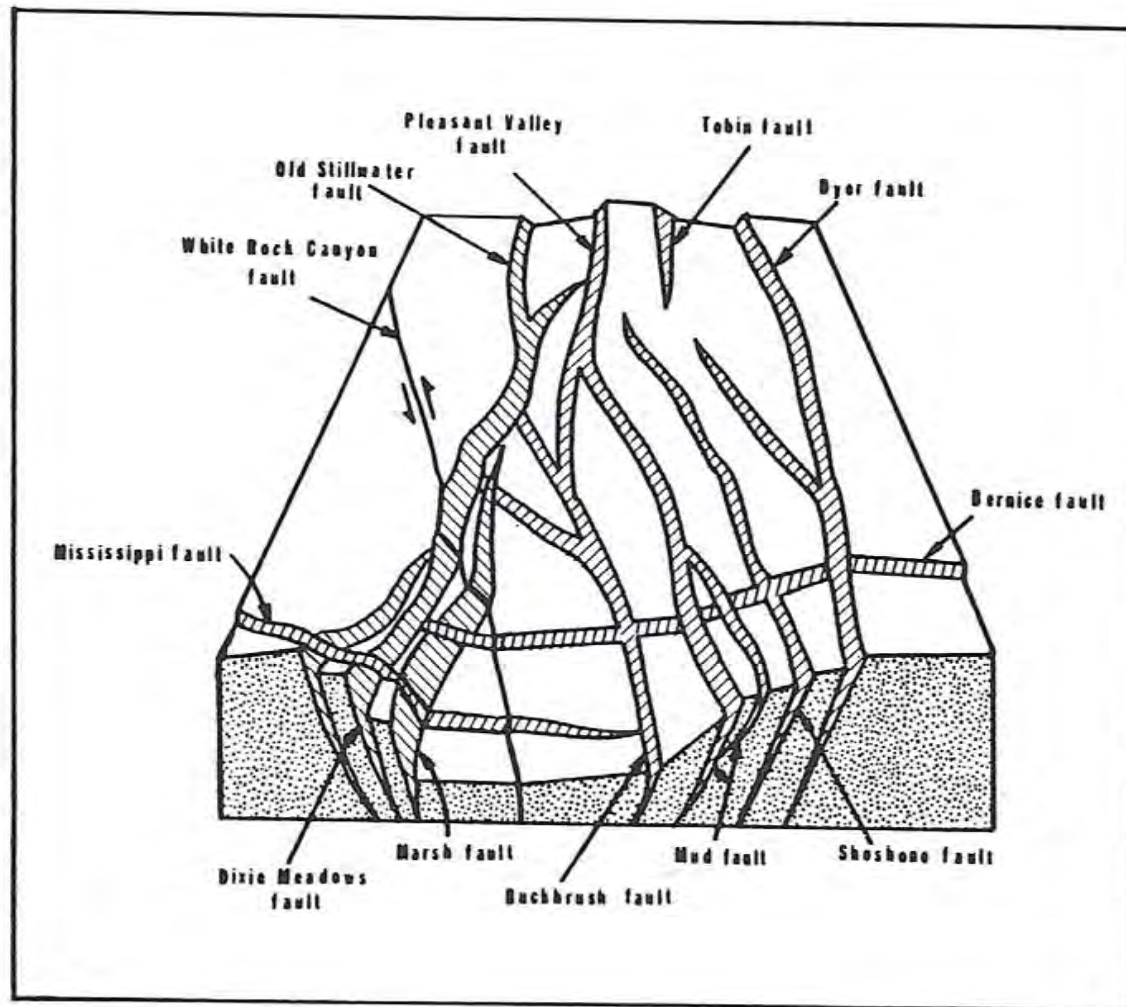


Figure 2-1. Structural model of northern Dixie Valley. Alluvium is removed and bedrock surfaces restored. Modified from Whitney (1980).

E X P L A N A T I O N

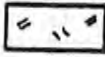

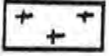

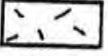




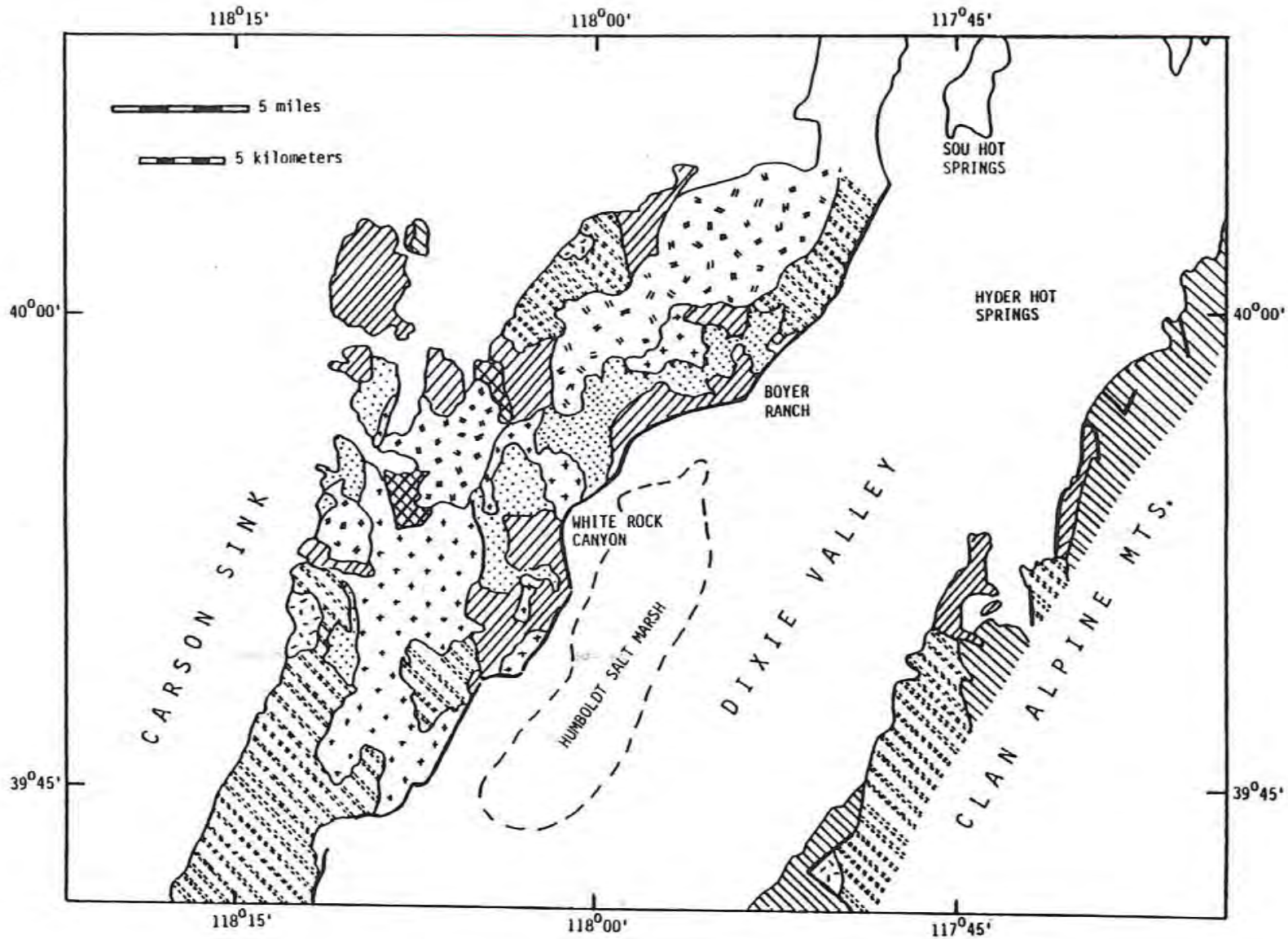
TERTIARY-- QUATERNARY	[Plio-Pleistocene		Basalt & Andesite
		Pliocene (?)		Non-Marine Sediments & Tuffs
		Miocene (?)		Tuffs & Flows of Rhyolite, Dacite
TERTIARY & MESOZOIC	[	Undifferentiated Pre-Lahontan Sediments & Volcanics
				Granitic Intrusives
MESOZOIC	[Upper Jurassic		Basalt Flows, Tuffs & Breccias
		Upper Jurassic		Gabbroic & Dioritic Intrusives
		Upper Triassic		Slate & Phyllite
			Fault	

Figure 2-2. Generalized geologic map of the Dixie Valley area.
Modified from Meister (1967); map on following page.



and the presence of hot water within 30 m (100 feet) of the surface.

Water chemistry data have shown major differences between hot spring systems at Sou, Hyder and Dixie Meadows, implying a general lack of communication between the springs. As yet however, the heat source and the precise structural or stratigraphic factors which may control the system have not been defined.

Exploratory drilling has confirmed the existence of a hot water-steam(?) system at depth. This conclusion is based on very limited information available for a group of wells (SW Lamb #1, #3 and #4) drilled by Sun Oil Company in the vicinity of the Boyer Ranch. Wells DF 45-14 and DF 66-21 have shown high temperature fluids in certain intervals, but flow rates have generally been low. The locations of the various deep exploratory wells are shown on Plate III.

2.3.3 Mineral Deposits

Numerous mineral deposits occur in or near the study area and, although production has been limited, they are significant to the present work. In the Corral Canyon area (Plate III) on the east front of the Stillwater Range, small gold deposits were worked in the 1920's and 1930's with minimal production and, more recently, some prospecting for titanium minerals was done along albite-calcite dikes which cut the gabbroic country rock (Willden and Speed, 1974). The Dixie Comstock Mine (Plate III) also lies along the eastern front of the Stillwater Range. Recently reopened, it was last worked for gold and silver prior to World War II. The ore body is a quartz vein in altered rhyolite. An amalgamation mill was erected in 1935 but is no longer standing. Production figures show \$293,304 from 17,880 tons of ore with a 2 to 3 gold/silver ratio (Willden and Speed, 1974). The mining was hampered by large quantities of hot water encountered less than 75 feet from the surface.

The upper reaches of both White Rock Canyon and, to a lesser extent, Cottonwood Canyon (Plate III) contain numerous mineral workings within the Table Mountain district. These include a nickel mine, a cobalt-nickel-copper mine, several copper prospects and several small iron deposits. The nickel-cobalt mines may be particularly significant to the study as the ore reportedly occurs as arsenides and sulfarsenides (Willden and Speed, 1974).

Just outside the study area on the west side of the Clan Alpine Range are deposits of mercury, silver and antimony in the Bernice Canyon district. Most important is the Red Bird Mine and associated prospects in Shoshone Canyon. Shoshone Creek which drains the area flows east and south into the Humboldt Salt Marsh. A portion of the study area is within the Shoshone Creek flood plain and the creek may be a source of anomalous concentrations of both mercury and arsenic. The Red Bird cinnabar deposit produced approximately 40 flasks of mercury prior to 1943 (Willden and Speed, 1974).

Numerous prospect pits are found throughout the entire Dixie Valley area and much small-scale mineralization is evident.

2.4 References

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Chapter 3. MERCURY AND ARSENIC SOIL GEOCHEMISTRY

By: Russell W. Junca1

3.0 MERCURY AND ARSENIC SOIL GEOCHEMISTRY

3.1 Introduction

3.1.1 Purpose and Scope

The purpose of the present study was to map and interpret the soil geochemical distribution of mercury (Hg) and arsenic (As) in Dixie Valley, Nevada, and to relate the observed distribution patterns, where possible, to the presence of a geothermal influence. The location and approximate extent of the study area are shown on Plates I through III.

The study utilized a broad sampling network to identify and delineate specific areas for more detailed study. Additionally, high density sampling was performed in the vicinity of the two exploratory wells and across specific structural features as outlined by earlier work (Whitney, 1980).

Dixie Valley is presently undergoing exploration for geothermal resources and much geologic, geophysical and hydrologic data, as well as shallow temperature gradient and deep exploratory drilling information is available and can be related to the geochemical results (Mackay Minerals Research Institute, 1980). The data have been correlated to known structures, producing wells, geophysical anomalies and sites of near-surface thermal activity to evaluate the cost-effectiveness of the technique for delineating permeable structures which may serve as drilling targets.

3.1.2 Methods and Analytical Techniques

3.1.2.1 Sampling

The initial sampling grid of approximately 400 points is shown on Plate I. The majority of the grid consists of parallel profile lines approximately 730 meters (2400 feet) apart bearing N60°W, essentially normal to the average valley trend. Sample points along these lines are approximately 305 m (1000 ft) apart. The data from the grid were evaluated and used as the basis for selection of follow-up profiles sampled at 30 m (100 ft) intervals which are indicated on Plate III. Certain known structures and the areas around two geothermal wells (DF 45-14, DF 66-21) were sampled at 30 to 180 m (100-600 ft) intervals (Plate III).

Points on the sampling grid were located by bearing and traverse

using a surveyor's wheel. Field notes at each sample point included a description of the soil, the surrounding vegetation, soil moisture condition, and additional comments including local morphology and proximity to roads or streams.

Soil samples were collected from a depth of 25 to 30 cm (10 to 12 in). The depth was chosen on the basis of three randomly selected vertical profiles from three major soil environments in the area: upper alluvial fan slopes, fan piedmont and playa. The results of the profiles show the Hg and As values generally increasing with depth, with a zone from 15 to 25 cm (6 to 10 in) where values tend to increase substantially and then level off. These results were taken as representative of the study area as a whole and indicated much higher As and Hg values roughly corresponding to the B horizon. This is not unexpected as it is commonly the site of accumulation of clays and iron/manganese oxides both of which are important scavengers of mercury and arsenic (Fang, 1978; Horsnail and others, 1969; Boyle and Jonasson, 1973). Organic material has been shown to be even more effective in scavenging mercury (Trost and Bisque, 1971; Fang, 1978), but this effect should not be important in Dixie Valley as the soils are generally only moderately to poorly developed with very little organic matter. Because identification of specific soil horizons is difficult in many places, and considering the study of Klusman and Landress (1979) which showed that variation of secondary soil parameters did not mask significant geothermal mercury anomalies, a standard sampling depth was chosen.

Dry samples were sieved in a stainless steel sieve at the site and the minus 80 mesh (0.177 mm) fraction sealed immediately in airtight glass vials. Wet samples were dried at ambient temperature before sieving.

3.1.2.2 Quantitative Analysis

A Jerome 301 Gold Film Mercury Detector was used for determining mercury concentrations in the soil samples. This analytical method is based on the change in electrical resistance undergone by gold film when mercury is adsorbed. Weighed samples are heated to 600^o C for approximately one minute. The mercury which is vaporized by this process is collected on a gold-plated coil. The coil is then heated to 600^o C on a 13 second timed cycle and the sample mercury is carried by

filtered air to the meter which consists of two gold films comprising opposite arms of a wheatstone bridge circuit. Before the air stream reaches the meter it is split and the mercury removed from one side by passing over steel wool impregnated with Palladium Black (PdCl_2). This stream serves as the reference for the other fraction from which the mercury is collected by the sensor film. The resultant difference in resistance between the bridge leg exposed to filtered air and that exposed to air containing mercury vapor reflects the amount of mercury in the sample and is displayed on a digital readout. The accuracy of the gold film detector used in this study was not determined, however, the mercury values of replicated samples which were reanalyzed by two cold vapor atomic absorption techniques including borohydride generation methods generally ran 60 to 80 percent of the AAS values.

The precision of the instrument under field conditions is reported as greater than 1 ppb (Matlick and Buseck, 1976; J. McNerney, Jerome Instrument Corporation, pers. commun., 1980), although a precision of 1 to 4 ppb is representative of the instrument used in this study. The reproducibility of the results is affected by several factors. The small sample size (approximately 0.1 gram) from within such an inhomogeneous medium as soil may not make a given sample representative, particularly in an area such as Dixie Valley where mercury mineralization is known. Other sources of variation include temporal changes in the soil mercury content and experimental errors. The former possibility was investigated by reoccupying five sample sites at two week intervals for eight weeks. The results, shown in Table 3-1, show some variability with time, although no consistent pattern is evident. The variance is similar to that observed in replicate samples used to determine the analytical variability. Matlick and Buseck (1976) showed very little variation in mercury content of four sites resampled periodically over a two month period. Capauno and Bamford (1978) report significant variations upon resampling a traverse two weeks later but note that the anomalous area remained the same.

Seventy percent of the samples were analyzed for mercury within 72 hours using the Gold Film Mercury Detector. Due to mechanical malfunction of the instrument, delays of up to 20 days occurred prior to analysis of some samples. However this delay was found to be insignificant.

Table 3-1
Sample Stations Reoccupied at Two-Week Intervals

<u>Sample number</u>	Mercury Values (in ppb)			
	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>
1	104	117	101	79
39	180	139	131	249
45	8	0	0	6
92	80	110	122	48
152	68	65	80	68

Upon rerunning 13 samples ranging from 28 to 272 ppb at one week intervals for four weeks to determine if mercury was escaping, no consistent pattern of declining mercury was observed (Table 3-2). Moreover, samples showed variations upon remeasurement similar to those of replicate samples rerun consecutively. Sample numbers, which correspond with sites indicated on Plate I, and their respective mercury concentrations are listed in Appendix A.

The arsenic analyses were performed by Rocky Mountain Geochemical Corporation of Salt Lake City using a colorimetric technique after Vasak and Sedivic (1952). In this method As is reduced to the trivalent state with potassium iodide and tin (II) chloride. The arsenic is then converted to its hydride by the action of nascent hydrogen, generated by adding zinc metal to the acid solution. The evolved arsenic hydride (arsine) is bubbled through a solution of silver diethyldithiocarbamate in a basic brucine-chloroform (or pyridine) solution. A brownish-red color is produced by the absorbing solution, the intensity of which is measured by a colorimeter and the arsenic concentration calculated. Results of the analysis indicating sample number and arsenic content are listed in Appendix A, with geographic distribution and values plotted on Plate II.

The arsenic content in many samples was not measured for up to two months from the time of sampling. Due to its lower volatility relative to mercury this was not expected to pose a problem. To verify this assumption, eight samples were rerun after more than four months in sealed storage. These samples showed no significant loss of arsenic as indicated by Table 3-3.

3.1.2.3 Data Presentation

Much of the data reduction and presentation has been performed by computer. The subroutines SRFACE and EZMXY from the National Center for Atmospheric Research graphics package (Wright, 1977) were used to make the chemical surface plots and profile line graphs, respectively. Subroutine CONRCM from the same package generated the contours of the mercury and arsenic values for the broad grid and well site sampling. The programs were run from time share and output to a Hewlett Packard 7202A plotter.

Table 3-2

Successive Mercury Analyses at One-Week Intervals

<u>Sample Number</u>	<u>Mercury Values (in ppb)</u>			
	<u>1</u>	<u>2</u>	<u>3</u>	<u>4</u>
15	88	82	50	63
36	232	170	305	280
65	28	10	21	15
85	116	109	124	125
98	376	308	366	343
125	52	87	48	85
160	272	240	350	248
193	156	135	130	120
242	40	25	50	20
287	156	165	120	115
310	28	55	38	35
362	84	70	62	75
401	16	10	10	12

Table 3-3

Successive Arsenic Analyses at Four-Month Interval

<u>Sample Number</u>	<u>Arsenic in ppm (9-6-1979)</u>	<u>Arsenic in ppm (3-4-1980)</u>
1-10	11	10
1-35	14	15
2-5	15	15
2-15	12	10
2-35	16	10
3-15	46	50
3-35	25	30
3-45	24	20

Statistical analysis of the data utilized portions of the Statistical Package for the Social Scientist (Nie and others, 1975). Subprogram CONDESCRIPTIVE was used to compute descriptive statistics which included mean, standard error, standard deviation, kurtosis and skewness. Subprogram FREQUENCIES provided histograms of the data, and computed absolute and relative frequencies as well as cumulative frequency (%). The cumulative frequency data were used in the preparation of log probability plots.

The relationship between the arsenic and mercury values was evaluated using the subprogram SCATTERGRAM. This routine draws an x-y plot and computes the correlation coefficient (Pearson's r), significance of r , standard error of the estimate and the equation of the regression line.

Analysis of replicate samples for temporal and analytical variance was performed by subprogram ONEWAY. Output from this routine is a standard analysis of variance table which includes degrees of freedom, sum of squares, mean square, F ratio and F probability.

The computer methods were supplemented with the use of log probability paper with a logarithmic ordinate scale and an abscissa scale of cumulative frequency (%) or probability. The probability axis is arranged such that a population that is lognormally distributed will plot as a straight line. An important consequence of this property is that a bimodal distribution, such as background plus anomaly, which is frequently encountered in geochemical sampling will show a sigmoid probability curve with two fairly linear segments joined through an inflexion point. Both the arsenic and mercury values from the broad grid sampling were plotted on log probability paper prior to the computer analyses which greatly aided their interpretation. For a more detailed discussion of probability plots including their use in partitioning mixed populations and selecting threshold values the reader is referred to Sinclair (1974).

3.1.3 Previous Work

A general association of arsenic and mercury with geothermal activity has been demonstrated by many workers. Weissberg and others (1979) reviewed a number of active geothermal systems where Hg and As precipitates occur and/or where thermal fluids contain high concentrations of

these elements. White (1967) showed the occurrence of Hg and As mineralization in numerous active or fossil hot spring systems. Robertson and others (1978) gave evidence of elevated Hg and As levels in the noncondensable gas, steam condensate and flashed brine phases of effluent from nine geothermal fields. Tonani (1970) showed high mercury levels in stream sediments around two Italian geothermal areas. Koga and Noda (1976) found anomalous Hg and As concentrations in both fumarole condensates and altered rocks in several Japanese thermal systems. Matlick and Buseck (1976), Phelps and Buseck (1978) and Klusman and Landress (1978, 1979) have shown broad Hg anomalies in the soils of several known geothermal areas. Klusman and others (1977) also showed broad arsenic soil anomalies in six Colorado thermal areas. More detailed soil surveys in known geothermal areas have shown mercury anomalies associated with faults and/or geophysical anomalies (Klusman and Landress, 1978; Souto, 1978; Capauno and Bamford, 1978). Surveys of this type may prove useful in selecting drilling targets.

Some mercury sampling of a reconnaissance nature was done in Dixie Valley by the Southland Royalty Company. This amounted to analysis of samples obtained during the installation of a one meter temperature survey net. The data will probably not be published but are mentioned briefly in this report.

3.1.4 Geochemistry of Mercury and Arsenic

Evaluation and interpretation of the field and laboratory data require consideration of certain aspects of the geochemistry of As and Hg. The elemental soil distribution patterns associated with geothermal activity in Dixie Valley are complicated by the presence of sulfide mineralization. Mercury and arsenic geochemical haloes associated with a variety of mineral deposits are well known. Numerous mineral exploration efforts using Hg and As content in soils have been conducted in the past and many of the geochemical principles developed in this work are applicable to geothermal exploration.

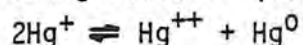
3.1.4.1 Mercury

Mercury can exist in three principal oxidation states, +2, +1 and the native metal (0). The relatively high vapor pressure of the native

element and, to lesser degrees, certain Hg compounds is unique among the metals. Mercury volatility is enhanced by the elevated temperatures within a geothermal system and its mobility in the vapor phase allows it to migrate to the surface through permeable zones. Robertson and others (1978) noted that elemental mercury was the dominant gaseous form of mercury in a study of effluent from eight geothermal areas.

While native mercury is known to occur in certain mineral deposits, including several within the Bernice Canyon district on the east side of Dixie Valley, generally it is found as the sulfide (cinnabar HgS) or as a trace within the lattice of other minerals. Various mechanisms have been proposed whereby mercury bound in this manner may be converted to the native element.

Figure 3-1 shows the relative stability fields of the important inorganic mercury compounds. The sulfide which has a restricted field of stability will yield native mercury in the zone of oxidation. Mercury contained within the lattice of other sulfides such as pyrite (FeS₂) or sphalerite (ZnS) may also be released as the native element upon oxidation. This involves the coupled oxidation of sulfur with the reduction of mercuric (+2) ion. The oxidation may occur in steps initially producing mercurous (+1) ions. The mercurous ions may then disproportionate by the following reaction (Jonasson and Boyle, 1972):



Natural reducing agents such as iron may account for Hg(0) release below the zone of oxidation. McNerney and Buseck (1973) proposed stoichiometric adjustments within sulfide lattices as a possible source of electrons which could reduce mercuric ion and release vaporous mercury. Khayretdinov (1971), citing the inert behavior of mercury at higher temperatures, indicates elemental mercury may be electrochemically bound to the surface of mineral grains at depth and migrate in response to electrochemical gradients. Dickson (1968) proposed a source of native mercury from thermal decomposition of existent cinnabar combined with the reaction of sulfur with certain other ionic species. The release of Hg(0) by the thermal dissociation of various inorganic mercury compounds has been shown by Koksoy and others (1967). They indicated complete decomposition of mercurous and mercuric chloride below 250^o C, and HgS below 340^o C. Detectable quantities of native mercury and the

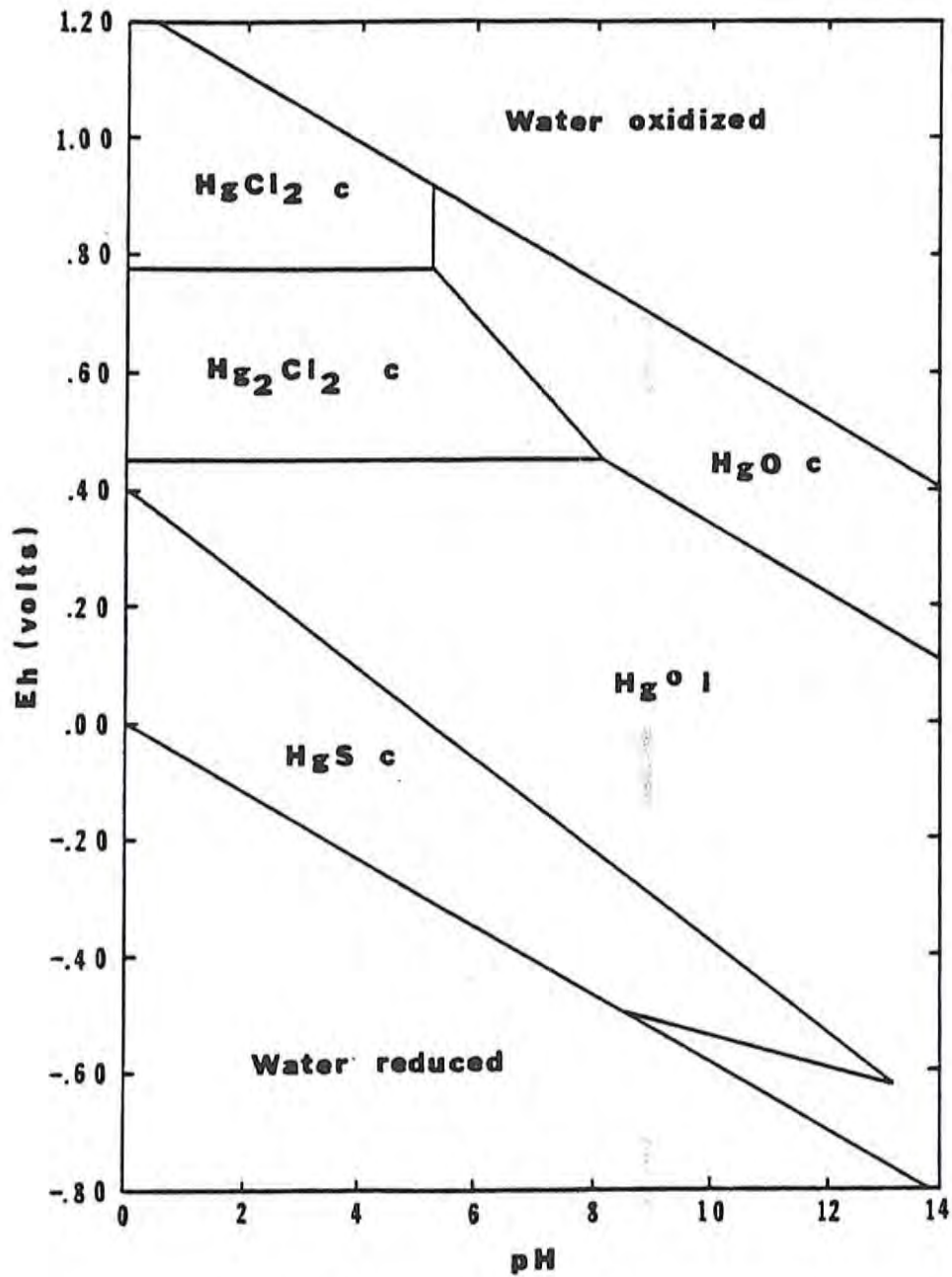


Figure 3-1. Stability fields of important inorganic mercury compounds at STP; total chloride = 36 ppm; total sulfur (as sulfate) = 96 ppm (from Hem, 1970a).

chlorides were liberated within 30 seconds at less than 80° C, detectable amounts of the sulfide were released at 210° C. Note that the samples for this experiment were crushed to pass a .074 mm (200 mesh) screen and that the amount of mercury released in 30 seconds is generally less for larger grain sizes.

Regardless of the mechanism of formation, elemental mercury will migrate in the vapor state through permeable zones toward the surface. Mercury in soil gas as an exploration technique has been demonstrated by McCarthy (1972), McNerney and Buseck (1973), and others. However, the soil-air mercury gradient is not fixed and may fluctuate temporally due to changes in temperature, barometric pressure, or moisture conditions. Because mercury tends to adsorb on clay and organic matter and to coprecipitate with hydrous oxides of iron and manganese it can accumulate in soils. A mercury anomaly in soil may be more consistent and pronounced than a corresponding soil gas anomaly.

Soil vapor anomalies are known over ore deposits as well as geothermal areas and the two can be difficult to distinguish from each other and from hydromorphic or mechanical dispersion haloes associated with mineralization. Anomalies related to both of the latter processes may occur in Dixie Valley.

Generally mechanical dispersion haloes are restricted in extent, occurring directly above or downslope from a mineralized area. The elements in this type of halo are held tightly by metallic bonds. Hydromorphic anomalies may be of greater extent and commonly occur as seepage anomalies at a break in slope. The elements in this type of anomaly are commonly adsorbed on clays or chelated by organic matter and not as tightly bound as mechanically dispersed haloes.

Figure 3-2 shows the stability fields and relative solubilities of the important inorganic aqueous species of mercury at STP in solution, with 36 ppm Cl^- and 96 ppm SO_4^{-2} . It can be seen that mercury is relatively insoluble under most circumstances. Additionally, the disproportionation reaction mentioned earlier can produce native mercury in water which may then volatilize and escape, effectively lowering its solubilities even further.

Oxygenated high chloride waters can carry a considerable amount of mercury if the waters are relatively acidic. This effect has been

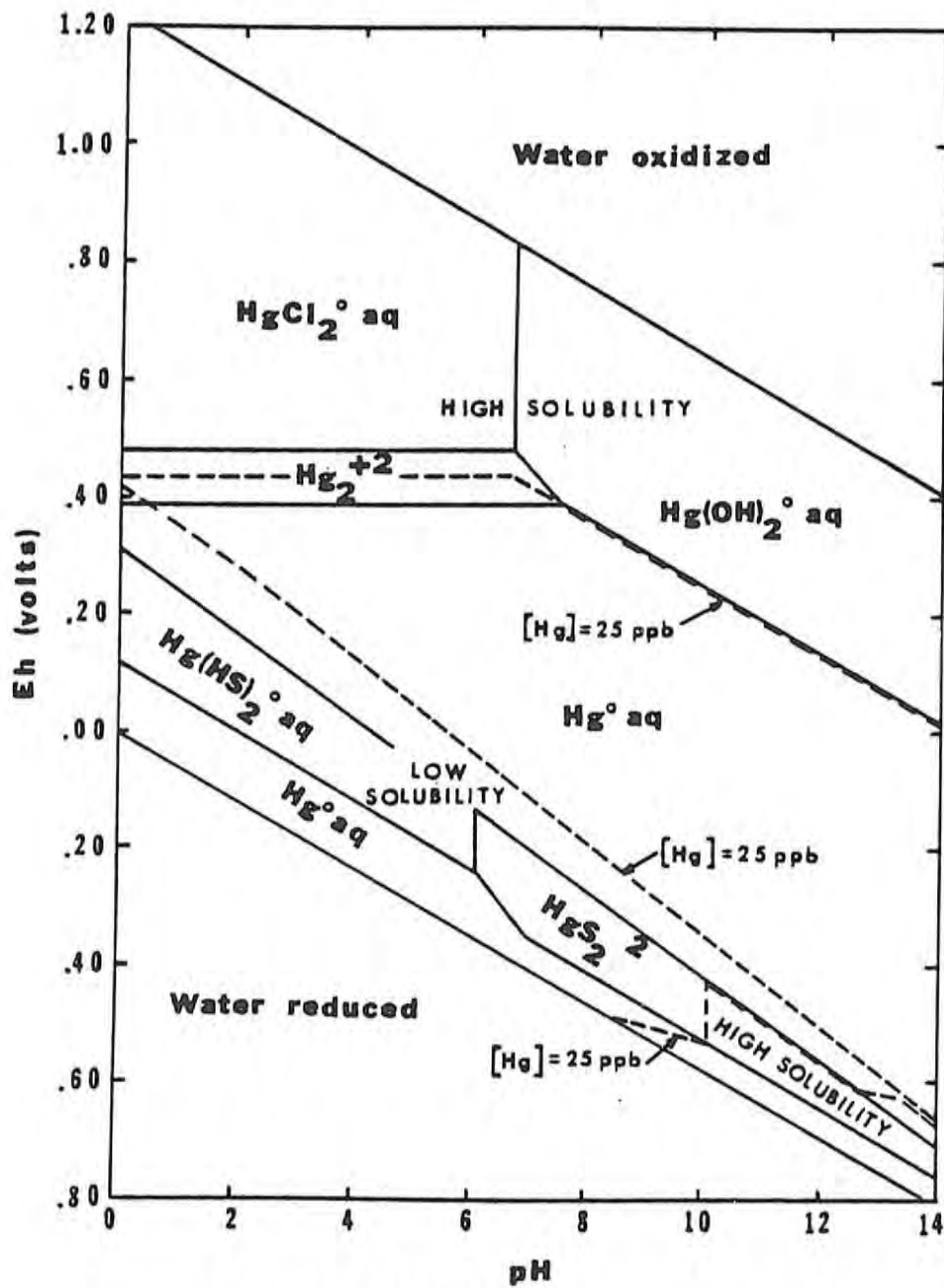


Figure 3-2. Stability fields of important inorganic aqueous mercury species at STP; total chloride = 36 ppm; total sulfur = 96 ppm (from Hem, 1970a).

observed in various geothermal areas where much higher mercury values are found in certain hot springs of high chloride and sulfate contents (Davy, 1974; Nakagawa, 1974). These types of springs probably represent supergene conditions and have been affected by surface oxidation and possibly evaporation. While relatively high chloride water is common in Dixie Valley, the deeper thermal waters exhibit low Eh values and the oxygenated surface waters are generally alkaline.

Most hot springs closely associated with the active hypogene transport and deposition of cinnabar carry only a few parts per billion mercury. The overall transporting capability of the springs is probably greater since the samples are representative of mercury concentration after deposition. Where total sulfur is high (approximately 3200 ppm) very large amounts of mercury can be held in solution as sulfide complexes. Mercury ore is typically deposited at temperatures of 100^o to 200^o C and pressures of 1 to 30 atmospheres from neutral to weakly alkaline waters of moderate to low salinity (Barnes, 1979). Above 200^o C mercury is mobile in steam even in the presence of excess sulfide (Ellis and Mahon, 1977). This effect has been observed at the Senator fumaroles in northern Dixie Valley where cinnabar and native sulfur are being actively deposited by H₂S bearing vapors with no liquid phase present (Lawrence, 1971).

Once in the surface environment in aqueous or vaporous form, the tendency of mercury to be adsorbed onto sediments will greatly restrict its mobility. In Dixie Valley where soils and waters are mostly alkaline, aqueous dispersion haloes should not persist for any great distance from the source of the mercury. However, sediments which have adsorbed mercury are susceptible to mechanical dispersion processes which may increase the size of a halo to some extent.

The adsorption behavior of mercury is central to the formation of soil anomalies. Numerous authors have demonstrated its ability to adsorb on clays and organic matter (Fang, 1978; Trost and Bisque, 1971; 1972; Rogers and McFarlane, 1978; Landa, 1978). High mercury values associated with the iron/manganese oxide fractions of some sediments suggest a coprecipitation or scavenging effect which has been demonstrated for other metals by Horsnail and others (1969). The possibility of secondary soil parameters masking mercury soil anomalies was

investigated by Klusman and others (1977) and Klusman and Landress (1978). This work indicated positive correlation between mercury content and soil pH, organic content, Fe/Mn oxides and aspect. Total carbon was found to be the most significant factor statistically. The authors concluded that, while significant, the controlling effects of these secondary parameters are overwhelmed in an area of prominent geothermal activity.

3.1.4.2 Arsenic

Arsenic exists naturally in three principal oxidation states, +3, +5 and 0. It may also occur in a -3 oxidation state in the very unstable hydride arsine (AsH_3) under extreme reducing conditions in natural waters (Sergeyeva and Khodakovskiy, 1969). The geochemistry of arsenic is complex and many factors bear on the mobility and fixation of the element in the geologic environment. Although the chemical states in which arsenic may migrate in a thermal system have not been characterized, several volatile phases are known. The native metal, AsCl_3 , AsO_3 and AsS , among others are known to be relatively volatile below 300°C . The chloride is highly volatile but hydrolyzes readily in the presence of water. It may be important in acid chloride vapors or thermal waters (Boyle and Jonasson, 1973). Despite its potential to migrate in the vapor phase, Robertson and others (1978) report low As levels (.003 to .016 mg/l) in the noncondensable gas fraction of effluent from eight geothermal areas. Steam condensates from that study showed somewhat higher As levels (.001 to .09 mg/l), but the majority (.028 mg/l to 10 mg/l) remained in the flashed brine.

Arsenic forms ions of high charge (+3, +5) which are rapidly hydrolyzed in aqueous solutions and rarely exist as cationic complexes. In natural processes, it commonly forms anionic complexes and polymers, primarily with oxygen and sulfur. Where oxidation potential is low, arsenite (AsO_2)⁻¹ or thioarsenite (AsS_2)⁻¹ species would predominate. In more oxidizing environments, arsenate (AsO_4)⁻³ and thioarsenate (AsS_4)⁻³ species would be stable.

Depending on its complexed form arsenic may be mobile in both alkaline and acidic waters and, like mercury, arsenic can form sulfide complexes of high solubility. Orpiment (As_2S_3) and realgar (AsS) are common hot spring precipitates, frequently associated with mercury

minerals. The hypogene transport of arsenic is not completely characterized but is probably similar to mercury (Barnes, 1979). High arsenic concentrations in thermal waters in general have been reported from many geothermal areas (Weissberg and others, 1979; Hem, 1970b).

In the weathering cycle arsenic enters surface and ground water mainly as soluble arsenates. Thermodynamic data indicates the species $(\text{H}_2\text{AsO}_4)^-$ and $(\text{HAsO}_4)^{-2}$ are the equilibrium forms over the pH range of most natural waters, with the former dominant below pH 7.2 and the latter above (Hem, 1970b). In reducing conditions HAsO_2 may be present.

Arsenic concentrations in surface water depend on cation concentrations as well as the solubility of individual arsenate species which could exist as precipitates. While magnesium and calcium form relatively soluble arsenates, certain trace metals such as copper can greatly limit its solubility. The solubility products of various metal arsenates indicate very low solubility in terms of the anion AsO_4^{-3} (Sillen and Martell, 1964); however, this form is dominant only above pH 11.4 which is much higher than most natural waters. This effect could have some bearing in part of Dixie Valley particularly in or near the playa, where extremely alkaline conditions prevail.

Arsenic also may be taken from solution by coprecipitation/adsorption by hydrous iron oxides or adsorption on clays or organic matter. The tendency of hydrous iron oxides as well as certain iron bearing colloids to concentrate As explains why many mineral exploration soil surveys have found B horizons, the site of hydrous oxide accumulation, to be the most effective for sampling.

Klusman and others (1977) in a study of six Colorado geothermal areas found anomalous arsenic values generally outlined the same area as the mercury anomalies though not as strongly. They concluded that arsenic was not as effective as mercury in delineating areas of geothermal influence due to its closer affinity for the liquid phase.

3.1.5 Acknowledgements

This work was sponsored by the U.S. Department of Energy whose support is gratefully acknowledged. Many thanks are due to Dr. Pat L. Beaulieu, Dr. Larry T. Larson, Mr. Dennis S. McMurdie, Ms. Elaine J. Bell and Ms. Mollie A. Stewart for scientific advice, desperately needed editorial assistance, and positive encouragement throughout this project.

3.2 ANALYTICAL RESULTS

3.2.1 Mercury

3.2.1.1 Geochemical Surface

The results of the broad grid sampling of mercury are shown by three dimensional plots in Figures 3-3 and 3-4. The z axis represents the soil mercury concentration in parts per billion. This geochemical surface reveals some of the important aspects of the distribution of mercury in Dixie Valley. Most notable is the isolated high at the Dixie Comstock Mine and the prominent spike and cluster of somewhat lesser peaks between the fumaroles and Cottonwood Canyon. Both areas exhibit mineralization along the range front. Also apparent is the trend toward lower values away from the range front, closer to the playa. The geochemical surface for mercury is contoured on Plate I with sample locations and values shown.

3.2.1.2 Frequency Distribution

The frequency distribution for mercury is shown in Table 3-4 along with certain descriptive statistics. The distribution is highly skewed with most of the values smaller than the mean, but with a few extreme values being much greater indicating a more log-normally distributed population. A log-normal distribution is common for trace elements in geochemistry (Ahrens, 1954).

Deviations of the data set from log normality may be explained by the presence of a polymodal distribution. The polymodal or mixed frequency distribution can arise in various ways. In geochemistry it is common for a trace element to have a regional distribution pattern of values that may then be mingled with another population associated with mineralization which will have a markedly different frequency distribution. This would produce a bimodal distribution indicated by two peaks (modes) on the histogram of values. Where several sources or dispersion mechanisms contribute to the abundance of a given element a polymodal distribution will occur.

The possibility of a mixed frequency distribution can be examined by graphing the data on log probability paper as shown in Figure 3-5. The graph suggests the presence of three log-normally distributed populations. These consist of a set of values ranging from 0 to 20 ppb comprising approximately 15% of the samples, an intermediate group of

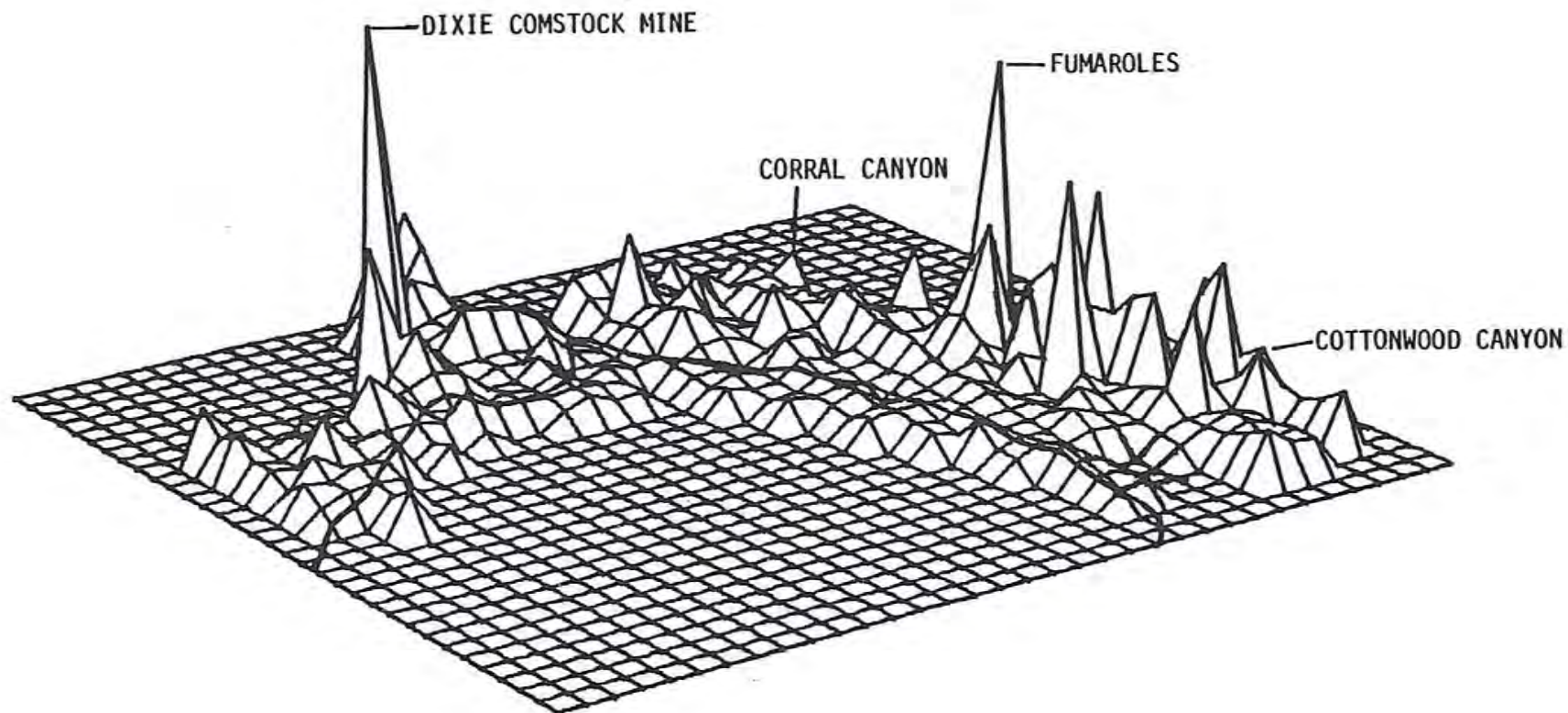


Figure 3-3. Mercury geochemical surface looking N15W.

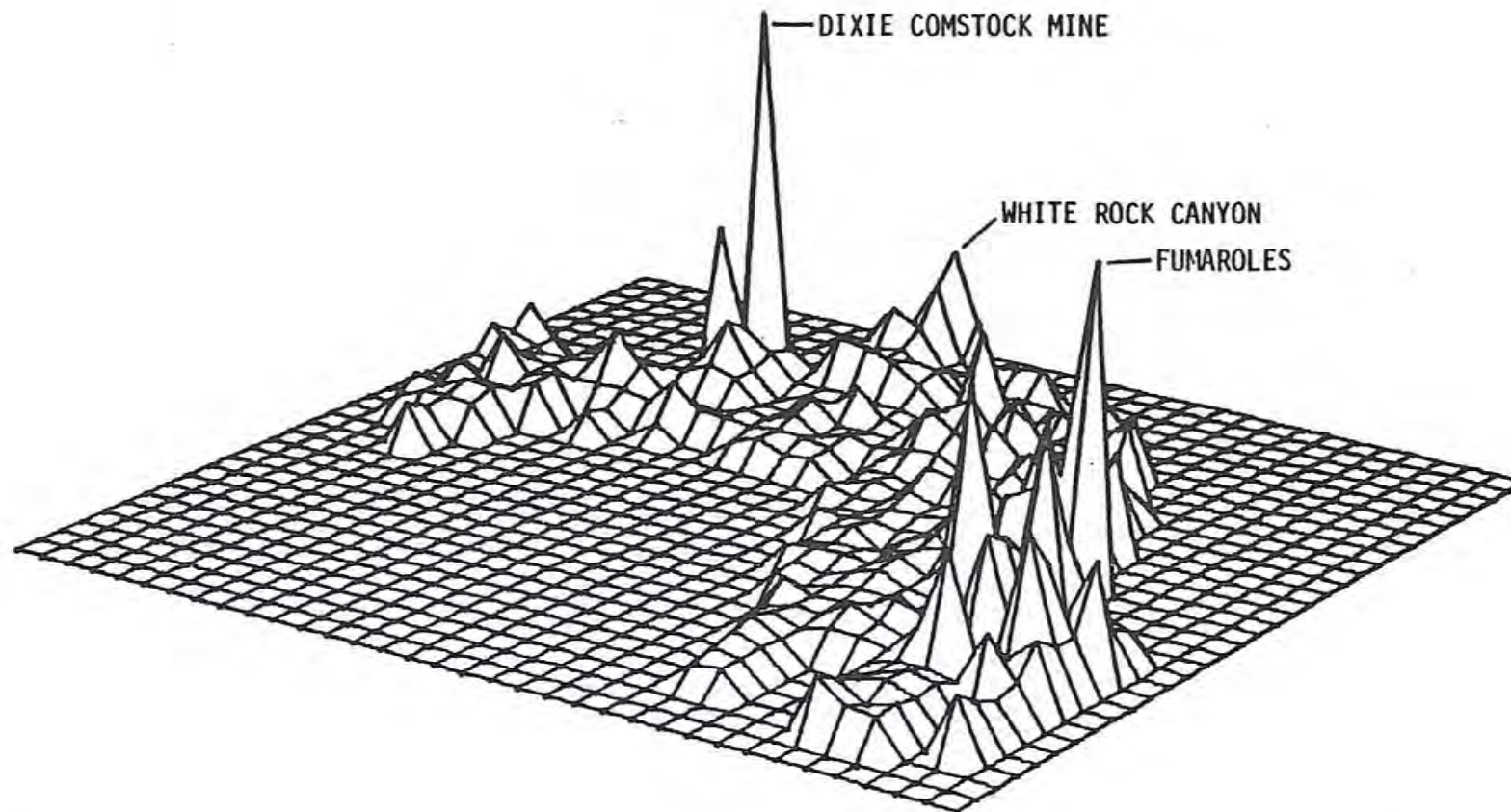


Figure 3-4. Mercury geochemical surface looking S75W.

Table 3-4
Frequency Distribution for Mercury

ppb Value	Freq.	Adj. Freq. %	Cum. Freq. %	Value	Freq.	Adj. Freq. %	Cum. Freq. %
0	2	0	0.5	144	2	0	87.4
4	6	1	1.0	148	1	0	88.2
8	13	4	4.7	152	2	0	88.7
12	14	3	8.6	156	5	1	89.8
16	16	4	12.6	160	5	1	91.0
20	6	1	14.1	164	2	0	91.5
24	18	4	18.6	168	3	1	92.3
28	22	5	24.0	176	3	1	93.0
32	20	5	29.0	180	1	0	93.3
36	16	4	32.9	188	2	0	93.8
40	28	7	39.8	204	1	0	94.1
44	10	2	42.3	212	1	0	94.3
48	19	5	47.7	220	1	0	94.6
52	6	1	48.2	232	2	0	95.1
56	24	6	54.2	248	1	0	95.3
60	10	2	56.6	256	2	0	95.8
64	13	3	59.9	272	2	0	96.3
68	13	3	63.1	284	1	0	96.6
72	11	3	65.8	288	1	0	96.8
76	8	2	67.8	296	1	0	97.0
80	7	2	69.5	364	1	0	97.3
84	7	2	71.2	376	1	0	97.5
88	9	2	73.5	388	1	0	97.8
92	4	1	74.5	392	1	0	98.0
96	8	2	76.7	400	1	0	98.3
100	3	1	77.4	480	1	0	98.5
104	7	2	79.2	488	1	0	98.7
108	8	2	81.2	560	1	0	99.0
112	2	0	81.7	976	1	0	99.3
116	6	1	83.1	1240	1	0	99.5
120	9	2	85.4	1400	1	0	99.8
124	1	0	85.7	2120	1	0	100.0
128	1	0	85.9				
132	3	1	86.6				
136	1	0	86.9				
140	2	0	87.4				

Mean	87.901	Variance	25432.764
Mode	40.000	Standard Deviation	159.477
Median	55.000	Standard Error	7.934
Minimum	0	Skewness	8.158
Maximum	2120.000	Kurtosis	84.794

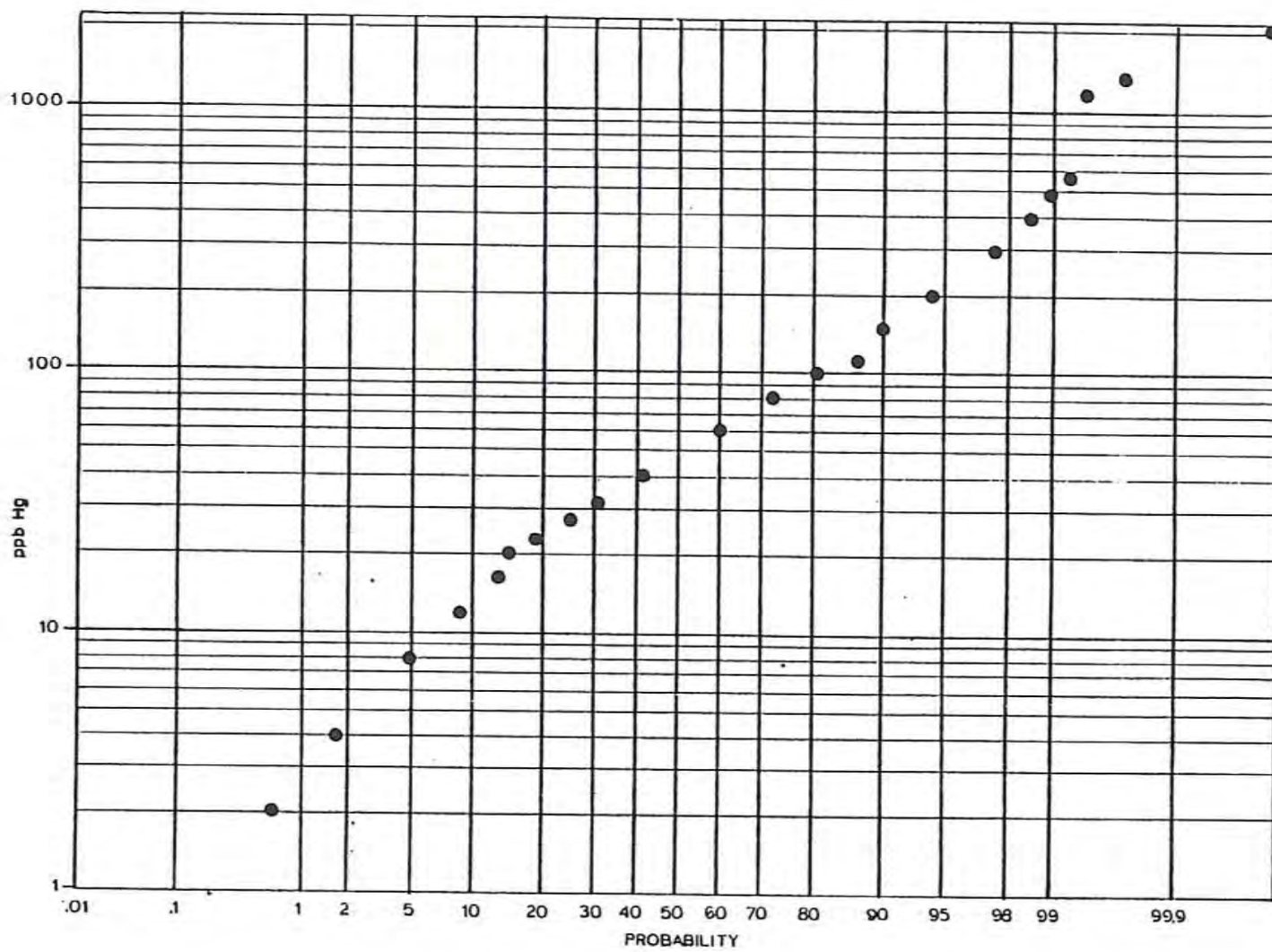


Figure 3-5. Log probability plot of broad grid mercury data.

values ranging from roughly 20 to 300 ppb making up about 83 percent of the samples and a small group of values above 300 ppb. The degree of overlap between groups is indicated by the steepness of the bend joining linear segments on the graph. A steep break between linear segments indicates a sharp demarcation, whereas a gentle break indicates considerable overlap between populations. The break between the lowest and intermediate populations is gentle and occurs over a small range of values indicating a large degree of overlap.

The group of lowest values probably represents a background population. It is comparable to background values for soil mercury in other geothermal areas as reported by Matlick and Buseck (1976) and Capauno and Bamford (1978). Four widely separated samples taken in the southern portion of the valley where there are no thermal manifestations and no nearby mineralization, all had mercury values in this range. The majority of the points within the background group are in or near the playa, which is not unexpected for several reasons. First, the playa area is away from the mineralized bedrock and mercury is not easily transported in neutral to alkaline surface waters such as those commonly found in Dixie Valley. Also the potential vaporous mobility of mercury in the playa is inhibited by the near-surface water table, by the plastic weakly permeable nature of the sediments which will tend to close potential migration conduits such as faults, and by the ability of clays below the surface to effectively adsorb large amounts of mercury.

The population of highest values on the probability plot is almost exclusively associated with areas of mineralization along the range front. This segment of the graph is not as well defined as the central segment due to far fewer points and because values in this range tend to tax the capabilities of the gold film detector making the results less reproducible. However, a distinct break from the middle segment is clear.

The well-defined central segment of mercury values on the graph is being tentatively attributed to the geothermal influence where no mineralization occurs, though other sources are possible. Motor vehicle traffic is known to account for anomalous concentrations of mercury, although anomalies would then be expected to trend along the main roads; this was not observed. Pesticides and seed are possible sources but considering the limited cultivation within the study area this too seems

unlikely. The nature of the break between the intermediate and high populations on the probability plot indicates somewhat less overlapping than observed for the low and intermediate populations. Some of the overlapping portion may represent areas where mercury previously concentrated by mineralizing processes has been remobilized by a thermal mechanism.

3.2.2 Arsenic

3.2.2.1 Geochemical Surface

The arsenic geochemical surface is depicted in Figures 3-6 and 3-7. There are some marked differences when compared to the mercury surface. The arsenic peak in the fumarole area is only moderate and the high near the Dixie Comstock is not nearly as anomalous as it is for mercury. Moreover, in contrast to the mercury values, high arsenic seems more prevalent towards the playa, particularly right at the playa margin. The N60°W trending zone of low mercury values in the area of the most mountainward advance of the playa is the site of highly anomalous arsenic values (Section 14). The geochemical surface for arsenic is contoured on Plate II with sample locations and values shown.

3.2.2.2 Frequency Distribution

Descriptive statistics for the arsenic distribution are listed in Table 3-5. Like the mercury it is roughly log-normally distributed. Figure 3-8 depicts the arsenic data using a log probability plot. The graph shows an apparent bimodal distribution with a large population ranging from 5 to approximately 35 ppm and a smaller group of values above 35 ppm. Unlike the mercury data the highest population of arsenic is not as clearly associated with mineralization. However many of the high values are associated with the drainage from mineralized areas and may be hydromorphic dispersion halos. This is consistent with the affinity of arsenic for the liquid phase, particularly in comparison to mercury. At least four relatively high arsenic values are associated with springs, and another possibly with drillhole discharge during testing. Thermal fluids from drillholes DF 45-14 and DF 66-21 showed rather high As contents (0.59 ppm and 2.1 ppm, respectively; Bohm and others, 1980) making source implications for these latter anomalies clear.

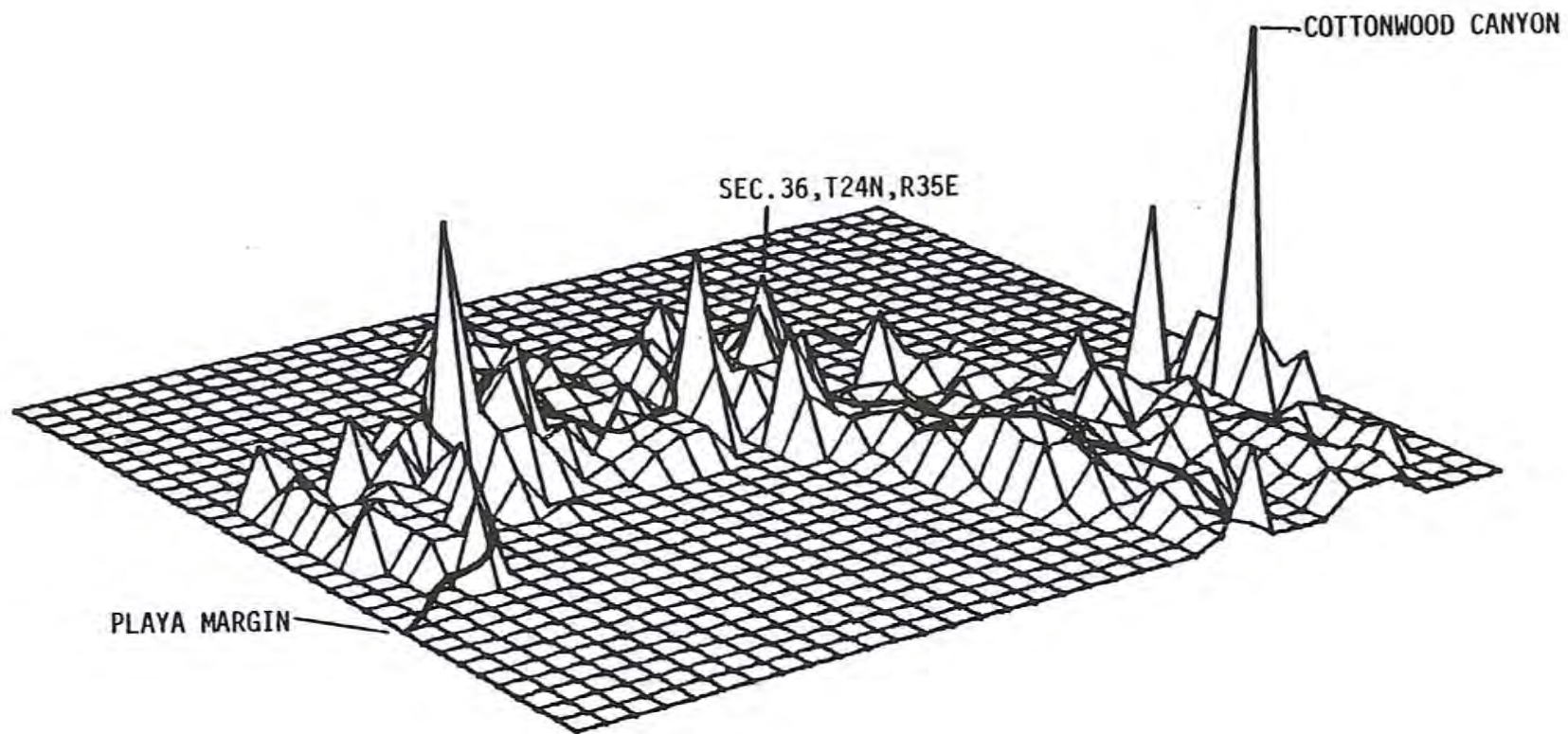


Figure 3-6. Arsenic geochemical surface looking N15W.

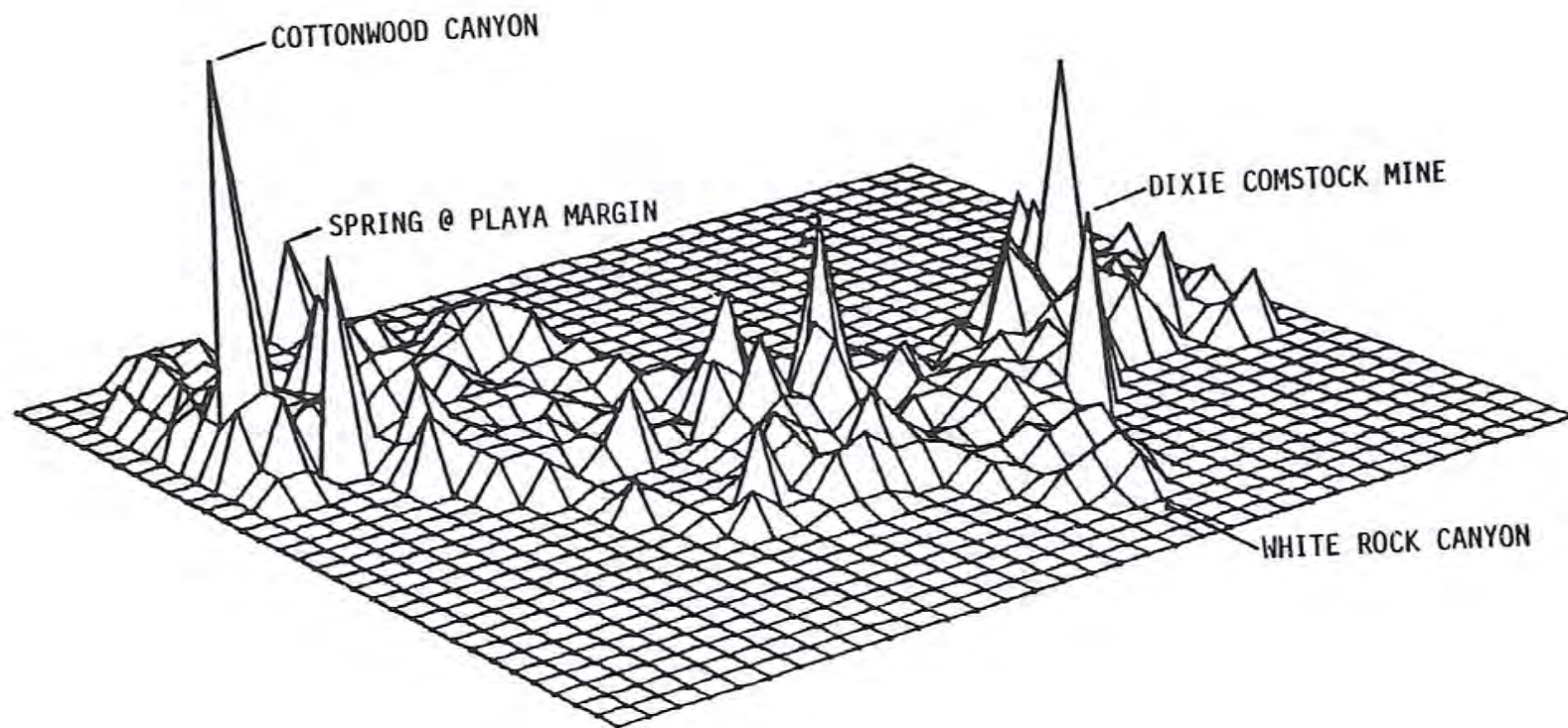


Figure 3-7. Arsenic geochemical surface looking S15E.

Table 3-5
Frequency Distribution of Arsenic

ppm Value	Freq.	Adj. Freq. %	Cum. Freq. %
1	4	.9	.9
5	46	9.8	10.7
10	139	29.7	40.4
15	137	29.3	69.7
20	34	7.3	76.9
25	45	9.6	86.5
30	16	3.4	90.0
35	21	4.5	94.4
40	9	1.9	96.4
45	5	1.1	97.4
50	3	.6	98.1
55	1	.2	98.3
60	1	.2	98.5
65	1	.2	98.7
80	2	.4	99.1
85	2	.4	99.6
115	1	.2	99.8
225	1	.2	100.0

Mean	18.010	Variance	272.794
Mode	15.000	Standard Deviation	16.516
Median	14.440	Standard Error	.822
Minimum	1.000	Skewness	6.270
Maximum	225.000	Kurtosis	65.184

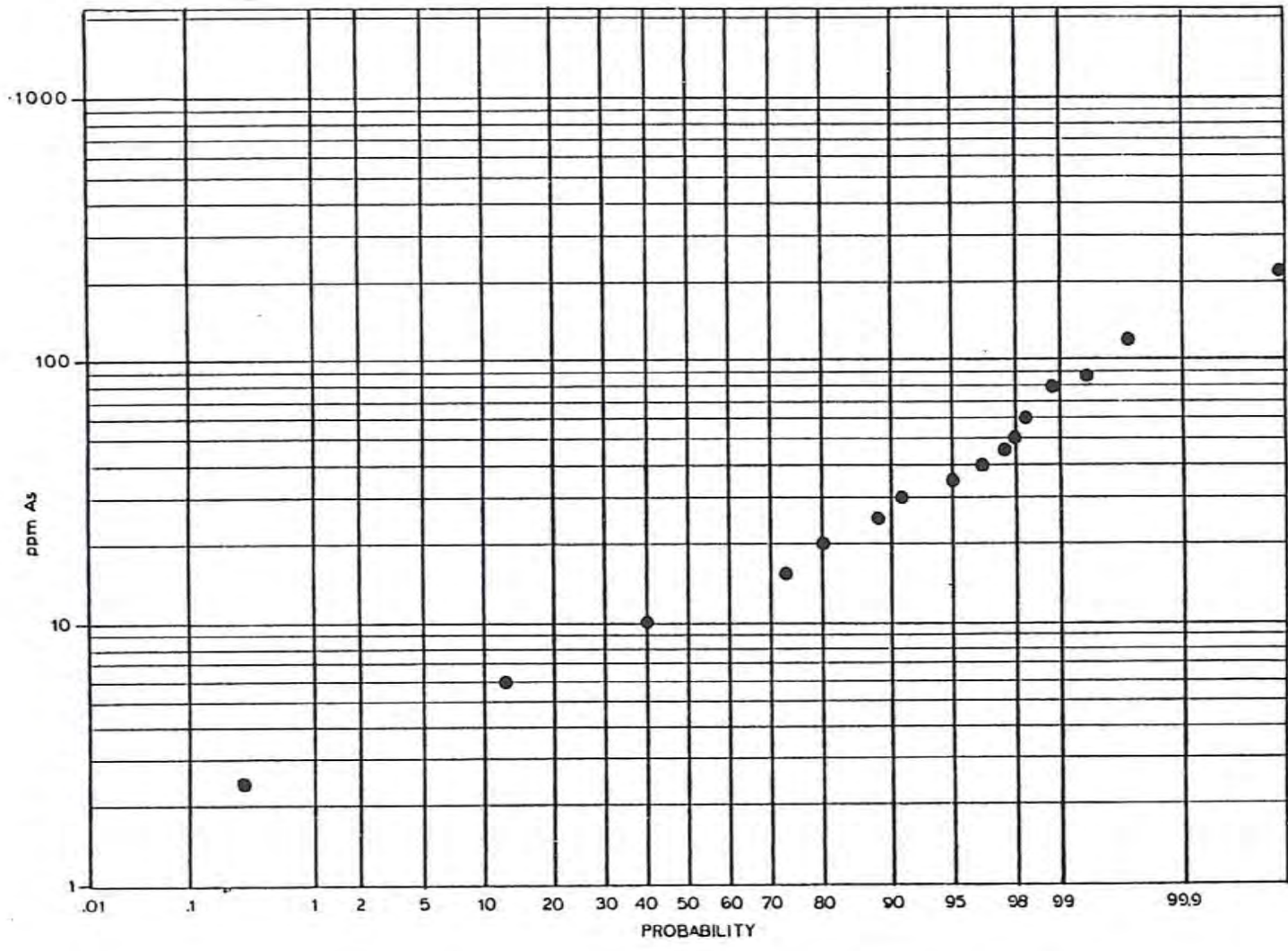


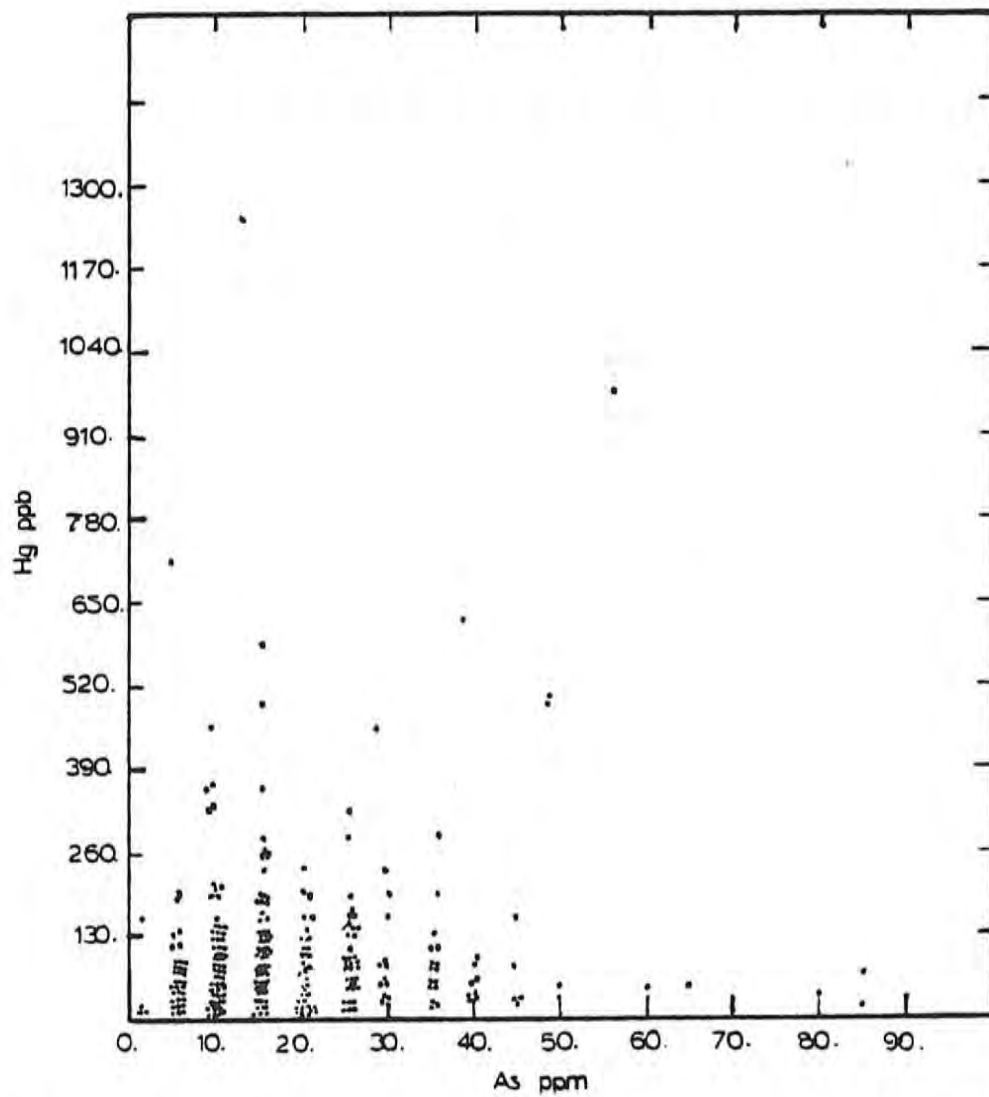
Figure 3-8. Log probability plot of broad grid arsenic data.

Plate II shows the approximate line where vegetation ends. This line also coincides closely with the site of several arsenic highs, most notably in the area south of the Dixie Comstock Mine. This may be due to a fairly abrupt chemical change in the near-surface environment. A rapid rise in alkalinity at the edge of the playa is indicated by crusts and salic horizons in the soil. This rapid pH change as surface water or near-surface ground water enters the area may cause deposition of arsenic. Perhaps a more likely source of these observed anomalies is the leakage of high arsenic waters from depth along faults at the margin of the playa. The active playa appears to be structurally controlled (Whitney, 1980), with numerous indicative geomorphic features found at its edge, including slumps, grabens and conspicuous springs, many of which show high arsenic values in the adjacent soil. This potential source of high arsenic values is discussed in more detail later.

The possibility exists that arsenic like mercury may be trimodally distributed. The sensitivity of the colorimetric method is limited to 5 ppm and it may be that a lognormally distributed background population exists in the 0 to 5 ppm range. Average arsenic values for "normal" soils from several data sources are given by Boyle and Jonasson (1973) as ranging from 3 to 12 ppm. Of four samples taken in the southern portion of the valley two showed less than 5 ppm, one 10 ppm and one 15 ppm. These results tend to support the above hypothesis. The thermal water and at least part of the nonthermal ground water of Dixie Valley is high in arsenic. Thus, the entire area may have a background population which is relatively high with no distinction between a geothermal and nonthermal source of arsenic. This possibility could be examined by reanalyzing with a more sensitive method such as hydride generation using AAS. The following discussion assumes a bimodal distribution of arsenic.

3.2.3 Correlation of Mercury and Arsenic

The correlation between the arsenic and mercury values for the grid sample locations is shown in Figure 3-9. The correlation (r) value indicates that they are very weakly correlated, with only two percent of the variation in mercury accounted for by the presence of arsenic. Generally away from the playa, particularly in areas of mineralization, the correlation is somewhat stronger but the presence of very high arsenic near the playa, probably from hydromorphic sources, obscures the



Correlation (r)	.14	Std. Error Est.	121.87	Sig. r	.001
Intercept (A)	125.51	Std. Error A	9.79	Sig. A	.0002
Slope (B)	1.50	Std. Error B	.46	Sig. B	.001
r^2	.02				

Figure 3-9. Correlation of broad grid arsenic and mercury data.

relationship. Mercury anomalies will not tend to coincide with these latter arsenic highs since it does not form strong hydromorphic anomalies, particularly in waters typical of Dixie Valley, and also the presence of vapor anomalies is limited by the nature of the playa sediments.

3.2.4 Anomalous Areas

3.2.4.1 Geochemical Thresholds

Threshold values above which a particular sample is considered anomalous have been calculated by the method of Sinclair (1974) using the 1st and 99th cumulative percentiles of partitioned populations.

The mercury distribution has been partitioned into three separate populations in Figure 3-10. For the present purposes of this study the discrimination of the background and intermediate geothermally influenced populations is most important since the highest population is very small and can also be indicative of geothermal manifestations. Thresholds between the two lower populations are indicated at 10 and 80 ppb. The set of mercury values above 80 ppb, referred to as high anomalous, contain approximately 35% of the anomalous values attributed to geothermal influences and will be almost entirely from the anomalous group.

Values in the range of 10 to 80 ppb, referred to as low anomalous, contain approximately 64% of the anomalous population and will be comprised of about 90% samples from the anomalous population. Values below 10 ppm are considered background.

Figure 3-11 shows the arsenic data partitioned into two separate log-normal populations. Thresholds corresponding to the 99th cumulative percentile of the lower population and the 1st cumulative percentile of the higher population divides the data into three groups. Values above 47 ppm are referred to as high anomalous. Eighty five percent of the anomalous population falls into this group and approximately 64% of the samples in this group are from the anomalous population. Values between 30 and 47 ppm are referred to as low anomalous. Approximately 14% of the anomalous values fall in this range and about 7% of the values in this group are from the anomalous population. Values below 30 ppm are considered background, but may contain 1% of the anomalous population.

Although the definition of threshold as used here is arbitrary it does provide a measure of the likelihood that a value has arisen due to

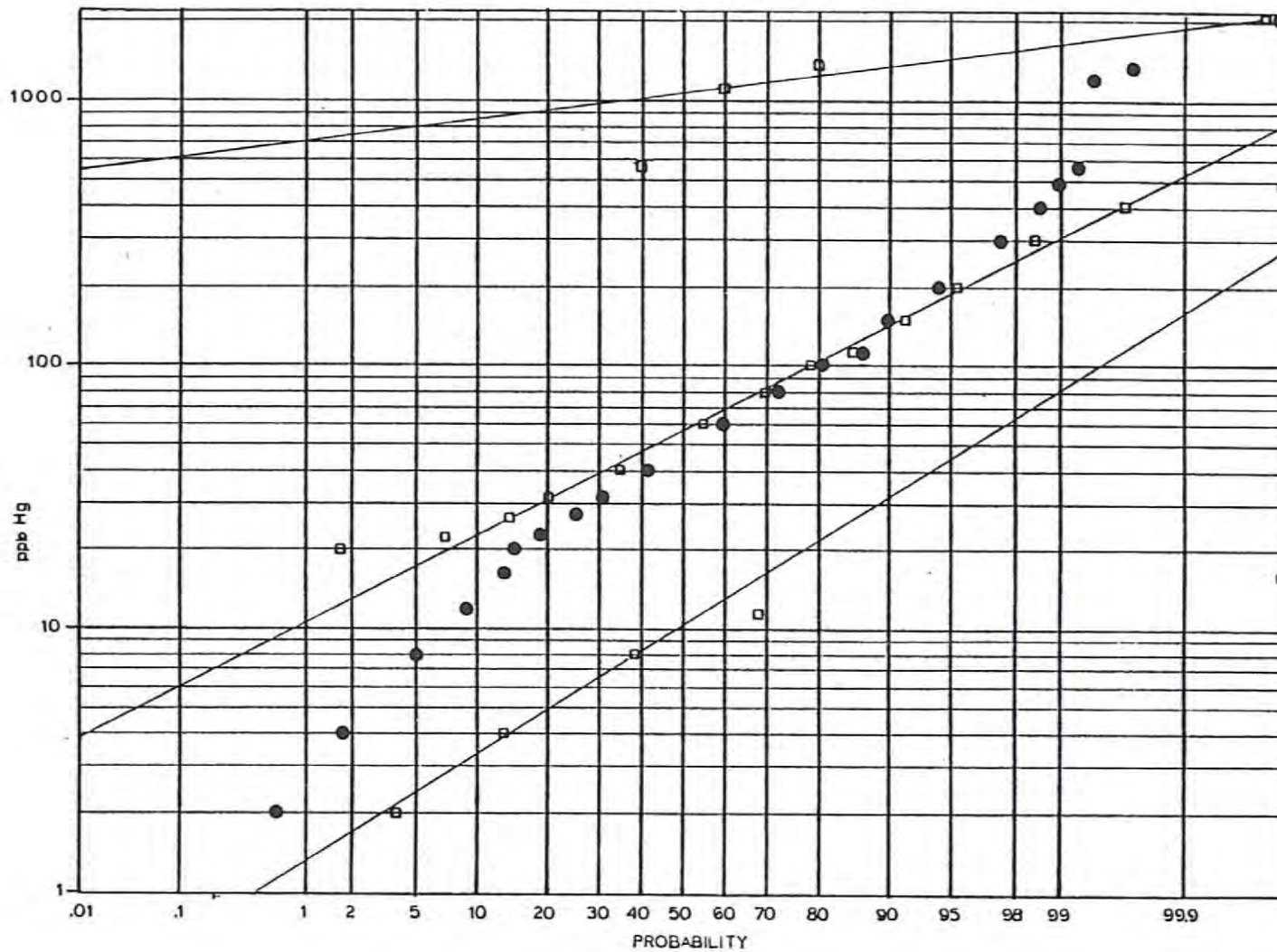


Figure 3-10. Partitioning of broad grid mercury data into three separate log-normal populations.

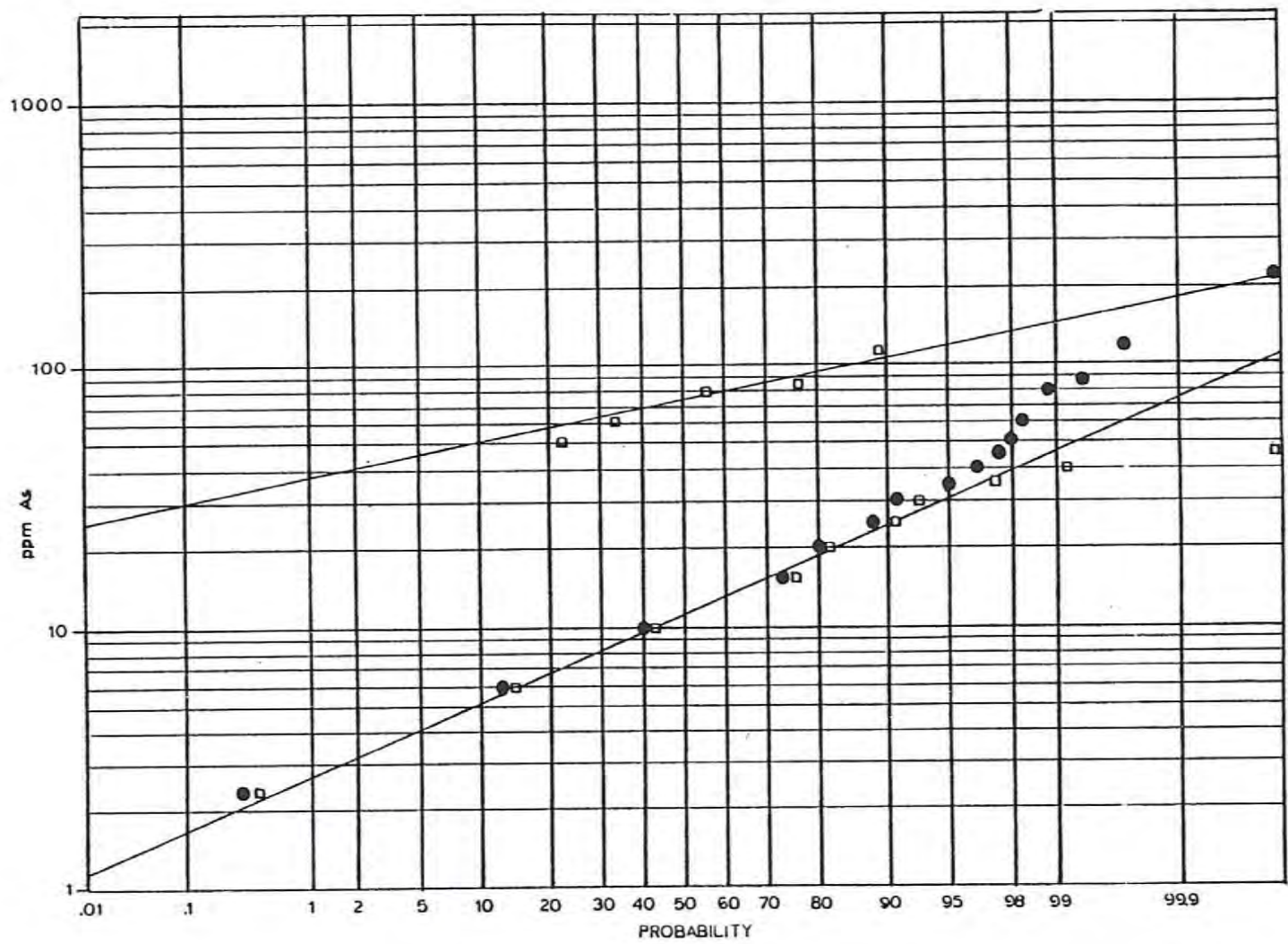


Figure 3-11. Partitioning of broad grid arsenic data into two separate log-normal populations.

a given source or mechanism. Also it serves to divide the data into priority groups for follow-up work. This is particularly important with regard to the mercury data where a majority of the samples fall in the 'geothermally influenced' population which overlaps significantly with the background population.

Much of the following discussion is based on Plates I through III. It must be noted that where isolated high values occur, contours around them may be misleading since the natural variability of the data can account for occasional extreme values.

3.2.4.2 Fumaroles and Corral Canyon

The highest mercury values are generally associated with mineral occurrences along the range front fault. The source of this mercury may be mechanically dispersed material from the deposits themselves with a geothermally related source also present. This seems likely in the area of the fumaroles in Section 15, T24N, R36E. The steam vents occur along the range front at an old mine site where two adits go into the hillside and a tailings dump is evident. Although it is not specifically mentioned in the literature the structural setting is similar to Corral Canyon where gold is found along the contact of albite/calcite dikes within the gabbroic country rock (Willden and Speed, 1974).

Extensive alteration and formation of red ferruginous clay has occurred adjacent to the ten fumaroles in the area and one of the adits is quite warm. Mercury values along the fumarole chain are much higher than those in the vicinity of the tailings dump at Corral Canyon. This suggests that the mercury is primarily related to the geothermal heat source, although the heat source could be mobilizing mercury which has already been preferentially concentrated by mineralizing processes. The size of the mercury halo around the fumaroles compared with Corral Canyon probably reflects the increased mobility of mercury in the vapor phase. Arsenic values at Corral Canyon and the fumaroles are similar, in the range of 25 to 35 ppm. High values of arsenic also exist down the fan slope from both of these areas, which is suggestive of a mechanically dispersed halo. Alternatively the geochemical distribution near the fumaroles, particularly for mercury, may be due to a north to northwest trending structure. While no obvious structure appears to intersect the range front at the fumaroles, it may be significant that

to the north the Senator fumaroles occur near the mapped intersection of the trace of the northern splay of the Buckbrush fault and the range front.

3.2.4.3 Dixie Comstock Mine

Although no fumaroles are present at the Dixie Comstock Mine, high heat flow and hot water very close to the surface are known. Very high anomalous mercury values may be due at least in part to the thermal influence. It should be noted that this mine is a considerably larger operation than the two small adits in the fumarole area, its tailings dump is many times larger, and at one time an amalgamation mill existed on the site. Therefore potential mercury anomalies associated with the mineralization and subsequent mining operation would be expected to be far more obvious here.

3.2.4.4 Cottonwood Canyon

Isolated high As and Hg values occur near the mouth of Cottonwood Canyon which contains known mineral deposits. The highest value near Cottonwood Canyon occurs in the area where drainage has been diverted to the Boyer Ranch reservoir and which was not affected by the flash flooding of July, 1979. Areas that were affected by the flooding generally showed background or low anomalous values of mercury and arsenic. The significance of this is not clear, although it may be that the debris of the flood contains most of its metal values in the larger grain sizes and/or does not possess an horizon of metal accumulation such as the typical soil. Because a standard sampling depth was employed only the flood debris and not the underlying soil was sampled. The thickness of the flood deposited material was too great to allow sampling of the underlying soil to test the above hypothesis. However, samples taken from the area prior to the flooding were analyzed for mercury by Southland Royalty Company and the values ran significantly higher. The arsenic value corresponding to the highest mercury was the largest recorded in the study. This may be due to the previously mentioned arsenide and sulfarsenide ores found three miles up the canyon, although adjacent values in the flood deposits from the canyon were at background levels.

3.2.4.5 White Rock Canyon

Several anomalous geochemical values occur near the mouth of White Rock Canyon. A high of 373 ppb mercury appears to be part of a somewhat linear trend of high anomalous values which is suggestive of a structurally controlled feature. The peak values of arsenic and mercury in the area coincide, although the arsenic values are generally low anomalous or background.

Although mineralization is known in the upper portions of the canyon, it is not extensive. A sample taken approximately 305 m (1000 ft) into the canyon where soil had formed on an older stream terrace, showed values of only 40 ppb mercury and 10 ppm arsenic. Several lineaments mapped in this area from low-sun-angle aerial photography (Whitney, 1980) are mimicked by the mercury contour trend. The White Rock Canyon fault, a major cross cutting structure in the valley intersects the range front Stillwater fault approximately at the mouth of the canyon (Plate III). A temperature gradient hole (H-2 on Plate III) drilled to a depth of 152 m (500 ft) near the mouth of the canyon showed a relatively low thermal gradient. A negative gradient is observed to a depth of 20 m (65 ft) before an approximately $2.5^{\circ}\text{C}/100\text{ m}$ ($1.4^{\circ}\text{F}/100\text{ ft}$) gradient is established to TD. It seems likely that shallow ground water flow is responsible for the negative temperature gradient in the upper 20 m (65 ft) east of the range front fault. The Stillwater fault is indicated on the drill logs by a zone of lost circulation which coincides with the gradient reversal. One meter temperature measurements in the area are inconclusive being neither significantly lower nor higher than adjacent areas. These results seem to indicate that any near-surface thermal effects along and just to the east of the range front are largely overwhelmed by the mountain front ground water recharge.

The possibility of a structure or structures further east of the range front which might conduct or previously conducted thermal fluids was investigated by sampling at 30 m (100 ft) intervals along a $\text{N}60^{\circ}\text{W}$ traverse (line B-B'; Plate III) approximately 500 m (1640 ft) south of White Rock Canyon. The results of this profile are shown in Figure 3-12.

A significant zone of high anomalous mercury values exists along the traverse which may be due to upward migration from a splay of one of the faults to the west or perhaps may be related to one of several

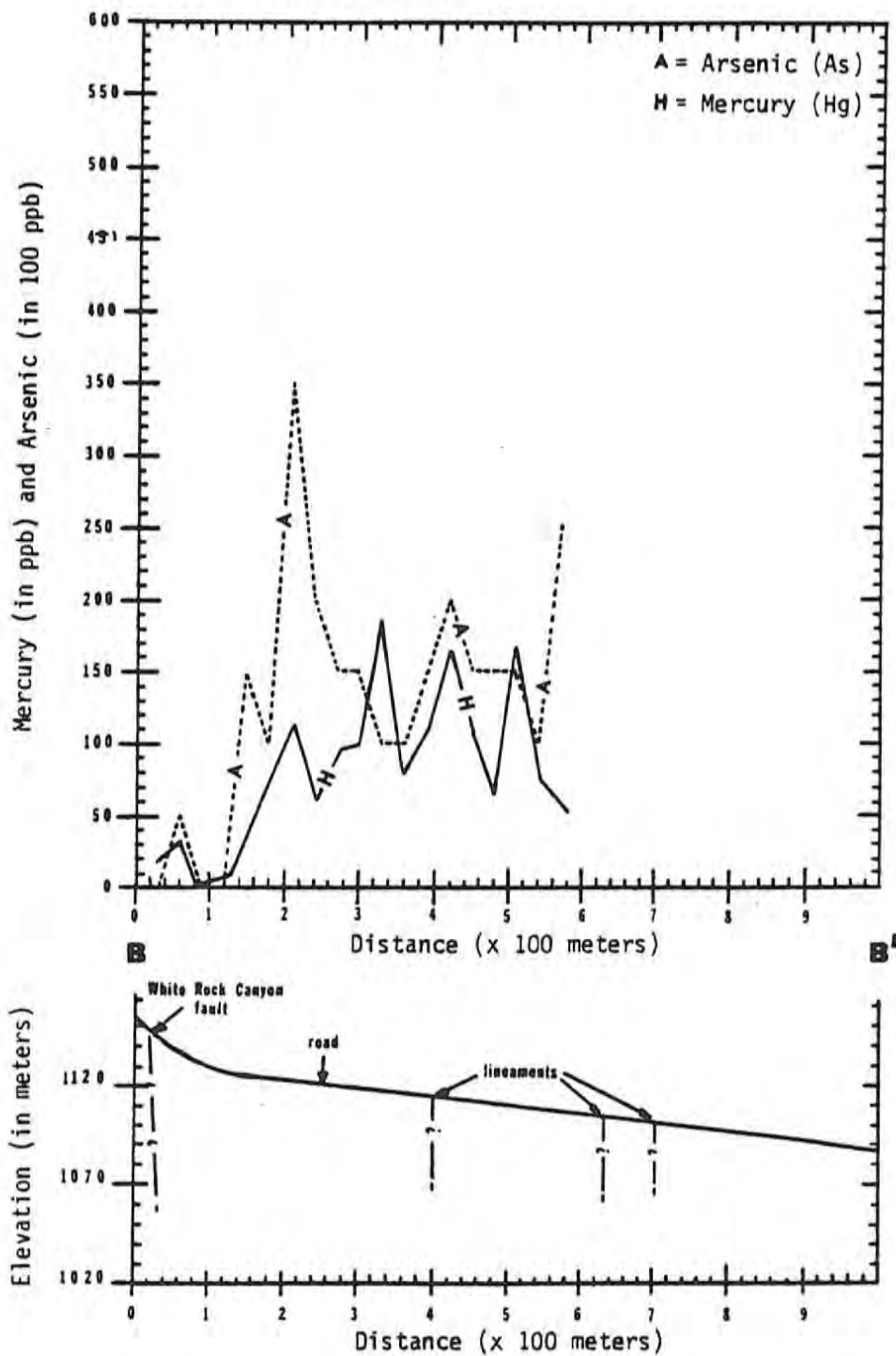


Figure 3-12. Profile of mercury and arsenic along traverse B-B' south of White Rock Canyon (see Plate III).

lineaments in the area representing older structures which are serving as paths for geochemical migration. It is conceivable that at one time the Stillwater fault broke across the fan emerging from White Rock Canyon and that these lineaments represent this older structure.

The correlation coefficient for the mercury and arsenic values along traverse B-B' is $-.07$ indicating the two are probably unrelated. However, the zone of highest arsenic occurred in an area where a well-developed, finely textured, damp soil existed. To the east this area changes abruptly into the gravelly, poorly developed soil typical of the upper portions of the fans. This transition could bear on the arsenic-mercury observations in several ways. First, a locally high water table, perched by faults, may promote more rapid soil development with the arsenic values associated with the seepage and mercury migration being suppressed by the water. Or second, the confluence of drainage from the several canyons may affect the local heat flow patterns with arsenic deposited at the range front break in slope. The mobility of mercury would be inhibited in areas of ground water influx while arsenic would have a greater tendency to be associated with areas of high water table. The abrupt change from coarse to fine material with a localized water condition is highly suggestive of a structural discontinuity. Possibly the arsenic is associated with the water bearing fault zone at the surface and mercury migrates through splays or more permeable zones just east of the fault. While largely speculative, a structurally controlled source for the high mercury values seems feasible.

3.2.4.6 Section 36, T24N, R35E

Another area where geochemical trends indicate possible structural controls occurs in the central part of the study area (Section 36, T24N, R35E) where the playa most closely approaches the range front, coming within approximately 610 m (2000 ft). A spring occurs at the base of the steep talus slope which joins the mountain front and playa. The valley drainage is also displaced closer to the western range front in this area so that streams flow nearly east-west toward this area before turning south toward the Humboldt Salt Marsh. This topographic low shows rather low mercury values similar to most other areas of fine-grained sediments. The area is, however, the site of a linear trend of high anomalous arsenic values. This arsenic could be related to the

spring issuing from the range front, or it may be due to water migrating up from deeper zones along permeable structures. This latter possibility is supported by the convergence of several of the major fault systems in the area: the Stillwater, Marsh and Dixie Meadows faults (Plate III). Additionally the southernmost, probably southeast trending, margin of the Humboldt gabbroic complex probably occurs very near this point in the valley. An area of abnormal aeromagnetic gradients also terminates at or near this area and may be related to the gabbro.

3.2.4.7 Dixie Meadows Fault (Section 32, T24N, R36E)

The Dixie Meadows fault describes an arcuate path to the northeast from Section 36 (T24N, R36E). In the southeast corner of Section 32 (T24N, R36E) an area of anomalous temperatures measured at one meter depth (Campana and others, 1980) occurs near the fault trace. This area is near the 455 m (1500 ft) gradient hole drilled by Southland Royalty Company in 1978 which showed an average gradient of approximately $7.5^{\circ}\text{C}/100\text{ m}$ ($4.1^{\circ}\text{F}/100\text{ ft}$). Both arsenic and mercury show high anomalous values in this area with fairly linear northwesterly trends (Plates I and II) near the postulated trace of the Dixie Meadows fault. Follow-up sampling was done in this area at 30 m (100 ft) intervals along a traverse (Line A-A'; Plate III) transecting the Dixie Meadows fault; the results are shown in Figure 3-13. A dominant mercury peak appears to be spatially associated with the approximate trace of the Dixie Meadows fault. Arsenic values tend to mimic the mercury particularly west of the fault trace. While indicated by only three points on the traverse mercury values appear to fall off rather rapidly to background levels east of the fault trace. This may be due to the presence of finer-grained sediments to the east. The correlation between arsenic and mercury is very weak for this traverse ($r = -0.02$) and, as at White Rock Canyon (Line B-B'; Plate III), may be due to the effects of hydromorphic processes upon arsenic.

The fault zone in this area may be relatively wide thus accounting for the rather broad band of anomalous mercury. This is supported by a wide area of lineaments and other fault related features identified on low-sun-angle aerial photography (Whitney, 1980). While structural control of the observed geochemical trend across profile A-A' is likely,

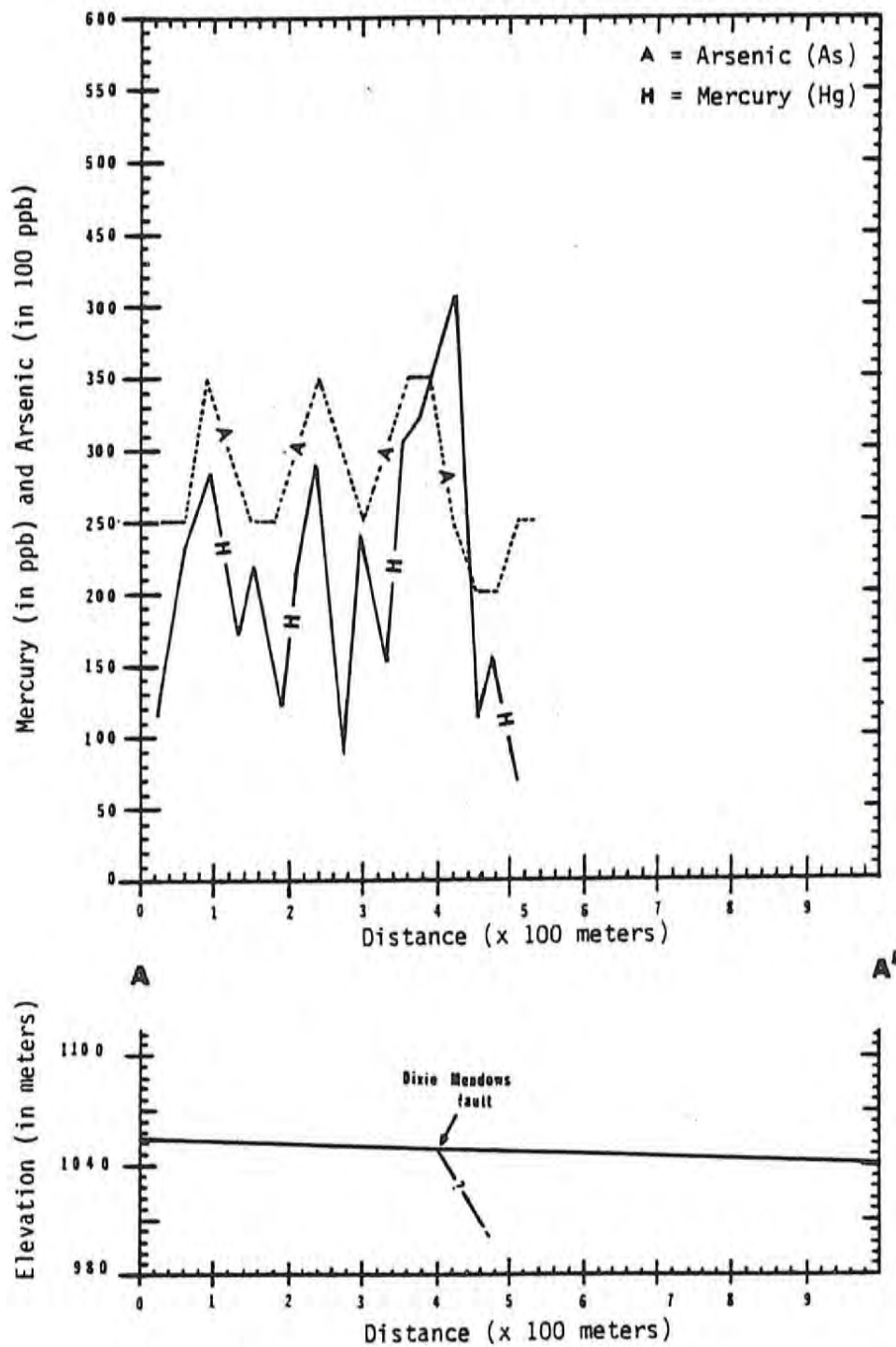


Figure 3-13. Profile of mercury and arsenic along traverse A-A', transecting the Dixie Meadows fault (see Plate III).

the effects of nearby mineralization may be at least partly responsible. The geochemical values, particularly arsenic, may be influenced by drainage from the Stillwater Range. The drainages of several mineralized zones along the range front fault flow through this area. However, considering the apparent structural interactions in this area, it seems more likely arsenic is coming from deeper water migrating upward along faults.

3.2.4.8 Dixie Meadows Fault (Section 14, T24N, R36E)

Further north along the trend of the Dixie Meadows fault (Plate III) are several coincident geochemical trends. Most notable is the linear zone of high mercury in and around Section 14 (T24N, R36E). Arsenic trends in the area are similar but not coincident. At the intersection of the Dixie Meadows fault trace and the range front is an extensively altered area, with high anomalous mercury and arsenic values, which appears to be an old landslide block. This strong correlation between the geochemical and structural trend is significant, although nearby range front mineralization may also contribute to the anomalous values.

Approximately 2 km (1.25 mi) southwest of this intersection in Section 15 is the chain of fumaroles previously discussed. Whether these two zones of high geochemical values indicate a structural relationship is not clear.

3.2.4.9 Buckbrush Fault Trace

As shown on Plate III, the southern splay of the Buckbrush fault appears to intersect the range front in the same area as the Dixie Meadows fault. A very conspicuous line of springs occurs along the fault trace in Section 26 (T24N, R36E) and is accompanied by low anomalous arsenic and mercury trends. The springs have not been sampled for arsenic but the possibility of communication with high arsenic thermal water exists. A similarity between the waters of the Buckbrush Seeps which occur along the Buckbrush fault to the south, and the Hyder Hot Springs to the north has been noted by Bohm and others (1980). Buckbrush Seeps have not been analyzed for arsenic but Hyder showed some of the highest values noted for surface water in the valley, approximately 30 ppb.

Another possible source of the arsenic is streams draining the Bernice District in the Clan Alpine Range. Both Spring and Shoshone

Creeks drain the area of the Red Bird mercury mine and could contribute anomalous amounts of arsenic and/or mercury.

The mercury values associated with the Buckbrush fault trace are generally low but the higher values are among the low anomalous group. Given the nature of the sediments and mercury values associated with the playa elsewhere, the areas above the 40 ppb contour along the Buckbrush fault trace must be considered anomalous. The path of this fault trace becomes less clear to the north due to cultural modifications to the land surface, but several other geochemical highs are associated with the postulated fault trace. While it is not clear on the contour map (Plate I), point 373 which is approaching the fault trace is a high anomalous mercury value. Profile C-C', discussed in more detail below, shows progressively higher values towards the fault trace.

3.2.4.10 Sun Oil Company Wells

Observed coincident geochemical and structural trends discussed above are significant in relation to the deep exploratory wells which have been drilled by Sun Oil Company in this area. At least three of the four wells are producers, with the depth of production from approximately 2200 to 2900 m (7200 to 9500 ft).

No obvious geochemical anomalies were identified near these wells by the 305 m (1000 ft) sampling. One potential reason for the generally low to moderate mercury values is the presence of very fine-grained sediment. Although the wells are considerably north of the furthest extent of the playa at present, the surface is quite likely underlain by older fine-grained playa material.

Since the Sun Oil Company Wells are proven producers and since a splay of the Buckbrush fault lies to the east, it was decided to profile across the zone of drilling at 30 m intervals (line C-C'; Plate III). The results of the traverse are shown in Figure 3-14. The correlation of arsenic and mercury along the traverse is stronger here ($r = .38$) than in other areas but is still rather weak. Similar to other areas in the valley, this may be due to a difference in sources or dispersion mechanisms for these two elements. The arsenic and mercury values become consistently higher to the east near the Buckbrush fault trace. Whether the fault, the geochemical values, and producing wells are related is not clear but comparison to nonproducing wells DF 45-14

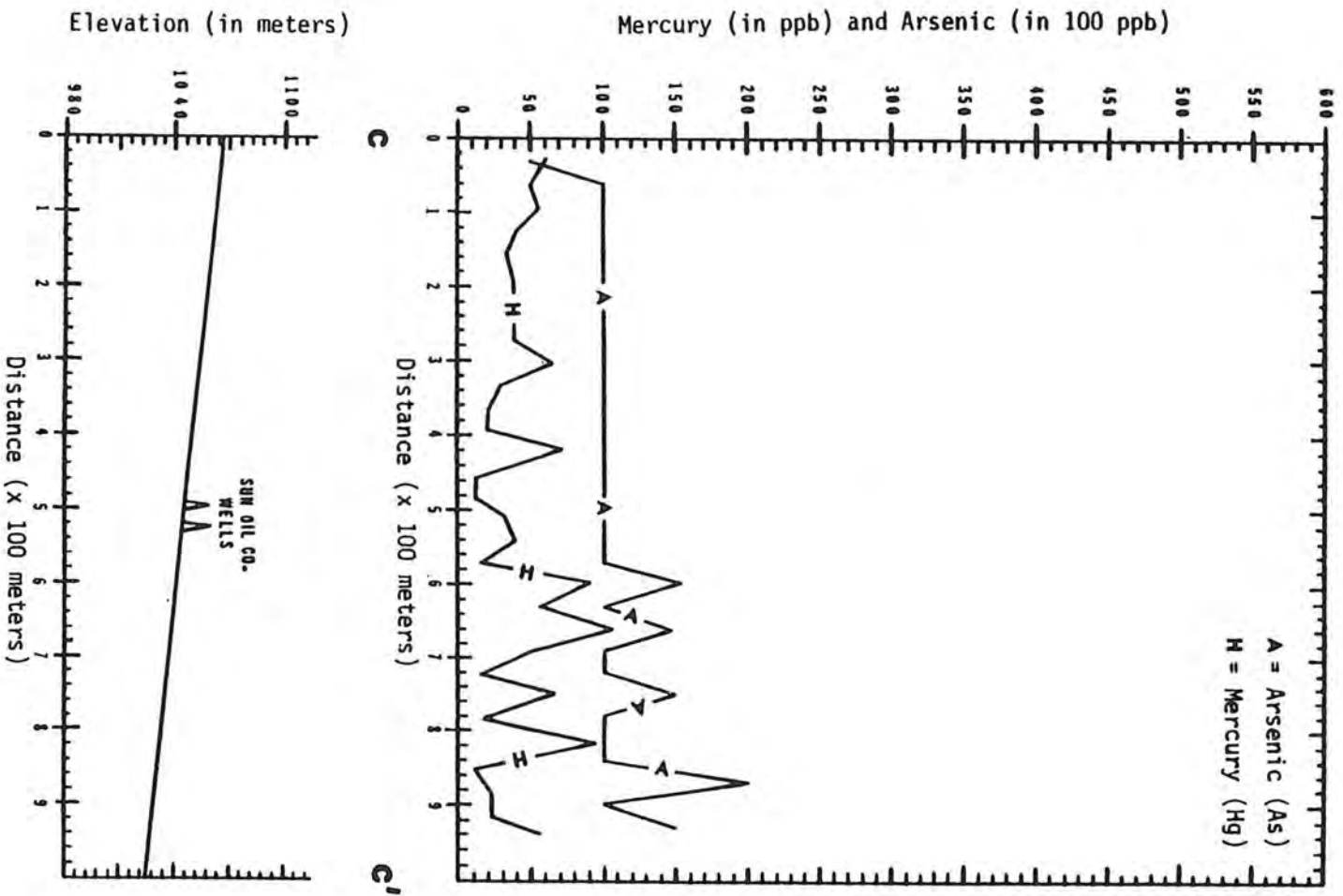


Figure 3-14. Profile of mercury and arsenic along traverse C-C' north of Boyer Ranch (see Plate III).

and DF 66-21 argues favorably for it.

3.2.4.11 DF 66-21

A square grid of points spaced at 183 m (600 ft) has been sampled around well DF 66-21 as shown in Figure 3-15 with geochemical contours imposed upon it. It can be seen that the values around the well site are considerably higher in places than any detected in the area by the broad grid sampling as shown on Plate I. This may be due to disturbance of the area during drilling and testing, an effect which has been inferred for other geothermal areas (Capauno and Bamford, 1978).

No particularly significant trends occur around the well, which is not unexpected since no thermal fluids were encountered near the surface here nor were any structures mapped (Whitney, 1980). The drillhole did penetrate zones of high temperature fluid but flow rates were generally low. Zones of thermal activity encountered in DF 66-21 would be expected to correlate with structures to the west, most notably the Stillwater fault. Where a zone of high pressure hot water found just below 1430 m (4700 ft) may connect to the surface is unclear. If the zone at 1430 m were postulated to be a basin and range type structure of 55° to 60° dip, its trace would coincide with an area of lineaments which have been mapped approximately 1000 m (3300 ft) west of DF 66-21 (Whitney, 1980). Perhaps significantly the trend of these lineaments intersects the range front at the fumaroles. Similarities between downhole mineralogy and mercury content for cuttings from the well (Chapter 4, this report) and samples taken from the fumarole chain to the northwest also suggest some association. No significant geochemical anomalies occur along the Stillwater fault in this part of the valley which may mean it has been sealed to some extent by mineral deposition. This is an area of the fault which has apparently been unaffected by the most recent tectonic events.

It also appears that DF 66-21 has been drilled west of a major structure that may border or communicate more directly with the geothermal reservoir which could account for the low flow rates. This is suggested by the coincident geochemical and geomorphic trends to the east along the Dixie Meadows and Buckbrush faults.

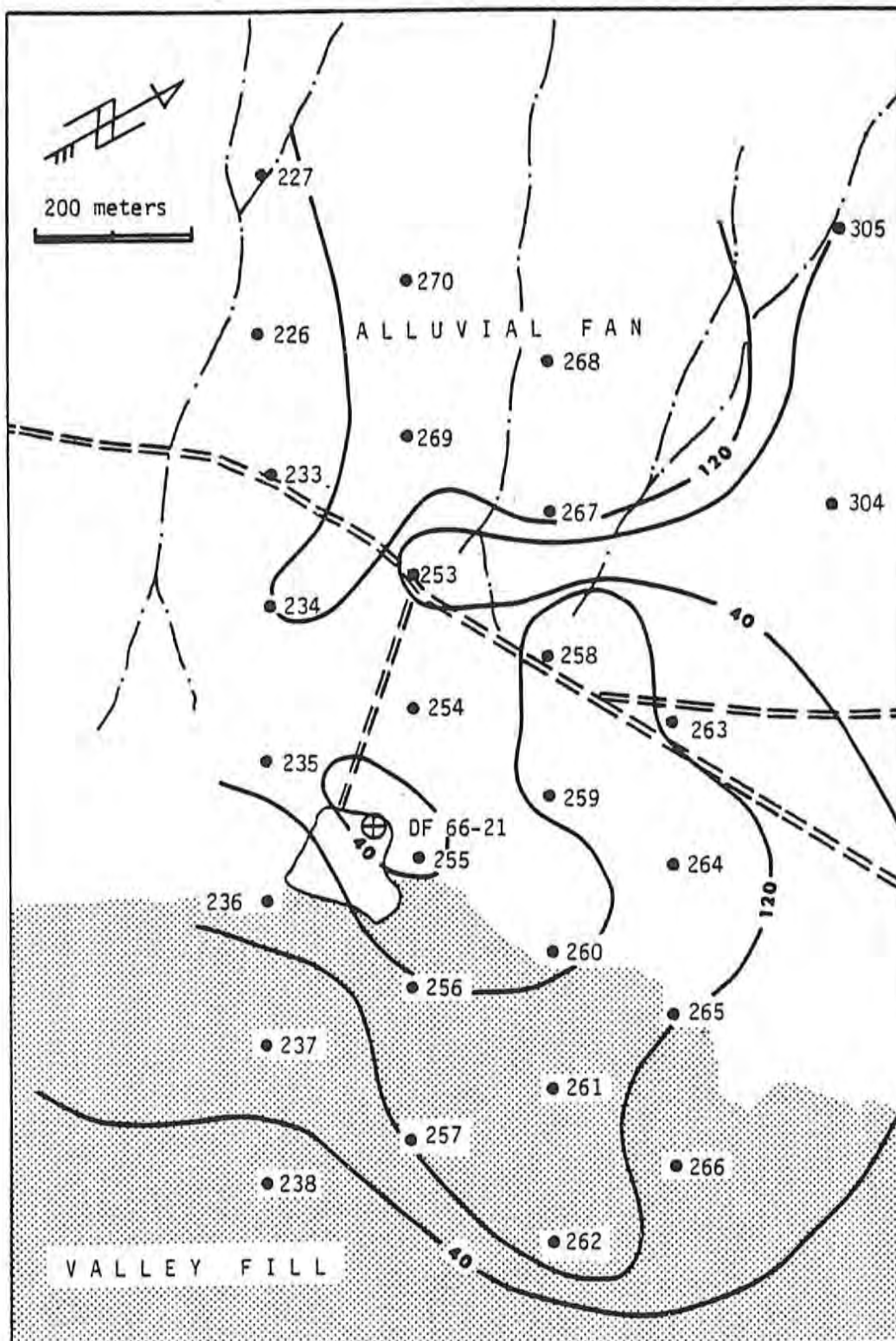


Figure 3-15a. Contoured mercury geochemical surface in vicinity of DF 66-21. Values indicated by contours; sampling sites indicated by solid circles; extent of geomorphic surfaces indicated by shading.

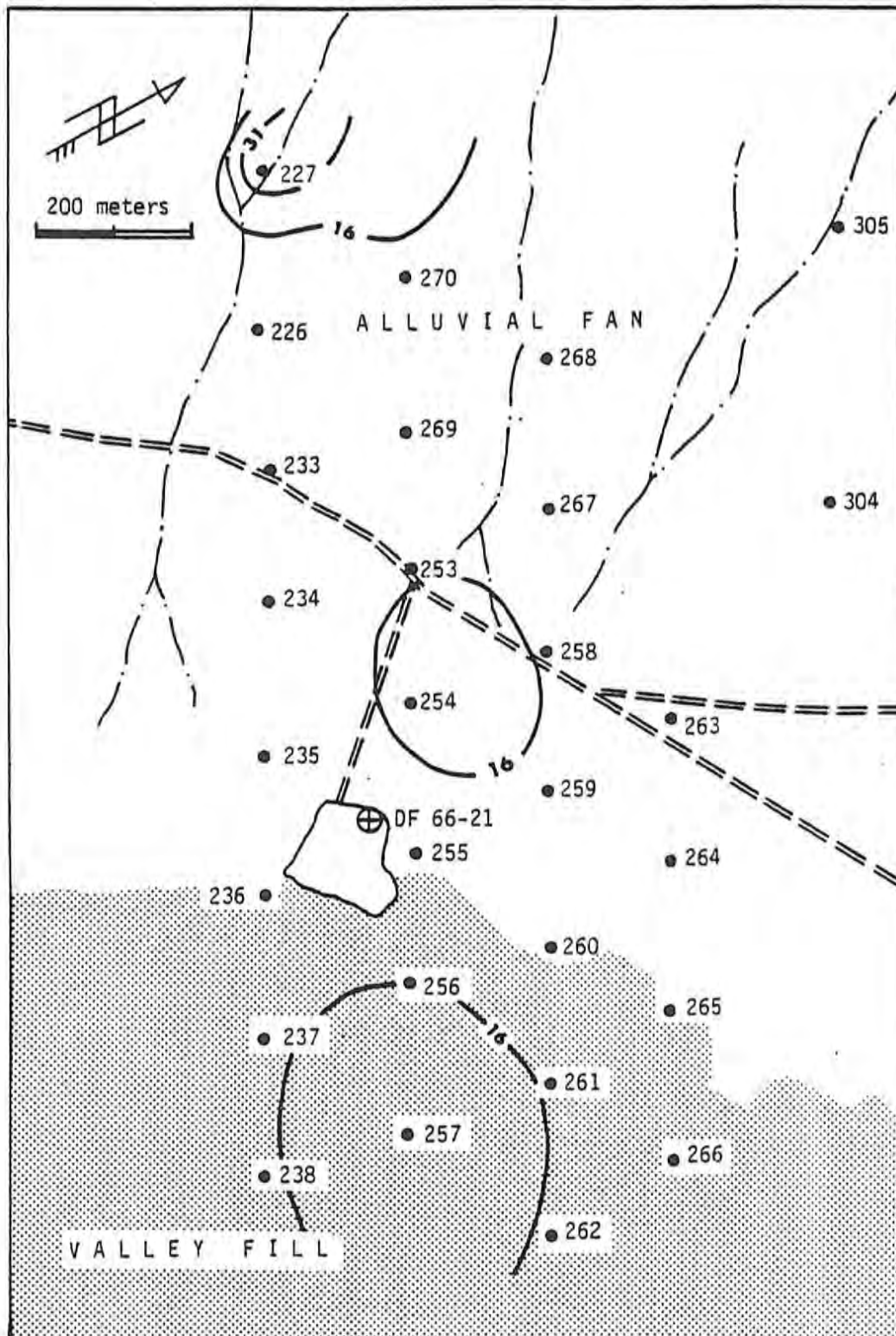


Figure 3-15b. Contoured arsenic geochemical surface in vicinity of DF 66-21. Values indicated by contours; sampling sites indicated by solid circles; extent of geomorphic surfaces indicated by shading.

3.2.4.12 DF 45-14

The results of the closely spaced sampling around DF 45-14 are shown by geochemical contours in Figure 3-16. Interpretation of the data is complicated by the presence of the Dixie Comstock Mine with its large associated mercury and arsenic values.

Away from the range front a rather obvious scarp occurs and is mimicked by the mercury trend. The significance of this structure is unclear but it may be an older splay of the Stillwater fault. Several structures appear to run through and to the east of the well site, and are associated with high anomalous arsenic and mercury values. The mercury values drop off quickly to the east, coincident with the occurrence of fine-grained sediments.

Similar to Section 36 (T24N, R35E) the playa comes quite near the range front in the vicinity of DF 45-14. High arsenic values at this site may be due to the convergence of structures with arsenic migrating upward. The graben, scarp and slump features east of the well are along the trace of the Dixie Meadows fault and are typically associated with profuse growth of rabbit brush, indicative of a higher than normal water table.

Any zones of fluid migration within DF 45-14 would be expected to show geochemical manifestations to the west since, like DF 66-21 to the north, it is west of the inner graben bounding faults that may be more directly connected to the geothermal reservoir. Like DF 66-21, well DF 45-14 hit high temperature fluids but flow rates have been low. This may indicate some sealing effects creating lower permeabilities (see Chapter 4; this report).

Some contamination may have occurred during flow tests of DF 45-14. The site of the highest arsenic values occurs directly downwind of the well where it may have been affected by wind carried effluent. However, the area around DF 66-21 does not exhibit high arsenic values even though its fluids contained significantly higher concentrations of the element.

3.3 SUMMARY AND CONCLUSIONS

While the soil geochemical values for arsenic and mercury have several sources of variation in Dixie Valley, it appears these sources may be identified. The Dixie Meadows fault shows numerous zones of high

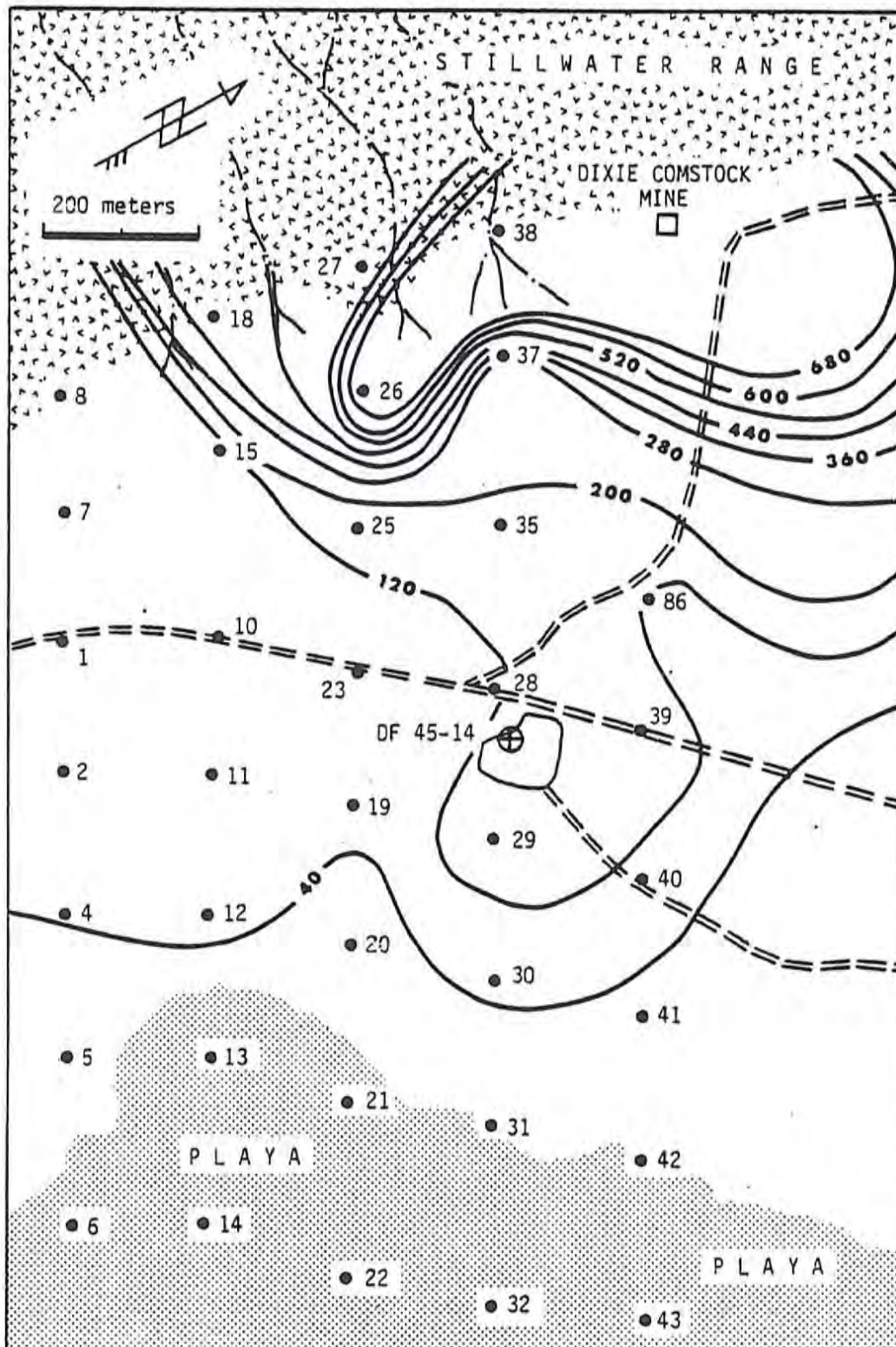


Figure 3-16a. Contoured mercury geochemical surface in vicinity of DF 45-14. Values indicated by contours; sampling sites indicated by solid circles; extent of geomorphic surfaces indicated by shading.

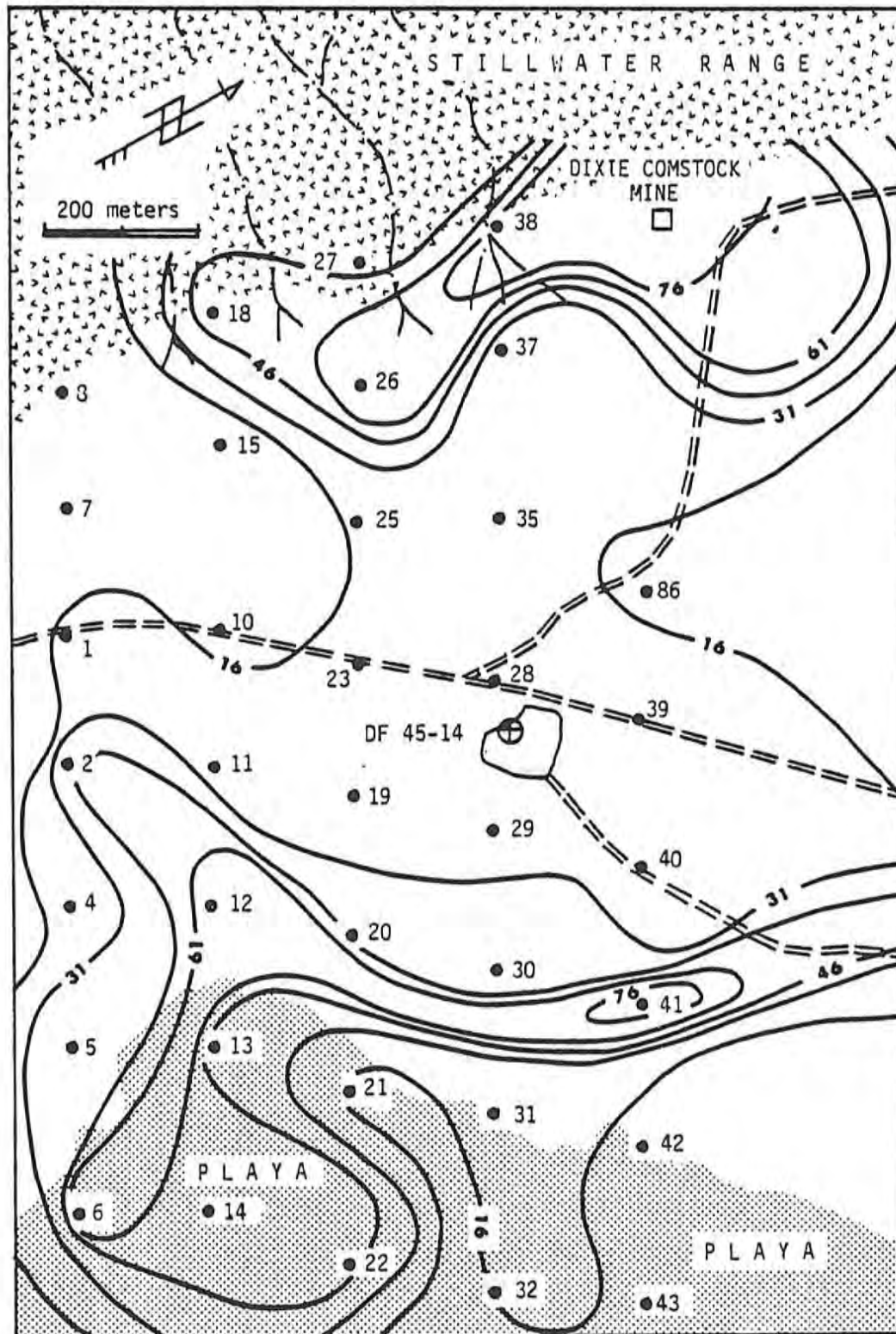


Figure 3-16b. Contoured arsenic geochemical surface in vicinity of DF 45-14. Values indicated by contours; sampling sites indicated by solid circles; extent of geomorphic surfaces indicated by shading.

anomalous arsenic and mercury along its trend. High anomalous arsenic values along the apparently structurally controlled playa margin, including several associated with springs, suggests some communication with arsenic rich thermal waters at depth. The anomalous arsenic values near the playa are seldom accompanied by high mercury concentrations but this may be due to the nature of the sediments and the affinity of arsenic for the liquid phase. Low anomalous values of mercury and arsenic also occur near and along the Buckbrush fault trace in the northern portion of the study area. Geochemical trends west of the Dixie Meadows fault also suggest some structural control. The association of geochemical anomalies with the more easterly fault traces is significant. Two low-production wells have been drilled to the west of these 'inner' faults whereas at least three productive wells have been drilled east of them.

Although contamination from mine tailings complicates the interpretations around DF 45-14 it appears that it has been drilled west of the two inner graben faults which may communicate with a reservoir at depth. Very high heat flow in this area is probably due to some fluid movement along the range front fault or a structure between the range front and the well.

DF 66-21 is also west of the inner graben structures as outlined by geochemical anomalies and geomorphic features. A zone of high pressure hot water occurs at 1430 m in this well and may represent a zone which communicates with reservoir conduits to the east. This zone is associated with a layer of red clay which is similar in mineralogy to that of the extensively altered, high mercury area at the fumarole chain northwest of the drill site. A possible structural link between the high heat flow zone and the fumaroles is suggested by a zone of lineaments which would correspond to the surface trace of a basin and range structure intersecting the drillhole at 1430 m. These lineaments if extended along trend intersect the range front at the fumaroles. Some anomalous surface geochemical values near the lineaments lend support to this possibility. Based on the available data it appears that the geothermal reservoir in Dixie Valley lies east of the Dixie Meadows fault and that communication with the Stillwater fault and other structures to the west is generally poor, perhaps due to sealing by mineral deposition.

Where follow-up sampling at 30 m (100 ft) intervals has been performed

a correlation with structures has been observed. More sampling at this interval should be conducted, particularly to the north in the area around Section 14 (T24N, R36E). An extension of profile C-C' further across the valley could be quite useful.

Because the nature of the Dixie Valley Geothermal System is still not completely understood, it is difficult to evaluate the effectiveness of mercury and arsenic soil geochemistry as an exploration tool. Tentatively it appears that the soil sampling together with previous structural interpretations has provided a plausible explanation for much of the observed drilling data, including the one meter and shallow gradient holes. It would also appear as though the broad grid sampling with follow-up work on close center could be useful in selecting drilling targets. Clearly, it rests with further deep drilling to confirm, modify or contradict the conclusions of this work.

The results of the survey are not clear cut and considerable interpretation has been necessary, particularly with regard to the geochemical behavior of the indicator elements. In many ways, As and Hg have complemented each other in this work. Their correlation is strongest in mineralized areas and this is a useful way of interpreting the origin of those type of anomalies.

Several other methods of determining the source of geochemical anomalies in surveys of this type might be a useful refinement. A selective leach method of analysis as proposed by Bradshaw and others (1974) would allow for the discrimination of different chemical forms of arsenic and mercury. The different chemical forms such as sulfide, chloride or adsorbed complex have clear source implications so that by selectively leaching, for instance the sulfide, a mechanically dispersed mineral origin for an anomaly could be detected.

Another method for discriminating chemical species and hence origins has been proposed for mercury by Koksoy and others (1967). The essence of the method is differentiation by the temperature of decomposition for different species.

Another useful refinement to the present work may be the inclusion of other elements in the analysis. Possibly suitable for this work would be boron which is found in elevated concentrations (approximately 10 ppm) in the thermal waters of DF 66-21 and DF 45-14. Boron is a volatile element commonly associated with geothermal fluids (Hem, 1970b; Koga and

Noda, 1976). Being generally unassociated with sulfide mineral deposits it might be a more reliable indicator in regions such as Dixie Valley.

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Chapter 4. PETROCHEMISTRY

By: Elaine J. Bell

4.0 PETROCHEMICAL ANALYSIS

4.1 Introduction

4.1.1 Purpose and Scope

The primary purpose of the petrochemical analysis of the subsurface drill chip samples from the two deep exploratory wells (DF 45-14 and DF 66-21) in Dixie Valley was to identify steam or hot water entries and to delineate the associated geochemical zonations. Secondly, the data allow for a limited re-evaluation of the integrated model of the Dixie Valley Geothermal System as well for as an evaluation of the technique as an exploration and analytical tool.

The present study is limited to the analysis of available drill chip samples from DF 45-14 and DF 66-21 in Dixie Valley, with elemental analysis for the following five trace elements as indicators of the activity of the geothermal system: lead (Pb), zinc (Zn), arsenic (As), antimony (Sb), and mercury (Hg). Both whole rock samples and heavy mineral fraction separates were analyzed, with the heavy mineral fractions showing geochemical enhancement. The resultant data were evaluated with respect to temperature, depth, lithology, and known or inferred permeable zones (i.e., fracture systems) significant to the fracture-controlled geothermal system.

4.1.2 Methods and Analytical Techniques

The petrochemical analysis to determine distribution of selected elements in samples from deep wells DF 45-14 and DF 66-21, relationship of these elements to geothermal activity, and minerals in which these elements are incorporated involved a four-step process: 1) preparation of whole rock and heavy mineral fraction samples, 2) mineralogic analysis of the heavy mineral fractions, 3) quantitative chemical analysis of whole rock and heavy mineral fraction samples for selected trace elements, and 4) evaluation and reduction of the data.

4.1.2.1 Sample Preparation

Limited volumes of processed drill cutting samples from both DF 45-14 and DF 66-21 were available for this study. These samples represent splits of grab samples originally collected at 10-foot intervals and cleaned and washed of drilling mud and other contaminants by the mud loggers. One-hundred foot composite samples were prepared by combining

five original samples in order to 1) minimize potential random sampling biases of individual grab samples; 2) reduce the number of samples that must be prepared and analyzed to a practical total while maintaining adequate spatial resolution for defining geochemical distribution patterns; and 3) maintain a consistent analytical base for comparison of the petrochemical data with the petrologic data generated by previous analyses of the drill cutting samples (Bard, 1980).

Both whole rock and heavy mineral fraction samples were prepared for each 100-foot composite interval. The heavy mineral fractions exhibit enhanced geochemical signatures relative to the whole rock samples. Rock-forming and alteration silicates, essentially lacking hydrothermally derived trace elements, are removed from the samples, leaving a greater than 3.3 specific gravity (+3.3) concentrate containing hydrothermal oxides and sulfides enhanced in the elements of interest. Analysis of whole rock samples for the same sampling intervals can confirm the validity of the heavy mineral fraction analyses and corroborate the location of geochemically anomalous zones.

Heavy mineral fractions were prepared by grinding and sieving each 100-foot composite sample to -80 mesh size to facilitate disaggregation of the particles. A hand magnet was used to separate the magnetic fraction consisting mostly of magnetite with minor residual drill bit shavings. The heavy mineral fraction was separated by settling in methylene iodide (specific gravity 3.3). A minimum of 500 mg of each heavy mineral fraction sample was required to conduct the quantitative analyses for the selected five elements. The amount of +3.3 heavy mineral fraction within a 100-foot composite sample is a function of both the nature (i.e., lithology) and the volume of well cuttings originally supplied to the MMRI for petrographic, petrologic and petrochemical analyses. The lack of sufficient heavy mineral fraction in many of the 100-foot composite samples necessitated the combining of samples into intervals representing as much as 500 feet to obtain the volume of material needed for analysis. This interval compositing results in bias of the sample data for those intervals.

Whole rock samples were prepared for each 100-foot interval by grinding and sieving to -80 mesh size. These samples were then split and, when necessary, combined to correspond to the composited intervals of the +3.3 heavy mineral fractions.

Control samples were prepared for selected whole rock samples and, when sufficient sample was available, for the heavy mineral fractions. These samples were submitted for analysis to provide a check on the accuracy of the analytical technique and to verify internal consistency of the data. In addition, samples of the magnetic fraction of the heavy mineral separates were prepared for analysis in order to evaluate their contribution to the elemental distribution.

4.1.2.2 Mineralogic Analysis

Limited mineralogic studies of the prepared heavy mineral fraction samples were conducted to identify specific minerals and their relative abundances in each interval. Mineral identification was based primarily on petrographic binocular examination under reflected light. Preliminary results were verified and modified by second observer visual scanning. In addition, several minerals were isolated and identified by using standard x-ray powder diffraction camera techniques.

4.1.2.3 Quantitative Elemental Analyses

All whole rock, +3.3 heavy mineral fractions and magnetic fractions, and control samples were submitted to Rocky Mountain Geochemical Corporation, Salt Lake City, for quantitative analysis for lead, zinc, arsenic, antimony, and mercury.

Lead and zinc were determined by atomic absorption; antimony was also determined by atomic absorption following bisulfate fusion. Arsenic determination was made by the colorimetric method discussed in Section 3.1.2. Mercury concentrations were determined by cold vapor generation atomic absorption. Limits of quantitative detection for each element are listed in Table 4-1; minimum reported values for whole rock and for heavy mineral fraction samples are a function of the sample size available for analysis or the minimum concentration of the element in the sample.

4.1.2.4 Data Presentation

4.1.2.4.1 Bar Graph Plots

The distribution of elements within both whole rock and heavy mineral fraction samples for each deep well are depicted by bar graph plots (Plates VI through IX). These histograms are plotted at a scale consistent with the revised data from the petrologic studies of the wells to facilitate

Table 4-1
 Detection Limits for Quantitative Trace Analysis

Element	Detection Limit	Minimum Value Reported*
Lead	5 ppm	- 5 ppm for WRS -25 ppm for HMF
Zinc	5 ppm	- 5 ppm for WRS -25 ppm for HMF
Arsenic	5 ppm	- 5 ppm for WRS -25 ppm for HMF
Antimony	1 ppm	- 1 ppm for WRS - 2 ppm for HMF
Mercury	10 ppb	75 ppb for WRS -10 ppb for HMF

* Minimum values reported are function of sample size available for analysis. WRS -- Whole Rock Samples; HMF -- Heavy Mineral Fractions. Minus (-) symbol should be read as 'less than'.

comparison and evaluation of relationships between the various elemental data and the petrology, structure, or other parameters (Plates IV and V).

4.1.2.4.2 Computer Data Reduction

Statistical reduction and analysis of the data utilized portions of the Statistical Package for the Social Sciences (Nie and others, 1975). Subprogram FREQUENCIES developed one-way frequency distributions as absolute, relative and cumulative frequency values; computed descriptive statistics, including mean, standard error, standard deviation, kurtosis and skewness; and generated histograms of the relative frequency distributions of each variable.

The SCATTERGRAM subprogram provides bivariate correlation analysis, generates bivariate scattergram plots, and computes a simple linear regression with the following associated statistical parameters: Pearson product-moment correlation, tests of statistical significance, and the standard error of estimate. Both actual value variables and computed arithmetic variables were analyzed by SCATTERGRAM.

4.1.2.4.3 Log Probability Graphs

Log probability graphs were plotted for each variable and sample type by using the data generated by the FREQUENCIES subprogram and by applying the plotting percentage technique of Koch and Link (1970). These log probability plots were used to verify the statistical data and to aid in the subjective evaluation and interpretations of the bar graph plots and other tabulated data. The log probability plots for each element in both heavy mineral fraction and whole rock samples for DF 45-14 are depicted in Figures 4-14 through 4-23 (Appendix B), and for DF 66-21 are depicted in Figures 4-24 through 4-33 (Appendix B).

4.1.2.4.4 Other

Tables were prepared listing the relative abundance and distribution of minerals in the heavy mineral fraction samples for each well (Tables 4-2 and 4-3, Section 4.2.1), the selected elements in heavy mineral fraction and whole rock samples for each well (Tables 4-5 through 4-8, Section 4.2.2), and the statistical distribution of each selected element in heavy mineral fraction and whole rock samples for DF 45-14 (Tables 4-13 through 4-22, Appendix B) and DF 66-21 (Tables 4-23 through 4-32, Appendix B).

4.1.3 Previous Work

The usefulness of collecting and analyzing solid samples for geochemical signatures and elemental zonations associated with ore deposits has been recognized in previous studies. The trace element zoning and ore deposition by hydrothermal solutions is a response to temperature, pressure or chemical gradients associated with a primary heat source. Skinner and others (1967), Browne and Ellis (1970), Ewers and Keays (1977), and Weissberg and others (1979) have shown that the spectrum of trace elements in geothermal systems is similar to ore-forming hydrothermal solutions.

Bamford (1978) developed techniques for using solid sampling geochemistry to identify trace elements that would be indicative of gradients and zoning around geothermal systems and that would be useful in exploration and evaluation schemes. Bamford and Christensen (1979) and Bamford and others (1980) have continued to test and refine the geochemical exploration assessment techniques for geothermal systems. These techniques are here applied to the Dixie Valley Geothermal System to test the selected elements as indicators of geothermal activity.

4.1.4 Geochemistry of Selected Trace Elements

The following five sections present brief outlines of the geochemical nature of the selected trace elements. While the specific behavior of these elements within geothermal systems (hydrothermal fluids) is not fully understood, elemental characteristics inferred from ore deposits are a valuable aid in interpreting the relationships and distribution of the elements with respect to the geothermal system. The reader is referred to other references, including Barnes (1967, 1979) and Wedepohl (1978), for more detailed discussion of the geochemical behavior of these elements.

4.1.4.1 Lead

Lead ions, naturally occurring in a +2 or +4 valence state, have a high electronegativity that generally relegates them to substitution for potassium or calcium in later formed minerals and concentration in hydrothermal fluids. The tendency of the element to form covalent bonds affects its distribution as a sulfide mineral such that it is not readily dispersed. Galena (PbS) is the primary sulfide mineral; however, about

20 different lead sulfates are recognized. Appreciable lead concentrations are present in fumarole sublimates as the mineral cotunnite (PbCl_2) which, under certain conditions, has a higher vapor pressure than the zinc chloride. The ionic radius of lead allows it to substitute for potassium, strontium, barium, and in some cases, calcium or sodium in minerals such as olivine ($(\text{Mg,Fe})_2\text{SiO}_4$), pyroxene, and amphibole. Lead is generally present as the mineral galena (PbS), but is also present in lead-arsenic sulfides as well as being a trace element in pyrite (FeS_2) and magnetite (Fe_3O_4).

4.1.4.2 Zinc

Zinc occurs as a +2 ion and forms halides, oxyacid salts or complexes that are highly soluble in water. Zinc forms covalent bonds and is a trace element in many common minerals in which it competes with iron (Fe^{++}). It is generally present in magmatic iron ores such as ilmenite (FeTiO_3) and magnetite (Fe_3O_4). In sulfides zinc is associated with very late hydrothermal bodies, being found with copper and lead in sphalerite (ZnS) and galena (PbS) and various complex sulfide minerals. Unlike lead, zinc is very mobile and is generally transported as a chloride complex. In rocks where carbon dioxide is dissolved, as much as 50 percent of the zinc may be in a mobile state. The dispersion patterns of zinc around areas of mineralization are normally distinct. Zinc, along with arsenic, is an indicator of gold deposition. Replacing iron and magnesium in mineral lattices, zinc is a common trace element in magnetite (Fe_3O_4) and diopside ($\text{CaMgSi}_2\text{O}_6$).

Lead and zinc are distinctly more abundant in acidic and intermediate rocks than in basic or ultrabasic igneous rocks. They generally are progressively deposited outward from an ore zone (halo effect) and may be accompanied by barite (BaSO_4). Metamorphism may result in remobilization of these trace elements. Zinc and lead can both be adsorbed from solution by clay minerals.

4.1.4.3 Antimony

Antimony may be present in solution in -3, +3, +5 or 0 valence states, although it is generally present in the +3 state as an ionic complex. This volatile element combines readily with sulfides and may substitute for arsenic and iron in minerals. Antimony becomes enriched in early

stages of magmatic differentiation in sulfide bodies and, in this manner, may be associated with gabbroic rocks. It also accumulates in late stage granitic rocks and in hydrothermal fluids. Precipitation from low-temperature alkaline hydrothermal solutions can be induced by 1) neutralization by carbon dioxide, 2) oxidation, 3) decreasing temperature, or 4) dilution. Antimony may occur as stibnite (Sb_2S_3) with cinnabar (HgS) and pyrite (FeS_2) or as sulfantimonides of copper, lead, zinc, silver or arsenic.

4.1.4.4 Arsenic

The geochemistry of arsenic is discussed in Section 3.1.4.2 with only significant parameters reiterated here. Occurring in three principal oxidation states (+3, +5, 0), this mobile element generally remains hydrolyzed in aqueous solutions. It commonly forms anionic complexes and polymers, particularly with sulfur and oxygen. Arsenic is frequently associated with mercury and, along with zinc, is indicative of gold deposition. It substitutes in mineral lattices for iron and titanium, commonly in magnetite and ilmenite, but also is found as a trace element in galena (PbS), pyrite (FeS_2), chalcopyrite (CuFeS_2), pyrrhotite (Fe_{1-x}S), and sphalerite (ZnS). Arsenopyrite (FeAsS) is the most abundant arsenic mineral.

4.1.4.5 Mercury

The geochemistry of mercury is discussed in Section 3.1.4.1. Mercury occurs in three principal oxidation states (+1, +2, 0). The relatively high vapor pressure of the element and its compounds is unique. Volatility of mercury is enhanced by the elevated temperatures of hydrothermal systems, and its mobility in the vapor phases allows it to migrate through permeable zones. Mercury is generally found as the sulfide and as a trace element within the lattice of minerals such as pyrite (FeS_2), tourmaline and sphalerite (ZnS).

4.1.5 Acknowledgements

Discussions with Russell W. Juncal, Dennis S. McMurdie and Odin B. Christensen contributed to the development of ideas and helped to focus

the analysis and interpretations of the petrochemical data expressed in this chapter. Dick and Sue Nosker and Mark Chandler are especially thanked for their meticulous efforts in sample preparation. The intensive efforts of Larry T. Larson and D. Burton Slemmons in reviewing and evaluating basic petrologic data were invaluable in developing the detailed base for interpreting the geochemical results. T. R. Bard assisted with data reduction. Appreciation is extended to Mollie A. Stewart for typing and text preparation and to Loretta Sabini for assistance with computer editing. Technical and editorial review by MMRI and SRC personnel, in particular Larry T. Larson, Russell W. Junca and Dennis S. McMurdie, significantly improved this chapter. And, to my husband John, a special expression of thanks for his assistance, support and patience.

4.2 Analytical Results

4.2.1 Mineralogic Occurrences

The relative abundance and distribution of minerals in the samples from DF 45-14 and DF 66-21 are listed in Tables 4-2 and 4-3, respectively. These semi-quantitative mineral compositions of the +3.3 heavy mineral fraction samples and the +3.3 magnetic separates were determined by petrographic methods.

Minerals frequently abundant in the +3.3 fraction samples include hematite (Fe_2O_3)/magnetite (Fe_3O_4), pyrite (FeS_2) and epidote ($\text{Ca}_2(\text{Al,Fe})_3(\text{OH})(\text{SiO}_4)_3$). In addition to these minerals, zircon (ZrSiO_4), tourmaline, barite(?) (BaSO_4), chalcopyrite(?) (CuFeS_2), rutile(?) (TiO_2), diopside ($\text{CaMgSi}_2\text{O}_6$), and garnet ($(\text{Mg,Fe})_3\text{Al}_2(\text{SiO}_4)_2$) are also common in either minor or trace amounts. Various rock-forming silicate minerals and steel shavings are present in minor or trace amounts contaminating the heavy mineral fraction samples. The +3.3 magnetic fraction samples are composed primarily of magnetite and steel shavings with pyrite, ilmenite(?), and various silicate minerals present in either minor or trace amounts as contaminants.

Ilmenite was not specifically identified in any of the heavy mineral fraction samples. However, its presence in at least several intervals is suspected since it was tentatively identified as a trace mineral in the magnetic fraction of DF 66-21 and leucoxene was identified during petrographic analysis of whole rock samples (Bard, 1980). Similarly, x-ray fluorescence data and the identification of leucoxene in whole rock

Table 4-2

RELATIVE ABUNDANCE AND DISTRIBUTION OF MINERALS IN DF 45-14 HEAVY MINERAL FRACTIONS

DEPTH	MAJOR	MINOR	TRACE
100-200	Hematite/Magnetite	Epidote	Tourmaline, Barite(?)
200-300	Hematite/Magnetite	Epidote	Pyrite, Tourmaline
300-400	Hematite/Magnetite	Epidote	Tourmaline, Garnet
400-500	Hematite/Magnetite	Epidote, Pyrite	Tourmaline, Zircon(?), Chalcopyrite(?)
500-700	Pyrite, Hematite/Magnetite	Tourmaline	Chalcopyrite(?)
700-800	Pyrite	Limonite	Epidote, Chalcopyrite(?)
800-900	Pyrite	Hematite/Magnetite	Epidote
900-1000	Pyrite, Magnetite/Hematite		Epidote, Chalcopyrite(?)
1000-1100	Magnetite/Hematite	Epidote, Pyrite	
1100-1300	Pyrite	Hematite/Magnetite	Epidote, Tourmaline, Garnet(?)
1300-1400	Pyrite	Hematite/Magnetite	Tourmaline(?), Chalcopyrite(?)
1400-1600	Pyrite	Hematite/Magnetite	(?) others
1600-1800	Pyrite, Hematite/Magnetite	Epidote	
1800-1900	Pyrite, Hematite/Magnetite	Epidote	
1900-2200	Pyrite, Hematite/Magnetite	Epidote	
2200-2400	Pyrite, Hematite/Magnetite	Epidote, Barite(?)	
2400-2700	Pyrite	Hematite/Magnetite	Epidote
2700-2900	Pyrite	Hematite/Magnetite	Tourmaline
2900-3000	Pyrite	Hematite/Magnetite	
3000-3200	Pyrite	Hematite/Magnetite	
3200-3300	Hematite	Pyrite, Tourmaline(?)	Magnetite
3300-3500	Pyrite	Hematite/Magnetite	Epidote
3500-3700	Pyrite	Hematite/Magnetite	Epidote
3700-4000	Pyrite	Epidote	Hematite/Magnetite
4000-4100	Epidote(?)	Magnetite	Pyrite, Tourmaline(?), Chalcopyrite(?)
4100-4200	Epidote(?)	Hematite/Magnetite	Pyrite
4200-4500	Hematite/Magnetite	Pyrite	Epidote, (?) other
4500-4900	Pyrite	Hematite/Magnetite	Epidote, Barite(?), (?) other
4900-5400	Pyrite	Hematite/Magnetite	Epidote, Barite(?)
5400-5700	Pyrite	Hematite/Magnetite	Epidote, Barite(?)
5700-5900	Pyrite	Hematite/Magnetite	Epidote
5900-6400	Pyrite	Hematite/Magnetite	Epidote
6400-6900	Pyrite	Hematite/Magnetite	Rutile(?), Zircon(?), Epidote
6900-7400	Pyrite	Hematite/Magnetite	Zircon, Garnet(?), (?) others
7400-7700	Pyrite	Hematite/Magnetite	Tourmaline, Epidote, (?) others
7700-8100	Pyrite, Magnetite/Hematite	Hematite/Magnetite	Tourmaline, Zircon, Epidote, (?) others
8100-8500	Pyrite	Hematite/Magnetite	Epidote
8500-9000	Pyrite	Hematite/Magnetite	Epidote, Barite(?)
Magnetic fraction (5000-7000)	Magnetite/Hematite	Pyrite	Tourmaline, (?) other

Note:

(?) following mineral name indicates tentative identification; (?) other indicates unidentified mineral present.

Table 4-3

RELATIVE ABUNDANCE AND DISTRIBUTION OF MINERALS IN OF 66-21 HEAVY MINERAL FRACTIONS

DEPTH	MAJOR	MINOR	TRACE
135-700	Hematite	Magnetite	Epidote, Pyrite, Diopside
700-900	Hematite	Magnetite	Epidote, Diopside
900-1200	Hematite	Magnetite	Epidote, Diopside
1200-1500	Hematite	Magnetite	Epidote, Diopside
1500-1700	Hematite	Magnetite	Epidote, Diopside
1700-2000	Hematite	Magnetite	Epidote, Diopside
2000-2400	Hematite	Magnetite	Epidote, Diopside
2400-2700	Hematite	Magnetite, Diopside	Epidote
2700-3000	Hematite	Magnetite	Epidote, Diopside
3000-3200	Hematite	Magnetite, Tourmaline	Epidote, Pyrite, Diopside
3200-3400	Hematite	Magnetite, Tourmaline	Epidote, Zircon, Diopside, Rutile(?), (?) other
3400-3700	Hematite	Magnetite, Tourmaline Diopside	Epidote, Zircon
3700-4000	Hematite	Magnetite, Tourmaline	Epidote, Diopside
4000-4100	Hematite	Magnetite, Tourmaline	Diopside
4100-4200	Hematite	Magnetite, Tourmaline	Epidote, Diopside
4200-4300	Hematite	Magnetite, Tourmaline	Pyrite
4300-4400	Hematite	Magnetite, Pyrite Tourmaline	
4400-4500	Pyrite, Tourmaline	Magnetite	Cinnabar(?), (?) other
4500-4600	Hematite/Magnetite	Pyrite	
4600-4800	Pyrite	Magnetite	Rutile, Garnet, Tourmaline
4800-4900	Hematite/Magnetite	Pyrite	Garnet(?)
4900-5200	Hematite/Magnetite	Pyrite	Gold, Tourmaline
5200-5400	Hematite/Magnetite	Pyrite, Epidote	Diopside, Tourmaline
5400-5800	Epidote, Hematite	Pyrite	Magnetite, Tourmaline, Diopside
5800-6100	Epidote, Tourmaline	Pyrite, Magnetite	(?) other
6100-6200	Epidote	Pyrite, Tourmaline	Diopside
6200-6500	Epidote, Pyrite	Tourmaline, Magnetite	Diopside, Zircon(?), (?) other
6500-6600	Epidote, Pyrite	Tourmaline, Hematite	Diopside, Hornblende(?), Corundum(?)
6600-6800	Pyrite	Tourmaline	Rutile, Epidote, (?) other
6800-7100	Pyrite	Epidote, Hematite	Tourmaline, Magnetite, Diopside
7100-7400	Pyrite	Epidote, Hematite	Tourmaline, Magnetite, Diopside
7400-7600	Pyrite	Epidote, Hematite	Magnetite
7600-7800	Pyrite		Epidote, Hematite
7800-8200	Pyrite		Diopside, Hematite
8200-8600	Pyrite		Hematite
8600-8900	Pyrite	Magnetite, Diopside	Epidote, Tourmaline, Zircon, (?) other
8900-9400	Pyrite		Epidote, Tourmaline, Diopside, Corundum(?)
9400-9780	Pyrite	Diopside, Tourmaline	Epidote, Magnetite
Magnetic fraction	Magnetite/Steel		Epidote, Tourmaline, Ilmenite(?)

Note:
(?) following mineral name indicates tentative identification; (?) other indicates unidentified mineral present.

samples (Bard, 1980) suggest the possible presence of ilmenite in DF 45-14. The presence of sphene (CaTiSiO_5) is also suspected since it too was identified during petrographic analysis of whole rock samples.

The presence of other significant minerals is suspected in a number of zones, particularly in DF 66-21, based upon the drill logs and petrographic analysis of the whole rock samples. These minerals and the respective approximate depths are indicated in Table 4-4.

Pyrite and magnetite are present as both primary and secondary minerals distributed throughout both wells. Primary pyrite is generally present as euhedral or subhedral grains in veins; primary magnetite is generally present as euhedral or subhedral igneous grains or in vein fillings. Secondary pyrite and magnetite are generally alteration products of ferromagnesian silicate minerals. Crystalline pyrite and magnetite are commonly partially or completely replaced by pseudomorphous hematite.

Epidote occurs within veins, as an alteration product of ferromagnesian silicates in association with magnetite, or within altered plagioclase grains.

Mineralogical hosts for the selected trace elements can be inferred from the geochemical behavior of these elements (Section 4.1.4). The trace elements have been adsorbed or incorporated into solid solution in pyrite and base metal sulfides or in oxide minerals (e.g., arsenic in pyrite or hematite), have formed discrete sulfide phases (e.g., mercury in cinnabar or lead in galena), or have been incorporated into silicate phases (e.g., zinc in diopside). The significantly enhanced concentrations of these trace elements in the heavy mineral fractions support the assumption that they are contained in sulfides, oxides and other +3.3 mineral phases.

The distribution and relative abundance of the mineral phases in the heavy mineral fraction samples (Tables 4-2 and 4-3) distinguish various zones or intervals within each well as summarized in Figures 4-1 and 4-2. In DF 45-14 (Figure 4-1) a shallow iron oxide-rich zone is replaced below 500 feet by a sulfide-rich zone in which pyrite is the major mineral. Within the pyrite zone, a number of subzones can be distinguished based on the changes in relative abundances of the various non-metallic minerals (i.e., silicates and barite). Significant intervals include 1600 to 2400 feet in which epidote is a minor rather than a

Table 4-4
Mineral Occurrences

<u>Well</u>	<u>Mineral</u>	<u>Depth(s) in feet*</u>	<u>Reference</u>
DF 45-14	Cinnabar (HgS)	900, 1300(?), 1600(?), 1900-2000(?), 4100, 5900(?), 9000(?)	Drill Log (DL)
	Arsenopyrite (FeAsS)	5300(?), 5700(?), 6300(?)	Petrographic Analysis (PA)
	Pyrrhotite (Fe _{1-x} S)	5600-5800(?)	PA
DF 66-21	Malachite (Cu ₂ (CO ₃)(OH) ₂)	700(?), 1000, 1400-1600, 1700, 4200	DL
	Galena (PbS)	1000 6600	DL PA
	Cinnabar (HgS)	1500-1600, 7600, 8800 6600, 7600	DL PA
	Bornite (Cu ₅ FeS ₄)	5000, 5500	DL
	Marcasite (FeS ₂)	5500	DL
	Arsenopyrite (FeAsS)	4600(?)	PA

* (?) following depth indicates tentative mineral identification rather than uncertain depth.

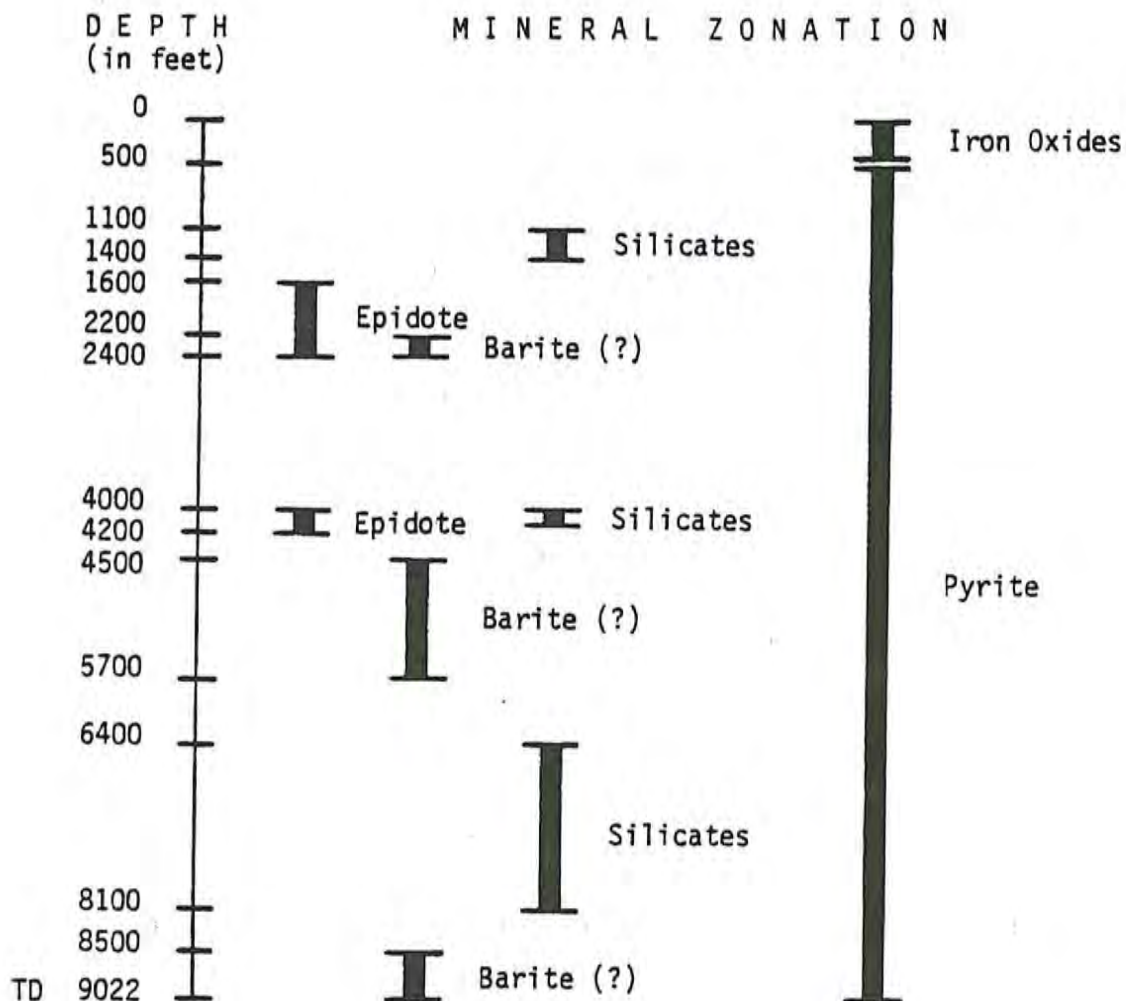


Figure 4-1. Zonation of mineral occurrences in DF 45-14 heavy mineral fraction samples.

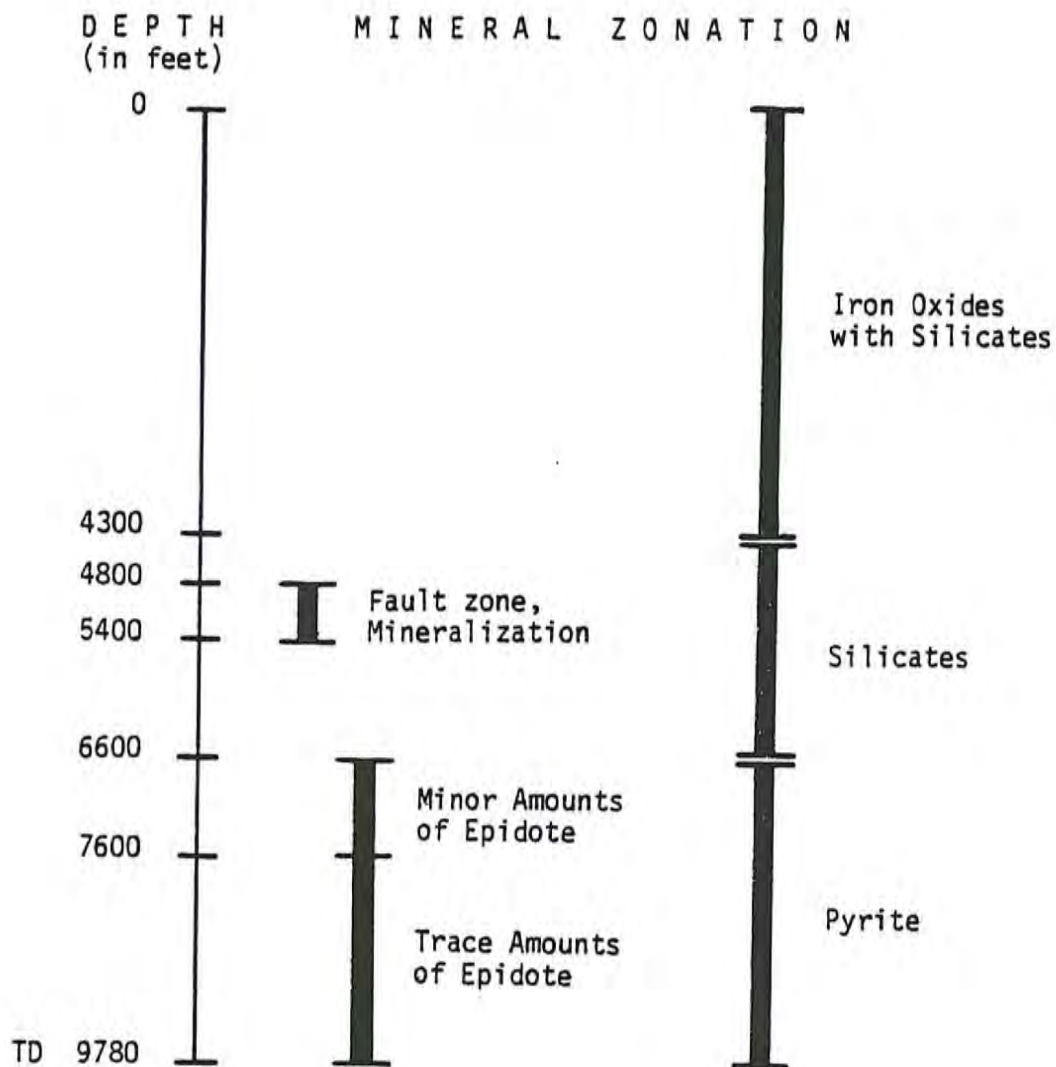


Figure 4-2. Zonation of mineral occurrences in DF 66-21 heavy mineral fraction samples.

trace mineral and 4000 to 4200 feet in which epidote is a major constituent. Barite is significant from 2200 to 2400 feet, from 4500 to 5700 feet and from 8500 feet to total depth of the well (9022 feet). Associations of silicate minerals including epidote, tourmaline, garnet, rutile, and zircon are notable in the following intervals: 1100 to 1400 feet, 4000 to 4100 feet, and 6400 to 8100 feet.

In DF 66-21 (Figure 4-2), an iron oxide-rich zone with minor or trace amounts of various silicate minerals (epidote, diopside, and tourmaline) extends to a depth of 4300 feet. With the appearance of significant pyrite from 4300 to 4400 feet, an intermediate zone starts which is dominated primarily by various silicate minerals to a depth of 6600 feet. Within this intermediate zone from 4800 to 5400 feet is a major mineralized fracture system. Below 6600 feet to total depth of the well is a pyrite dominated zone in which a subdivision at 7600 feet can be based on a decrease in epidote from minor to trace amounts. The possible significance of these mineral distribution zones in both DF 45-14 and DF 66-21 will be discussed in a later section.

4.2.2 Elemental Distribution

4.2.2.1 General

Distributions of the selected trace elements in +3.3 heavy mineral fraction and whole rock composite samples were determined for wells DF 45-14 and DF 66-21 to allow for comparison of the geochemical signature patterns and the known physical characteristics of the geothermal system. For DF 45-14 composite interval samples from 100 feet to a depth of 9022 feet were analyzed; Table 4-5 lists the results for the heavy mineral fraction samples and Table 4-6 lists the results for the whole rock samples. For DF 66-21 composite interval samples from 135 feet to a depth of 9780 feet were analyzed; Table 4-7 and Table 4-8 list the results for the heavy mineral fraction and whole rock samples, respectively. Plates VI through IX present bar graph plots of these data.

Frequency distributions of these data as plotting percentages are presented on log probability plots in Figures 4-14 through 4-33 (Appendix B), with Tables 4-13 through 4-32 (Appendix B) listing corresponding statistical data and specific lithologies for each value increment. Anomalous population threshold values (Table 4-9) for each trace element in both heavy mineral fraction and whole rock samples for each well were

Table 4-5

Distribution of Selected Elements in DF 45-14 Heavy Mineral Fractions

Depth Interval (in Feet)	Lead (ppm)	Zinc (ppm)	Arsenic (ppm)	Antimony (ppm)	Mercury (ppm)	Temperature (°F)
0100 0200	00425	00375	00220	0014	00.125	105
0200 0300	00075	00400	00220	0022	00.010	112
0300 0400	00125	00075	00200	0016	00.010	121
0400 0500	00025	00375	00620	0192	07.400	130
0500 0700	00050	00125	00780	0338	04.800	141
0700 0800	00050	00025	01000	0836	13.800	151
0800 0900	00075	00025	00920	0384	16.700	156
0900 1000	00075	00025	00550	0104	01.700	163
1000 1100	00075	00050	00780	0514	00.100	169
1100 1300	00075	00075	06000	0500	02.100	175
1300 1400	00475	00075	06500	0338	00.270	180
1400 1600	00100	00075	03800	0214	09.100	187
1600 1800	00075	00100	00880	0074	00.375	197
1800 1900	00050	00150	01600	0106	01.100	205
1900 2200	00075	00075	00580	0046	00.570	213
2200 2400	00100	02000	00500	0024	-ND-	223
2400 2700	00175	00650	00900	0032	00.010	231
2700 2900	00150	00100	00650	0020	00.010	240
2900 3000	00150	00100	00750	0014	00.010	246
3000 3100	00075	00175	00780	0012	00.010	248
3100 3200	00150	00075	00420	0002	00.010	251
3200 3300	00150	00050	00175	0002	00.200	255
3300 3500	00250	00175	00200	0002	00.270	260
3500 3700	00200	00300	00650	0008	00.235	268
3700 4000	00125	00175	00800	0036	00.360	277
4000 4100	00050	00150	00400	0014	00.185	284
4100 4200	00025	00125	00125	0002	00.185	287
4200 4500	00325	00325	01500	0022	00.270	298
4500 4900	00425	00200	01300	0020	00.380	314
4900 5400	00475	00600	01000	0016	00.585	327
5400 5700	02200	04100	00400	0020	00.395	336
5700 5900	00200	01200	00200	-ND-	00.185	340
5900 6400	02200	00800	00600	0006	00.965	346
6400 6900	03000	00875	00450	0004	01.400	353
6900 7400	02200	01200	00275	0002	01.800	360
7400 7700	00325	00175	00250	0002	00.505	366
7700 8100	00680	00200	00225	0002	01.600	372
8100 8500	03700	00800	00520	0012	01.600	377
8500 9022	02500	00300	00480	0420	-ND-	379
Magnetic fraction (5000-7000)	00025	00025	00800	0304	09.600	

ND = No Data

Table 4-6

Distribution of Selected Elements in DF 45-14 Whole Rock Samples

Depth Interval (in Feet)	Lead (ppm)	Zinc (ppm)	Arsenic (ppm)	Antimony (ppm)	Mercury (ppm)	Temperature (°F)
0300 0400	00035	00050	00020	0002	00.300	121
0400 0600	00020	00065	00050	0010	03.100	130
0600 0700	00015	00045	00085	0024	06.000	141
0700 0800	00015	00045	00145	0046	07.400	151
0800 0900	00010	00040	00070	0021	03.300	156
0900 1000	00015	00040	00060	0012	01.400	163
1000 1100	00015	00025	00080	0029	00.460	169
1100 1300	00015	00020	00350	0036	02.900	175
1300 1400	00025	00135	00760	0022	02.300	180
1400 1600	00010	00020	00230	0016	00.940	187
1600 1800	-ND-	-ND-	-ND-	-ND-	-ND-	197
1800 1900	00010	00025	00065	0006	00.220	205
1900 2200	00015	00030	00025	0003	00.220	213
2200 2300	00010	00025	00005	0001	00.225	220
2300 2400	00005	00025	00005	0001	00.240	224
2400 2700	00010	00025	00015	0001	00.245	231
2700 2900	00010	00025	00010	0001	00.245	240
2900 3000	00015	00020	00025	0001	00.270	246
3000 3100	00015	00025	00030	0001	00.250	248
3100 3200	00020	00055	00010	0001	00.190	251
3200 3300	00025	00070	00005	0001	00.180	255
3300 3500	00020	00035	00005	0001	00.220	260
3500 3700	00025	00060	00025	0001	00.180	268
3700 4000	00015	00055	00020	0001	00.225	277
4000 4100	00015	00050	00010	0001	00.240	284
4100 4200	00015	00060	00070	0001	00.240	287
4200 4500	00025	00070	00040	0001	00.245	298
4500 4900	00020	00075	00015	0001	00.195	314
4900 5400	00030	00100	00010	0001	00.310	327
5400 5700	-ND-	-ND-	-ND-	-ND-	-ND-	336
5700 5900	00030	00120	00010	0001	00.360	340
5900 6400	00045	00085	00010	0001	00.320	346
6400 6900	00030	00070	00010	0001	00.345	353
7000 7100	00025	00060	00005	0001	00.350	357
7300 7400	00115	00085	00005	0001	00.190	360
7400 7700	00025	00090	00005	0001	00.185	366
7700 8100	00035	00055	00005	0001	00.225	372
8100 8500	00040	00070	00005	0001	00.075	377
8500 9022	00030	00060	00005	0001	00.260	379

ND = No Data

Table 4-7

Distribution of Selected Elements in DF 66-21 Heavy Mineral Fractions

Depth Interval (in Feet)	Lead (ppm)	Zinc (ppm)	Arsenic (ppm)	Antimony (ppm)	Mercury (ppm)	Temperature (°F)
0135 0700	00025	00125	00075	0014	02.300	206
0700 0900	00025	00100	00050	0016	01.300	232
0900 1200	00025	00250	00025	0006	00.400	241
1200 1500	00025	00100	00050	0008	00.285	250
1500 1700	00025	00125	00050	0002	01.400	254
1700 2000	00025	00175	00050	0002	00.375	259
2000 2400	09500	00250	00175	0004	13.000	267
2400 2700	00025	00050	00050	0006	00.435	275
2700 3000	00050	00050	00250	0014	05.500	283
3000 3200	00050	00050	00125	0008	00.225	291
3200 3400	00050	00050	00075	0006	00.315	294
3400 3700	00050	00050	00100	0002	01.800	298
3700 4000	00075	00075	00100	0016	42.000	303
4000 4100	00025	00025	00050	0024	43.000	306
4100 4200	00025	00025	00075	0024	21.000	307
4200 4300	00025	00050	00250	0008	37.000	308
4300 4400	00025	00025	00025	0004	02.500	310
4400 4500	00025	00100	00200	0088	16.000	311
4500 4600	00025	00050	00050	0002	10.300	313
4600 4800	00025	00050	00050	0156	19.000	315
4800 4900	00025	00050	00025	0002	00.260	317
4900 5200	01200	11500	00225	0048	00.105	318
5200 5400	00025	00050	00025	0002	00.680	319
5400 5800	00250	00100	00005	-ND-	00.435	321
5800 6100	00025	00025	00780	0010	03.500	324
6100 6200	00025	00025	00025	0002	00.695	326
6200 6500	00025	00125	00075	0002	00.750	327
6500 6600	00025	00025	00050	0002	02.300	329
6600 6800	00025	00025	00125	0006	01.600	330
6800 7100	00025	00100	01200	0014	00.660	332
7100 7400	00075	00050	00175	0020	01.800	333
7400 7600	13500	00520	00200	1600	13.400	334
7800 8200	01000	00950	00520	0068	07.200	335
8200 8600	06000	00125	00480	0048	00.730	336
8600 8900	00650	00175	00320	0008	00.305	336
8900 9400	01100	00150	00420	0002	00.995	-ND-
9400 9780	01200	01300	00100	0018	08.600	-ND-
Magnetic fractions	00025	00125	00025	0002	05.500	
	00025	00125	00025	0002	06.100	

ND = No Data

Table 4-8

Distribution of Selected Elements in DF 66-21 Whole Rock Samples

Depth Interval (in Feet)	Lead (ppm)	Zinc (ppm)	Arsenic (ppm)	Antimony (ppm)	Mercury (ppm)	Temperature (°F)
0135 0700	00020	00050	00005	0001	00.130	206
0700 0900	00015	00035	00005	0001	00.180	232
0900 1200	00010	00035	00010	0001	00.095	241
1200 1400	00005	00040	00005	0001	00.165	250
1400 1700	00005	00045	00010	0001	00.130	254
1700 2000	00005	00050	00005	0001	00.200	259
2000 2400	00065	00075	00010	0001	00.120	267
2400 2700	00005	00050	00010	0001	00.140	275
2700 3000	00010	00055	00005	0001	00.160	283
3000 3200	00010	00055	00005	0001	00.150	291
3200 3400	00010	00040	00010	0001	00.195	294
3400 3700	00010	00060	00005	0001	00.140	298
3700 4000	00010	00060	00005	0001	00.120	303
4000 4100	00015	00050	00010	0003	00.130	306
4100 4200	00020	00065	00015	0002	00.190	307
4200 4300	00015	00045	00020	0002	00.280	308
4300 4400	00010	00035	00010	0002	00.300	310
4400 4500	00010	00025	00010	0006	00.720	311
4500 4600	00010	00020	00005	0007	00.460	313
4600 4800	00020	00015	00005	0002	00.235	315
4800 4900	00015	00020	00005	0005	00.180	317
4900 5200	00045	00055	00005	0004	00.150	318
5200 5400	00020	00010	00005	0002	00.105	319
5400 5800	00010	00010	00005	0001	00.085	321
5800 6100	00020	00010	00050	0001	00.125	324
6100 6200	00020	00010	00005	0001	00.255	326
6200 6500	00020	00015	00005	0001	00.125	327
6500 6600	00020	00010	00010	0001	00.685	329
6600 6800	00005	00005	00005	0001	00.265	330
6800 7100	00005	00025	00030	0003	00.330	332
7100 7400	00010	00020	00005	0002	00.260	333
7400 7600	00015	00035	00005	0001	00.200	334
7600 7800	00020	00065	00010	0001	00.130	335
7800 8200	00020	00060	00005	0002	00.115	335
8200 8600	00040	00075	00005	0001	00.175	336
8600 8900	00030	00070	00005	0001	00.105	336
8900 9400	00030	00080	00005	0001	00.175	-ND-
9400 9780	00025	00065	00005	0001	00.155	-ND-

ND = No Data

Table 4-9
Threshold Values of Anomalous Populations

<u>Well</u>	<u>Sample Type</u> *	<u>Element</u>	<u>Threshold Value</u>
DF 45-14	HMF	Lead	1500 ppm
		Zinc	** (1500 PPM)
		Arsenic	2000 ppm
		Antimony	80 ppm or 250 ppm
		Mercury	2500 ppb
DF 45-14	WR	Lead	60 ppm
		Zinc	** (100 ppm)
		Arsenic	100 ppm
		Antimony	15 ppm
		Mercury	600 ppb
DF 66-21	HMF	Lead	400 ppm or 2000 ppm
		Zinc	700 ppm
		Arsenic	** (275 ppm)
		Antimony	300 ppm
		Mercury	4 ppm
DF 66-21	WR	Lead	** (45 ppm)
		Zinc	**
		Arsenic	** (15 ppm)
		Antimony	** (5.5 ppm)
		Mercury	400 ppb

** Double asterisk denotes a log normal population; when a (value) is provided, it signifies a possible anomalous population threshold with uncertainty based upon the small size of the data set.

* Sample types include HMF = Heavy Mineral Fractions and WR = Whole Rock samples.

determined from the log probability plots. It should be noted that a number of plots are interpreted as indicating only one log-normal population. However, the small size of the data sets in some cases makes identification of anomalous populations uncertain; tentative threshold values are indicated for these log probability plots.

By applying these threshold values to their respective data sets, anomalous intervals can be identified and evaluated. Tables 4-10 and 4-11 list the anomalous trace element concentrations in DF 45-14 and DF 66-21, respectively. The spatial relationships and relative values of these anomalous intervals are depicted in Figures 4-3 through 4-6.

4.2.2.2 DF 45-14

Three distinct zones are distinguishable in DF 45-14 (Figure 4-3). The first, extending from 400 feet to a depth of 1600 feet, is characterized by concentrations of arsenic, antimony and mercury. Arsenic is restricted to below 1100 feet; antimony is fairly evenly distributed throughout the zone but shows a maximum anomaly occurring above 1000 feet; and mercury is restricted to the interval above 1000 feet with the exception of a 6.6 ppm anomaly from 1400 to 1600 feet. Similar relationships among the anomalous concentrations of these three elements within the whole rock samples also define this zone (Figure 4-4). This spatial-value relationship is consistent with the geochemical behavior of these elements. The mercury, being more volatile, occurs as a halo above the arsenic, with antimony reflecting an affinity for both elements.

The second zone extends from 5400 feet to a depth of 7400 feet and is characterized by anomalous lead concentrations in the heavy mineral fraction samples (Figure 4-3). Secondary confirmation of this zone is provided by tentative zinc anomalies in both heavy mineral fraction and whole rock samples from 5400 to 5700 feet and a 55 ppm lead anomaly in the whole rock sample from 7300 to 7400 feet (Figure 4-4).

The third zone extends from 8100 feet to the total depth of the well (9022 feet) and is defined solely by anomalous lead concentrations in the heavy mineral fraction samples (Figure 4-3). No anomalous arsenic, antimony or mercury concentrations occur in either the second or third zones. This is consistent with the geochemical behavior of these elements.

Table 4-10

Anomalous Trace Element Concentrations in DF 45-14

Element	Threshold Values* HMF; WR in ppm	Depth Interval in feet	Anomalous Concentrations* (Actual - Threshold Value) in ppm	
			HMF	WR
Lead	1500; 60	5400 - 5700	700	--
		5900 - 6400	700	--
		6400 - 6900	1500	--
		6900 - 7400	700	--
		7300 - 7400	--	55
		8100 - 8500	2200	--
		8500 - TD	1000	--
Zinc	(1500); (100)	1300 - 1400	--	(35)
		2200 - 2400	(500)	--
		5400 - 5700	(2600)	--
		5700 - 5900	--	(20)
Arsenic	2000; 100	700 - 800	--	45
		1100 - 1300	4000	250
		1300 - 1400	4500	660
		1400 - 1600	1800	130
Antimony	250; 15	500 - 700	88	--
		600 - 700	--	9
		700 - 800	586	31
		800 - 900	134	6
		1000 - 1100	264	14
		1100 - 1300	250	21
		1300 - 1400	88	7
		1400 - 1600	--	1
Mercury	2.5; 0.6	400 - 500	4.9	--
		400 - 600	--	2.5
		500 - 700	2.3	--
		600 - 700	--	5.7
		700 - 800	11.3	6.8
		800 - 900	14.2	2.7
		900 - 1000	--	0.8
		1100 - 1300	--	2.3
		1300 - 1400	--	1.7
		1400 - 1600	6.6	0.34

* Values given in parentheses indicate tentative anomalous population; see Table 4-9. HMF - Heavy Mineral Fraction; WR - Whole Rock.

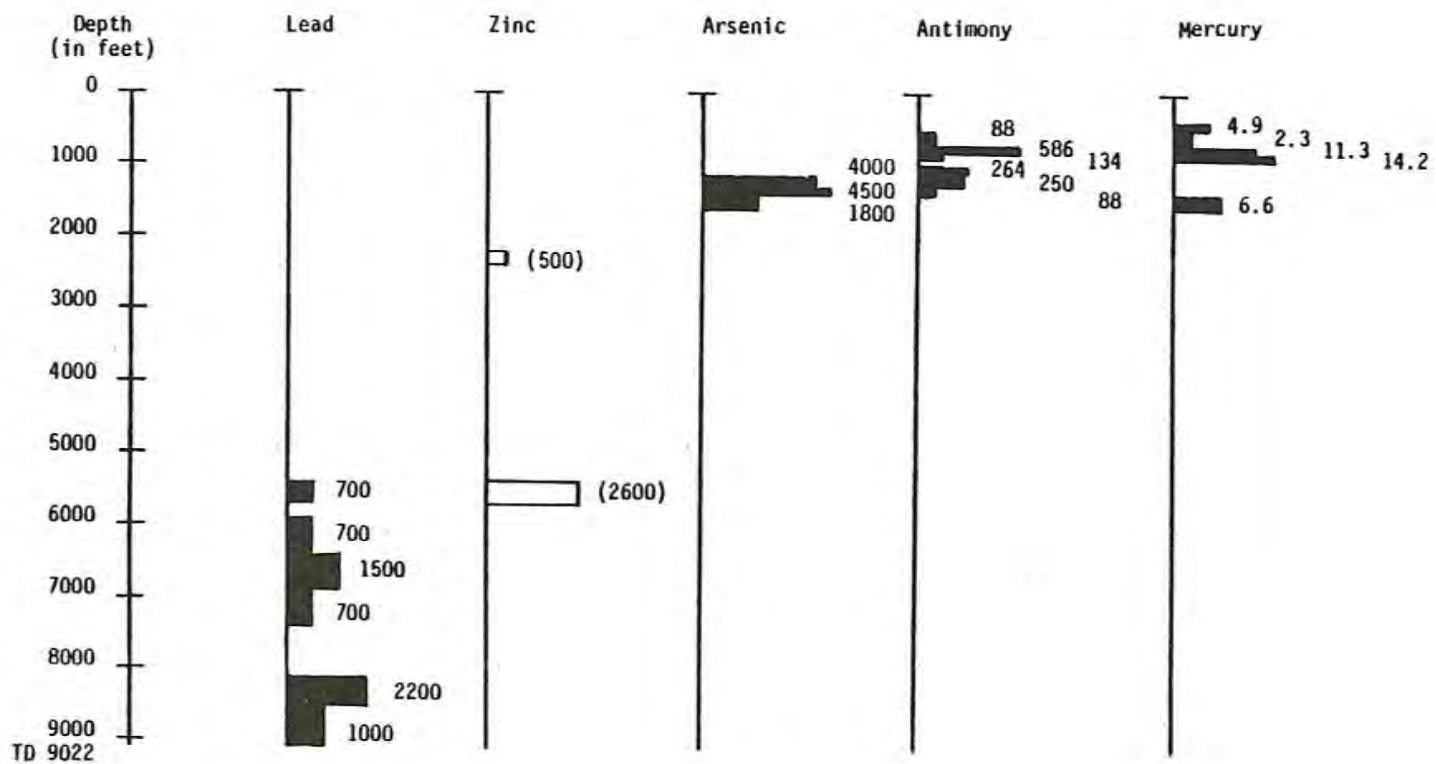
Table 4-11

Anomalous Trace Element Concentrations in DF 66-21

Element	Threshold Values* HMF; WR in ppm	Depth Interval in feet	Anomalous Concentrations* (Actual - Threshold Value)	
			HMF	WR in ppm
Lead	2000; (45)	2000 - 2400	--	(20)
		2200 - 2400	7500	--
		7400 - 7600	11500	--
		8200 - 8600	4000	--
Zinc	700; (**)	4900 - 5200	10800	(**)
		7800 - 8200	250	(**)
		9400 - TD	600	(**)
Arsenic	(275); (15)	4200 - 4300	--	(5)
		5800 - 6100	(515)	(35)
		6800 - 7100	(925)	(15)
		7800 - 8200	(245)	--
		8200 - 8600	(205)	--
		8600 - 8900	(45)	--
Antimony	300; (5.5)	4400 - 4500	--	(0.5)
		4500 - 4600	--	(1.5)
		7400 - 7600	1300	--
Mercury	4.0; 0.400	2200 - 2400	9.0	--
		2700 - 3000	1.5	--
		3700 - 4000	38.0	--
		4000 - 4100	39.0	--
		4100 - 4200	17.0	--
		4200 - 4300	33.0	--
		4400 - 4500	12.0	0.320
		4500 - 4600	6.3	0.620
		4600 - 4800	15.0	--
		6500 - 6600	--	0.285
		7400 - 7600	9.4	--
7800 - 8200	3.2	--		
9400 - TD	4.6	--		

* Values given in parentheses indicate tentative anomalous populations; see Table 4-9. HMF - Heavy Mineral Fraction; WR - Whole Rock.

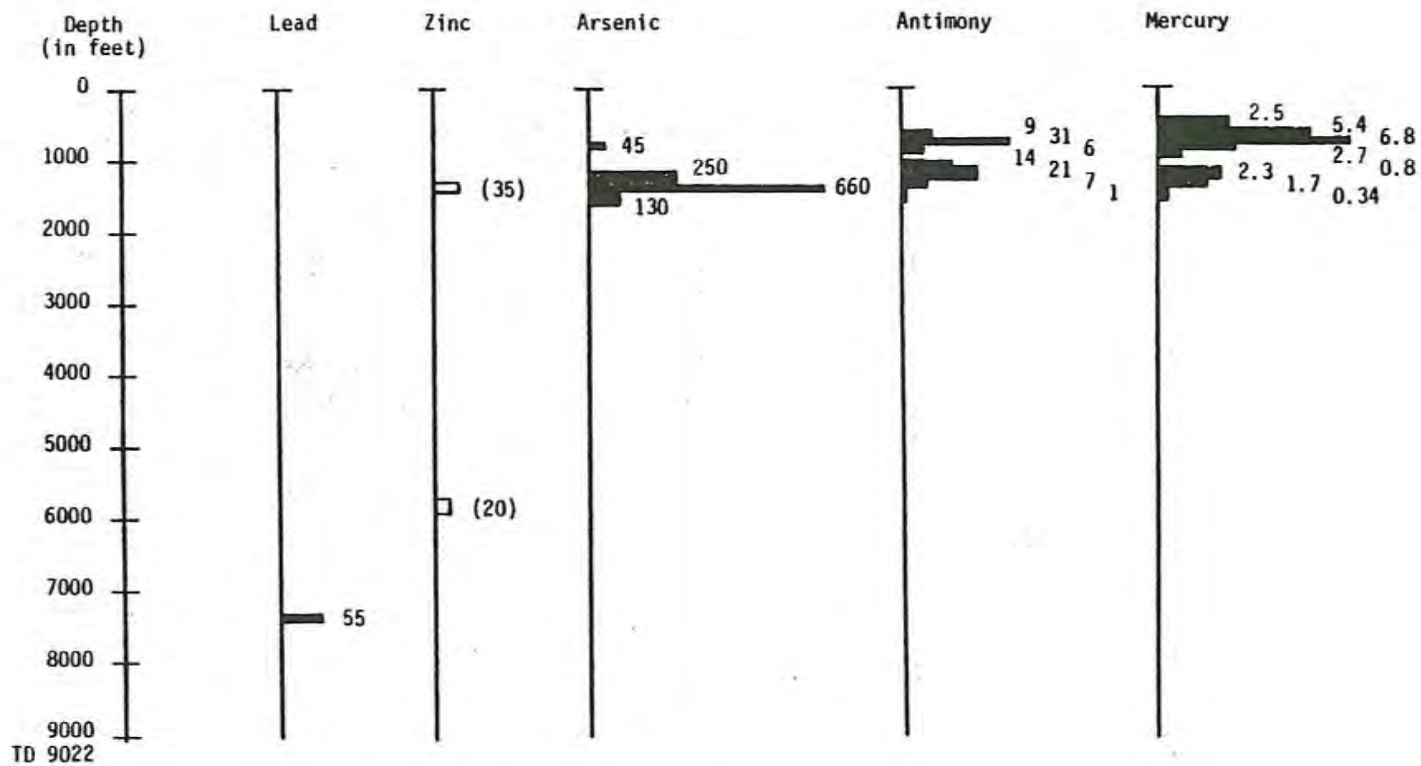
** Indicates single log-normal population; no anomalous values.



Anomalous Concentration and Value in ppm.
 Tentative Anomalous Concentration and Value in ppm. See Table 4-10.

Note: Value scales for concentrations are consistent within each column.

Figure 4-3. Anomalous concentrations of selected trace elements in DF 45-14 heavy mineral fraction samples.



■ Anomalous Concentration and Value in ppm.
 □ Tentative Anomalous Concentration and Value in ppm. See Table 4-10.
 Note: Value scales for concentrations are consistent within each column.

Figure 4-4. Anomalous concentrations of selected trace elements in DF 45-14 whole rock samples.

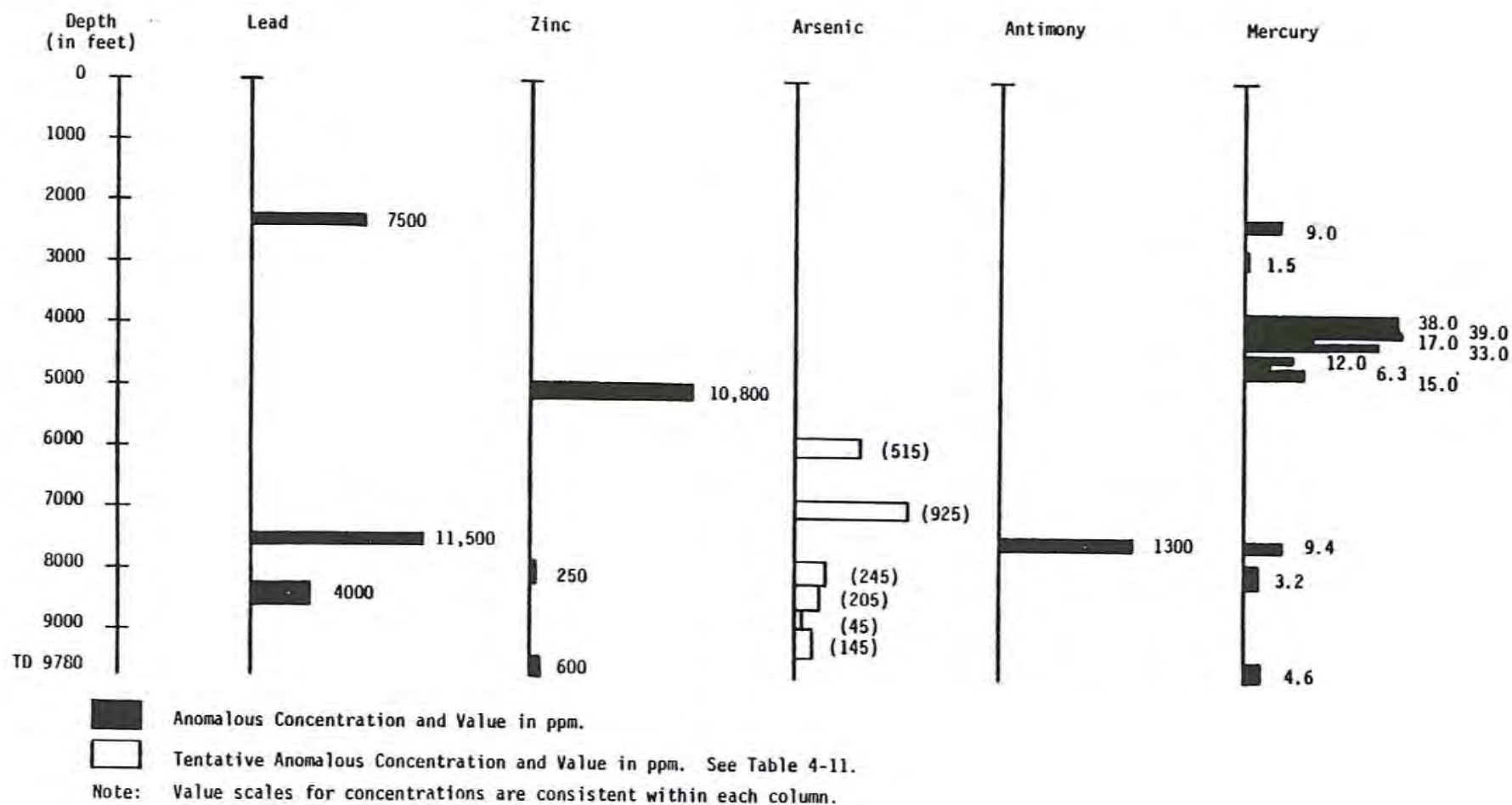


Figure 4-5. Anomalous concentrations of selected trace elements in DF 66-21 heavy mineral fraction samples.

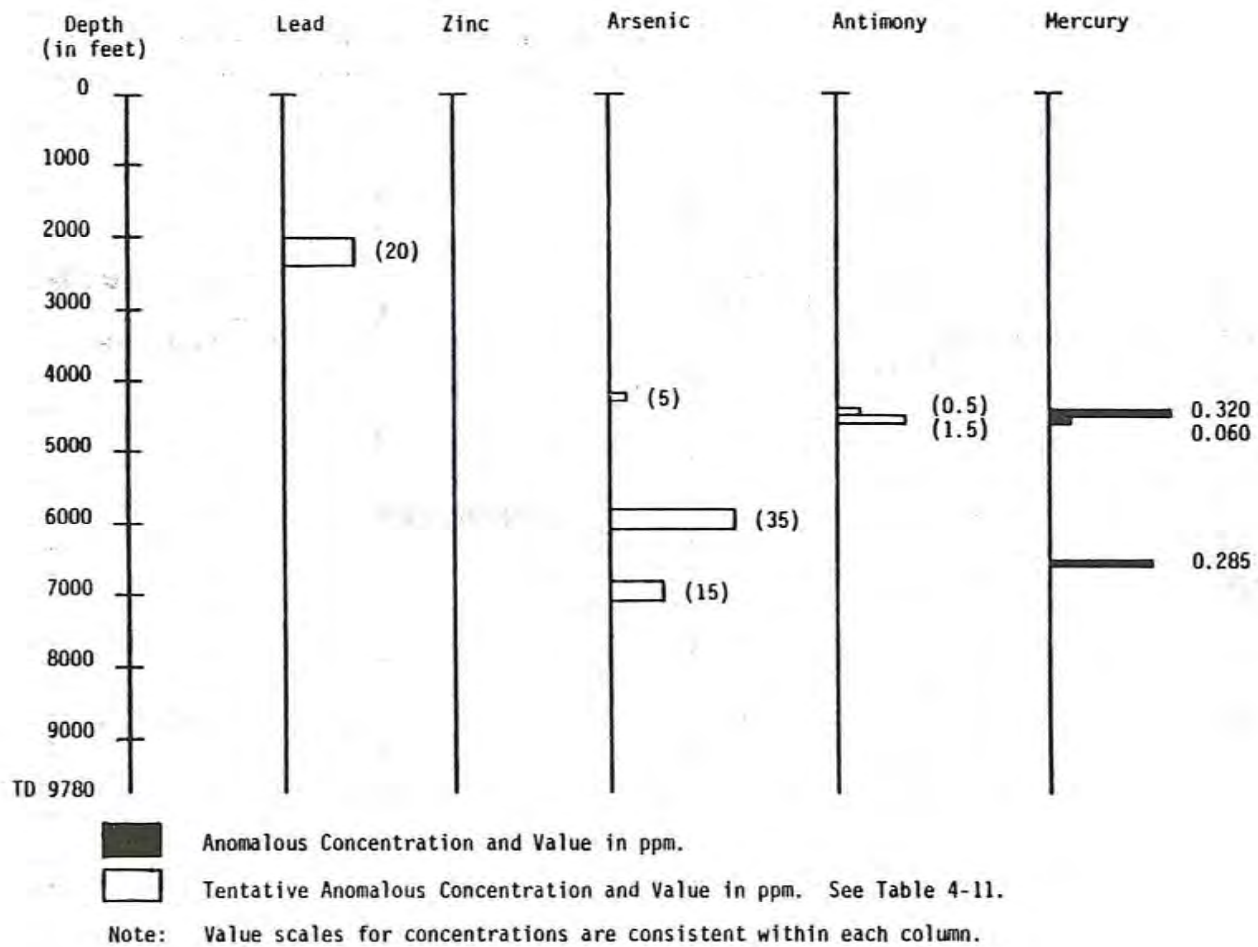


Figure 4-6. Anomalous concentrations of selected trace elements in DF 66-21 whole rock samples.

4.2.2.3 DF 66-21

In DF 66-21, a number of zones can also be defined by the spatial distribution and correspondence of anomalous concentrations of the trace elements (Figures 4-5 and 4-6). It should be noted that the patterns are not as distinct as in DF 45-14, nor as simple. In addition, the respective trace element abundances for whole rock samples in DF 66-21, with the exception of mercury, appear to be log-normal populations (Figure 4-29 through 4-33, Appendix B), that is, that no anomalous concentrations are present and only very tentative interpretations could be made in identifying possible lead, arsenic and antimony anomalies. This is also true for arsenic concentrations in the heavy mineral fraction samples (Figure 4-26, Appendix B).

The following zones can be identified in DF 66-21 (Figure 4-5):

- 1) a zone from 2200 to 2400 feet characterized by anomalous lead and mercury concentrations;
- 2) a zone from 3700 to 4800 feet characterized by anomalous mercury concentrations ranging up to 39 ppm in heavy mineral fraction samples, with anomalous concentrations of mercury also occurring in whole rock samples between 4400 and 4600 feet;
- 3) a zone from 4900 to 5200 feet characterized by a 10,800 ppm anomalous zinc concentration;
- 4) a zone from 7400 to 7600 feet characterized by anomalous concentrations of lead (11,500 ppm), antimony (1300 ppm) and mercury (9.4 ppm);
- 5) a zone from 7800 to 8200 feet characterized by anomalous concentrations of zinc (250 ppm) and mercury (3.2 ppm);
- 6) a zone from 8200 to 8600 feet characterized by a 400 ppm anomalous lead concentration; and
- 7) a zone from 9400 feet to total depth of the well (9780 feet) characterized by anomalous concentrations of zinc (600 ppm) and mercury (4.6 ppm). In addition, a zone can also be distinguished at 6500 to 6600 feet characterized by a minor (0.285 ppm) anomalous mercury concentration in whole rock samples (Figure 4-6). Unlike DF 45-14, these zones exhibit apparent discordant trace element geochemistry that poses a problem of interpretation. The significance of these anomalous trace element concentration zones is discussed in a later section.

4.2.3 Reservoir Zoning

4.2.3.1 General

The distribution of the selected trace elements may also be viewed using the concepts of the reservoir, self-sealed and peripheral zones.

As applied by Bamford and others (1980), the zones may generally be characterized as follows:

The reservoir zone exhibits temperatures greater than 206°C (400°F) with predominantly convective heat transfer. Arsenic and zinc concentrations are enriched, while lead is constant or only slightly enriched, and mercury concentrations remain low.

The self-sealed zone is characterized physically in the rock by alteration, limonite and sulfide accumulation, and/or siliceous and carbonate cement, as well as by the absence of significant surface leakage of fluids. Chemically the outer portion of the self-sealed zone is characterized by high zinc enrichments and arsenic enrichment, with the inner self-sealed zone characterized by arsenic and lesser amounts of zinc. The entire self-sealed zone is characterized by very high mercury enrichment. In general, an outer halo assemblage of erratic low-level mercury and arsenic anomalies, usually associated with locally anomalous concentrations of zinc (and manganese), grades into a zone of increasing mercury concentrations to a region of continuous anomalous arsenic with minor zinc.

The overlying peripheral zone generally exhibits zinc enrichment and physically exhibits slight to moderate evidence of hydrothermal activity. Localized zoning around hot water entries or sealed fracture zones may occur in any or all of these zones.

The distribution of the selected trace elements in DF 45-14 (Plates VI and VII) and DF 66-21 (Plates VIII and IX) and the known physical characteristics of the wells (Plates IV and V) can be subjectively interpreted with respect to the reservoir, self-sealed and peripheral zone concepts. Figure 4-7 depicts these zones as interpreted for the two wells.

4.2.3.2 DF 45-14

In DF 45-14, a peripheral zone, extending to approximately 4100 feet, is characterized by low lead and zinc with arsenic, antimony and mercury concentrations localized with respect to a series of interbedded tuffaceous units which display evidence of secondary mineralization, recrystallization and fracture filling -- i.e., a localized self-sealed zone.

The self-sealed zone extends from about 4100 feet to total depth of the well. This zone is characterized by high lead concentrations with

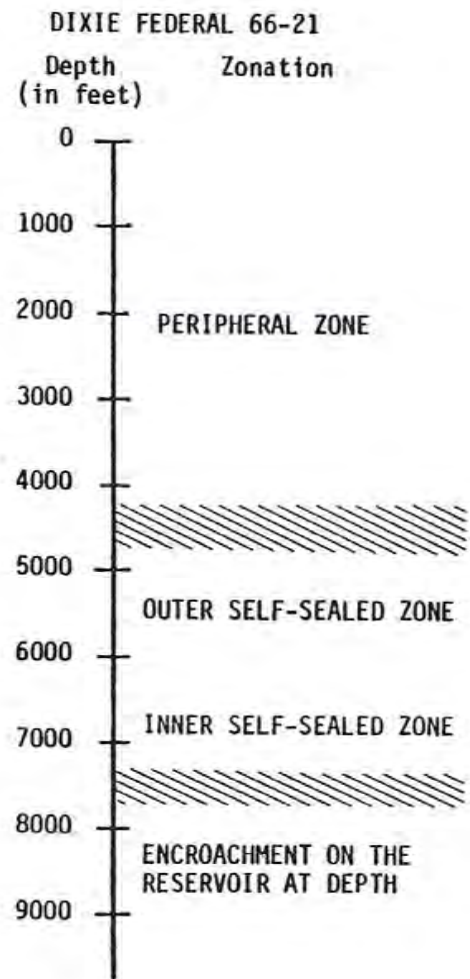
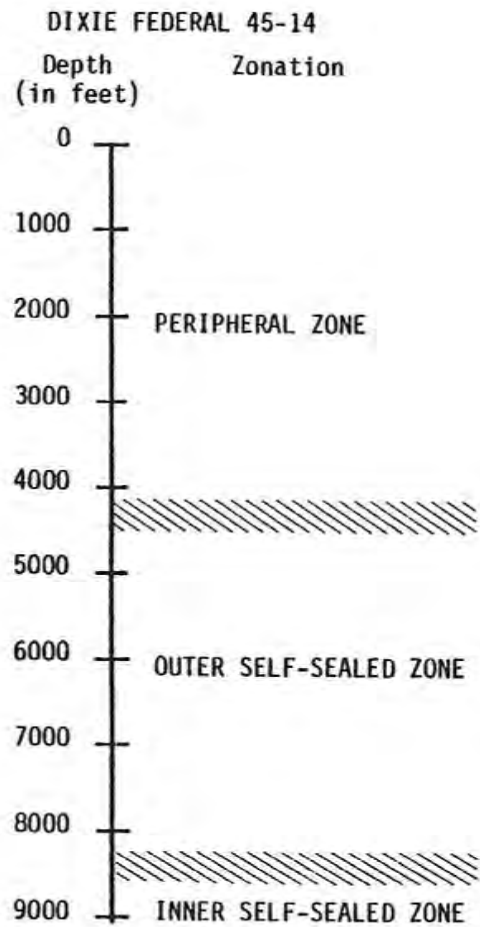


Figure 4-7. Reservoir zoning model for DF 45-14 and DF 66-21.

zinc. The arsenic is high but decreasing somewhat with depth as antimony increases. Mercury concentrations increase with depth. A tentative gradational boundary from outer to inner self-sealed zone is placed between 8100 to 8500 feet. Lead and zinc remain fairly constant and both arsenic and antimony are enriched. Physical evidence for hydrothermal activity within the well (Bard, 1980) is consistent with the zoning concepts. In addition, the presence of fracture zones (Plate IV) with barite (?) deposition (Figure 4-1) and very limited production within the self-sealed zone suggest that self-sealing has occurred and may be continuing.

4.2.3.3 DF 66-21

In DF 66-21, a peripheral zone is identified above about 4300 feet and is characterized by low lead concentrations and constant concentrations of zinc, arsenic, antimony and mercury. Anomalous lead and mercury concentrations, with some enrichment of arsenic, occur between 2000 and 2400 feet associated with the upper boundary of a red clay-alluvial sequence. This concentration may reflect the low permeability of the clay with deposition from aqueous solutions or adsorption by clays of the vapor transported elements. The very high mercury concentration near the base of the peripheral zone probably reflects adsorption on the fine-grained alteration minerals from remobilization or direct vapor transport generated by a major fracture system at 4700 to 5100 feet (Bard, 1980).

The self-sealed zone extends below about 4300 feet with a very broad gradation from outer to inner self-sealed zone. Mercury, zinc and arsenic concentrations characterize the outer self-sealed zone. In the inner self-sealed zone mercury and arsenic increase somewhat while zinc concentrations remain fairly constant. Below about 7500 feet the self-sealed zone begins to encroach on a possible reservoir zone at depth. This encroachment is evidenced by a general increase in mercury, high arsenic and lead concentrations, and significant concentrations of zinc. Localized zoning present within the self-sealed zone is associated with active and sealed fractures (e.g., about 5000 feet, 8000 to 8400 feet, 9400 to 9700 feet). Physical evidence for hydrothermal activity within the well (Bard, 1980) is consistent with the zoning concepts.

The zoning depicted in Figure 4-7 for DF 45-14 and DF 66-21 was tested by scattergram statistical analysis. Figures 4-8 and 4-9 depict the scattergrams for depth versus zinc bivariate analysis for whole rock

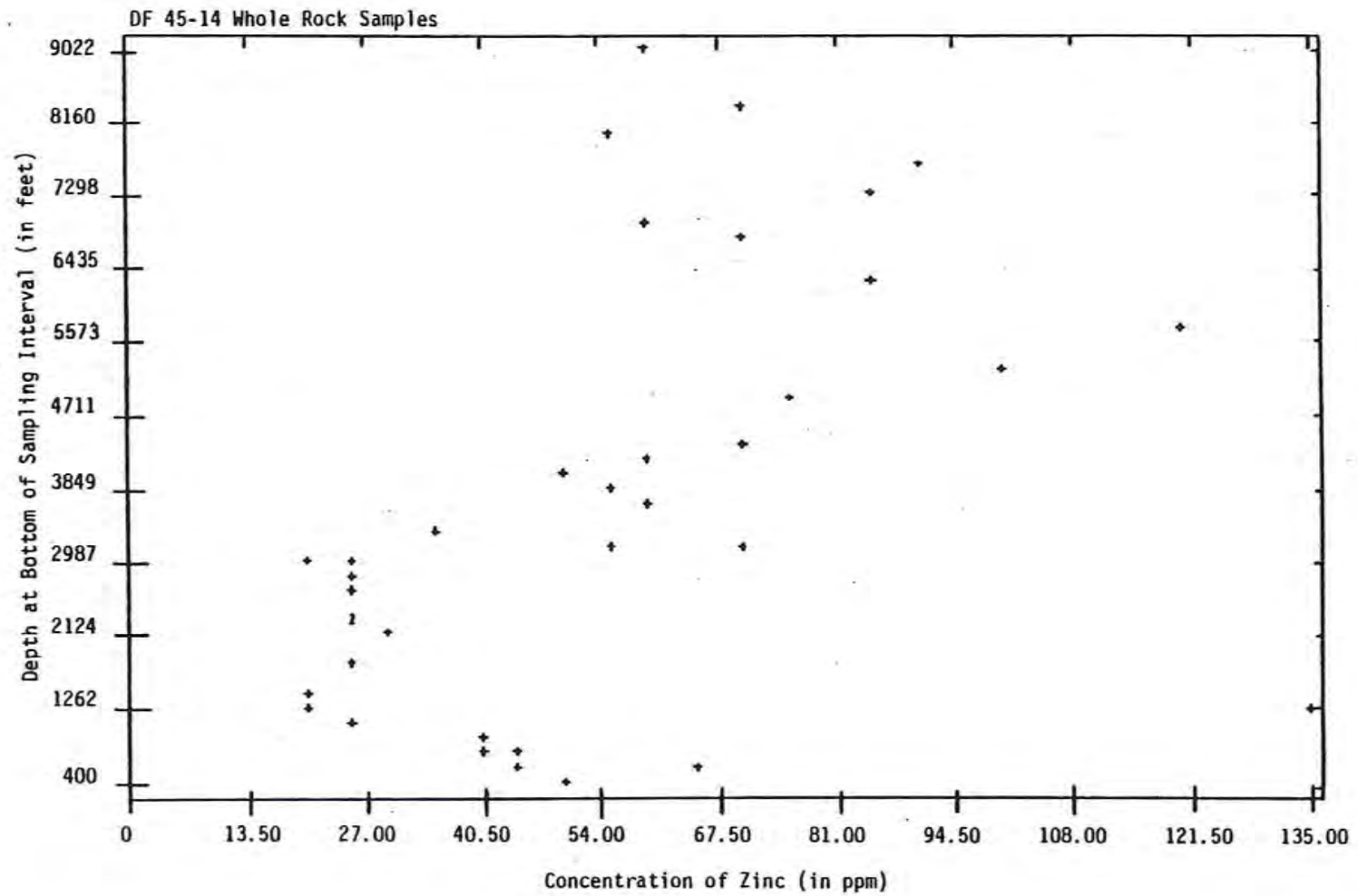


Figure 4-8. Scattergram plot of bivariate analysis of zinc with respect to depth (as bottom of sampling interval) for DF 45-14 whole rock samples.

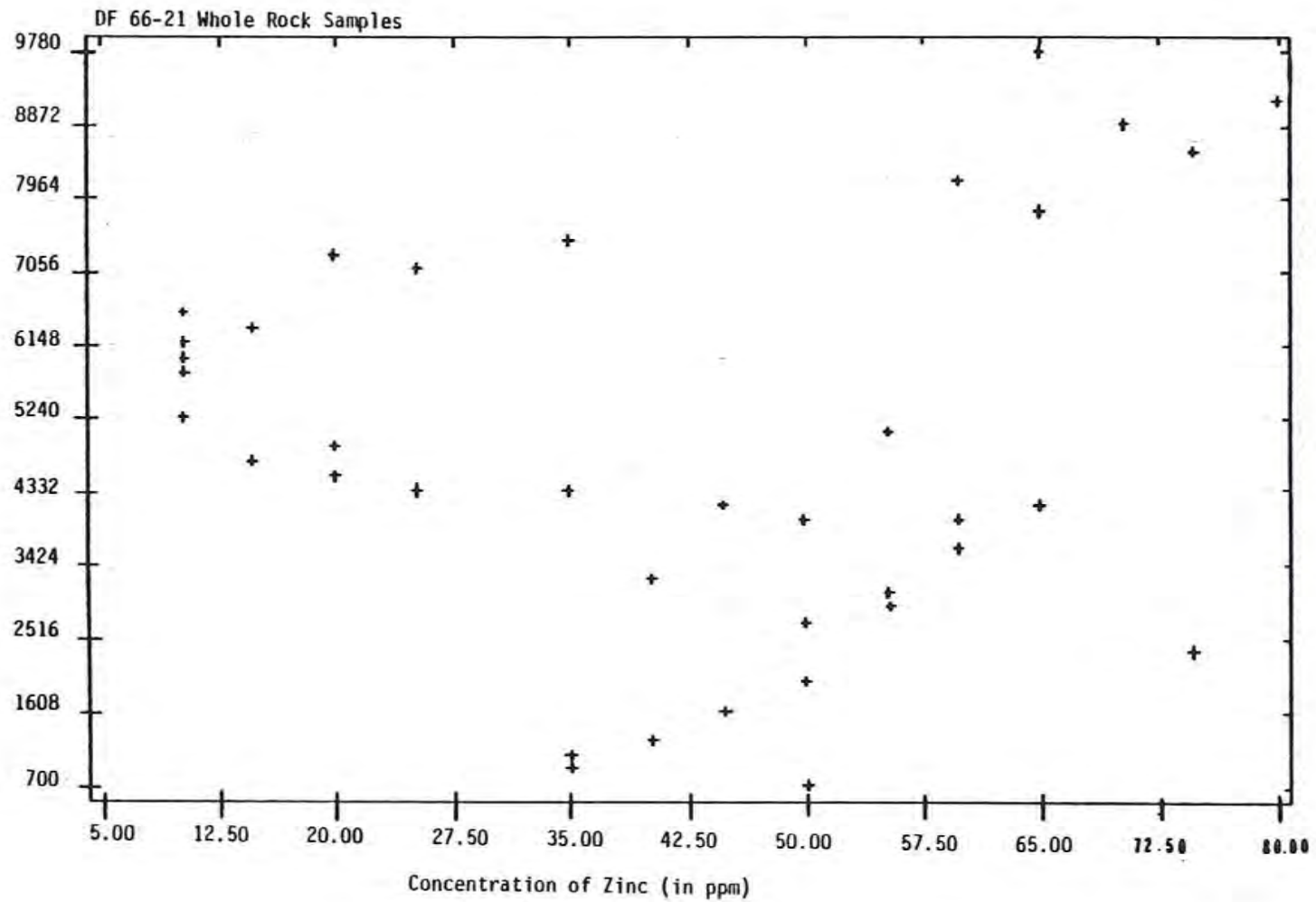


Figure 4-9. Scattergram plot of bivariate analysis of zinc with respect to depth (as top of sampling interval) for DF 66-21 whole rock samples.

samples in the two wells. While the scatter and overlap on the diagrams is a function of gradational boundaries and localized zoning, the divisions between the generalized zones are consistent with those subjectively determined. Similar results are obtained by using computed algebraic variables, such as zinc + arsenic + mercury as depicted in Figures 4-10 and 4-11. Since the general trends of the element concentrations in the heavy mineral fraction samples are reflected by the whole rock sample values, similar scattergram plots can be generated for corresponding data sets but are not included here.

4.2.4 An Alternate Interpretation

4.2.4.1 General

The reservoir zoning interpretation (Figure 4-7) is based on general trends and corresponding concentrations of the selected trace elements. This interpretation is not, however, totally compatible with a zonation based on mineral occurrences in the heavy mineral fraction samples (Figures 4-1 and 4-2) or anomalous concentrations of the selected trace elements (Figures 4-3 through 4-6). For this reason, and because the reservoir zoning technique is still in a formative phase, an alternate interpretation of the wells based on the petrochemical data as well as the physical characteristics of the wells is offered.

Figures 4-12 and 4-13 depict composite logs of DF 45-14 and DF 66-21, respectively, including selected physical parameters and both the mineral occurrences and anomalous concentrations of trace elements in the heavy mineral fraction samples.

4.2.4.2 DF 45-14

The pyrite dominated portion of DF 45-14 reflects the pervasive influence of the geothermal system. Within the pyrite-rich zone, intervals characterized secondarily by epidote and/or barite(?) correlate with intervals of fracture systems (Figure 4-12) and indicate that more intense hydrothermal activity has occurred in the fracture systems and rocks adjacent to them.

The zones of intense hydrothermal activity, as reflected by anomalous concentrations of the trace elements, suggest two different influences. The 400 to 1600 feet interval is characterized by high arsenic concentrations and an overlying halo of high mercury concentration with

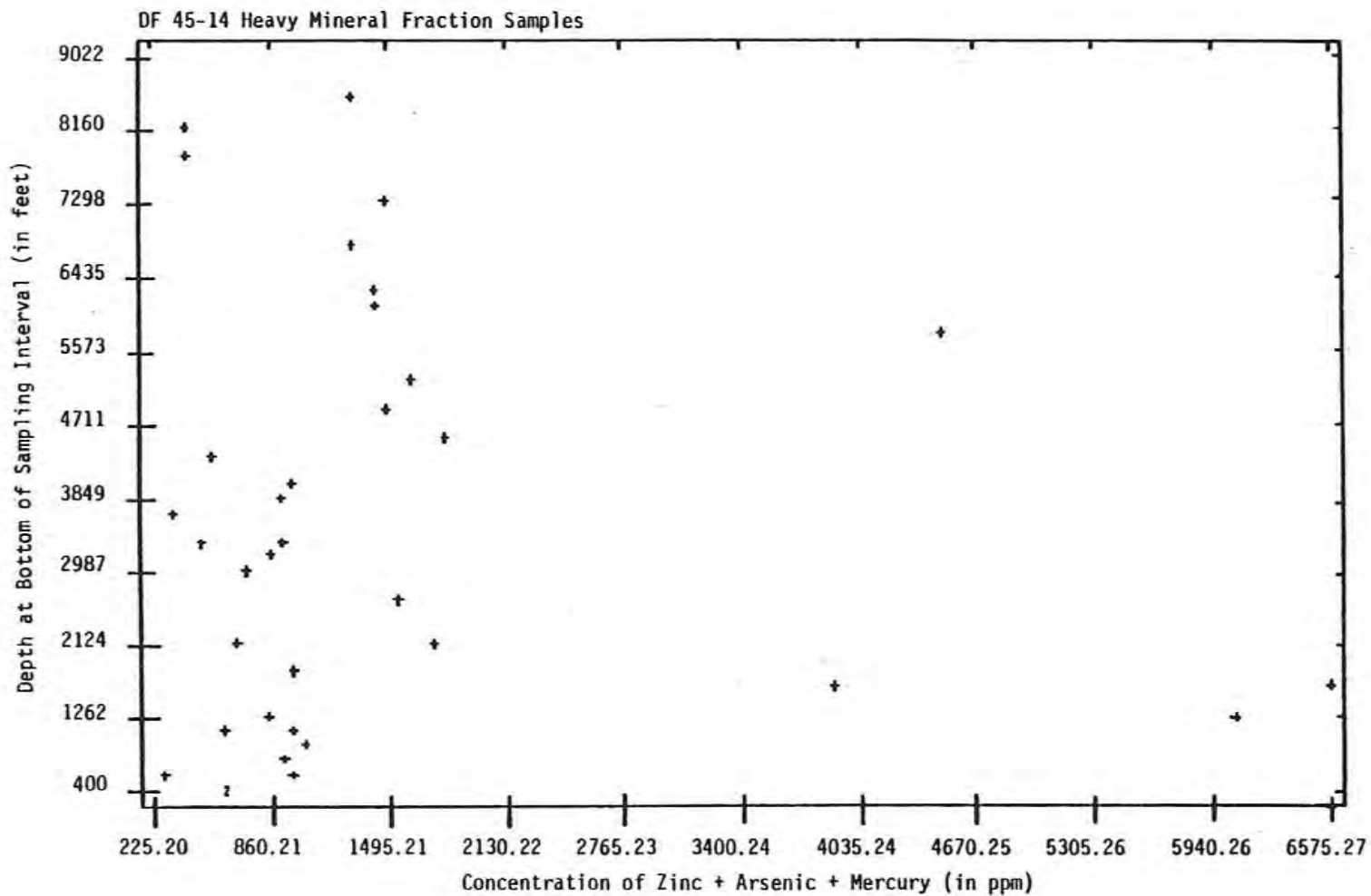


Figure 4-10. Scattergram plot of computed variable (zinc + arsenic + mercury) with respect to depth (as bottom of sampling interval) for DF 45-14 heavy mineral fraction samples.

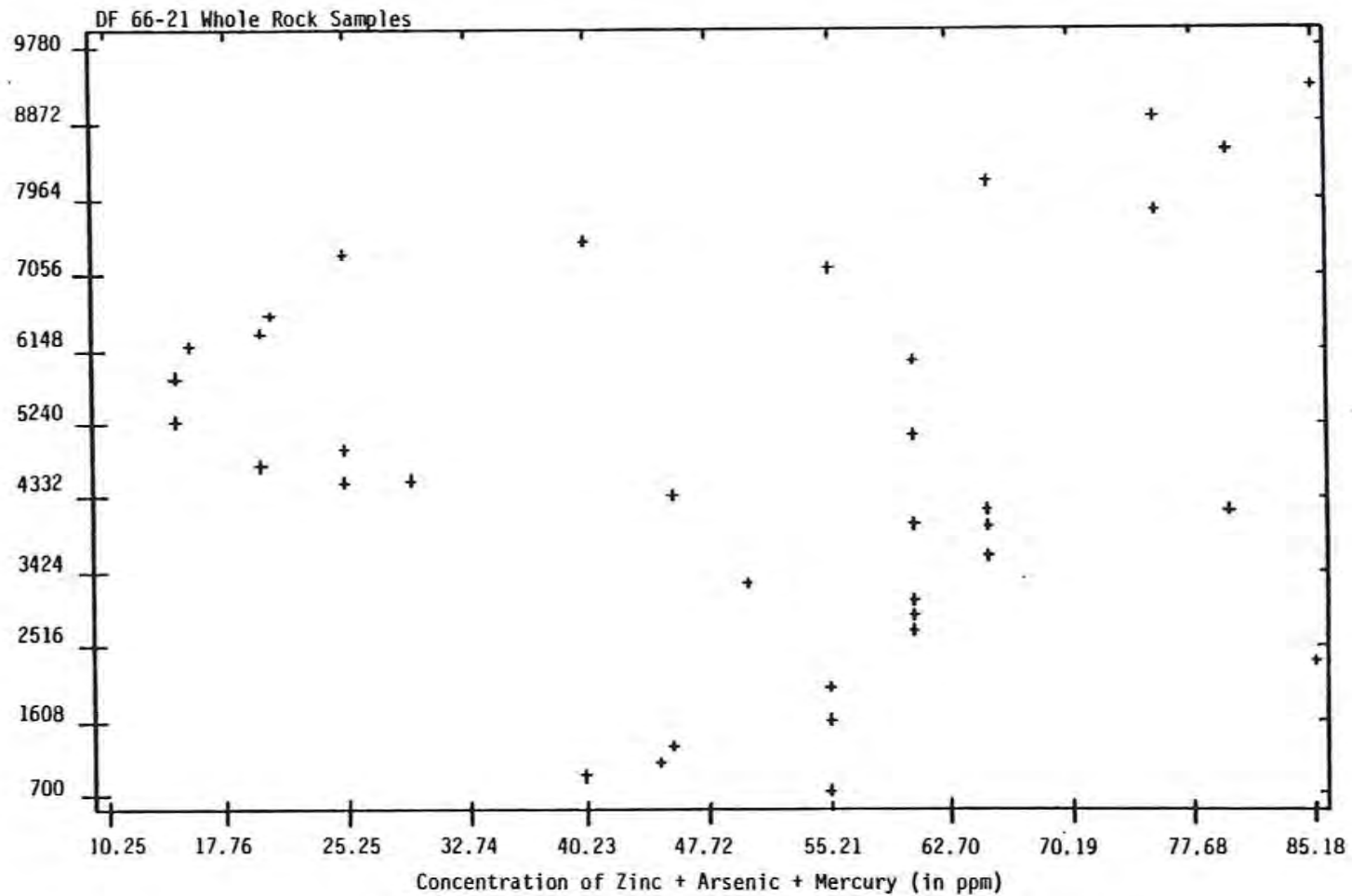


Figure 4-11. Scattergram plot of computed variable (zinc + arsenic + mercury) with respect to depth (as bottom of sampling interval) for DF 66-21 whole rock samples.

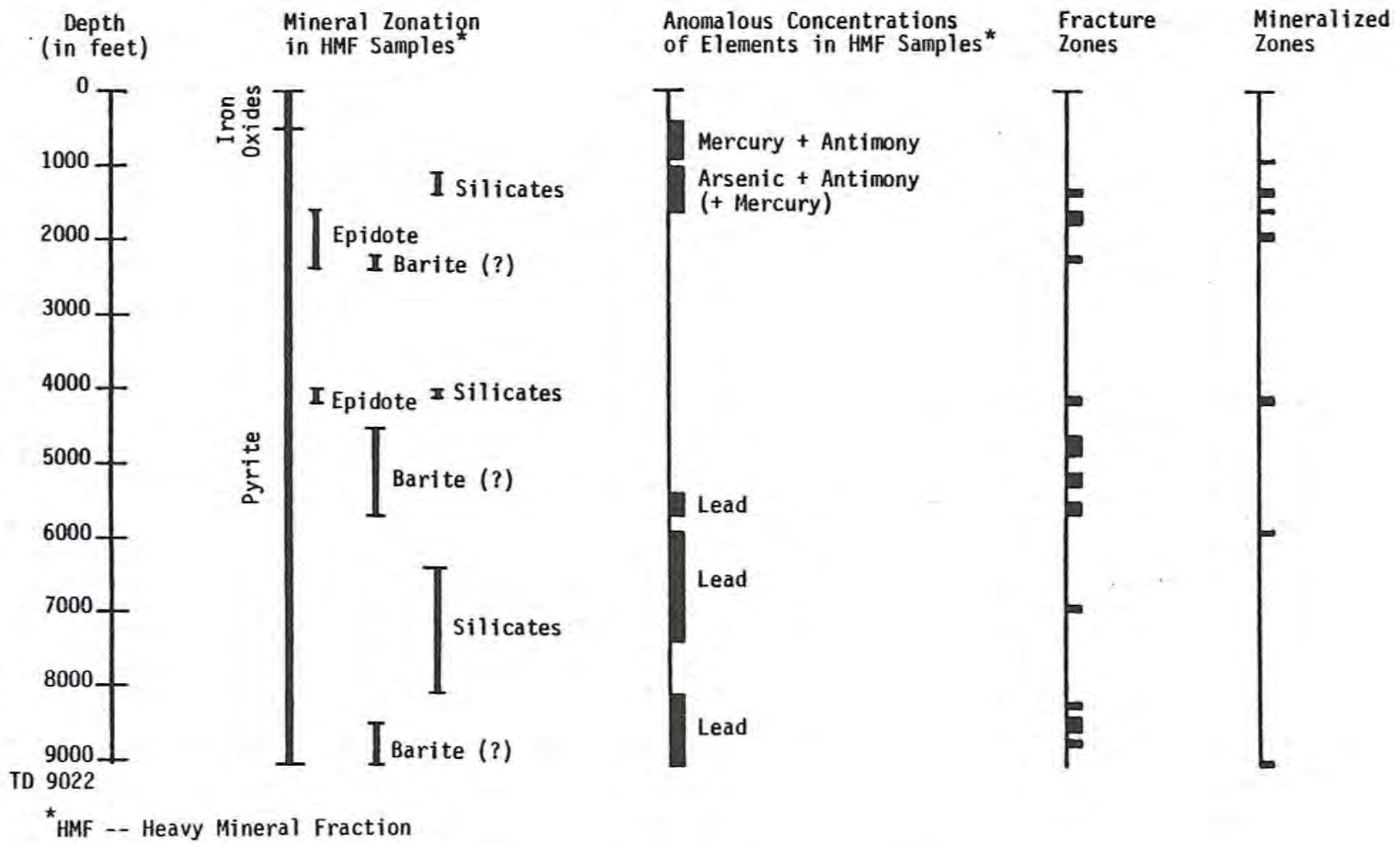


Figure 4-12. Composite of logs for DF 45-14 indicating selected physical parameters, mineral occurrences and anomalous element concentrations in heavy mineral fraction samples.

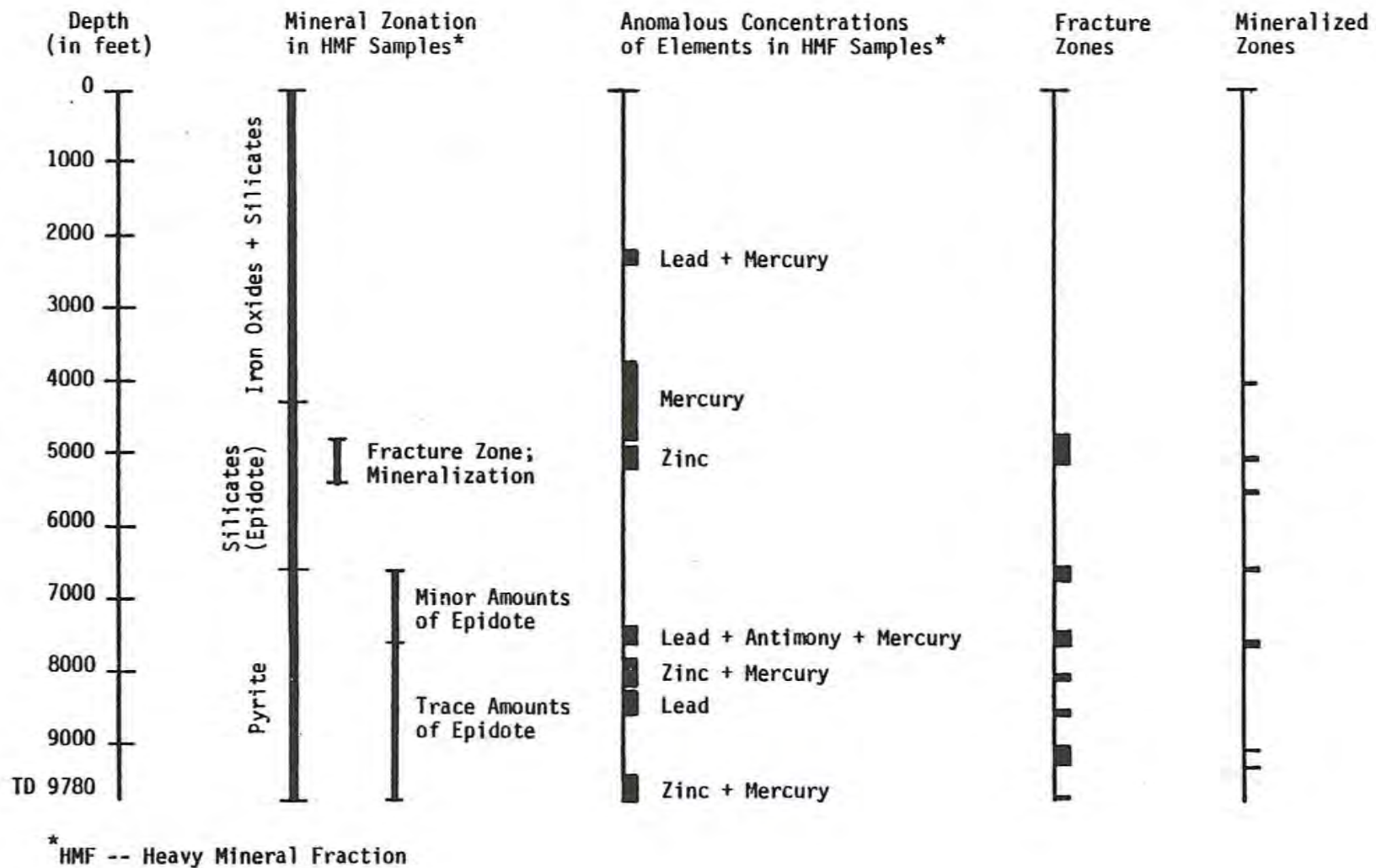


Figure 4-13. Composite of logs for DF 66-21 indicating selected physical parameters and the mineral occurrences and anomalous concentrations of elements in heavy mineral fraction samples.

antimony fairly uniformly distributed throughout the interval. This zone is interpreted as being a product of fracture-controlled permeability with the hydrothermal fluids migrating from the reservoir at depth through fractures associated with the tuffaceous units between 1300 and 1800 feet. Because of the geochemical behavior of the selected trace elements, it is believed that they were probably carried by the fluids through the fracture zone, and deposited in sulfide and silicate +3.3 mineral phases. The mercury, being more volatile and subject to vapor transport, occurs as a halo in the overlying interval (400-900 feet) probably due to adsorption by, or incorporation into, the +3.3 mineral phase.

A second probable influence of the geothermal system in DF 45-14 is reflected in the anomalous concentrations of lead present in the lower portions of the well (Figure 4-12): 700 to 1500 ppm in the interval from 5400 to 7400 feet and 1000 to 2200 ppm in the interval from 8100 to 9022 feet. These lead concentrations are interpreted as lead halos overlying the reservoir or heat source at depth, just as lead halos are similarly associated with ore deposits. This idea is further supported by the presence of barite(?) in fracture zones corresponding to portions of the anomalous lead intervals. The magnitude of the lead anomalies suggests that they probably developed over a relatively long time period, and that the larger anomaly (up to 2200 ppm) from 8100 to 9022 feet may be due to secondary enrichment by halo migration toward the heat source as it cooled. This cooling of the heat source may have been caused by age or by a decrease in connection to the heat source whether from self-sealing of the conduits or possibly from tectonic adjustments that may have closed the conduits. Alternatively, the high lead concentrations may reflect enrichment from fluids migrating through the fracture zones within the interval from 8100 to 9022 feet.

The degree of cementation and fracture-filling within most of the recognized major fault zones and in the numerous minor fractures encountered in the well is also indicative of significant decrease in connection with the reservoir at depth, with the exception of several intervals, most notably from 8100 feet to the bottom of the well.

4.2.4.3 DF 66-21

In contrast, DF 66-21 is interpreted as a more dynamic system with significant connection with the reservoir and a developing geochemical

signature. The anomalous lead (7500 ppm) and mercury (9.0 ppm) concentrations in the interval 2200 to 2400 feet (Figure 4-13) associated with the top of a red clay-alluvial sequence are interpreted as being deposited from aqueous solutions or adsorbed from vapor transport by +3.3 heavy mineral phases at the low-permeability boundary produced by the clay content. The source for the lead and mercury is uncertain; there may be an unrecognized fracture system in the overlying alluvial sequence, or more likely, thermal fluids released in the alluvium may be flowing through the ground water system with downward percolation inhibited by the impermeable clay.

From the lowermost portion of the clay-alluvial sequence to the total depth of the well at 9780 feet, the rocks show an increasingly pervasive influence of the geothermal system: silicate minerals (primarily epidote) with minor pyrite are replaced by a pyrite dominated assemblage. Within the silicate dominated zone is a major active fault system between 4800 and 5400 feet. An anomalous concentration of zinc (10,800 ppm) marks the most active portion of the fault zone where, during drilling, hot fluids under high pressure were encountered. The most significant indication of mineralization in this interval is the presence of flakes of native gold in the heavy mineral fraction sample. A halo of mercury with anomalous concentrations ranging from 6.3 to 39.0 ppm overlies the fault zone, with maximum concentrations occurring in the lowermost portion of the red clay-alluvium sequence. These maximum mercury concentrations are the result of adsorption of the vapor transported mercury by fine-grained +3.3 mineral phases in the base of the clay sequence, with upward migration of the volatile mercury limited by the impermeability of the clay.

Below 5400 feet are four zones containing anomalous element concentrations (Figure 4-13) interpreted as being the result of active migration of thermal fluids through fracture zones. Anomalous lead (11,500 ppm), antimony (1300 ppm) and mercury (9.4 ppm) concentrations are associated with a mineralized fracture zone in the interval 7400 to 7600 feet. A 250 ppm zinc and 3.2 ppm mercury anomaly is associated with a fracture zone identified at 8000 to 8100 feet. The fracture zone encountered at the bottom of the well (9780 feet) is associated with anomalous concentrations of zinc (600 ppm) and mercury (4.6 ppm). The range of concentrations for lead (4000 to 11,500 ppm), zinc (250 to 10,800 ppm), and

mercury (3.2 to 39 ppm) is believed to be related to possible variations in fluid compositions and may also reflect the significance of the fracture zone as a conduit; i.e., the more open a fracture zone, the greater volume of fluid it conducts and the greater the concentration of elements. For example, the mineralized fracture zone at 6500 to 6750 feet exhibits no anomalous element concentrations and is basically closed by calcite fracture-filling and veining.

4.3 Conclusions and Recommendations

4.3.1 Significance of the Interpretations

Both reservoir zoning and petrochemical analysis can be used as methods of interpreting the trace element geochemistry exhibited by DF 45-14 and DF 66-21 in Dixie Valley. In DF 45-14, distinct peripheral and inner and outer self-sealed zones can be identified based on general trends and corresponding concentrations of the trace element geochemistry. The outer self-sealed zone is approximately 4000 feet thick and the inner self-sealed zone extends from about 8500 feet to beyond the bottom of the well (9022 feet). Self-sealing has occurred with either partial or complete sealing of localized zones within the peripheral and self-sealed zones of DF 45-14.

In DF 66-21, the peripheral zone is distinct and extends to a depth of approximately 4300 feet. Below this depth lies the self-sealed zone with the boundary between the outer and inner self-sealed zones being gradational and poorly defined within the interval from 4300 to 7500 feet where the inner self-sealed zone begins encroaching upon the reservoir at depth. This encroachment zone extends beyond the bottom of the well (9780 feet). Self-sealing of localized zones is generally restricted to minor fracture systems, with most major fracture systems still serving as active fluid conduits.

By applying petrochemical analysis, relationships between the trace element geochemistry and the physical characteristics of the well, in particular the locations of fracture and mineralized zones, can be defined. In DF 45-14, two diverse effects of the geothermal system are identified: the first is a fracture-controlled arsenic-mercury concentration in the upper portion of the well and the second is a broad halo of lead concentrations overlying the reservoir, with possible secondary enrichment as a result of fracture-controlled fluid migration.

In DF 66-21 a number of zones can be defined with various combinations of anomalous trace elements; these are essentially fracture-controlled. Most of the fractures are still active conduits for the thermal fluids with the magnitude of the anomalous concentration values possibly related to the fluid composition or to the significance of the conduit.

The reservoir zoning and the petrochemical analysis techniques generate models of the two deep wells that are generally consistent and significant with respect to developing an understanding of the history of the Dixie Valley Geothermal System. In general, DF 45-14 represents a longer history of geothermal activity: the reservoir zones are more distinct and extend over larger depth intervals; self-sealing is more complete; a halo of anomalous lead concentrations has developed in the lower portions of the well that, while it may be secondarily enriched by fracture-controlled fluids, is a pervasive concentration; and fluid migration through fracture systems is limited. DF 66-21 exhibits a younger, more dynamic history of geothermal activity: while the peripheral zone is well defined, the self-sealed zone extends over a relatively narrow depth interval and has not yet developed geochemical signatures which differentiate well-defined outer and inner self-sealed zones; the inner self-sealed zone encroaches on the reservoir at depth; fracture zones are generally not self-sealed and actively conduct significant volumes of thermal fluids; and anomalous trace element concentrations exist in fairly narrow intervals associated primarily with active fracture systems.

This interpretation is based solely on the present level of knowledge of the two deep wells. It can not be over-emphasized that DF 45-14 and DF 66-21 represent a very limited sampling of the Dixie Valley Geothermal System. With the collection of additional data as suggested in the following section or with refinement of the data presently available, these interpretations can and should be re-evaluated.

4.3.2 Recommendations

- 1) Core samples should be taken at selected intervals in the wells. This would provide definitive identification of the specific mineral assemblages and the inter-granular relationships among the various mineral phases, and would allow for estimation of the amount of mixing and contamination that affects drill chip samples.

2) Other trace elements in the samples should be examined. Based on the models of Bamford and others (1980) for Roosevelt Hot Springs, KGRA, manganese may prove useful. In addition, since the thermal waters in DF 45-14 and DF 66-21 contain relatively high concentrations of boron (Bohm and others, 1980) it might prove to be a useful indicator element for the Dixie Valley system. Since boron is not selectively concentrated in sulfide phases, whole rock sampling may prove sufficient.

3) While an examination and interpretation of the trace element geochemistry of DF 45-14 and DF 66-21 has proven quite useful in understanding these portions of the reservoir and confirming prior data, testing of the technique should continue. Other methods, besides reservoir zoning and petrochemical analysis, should be developed for interpreting this type of data as it is applied to new wells drilled in Dixie Valley.

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Chapter 5. DIXIE VALLEY GEOTHERMAL SYSTEM

By: Elaine J. Bell

5.0 DIXIE VALLEY GEOTHERMAL SYSTEM

5.1 Introduction

The purpose of the MMRI program of investigation was to develop a model for the geothermal system in Dixie Valley. An initial model based on existing data, was modified and refined to develop an integrated model incorporating data derived from four fields of study: structure-tectonics, petrology, hydrology and hydrogeochemistry, and shallow temperature survey (Mackay Minerals Research Institute, 1980). With the development of new data derived from soil geochemistry and petrochemistry, as presented in the preceding chapters of this report, the integrated model of the Dixie Valley Geothermal System can be re-evaluated.

A brief summary of the integrated model, an evaluation of the model based on the soil geochemistry and petrochemistry and recommendations for investigations considered necessary for further refinement and verification of the model are presented in the following sections.

5.2 Integrated Model of the Dixie Valley Geothermal System

The Dixie Valley Geothermal System appears to be the result of regionally high mantle heat flow through a thin crust, with preferential conduits for fluid migration structurally controlled by major basement faults and by Basin and Range extensional faults. Figure 5-1 depicts the spatial relationships among the major structural elements in Dixie Valley, with a generalized cross-section of the model shown in Figure 5-2. Production zones may be found in association with any of several lithologies, such as the Humboldt gabbroic complex or Triassic metasediments, if there is communication by fracture or fault zones with the reservoir at depth and sufficient fluid to transport the heat. A detailed discussion of the model is presented by Bell and others (1980a).

5.3 Evaluation of the Integrated Model

The data and conclusions derived from the soil geochemistry and petrochemistry are generally consistent with the integrated model of the Dixie Valley Geothermal System. Mercury and arsenic soil geochemical distributions confirm the location, trend and significance of known structures as conduits for migration of thermal fluids to the surface. In addition, geochemical trends also suggest the location and orientation of suspected or inferred structures within the study area that have

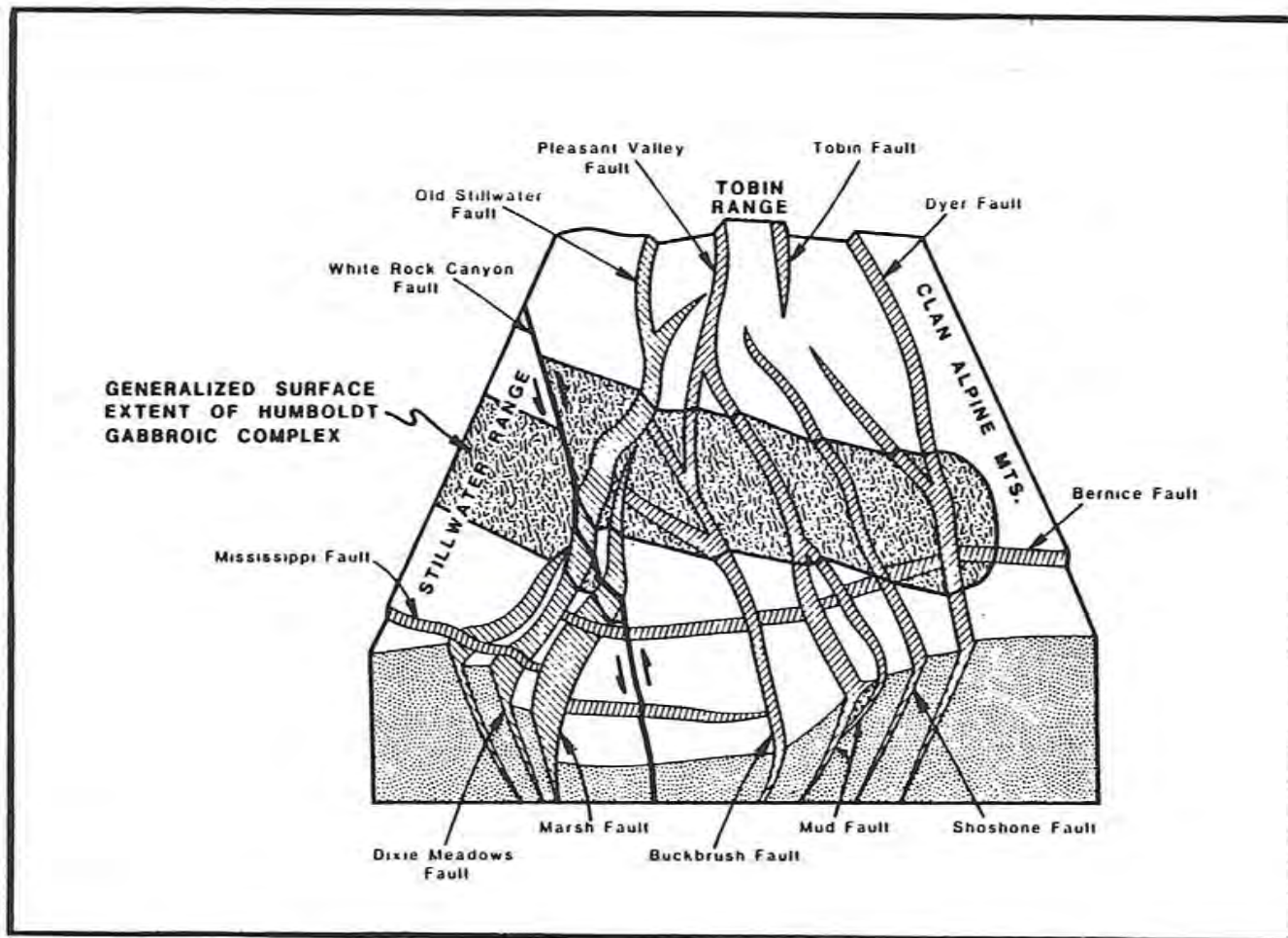


Figure 5-1. Three dimensional view of integrated model of the Dixie Valley Geothermal System.

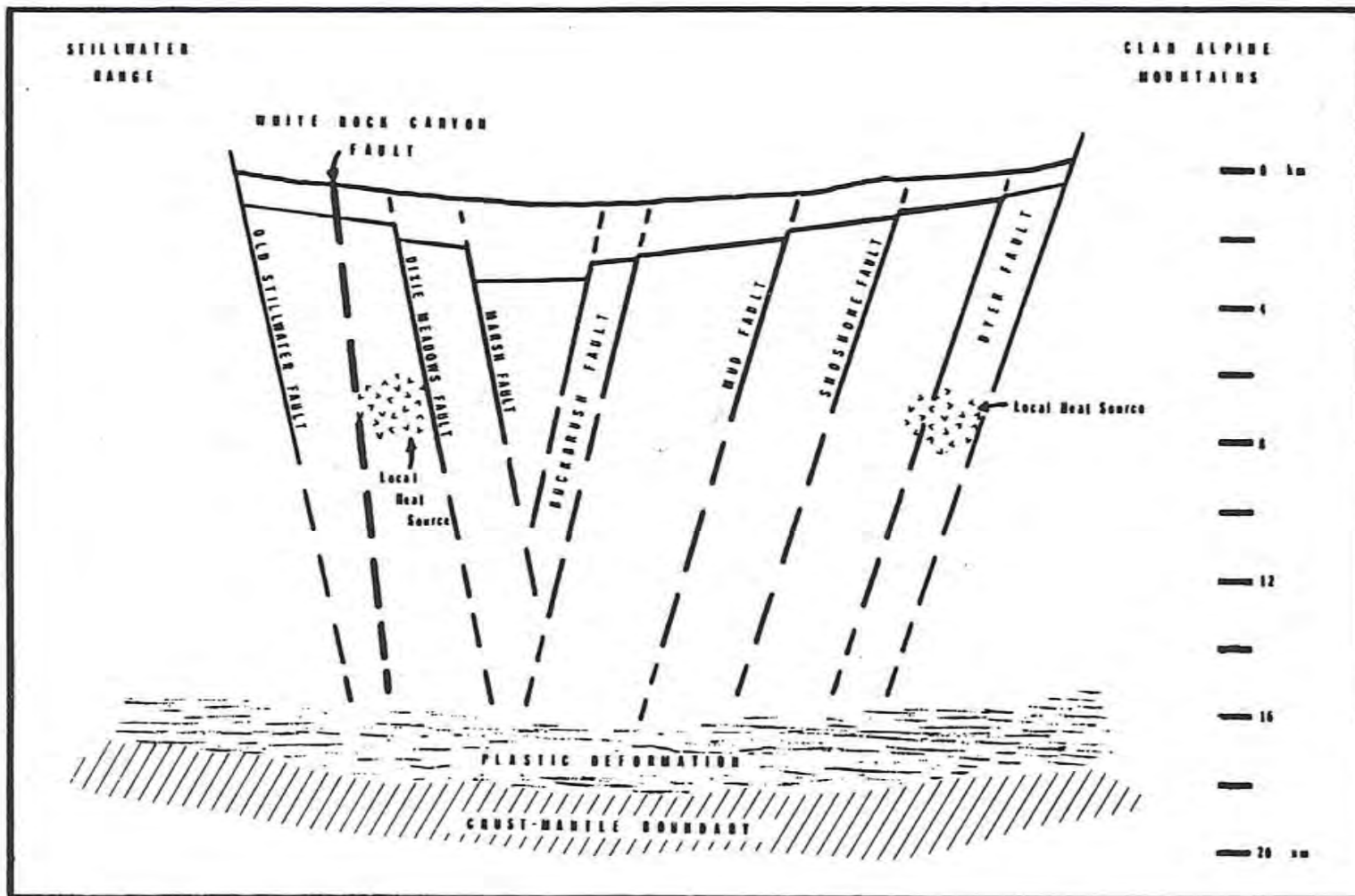


Figure 5-2. Generalized east-west cross-section of the integrated model of the Dixie Valley Geothermal System. Structural elements (faults) are propagated downward from the surface through the valley fill (alluvium with intercalated volcanics) and into the underlying undifferentiated bedrock and then projected through the zone of brittle failure.

significance with respect to proven production areas or natural surface manifestations of the geothermal system.

The general distribution of selected trace elements (reservoir zoning) and the specific zones distinguished by anomalous trace element concentrations (petrochemical analysis) confirm the significance of structural controls of fluid migration in the subsurface. Additionally, the petrochemistry suggests a possible interpretation of the dynamics of the reservoir. DF 45-14 is located in an area with a longer history of geothermal activity that has resulted in reduced interaction with and connection to the reservoir. DF 66-21, however, exhibits a younger, more dynamic history with significant fracture zones actively conducting thermal fluids migrating from the reservoir at depth.

5.4 Recommendations

Recommendations for further studies necessary to test and refine the model of the Dixie Valley Geothermal System have previously been made (Bell and others, 1980b). The importance of conducting these types of investigations is emphasized here, along with the recommendation that both soil geochemical and petrochemical studies be continued and expanded in Dixie Valley as new wells are drilled.

5.5 References

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Appendix A. SOIL GEOCHEMICAL DATA

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
1	104	25
2	56	45
3	96	15
4	40	30
5	32	35
6	28	65
7	72	10
8	64	15
9	96	10
10	60	15
11	68	25
12	48	60
13	8	40
14	16	45
15	88	15
16	56	15
17	116	20
18	392	50
19	76	20
20	16	40
21	24	15
22	8	45
23	56	25
24	60	20
25	176	20
26	976	55
27	488	40
28	108	25
29	168	30
30	80	35
31	8	25
32	16	30
33	60	40
34	28	15
35	152	30
36	232	35
37	272	25
38	2120	80
39	180	30
40	88	20
41	16	85
42	20	10
43	4	15
44	24	25
45	8	30
46	24	15
47	40	25
48	56	25
49	64	30
50	56	35
51	104	25

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
52	36	25
53	76	15
54	16	30
55	68	15
56	76	25
57	44	15
58	8	15
59	28	20
60	16	15
61	8	15
62	8	20
63	16	15
64	28	5
65	28	10
66	28	10
67	44	15
68	44	15
69	52	15
70	64	20
71	108	10
72	36	10
73	32	25
74	36	20
75	16	15
76	24	25
77	24	20
78	4	10
79	144	25
80	72	15
81	80	10
82	156	20
83	96	10
84	12	20
85	116	15
86	128	25
87	60	10
88	108	10
89	188	15
90	120	20
91	68	15
92	80	10
93	88	15
94	80	15
95	116	15
96	88	15
97	52	10
98	376	30
99	120	15
100	24	10
101	96	15
102	40	10

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
103	132	15
104	36	15
105	40	15
106	68	10
107	44	10
108	28	15
109	4	20
110	108	10
111	24	5
112	92	15
113	12	5
114	8	10
115	8	15
116	32	15
117	48	25
118	36	10
119	32	5
120	32	5
121	24	5
122	16	5
123	108	10
124	84	10
125	52	10
126	24	5
127	12	5
128	32	5
129	12	80
130	12	35
131	8	35
132	28	5
133	72	5
134	20	5
135	24	5
136	12	20
137	48	15
138	84	10
139	52	10
140	28	10
141	8	20
142	8	25
143	12	40
144	12	45
145	36	15
146	20	15
147	60	5
148	92	35
149	132	15
150	44	10
151	68	10
152	68	5
153	64	20

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
154	24	20
155	24	25
156	16	10
157	12	50
158	32	25
159	104	15
160	272	10
161	84	15
162	136	25
163	56	10
164	32	5
165	120	25
166	140	45
167	64	25
168	20	25
169	28	25
170	12	15
171	8	15
172	28	10
173	164	15
174	232	25
175	104	20
176	92	15
177	148	20
178	60	10
179	152	1
180	100	20
181	68	35
182	8	10
183	16	15
184	16	10
185	8	10
186	12	15
187	24	20
188	24	15
189	12	20
190	28	15
191	56	10
192	56	10
193	156	10
194	68	15
195	84	5
196	32	5
197	16	10
198	36	5
199	36	10
200	96	10
201	64	5
202	40	5
203	36	10
204	48	15
205	28	15

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
206	40	30
207	28	25
208	176	5
209	48	10
210	36	5
211	56	5
212	120	10
213	40	35
214	120	10
215	56	15
216	88	15
217	40	10
218	40	10
219	24	10
220	0	5
221	28	15
222	40	15
223	16	35
224	60	25
225	80	10
226	76	5
227	112	35
228	100	5
229	104	5
230	68	5
231	56	5
232	108	5
233	104	10
234	120	10
235	56	5
236	168	10
237	48	15
238	40	15
239	48	15
240	40	10
241	36	10
242	40	25
243	32	15
244	28	15
245	32	10
246	40	5
247	56	15
248	44	10
249	28	10
250	12	5
251	20	40
252	28	15
253	36	10
254	72	20
255	24	10
256	120	15
257	156	25

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
258	132	15
259	164	10
260	100	10
261	120	15
262	160	15
263	116	15
264	160	10
265	124	10
266	108	15
267	140	10
268	160	15
269	176	10
270	160	15
271	144	20
272	68	15
273	52	40
274	40	15
275	72	15
276	48	15
277	40	15
278	72	35
279	48	20
280	48	15
281	56	30
282	56	15
283	36	15
284	64	15
285	40	40
286	40	115
287	156	35
288	4	15
289	56	30
290	48	30
291	32	10
292	76	25
293	72	35
294	24	5
295	32	10
296	32	15
297	44	15
298	88	20
299	96	20
300	56	10
301	108	25
302	56	25
303	12	10
304	32	15
305	32	10
306	0	10
307	48	10
308	160	15
309	68	15

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
310	28	10
311	36	15
312	16	15
313	88	10
314	40	10
315	32	15
316	40	10
317	44	15
318	72	25
319	36	15
320	96	15
321	48	10
322	76	15
323	212	15
324	6280	10
325	1080	20
326	1400	25
327	12760	15
328	400	50
329	1720	35
330	1240	10
331	388	15
332	364	10
333	48	35
334	248	15
335	76	25
336	64	10
337	40	10
338	32	10
339	84	10
340	24	10
341	40	15
342	80	15
343	36	15
344	72	15
345	28	35
346	88	25
347	40	10
348	20	10
349	88	10
350	44	40
351	156	15
352	84	15
353	116	15
354	120	15
355	60	15
356	204	15
357	116	15
358	56	25
359	56	15
360	168	5

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
361	560	5
362	84	15
363	44	5
364	220	15
365	40	15
366	56	10
367	92	10
368	96	10
369	40	10
370	40	5
371	48	5
372	72	15
373	104	10
374	80	15
375	288	10
376	68	10
377	96	15
378	56	15
379	64	5
380	284	10
381	256	10
382	188	20
383	480	15
384	64	85
385	60	15
386	296	15
387	40	15
388	64	15
389	64	5
390	32	10
391	56	10
392	28	10
393	48	10
394	40	15
395	52	15
396	48	10
397	72	15
398	48	10
399	68	10
400	28	10
401	16	10
402	112	15
403	64	10
404	32	10
405	56	10
406	256	225
407	76	5
408	240	40
409	16	15
410	32	15
411	60	5
412	48	10

<u>Sample #</u>	<u>Hg ppb</u>	<u>As ppm</u>
413	56	10
414	40	10
415	28	10
416	36	10
417	36	10
418	12	10
419	40	10
420	68	10
421	28	10
422	24	10
423	24	10
424	76	10
425	16	10
426	16	10
427	36	10
428	44	10
429	16	10
430	84	15
431	64	10
432	104	15
433	56	10
434	16	10
435	68	15
436	40	10
437	96	10
438	12	10
439	40	20
440	40	10
441	56	15
442	56	20
443	112	35
444	48	10
445	36	15
446	4	1
447	0	1
448	36	5
449	16	1
450	84	15
451	92	15
452	184	10
453	68	10
454	104	15
455	172	20
456	104	15
457	56	15
458	172	15
459	72	10
460	52	25
461	232	25
462	284	35
463	172	30
464	216	25

Appendix B. PETROCHEMICAL DATA

Table 4-12
Lithologic Symbols

Symbol	Lithologic Symbol
A	Alluvium
A/c	Alluvium with clay
T	Tuff
A/B	Andesite/Basalt
G	Granodiorite Intrusive
QA	Quartz Arenite
M/M	Metasiltstone/Metashale
G/D	Gabbro/Diorite Intrusive
MZ	Mineralized Zone
FZ	Fault Zone

Table 4-13
Lead Distribution in DF 45-14 Heavy Mineral Fractions

Lead (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
25	2	4.2	A
50	4	14.4	A; T; G/D
75	8	34.7	T; A/B; A; M/M
100	2	39.8	T; A/B
125	2	44.9	M/M; A; G/D
150	4	55.1	M/M
175	1	57.6	M/M
200	2	62.7	M/M; FZ
250	1	65.2	M/M
325	2	70.3	M/M; QA; G/D; FZ
425	2	75.4	M/M; A; G/D; FZ
475	2	80.5	M/M; T; A/B; FZ
680	1	83.0	M/M; G/D
2200	3	90.7	M/M; QA; G/D; T; FZ
2500	1	93.2	QA; M/M; FZ
3000	1	95.8	QA; M/M; FZ
3700	1	98.3	M/M; G/D; FZ

Mean 555.2 Std Err 149.9 Std Dev 936.0 Median 146.9
Kurtosis 3.6 Skewness 2.2

*Refer to Table 4-12 for explanation of symbols.

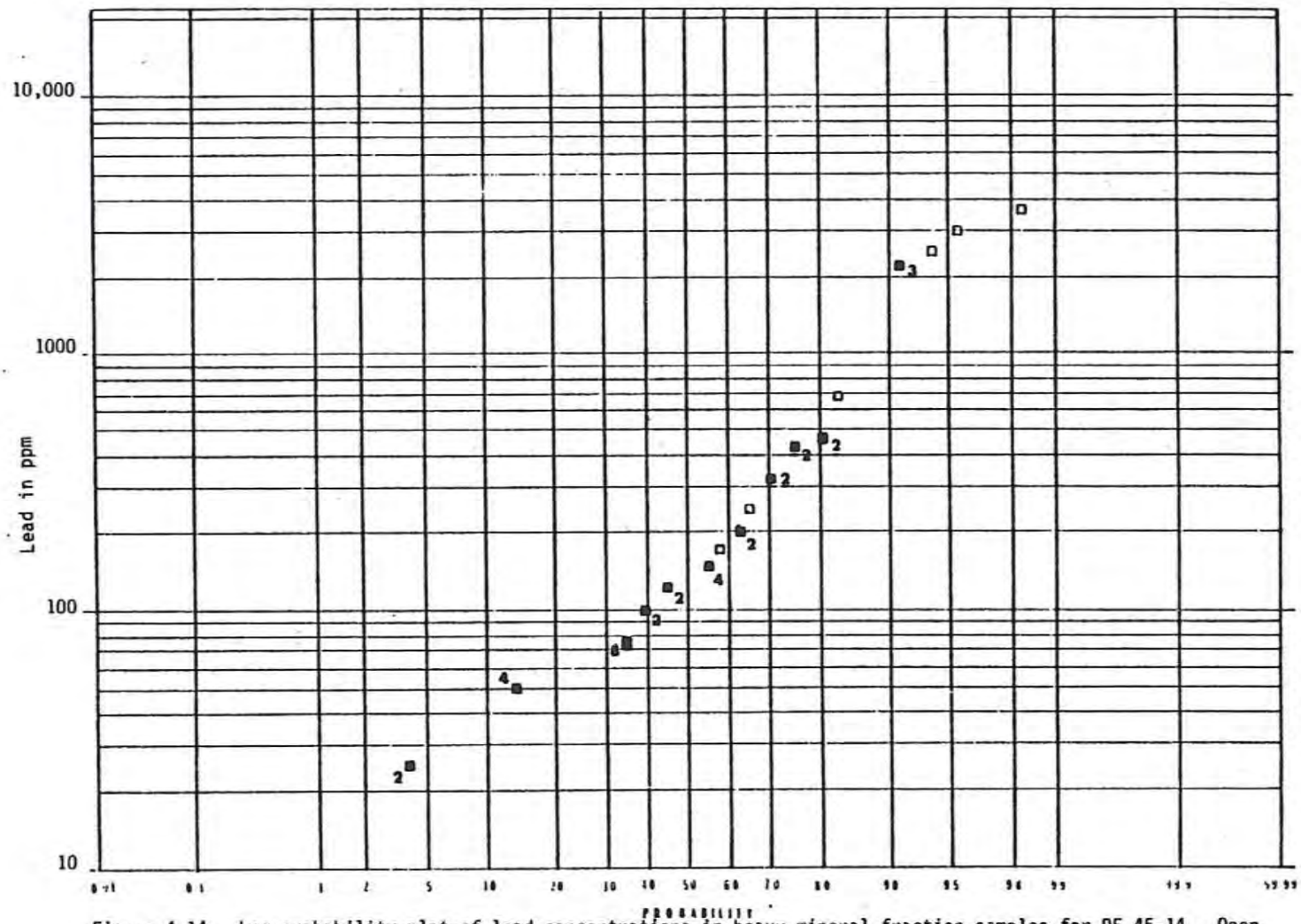


Figure 4-14. Log probability plot of lead concentrations in heavy mineral fraction samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-14

Zinc Distribution in DF 45-14 Heavy Mineral Fractions

Zinc (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
25	3	6.8	A; T
50	2	11.9	A; T; M/M
75	6	27.1	T; A/B; A; M/M
100	3	34.7	M/M; T; A/B
125	2	39.8	A; G/D; M/M
150	2	44.9	T; A/B; G/D
175	4	55.1	M/M; QA; G/D
200	2	60.2	M/M; G/D; FZ
300	2	65.2	M/M; QA; G/D; FZ
325	1	67.8	M/M; G/D
375	2	72.9	A
400	1	75.4	A
600	1	77.9	M/M; FZ
650	1	80.5	M/M; T
800	2	85.6	M/M; FZ
875	1	88.1	QA; M/M; FZ
1200	2	93.2	M/M; FZ
2000	1	95.8	T; A/B
4100	1	98.3	M/M; FZ

Mean 432.7 Std Err 117.0 Std Dev 730.8 Median 171.9
Kurtosis 17.1 Skewness 3.8

*Refer to Table 4-12 for explanation of symbols.

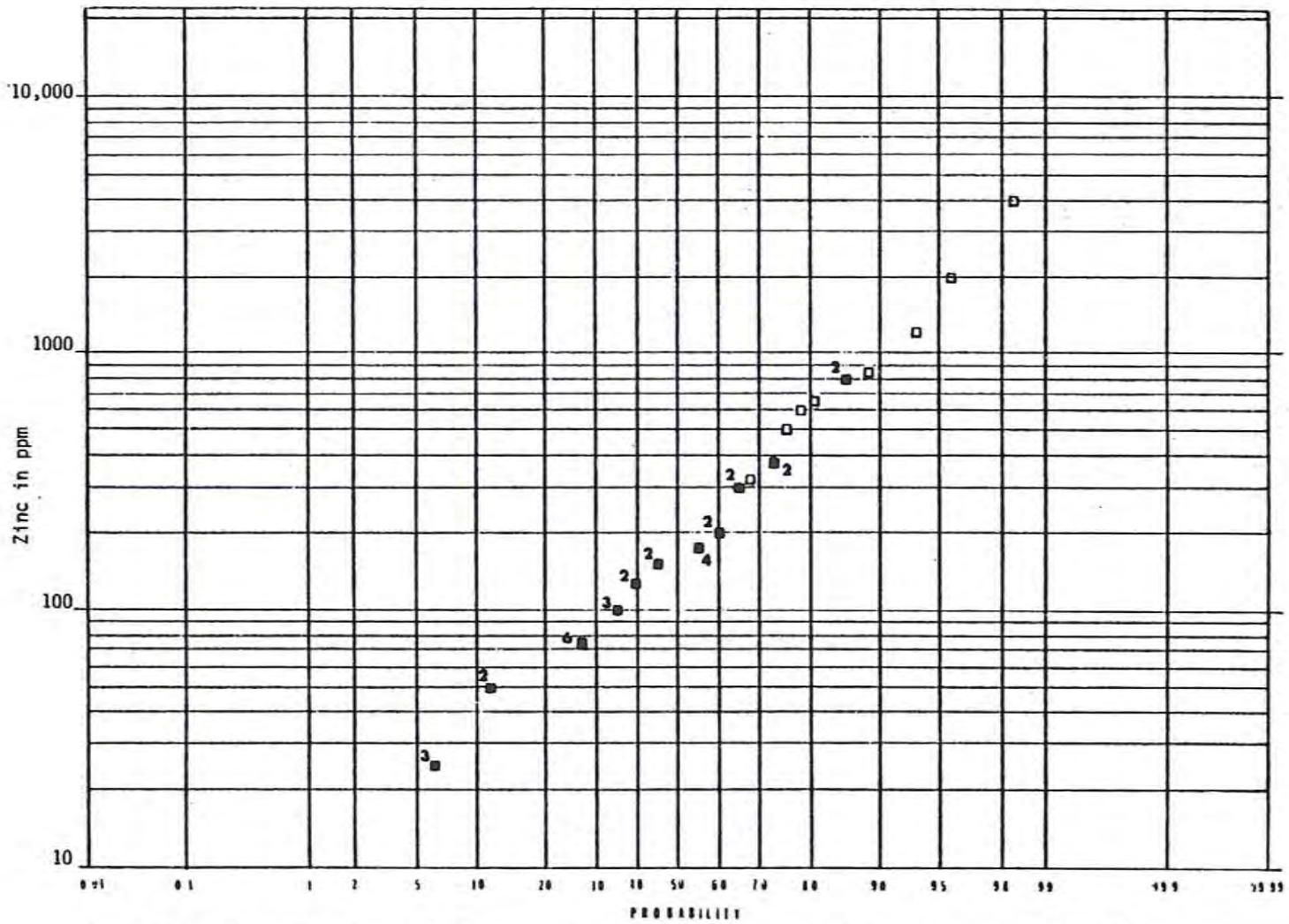


Figure 4-15. Log probability plot of zinc concentrations in heavy mineral fraction samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-15

Arsenic Distribution in DF 45-14 Heavy Mineral Fractions

Arsenic (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
125	1	1.7	G/D
175	1	4.2	M/M
200	3	11.9	M/M; A; FZ
220	2	16.9	A
225	1	19.5	M/M; G/D
250	1	22.0	M/M; QA; G/D
275	1	24.6	QA; T; FZ
400	2	29.7	M/M; G/D; FZ
420	1	32.2	M/M
450	1	34.7	QA; M/M
480	1	37.3	QA; M/M; FZ
500	1	39.8	T; A/B
520	1	42.2	M/M; FZ
550	1	44.9	A
580	1	47.4	T; A/B
600	1	50.0	M/M; QA; G/D
620	1	52.5	A
650	2	57.6	M/M
750	1	60.2	M/M
780	3	67.8	A; M/M; T
800	1	70.3	M/M; G/D
880	1	72.9	T; A/B
900	1	75.4	M/M; T
920	1	77.9	A
1000	2	83.0	T; A
1300	1	85.6	M/M; G/D; FZ
1500	1	88.1	M/M; G/D
1600	1	90.7	T; A/B
3800	1	93.2	T; A/B
6000	1	95.8	T; A
6500	1	98.3	T; A/B; A

Mean 979.5 Std Err 222.5 Std Dev 1389.5 Median 600.0
Kurtosis 10.2 Skewness 3.2

*Refer to Table 4-12 for explanation of symbols.

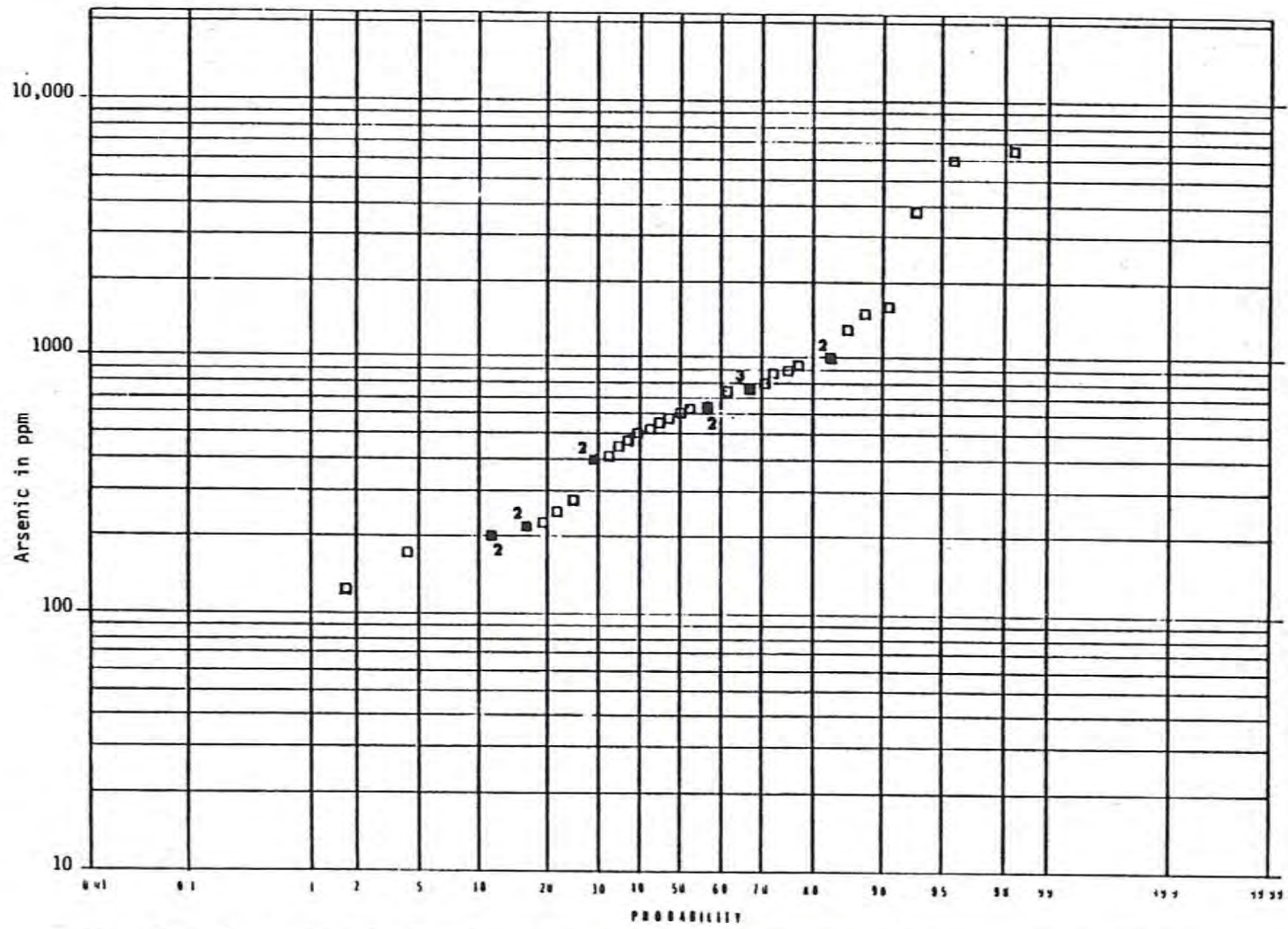


Figure 4-16. Log probability plot of arsenic concentrations in heavy mineral fraction samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-16

Antimony Distribution in DF 45-14 Heavy Mineral Fractions

Antimony (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
2	7	17.4	M/M; QA; G/D; FZ
4	1	20.0	M/M; QA
6	1	22.6	M/M; QA; T
8	1	25.2	M/M
12	2	30.4	M/M; FZ
14	3	38.3	M/M; G/D; A
16	2	43.5	M/M; A; FZ
20	3	51.3	M/M; G/D; FZ
22	2	56.5	M/M; A; G/D
24	1	59.1	T; A/B
32	1	61.7	M/M; T
36	1	64.3	M/M; G/D
46	1	66.9	T; A/B
74	1	69.5	T; A/B
104	1	72.2	A
106	1	74.8	T; A/B
192	1	77.4	A
214	1	80.0	T; A/B; A
338	2	85.2	A; T; A/B
384	1	87.8	A; T
420	1	90.4	QA; M/M; FZ
500	1	93.0	A; T
514	1	95.7	A
836	1	98.3	T

Mean 115.6 Std Err 31.3 Std Dev 192.9 Median 20.3
 Kurtosis 4.5 Skewness 2.1

*Refer to Table 4-12 for explanation of symbols.

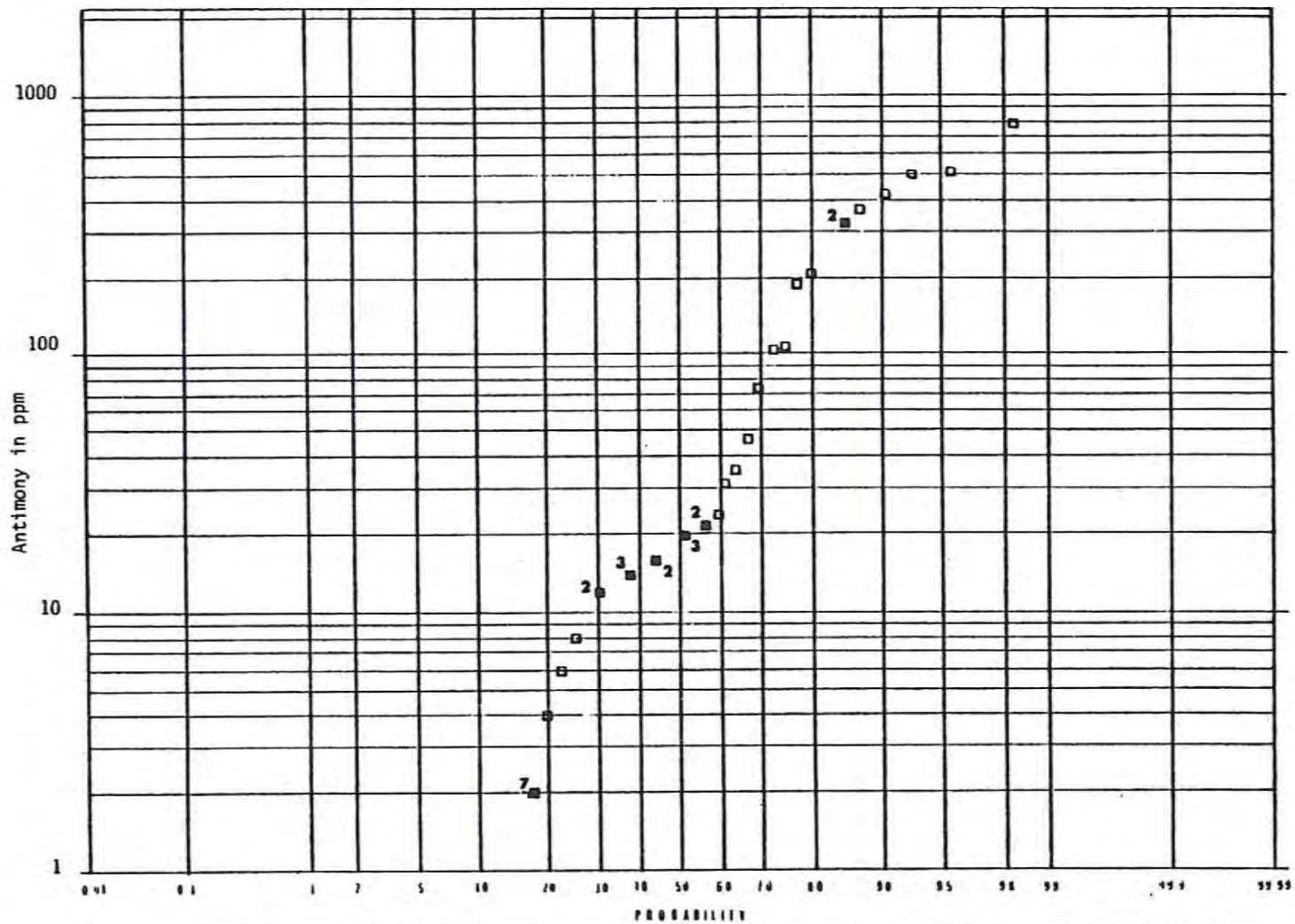


Figure 4-17. Log probability plot of antimony concentrations in heavy mineral fraction samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-17

Mercury Distribution in DF 45-14 Heavy Mineral Fractions

Mercury (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
0.01	7	17.9	A; M/M
0.1	1	20.5	A
0.125	1	23.2	A
0.185	3	31.3	G/c; M/M; FZ
0.200	1	33.9	M/M
0.235	1	36.6	M/M
0.270	3	44.6	M/M; A; T
0.360	1	47.3	M/M; G/D
0.375	1	50.0	T; A/B
0.380	1	52.7	M/M; G/D; FZ
0.395	1	55.4	M/M; FZ
0.505	1	58.0	M/M; QA; G/D
0.570	1	60.7	T; A/B
0.585	1	63.4	M/M; G/D; T; FZ
0.965	1	66.1	M/M; QA; T
1.1	1	68.8	T; A/B
1.4	1	71.4	QA; M/M; FZ
1.6	2	76.8	M/M; G/D
1.7	1	79.5	A
1.8	1	82.1	QA; T; FZ
2.1	1	84.8	A; T
4.8	1	87.5	A
7.4	1	90.2	A
9.1	1	92.9	T; A/B; A
13.8	1	95.5	T
16.7	1	98.2	A; T

Mean 1.9 Std Err 0.6 Std Dev 3.8 Median 0.4
Kurtosis 8.1 Skewness 2.9

*Refer to Table 4-12 for explanation of symbols.

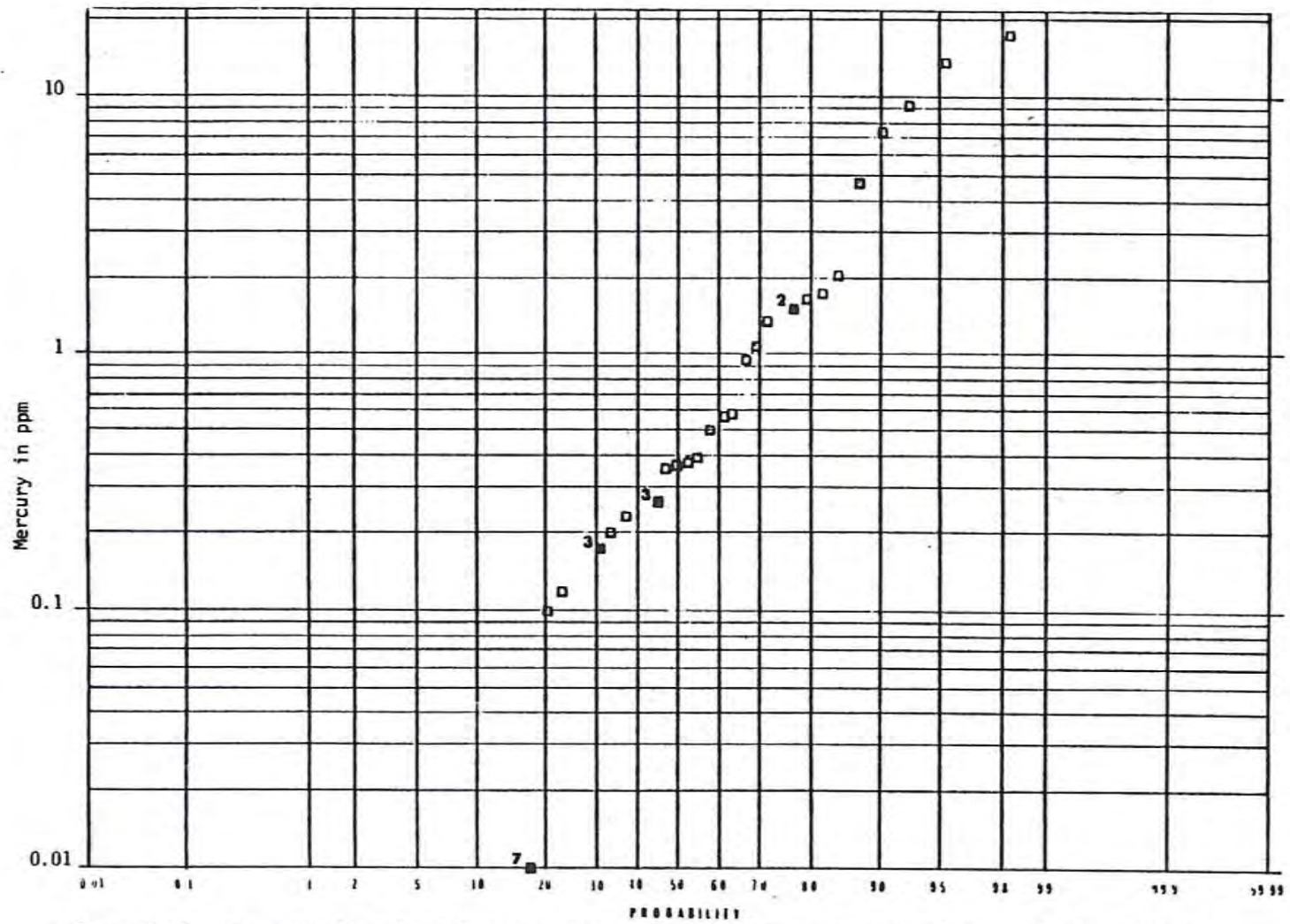


Figure 4-18. Log probability plot of mercury concentrations in heavy mineral fraction samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-18

Lead Distribution in DF 45-14 Whole Rock Samples

Lead (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
5	1	1.8	T
10	6	17.9	T; A/B; A; M/M
15	11	47.3	M/M; A; T; A/B
20	4	58.0	M/M; A; G/D; FZ
25	6	74.1	M/M; QA; G/D; T; A/B
30	4	84.8	M/M; QA; G/D; FZ
35	2	90.2	M/M; A; G/D
40	1	92.9	M/M; G/D; FZ
45	1	95.5	M/M; QA/ T
115	1	98.3	QA

Mean 22.9 Std Eff 2.9 Std Dev 18.1 Median 18.1

Kurtosis 18.8 Skewness 3.8

*Refer to Table 4-12 for explanation of symbols.

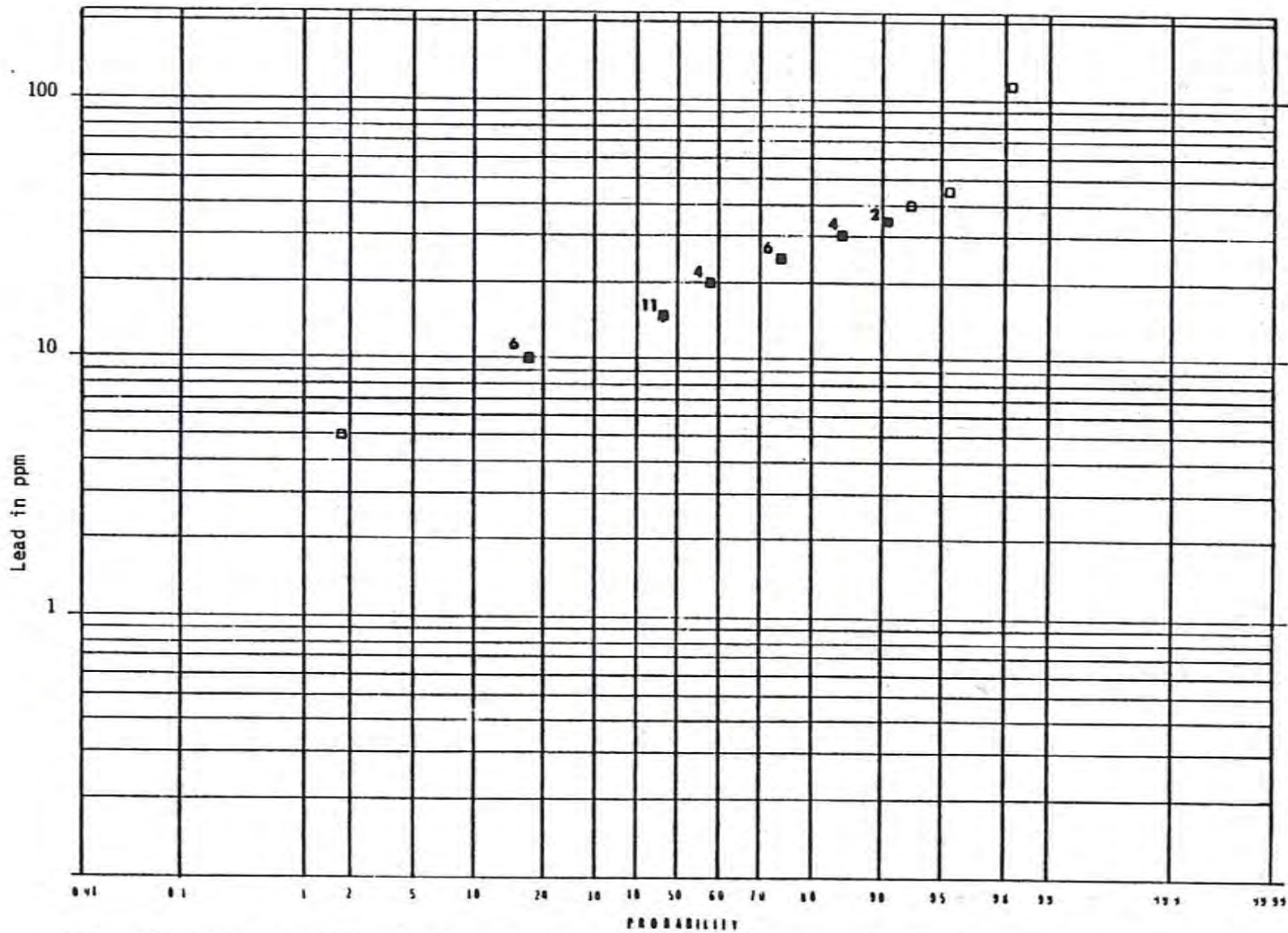


Figure 4-19. Log probability plot of lead concentrations in whole rock samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-19
Zinc Distribution in DF 45-14 Whole Rock Samples

Zinc (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
20	3	7.1	M/M; T; A/B; A
25	7	25.9	M/M; T; A/B; A
30	1	28.6	T; A/B
35	1	31.2	M/M
40	2	36.6	A; T
45	2	42.0	T; A
50	2	47.3	A; G/D
55	3	55.4	M/M; G/D
60	4	66.1	M/M; QA; G/D; FZ
65	1	68.8	A
70	4	79.5	M/M; QA; G/D; FZ
75	1	82.1	M/M; G/D; FZ
85	2	87.5	QA; M/M; T
90	1	90.2	QA; M/M
100	1	92.9	M/M; FZ
120	1	95.5	M/M; FZ
135	1	98.2	T; A/B; A

Mean 54.3 Std Err 4.6 Std Dev 28.3 Median 53.3
Kurtosis 0.8 Skewness 0.9

*Refer to Table 4-12 for explanation of symbols.

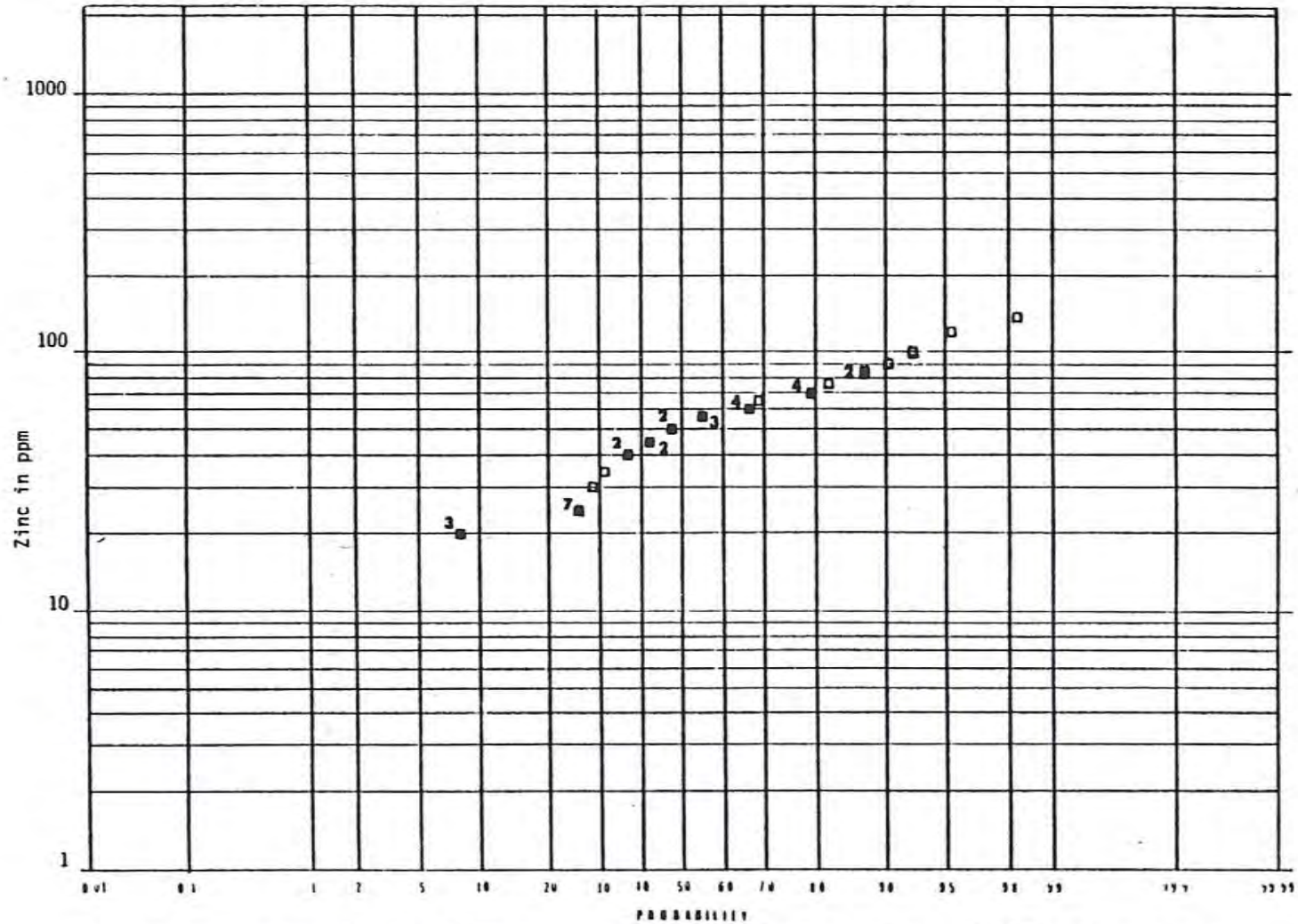


Figure 4-20. Log probability plot of zinc concentrations in whole rock samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-20

Arsenic Distribution in DF 45-14 Whole Rock Samples

Arsenic (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
5	10	25.9	M/M;QA;T;A/B;G/D;FZ
10	7	44.6	M/M; QA; T; G/D; FZ
15	2	50.0	M/M; T; G/D; FZ
20	2	55.4	M/M; A; G/D
25	3	63.4	M/M; T; A/B
30	1	66.1	M/M
40	1	68.8	M/M; G/D
50	1	71.4	A
60	1	74.1	A
65	1	76.8	T; A/B
70	2	82.1	T; G/D; A
80	1	84.8	A
85	1	87.5	A
145	1	90.2	T
230	1	92.9	T; A/B
350	1	95.5	T; A
760	1	98.2	T; A/B; A

Mean 62.2 Std Err 22.5 Std Dev 136.6 Median 16.3
 Kurtosis 19.9 Skewness 4.2

*Refer to Table 4-12 for explanation of symbols.

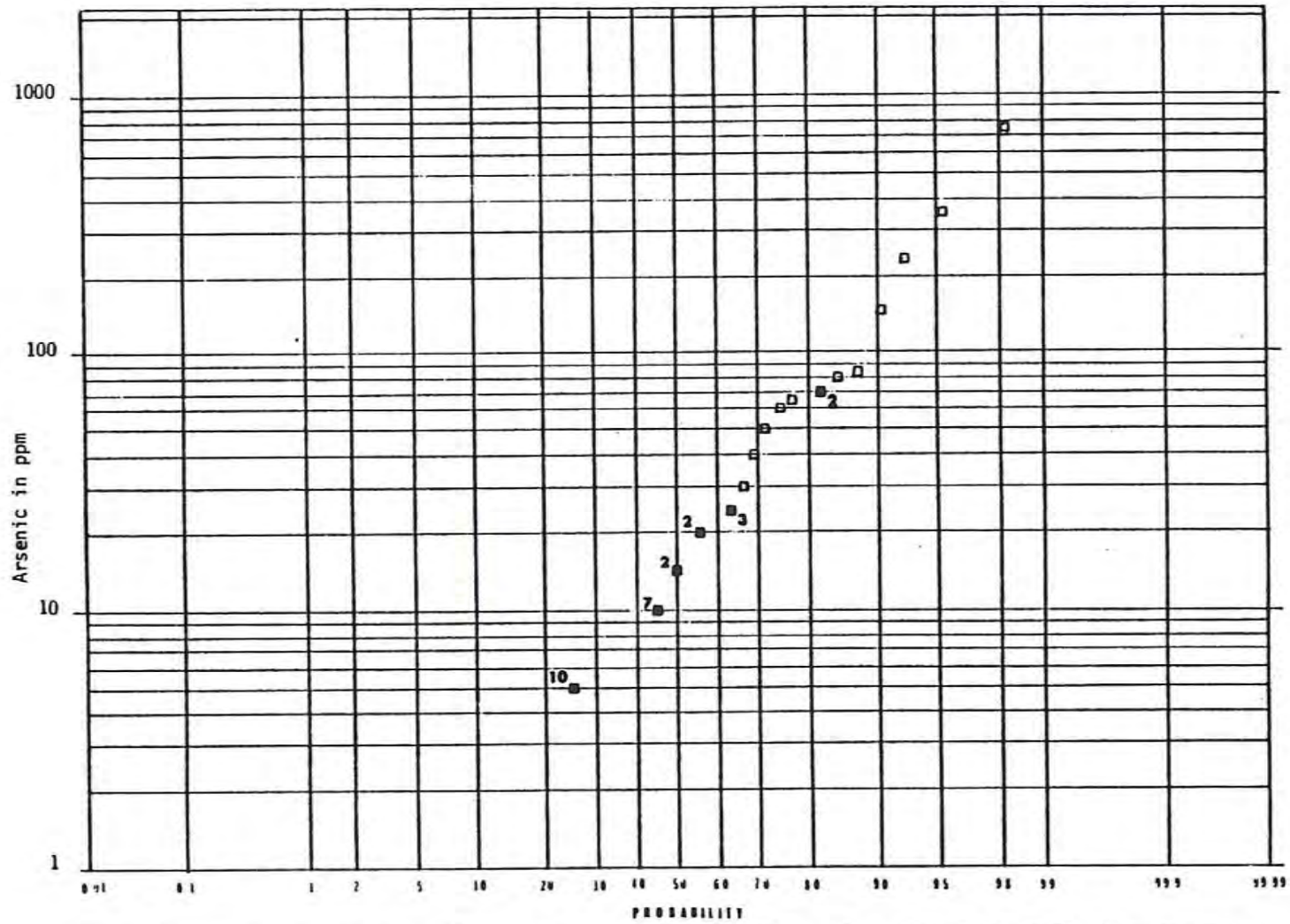


Figure 4-21. Log probability plot of arsenic concentrations in whole rock samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-21

Antimony Distribution in DF 45-14 Whole Rock Samples

Antimony (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
1.0	25	66.1	M/M;Q/A;T;A/B;G/D;FZ
2.0	1	68.8	A
3.0	1	71.4	T; A/B
6.0	1	74.1	T; A/B
10.0	1	76.8	A
12.0	1	79.5	A
16.0	1	82.1	T; A/B
21.0	1	84.8	A; T
22.0	1	87.5	T; A/B; A
24.0	1	90.2	A
29.0	1	92.9	A
36.0	1	95.5	A; T
46.0	1	98.2	T

Mean 6.8 Std Err 1.9 Std Dev 11.3 Median 1.2
Kurtosis 3.8 Skewness 2.1

*Refer to Table 4-12 for explanation of symbols.

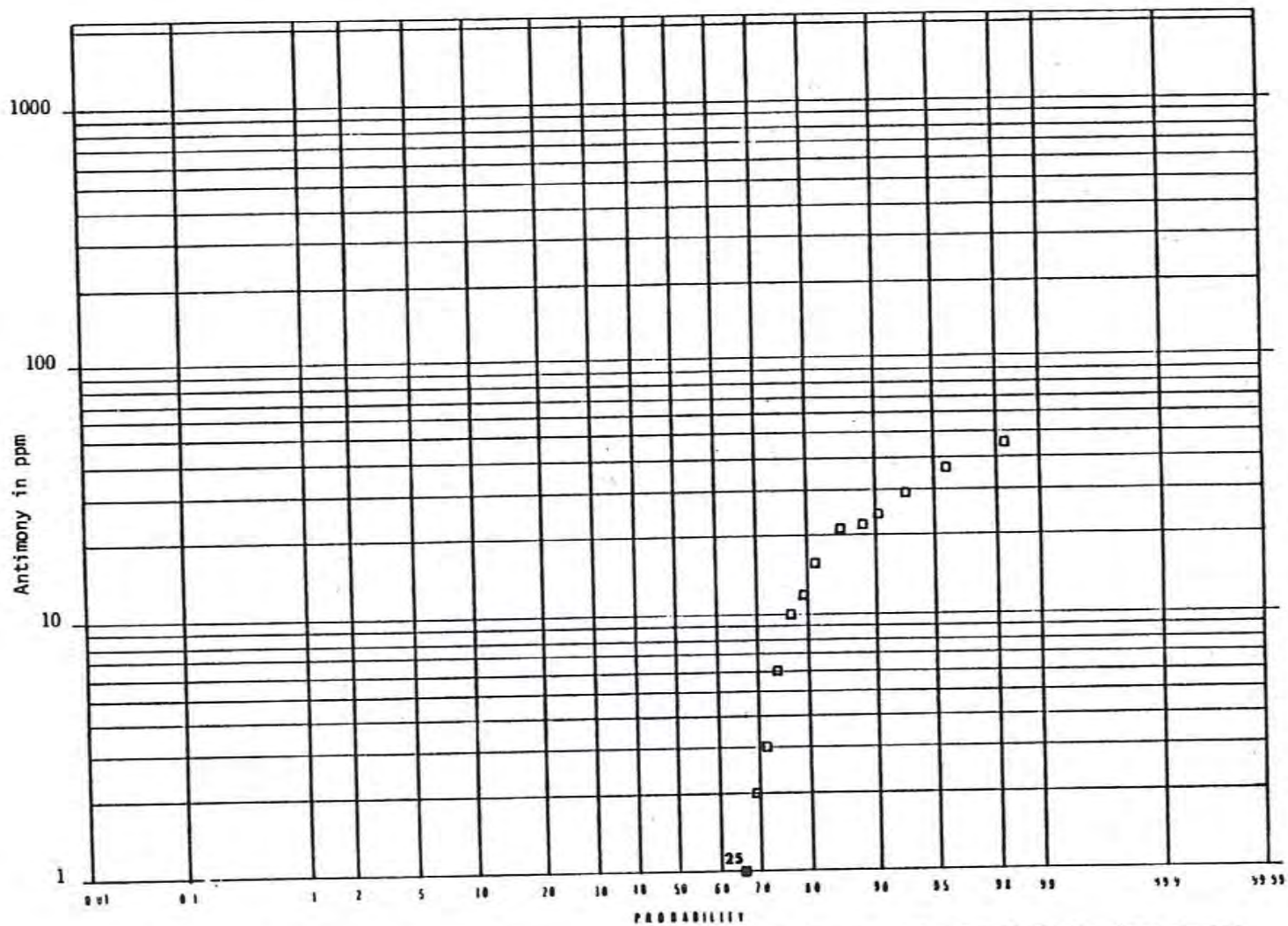


Figure 4-22. Log probability plot of antimony concentrations in whole rock samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-22

Mercury Distribution in DF 45-14 Whole Rock Samples

Mercury (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
0.075	1	1.8	M/M; FZ
0.180	2	7.1	M/M
0.185	1	9.8	QA; M/M; G/D
0.190	2	15.1	M/M; QA
0.195	1	17.9	M/M; G/D; FZ
0.220	3	25.9	T; A/B; M/M
0.225	3	33.9	M/M; T; A/B; G/D
0.240	3	42.0	T; A/B; M/M; G/D
0.245	3	50.0	M/M; G/D
0.250	1	52.7	M/M
0.260	1	55.4	M/M; QA; FZ
0.270	1	58.0	M/M
0.300	1	60.7	A
0.310	1	63.4	M/M; F/D; R; FZ
0.320	1	66.1	M/M; QA; G/D; T
0.345	1	68.8	QA; M/M
0.350	1	71.4	QA; T
0.360	1	74.1	M/M; T; FZ
0.460	1	76.8	A
0.940	1	79.5	T; A/B; A
1.4	1	82.1	A
2.3	1	84.8	T; A
2.9	1	87.5	A; T
3.1	1	90.2	A
3.3	1	92.9	A; T
6.0	1	95.5	A
7.4	1	98.2	T

Mean 0.9 Std Err 0.3 Std Dev 1.6 Median 0.3
Kurtosis 7.9 Skewness 2.8

*Refer to Table 4-12 for explanation of symbols.

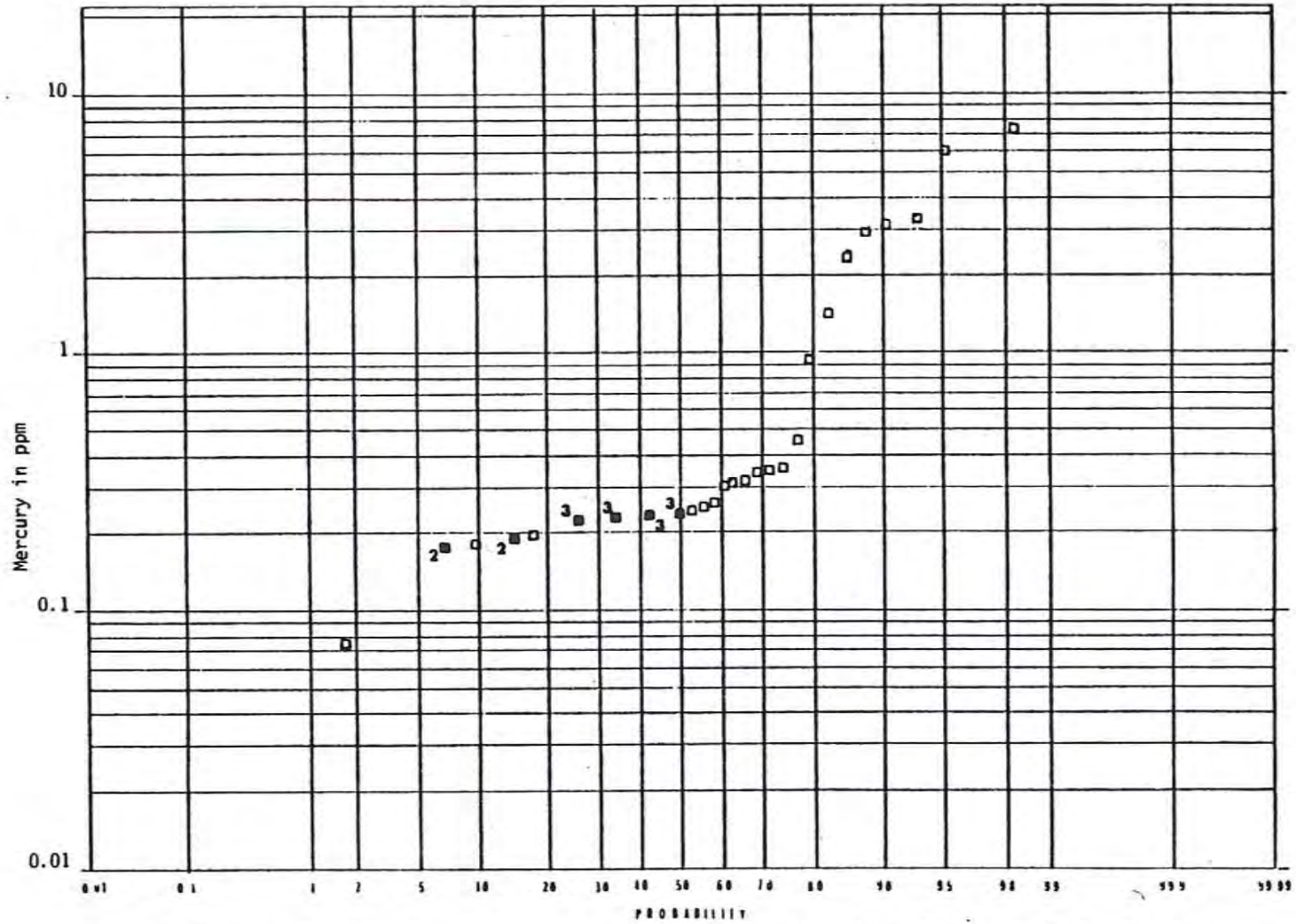


Figure 4-23. Log probability plot of mercury concentrations in whole rock samples for DF 45-14. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-23
Lead Distribution in DF 66-21 Heavy Mineral Fractions

Lead (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
25	22	58.0	A; A/c; T; A/B; G
50	4	68.7	A/c
75	2	74.1	A/c; G; QA; FZ
250	1	76.8	G
650	1	79.5	M/M
1000	1	82.1	M/M; G
1100	1	84.8	M/M; G/D; FZ; MZ
1200	2	90.2	M/M; G
6000	1	92.8	M/M
9500	1	95.5	A; A/c
13500	1	98.3	QA; MZ; FZ

 Mean 954.1 Std Err 458.8 Std Dev 2791.0 Median 33.5
 Kurtosis 13.4 Skewness 3.7

*Refer to Table 4-12 for explanation of symbols.

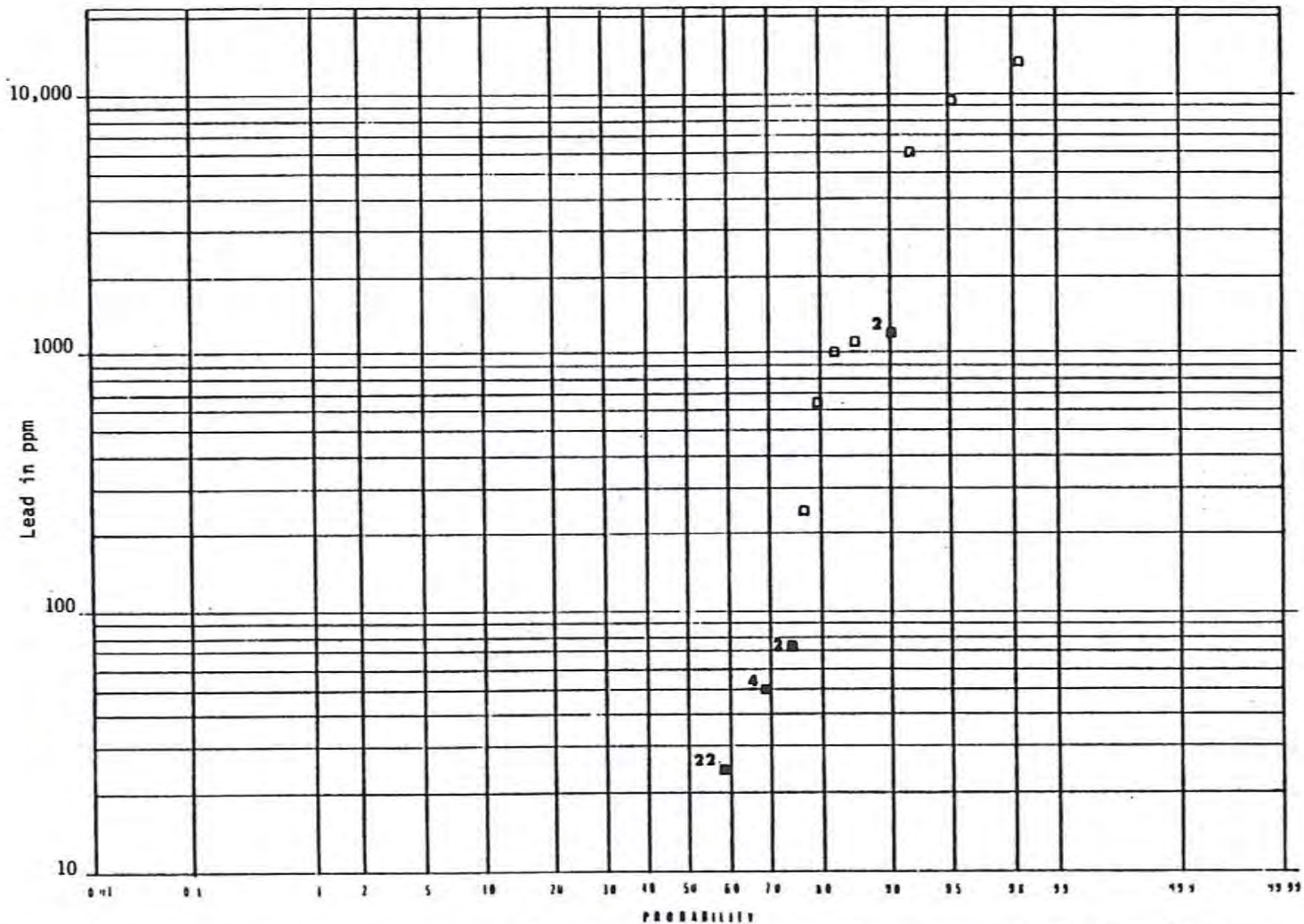


Figure 4-24. Log probability plot of lead concentrations in heavy mineral fraction samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-24
Zinc Distribution in DF 66-21 Heavy Mineral Fractions

Zinc (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
25	7	17.8	G; A; A/B; T; MZ; FZ
50	11	47.3	A/c; A/B; T; FZ; G
75	1	50.0	A/B; T; FZ
100	5	63.4	G; A; QA; A/B
125	4	74.1	A; G; M/M
150	1	76.8	M/M; G; G/D; MZ; FZ
175	2	82.1	M/M; A; G
250	2	87.5	A; A/c
520	1	90.2	G; QA; FZ
950	1	92.8	M/M; G
1300	1	95.5	M/M; G; G/D
11500	1	98.3	A/B; T; FZ

Mean 461.4 Std Err 310.0 Std Dev 1882.6 Median 75.0
Kurtosis 35.5 Skewness 5.9

*Refer to Table 4-12 for explanation of symbols.

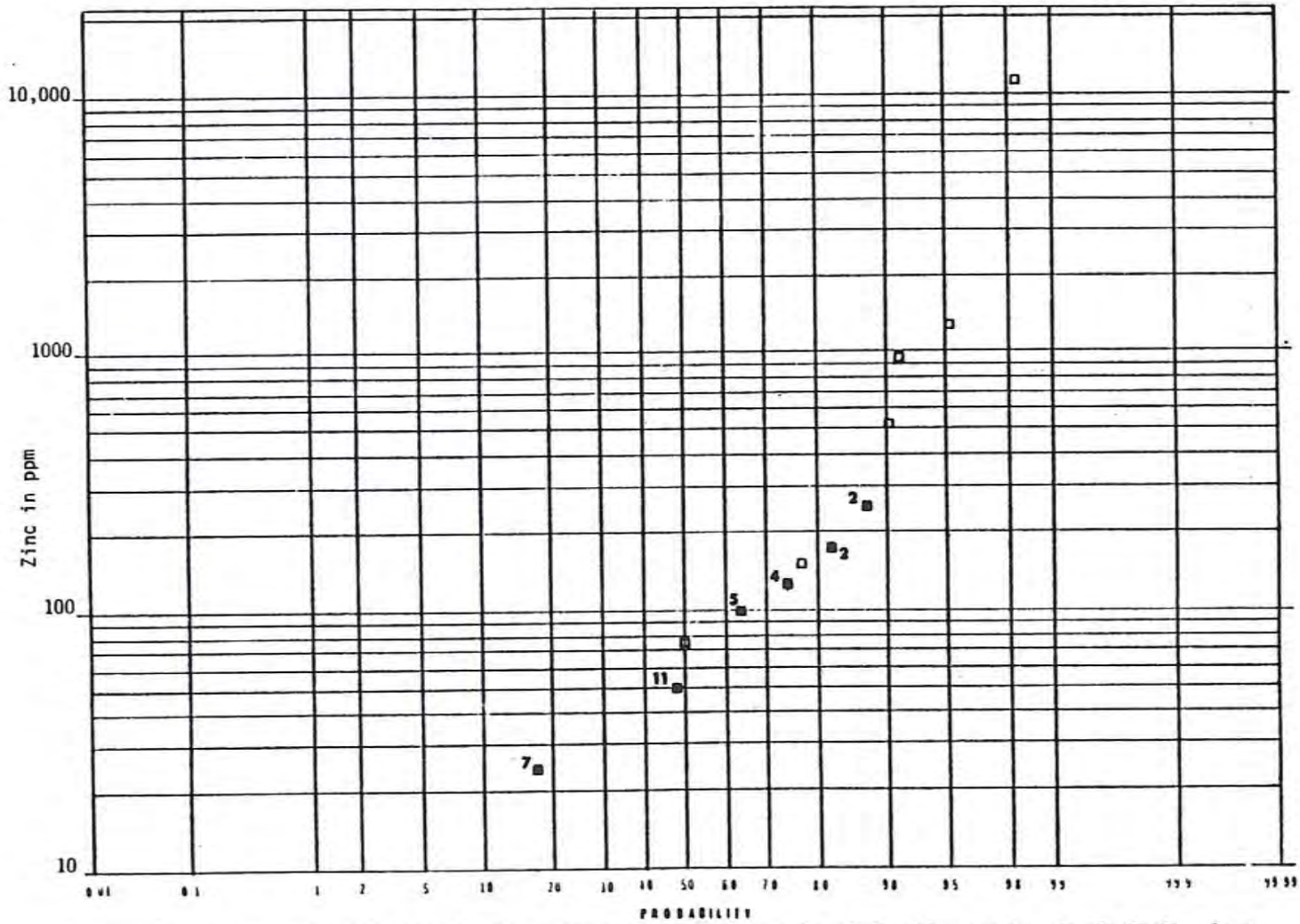


Figure 4-25. Log probability plot of zinc concentrations in heavy mineral fraction samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-25

Arsenic Distribution in DF 66-21 Heavy Mineral Fractions

Arsenic (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
5	1	1.7	G
25	5	15.2	A; A/B; T; G; FZ
50	9	39.3	A; A/c; A/B; T; FZ; G
75	4	50.0	A; A/c; G
100	3	58.0	A/c; M/M; G
125	2	63.4	A/c; QA; MZ; FZ
175	2	68.8	G; QA; A; A/c
200	2	74.1	G; QA; A/B; T; FZ
225	1	76.8	A/B; T; FZ
250	2	82.1	A/c; A/B; T
320	1	84.8	M/M; G
420	1	87.5	M/M; G; G/D; MZ; FZ
480	1	90.2	M/M
520	1	92.9	M/M; G
780	1	95.5	G
1200	1	98.3	QA; G; MZ; FZ

Mean 179.1 Std Err 39.4 Std Dev 239.7 Median 82.5
 Kurtosis 9.2 Skewness 2.8

*Refer to Table 4-12 for explanation of symbols.

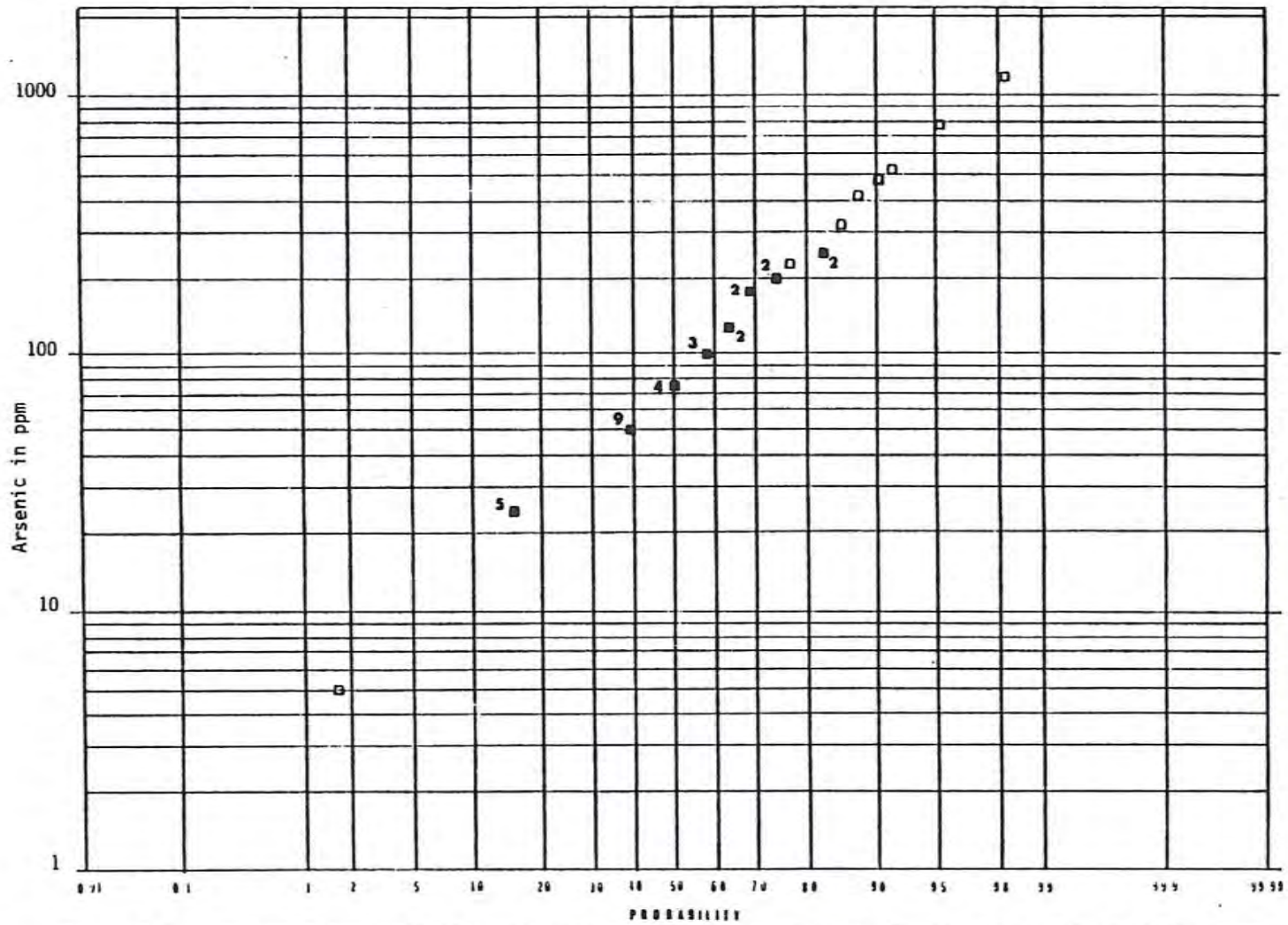


Figure 4-26. Log probability plot of arsenic concentrations in heavy mineral fraction samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-26

Antimony Distribution in DF 66-21 Heavy Mineral Fraction

Antimony (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
2	10	26.6	G;A;A/c;A/B;T;M/M;MZ;FZ
4	2	32.1	A; A/c; A/B; T
6	4	43.1	A; A/c; QA; MZ; FZ
8	4	54.1	A; A/c; A/B; T; M/M
10	1	56.9	G
14	3	65.1	A; A/c
16	2	70.6	A/c
18	1	73.4	M/M; G
20	1	76.1	G; QA
24	2	81.7	A/c
48	2	87.2	M/M; A/B; T; FZ
68	1	89.9	M/M; G
88	1	92.7	A/B; T; FZ
156	1	95.4	A/B; T; FZ
1600	1	98.3	G; QA; FZ

Mean 62.8 Std Err 44.2 Std Dev 265.3 Median 8.0
 Kurtosis 35.0 Skewness 5.9

*Refer to Table 4-12 for explanation of symbols.

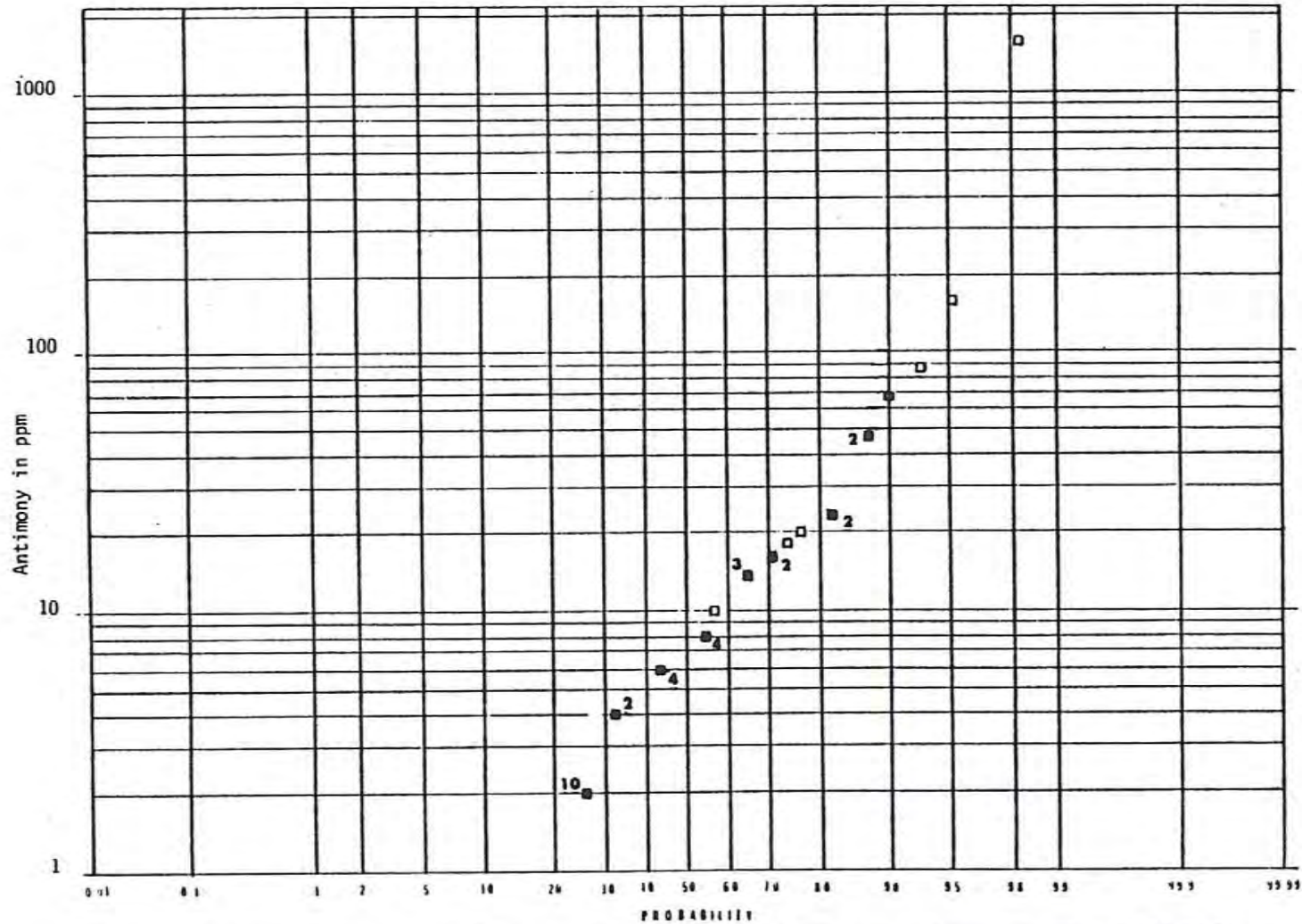


Figure 4-27. Log probability plot of antimony concentrations in heavy mineral fraction samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-27

Mercury Distribution in DF 66-21 Heavy Mineral Fractions

Mercury (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
0.105	1	1.8	A/B; T; FZ
0.225	1	4.5	A/c
0.260	1	7.1	A/B; T; FZ
0.285	1	9.8	A
0.305	1	12.5	M/M; G
0.315	1	15.1	A/c
0.375	1	17.9	A
0.400	1	20.5	A
0.435	2	25.9	A; A/c; G
0.660	1	28.6	G; QA
0.680	1	31.3	G
0.695	1	33.9	G
0.730	1	36.6	M/M
0.750	1	39.3	G; FZ
0.995	1	42.0	M/M; G; G/D; MZ; FZ
1.3	1	44.6	A
1.4	1	47.3	A
1.6	1	50.0	QA; FZ; MZ
1.8	2	55.4	A/c; G; QA
2.3	2	60.7	A; G; QA; MZ; FZ
2.5	1	63.4	A/B; T; A/c
3.5	1	66.1	G
5.5	1	68.8	A; A/c
7.2	1	71.4	M/M; G
8.6	1	74.1	M/M; G
10.3	1	76.8	A/B; T
13.0	1	79.5	A; A/c
13.4	1	82.1	QA; MZ; FZ
16.0	1	84.8	A/B; T; A/c
19.0	1	87.5	A/B; T; A/c; FZ
21.0	1	90.2	A/c; A/B; T
37.0	1	92.9	A/B; T; A/c
42.0	1	95.5	A/c
43.0	1	98.3	A/c

Mean 7.1 Std Err 1.9 Std Dev 11.6 Median 1.6
Kurtosis 4.1 Skewness 2.2

*Refer to Table 4-12 for explanation of symbols.

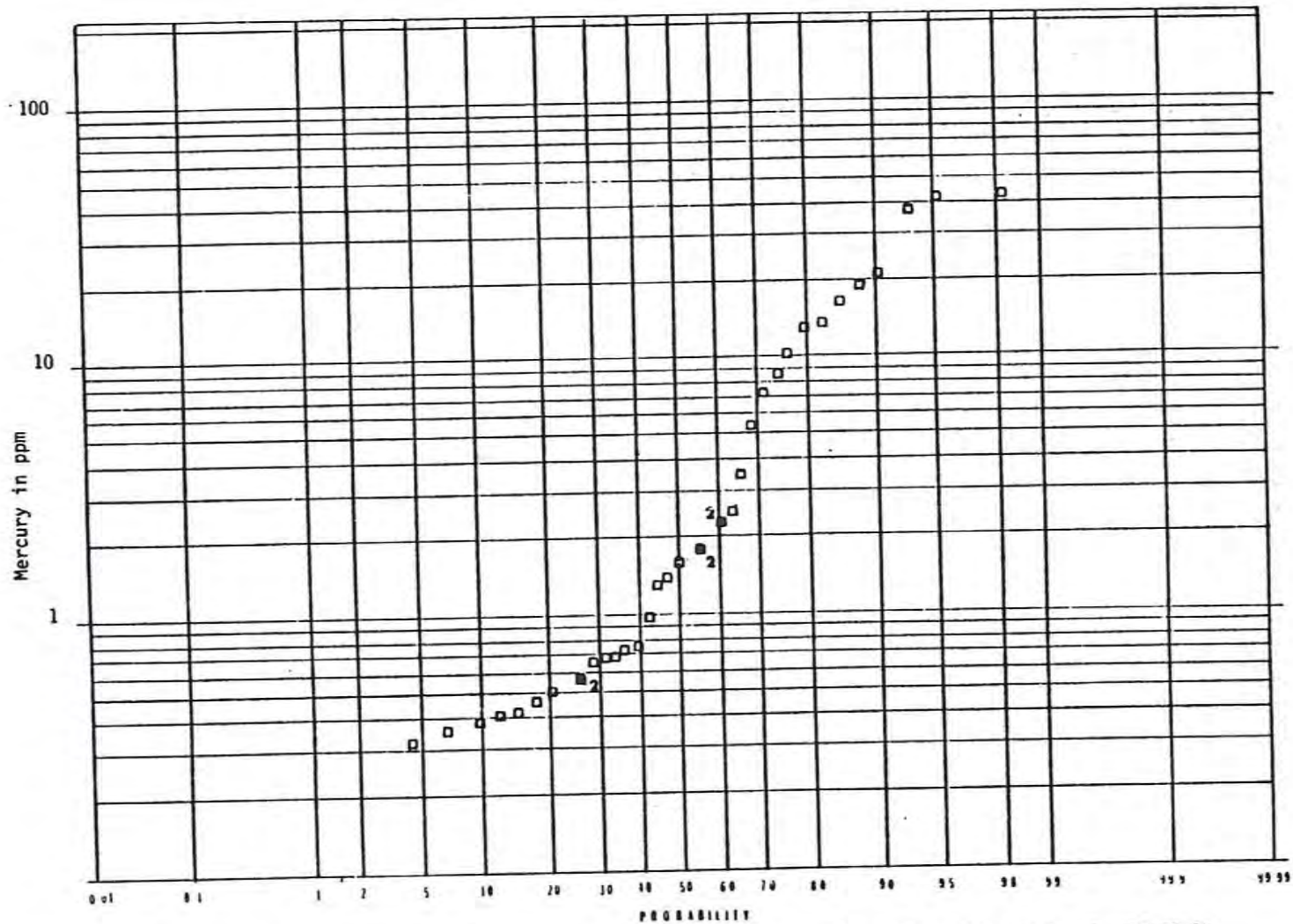


Figure 4-28. Log probability plot of mercury concentrations in heavy mineral fraction samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-28

Lead Distribution in DF 66-21 Whole Rock Samples

Lead (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
5	6	14.8	A; QA; G; MZ; FZ
10	11	43.5	A/c; A; A/B; T; G; QA
15	5	56.5	A/c;A;A/B;T;G;QA;FZ;MZ
20	10	82.6	G;A;QA;A/B;T;A/c;FZ;MZ
25	1	85.2	M/M; G
30	2	90.4	M/M; G
40	1	93.0	M/M
45	1	95.7	A/B; T; FZ
65	1	98.3	A; A/c

Mean 17.1 Std Err 2.0 Std Dev 12.2 Median 14.5
 Kurtosis 5.8 Skewness 2.1

*Refer to Table 4-12 for explanation of symbols.

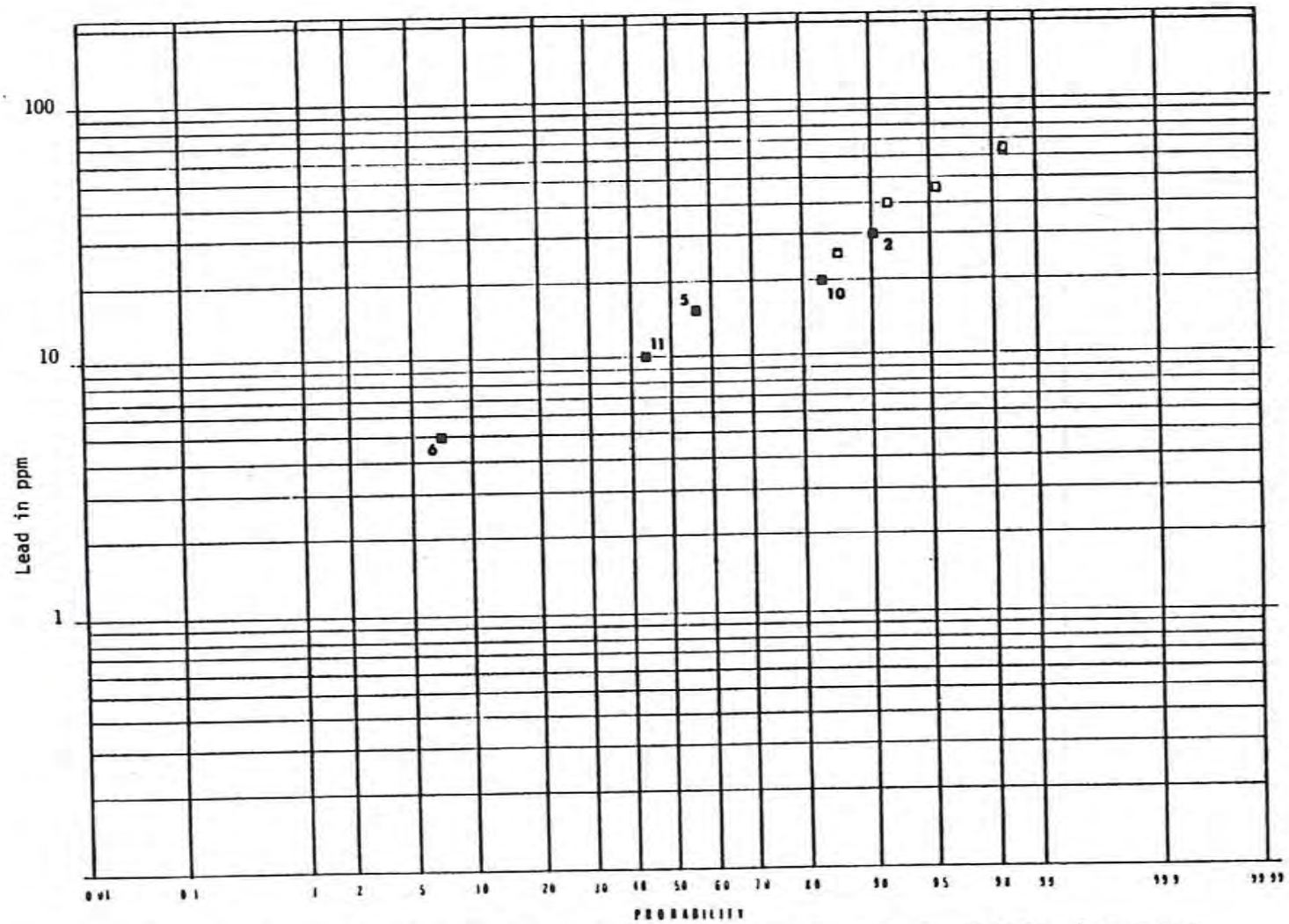


Figure 4-29. Log probability plot of lead concentrations in whole rock samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-29

Zinc Distribution in DF 66-21 Whole Rock Samples

Zinc (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
5	1	1.7	QA; MZ; FZ
10	5	14.8	G; QA; MZ; FZ
15	2	20.0	G; A/B; T; FZ
20	3	27.8	G; QA; A/B; T; FZ
25	2	33.0	G; QA; A/B; T
35	4	43.5	A; G; QA; A/B; T; MZ; FZ
40	2	48.7	A; A/c
45	2	53.9	A; A/c; A/B; T
50	4	64.3	A; A/c
55	3	72.2	A/c; A/B; T; FZ
60	3	80.0	A/c; M/M; G
65	3	87.8	M/M; G; A/c; A/B
70	1	90.4	M/M; G
75	2	95.6	M/M; A; A/c
80	1	98.3	M/M; G/D; G; MZ; FZ

Mean 40.7 Std Err 3.6 Std Dev 22.0 Median 42.5
Kurtosis -1.2 Skewness -0.04

*Refer to Table 4-12 for explanation of symbols.

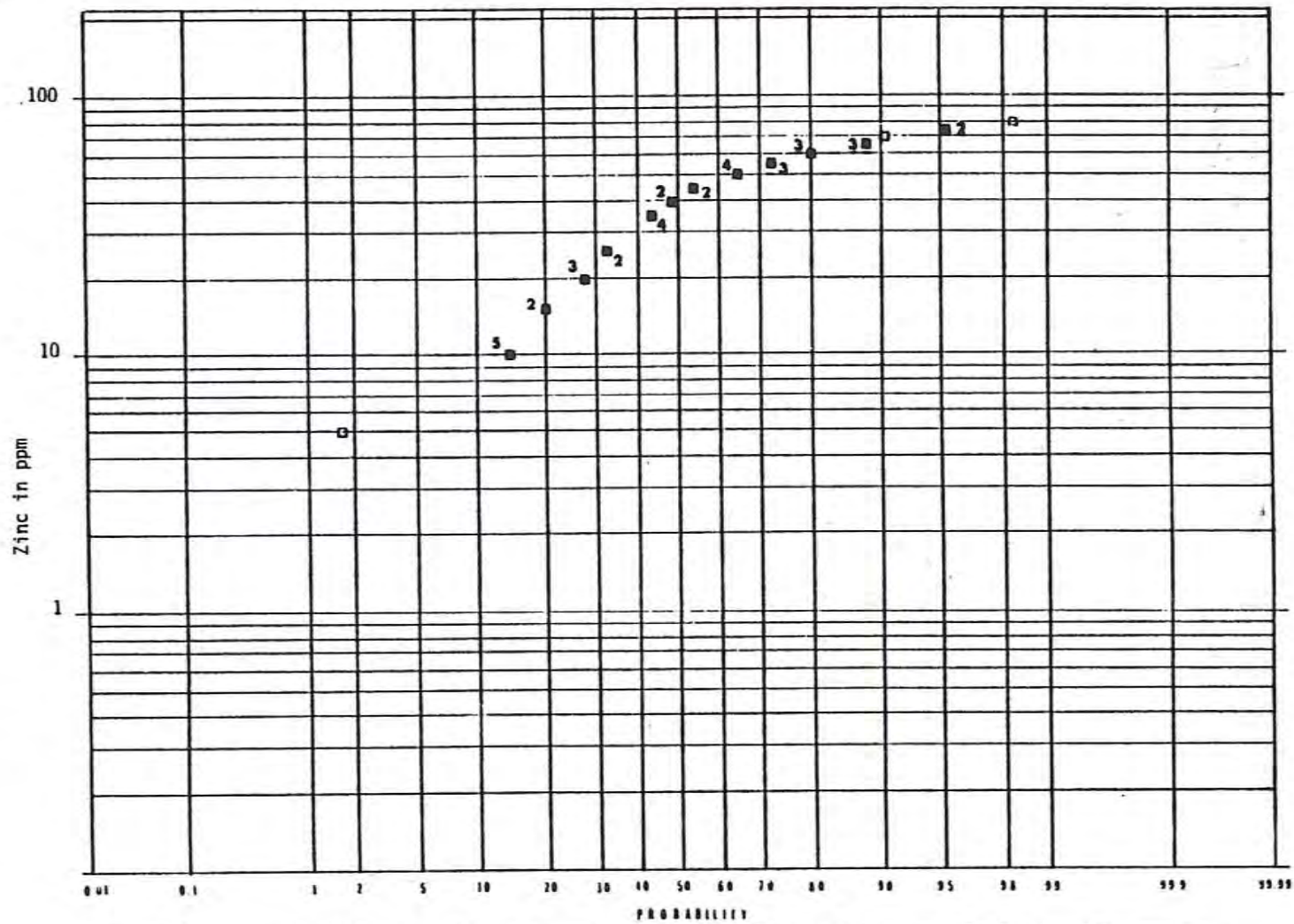


Figure 4-30. Log probability plot of zinc concentrations in whole rock samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-30

Arsenic Distribution in DF 66-21 Whole Rock Samples

Arsenic (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
5	24	61.7	G;M/M;A;A/c;A/B;T;QA;FZ;MZ
10	10	87.8	A;A/c;A/B;T;G;QA;FZ
15	1	90.4	A/c; A/B; T
20	1	93.0	A/B; A/c; T
30	1	95.6	M/M; G
50	1	98.3	G

Mean 8.8 Std Err 1.4 Std Dev 8.5 Median 6.5
 Kurtosis 15.6 Skewness 3.7

*Refer to Table 4-12 for explanation of symbols.

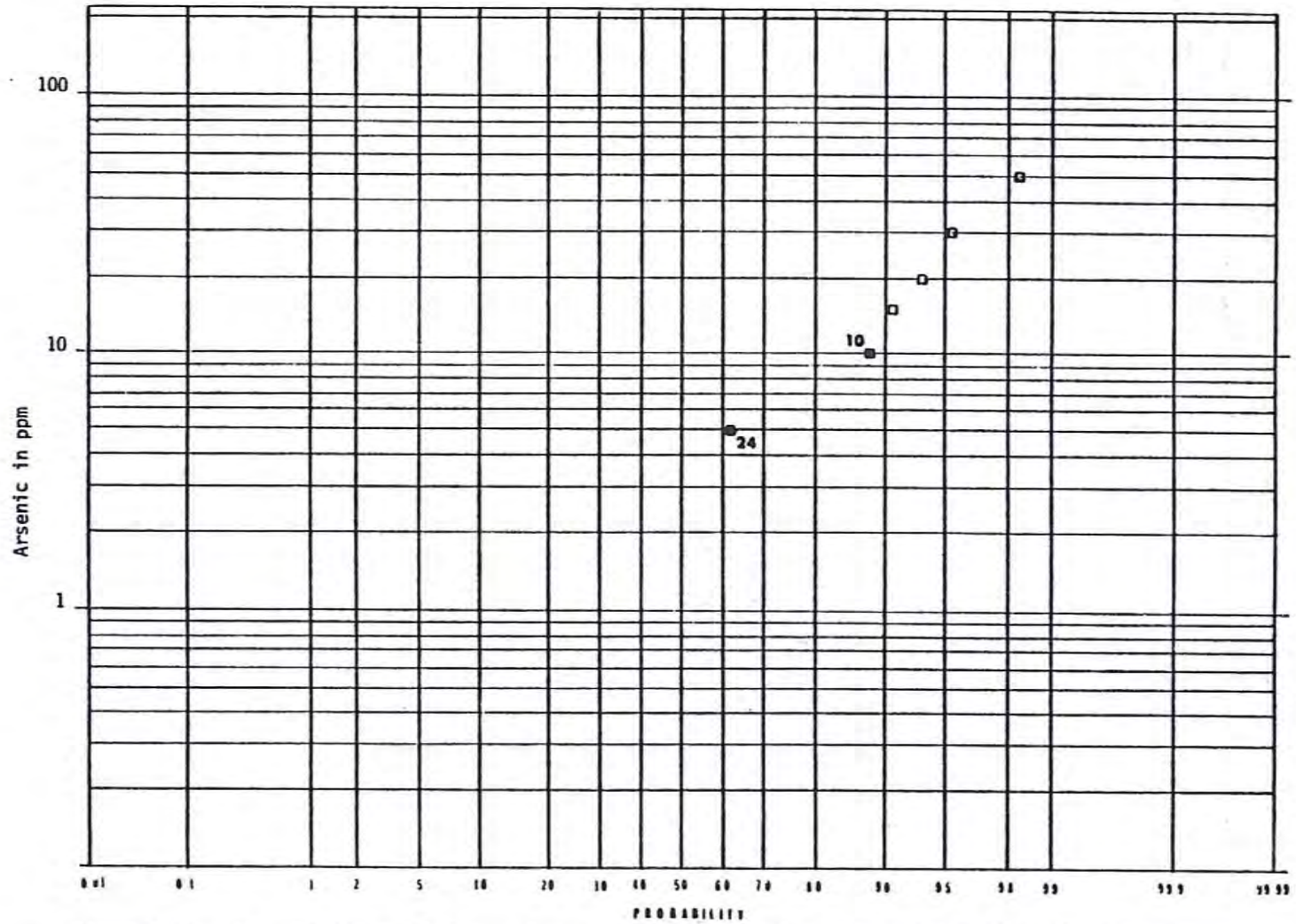


Figure 4-31. Log probability plot of arsenic concentrations in whole rock samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-31
Antimony Distribution in DF 66-21 Whole Rock Samples

Antimony (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
1	25	64.3	A;A/c;G;QA;M/M;G/D;MZ;FZ
2	7	82.6	A/B; T; M/M; G; FZ
3	2	87.8	A/c; A/B; T; G; QA
4	1	90.4	A/B; T; FZ
5	1	93.0	A/B; T; FZ
6	1	95.6	A/B; T
7	1	98.3	A/B; T

Mean 1.8 Std Err 0.2 Std Dev 1.5 Median 1.3
Kurtosis 5.3 Skewness 2.4

*Refer to Table 4-12 for explanation of symbols.

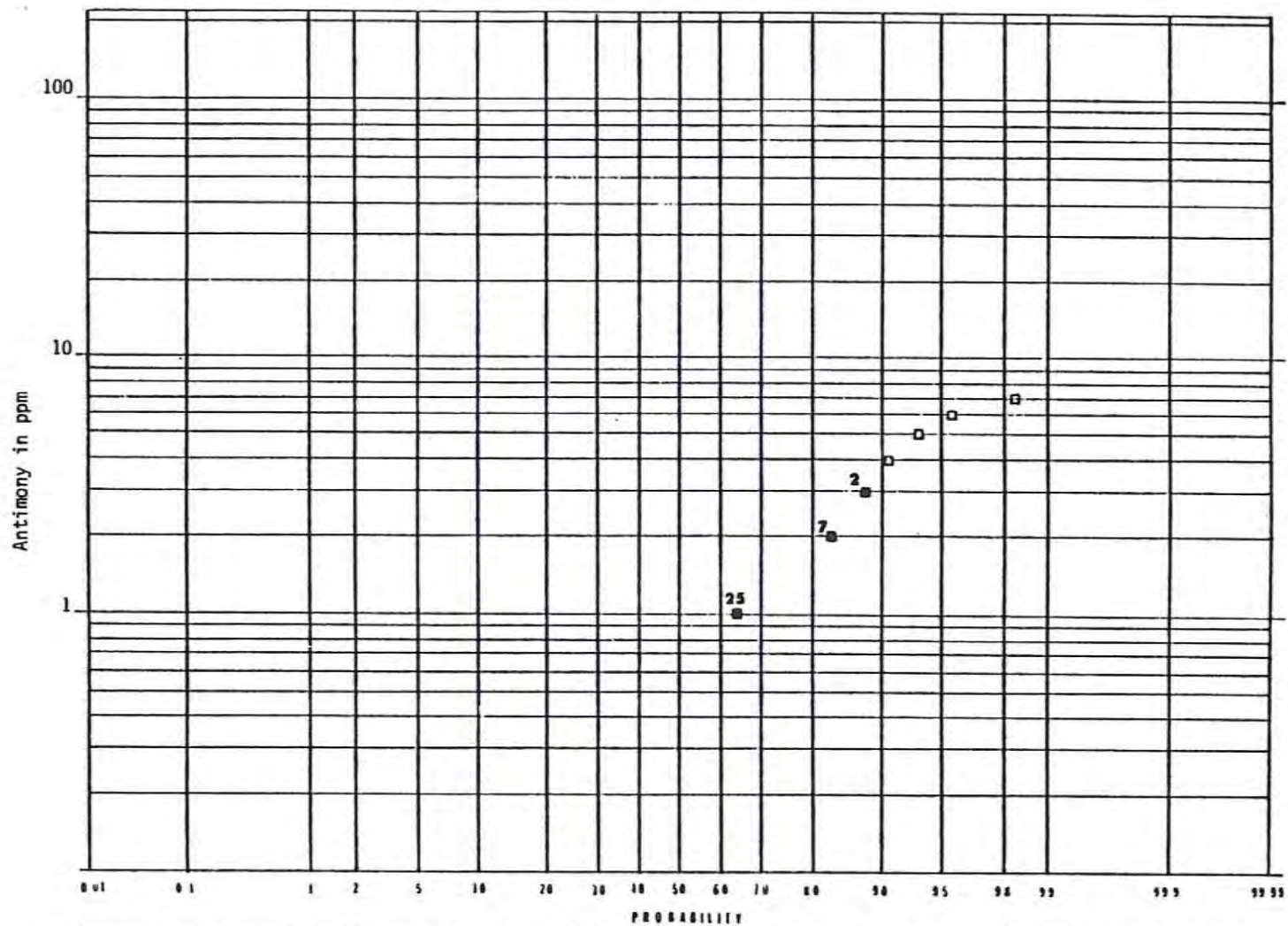


Figure 4-32. Log probability plot of antimony concentrations in whole rock samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

Table 4-32

Mercury Distribution in DF 66-21 Whole Rock Samples

Mercury (ppm)	Absolute Frequency	Plotting Percentage	Lithology*
0.085	1	1.7	G
0.095	1	4.3	A
0.105	2	9.5	M/M; G
0.115	1	12.2	M/M; G
0.120	2	17.4	A; A/c
0.125	2	22.6	G
0.130	4	33.0	A; A/c; G; QA; FZ
0.140	2	38.3	A; A/c
0.150	2	43.5	A/c; A
0.155	1	46.1	M/M; G
0.160	1	48.7	A; A/c
0.165	1	51.3	A
0.175	2	56.5	M/M; G; G/D; MZ; FZ
0.180	2	61.7	A; A/B; T; FZ
0.190	1	64.3	A/c; A/B; T
0.195	1	66.9	A/c; A
0.200	2	72.2	A; G; QA; FZ
0.235	1	74.8	A/B; T; FZ
0.255	1	77.4	G
0.260	1	80.0	G; QA
0.265	1	82.6	QA; MZ; FZ
0.280	1	85.2	A/B; T; A/c
0.300	1	87.8	A/B; T; A/c
0.330	1	90.4	G; QA
0.460	1	93.0	A/B; T
0.685	1	95.7	G; QA; MZ; FZ
0.720	1	98.3	A/B; T

Mean 0.2 Std Err 0.02 Std Dev 0.1 Median 0.2

Kurtosis 7.1 Skewness 2.6

*Refer to Table 4-12 for explanation of symbols.

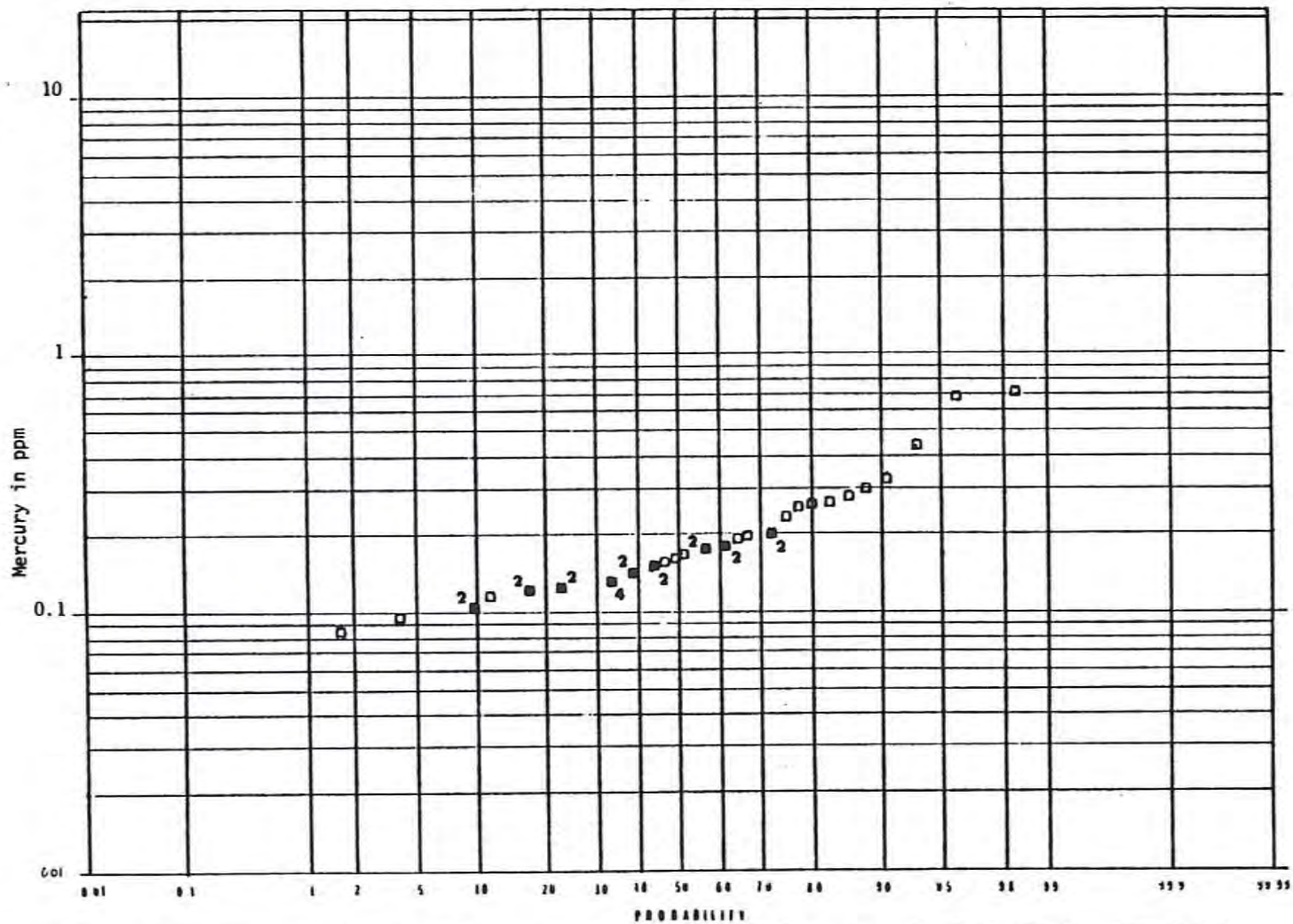


Figure 4-33. Log probability plot of mercury concentrations in whole rock samples for DF 66-21. Open squares indicate single sample value; solid squares indicate multiple sample values.

APPENDIX F-3

Haskins • Pfeiffer Inc.

International Geophysical Consultants

1449 Denver Club Building
Denver, Colorado 80202
(303) 573-8958



October 7, 1980

Southland Royalty Company
1000 Fort Worth Club Tower
Fort Worth, Texas 76102
Attn: Mr. Jere Denton

RE: Dixie Valley Area
Churchhill & Pershing
Counties, Nevada

Dear Mr. Denton:

Seismic data for the Dixie Valley Prospect Area were acquired by Petroleum Geophysical Co. party No. 302. The data were recorded in SEG-B format by Texas Instrument's DFS-5 amplifiers. An 8-128Hz filter was used on recording at a sample rate of 2 ms.

Energy source consisted of four Mertz-11 vibrators equipped with Peltan Electronics Advance 1 Model 4 instruments. Sixteen sweeps were used per V.P. with sweep pilot of 12-60Hz. V.P. interval was 220 feet with a group interval of 110 feet. Stack is 2400% from 96 trace instruments with a spread geometry 5280'-440'-0'-440'-5280'.

Data processing was performed by Western Geophysical Co. Complex and rapidly changing geologic environment introduced difficult velocity and stacking problems. These problems were resolved by Western's diligence in the analysis of the stacking velocities and design of deconvolution parameters to equalize the frequency spectrum to attenuate short period multiples. Since the dips in the area were generally less than 20 degrees, "Finite Difference" digital migration was used to collapse diffractions generated by faulting and to place reflected energy in the true time domain position.

It should be noted that any additional shooting in the area will be more effective from the standpoint of migration techniques if line orientation is designed normal to the strike shown by existing control.

From available control, three maps have been constructed. These are :

- 1) Migrated Producing Aquifer
- 2) Migrated Pluton
- 3) Non-Migrated Shallowest Flow

Southland Royalty Company
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October 7, 1980

Identification is from a synthetic seismogram constructed using the sonic log of the Thermal Power; Dixie Fed. 66-21 in Twp. 24N, Rge. 36E, Sec. 21, or by geologic definition of the mapped event. A discussion of the maps submitted follows:

MIGRATED PRODUCING AQUIFER

The event identified as the producing aquifer in the Thermal Power Well Dixie Fed. 66-21 is coincident with a high amplitude event generated on the synthetic seismogram from that well. A reasonable correlation exists between the synthetic and the recorded seismic data at the well. This event, in its migrated position is the basis for the map under discussion.

Because the seismic lines cross from the sedimentary basin to positions overlying the pluton, interruptions occur in the continuity of data from the aquifer. In such cases it becomes necessary to re-establish continuity through correlation across the data gaps.

The overall configuration of the event mapped indicates the Thermal Power well encountered the aquifer in an overthrust block off the forefront of the pluton. The aquifer terminates both against the pluton and basinward. A series of events with character similar to the aquifer can be noted on the sections. Because of the limited areal extent of the individual events, the map has been constructed using the event with characteristic response in the nearest equivalent stratigraphic position. Termination of events is marked by hatchers.

MIGRATED PLUTON

The purpose of this map is to show the configuration of the forefront of the pluton so that the contact between the aquifer and the pluton can be viewed in perspective. The criteria used for defining the forefront is the zone of termination of coherent energy.

NON-MIGRATED SHALLOWEST FLOW

Since no appreciable geothermal gradient is indicated in the Thermal Power well, it is possible that any potential aquifer would be suitable for production without regard to depth of burial. This map is constructed to show the shallowest position at which an event with character similar to the identified aquifer, could be expected.

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RECOMMENDATIONS

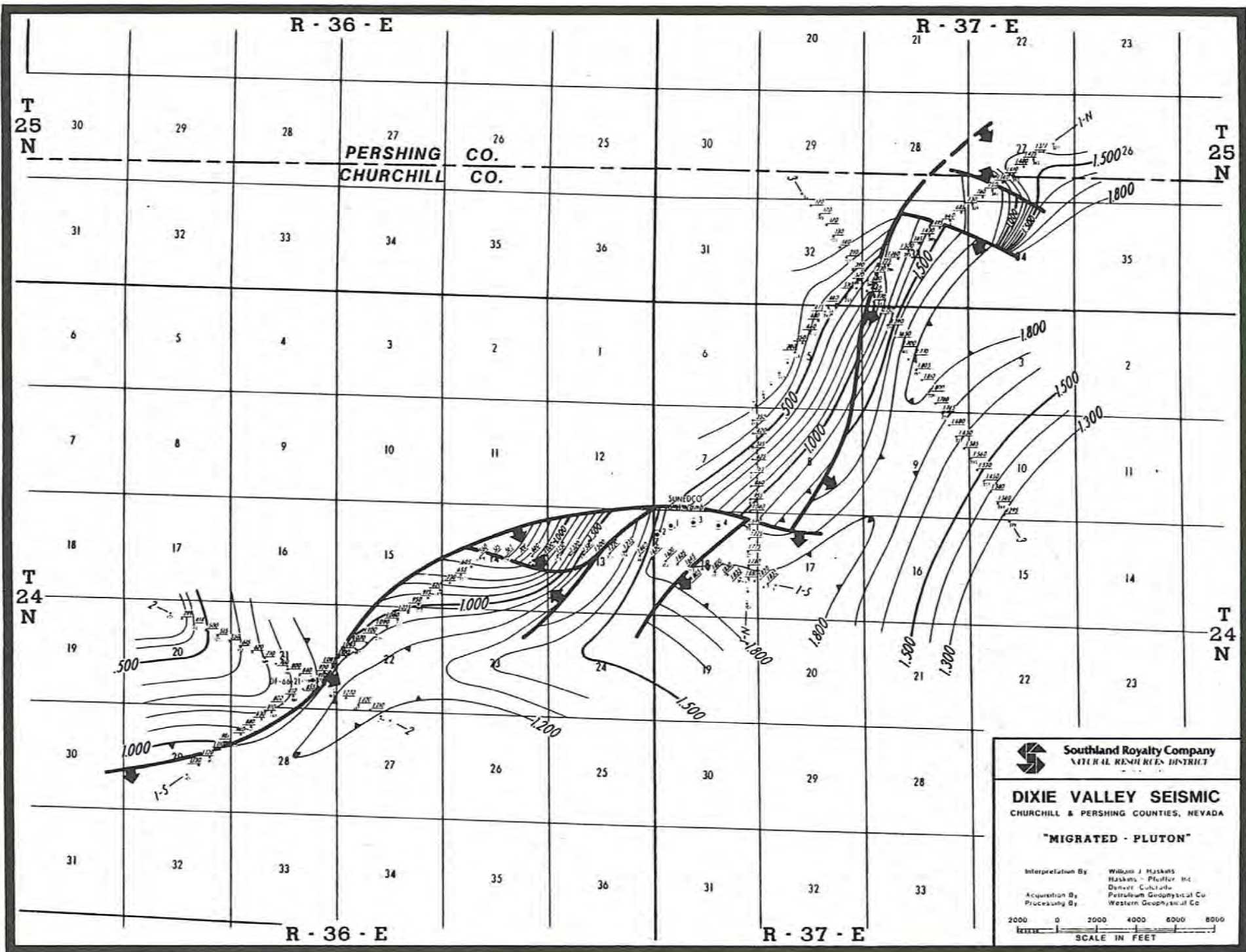
At this stage of exploration, plans should be made to increase the density of seismic control in the areas available for drilling. This additional control should use parameters compatible with the existing data. Care should be taken to keep all new lines normal to strike to aid migration. Processing should be carefully supervised so that data quality will be compatible with that now at hand.

Yours truly,

Haskins-Pfeiffer, Inc.


William J. Haskins

WJH/1f

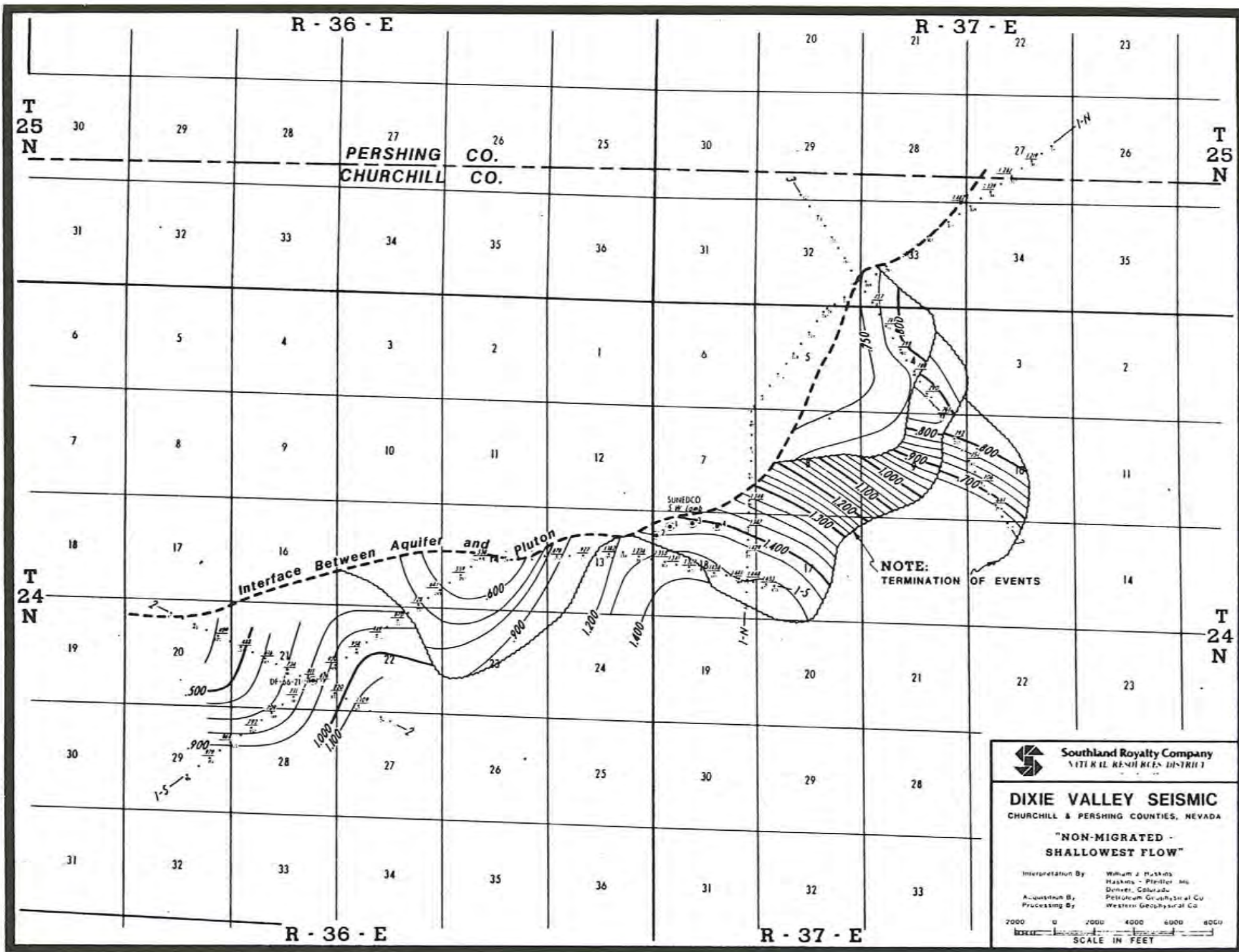


 **Southland Royalty Company**
NATURAL RESOURCES DISTRICT

DIXIE VALLEY SEISMIC
CHURCHILL & PERSHING COUNTIES, NEVADA

"MIGRATED - PLUTON"

Interpretation By: William J. Huskins
Huskins - Pfeiffer Inc.
Denver, Colorado
Acquisition By: Petroleum Geophysical Co.
Processing By: Western Geophysical Co.




Southland Royalty Company
 NATURAL RESOURCES DISTRICT

DIXIE VALLEY SEISMIC
 CHURCHILL & PERSHING COUNTIES, NEVADA

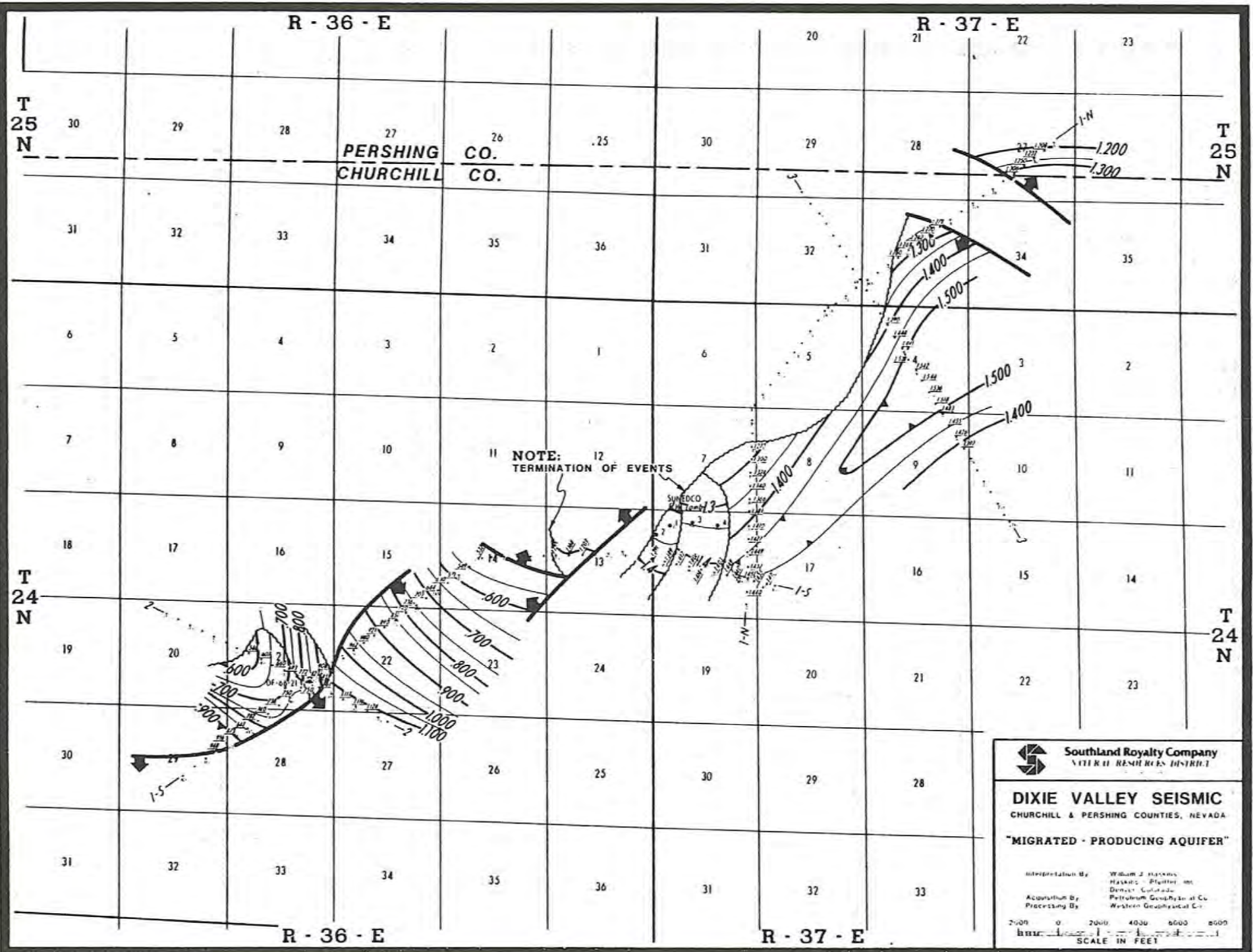
"NON-MIGRATED - SHALLOWEST FLOW"

Interpretation By: William J. Huxford
 Huxford - Pfeiffer & Co., Denver, Colorado

Acquisition & Processing By: Petroleum Geophysical Co. Western Geophysical Co.

2000 0 2000 4000 6000 8000
 SCALE IN FEET

U



PERSHING CO.
CHURCHILL CO.

NOTE:
TERMINATION OF EVENTS

SUNEDCO
13,000

 Southland Royalty Company
NATURAL RESOURCES DISTRICT

DIXIE VALLEY SEISMIC
CHURCHILL & PERSHING COUNTIES, NEVADA

"MIGRATED - PRODUCING AQUIFER"

Interpretation By: William J. Hawkins
Haskins - Platteau, Inc.
Denver, Colorado
Acquisition By: Petroleum Geophysical Co.
Processing By: Western Geophysical Co.

2000 0 2000 4000 6000 8000
SCALE IN FEET

APPENDIX F-4

ENVIRONMENTAL ISOTOPE HYDROLOGY OF THE DIXIE VALLEY
GEOHERMAL SYSTEM

Conducted for:

U.S. Department of Energy
Contract Number DE-AC08-79ET27006

Work Performed under Subcontract to:

Southland Royalty Company
Fort Worth, Texas 76102

Conducted by:

Burkhard Bohn
Neil Ingraham
Michael Campana
Roger Jacobson

Desert Research Institute
University of Nevada, Reno, Nevada

H.0 ENVIRONMENTAL ISOTOPE HYDROLOGY OF THE DIXIE VALLEY GEOTHERMAL SYSTEM

H.1 Introduction

H.1.1 Purpose and Scope

The purpose of this portion of the study is to use inferences from the distribution of naturally-occurring isotopes in the waters of the study area to provide insight into the Dixie Valley geothermal system and input to the formulation of a conceptual model of the Dixie Valley hydrothermal system. The isotope distributions and their interpretations can provide information on recharge to the system and can supplement the information derived from the hydrogeochemistry, which was discussed in an earlier report. Most of the emphasis was placed upon the collection and interpretation of the stable isotopes ^2H (deuterium or D) and ^{18}O . These isotopes are useful in delineating recharge areas and functioning as fluid flow tracers. The radioactive isotope ^3H (tritium) was also utilized to a minor extent. At the inception of the study, it was anticipated that the radioactive isotope ^{14}C would be utilized as well; this environmental isotope would have been used to date the waters from the various parts of the study area. The spatial distribution of fluid ^{14}C dates could then be used to estimate fluid residence times and reservoir volume. However, delays in receiving stable isotope results precluded the collection of ^{14}C data, since the stable isotope information is essential to the proper application of ^{14}C techniques and interpretation of the ^{14}C data.

The study area encompassed the same region as the hydrogeochemical investigations. Most of the work focused on the area of greatest interest to the Southland Royalty Company: the northern half of Dixie Valley (between Dixie Meadows and Sou Hot Springs). Emphasis was placed on the western half of this particular area, i.e., in the vicinity of the Stillwater Range. However, sampling outside this specific area was performed.

H.1.2. Methods and Analytical Techniques

Approximately 100 samples of thermal and non-thermal waters were collected for stable isotope analyses from the study area. In addition, 27 samples were collected for ^3H analyses. All isotope samples

were collected in 125 ml glass bottles with Teflon-lined plastic caps. The bottles were completely filled and capped underwater (where possible) to insure that no air bubbles would be entrapped. The bottles were then sealed with Parafilm and plastic tape.

Both enriched and unenriched ^3H analyses were performed at the Water Resources Center of the Desert Research Institute. Liquid scintillation was used to measure ^3H activity; selected samples were electrolytically enriched before analysis. The stable isotope (D and ^{18}O) samples were analyzed by the Laboratory of Isotope Geochemistry at the University of Arizona. Mass spectroscopy was used for the analyses.

It should be noted that isotope samples were collected from the same locations (with a few exceptions) as the chemical samples and are labeled accordingly. However, because of cost, not all the collected isotope samples were analyzed. Plate H-1 shows the locations of all chemical and isotope samples collected from the study area.

H.1.3. Previous Work

Very little previous work on the hydrology of Dixie Valley exists. A paper by Zones (1957) describes some of the hydrologic effects of the 1954 Dixie Valley earthquake. He noted that wells in the valley showed temporary increases in artesian heads and flow rates which he attributed to a westward tilting of the confined and semi-confined aquifers penetrated by the wells, a reduction in the porosity of the unconsolidated sediments caused by rearrangement of the grains during the earthquake, and local fracturing of the alluvium which may have increased the upward flow of groundwater from depth. Zones did not observe a rise in water temperatures, so he concluded that if water did move from depth along new fracture zones, it did not come from a great depth. He also noted a temporary increase in the flow rate of Mud Springs in T20N, R34E, Section 5. This was attributed to new movement on the fault zone along which water had been rising. This movement opened fractures which permitted more water to move upward and/or lowered the land surface at the spring areas, which produced a temporary increase in artesian head at the spring orifices. Discharge at Dixie Hot Springs did not

appear to be affected by the earthquake; the temperature at one of the orifices was the same (59.4°C) both before and after the earthquake.

A reconnaissance report by Cohen and Everett (1963) provides an overview of the groundwater hydrology of Dixie and Fairview Valleys. Their report described a seven-valley hydrologic system (Dixie, Fairview, Pleasant, Jersey, Eastgate, Cowkick and Stingaree Valleys) that forms a closed hydrologic unit. Dixie Valley is the major groundwater sink for the system; it receives underflow from the other six valleys in the system. The Cohen and Everett report provides information on groundwater chemistry, well yields and inventories, recharge, discharge, aquifer storage and transmissive properties and related data. The report does not purport to be a detailed description of the groundwater hydrology of Dixie Valley and environs.

Additional work of a limited nature was conducted by Mariner, Rapp, Willey and Presser (1974) and Mariner, Presser, Rapp and Willey (1975) who collected chemical, gas and stable isotope data from Dixie Hot Springs, Sou Hot Springs and a few other locations in the study area. No interpretations of the data are given in their reports. Studies were also conducted by GeothermEx, Inc. (1976) and by Keplinger and Associates (1977; 1978). A substantial amount of geochemical sampling, analyses and interpretation were conducted in northern Dixie Valley as part of a multidisciplinary study by the Mackay Minerals Research Institute (1980); the work described herein was performed as part of this investigation.

H.1.4. Hydrologic Framework

The following discussion is a brief description of the groundwater hydrology of Dixie Valley, which will be useful in understanding the interpretations of and inferences from the isotope data. Most of the information can be found in Cohen and Everett (1963).

Insofar as groundwater hydrology is concerned, Dixie Valley is one of seven valleys forming a single closed hydrologic unit. Dixie Valley is topographically the lowest of these valleys; indeed, it contains the lowest point in northern Nevada. The valley is the major groundwater sink for the seven-valley system; it is believed

to receive underflow either directly or indirectly from the other six valleys. The groundwater flow into Dixie Valley is southward from Jersey and Pleasant Valleys; westward from Eastgate, Cowkick and Stingaree Valleys into Dixie Valley and then northward to the Humboldt Salt Marsh; and northward from Fairview Valley. In Dixie Valley itself, groundwater moves radially from the valley margins toward the Humboldt Salt Marsh.

Groundwater occurs under confined (artesian), semi-confined and unconfined (water-table) conditions in the unconsolidated sediments comprising the valley fill. The largest area of artesian conditions lies just south of the Humboldt Salt Marsh in the major groundwater discharge area. Semi-confined groundwater occurs in scattered areas throughout the valley. Most of the groundwater occurs in and moves through the interstices of the valley-fill sediments. However, a certain amount undoubtedly occurs as fracture-flow in the consolidated rocks. It is believed that the underflow from Pleasant Valley occurs in this manner. This type of flow is more important with respect to the geothermal reservoir than it is to the shallow groundwater reservoir.

Average annual precipitation in Dixie Valley is estimated to be about 456,000 acre-feet. Of this amount, 6,000 acre-feet is believed to reach the groundwater reservoir as recharge. Most of this recharge is believed to originate as precipitation in the mountainous regions: the Stillwater Range and Clan Alpine Mountains. Some of the mountainous regions receive more than 20 inches of precipitation annually, as compared to about 5 inches on the valley floor. In addition, the valley is believed to receive approximately 7,000 acre-feet of recharge annually as subsurface flow from the six surrounding valleys. Average annual natural discharge by evapotranspiration in Dixie Valley is approximately 16,200 acre-feet. The major discharge area is the Humboldt Salt Marsh, where groundwater is discharged by transpiration and direct evaporation from the water table. It should be noted that these recharge and discharge estimates from Cohen and Everett (1963) result in an average annual discrepancy of 3,200 acre-feet between discharge (16,200 acre-feet) and total recharge (13,000 acre-feet). This discrepancy results from the

limited data and the uncertainties inherent in the methods used to estimate natural groundwater discharge and recharge. Since recharge is less than discharge, Dixie Valley which theoretically was in a state of equilibrium at the time of the Cohen and Everett report, could be receiving more recharge via precipitation and/or subsurface inflow from the surrounding valleys. However, it is difficult to draw any substantive conclusions since neither the recharge nor the discharge figure is particularly accurate.

Springs are numerous in the valley; most of them discharge thermal water. In terms of this study, the most important springs are the three major hot spring systems: Dixie Hot Springs, Hyder Hot Springs and Sou Hot Springs. The Dixie Hot Springs system is comprised of about 35 springs and seeps, which emerge from alluvium over an area of about four square miles. The Hyder Hot Springs system emerges in the middle of the valley where the bedrock is apparently covered by thick alluvium. Travertine deposits up to 100 feet thick have been deposited here. The Sou Hot Springs system is found at the northern end of the study area, in a region characterized primarily by bedrock with hot spring deposits.

H.2 Analytical Results

H.2.1 Isotopic Characteristics of Dixie Valley and Surrounding Waters

Table H-1 lists the isotopic (^2H or D, ^{18}O and ^3H) characteristics of Dixie Valley and surrounding waters. The data for D and ^{18}O are reported as per mil (parts per thousand) variations from Standard Mean Ocean Water (SMOW). The analytical precision is $\pm 0.2\text{‰}$ for $\delta^{18}\text{O}$ and $\pm 3\text{‰}$ for δD . Tritium data are reported in tritium units (TU); one TU equals one atom of ^3H per 10^{18} atoms of ^1H .

The geographical distribution of samples can be ascertained by referring to Plate H-1. This plate covers only the area of greatest interest to the Southland Royalty Company. Sample locations not shown on Plate H-1 are listed in Table H-2.

Throughout the rest of this report, references will be made to the hydrogeochemistry of the area. The hydrogeochemical study was

Table H-1. Isotopic Analyses of Dixie Valley and Surrounding Waters.

Code	Sample Number	Sampling Date	T (°C)	δD (‰)	$\delta^{18}O$ (‰)	3H (TU)
Intermediate temperature springs in the vicinity of McCoy Ranch						
M	DV5	3-20-79	39.0	-131	-15.8	<10
M	DV6	3-20-79	29.0	-135	-14.6	<100
M	DV7	3-20-79	--	-130	-15.8	<100
M	DV8	3-20-79	--	-134	-15.6	<100
M	DV9	3-20-79	--	-131	-15.2	<100
M	DV11	3-20-79	43.0	-124	-15.4	<100
M	DV116	3-20-79	40.0	-132/-136	-15.7	--
Exploration well DF45-14						
1	DV69	7-9-79	68.5	-109	-10.5/-10.6	--
1	DV70	7-9-79	75.5	-117	-12.6	--
1	DV71	7-9-79	82.0	-113	-12.8	--
1	DV72A	7-9-79	76.5	-109	-12.6	--
1	DV72B	7-9-79	76.5	-119	-12.6	--
3	DV90	9-12-79	94.0	-114/-117	-14.4	<10
1	DV126	12-5-79	86.5	-121/-118	-14.6	<5
1	DV128	3-15-80	48.0	-119	-14.8	--
Exploration well DF66-21 and thermal well SR2-A (DV30)						
2	DV81	8-14-79	51.0	-124	-12.6/-12.5	<100
2	DV82	8-14-79	86.0	-126	-12.7	--
2	DV93	9-14-79	83.0	-101	-11.7	--
4	DV94	9-15-79	--	--	--	<10
2	DV125	12-5-79	32.5	-120	-13.9	<5
T	DV30	5-15-79	64.5	-123	-14.3/14.8	--
Dixie Hot Springs						
Code	Sample Number	Sampling Date	T (°C)	δD (‰)	$\delta^{18}O$ (‰)	3H (TU)
D	DV23	3-20-79	73.0	-120	-15.3	<10
D	DV24	3-20-79	55.0	-126	-15.3	--
D	DH6	7-23-79	73.0	-125	-15.4	--
D	DH9	7-23-79	73.0	-134	-15.8	--
D	DH15	7-23-79	52.0	-133	-15.8/-15.5	--
D	DH16	7-23-79	26.0	-129	-15.4	--
D	DHS17	7-23-79	20.0	-121	-15.8	--

Table H-1 (con't.)

Code	Sample Number	Sampling Date	T (°C)	δD (‰)	δ ¹⁸ O (‰)	³ H (TU)
Sou Hot Springs						
S	DV1	3-20-79	55.0	-131	-15.0	300
S	DV2	3-20-79	51.0	-137	-14.7	<100
S	DV35	5-15-79	53.5	-127	-15.3	--
S	DV36	5-15-79	73.0	-123	-15.3	--
Hyder Hot Springs						
H	DV4	3-20-79	63.0	-132	-15.0	<10
H	DV33	5-15-79	65.5	-135	-15.0	--
H	HHS1	7-23-79	75.0	-130	-15.8	--
Artesian wells in the vicinity of Dixie Settlement						
A	DV13	3-20-79	19.0	-134	-16.0	<10
A	DV14	3-20-79	19.0	-133	-16.5	<100
A	DV43	5-15-79	18.3	-127	-16.3	--
A	DV44	5-15-79	18.3	-126	-16.4	--
A	DV65	6-20-79	20.0	-130	-16.7/-16.8	--
Irrigation wells in the northern part of Dixie Valley						
I	DV34	5-15-79	19.5	-122	-15.3	--
I	DV37	5-15-79	16.0	-112/-119	-15.4	--
I	DV38	5-15-79	18.5	-125/-130	-16.0	--
I	DV39	5-15-79	22.5	-125	-16.4	--
I	DV40	5-15-79	15.5	-115	-15.7	--
I	DV100	6-25-79	19.0	-126	-15.4	--
Springs/streams from the eastern slopes of the Stillwater Range						
Code	Sample Number	Sampling Date	T (°C)	δD (‰)	δ ¹⁸ O (‰)	³ H (TU)
W	DV21	3-20-79	13.0	-117	-14.0	<100
W	DV22	3-20-79	11.0	-112/-115	-15.0	<100
W	DV45	5-15-79	14.0	-116	-14.7	--
W	DV46	5-15-79	15.75	-114	-14.2	--
W	DV47	5-15-79	18.0	-108	-14.0	--
W	DV48	5-15-79	13.5	-107	-14.3	--
W	DV49	5-15-79	15.75	-107	-14.1	--
W	DV53	6-25-79	18.0	-124	-14.1	--
W	DV67	6-25-79	16.0	-114	-14.0/-14.6	--
W	DV78	7-23-79	22.0	-125	-14.0	--
W	DV92	9-12-79	43.0	-98	-12.0	--
W	DV102	6-25-79	14.0	-115	-13.5	--
W	DV103	6-25-79	14.0	-121	-14.5	--
W	DV110	6-25-79	15.5	-129	-14.9	--

Table H-1 (con't.)

Code	Sample Number	Sampling Date	T (°C)	δD (‰)	$\delta^{18}O$ (‰)	3H (TU)
Spring/stream from the western slopes of the Stillwater Range (Carson Sink side)						
C	CS1	7-23-79	15.0	-125/-123	-15.4/-15.0	--
C	CS2	7-23-79	16.0	-123	-15.9	--
C	CS3	7-23-79	21.0	-122	-14.7	--
C	CS4	7-23-79	16.0	-122	-15.3/-15.0	--
C	CS6	7-23-79	21.0	-122	-15.7	--
C	CS8	7-23-79	24.0	-121	-14.2	--
Buckbrush seeps						
B	*DV41	5-15-79	28.0	-77/-82	-4.0/-4.2	--
B	DV42	5-15-79	15.0	-124	-16.3	--
Spring/stream from the Clan Alpine and Augusta Mountains						
Code	Sample Number	Sampling Date	T (°C)	δD (‰)	$\delta^{18}O$ (‰)	3H (TU)
E	DV3	3-20-79	11.0	-121	-15.2/-14.9	<100
E	DV10	3-20-79	15.0	-121/-123	-15.3/-15.0	<100
E	DV12	3-20-79	11.0	-117	-14.3	58
E	DV60	6-25-79	9.0	-113	-14.7	--
E	DV63	6-25-79	18.0	-117/-118	-15.2	--
E	DV64	6-25-79	--	-120	-13.6/-13.3	--
E	DV83	8-10-79	28.0	-128	-16.1/16.1	--
Miscellaneous samples (RS = Dixie Valley rain; DV20 = spring in Pleasant Valley; DV120 = spring in Fairview Valley; ECV = springs in Edwards Creek Valley)						
DR	RS1	11-3-79	--	- 93	-14.0	--
DR	RS2	11-17-79	--	-114	-14.9	--
DR	RS3	12-3-79	--	- 80	-10.7	--
EC	ECV1	10-19-79	13.5	-117	-17.2	--
EC	ECV2	10-19-79	13.0	-112	-15.5	--
EC	ECV3	10-19-79	13.0	-114	-16.6	--
P	DV20	3-20-79	21.0	-126	-15.3	--
F	DV120	10-19-79	12.0	-111	-15.6/-15.7	--

*Stagnant surface water; evaporation undoubtedly responsible for enrichment

published as a chapter (Chapter 5) of the Mackay Minerals Research Institute (MMRI, 1980).

Table H-2. Sample Locations Outside The Area Covered by Plate H-1.

Sample Number	Location
DV 66	T21N,R35E, Sec. 7
DV 67, LL 109	T21N,R34E, Sec. 6
DV 120	T17N,R35E
ECV1	T21N,R39E, Sec. 3
ECV2	T21N,R39E, Sec. 18
ECV3	T19N,R37E, Sec. 11
CS1	T27N,R37E, Sec. 30dd
CS2	T26N,R37E, Sec. 5ba
CS3	T26N,R36E, Sec. 18b
CS4	T26N,R37E, Sec. 5ba
CS6	T23N,R34E, Sec. 20
CS7	T23N,R33E, Sec. 24ad
CS8	T22N,R33E, Sec. 11ad

Figure H-1 shows the meteoric water line (MWL), $\delta D = 8(\delta^{18}O) + 10$ by Craig (1961) for world wide precipitation. The exact composition is dependent upon the temperature of water condensation and, as is the case with rain in arid environments, how much evaporation the precipitation is subjected to during its descent. Precipitation under cold conditions (high altitude snow) is more depleted with respect to both D and ^{18}O and plots on the lower portion of the MWL (Figure H-1) conversely precipitation under warmer conditions (low elevation rain) is more enriched with respect to D and ^{18}O and plots higher on the MWL (Figure H-1).

The isotopic composition of water can be altered by several processes. The most common is by evaporation. Figure H-1 shows the evaporation trend lines for a water at point A being subjected to evaporation under atmospheric conditions. The resulting water will lie on a line emanating from A. The particular slope of the line is dependent upon several factors, the most important of which is the temperature at which evaporation takes place.

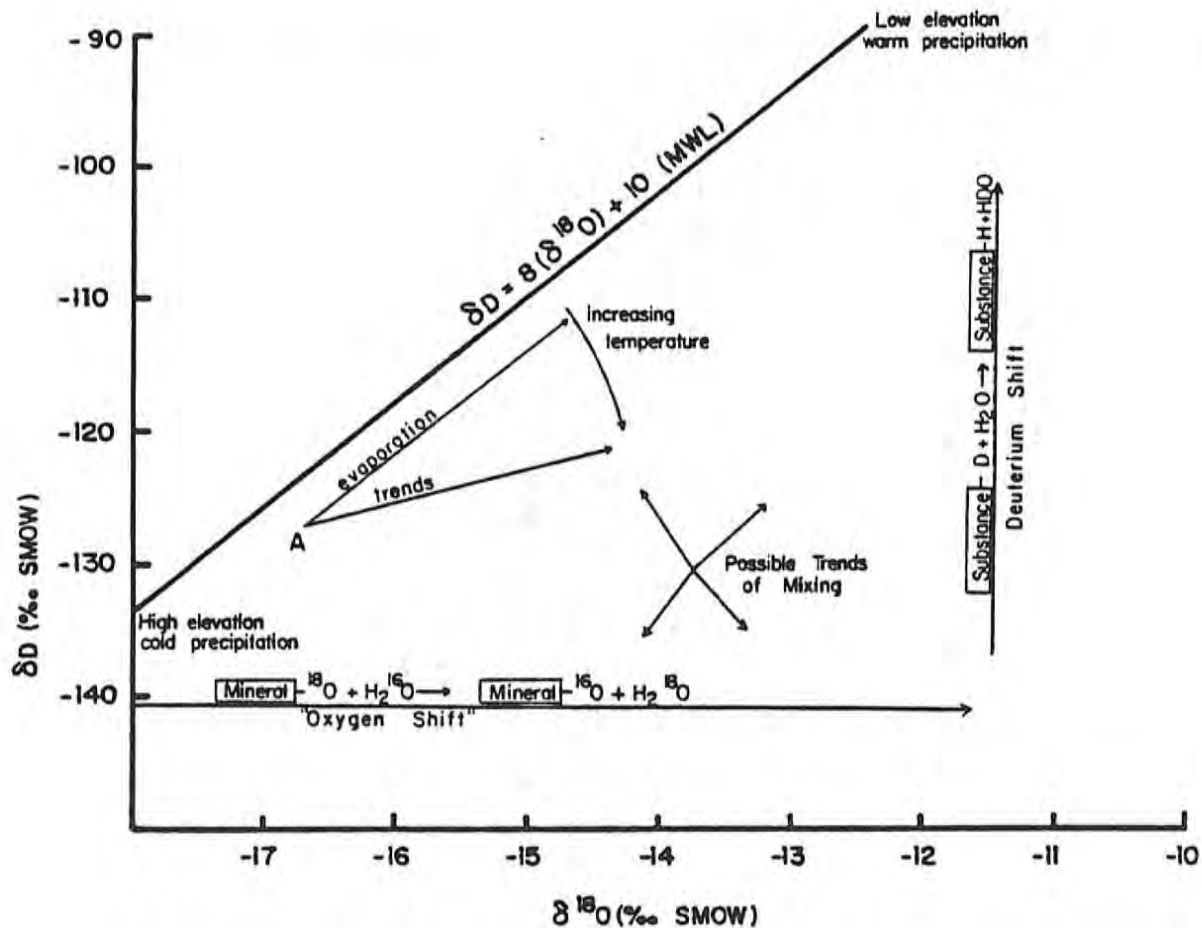


Figure H-1. General Principles of Stable Isotope Data Interpretation.

Once a water has been infiltrated and incorporated into the groundwater system, the isotopic composition will not change unless it is subjected to high temperature such as in a geothermal system. The most common, and widely accepted process which can alter the isotopic composition, occurs during water-rock reaction with respect to Oxygen-18 as shown on Figure H-1. A less common process which is also less accepted is a so-called deuterium shift. The chemical reaction is analogous to the oxygen shift. The water must react with hydrogen-bearing rocks (clays), H_2S or organic substances.

Another process by which isotopic composition can be altered is through mixing. The mixing of two waters of a different isotopic composition will produce a water of an intermediate isotopic composition. If mixing is done in varying proportions mixing lines can be developed on a δD vs $\delta^{18}O$ graph which are usually linear. Mixing can involve any water of any isotopic composition.

H.2.2. Dixie Hot Springs

Seven samples from Dixie Hot Springs were analyzed for stable isotopes. The springs have wide variations in temperature and earlier work (MMRI, 1980) indicated wide variations in chemistry (TDS levels ranging from 638 to 1470 ppm). The $\delta^{18}\text{O}$ values show relatively small variations, whereas variations in δD are significant, although no boiling springs occur. Figure H-2 shows a plot of deuterium and oxygen-18 compositions. Spring temperatures are in parenthesis next to the sample number. Deuterium values range between those characteristic of high elevation waters such as the artesian wells (Section H.2.5., Figure H-5) and low elevation waters, such as those from the Still-water Range (Section H.2.6, Figure H-6).

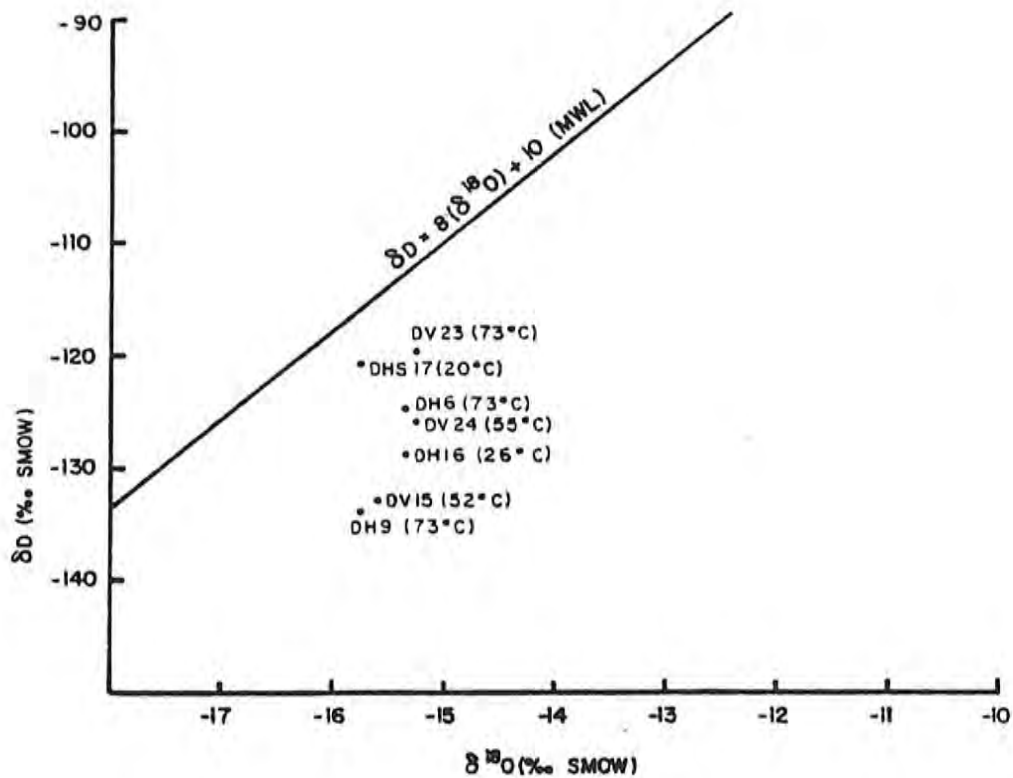


Figure H-2. Deuterium and Oxygen-18 Compositions of Dixie Hot Springs.

Apparently a high elevation water similar in isotopic composition to that from the artesian wells is heated, resulting in a slight oxygen shift. This type of water is probably the original contributor to the Dixie Hot Springs system, which undergoes various degrees of mixing with cold water from the Stillwater Range. Variations in the process of cooling probably account for the differences in temperature.

The Dixie Hot Springs system is located at the intersection of the Dixie Meadows and Mississippi faults (Plate IV, MMRI, 1980). Since the artesian wells are also in the vicinity of the Mississippi fault it lends support to the hypothesis that waters of an isotopic composition similar to that of the artesian wells is mixed with water from the Stillwater Range.

H.2.3. Hyder Hot Springs

The Hyder Hot Springs system is a group of approximately twenty hot springs issuing from a travertine mound in the north central part of Dixie Valley. A thick cover of alluvium apparently exists in this portion of the valley. Hydrogeochemical evidence (relatively constant TDS contents and ionic concentrations among the various orifices) indicated that the Hyder system is relatively homogeneous with little mixing. The springs probably issue from a single source via a well defined fault-controlled flow system. The isotopic evidence (Figure H-3) tends to support this hypothesis, although there are only three samples. There is little variation in δD or $\delta^{18}O$, indicating a probable lack of mixing with shallow groundwater.

H.2.4 Hot Springs in the Vicinity of McCoy Ranch

The isotopic compositions of these waters are relatively homogeneous (Figure H-4). The small variations are probably due to mechanisms similar to those described for the Dixie Hot Spring system. All the springs in this area seem to be associated with the Dyer Fault system (Plate IV in MMRI, 1980), which probably accounts for their isotopic and chemical homogeneity. Most of the springs seem to be similar in chemistry and deuterium composition to Hyder Hot Springs, although Hyder is apparently related to a different fault system.

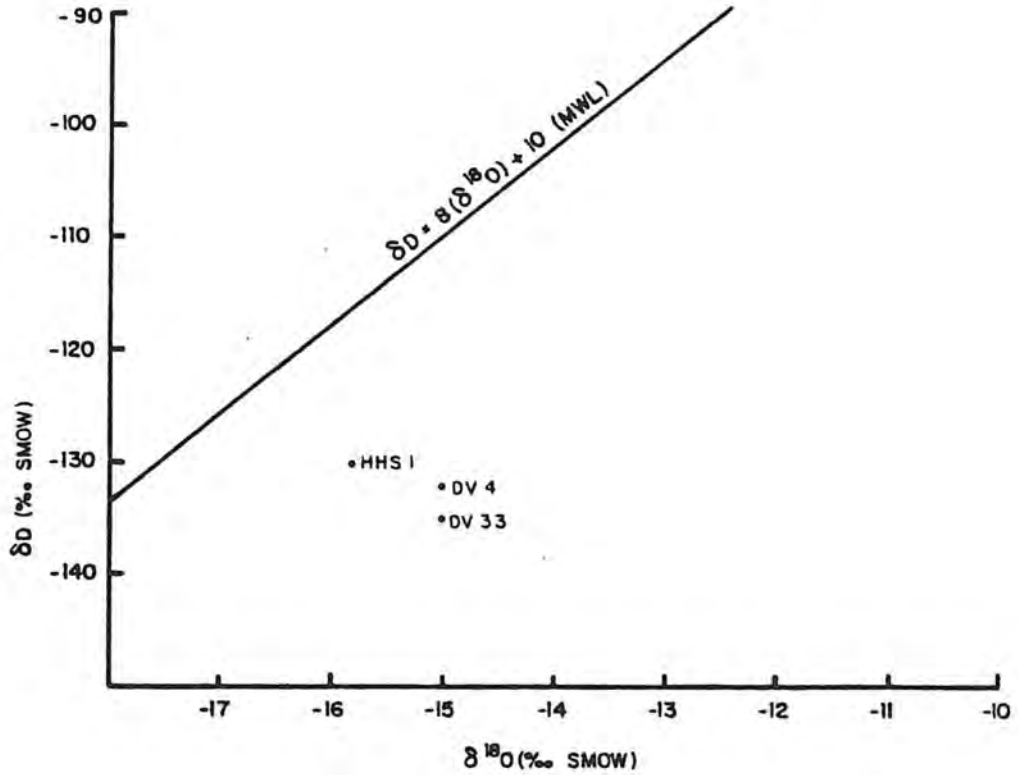


Figure H-3. Deuterium and Oxygen-18 Compositions of Hyder Hot Springs.

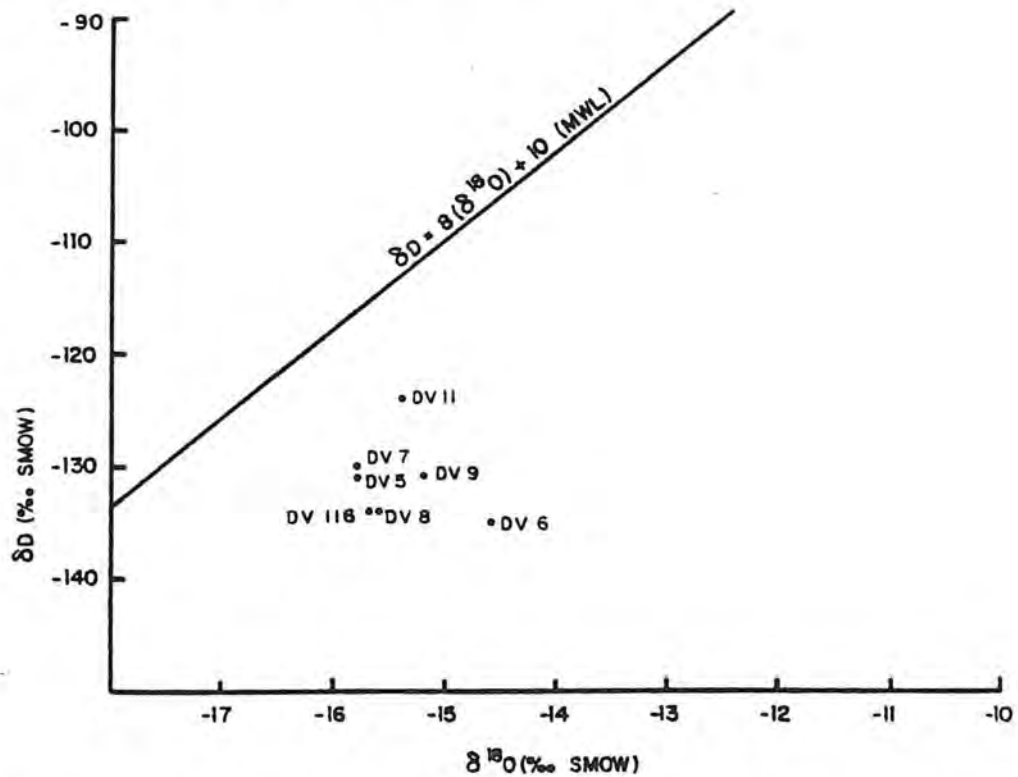


Figure H-4. Deuterium and Oxygen-18 Compositions of Hot Springs in the Vicinity of McCoy Ranch.

H.2.5 Artesian Wells at Dixie Settlement and Waters from the Eastern Mountain Ranges.

Both sets of waters were originally considered similar with respect to their chemical composition (MMRI, 1980). For this reason the Clan Alpine Mountains were, at least partially, considered as the recharge area for the waters discharged from the artesian wells. The isotopic composition of waters from the eastern mountains is variable and indicates low elevation waters (Figure H-5). Only one sample (DV83 from Shoshone Spring) plots within the range of the artesian wells, which are all high elevation waters.

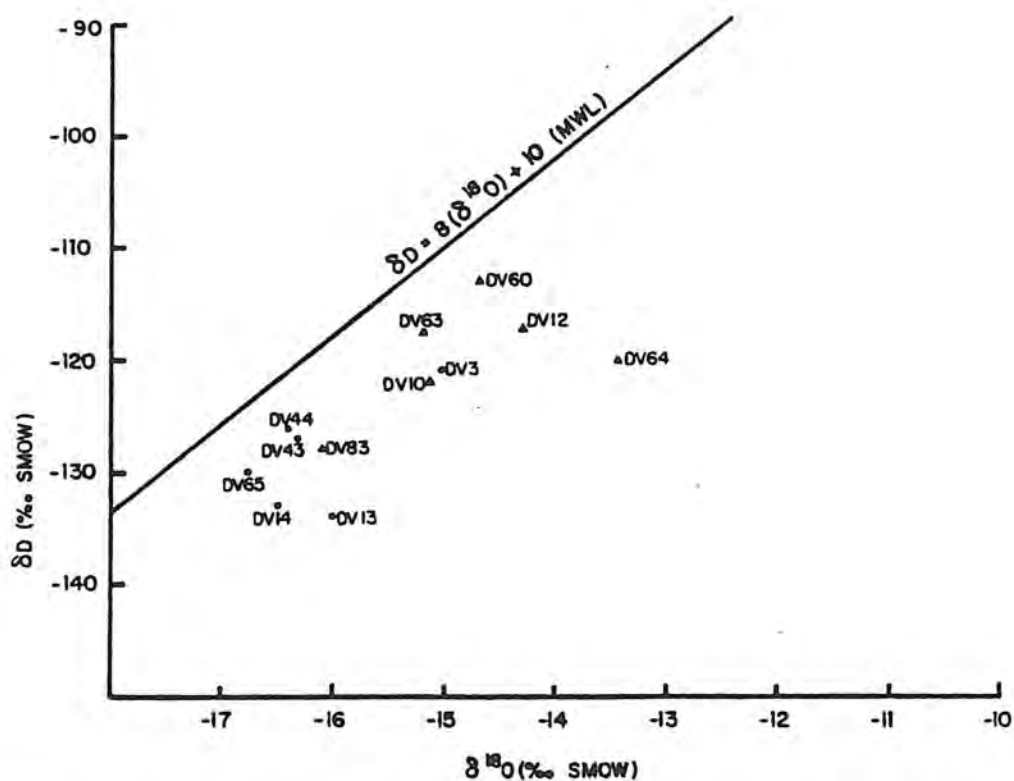


Figure H-5. Deuterium and Oxygen-18 Compositions of Artesian Wells at Dixie Settlement (•) and Waters from the Eastern Mountain Ranges (*).

According to water chemistry Shoshone Spring is an anomaly within all the other waters in the eastern mountains (DV91 in MMRI, 1980). It is a warm spring (32°C) and its chemical similarity to the artesian wells originally led to the conclusion that the artesian wells are recharged from the Clan Alpine Mountains. However, despite

their chemical and isotopic similarity it remains questionable whether there is a hydrologic connection between Shoshone Spring and the artesian wells, since they are about fifteen miles apart on opposite sides of a mountain range. Shoshone Spring remains an anomaly in the Clan Alpine Mountains, in that it constitutes a manifestation of thermal water at an altitude 1500 feet above the bottom of Dixie Valley.

H.2.6 Waters from the Stillwater Range

Variations in $\delta^{18}\text{O}$ and δD are significant in this area (Figure H-6). Some of the variations can be explained readily. DV92 is very likely affected by evaporation due to heating since it was diverted across the valley in a PVC-pipe to the site of DF66-21. Variations due to evaporation might also apply to some of the samples from Cottonwood Creek and the west side of the Stillwater Range. On the other hand, samples which were also likely subjected to evaporation, such as those from White Rock Canyon Creek (e.g. DV45 to DV49) do not

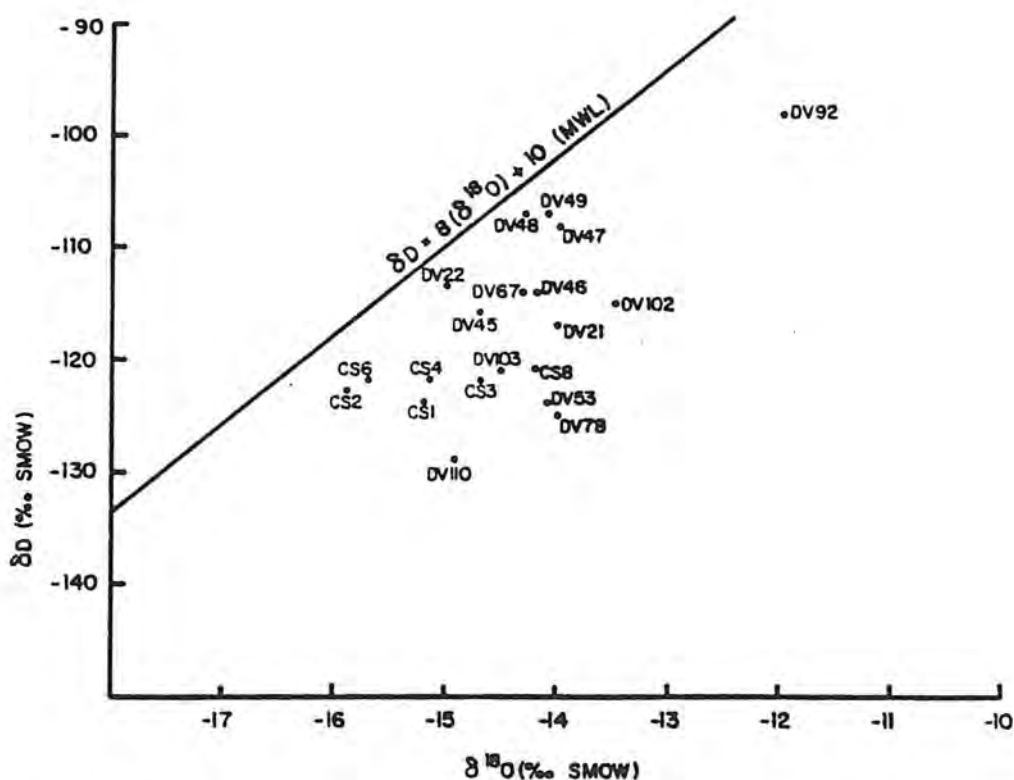


Figure H-6. Deuterium and Oxygen-18 Compositions of Waters from the Stillwater Range.

display variations that could be interpreted as evaporation. Generally variations in isotopic compositions from the Stillwater Range are due to evaporation and different elevations of precipitation and no oxygen shifts due to temperature can be recognized.

In the hydrogeochemical investigation it was found that all the streams in the Stillwater Range deposit CaCO_3 and that their TDS and calculated CO_2 -pressures were anomalously high. However, the isotopic values are as expected for this type of terrain. Nevertheless, because of the chemical anomalies more work on hydrogeochemistry and isotopes should be undertaken.

H.2.7 Sou Hot Springs

The four samples from Sou Hot Springs (DV1, DV2, DV35 and DV36) exhibit little variation in $\delta^{18}\text{O}$ among themselves, but large variations in δD do exist--as much as in Dixie Hot Springs (Figure H-7). DV2 is the most depleted in δD of any individual hot spring sample. The low δD values indicate a high elevation recharge source for the Sou

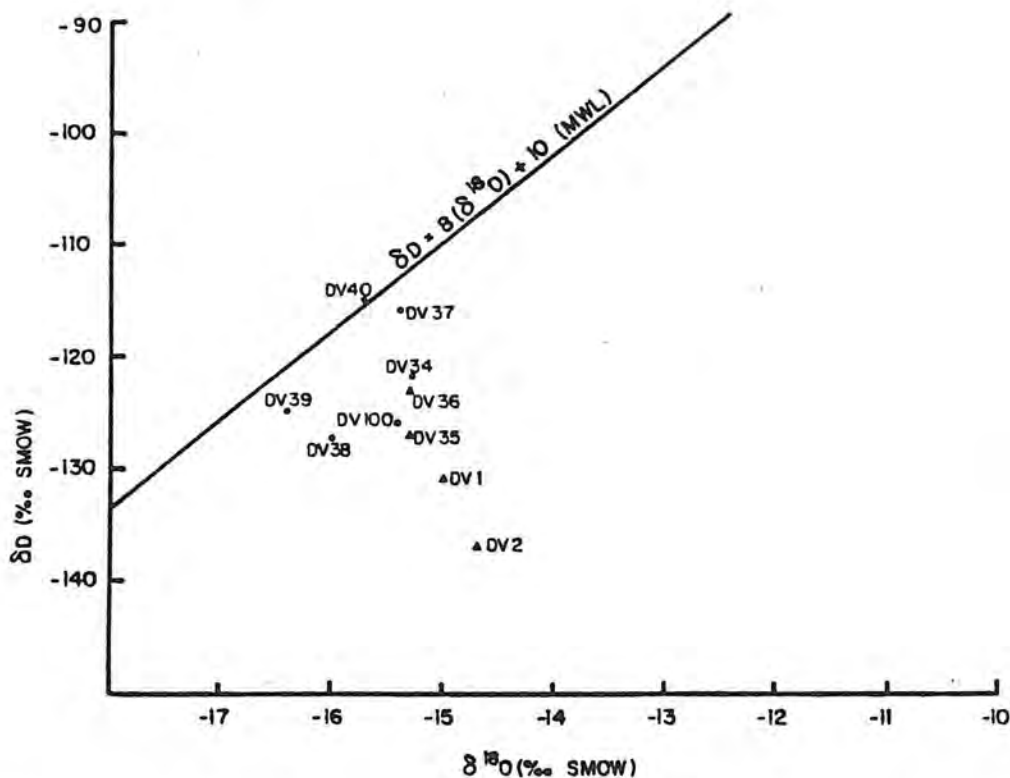


Figure H-7. Deuterium and Oxygen-18 Compositions of the Irrigation Wells (•) and Sou Hot Springs (▲).

system. The variation in δD between DV2 and DV35, which were taken from the same orifice two months apart, implies different degrees of mixing between thermal ^{18}O shifted waters and shallow groundwater at different times. The groundwater found in the irrigation wells to the east of Sou Hot Springs (DV34, DV37 and DV40) may be similar to the water that mixes with the thermal water derived at high elevations.

H.2.8. Deep Wells DF45-14 and DF66-21

Isotopic analyses of representative samples from these two wells would undoubtedly provide a wealth of information on the nature of the Dixie Valley geothermal system. A total of 12 samples (7 from DF45-14 and 5 from DF66-21) were collected from these wells. Unfortunately the validity of some of these samples is questionable. Possible casing ruptures and subsequent contamination by shallow groundwater, drilling fluid contamination, incomplete flushing of the wells, etc. cast doubts on the validity of the isotopic analyses from these wells. The following discussion must therefore be read in light of the aforementioned potential sources of contamination.

The initial sample (DV69) from DF45-14 exhibited the largest ^{18}O shift of any sample collected during the project (Figure H-8). However, DV69 as well as all the other samples taken from DF45-14 on that same day (DV70 to DV72) were very likely contaminated with water from temperature gradient hole SR2-A, which was used for flushing. Subsequent sampling of this well showed a gradual diminishing in the magnitude of this shift. The last sample from this well, DV126 (collected in December 1979) showed little ^{18}O shift. This same trend was observed in DF66-21, although the shift was not as pronounced. This reduction in ^{18}O shift probably represents a mixing/contamination of the original geothermal fluid with shallow groundwater, spring waters and/or drilling fluids (DV93 is probably contaminated with water represented by DV92-Figure H-8). Little change is apparent in samples collected between September (DV90) and December (DV126) of 1979. This implies that the mixing/contamination has reached equilibrium.

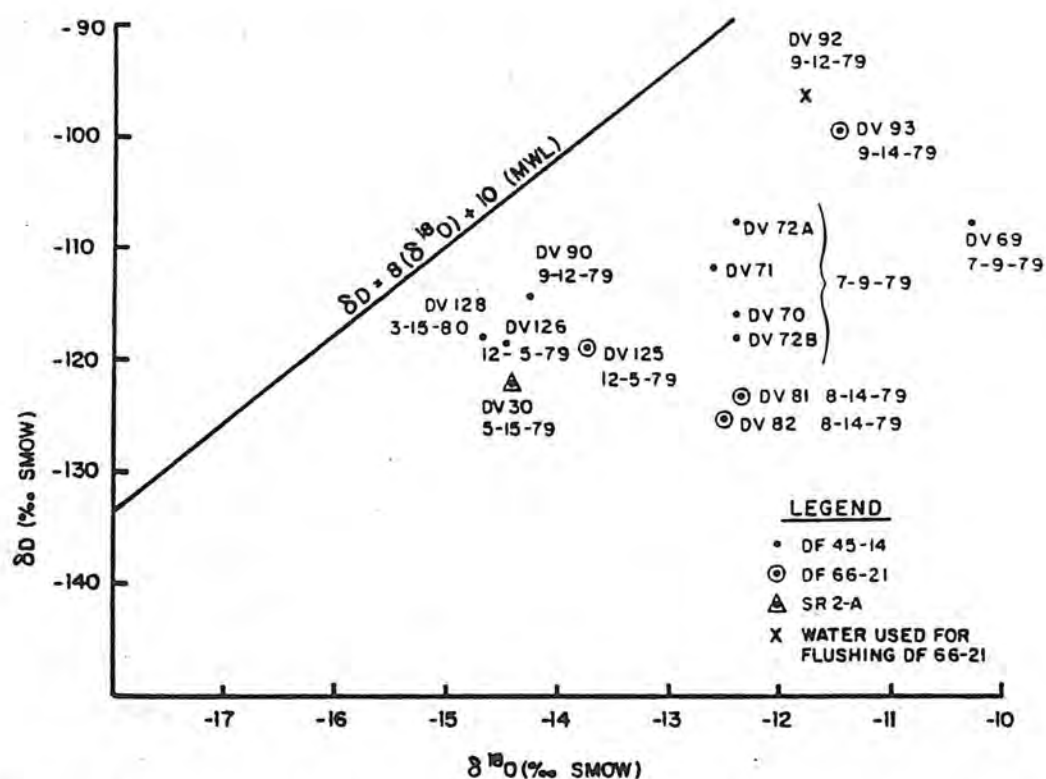


Figure H-8. Deuterium and Oxygen-18 Compositions of Deep Wells DF45-14 and DF66-21.

The last samples from these wells (DV90, DV125, DV126) plot near the waters from the eastern slopes of the Stillwater Range (DV45, DV46, DV67) and waters from the Clan Alpine Mountains (DV10, DV3, DV63). This indicates that waters of this isotopic composition are mixing with/contaminating the geothermal fluids.

The initially high deuterium contents in DF45-14 could be the results of a "deuterium shift" (Panichi and Confiantini, 1978). Shifts in deuterium contents are generally not as common as ¹⁸O shifts because rocks normally contain more oxygen than hydrogen. Notable exceptions are rocks with clay and micaceous minerals (Ellis and Mahon, 1977) and those containing organic matter. The fracture zones (see Bard, MMRI 1980) encountered in DF45-14 were encountered in the organic-rich fine-grained metasilstone/metashale, which could be responsible for the initially high levels of deuterium in this well. However, deuterium shifts are not well accepted or documented. White (1973, p. 553) stated that water-rock reactions and membrane filtration of isotopes have not been effective in changing the deuterium content of meteoric water that flows through these systems.

Isotopic exchange between H_2S and H_2O may occur in nature (Hitchon, 1963) thereby providing another mechanism for deuterium content variations. Hydrogen sulfide was noticeable in DF45-14, although no quantitative measurements of H_2S concentrations were made. Nevertheless, the possibility of contamination with shallow groundwater must still be kept in mind.

Assuming that the fluid from DF45-14 underwent a deuterium shift (DV69) to a more deuterium enriched water, and that fluids from DF66-21 were mixed with shallow groundwater, the trend lines of changing composition can be extrapolated backwards to determine an original composition for the fluid (Figure H-9). When this is done, the intersection of these two trend lines yields a hypothetical composition of an original geothermal fluid (H in Figure H-9). This hypothetical fluid has a deuterium composition which is similar to the fluids derived from the hot springs and the artesian wells. This implies recharge for the deep geothermal system from high elevation waters which underwent a significant oxygen shift (a horizontal shift is inferred assuming no boiling occurred at depth).

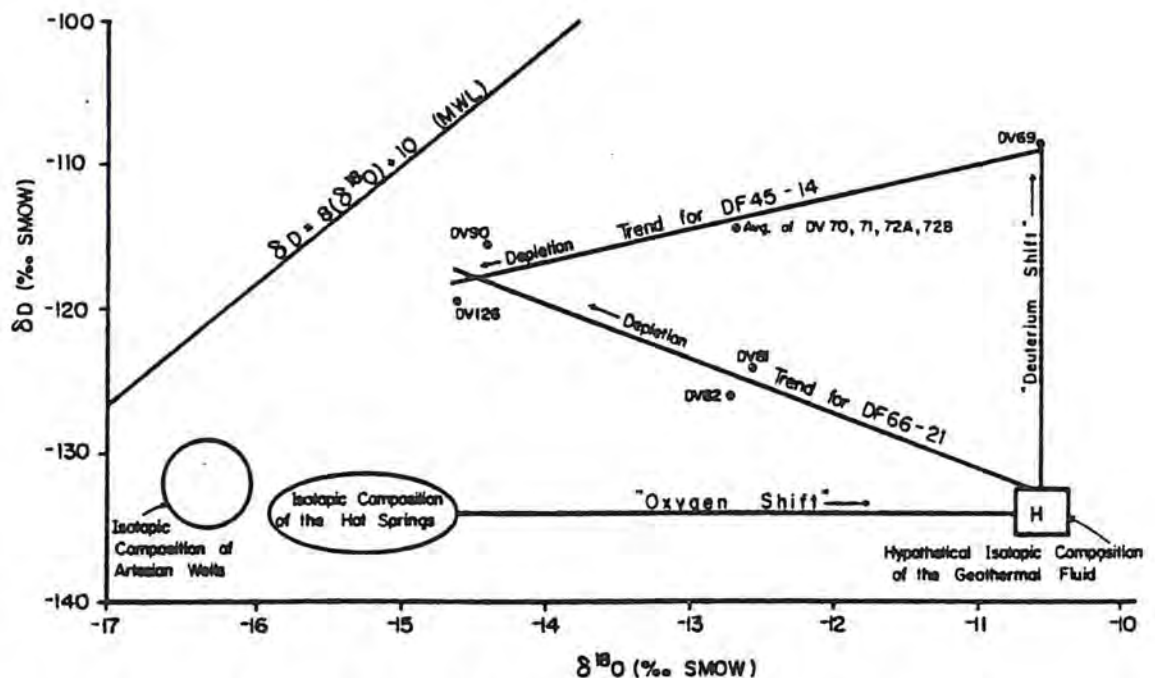


Figure H-9. Inferred Trend Lines of Changes in Isotopic Composition of Samples taken from DF45-14 and DF-66-21, and the Isotopic Composition of a Hypothetical Geothermal Fluid.

H.2.9 Interpretations of Stable Isotope and Hydrogeochemical Data.

Figure H-10 is a plot containing all isotope data collected (see Table H-3 for location codes). It allows comparisons between the various geographical groups of samples. Aside from seasonal variations, a geographical grouping is vaguely discernible.

Assuming that lower δD values indicate relatively high elevations (Figure H-1) as sources of groundwater and vice-versa, changes on the vertical axis should permit some conclusions on the sources of water and the hydrologic system as such. Hyder Hot Springs, Sou Hot Springs, the artesian wells at Dixie Settlement, most of the hot springs in the vicinity of McCoy Hot Springs and parts of Dixie Hot Springs have, in comparison to all the other samples, the highest elevations as sources. They are all discharged in the lower part of Dixie Valley. However, the samples from the surrounding mountain ranges have lower elevations as sources. Unfortunately, no snow samples from the tops of these mountains have been collected. Thus they could possibly have deuterium values comparable with those found in the waters from the lower parts of the valley.

The fact that high elevation waters discharge in the lower parts of the valley is in agreement with the classical models of arid basin hydrogeology (e.g. Maxey, 1968 and Mifflin, 1968). Figure H-11 shows that high elevation waters (with long flow paths) discharge in the center of the valley. All waters that are apparently derived from high elevations have one or another feature that indicates geothermal activity, which is another hint of deep flow (Figure H-11). Thus, it is very likely that recharge waters from the Stillwater Range, the Clan Alpine and Augusta Mountains penetrates to various depths and undergoes varying oxygen shifts under different temperature regimes. This water mixes to various degrees in the shallow subsurface with water originating from relatively lower elevations, and discharges in hot springs and wells in Dixie Valley. Systematic collection of stable isotope data of snow at various elevations from the surrounding mountains could substantiate this hypothesis.

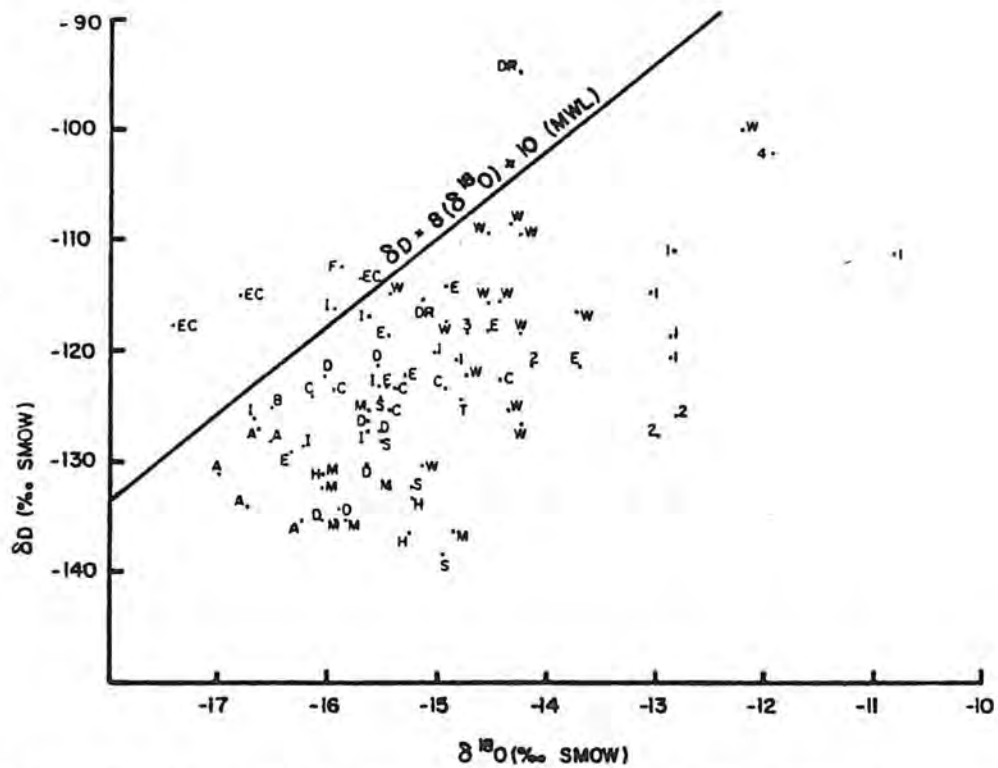


Figure H-10. Deuterium and Oxygen-18 Compositions of Dixie Valley and Surrounding Waters.

Table H-3. Location Codes for Figure H-10.

<u>Code</u>	<u>Description</u>
A	Artesian wells in the vicinity of Dixie Settlement
B	Buckbrush Seeps
C	Carson Sink (western slopes of the Stillwater Range)
D	Dixie Hot Springs
DR	Dixie Valley rain
E	Spring or stream from the eastern mountain ranges (Clan Alpine and Augusta Mountains)
EC	Edwards Creek Valley
F	Fairview Valley
H	Hydar Hot Springs
I	Irrigation wells in the northern part of Dixie Valley
M	Intermediate temperature springs around McCoy Ranch
P	Pleasant Valley
S	Sou Hot Springs
T	Temperature Gradient Hole SR2-A
W	Spring or stream from the Stillwater Range
1	DF 45-14
2	DF 66-21
3	DF 45-14
4	DF 66-21

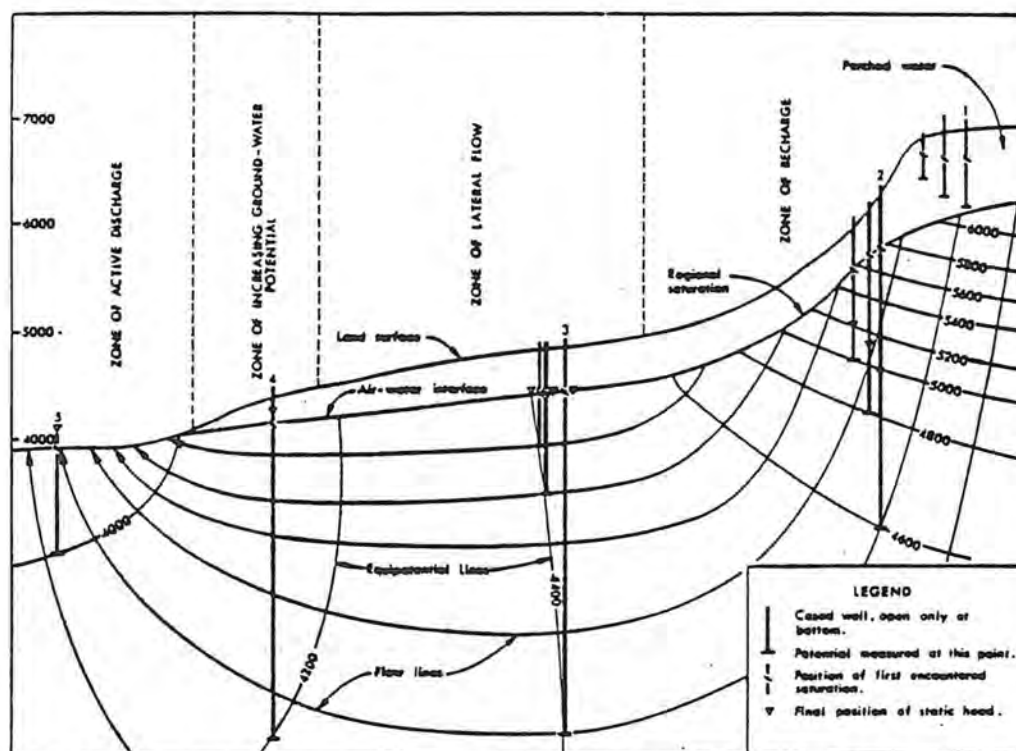


Figure H-11. Sketch of Observed Relationship in a Typical Desert Basin Flow System (after Mifflin, 1968)

However, it should be pointed out that similar deuterium values for the hot springs and geothermal wells does not necessarily imply that they are hydraulically connected. It rather implies that they are all recharged from one common source or a set of similar sources.

All waters that display geothermal characteristics are also associated with various major fault systems (Plate IV in MMRI, 1980). In general the chemical characteristics between the hot spring systems are variable, whereas their deuterium levels are similar. An exception is Hyder Hot Springs which is chemically similar to the hot springs distributed across the northeastern part of the valley, although Hyder is associated with a different fault system.

In several cases the grouping displayed in Figure H-11 has a surprising similarity with the one used earlier in the hydrogeochemical interpretation (MMRI, 1980). This supports the hypothesis that particular geographic areas are dominated by certain hydrologic, chemical and thermal regimes.

A comparison between isotopes and hydrogeochemistry also poses some contradictions. Waters from the Stillwater Range and some waters from the eastern mountains (DV10) have comparatively high TDS levels for mountain spring waters. Contrary to that, the waters discharged in the artesian wells in the center of the valley have some of the lowest TDS levels in the area. This is surprising, since the low deuterium values indicate high elevation waters with relatively long flow paths (Figure H-11) that would cause high levels of TDS.

The occurrence of a warm spring in the mountains (Shoshone Spring) discharging water that was derived from a higher elevation than the surrounding mountain springs suggests the possibility of the emergence of diluted thermal waters in some of the mountain canyons. The unusually high CO_2 pressures in most mountain springs and extensive sinter deposition in the Stillwater Range seem to support this hypothesis.

H.2.10 Tritium Concentrations

Tritium (^3H) samples were collected along with the stable isotope and chemical samples at each collection site. Due to the lack of detectable tritium in the samples and the cost of tritium analyses, they were soon terminated. Twenty-seven samples were run unenriched, a process yielding lower detection limits of 100 tritium units (TU). Only one sample, DV1, had detectable amounts. DV1 was taken from a pool in a large travertine crater at Sou Hot Springs which has no apparent discharge. It is assumed that the high reading is due to the pooling of precipitation of higher tritium content than the hot spring water and evaporation. This analysis also has an inherently large analytical error (300 ± 240 TU), which makes conclusive interpretation almost impossible.

A spring in Dyer Canyon on the western slope of the Clan Alpine Mountains did have a significant tritium level (58 ± 8 TU). No hot spring, artesian well, or deep well showed detectable quantities, indicating that the deep ground water and thermal systems are pre-bomb test (1954) waters. Some mountain spring and shallow groundwater, however, must be post-bomb water as the spring in Dyer Canyon (DV12 indicates).

H.3 Conclusions

Two of the three major hot spring systems, Dixie and Sou Hot Springs, discharge a mixture of two waters, each of which is characterized by its own isotopic composition. One type of water is more depleted with respect to deuterium ($\delta D = -134^0/00$) and originates at high elevations (> 7000 feet). This water moves downward, is heated and shifts toward a more enriched ^{18}O content. The second type of water is a non- ^{18}O shifted water that plots close to the MWL at about $\delta D = 122^0/00$. This water is derived from precipitation at lower elevations (< 7000 feet) than the first water. These waters mix in differing proportions, resulting in the composition of the waters discharging at Dixie and Sou Hot Springs. The waters from the artesian wells near Dixie Settlement and the hot springs in the vicinity of McCoy Hot Springs also exhibit this same pattern. Hyder Hot Springs is dissimilar to the above systems. The isotopic results are similar to the hydrogeochemical results in that they suggest that the Hyder system is relatively homogeneous, isolated from the other systems and emanates from a different reservoir. Very little mixing with shallow groundwater occurs in the Hyder system. Recharge is derived from high elevation snows, probably at about the same elevation as the deuterium-depleted component of the waters at Dixie and Sou Hot Springs.

The waters from DF45-14 and DF66-31 are derived from high elevations (> 7000 feet) snows. These waters become heated at depth and acquire the characteristic ^{18}O shift. In addition, the water from DF45-14 may experience a deuterium shift after reacting with clay minerals, hydrogen sulfide and organic matter. Waters from both wells probably mix with shallow groundwater, either naturally and/or by contamination during drilling. It is difficult to draw valid inferences from the DF45-14 and DF66-21 samples because of possible contamination; the samples obtained may not be representative of the geothermal fluids in the system.

In summary Dixie Valley represents a major groundwater sink for the surrounding basins. Recharge occurs in the mountains. Water infiltrated at the highest elevations is discharged in the lowest parts of the valley either in hot springs, geothermal wells, in

low temperature artesian wells and by evapotranspiration in the Humboldt Marsh.

Contrary to accepted beliefs, here groundwaters with high TDS contents are discharged in the mountains whereas low TDS waters are discharged in some parts of the central valley. It is also possible that waters with geothermal features are discharged in some parts of the mountains.

H.4 References

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APPENDIX F-5

SUMMARY OF LITHOLOGY AND ALTERATION IN SR-3,
DIXIE VALLEY, CHURCHILL COUNTY, NEVADA

M. J. Sweeney, July 1980

INTRODUCTION

Samples of washed, rotary cuttings taken every ten feet from drill-hole SR-3 were examined under a stereomicroscope. A thin-section of the cuttings was described from every 100 feet throughout the 1500 foot length of SR-3. Detailed descriptions of the cuttings and thin-sections are attached to this report. Photomicrographs of the thin-sections are also attached.

LITHOLOGIES

SR-3 is in gravels for its entire 1500 foot depth. All the gravel was derived from a common source. It is of nearly uniform composition for 1500 feet, i.e., the same types of clasts occur in about the same proportions throughout the entire hole. Metasandstones and phyllites are the most common lithic clasts. Locally the gravels are cemented by a mixture of clay[±]calcite containing sand-sized lithic and crystal clasts; this clay-sand material may be derived from volcanic ash.

The rock types present in the gravel are listed below.

- 1) Limestone: Black, foliated, very fine-grained, carbonaceous, locally phlogopitic limestone.
- 2) Phyllites: Black, foliated, carbonaceous phyllites composed of varying proportions of biotite, sericite and silt-sized quartz and feldspar. Occasionally porphyroblasts of biotite or of clinozoisite occur in some of the phyllite particles. Often disseminated anhedral magnetite is present. Pyrite was observed rarely.
- 3) Diorite: Dark grey to dark grey green, fine-grained diorite or diabase. The original mafic minerals in this rock type were pyroxenes; in some particles, unaltered pyroxene remains, but most often it has been replaced by chlorite and/or montmorillonite or sericite[±]carbonate. Black Fe-oxides, most probably mixtures of ilmenite/magnetite, are present in the diorite; locally ilmenite/magnetite is oxidized to earthy red hematite. Pyrite was not observed in any diorite fragments.
- 4) Metasandstones: White, buff, pale pink, orange and greenish. These particles are well indurated and never friable. A small proportion of the metasandstones in most every sample contain goethite pseudomorphing a disseminated cubic mineral (pyrite??). Only rarely was unoxidized pyrite observed in metasandstone particles.

It is composed 1) of clay which appears in thin-section to be a montmorillonite, 2) of angular sand grains of quartz, feldspar, mafic minerals, and carbonate, 3) of very fine-grained calcite cement, and 4) of sand-, pebble- and cobble-sized lithic clasts. The lithic clasts include all of the rock types described above in this report.

The clay-sand-lithic material is locally abundant in the hole; its abundance seems to correlate with how clayey the unwashed samples are. The letters, n, s, m and st in the depth column of the stratigraphic log indicate how clayey the unwashed samples are. The symbol "n" indicates that the sample was fairly free of clay, "s" slightly clayey, "m" moderately clayey and "st" strongly clay-cemented. The clay-sand-lithic material is compositionally and behaviorly similar to the bentonite-gel drilling mud used in SR-3; there is no definitive way of deciding which type of material, drilling mud or clay-sand-lithic material, is dominant in the clayey horizons. Some of the clay-rich samples, particularly those at 600', contain pyritic mudstone; the mudstone may have been deposited originally as a lacustrine bed.

The clay-sand material may have originated as volcanic ash. This is suggested by the heterogeneity of the crystal clasts and by the possibility that the clay could be a product of altered glass shards. The clay-sand material and the lithic clasts are thoroughly mixed together; the clasts and ash were probably transported together to the site of SR-3.

Thin (less than 1 mm wide) calcite veins were seen in consolidated fragments of clay-sand-lithic material which survived the washing process. Only once was a pyrite crystal seen in the calcite veins; the pyrite was attached to the surface of the vein. No disseminated pyrite was seen in the clay-sand material.

14) Pyrite crystals[±]clay: Loose pyrite crystals, which occasionally are associated with a white clay (kaolinite?), very fine-grained sericite? and calcite occur in all samples below 590 feet. Frequently these crystals are oxidized. Possibly, these loose crystals were derived from pyrite deposited in open fractures in the consolidated gravels.

Alteration

Unraveling the alteration history of gravels, especially those sampled by rotary drilling, is not straightforward. The lithic clasts may contain alteration assemblages developed originally at their sources and not at the present site. Soft or friable assemblages and veins are easily disaggregated by the process of drilling rotary holes and these same assemblages are removed from the sample by washing.

Pyritization. Pyrite occurs 1) as loose particles of subhedral crystal aggregates, sometimes associated with a white clay and calcite, 2) as loose euhedral crystals 1 mm or less across, and 3) as disseminations in siltstone, metasandstones, phyllites and mudstones. Nearly all of the pyrite? in the metasandstones has been replaced by goethite. A smaller proportion of the pyrite in siltstones has been oxidized and even less of that in mudstones has been oxidized. The significance of the disseminated pyrite cannot be evaluated without knowing if pyrite is present at the source area of these clasts.

The loose pyrite crystals are possibly from open veins or from a mudstone disaggregated by drilling/sample washing. The total volume of pyrite in the samples from SR-3 ranges between 0.1 and 0.5 volume % through the hole, including both disseminated and loose pyrite crystals. There is a fair probability that little of this pyrite was deposited as the result of hydrothermal processes occurring at the site of this drill hole, SR-3. The possibility that little of the pyrite was locally deposited is supported by the following observations: 1) only a trace of pyrite was observed in a vein in the clay-sand-lithic matrix material, and 2) no pyrite was seen in the diorite/diabase fragments. Both of these lithologies are usually more susceptible to pyritization than quartz-rich rocks.

Calcite Veining. Thin calcite veins occur in particles of clay-sand-lithic matrix material. Only once was a small pyrite crystal observed on the surface of one of these calcite veins. Very fine-grained calcite was observed occasionally on the surface of lithic clasts; this material is possibly caliche.

Quartz Veining. Drusy quartz veins occur in metasandstones and phyllites. Most probably this veining is related to events at the source area of these particles.

Opal(?) Veining. Opal-quartz occurs in thin veins in clay-sand matrix material. They were observed only in upper 600' of the drill hole.

Summary. Drill hole SR-3 has sampled 1500 feet of gravels locally cemented by sand-clay-carbonate material that is possibly derived from volcanic ash. Many of the clasts in the gravel are weakly pyritized; the pyrite possibly was deposited initially at the source area of these clasts.

The only alteration definitely related to the site of the drill hole is weak calcite as well as opal-quartz veining developed in the sand-clay matrix material.

Oxidation of pyrite is fairly complete down to 1500', the total depth of the hole.

Clay Analysis

An unwashed sample of strongly clay-cemented drill cuttings was prepared for clay analysis. The sample selected for analysis is from the 1130'-1140' interval. The sample of chips was washed in de-ionized water in a blender. Sodium tripolyphosphate was added to peptize suspended clays. The suspended material was then centrifuged at 1000 r.p.m. to remove larger than clay-sized material. Suspended clay was dropped by centrifuging at 4000 r.p.m. A portion of the dropped clay fraction was smeared on a glass slide and air dried. This sample was analyzed by x-ray diffraction. The clay fraction from the 1130'-1140' interval contains calcite, montmorillonite, illite and kaolinite in order of decreasing abundance.

Clay analysis was also made of a composite of clay-sand fragments which had survived washing. The sample consisted of fragments collected from throughout the hole during chip logging. X-ray diffraction analysis shows that this sample also contains calcite, montmorillonite, illite and kaolinite. Relative peak intensities on this XRD pattern and the previously discussed pattern are very similar. Drilling mud does not appear to have strongly affected peak intensities obtained on the 1130'-1140' sample.

The two patterns are attached to this report.

DESCRIPTIONS OF THIN-SECTIONS FROM SR-3,
DIXIE VALLEY, CHURCHILL CO., NEVADA

100'-110'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u> : Mostly very fine-grained; grain size 0.1 mm or less, usually 0.01 mm. Rock exhibits fine-scale foliation. Contains porphyroblasts? or detrital flakes of phlogopite; these mica flakes oriented parallel to rock foliation. Very fine-grained carbonaceous? particles (=black dust) occur throughout rock, distributed both inter- and intragranularly. One of the limestone chips contains a vein filled with elongate quartz crystals oriented perpendicularly to vein walls. Ghosts of fossils are still recognizable in most chips.	7	33
<u>Phyllite</u> : Very fine-grained. Composed of foliated biotite, muscovite and silt-sized quartz and feldspar grains.	2	10
<u>Diorite/Diabase</u> : One unaltered chip is composed of unoriented plagioclase laths, and lesser pyroxene and possibly olivine. In another chip, mafic minerals have been replaced by chlorite. In another, montmorillonite has replaced mafic minerals, and kaolinite has replaced plagioclase.	3	14
<u>Metasandstone</u> : One grain composed of sand-sized quartz grains and muscovite-chlorite porphyroblasts. One composed of subangular to subrounded quartz grains about 0.1 mm across; sorting good. Silica, clay and calcite occur as cement.	2	10
<u>Marble</u> : Grain size about 1 mm; granoblastic. One chip contains patches of chert. Another contains patches of more coarsely crystalline quartz (recrystallized chert?). The latter also contains patches of montmorillonite after an unidentified mineral (low birefringence=first order grey; low 2V, 10-30°; opt +; good cleavage).	3	14
<u>Chert</u> : Recrystallized; quartz occurs as fine-grained feathery crystals.	3	14
<u>Silicified Cataclasite</u> : Original rock type not known; now composed of subhedral quartz of variable grain size. Patches of brown clay (montmorillonite?) form 5-10% of chip.	1	5
Total Number of Chips Examined	21	100%

190'-200'

Rock Type	Number of Particles	Percentage of Sample
<p><u>Phyllite:</u> This category also includes a few clasts of hornfelsed muddy siltstones; the only significant difference between hornfelses and phyllites is the degree or complete lack of foliation of mica in the phyllites.</p> <p>Very fine-grained, metamorphosed clayey siltstones and silty mudstones; usually composed of subequal amounts of silt-sized quartz and feldspar and of very fine-grained white micas, lesser biotite and of variable amounts of chlorite (0-50%). Micas are usually well foliated. In some chips, biotite occurs in sparse, small clots of unoriented crystals; these are spotted hornfelses. Porphyroblasts of clinozoisite also occur in some biotite-spotted hornfelses.</p> <p>Variable amounts of very fine-grained, black (= carbonaceous material?, magnetite?) are disseminated throughout these rocks. Trace amounts of goethite occur (after disseminated magnetite?, pyrite?).</p>	27	73
<p><u>Diorite:</u> Composed of 0.3 to 1.5 mm long laths of plagioclase, interstitial chlorite-clay altered mafic crystals which comprise 7 to 15% of rock. Interstitial quartz forms 2 to 4% of rock. Alteration of plagioclase varies from chip to chip; sericite-clay[±]carbonate replacement of feldspar ranges between 20 and 70%. In the least altered chip, disseminated magnetite/ilmenite forms 1 to 2% of rock.</p>	4	11
<p><u>Metasandstones:</u> Fine-grained; sand grains of quartz; one contains 5-10% carbonate as cement; one is moderately foliated.</p>	3	8
<p><u>Marble:</u> Fine- to medium-grained.</p>	2	5
<p><u>Clay-Sand-Lithic Matrix Material:</u> Clayey matrix containing sand-sized, angular crystal fragments of quartz, feldspar and pyroxene. This material is attached to one of the phyllite chips; presumably, the clay-sand material is the matrix for the alluvial clasts.</p>	1	3
Total Number of Particles Examined	37	100%

290'-300'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u> : Very fine-grained; foliated.	11	10
<u>Phyllite</u> : Usually composed of foliated, fine-grained sericite and biotite in variable proportions and of silt-sized quartz and feldspar. Also contains very fine-grained, disseminated, black material (=graphite?, magnetite?).	20	18
<u>Diorite</u> : Mafic minerals altered to chlorite, montmorillonite and calcite.	14	12
<u>Metasandstone</u> : Composed of fine-grained, well sorted quartz; calcite cement present in some particles.	25	22
<u>Siltstone</u>	1	1
<u>Clayey Siltstone</u>	2	2
<u>Mudstone</u> : Composed of very fine-grained sericite/illite, sometimes with biotite porphyroblasts. Traces of goethite (after pyrite?) in a few chips.	14	12
<u>Volcanics</u> : Rhyolite?, composed of feathery feldspar crystals and quartz.	3	3
<u>Marble</u>	5	4
<u>Chert</u> : Recrystallized. Composed of very fine-grained, feathery crystals of quartz; contains variable proportions of silt-sized quartz grains.	16	14
<u>Vein Quartz</u>	2	2
Total Number of Particles Examined	113	100%

390'-400'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u> : Very fine-grained; foliated; \pm mica; contains abundant, fine-grained, carbonaceous? material. Black or dark grey in hand-specimen.	2	2
<u>Phyllite</u> : Fine-grained; foliated; contains variable proportions of sericite and biotite, as well as, silt-sized grains of quartz and feldspar. Very fine-grained, black material (graphite?, Fe-oxides?) is disseminated throughout most clasts.	43	33
<u>Diorite</u> : Sericite-chlorite-clay altered.	2	2
<u>Metasandstones</u> : Such particles range in composition from nearly pure quartzites to containing 30% sericite \pm chlorite and/or calcite. Goethite/hematite (after pyrite?) porphyroblasts(?) poikilitically encloses quartz grains in one chip; no other chips contained disseminated goethite and/or pyrite.	65	50
<u>Clayey Siltstone</u> : Contains abundant sericite/illite as well as silt.	2	2
<u>Mudstone</u> : Composed of very fine-grained unfoliated sericite and chlorite in varying proportions; also contains 0 to 30% silt.	3	2
<u>Granite</u>	1	1
<u>Marble</u> : Fine- to medium-grained.	4	3
<u>Chert</u> : Recrystallized to feathery quartz.	6	5
<u>Pyrite</u> : Unoxidized; a free, 0.5 mm grain.	1	1
<u>Chlorite Schist</u>	2	2
Total Number of Particles Examined	131	100%

490'-500'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u> : Very fine-grained; usually foliated; carbonaceous; rarely contains phlogopite flakes.	15	19
<u>Phyllite</u> : Very fine-grained; composed of variable proportions of sericite, biotite and silt-sized quartz and feldspar. Foliated. Contains very fine-grained, disseminated black material, graphite?, Fe-oxides?. Coarse-grained Fe-oxides, magnetite [±] hematite, occur in trace amounts as disseminations.	23	29
<u>Metasandstone</u>	14	18
<u>Siltstone</u> : Very fine-grained; contains silt-sized quartz and feldspar as well as variable amounts of sericite/illite [±] chlorite.	3	4
<u>Marble</u>	12	15
<u>Chert</u> : Composed of fine-grained feathery quartz with variable amounts of carbonate and silt- and/or sand-sized quartz grains.	10	13
<u>Clay-Sand-Lithic Matrix Material</u> : Sand- and pebble-sized clasts in a clay-rich matrix; this material is probably the matrix of the alluvium.	2	3
Total Number of Particles Examined	79	100%

590'-600'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u>	5	6
<u>Phyllite</u>	28	36
<u>Diorite</u>	3	4
<u>Metasandstones</u>	26	33
<u>Siltstone</u>	4	5
<u>Mudstone</u> : Pyritic; very soft; pale grey-green in hand-specimen; disaggregates on wetting; forms about 30% of washed sample, most chips were destroyed in thin-section making process so they are rare in thin-section.	1	1
<u>Volcanics</u> : Latite?	1	1
<u>Marble</u>	2	3
<u>Chert</u>	6	8
<u>Vein Quartz</u>	1	1
<u>Clay-Sand-Lithic Matrix Material</u>	1	1
Total Number of Particles Examined	78	100%

790'-800'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u>	10	6
<u>Phyllite</u> : Composed of very fine-grained foliated sericite, biotite (10-70%) and silt-sized grains of quartz and feldspar. Contains 0 to 1% disseminated pyrite. Rarely contains porphyroblasts of feldspar and/or biotite. Almost always contains finely disseminated black material, graphite?	46	25
<u>Diorite</u> : Mafic minerals have gone to montmorillonite.	3	2
<u>Metasandstone</u> : Goethite (after pyrite?) in a few particles.	49	27
<u>Siltstone</u> : Usually sericitic.	12	7
<u>Mudstone</u> : Composed of very fine-grained, clay-sized material. Contains 0.1-0.3% pyrite (usually oxidized to goethite); contains 0 to 15% silt-sized quartz and feldspar.	26	14
<u>Volcanics</u> : Andesite, illite-replaced.	2	1
<u>Marble</u>	14	8
<u>Chert</u>	11	6
<u>Vein Quartz</u>	1	1
<u>Vein Calcite</u>	1	1
<u>Quartz and/or Feldspar Schist</u>	4	2
<u>Chlorite Schist</u>	1	1
<u>Sand-Clay Matrix Material</u>	1	1
Total Number of Particles Examined	181	100%

890' - 900'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u>	14	12
<u>Phyllite</u>	17	14
<u>Diorite</u>	1	1
<u>Metasandstones</u>	40	33
<u>Siltstones</u>	14	12
<u>Mudstone</u>	6	5
<u>Volcanics</u> : Andesite?, altered to clay and hematite.	1	1
<u>Marble</u>	10	8
<u>Chert</u> : Recrystallized.	5	4
<u>Vein Quartz</u> : One piece coated with limonite and quartz; one vein cuts marble; one contains vugs filled with clay-pyrite.	5	4
<u>Vein Calcite</u> : Associated with hematite and quartz.	1	1
<u>Schist</u>	2	2
<u>Clay-Sand-Calcite Matrix Material</u>	4	3
Total Number of Particles Examined	120	100%

990'-1000'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestones</u> : Black; foliated; very fine-grained; sometimes contains phlogopite flakes; contains black, carbonaceous? material.	27	13
<u>Phyllites</u>	25	12
<u>Diorite</u>	1	0.5
<u>Metasandstones</u>	91	45
<u>Siltstone</u>	8	4
<u>Mudstone</u> : Occasionally contains large flakes of phlogopite.	8	4
<u>Marble</u>	27	13
<u>Chert</u>	12	6
<u>Vein Quartz</u>	2	1
<u>Schist</u>	2	1
<u>Clay-Sand-Lithic Matrix Material</u>	1	0.5
Total Number of Particles Examined	204	100%

1090'-1100'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u>	13	8
<u>Phyllite</u>	20	12
<u>Diorite</u>	1	1
<u>Metasandstone</u>	87	51
<u>Siltstone</u> : Some are chloritic.	8	5
<u>Mudstone</u> : Composed of very fine-grained clay-sized material.	9	5
<u>Volcanics</u> : Chlorite-carbonate replaced.	1	1
<u>Marble</u>	13	8
<u>Chert</u>	12	7
<u>Vein Quartz</u>	5	3
<u>Clay-Sand-Lithic Matrix Material</u>	3	2
Total Number of Particles Examined	172	100%

1190' - 1200'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u>	25	15
<u>Phyllite</u>	17	10
<u>Diorite</u>	2	1
<u>Metasandstone</u>	81	50
<u>Siltstone</u>	7	4
<u>Mudstone</u>	11	7
<u>Volcanics</u> : Andesite; plagioclase replaced by montmorillonite and calcite.	1	0.5
<u>Marble</u>	8	5
<u>Chert</u> : Recrystallized.	9	6
<u>Vein Quartz</u>	1	0.5
<u>Quartz Schist</u>	1	0.5
Total Number of Particles Examined	163	100%

1290' - 1300'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestones</u>	13	8
<u>Phyllites</u>	32	20
<u>Metasandstones</u>	65	41
<u>Siltstones</u>	9	6
<u>Mudstones</u>	7	4
<u>Volcanics</u> : Andesites, clay-chlorite and clay-hematite replaced.	4	3
<u>Marble</u>	12	8
<u>Chert</u>	9	6
<u>Vein Quartz</u>	7	4
<u>Quartz-Feldspar Schist</u>	1	0.6
<u>Sericite Schist</u>	1	0.6
Total Number of Particles Examined	160	100%

1390'-1400'

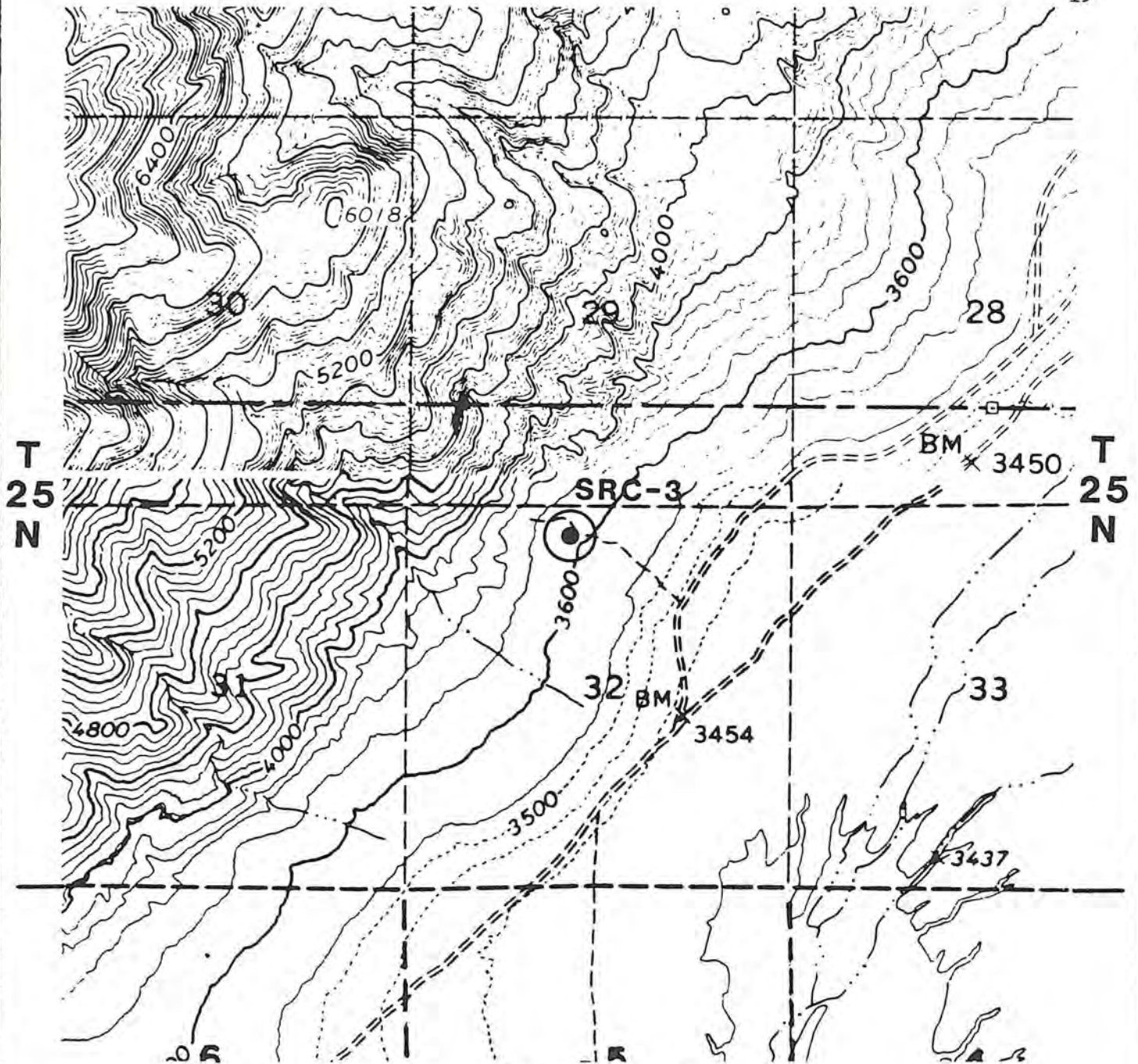
Rock Type	Number of Particles	Percentage of Sample
<u>Limestones</u>	13	7
<u>Phyllites</u>	38	19
<u>Diorite:</u> Primary pyroxene preserved in one chip; in most, montmorillonite replaced mafic minerals.	5	3
<u>Metasandstones</u>	96	49
<u>Siltstone</u>	17	9
<u>Mudstone</u>	14	7
<u>Volcanics:</u> Andesites; one replaced by hematite-clay; another by montmorillonite-illite.	2	1
<u>Marble</u>	7	4
<u>Chert</u>	4	2
Total Number of Particles Examined	196	100%

1490' - 1500'

Rock Type	Number of Particles	Percentage of Sample
<u>Limestone</u>	14	6
<u>Phyllite</u>	35	16
<u>Diorite</u> : Mafic minerals replaced by chlorite [±] clays [±] carbonate.	8	4
<u>Metasandstone</u>	88	41
<u>Siltstone</u>	18	8
<u>Mudstone</u>	11	5
<u>Volcanics</u> : Some are partially sericitized.	2	1
<u>Marble</u>	27	13
<u>Chert</u>	11	5
<u>Vein Quartz</u> : In phyllite	1	0.5
<u>Chlorite Schist</u>	1	0.5
Total Number of Particles Examined	216	100%

R - 37 - E

19



R - 37 - E



Southland Royalty Company
 NATURAL RESOURCES DISTRICT
 FORT WORTH, TEXAS

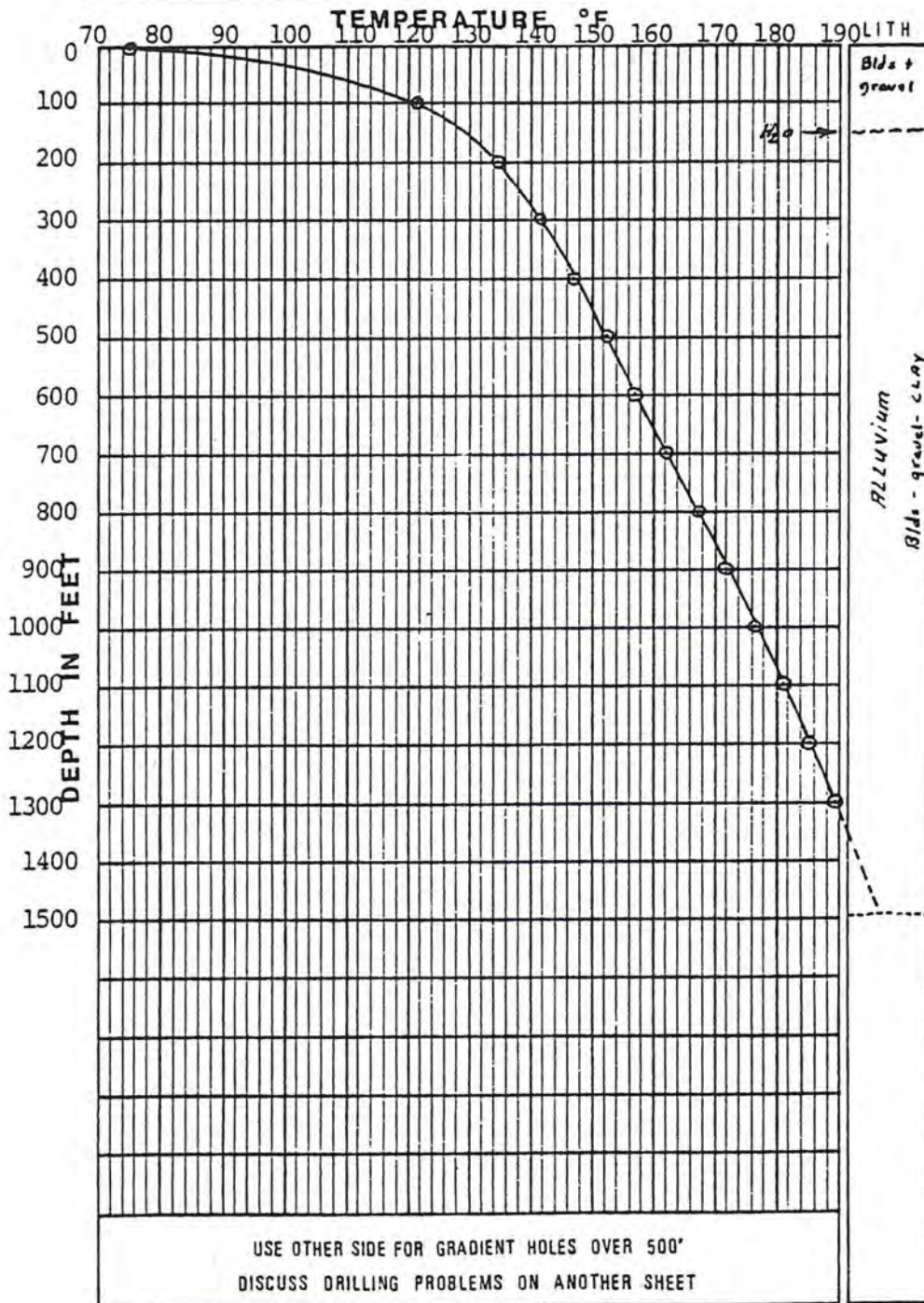
FIGURE 1
 DIXIE VALLEY, NEVADA
 LOCATION AND INDEX MAP
 SRC-3

Scale: 1" = 2000'

SHALLOW TEMPERATURE GRADIENT HOLE RECORD



<u>SR-3</u> GRADIENT HOLE NO.	<u>Dixie Valley</u> PROSPECT	Southland Royalty Company	
<u>NW NE</u> LOCATION	<u>S 32 T 25N R 37E</u> SEC. TOWNSHIP RANGE	<u>Churchill</u> COUNTY	<u>Nevada</u> STATE
<u>01/27/80</u> SPUD DATE	<u>02/14/80</u> COMPLETION DATE	<u>1474</u> AT _____ FT.	<u>4.38°</u> F/100'
<u>1500</u> TOTAL DEPTH	<u>Richard Jodry</u> GEOLOGIST OR PERSON READING TEMP. Reading for 05/29/80	GRADIENT <u>500 TO 1474</u> <u>J. D. Christiansen Drilling,</u> COMPANY DRILLING HOLES <u>Ely</u>	



FOR HOLES 0-500'				
TEMP	°C	°F	°F	°F
DATE OF SURVEY	5-29-80	5-29-80		
TIME SINCE COMPLETION	105 DAYS	105 DAYS		
0	24.0	75.2		
100	49.7	121.2		
200	57.0	134.6		
300	61.2	149.6		
400	64.1	149.4		
500	66.9	152.4		
600	69.5	157.4		
700	72.3	162.1		
800	75.2	167.4		
900	77.8	182.0		
1000	80.3	176.8		
1100	82.8	181.0		
1200	85.2	185.4		
1300	87.5	189.4		
1400	89.8	187.5	193.6	
1474	91.2	196.2		
LOST CIRCULATION ZONES:				
WATER ENTRIES: 150'-160'				
Gravel bed.				

Prep. by *asw.*

SUMMARY OF LITHOLOGIES AND ALTERATION IN SR-4
DIXIE VALLEY, CHURCHILL COUNTY, NEVADA

M. J. Sweeney, September 1980

INTRODUCTION

Samples of washed, rotary cuttings taken every ten feet from drill-hole SR-4 were examined under a stereomicroscope. A thin-section of the cuttings was described from every 100 feet throughout the 1500 foot length of SR-4. Detailed descriptions of the cuttings and thin-sections are attached to this report. Photomicrographs of the thin-sections are also attached. X-ray diffraction patterns of two tuff samples are enclosed.

LITHOLOGIES

SR-4 intersects 900 feet of rhyolitic, vitric-crystal-lithic tuffs and 600 feet of carbonate rocks, limestones and dolomites, which are interbedded with narrow intervals of calcareous siltstones and sandstones. The tuffs overly the carbonate rocks.

Rhyolitic tuffs: The tuffs were originally composed largely of volcanic glass which was deposited as vitric shards, dust and pumice fragments. Shards and dust occurring in about subequal amounts usually comprise 75 to 85% of the tuffs and pumice fragments 5 to 15%. Devitrification of the volcanic glass is variably developed throughout this volcanic section.

Crystal clasts of sanidine, plagioclase and biotite occur throughout the tuff section. The abundances of the crystal clasts are variable; biotite is very rare in some intervals. Sanidine clasts usually comprise 5 to 7% of the tuff, plagioclases 1 to 5% and biotites 0 to 3%. Quartz crystal clasts were seen only in the samples at the bottom of the volcanic section. The feldspar crystal clasts usually range between 0.5 and 2 mm in length. They were originally subhedral to euhedral in shape, but most have been broken prior to deposition. The biotites, 0.2 to 1 mm across, exhibit very dark brown pleochroism, suggesting that they are very Fe-rich. Magnetite is locally exsolved from biotite.

Lithic clasts comprise 5 to 20% of the tuff; their abundance is variable throughout this section. Clasts of andesites, basalts, latites, quartzite and carbonate were seen; the first three types are the most ubiquitous. The lithic clasts observed are less than 10 mm in diameter.

The entire 900' section of this volcanic unit is very similar. Primary mineralogy and textures were virtually identical. One of the major variations seen in the tuff now is hardness. Most of the tuff is relatively soft and also gritty to the touch. However, between the depths of 520' and 640' the tuff is hard, exhibits a vitreous luster and fractures conchoidally. Examination of the thin-section at 590'-600' shows that the vitric components have been well compacted and welded. Devitrification to Kspar and quartz is complete. This well welded interval may be the base of a thick ignimbrite unit. Other intervals in the tuff have also been well devitrified, but no other intervals appear to have been as intensely compacted as the interval between 520 and 640 feet.

Dolomites, Limestones, Sandstones, Siltstones: Beneath the tuffs, SR-4 intersected dolomites, limestones (both of which are locally cherty), and calcareous sandstones and siltstones.

Cherty dolomites occur in the first 150 feet below the tuffs. Limestones occur in the next 60 feet, calcareous siltstones and sandstones in the next 110 feet, and then silty, locally cherty or argillaceous, limestones occur in the lowermost 280 feet of the drill hole.

The dolomites and limestones are dominantly fine-grained; they are micritic. Fossils were seen rarely.

The siltstones and sandstones are well sorted; the calcite in them is of detrital origin.

ALTERATION

Rhyolitic tuffs: Hydrothermal alteration in the tuffs is seen mainly in the glass and in the mafic minerals in the lithic clasts. Only rarely have crystal clasts suffered any alteration.

The volcanic glass has been replaced most commonly by clays and lesser calcite. Local occurrences of jarosite (90'-100'), gypsum (40'-50', 90'-100') and manganese-oxides (10'-20', 180'-240') were observed. Trace amounts of hematite and goethite occur throughout the tuff. These limonites replace disseminated and vein magnetite and sulfides; it was not often possible to determine which of these was the precursor of the limonites. Trace amounts of zeolites may occur in vugs in the upper portion of the tuff.

The extent and type of clay replacement is highly variable. Replacement may range between 5 and 100%. The most intensely clay-altered intervals are easily identified by observing the degree of rock disaggregation during washing; easily disaggregated rocks are intensely clay-replaced. The clay types present appear to be variable. X-ray diffraction analysis of clays in the sample at 590'-600' gives a strong 12.6Å⁰ peak possibly indicative of the presence of a mixed layer clay or possibly of hydrobasaluminite. X-ray analysis of sample 930'-940' indicates that the abundant white clay is montmorillonite; this rock disaggregates completely on wetting.

The present color of the tuffs is controlled by the type and distribution of limonites and/or clays present. Red, pink and orange colored tuffs are the result of limonite staining. Bright green tuffs contain a bright kelly-green clay mineral, probably celadonite.

In the interval between 860' and 900', the tuff has been brecciated. The spaces among the fragments have been filled with microcrystalline quartz, celadonite (?) and calcite.

Only rarely was fresh pyrite seen in the tuffs (at 520' to 530').

Slickensided chips derived from fault zones are sporadically present throughout the tuff; the gouge chips are usually white and weakly stained with hematite.

Dolomites, Limestones, Sandstones, Siltstones: Evidence of hydrothermal alteration in the carbonates and sandstones is largely confined to veins. Thin calcite veins, less than 1-2 mm wide, with and without pyrite occur throughout this interval. Pyrite also is disseminated in these rocks; it usually occurs as very fine-grained crystals forming less than 0.2% of the rock. This disseminated pyrite may be syngenetic in origin. Traces of magnetite also occur as disseminations. Total pyrite content throughout the carbonate-siltstone interval never exceeds 0.3 volume % and is usually less.

Evidence of faulting occurs throughout the carbonate-siltstone interval. Slickensided chips are present in nearly every sample of every rock type. Weak hematite stains are present on these slickensided surfaces.

THIN-SECTION DESCRIPTIONS
OF SR-4 SAMPLES



SR-4 90'-100' Devitrified Vitric-Crystal Tuff

This tuff contains sanidine crystal clasts, 0.3 to 1 mm long, most of which are angular, broken fragments; they form 5-7% of rock. These sanidines are totally unaltered. A few phenocrysts of albite-twinned plagioclase are present, 1-2%; they are unaltered. The groundmass is composed of devitrified glass shards; the shard outlines are preserved by axiolitic devitrification structures. Spherulitic devitrification structures occur occasionally. Traces of zeolite minerals may be present in some vugs.

Jarosite occurs in this tuff; it pseudomorphically replaces individual shards and occurs in irregular blebs disseminated throughout rock. Jarosite chips as large as 0.5 mm across are present. There is no evidence as to the origin of the jarosite; it does not occur in pyrite-like pseudomorphs. Jarosite also lines vugs.

Traces of hematite occur as disseminations (after magnetite). A few, free chips of gypsum occur in the section.

PRIMARY MINERALS:

Quartz	10-30%	Anhedral to subhedral; occurs only in groundmass. Grain size 0.01 mm or less.
Sanidine	5-7%	Crystal clasts: 0.2 to 1 mm long; unaltered.
	50-70%	Groundmass: anhedral to subhedral; grain size 0.01 mm or less.
Plagioclase	1-2%	Crystal clasts: 0.2 to 1 mm long; unaltered; albite-twinned.

SECONDARY MINERALS:

Jarosite	3-5%	Pseudomorphically replaces vitric shards and biotites?; also in irregular blebs disseminated throughout tuff. Very fine-grained crystals, 0.005 mm or less.
Gypsum	tr	Free crystals in sample.
Zeolites??	1-5%	In vugs; minerals with birefringence up to first-order red occur in vugs. Colorless; lath shaped.
Hematite	tr	Disseminated.

SR-4 190'-200' Devitrified Vitric-Crystal Tuff

This tuff originally was composed largely of vitric shards; it also contains crystal clasts of: sanidine, 0.3-0.6 mm, 5-7%; plagioclase, 0.3-1 mm long, albite-twinned (An_{30-40}), 2-3%; and of biotite, 0.3-1 mm, very dark brown pleochroism, 1-3%. The phenocrysts (crystal clasts) are unaltered. The glass shards have devitrified to quartz, Kspar and clays. Irregular aggregates of Mn-oxides are distributed throughout rock; they form 2-3% of rock.

PRIMARY MINERALS:

Quartz	10-35%	Anhedral to subhedral; occurs in groundmass intergrown with Kspar.
Kspar	5-7%	Sanidine crystal clasts: angular, crystal fragments.
	50-70%	Groundmass: anhedral to subhedral; grain size less than 0.01 mm.

Plagioclase	2-3%	Crystal clasts: unaltered.
Biotite	1-3%	Crystal clasts: very dark brown pleochroism.

SECONDARY MINERALS:

Clay	5-20%	Very fine-grained; occurs in matrix.
Mn oxides	2-3%	Black reflecting; brown streak, disseminated throughout tuff in clots up to 0.1 mm across.

SR-4 290'-300' Vitric-Crystal Tuff

This tuff is composed of: glass shards and even finer-grained, glass dust both of which form 80% of rock; glassy pumice fragments, 0.5 to 4 mm, 10% of rock; crystal clasts of sanidine, 0.1-0.5 mm, 5-7% of rock; crystal clasts of plagioclase, 0.1-0.5 mm, 1-2%; and crystal clasts of very dark brown biotite, 0.1-1.5 mm across, 2-4%. A very few, tiny lithic fragments of very fine-grained basalt are also present. Magnetite is disseminated throughout rock; magnetite replaces some of the biotite phenocrysts.

The glass particles have been less than 50% devitrified. Devitrification of pumice fragments is more advanced than that of vitric shards and ash. Spherulitic devitrification structures occur most commonly in pumice fragments. Devitrification products include Kspar, quartz and clay.

Calcite occurs in narrow veins (less than 0.1 mm wide); it also replaces feldspar phenocrysts, pumice fragments and vitric shards. Traces of celadonite (bright green) are disseminated throughout rock.

PRIMARY MINERALS:

Quartz	10-20%	Groundmass: devitrification product of glass. Very fine-grained.
Kspar	5-7%	Crystal clasts.
	10-20%	Groundmass: devitrification product of glass. Very fine-grained.
Plagioclase	1-2%	Crystal clasts: 0.2-1 mm long; 10-90% (average 15% replaced by calcite).
Biotite	2-4%	Crystal clasts: very dark brown; 0.1-1 mm across; locally, partially replaced by magnetite.
Magnetite	0.5-1%	Disseminated; also common in biotite sites.
Glass	30-40%	Vitric shards, ash, pumice fragments.

SECONDARY MINERALS:

Celadonite	tr-1%	Bright green; disseminated throughout groundmass.
Calcite	3%	Vein and disseminated.
Clay	5-20%	Alteration product of vitric clasts.

SR-4 390'-400' Vitric-Lithic-Crystal Tuff

This tuff is composed mostly of glass which includes shards, pumice fragments and very fine-grained vitric dust; glass forms 65 to 75% of the tuff. The fine dust has been altered to illite/sericite. Most of the shards and pumice fragments are unaltered. Crystal clasts of sanidine are present; they are

completely unaltered and form 7-10% of rock. Crystal clasts of unaltered plagioclase comprise 1-2% of rock. Crystal clasts of biotite constitute 2-4% of the tuff; the biotite exhibits dark brown pleochroism and is not altered.

Lithic fragments of andesite, basalt(?), quartzite(?), latite and carbonate occur; they are listed in order of abundance. The lithic fragments form about 10-20% of the tuff; most of these fragments are less than 5 mm across. The plagioclases, both phenocrysts and groundmass laths, are unaltered in these fragments. Original mafic minerals, excepting biotite phenocrysts, have been replaced by Fe-oxides and clays (montmorillonite?); biotite phenocrysts are unaltered. Primary magnetite is disseminated throughout the lithic fragments; it forms 0.5-1% of the fragments.

In one latite fragment, the feldspar phenocrysts have been replaced by epidote.

SECONDARY MINERALS:

Illite/sericite	10-30%	Alteration product of glass dust.
Montmorillonite?	5-7%	Alteration product of pyroxenes/hornblendes.
Fe-oxides (goethite/ hematite)	1%	Alteration product of pyroxenes/hornblendes.

SR-4 490'-500' Clay-altered Vitric-Crystal-Lithic Tuff

The tuff contains crystal clasts of sanidine which occur as either broken crystal fragments or euhedral, Carlsbad-twinned crystals up to 3 mm in length; they form 5 to 7% of rock. These sanidines are completely unaltered. Biotite crystals are rare, forming less than 0.5% of the rock. Plagioclase crystal clasts, up to 4 mm in length, form less than 1% of the tuff; they are completely unaltered. Lithic fragments of other tuffs and andesites are usually less than 2 mm across; they form about 3% of the tuff.

The groundmass or matrix of this tuff was originally composed of vitric shards and dust. The original glass is now composed mostly of fine-grained quartz and Kspar; locally devitrified shards exhibit axiolitic structures. Original vitric dust is locally replaced by fine-grained clays, celadonite?(green) and illite(colorless). Celadonite also occurs in aggregates filling gas bubbles and possibly replacing primary biotites. Minor amounts of carbonate are irregularly distributed throughout groundmass.

SECONDARY MINERALS:

Illite(?)	10-30%	Replaces glass in matrix.
Celadonite	5%	Occurs in groundmass; also fills vugs and may replace biotite phenocrysts.
Carbonate	1-3%	Occurs in irregular patches throughout groundmass.

SR-4 590'-600' Devitrified Welded Vitric-Crystal-Lithic Tuff

This tuff contains crystal clasts of: sanidine, 0.2-2mm long, 3 to 5% of rock; plagioclase, 0.2 to 1 mm long, 1-2% of rock; and of biotite, 0.1-0.5 mm across, 0.5-1% of rock, exhibits very dark brown pleochroism. None of these crystal clasts have suffered any alteration.

The groundmass was once composed of flattened pumic fragments, vitric shards

and dust. The glass has been devitrified to fine-grained, anhedral quartz, feldspar and very fine-grained disseminated magnetite. Clay also (celadonite? =green; montmorillonite?=orange) occurs in the groundmass; clay spottily replaces 10 to 40% (average 20%) of groundmass.

SECONDARY MINERALS:

Celadonite	7-10%	Alteration product of glass; bright kelly green.
Montmorillonite?	7-10%	Alteration product of glass; dirty orange.
Pyrite	0.1%	Vein and disseminated. Occurs in a vein in an intensely clay-replaced chip.

SR-4 690'-700' Vitric-Lithic-Crystal Tuff

This tuff is composed dominantly of vitric shards and dust. The dust has been partially devitrified to clay (sericite) and to very fine-grained orthoclase and quartz. It also contains crystal clasts of: sanidine, up to 2 mm long, 3 to 5% of rock; plagioclase, up to 2 mm long--these plagioclase crystals are strongly zoned, they form 5 to 7% of rock; and of biotites which exhibit very dark brown pleochroism--they form 1-2% of rock. None of the crystal clasts has suffered any alteration.

Lithic clasts, fine-grained andesites/basalts?, are usually less than 2 mm in diameter. Such clasts form 5-10% of the tuff. The feldspars are unaltered; mafic minerals have been replaced by clays (montmorillonite?, illite?) and Fe-oxides (goethite, hematite).

SECONDARY MINERALS

Sericite/clay	5-20%	Replaces glass dust; also replaces mafic minerals in lithic clasts.
Goethite/hematite	0.5-1%	Alteration product of mafic minerals; some primary magnetite is partially oxidized.
Carbonate	2-5%	Locally floods rock--replaces all components.

SR-4 790'-800' Devitrified Welded Vitric-Crystal Tuff

This rock was formed as a welded vitric-crystal tuff. The glass has been completely devitrified. Spherulites, usually about 0.5 mm across, are common devitrification structures. Axiolitic structures developed in shards are also common.

Crystal clasts of sanidine form 3-5% of rock; they are unaltered. Clasts of plagioclase(?) formed 1-3% of rock; they have been replaced by celadonite and illite. Biotites have also been replaced by celadonite. Primary magnetite is disseminated in trace amounts. A weak stain of hematite/goethite is disseminated throughout many chips.

Lithic clasts of latite/andesite form less than 50% of sample.

This tuff has been subjected to cataclasis or brecciation prior to devitrification. The breccia fragments are 0.1 to 2 mm across. The spaces among the fragments have been filled with microcrystalline quartz, celadonite and calcite.

SECONDARY MINERALS:

Sericite	2-5%	Replaces plagioclase clasts.
Celadonite	3-5%	Replaces mafic minerals; fills vugs; occurs with chert in interbreccia spaces.
Quartz	1-10%	Fills spaces among breccia clasts; microcrystalline or cherty.
Calcite	1-2%	Occurs in interbreccia spaces; occurs in post-brecciation fractures.
Hematite/goethite	1-2%	Stains rock.

SR-4 890'-900' Devitrified Crystal-Vitric Tuff

This is the first tuff sample from this drill hole to contain quartz crystal clasts. Several of those present are rounded and embayed by resorption. The quartz clasts are usually about 1 mm in diameter; they comprise about 5% of the tuff. Crystal clasts of sanidine form about 10% of rock; those of plagioclase about 1%. Biotite phenocrysts occur in trace amounts. None of these phenocrysts have suffered any alteration.

The groundmass once composed of vitric shards and dust is now composed mostly of anhedral quartz and Kspar; spherulites occur occasionally. Clay alteration (sericite/illite) of groundmass is weak to moderate. Trace amounts of magnetite are disseminated throughout groundmass.

Calcite and rarely microcrystalline quartz occur as fracture fillings.

Lithic fragments of andesites/basalts form less than 5% of this rock.

SECONDARY MINERALS:

Sericite/clay	10-25%	Very fine-grained material occurs in groundmass.
Calcite	3-7%	Vein and disseminated.
Quartz	1-3%	Microcrystalline; occurs in fracture fillings with calcite.

SR-4 990'-1000' Carbonates (limestones/dolomites), Cherts and Devitrified Welded Tuffs

This thin-section contains chips of fine- to medium-grained, recrystallized carbonates; these chips form 50-60% of the section. A few of these chips contain disseminated chert. Pyrite fills fractures in a few of the carbonate chips.

The next most abundant rock type is chert; most of these chips are carbonate-free. Chert comprises 30-40% of rock chips in this thin-section.

Quartz "eye" tuff constitutes 5-10% of the chips. The matrix and plagioclase phenocrysts are weakly clay (illite)-altered.

SECONDARY MINERALS:

Pyrite	tr	Fills or coats fractures in carbonate rocks.
Illite	2-5%	Alteration product of tuff.

SR-4 1090'-1100' Limestones, Sandy Limestones and Cherty Limestones

This sample contains pure limestones, sandy limestones and cherty limestones.

About 30% of the chips are composed of pure carbonate, mostly calcite judging by how reactive the rock is to cold, dilute HCL. The calcite is mostly very fine-grained (micritic). A few chips contain relict fossil structures (bryozoans?, shell fish). A small proportion of these chips have been recrystallized to medium-grained marbles.

About 50% of the limestone chips contain 10-15% silt and fine-sand sized quartz. Disseminated pyrite occurs most commonly in these chips; pyrite occurs in trace amounts up to 0.5 volume %.

About 15% of the limestone chips contain chert which occurs as microcrystalline quartz or fibrous chalcedony. Chert occurs in limestones with and without sand. Pyrite also occurs in cherty limestones.

Thin calcite veins occur in all of the above described chips.

SECONDARY MINERALS:

Pyrite	0.5%	Vein and disseminated.
Calcite	2-3%	Vein.

SR-4 1190'-1200' Calcareous Siltstone

This rock is composed largely of angular, detrital quartz grains which average 0.04 mm in size. Quartz comprises about 50% of the rock. Silt-sized orthoclase grains form about 5% of the rock. The remainder of the rock is composed of calcite. This calcite is also fine-grained; it is disseminated throughout the sandstone. The calcite is usually orange or brown in color, presumably from exsolved iron. Locally illite is common as a matrix for quartz grains; it may form 5-15% of rock.

Thin calcite and calcite-quartz veins crosscut the sandstone. Vein calcite is colorless. Euhedral cubes of hematite-replaced pyrite (or magnetite) up to 0.1 mm across are disseminated through the sandstone.

SECONDARY MINERALS:

Pyrite	tr	
Calcite	1-2%	In veins by itself and with quartz.
Quartz	1-2%	In veins with calcite.

SR-4 1290'-1300' Silty Limestones, Argillaceous Calcareous Siltstones and Limestones

Silty limestone is the most common rock type in this sample. Angular, silt-sized, quartz grains comprise 5 to 30% of these limestones. Most of the calcite in these chips is colorless in thin-section; however, orange or brown-stained calcite constitutes up to 30% of some rock chips.

The argillaceous rocks contain 10 to 40% clays (illite? and kaolinite?). Silt and fine-grained calcite, present in subequal amounts, are the other constituents of the clayey rocks; such chips comprise about 25% of this sample.

A few pure calcite chips are present. The calcite in these chips is mostly very fine-grained (micritic) and colorless.

Traces of pyrite, both fresh and oxidized, occur in silty limestones and argillaceous siltstones. Pyrite occurs in veins and as disseminations.

SECONDARY MINERALS:

Pyrite/hematite 0.1-0.2% Vein and disseminated.

SR-4 1390'-1400' Silty Limestones

These rocks are composed dominantly of very fine-grained calcite. Angular, detrital, silt-sized grains of quartz and feldspar form 5 to 20% of these limestones. Clay (illite, kaolinite) can comprise near 15% of the limestone; usually clay are present in small amounts.

Trace amounts of magnetite are disseminated throughout the limestones. Pyrite occurs in trace amounts in calcite veins.

SECONDARY MINERALS:

Calcite	1-2%	Occurs in veins.
Pyrite	tr	Occurs in veins with calcite.

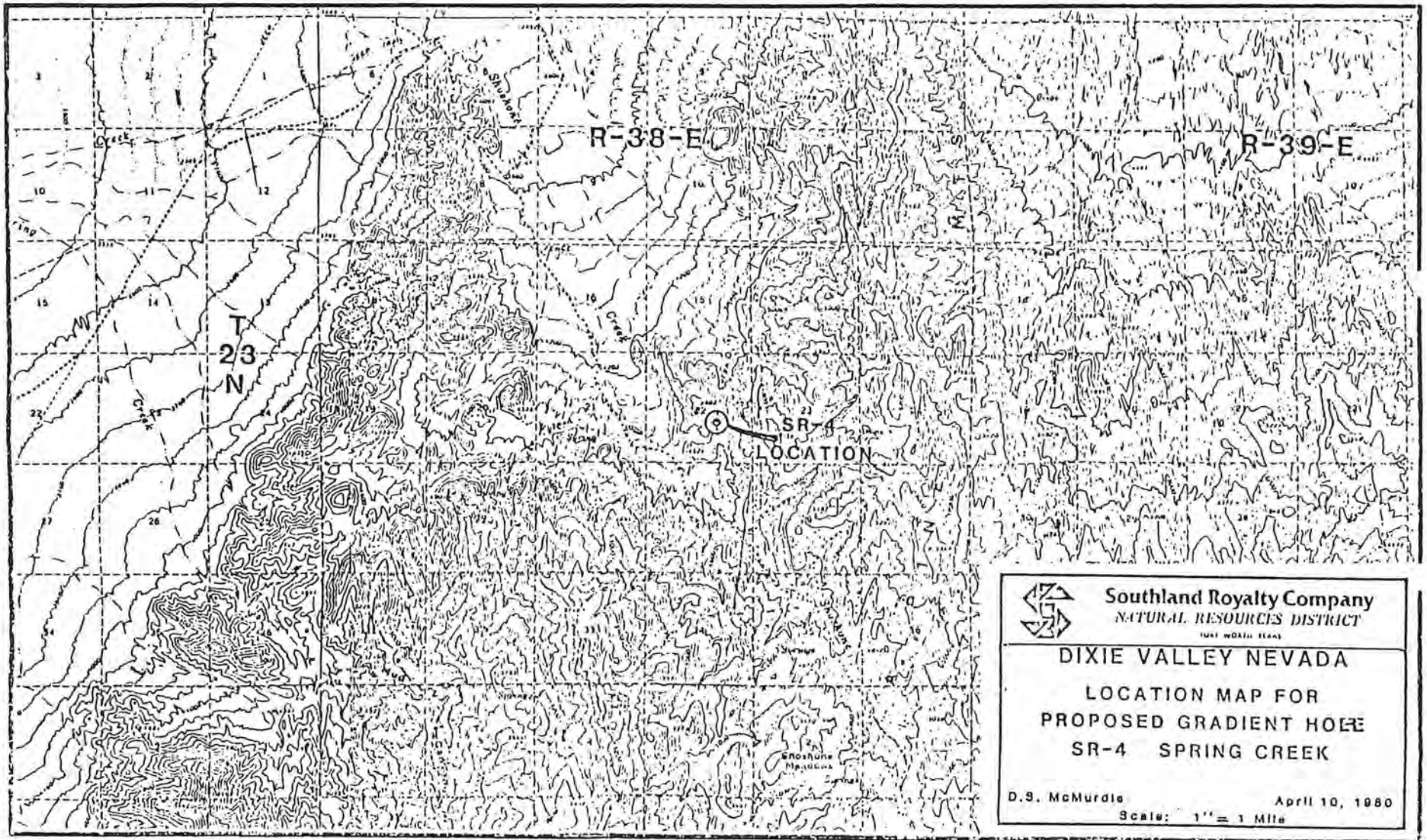
SR-4 1490'-1500' Limestones

The rocks in this sample are micritic limestones which contain 2 to 20% silt. Clay comprises 5 to 25% (average 10% or less) of these limestones.

Thin calcite veins, some with goethite (after pyrite?), occur. Traces of very fine-grained pyrite and magnetite are disseminated throughout these limestones; these disseminated minerals are probably syngenetic.

SECONDARY MINERALS:

Calcite	2-3%	Veins, sometimes with goethite (after pyrite?).
Pyrite/goethite	0.2%	Vein and disseminated; the disseminated pyrite may be of syngenetic origin.



Log No.
 Permit No.
 Basin.

WELL DRILLERS REPORT

Please complete this form in its entirety

1. OWNER Southland Royalty Company ADDRESS 1500 First National Bank Building
Fort Worth, Texas 76102

2. LOCATION NW 1/4 SE 1/4 Sec. 22 T. 23 N 1/2 R. 38 E Churchill County
 PERMIT NO. Thermal Gradient Hole No. SR-4

3. TYPE OF WORK
 New Well Recondition
 Deepen Other

4. PROPOSED USE
 Domestic Irrigation Test
 Municipal Industrial Stock

5. TYPE WELL
 Cable Rotary
 Other

6. LITHOLOGIC LOG

Material	Water Strata	From	To	Thick-ness
Top Soil		0	2	2
Rhyolite		2	310	308
Fault Gouge		310	330	20
Granite		330	340	10
Rhyolite		340	390	50
Granite		390	420	30
Rhyolite		420	530	110
Granite		530	650	120
Sandstone		650	660	10
Fault Gouge		660	670	10
Rhyolite		670	930	260
Shale		930	940	10
Rhyolite		940	1150	210
Shale		1150	1160	10
Rhyolite		1160	1180	20
Fault Gouge		1180	1190	10
Shale		1190	1210	20
Rhyolite		1210	1240	30
Shale		1240	1280	40
Granite		1280	1290	10
Shale		1290	1470	180
Rhyolite		1470	1500	30

8. WELL CONSTRUCTION
 Diameter hole 9-7/8 inches Total depth 1,500 feet
 Casing record 7-5/8" 0 - 151 Ft.
 Weight per foot 20 Lbs. Thickness.....
 Tubing 1" 0 - 1,503 Ft.
 Diameter From To
6-1/4 inches 160 feet 1,260 feet
6-1/8 inches 1,260 feet 1,500 feet
 inches feet feet
 inches feet feet
 inches feet feet
 inches feet feet
 Surface seal: Yes No Type Cement
 Depth of seal 1" Tubing Cemented 32 feet
 Gravel packed Yes No
 Gravel packed from 1,500 feet to 1,468 feet
 Perforations: None
 Type perforation.....
 Size perforation.....
 From feet to feet
 From feet to feet
 From feet to feet
 From feet to feet
 From feet to feet

Date started June 8, 19 80
 Date completed June 27, 19 80

7. WELL TEST DATA

Pump RPM	G.P.M.	Draw Down	After Hours Pump

BAILER TEST
 G.P.M. Draw down feet hours
 G.P.M. Draw down feet hours
 G.P.M. Draw down feet hours

9. No Water WATER LEVEL
 Static water level 0 Feet below land surface 0
 Flow G.P.M.
 Water temperature ° F. Quality.....

10. DRILLERS CERTIFICATION
 This well was drilled under my supervision and the report is true to the best of my knowledge.
 Name Jerrold D. Christiansen
 Address 557 Ely Ave., Ely, Nevada 89301
 Nevada contractor's license number 14700
 Nevada driller's license number 641
 Signed Jerrold D. Christiansen
 Date July 5, 1980



R - 35 - E

R - 36 - E

R - 37 - E

T 24 N

T 24 N

T 23 N

T 23 N




T 22 N

T 22 N

R - 35 - E

R - 36 - E

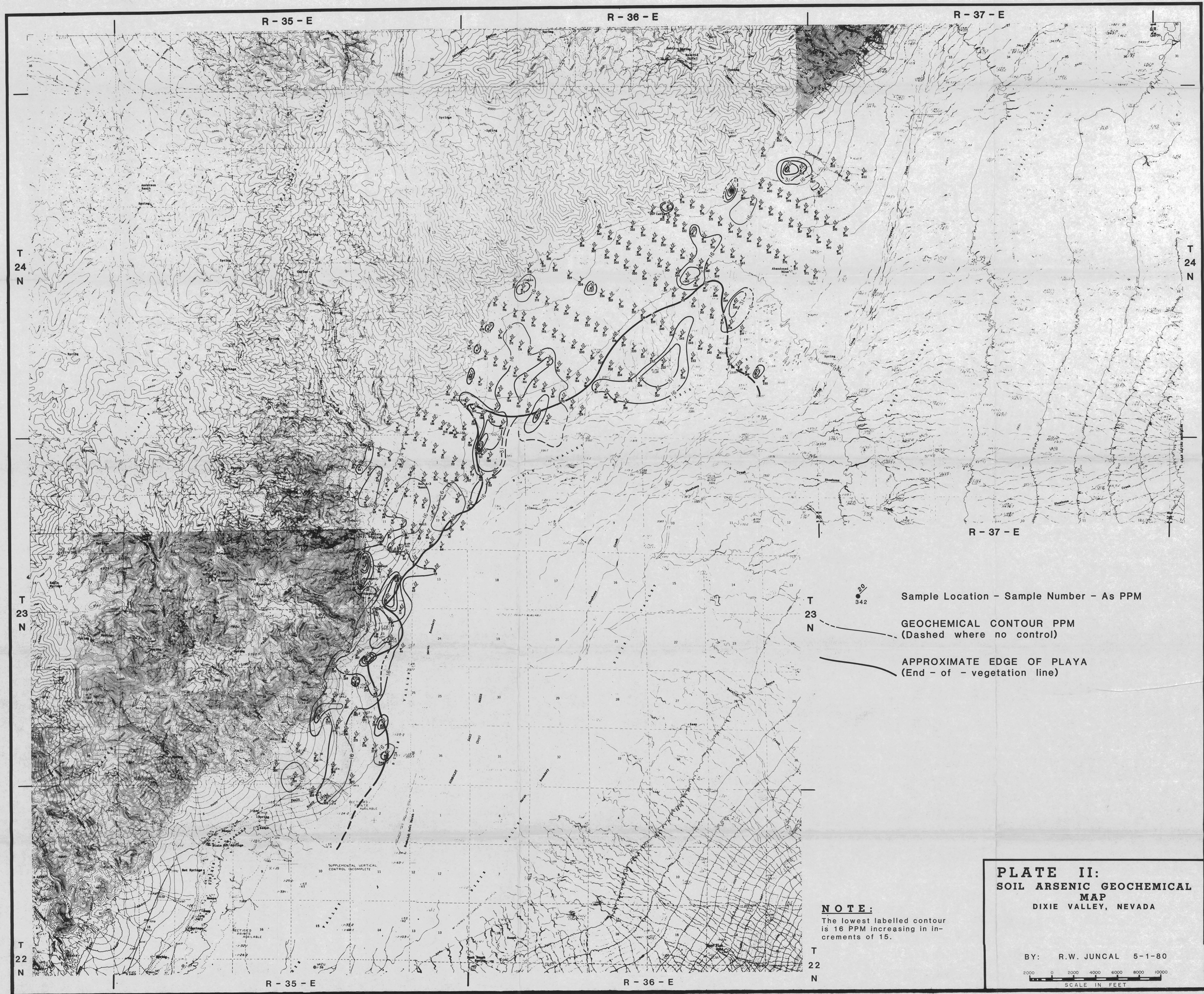
R - 37 - E




-  Sample Location - Sample Number - Hg PPB
-  GEOCHEMICAL CONTOUR PPB
(Dashed where no control)
-  APPROXIMATE EDGE OF PLAYA
(End - of - vegetation line)

NOTE:
The lowest labelled contour is 40 PPB increasing in increments of 80. Where single point anomalies go into the PPM range a maximum contour of 600 is adequate.

PLATE I:
SOIL MERCURY GEOCHEMICAL
MAP
DIXIE VALLEY, NEVADA

BY: R.W. JUNCAL 5-1-80
SCALE IN FEET
0 2000 4000 6000 8000 10000

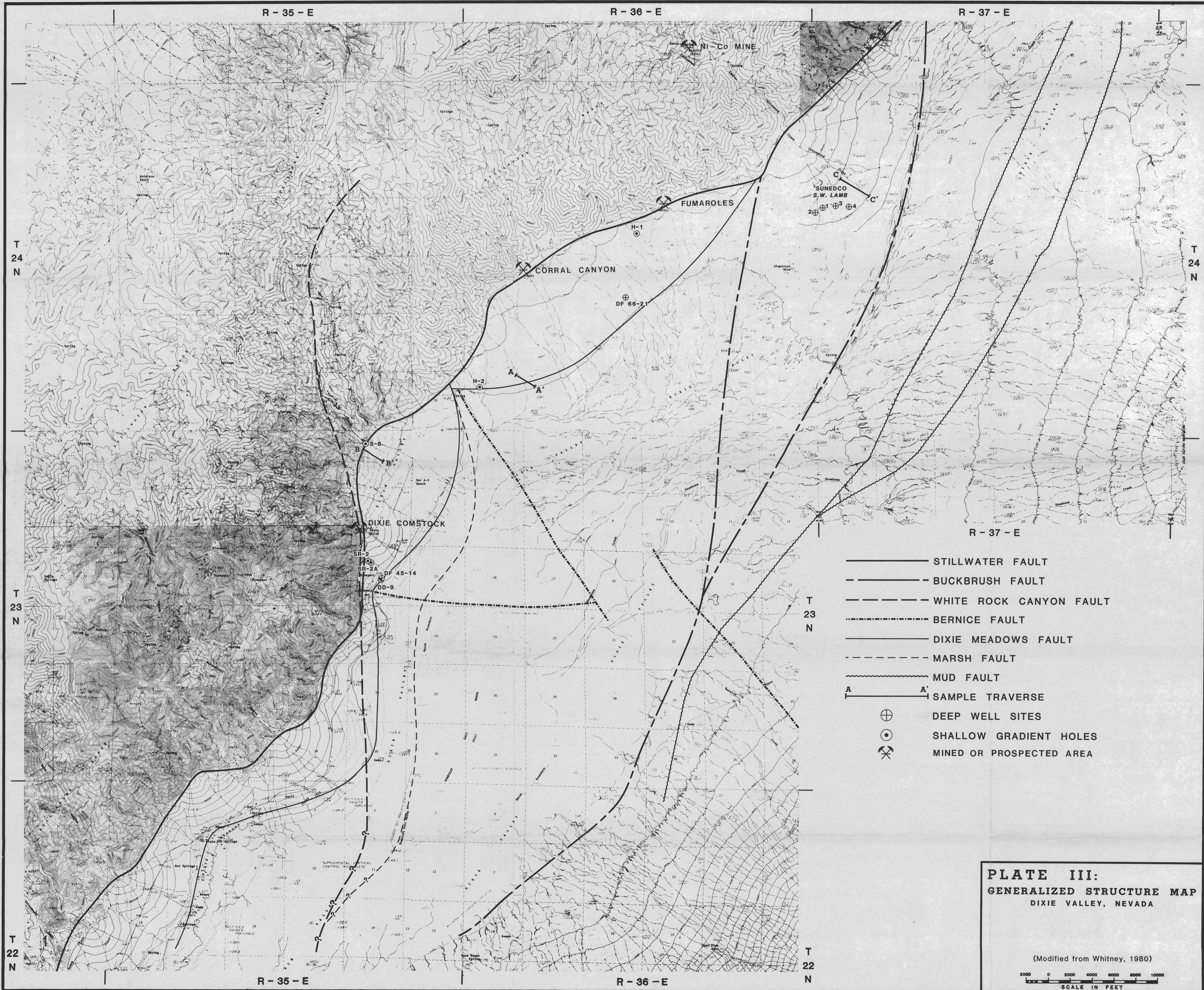


-  Sample Location - Sample Number - As PPM
-  GEOCHEMICAL CONTOUR PPM
(Dashed where no control)
-  APPROXIMATE EDGE OF PLAYA
(End - of - vegetation line)

**PLATE II:
SOIL ARSENIC GEOCHEMICAL
MAP
DIXIE VALLEY, NEVADA**

NOTE:
The lowest labelled contour
is 16 PPM increasing in in-
crements of 15.

BY: R.W. JUNCAL 5-1-80
SCALE IN FEET
2000 0 2000 4000 6000 8000 10000



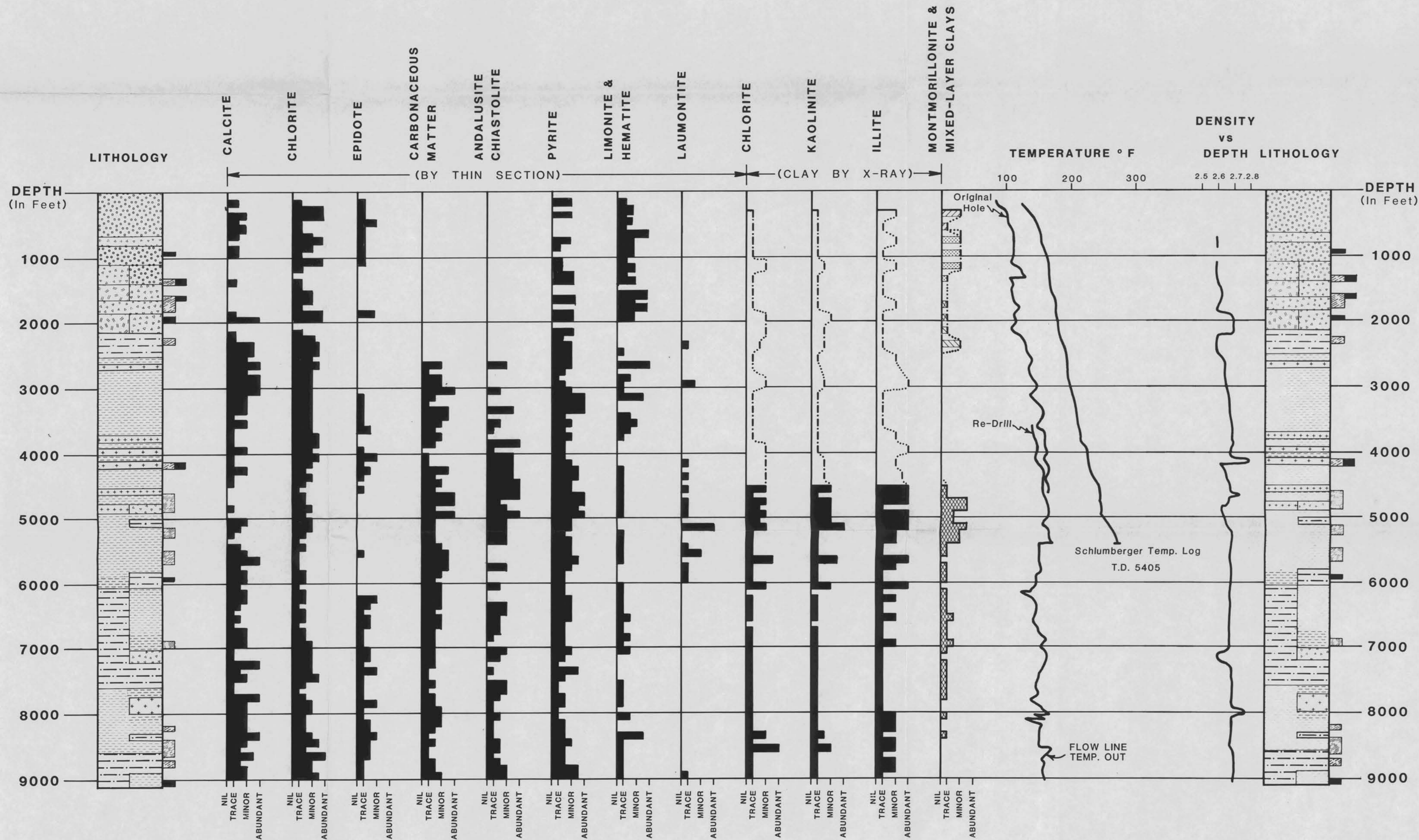
- STILLWATER FAULT
- - - - - BUCKBRUSH FAULT
- - - - - WHITE ROCK CANYON FAULT
- - - - - BERNICE FAULT
- DIXIE MEADOWS FAULT
- - - - - MARSH FAULT
- ~~~~~ MUD FAULT
- A ——— A' SAMPLE TRAVERSE
- ⊕ DEEP WELL SITES
- ⊙ SHALLOW GRADIENT HOLES
- ⚒ MINED OR PROSPECTED AREA

PLATE III:
GENERALIZED STRUCTURE MAP
 DIXIE VALLEY, NEVADA

(Modified from Whitney, 1980)

2000 0 2000 4000 6000 8000 10000
 SCALE IN FEET

PLATE IV: RELATIVE ABUNDANCE OF SELECTED MINERAL SPECIES IN DF 45-14



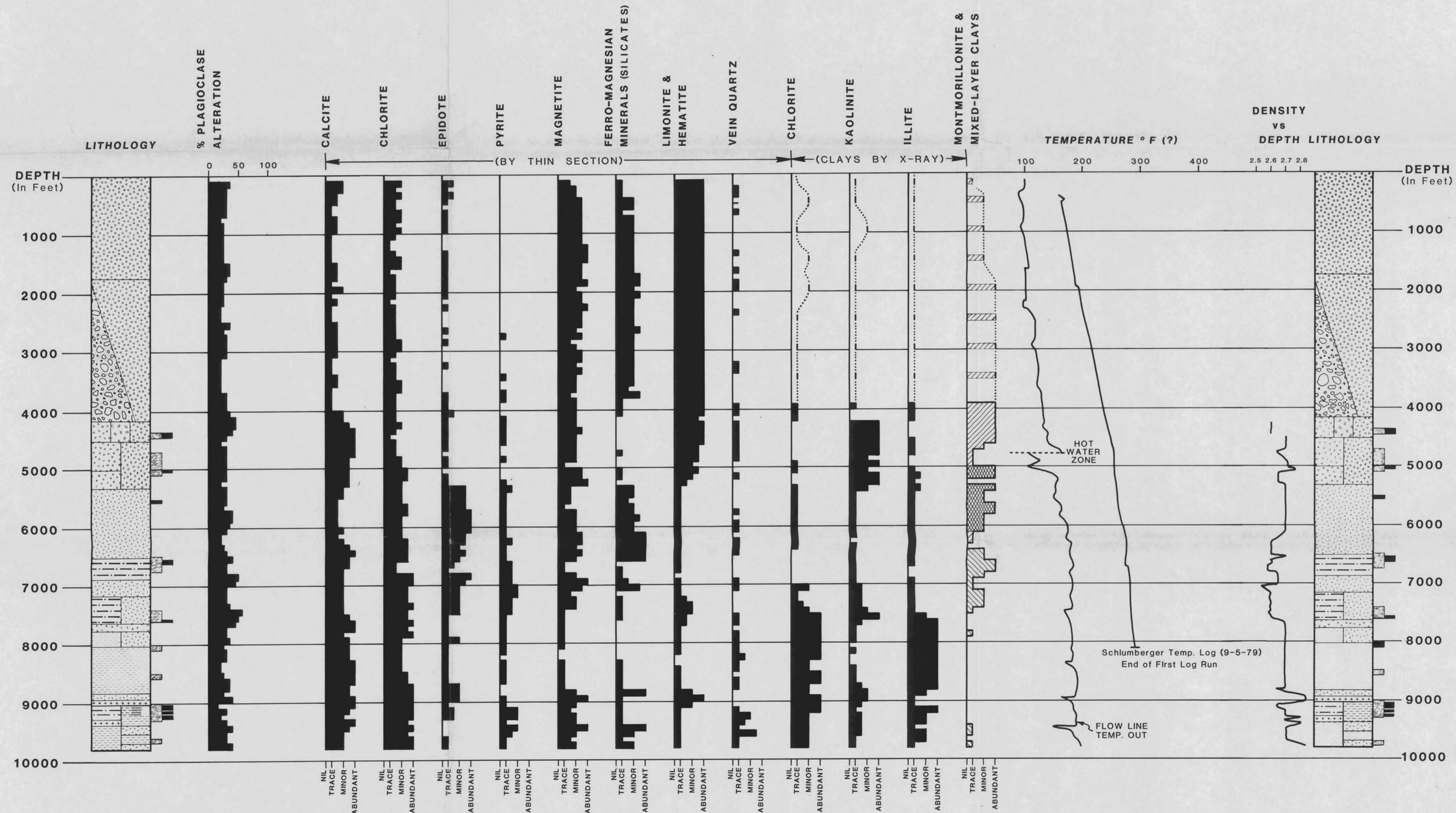
EXPLANATION - MINERAL ANALYSIS

- Randomly Interstratified Illite/Montmorillonite with Montmorillonite < 30%
- Calcium/Magnesium-Montmorillonite
- Sodium-Montmorillonite
- Regular Mixed-Layer Chlorite/Vermiculite
- Inferred

EXPLANATION - LITHOLOGY

- Alluvium
- Tuff
- Andesite/Basalt
- Metasiltstone/Metashale
- Gabbro/Diorite Intrusive
- Metarenite/Quartz Arenite
- Postulated Fracture/Fault Zones
- Mineralized Zone

PLATE V: RELATIVE ABUNDANCE OF SELECTED MINERAL SPECIES IN DF 66-21



EXPLANATION - MINERAL ANALYSIS

- Randomly Interstratified Illite/Montmorillonite with Montmorillonite > 30%
- Randomly Interstratified Illite/Montmorillonite with Montmorillonite < 30%
- Regular Mixed-Layer Chlorite/Vermiculite
- Inferred

EXPLANATION - LITHOLOGY

- Alluvium
- Alluvium and Clay
- Tuff
- Andesite/Basalt
- Granodiorite Intrusive
- Quartz Arenite
- Metasiltstone/Metashale
- Gabbro/Diorite Intrusive
- Mineralized Zones
- Postulated Fracture/Fault Zones

PLATE VI:

DISTRIBUTION OF SELECTED ELEMENTS IN DIXIE FEDERAL 45-14 HEAVY MINERAL FRACTIONS DIXIE VALLEY, CHURCHILL COUNTY, NEVADA

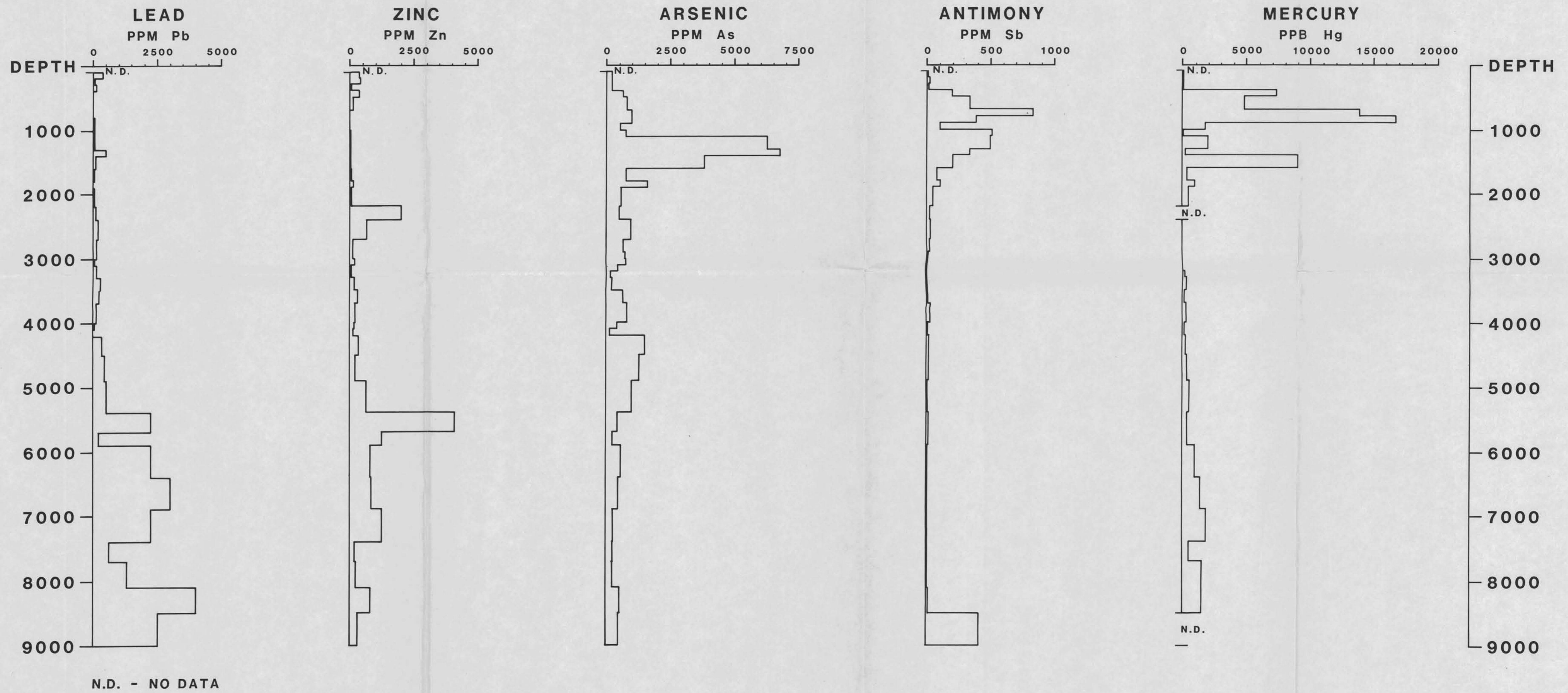


PLATE VII:

DISTRIBUTION OF SELECTED ELEMENTS IN DIXIE FEDERAL 45-14 WHOLE ROCK SAMPLES DIXIE VALLEY, CHURCHILL COUNTY, NEVADA

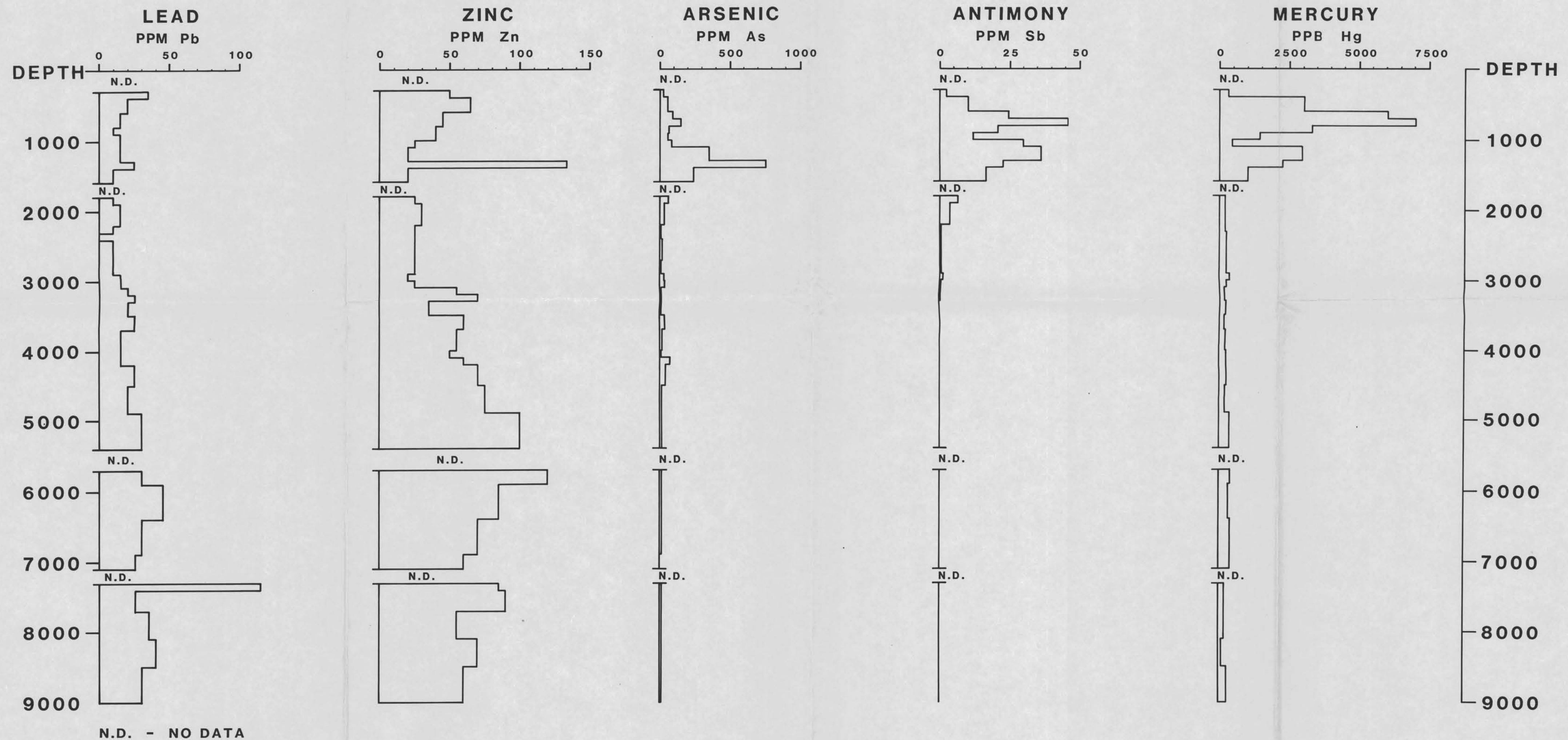


PLATE VIII:

DISTRIBUTION OF SELECTED ELEMENTS IN DIXIE FEDERAL 66-21 HEAVY MINERAL FRACTIONS DIXIE VALLEY, CHURCHILL COUNTY, NEVADA

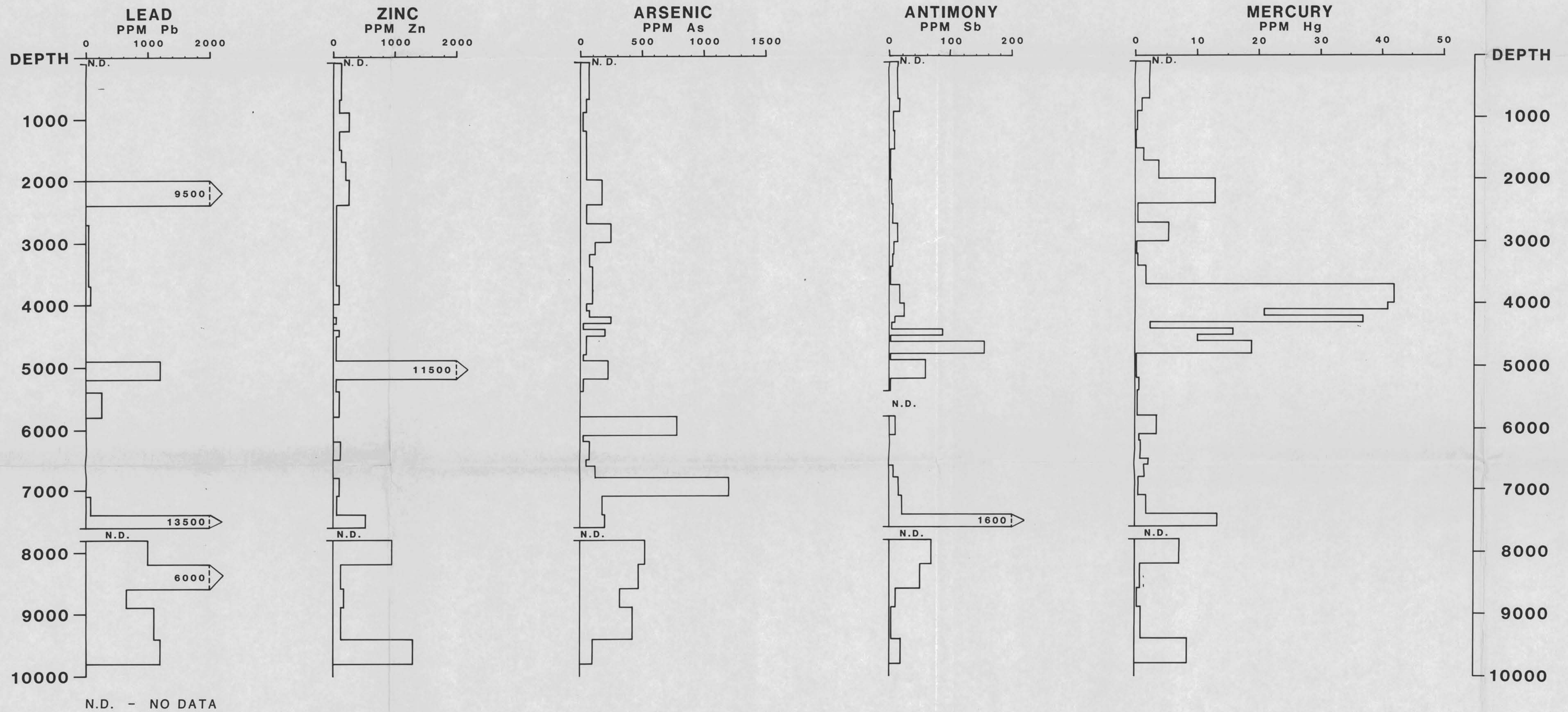
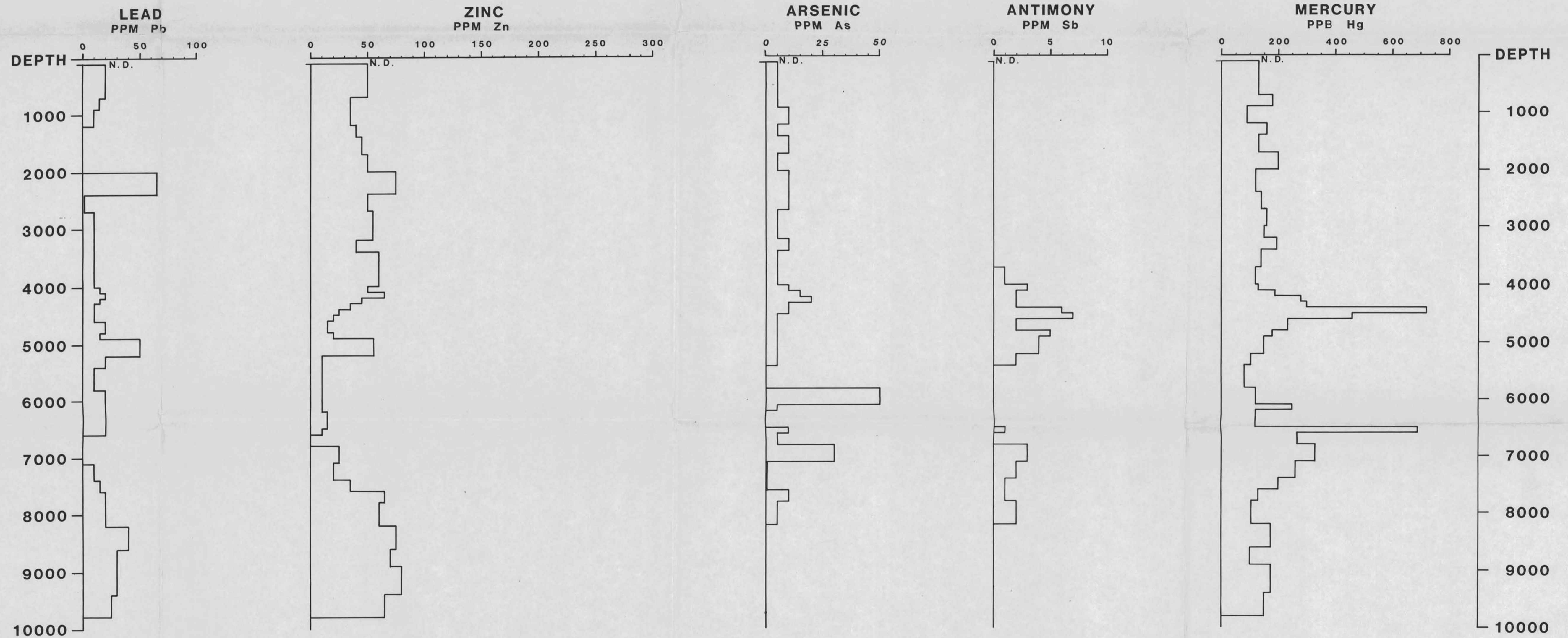


PLATE IX:

DISTRIBUTION OF SELECTED ELEMENTS IN DIXIE FEDERAL 66-21 WHOLE ROCK SAMPLES DIXIE VALLEY, CHURCHILL COUNTY, NEVADA

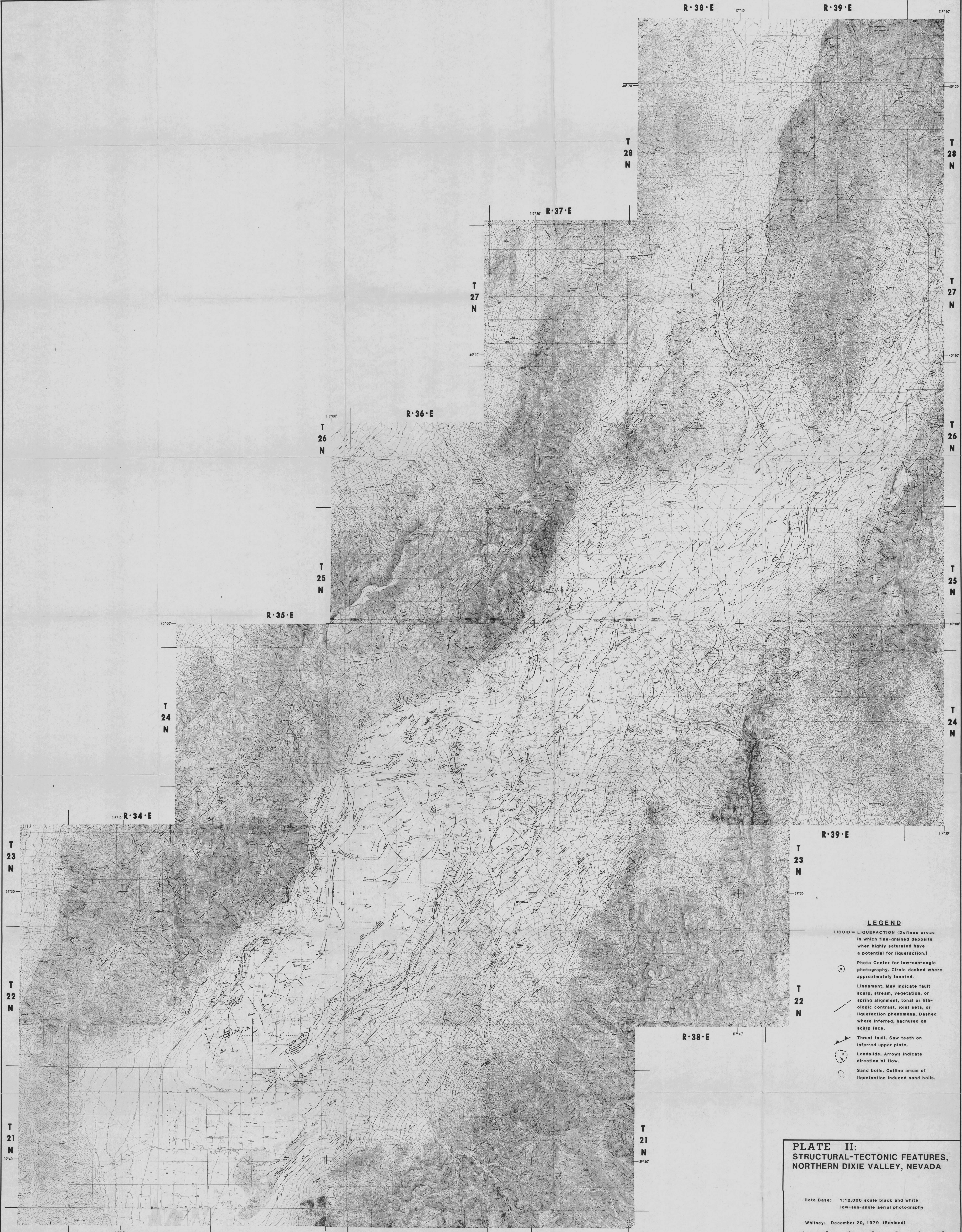


N.D. - NO DATA

REVISED: E.J. BELL JUNE 13, 1980
BY: T.R. BARD APRIL 24, 1980



DIXIE VALLEY PROSPECT
 NEVADA
 PLATE H-1
 ISOTOPIC AND HYDRO-CHEMICAL SAMPLE
 LOCATIONS IN DIXIE VALLEY
 SCALE 1"=1 MILE



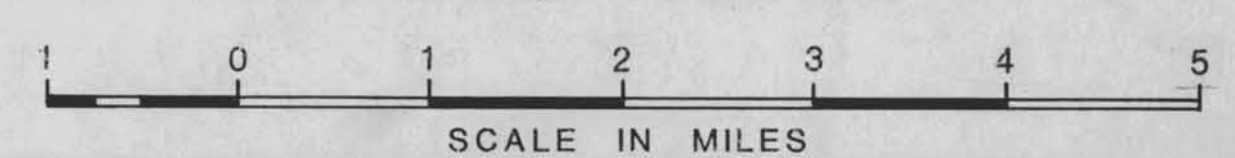
LEGEND

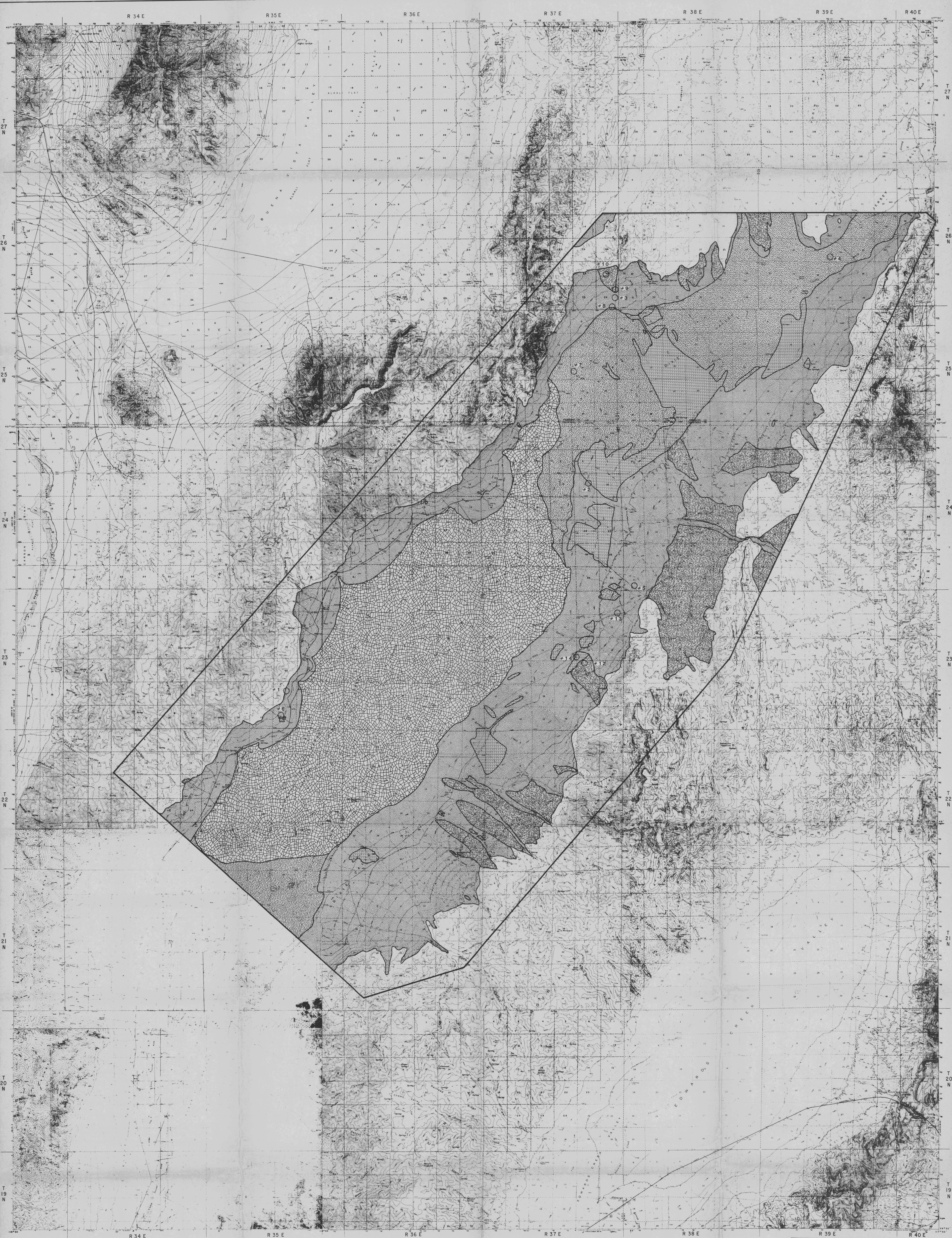
- LIQUID = LIQUEFACTION (Defines areas in which fine-grained deposits when highly saturated have a potential for liquefaction.)
- Photo Center for low-sun-angle photography. Circle dashed where approximately located.
- Lineament. May indicate fault scarp, stream, vegetation, or spring alignment, tonal or lithologic contrast, joint sets, or liquefaction phenomena. Dashed where inferred, hachured on scarp face.
- Thrust fault. Saw teeth on inferred upper plate.
- Landslide. Arrows indicate direction of flow.
- Sand boils. Outline areas of liquefaction induced sand boils.

**PLATE II:
STRUCTURAL-TECTONIC FEATURES,
NORTHERN DIXIE VALLEY, NEVADA**

Data Base: 1:12,000 scale black and white low-sun-angle aerial photography

Whitney: December 20, 1979 (Revised)





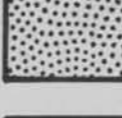





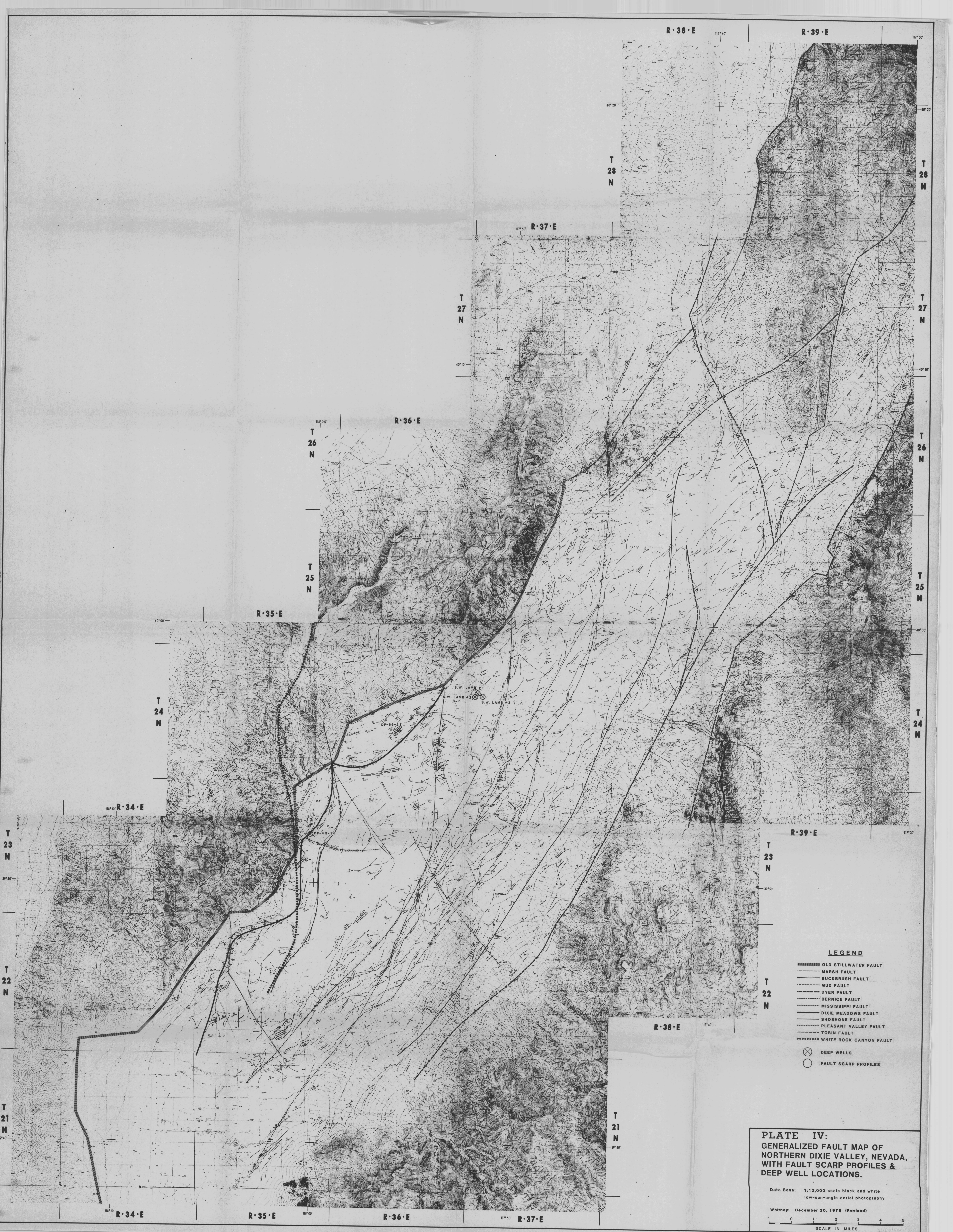
- LEGEND**
-  ACTIVE ALLUVIAL FANS
 -  INACTIVE ALLUVIAL FANS
 -  LACUSTRINE SURFACES
 -  COMPOSITE SURFACES
 -  ACTIVE PLAYA
 -  SAMPLE SITES FOR GRAIN-SIZE ANALYSIS

PLATE III:
GENERALIZED GEOMORPHIC MAP,
NORTHERN DIXIE VALLEY, NEVADA

By: Whitney Date: November 22, 1979

Scale (approx.): 1" = 1 Mile



LEGEND

- OLD STILLWATER FAULT
- MARSH FAULT
- BUCKBRUSH FAULT
- MUD FAULT
- DYER FAULT
- BERNICE FAULT
- MISSISSIPPI FAULT
- DIXIE MEADOWS FAULT
- SHOSHONE FAULT
- PLEASANT VALLEY FAULT
- TOBIN FAULT
- WHITE ROCK CANYON FAULT
- ⊗ DEEP WELLS
- FAULT SCARP PROFILES

**PLATE IV:
GENERALIZED FAULT MAP OF
NORTHERN DIXIE VALLEY, NEVADA,
WITH FAULT SCARP PROFILES &
DEEP WELL LOCATIONS.**

Data Base: 1:12,000 scale black and white
low-sun-angle aerial photography

Whitney: December 20, 1979 (Revised)

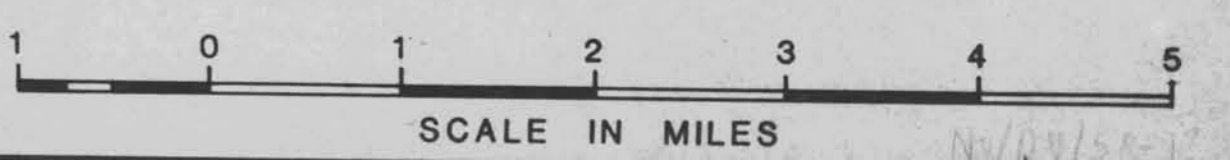
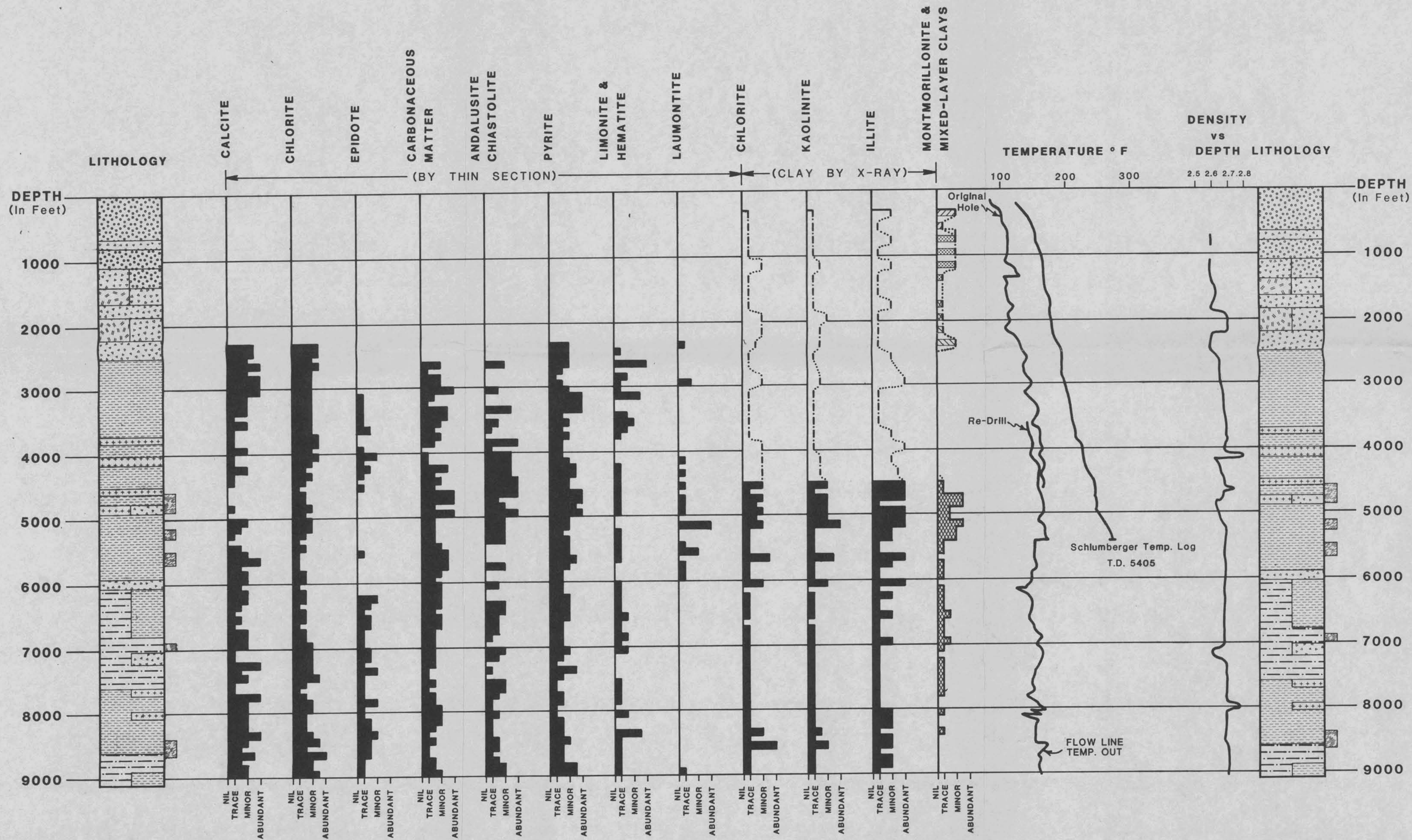


PLATE V: RELATIVE ABUNDANCE OF SELECTED MINERAL SPECIES IN DF-45-14



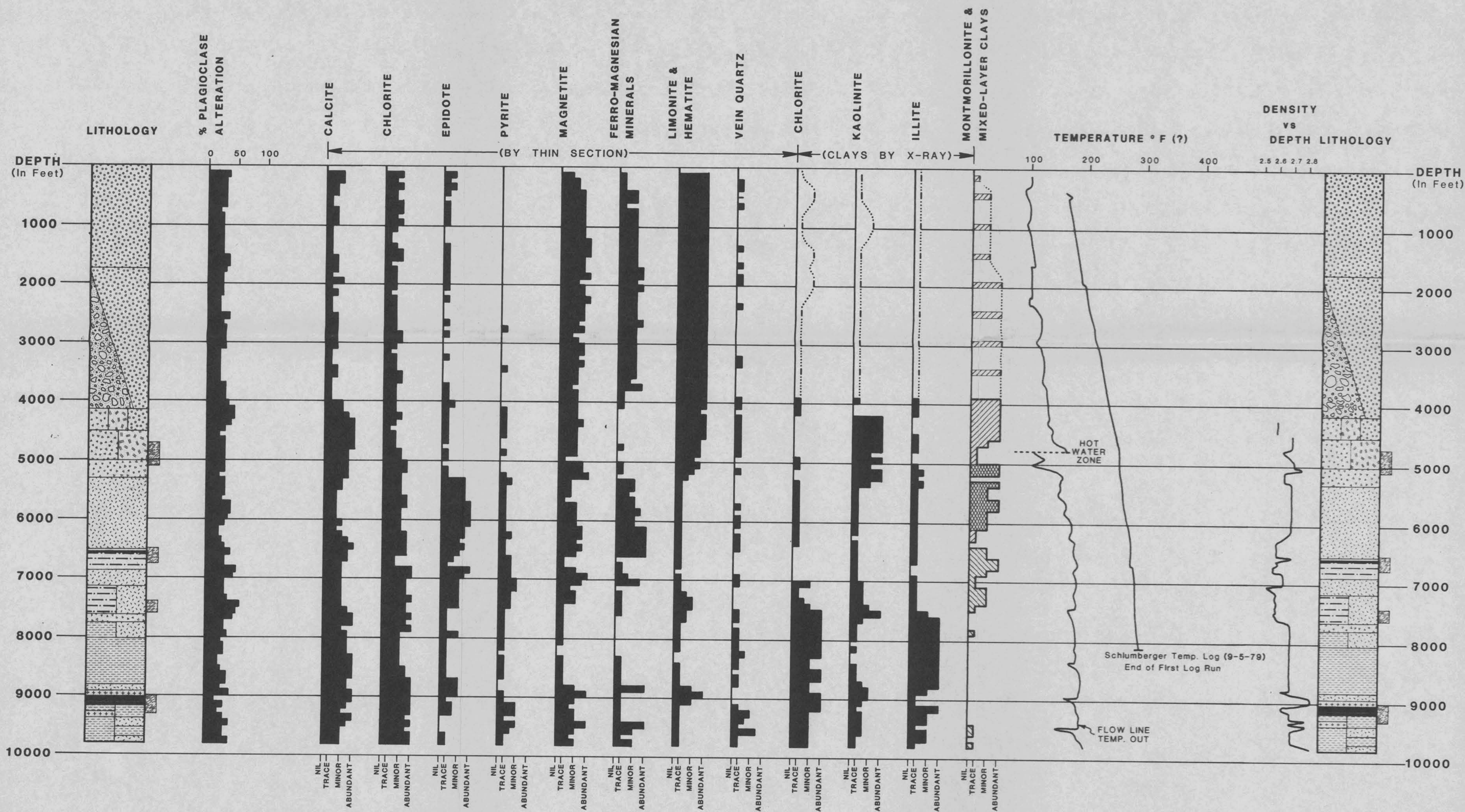
EXPLANATION - MINERAL ANALYSIS

- Randomly Interstratified Illite/Montmorillonite with Montmorillonite < 30%
- Calcium/Magnesium-Montmorillonite
- Sodium-Montmorillonite
- Regular Mixed-Layer Chlorite/Vermiculite
- Inferred

EXPLANATION - LITHOLOGY

- Alluvium
- Tuff
- Andesite/Basalt
- Metasiltstone/Metashale
- Gabbro/Diorite Intrusive
- Metarenite/Quartz Arenite
- Postulated Fracture/Fault Zones

PLATE VI: RELATIVE ABUNDANCE OF SELECTED MINERAL SPECIES IN DF-66-21



EXPLANATION - MINERAL ANALYSIS

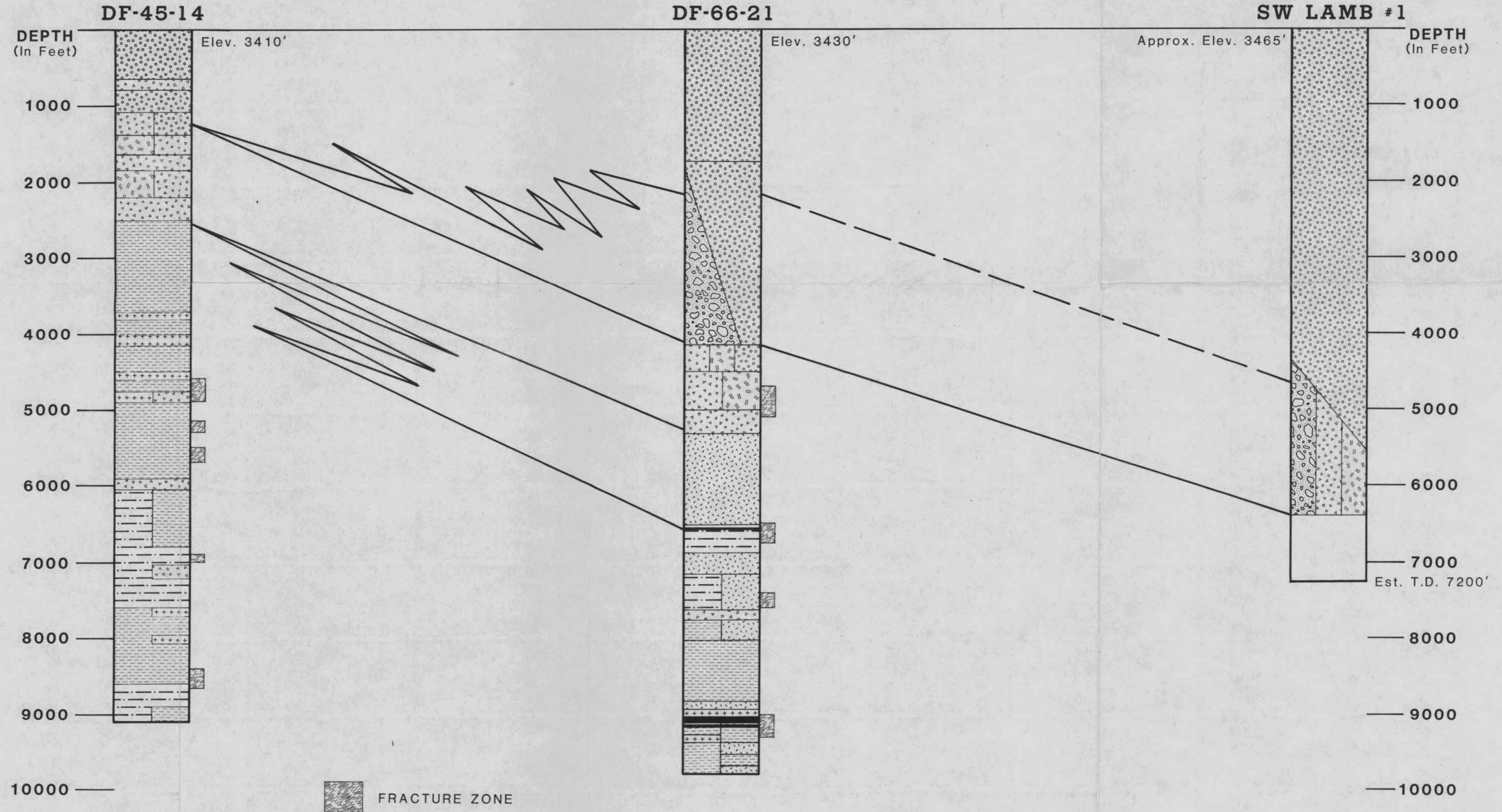
- Randomly Interstratified Illite/Montmorillonite with Montmorillonite > 30%
- Randomly Interstratified Illite/Montmorillonite with Montmorillonite < 30%
- Regular Mixed-Layer Chlorite/Vermiculite
- Inferred

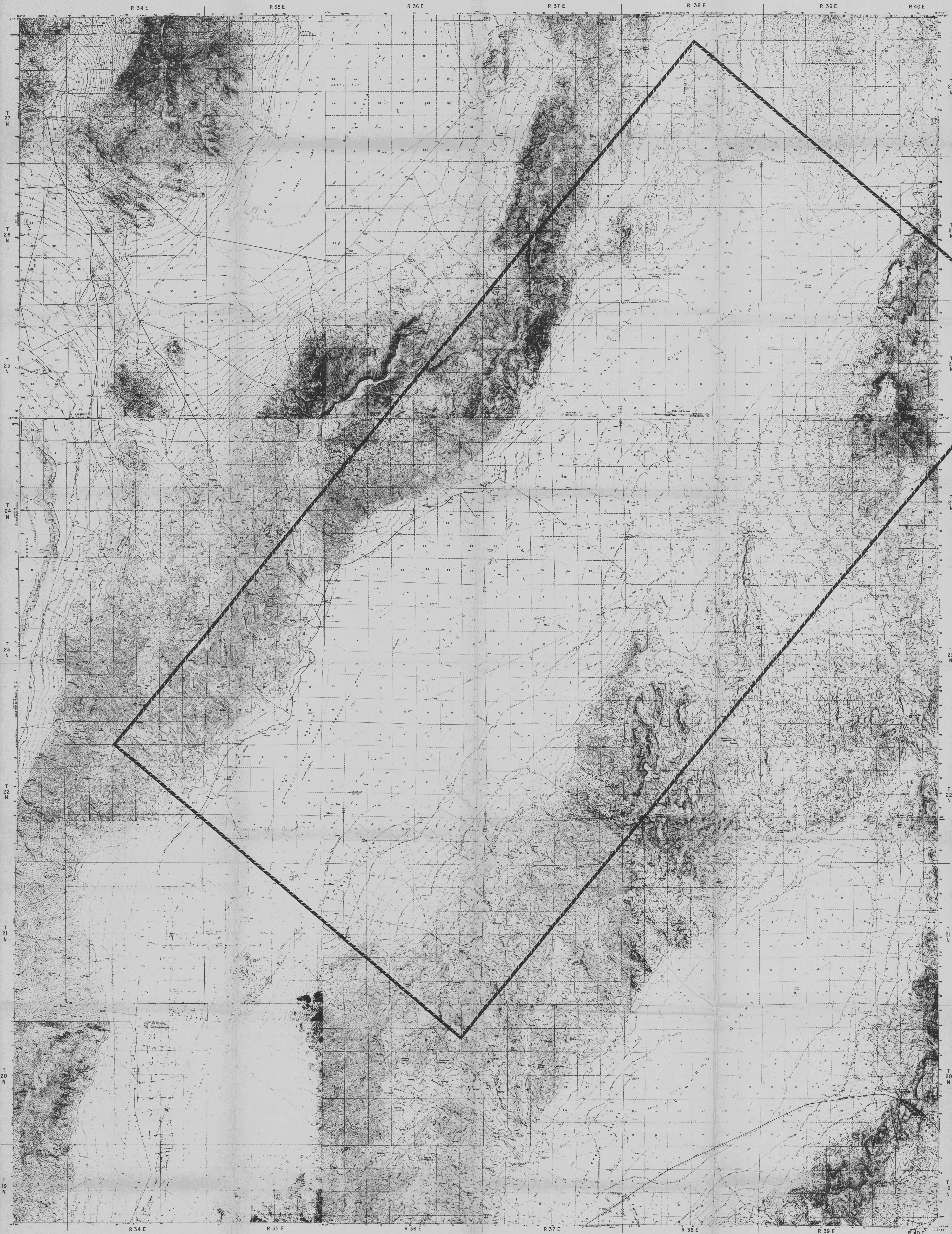
EXPLANATION - LITHOLOGY

- Alluvium
- Alluvium and Clay
- Tuff
- Andesite/Basalt
- Granodiorite Intrusive
- Quartz Arenite
- Metasiltstone/Metashale
- Gabbro/Diorite Intrusive
- Mineralized Zones
- Postulated Fracture/Fault Zones

PLATE VII: WELL CORRELATIONS

(Refer to Plates V & VI for Lithologic Descriptions)





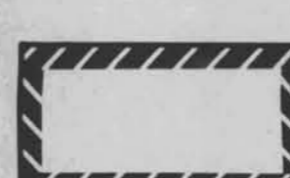
 STUDY AREA

PLATE 1:
DIXIE VALLEY PROSPECT
NEVADA

SCALE: 1" = 1 Mile