

## INFORMATION TO USERS

This reproduction was made from a copy of a document sent to us for microfilming. While the most advanced technology has been used to photograph and reproduce this document, the quality of the reproduction is heavily dependent upon the quality of the material submitted.

The following explanation of techniques is provided to help clarify markings or notations which may appear on this reproduction.

1. The sign or "target" for pages apparently lacking from the document photographed is "Missing Page(s)". If it was possible to obtain the missing page(s) or section, they are spliced into the film along with adjacent pages. This may have necessitated cutting through an image and duplicating adjacent pages to assure complete continuity.
2. When an image on the film is obliterated with a round black mark, it is an indication of either blurred copy because of movement during exposure, duplicate copy, or copyrighted materials that should not have been filmed. For blurred pages, a good image of the page can be found in the adjacent frame. If copyrighted materials were deleted, a target note will appear listing the pages in the adjacent frame.
3. When a map, drawing or chart, etc., is part of the material being photographed, a definite method of "sectioning" the material has been followed. It is customary to begin filming at the upper left hand corner of a large sheet and to continue from left to right in equal sections with small overlaps. If necessary, sectioning is continued again beginning below the first row and continuing on until complete.
4. For illustrations that cannot be satisfactorily reproduced by xerographic means, photographic prints can be purchased at additional cost and inserted into your xerographic copy. These prints are available upon request from the Dissertations Customer Services Department.
5. Some pages in any document may have indistinct print. In all cases the best available copy has been filmed.

**University  
Microfilms  
International**

300 N. Zeeb Road  
Ann Arbor, MI 48106

1320159

SZECSODY, JAMES EDWARD

USE OF MAJOR ION CHEMISTRY AND ENVIRONMENTAL ISOTOPES TO DELINEATE  
SUBSURFACE FLOW IN EAGLE VALLEY, NEVADA

UNIVERSITY OF NEVADA, RENO

M.S. 1982

University  
Microfilms  
International 300 N. Zeeb Road, Ann Arbor, MI 48106

PLEASE NOTE:

In all cases this material has been filmed in the best possible way from the available copy. Problems encountered with this document have been identified here with a check mark .

1. Glossy photographs or pages
2. Colored illustrations, paper or print \_\_\_\_\_
3. Photographs with dark background
4. Illustrations are poor copy \_\_\_\_\_
5. Pages with black marks, not original copy \_\_\_\_\_
6. Print shows through as there is text on both sides of page \_\_\_\_\_
7. Indistinct, broken or small print on several pages
8. Print exceeds margin requirements \_\_\_\_\_
9. Tightly bound copy with print lost in spine \_\_\_\_\_
10. Computer printout pages with indistinct print
11. Page(s) \_\_\_\_\_ lacking when material received, and not available from school or author.
12. Page(s) \_\_\_\_\_ seem to be missing in numbering only as text follows.
13. Two pages numbered \_\_\_\_\_. Text follows.
14. Curling and wrinkled pages \_\_\_\_\_
15. Other \_\_\_\_\_

University  
Microfilms  
International

University of Nevada  
Reno

Use of Major Ion Chemistry and Environmental Isotopes  
to Delineate Subsurface Flow  
in Eagle Valley, Nevada

A thesis submitted in partial fulfillment of the  
requirements for the degree of Master of Science,  
in Hydrology

by

James Edward Szecsody

October 1982

The thesis of James Edward Szecsody is approved:

Michael E. Campana  
Thesis Advisor

L. F. Larson  
Department Chairman

Philip H. Keller  
Dean, Graduate School

University of Nevada  
Reno

October 1982

## ACKNOWLEDGEMENTS

I would like to thank the following Desert Research Institute graduate assistants for helping me with the field work for this project: Neil Ingraham, Ken Shump, Tony Truschel, Marc Herman, Carol Janecke, Pete Ornstein, Chris Koltermann, Jon Rhodes, Jim Rumbaugh, Jack Dowden, and Lee Huckins. Special thanks go to Dr. Roger Jacobson, Dr. Burkhard Bohm, Dr. Michael Campana, Mr. John Fordam, and Dr. John Bird for their guidance in formulating my thesis. Commendation goes to the Desert Research Institute, Water Resources Center for providing field equipment and vehicles, computer facilities, and laboratory facilities without which this thesis would not be as complete.

I would like to give special thanks to my wife, Gail Cordy Szecsody, for her continued support, advice, and patience throughout the past two years, and my parents for their insight and perseverance over the past 24 years.

Financial support was provided by the Desert Research Institute, Water Resources Center, through grant #14-34-0001-1244 from the Office of Water Resources and Technology, U.S. Department of Interior.

## ABSTRACT

Snow sampling in the main recharge mountains of Eagle Valley indicates non-equilibrium fractionation of stable isotopes from:  $\delta D/H = 6.0(\delta^{18}O/^{16}O) - 14$ . There is significant correlation of isotopic depletion with elevation indicated by:  $\delta D/H = -9.9\%/1000$  ft and  $\delta^{18}O/^{16}O = -1.48\%/1000$  ft. Electrical conductivity and chloride enrichment from snow to lysimeter water produces 4000 acre-ft/year of potential recharge from the Carson Range. Monitoring of lysimeters throughout the winter shows that ground is not frozen under snow. Contour maps of sulfate, chloride, sodium, carbon-13, and deuterium concentrations in the valley aquifer suggest thermal/non-thermal water mixing. Aquifer stable isotope values and recharge area values indicate recharge to the aquifer via stream channel infiltration and deep percolation. Tritium and carbon-14 dating of non-thermal water in the aquifer shows recharge from the Carson Range and near the Carson River, and the oldest non-thermal water in the basin center. Carson Hot Springs is about 12,500 years old and is isotopically depleted, whereas Prison Hot Springs contains recent water and is not depleted.

## TABLE OF CONTENTS

	page
ACKNOWLEDGEMENTS . . . . .	ii
ABSTRACT. . . . .	.iii
LIST OF FIGURES . . . . .	.vii
LIST OF TABLES. . . . .	ix
1. INTRODUCTION . . . . .	1
2. ENVIRONMENTAL SETTING	
2.1 GEOGRAPHICAL SETTING . . . . .	5
2.2 METEOROLOGY . . . . .	6
2.2.1 Regional Weather Systems. . . . .	6
2.2.2 Local Weather Systems. . . . .	7
2.3 VEGETATION . . . . .	12
2.4 GEOLOGY . . . . .	13
2.5 HYDROLOGY. . . . .	15
2.5.1 Physical Hydrology. . . . .	15
2.5.2 Water Chemistry. . . . .	19
3. THEORY OF HYDROGEOCHEMICAL INVESTIGATIONS	
3.1 MAJOR CONSTITUENT CHEMISTRY. . . . .	22
3.2 STABLE ISOTOPE CHEMISTRY. . . . .	27
3.2.1 Introduction. . . . .	27
3.2.2 Stable Isotopes during Evaporation and Vapor Transport . . . . .	32
3.2.3 Stable Isotopes during Condensation . . . . .	37
3.2.4 Stable Isotopes during Infiltration . . . . .	42
3.2.5 Stable Isotopes in Trees. . . . .	45
3.2.6 Stable Isotopes in Groundwater Flow . . . . .	47
3.3 TRITIUM DATING OF GROUNDWATER . . . . .	49
3.4 RADIOCARBON DATING OF GROUNDWATER. . . . .	52
4. FIELD METHODS	
4.1 INTRODUCTION. . . . .	59



## TABLE OF CONTENTS Continued

	page
4.2 RECHARGE AREA SAMPLING . . . . .	59
4.2.1 Snow Sampling . . . . .	59
4.2.2 Lysimeter Water Sampling. . . . .	63
4.2.3 Soil Sampling . . . . .	67
4.2.4 Tree Water Sampling . . . . .	67
4.3 STREAM GAGING AND SAMPLING . . . . .	70
4.4 DISCHARGE AREA SAMPLING . . . . .	72
4.4.1 Well and Spring Sampling. . . . .	72
4.4.2 Stream Gaging . . . . .	74
5. LABORATORY METHODS	
5.1 MAJOR ION ANALYSIS. . . . .	75
5.2 STABLE ISOTOPE ANALYSIS . . . . .	77
5.2.1 Introduction. . . . .	77
5.2.2 Soil and Tree Water Distillation . . . . .	77
5.2.3 Stable Isotope Sample Preservation . . . . .	81
5.2.4 Stable Isotope Analysis . . . . .	81
5.2.4.1 Sample Preparation. . . . .	82
5.2.4.2 Measurement by Mass Spectrometer . . . . .	83
5.3 TRITIUM ANALYSIS . . . . .	86
5.3.1 Introduction. . . . .	86
5.3.2 Analysis Method. . . . .	86
5.4 CARBON ISOTOPE ANALYSIS . . . . .	89
5.4.1 Introduction. . . . .	89
5.4.2 Carbonate Precipitation . . . . .	90
5.4.3 Analysis Method. . . . .	93
6. RESULTS AND DISCUSSION	
6.1 CHEMISTRY OF PRECIPITATION . . . . .	95
6.1.1 Stable Isotope Chemistry. . . . .	95
6.1.2 Radioisotope Chemistry . . . . .	.101

## TABLE OF CONTENTS Continued

	page
6.2 CHEMISTRY OF INFILTRATED WATER. . . . .	101
6.2.1 Major Ion Chemistry . . . . .	101
6.2.2 Stable Isotope Chemistry. . . . .	105
6.2.2.1 Snow to Lysimeter Water . . . . .	105
6.2.2.2 Lysimeters . . . . .	111
6.2.2.3 Tree Water . . . . .	114
6.2.3 Radiocarbon Sampling in Recharge Areas .	123
6.3 STREAMFLOW IN RECHARGE AREAS . . . . .	123
6.4 AQUIFER RECHARGE WATER CHEMISTRY . . . . .	129
6.5 WATER CHEMISTRY OF THE EAGLE VALLEY AQUIFER .	134
6.5.1 Introduction. . . . .	134
6.5.2 Aquifer Physical Parameters. . . . .	137
6.5.3 Major Ion Chemistry . . . . .	140
6.5.3.1 Major Anions. . . . .	146
6.5.3.2 Major Cations . . . . .	153
6.5.4 Stable Isotope Chemistry. . . . .	158
6.5.5 Groundwater Age Dating . . . . .	167
6.5.5.1 Tritium Age Dating. . . . .	167
6.5.5.2 Carbon Isotope Age Dating . . . . .	170
7. CONCLUSIONS AND RECOMMENDATIONS	
7.1 RECHARGE AREAS . . . . .	176
7.2 STREAM CHEMISTRY . . . . .	179
7.3 AQUIFER CHEMISTRY . . . . .	180
REFERENCES CITED . . . . .	184
APPENDIX 1 EAGLE VALLEY - MAJOR ION ANALYSES. . . . .	192
APPENDIX 2 EAGLE VALLEY - STABLE ISOTOPE ANALYSES . . . . .	193
APPENDIX 3 TRITIUM ANALYSES . . . . .	194
APPENDIX 4 ASH CANYON CREEK - FLOW AND ISOTOPE DATA .	194
APPENDIX 5 PRECIPITATION DATA FROM THE CARSON RANGE .	194
APPENDIX 6 DERIVATION OF THE INFILTRATION EQUATION .	195

## LIST OF FIGURES

	page
Figure 1.1 Eagle Valley Basin Location . . . . .	2
Figure 2.1 Annual Precipitation versus Elevation in the Carson Range . . . . .	9
Figure 3.1 Hydrogen and Oxygen Isotope Fractionation under Equilibrium Conditions . . . . .	34
Figure 3.2 Calculated Isotopic Composition of Vapor and Precipitation in the Sierra Nevadas . . . . .	39
Figure 3.3 Tritium Concentration in Precipitation. . . . .	51
Figure 3.4 The Carbon Isotopic Cycle . . . . .	57
Figure 4.1 Snow Sampling with Modified PVC Pipe . . . . .	61
Figure 4.2 Sampling a Pressure-Suction Lysimeter . . . . .	61
Figure 4.3 SrCO <sub>3</sub> Precipitate from Soil-CO <sub>2</sub> . . . . .	68
Figure 4.4 Tree Coring. . . . .	68
Figure 4.5 Stream Gaging with Transducer Flow Meter in Ash Canyon Creek at 7800 ft . . . . .	71
Figure 5.1 Soil and Tree Water Distillation Apparatus	79
Figure 5.2 Carbonate Precipitation Apparatus . . . . .	91
Figure 6.1 $\delta D/H$ versus $\delta^{18}O/^{16}O$ for Snow Sampled in Spring, 1981 . . . . .	96
Figure 6.2 $\delta D/H$ versus Elevation for Spring, 1981 Snow Samples . . . . .	97
Figure 6.3 $\delta^{18}O/^{16}O$ versus Elevation for Spring, 1981 Snow Samples. . . . .	98
Figure 6.4 Percent Potential Soil Zone Recharge versus Elevation . . . . .	102
Figure 6.5 Volume of Potential Soil Zone Recharge versus Elevation . . . . .	104
Figure 6.6 Snow to Soil Water Isotopic Fractionation at 8500 ft in Ash Canyon . . . . .	107
Figure 6.7 Isotopic Shift from Snow to Soil Water in Ash Canyon . . . . .	110
Figure 6.8 Soil Water $\delta D/H$ Elevational Shift . . . . .	112
Figure 6.9 Soil Water $\delta^{18}O/^{16}O$ Elevational Shift. . . . .	113
Figure 6.10 Tree Xylem Water $\delta D/H$ Elevational Shift . . . . .	115
Figure 6.11 Tree Xylem Water $\delta^{18}O/^{16}O$ Elevational Shift. . . . .	116
Figure 6.12 Comparison of Soil and Tree Water $\delta D/H$ Elevational Shift . . . . .	120

## LIST OF FIGURES Continued

	page
Figure 6.13 Comparison of Soil and Tree Water $\delta^{18}/^{16}$ Elevational Shift . . . . .	121
Figure 6.14 Results of August 27, 1981 Stream Gaging on Ash Canyon Creek . . . . .	126
Figure 6.15 Results of January 29, 1981 Stream Gaging on Ash Canyon Creek . . . . .	128
Figure 6.16 Ash Canyon Creek Estimated Isotopic Composition . . . . .	133
Figure 6.17 Aquifer Sample Locations . . . . .	135
Figure 6.18 Temperature ( $^{\circ}\text{C}$ ), 46 Wells . . . . .	138
Figure 6.19 Electrical Conductivity ( $\mu\text{mhos/cm}$ ). . . . .	139
Figure 6.20 Electrical Conductivity Trend Surface Maps ( $\mu\text{mhos/cm}$ ) . . . . .	141
Figure 6.21 Piper Diagram of Waters in Eagle Valley . . . . .	144
Figure 6.22 Major Ion Changes Eastward across Eagle Valley . . . . .	145
Figure 6.23 Density Plot of Piper Diagram . . . . .	147
Figure 6.24 Alkalinity ( $\text{HCO}_3^- + \text{CO}_3^{2-}$ , in ppm), 59 Wells. . . . .	149
Figure 6.25 Sulfate (ppm), 54 Wells . . . . .	150
Figure 6.26 Chloride (ppm), 58 Wells . . . . .	152
Figure 6.27 Calcium (ppm), 55 Wells . . . . .	154
Figure 6.28 Sodium (ppm), 56 Wells. . . . .	156
Figure 6.29 Oxygen versus Hydrogen Isotopes of Wells in Eagle Valley . . . . .	159
Figure 6.30 $\delta^{18}/^{16}$ (‰), 22 Wells . . . . .	160
Figure 6.31 $\delta^{18}/^{16}$ (‰), 22 Wells, Corrected for Laboratory Differences. . . . .	161
Figure 6.32 $\delta\text{D}/\text{H}$ (‰), 22 Wells. . . . .	164
Figure 6.33 $\delta\text{D}/\text{H}$ (‰), 22 Wells, Corrected for Laboratory Differences. . . . .	165
Figure 6.34 Tritium Data (TU) . . . . .	168
Figure 6.35 Apparent Groundwater Age from Carbon Isotope Data (years) . . . . .	173

## LIST OF TABLES

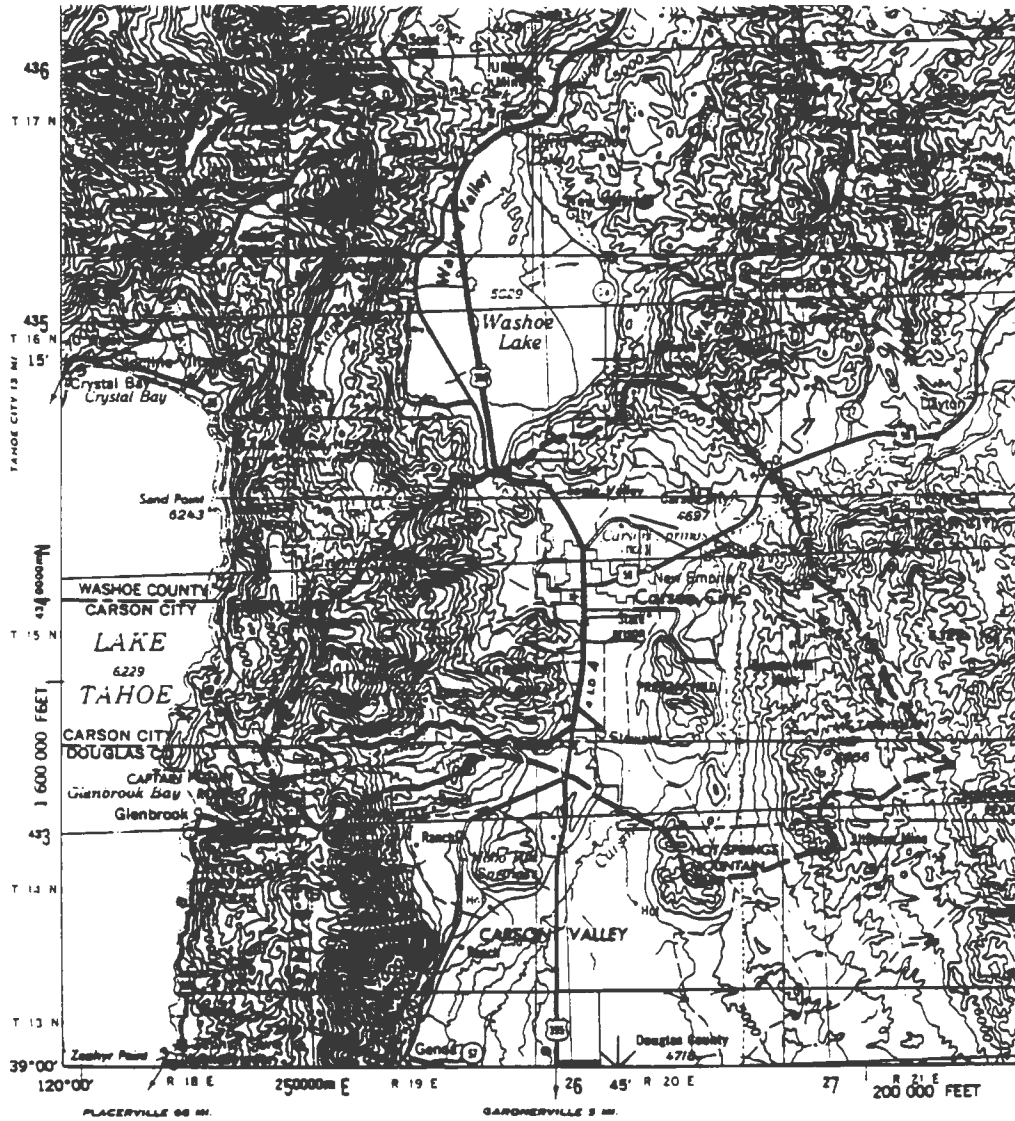
	page
Table 2.1 Precipitation Data in Eagle Valley . . .	10
Table 2.2 Temperature Data of Eagle Valley and Surrounding Areas . . . . .	10
Table 6.1 Chloride and Electrical Conductivity Data . . . . .	102
Table 6.2 Potential Recharge Data . . . . .	104
Table 6.3 Major Ion Changes over Time. . . . .	142
Table 6.4 Carbon Isotope Analysis . . . . .	172

## CHAPTER 1

## INTRODUCTION

Eagle Valley is located along the western border of Nevada at the foot of the Sierra Nevadas. This basin (see Figure 1.1), which covers about 117 square miles, encompasses Nevada's capital, Carson City, and the surrounding communities of New Empire and Stewart. The population of Carson City has increased from 15,000 in 1965 to 37,000 in 1980, which has led to increased water use. In early years, water demand was small compared to the amount available. As the population of Eagle Valley grew, the demand for water had increased to the point where groundwater pumping became necessary. Previous investigators studying the available water resources in Eagle Valley (Worts and Malmberg, 1966) concluded that surface and groundwater resources would be fully developed by 1980. It is, therefore, necessary to have updated information on the groundwater system in Eagle Valley. The purpose of this investigation is to study the water chemistry changes of both recharge and discharge areas of the Eagle Valley groundwater system by use of major ion chemistry, hydrogen and oxygen isotope chemistry, tritium and carbon isotope chemistry.

Recent work devoted to development of the groundwater system was done by the U.S. Geological Survey (Arteaga and Durbin, 1978), where again, the statement was made that the demand for water is approaching the limits of the resources.



--- Basin Boundary

Figure 1.1 Eagle Valley Basin Location

Further study of the geothermal part of the groundwater system (Trexler, Koenig, Flynn, and Bruce, 1980) yielded information on the origin of the geothermal fluids. However, there are no recent studies of the water chemistry of the entire flow cycle in Eagle Valley, as is done in this study.

The present study is devoted to describing and quantifying the changes in water chemistry in the hydrologic cycle from precipitation, to infiltration, runoff, and finally groundwater flow. Previous investigators have concluded that most recharge is from the Carson Range, in the western part of Eagle Valley. Therefore, a flow path from Snow Valley Peak in the west to the Carson River in the east was studied in detail, and is augmented by a more general study of chemistry changes in the entire basin.

Stable isotope chemistry and chloride ion changes were used in the recharge portion of the basin to: 1) calculate the volume of water available for recharge from different elevations after evaporation; 2) calculate the average recharge elevation; and 3) estimate an average isotopic composition of water that is potential recharge to the groundwater system. Stable isotope content of wells in the valley alluvium are compared to the calculated potential input to the aquifer in order to delineate recharge flow paths.

Major ion chemistry is used in conjunction with stable isotope chemistry of water in the valley to trace



groundwater flow. In other studies, (Freeze and Cherry, 1979) major ions have successfully been used to indicate groundwater flow direction and relative velocity.

Estimation of the groundwater age was made by tritium and carbon isotope dating. Tritium, a radioactive isotope of hydrogen, is produced in the atmosphere by cosmic ray bombardment and from atmospheric thermonuclear testing. Tritium becomes part of the groundwater system when removed from surface water, and decays at a constant rate. Variations in tritium levels in a groundwater system can be used to estimate the age of water to a maximum of about 50 years.

Carbon isotopes are also produced in the atmosphere by cosmic ray bombardment. Radiocarbon age dating works on the principle of measuring the activity of dissolved carbonate in groundwater and comparing it to modern activity. Unfortunately, carbonate in water comes from two different sources: initially from dissolved  $\text{CO}_2$  in the soil zone in recharge areas and from dissolving carbonate along the groundwater flow path. Soil zone  $\text{CO}_2$  can be measured and a stable isotope of carbon (C-13) can be used to correct for contamination by dissolving carbonate.

## CHAPTER 2

### ENVIRONMENTAL SETTING

#### 2.1 GEOGRAPHICAL SETTING

Eagle Valley is located along the western border of Nevada at the eastern foot of the Sierra Nevada. The basin, as described in this report (Plate 1), is defined as the area in which surface water drains into the Carson River from T14N, R20E, Sec 9, to T15N, R21E, Sec 5. This corresponds to a 117 square mile area with surface runoff from the Carson Range of the Sierras to the west, the Virginia Range in the north, and the Pine Nut Range to the east. A low alluvial divide separates Eagle Valley from Carson Valley to the south. This area also defines the boundary for potential flow into the Eagle Valley groundwater system, similar to the area defined by Mifflin (1968), although flow into the basin does not extend as far east as Bismark Peak as Mifflin describes.

The valley floor of Eagle Valley is at 4600 to 4800 feet in elevation with mountains rising to 9200 feet in the Carson Range, 7000 feet in the Virginia Range and 6700 feet in the Pine Nut Mountains. The Carson River flows northward through the valley, along the eastern margin of the alluvium. The major population centers in Eagle Valley are the state capital, Carson City, Stewart, and New Empire.

## 2.2 METEOROLOGY

### 2.2.1 Regional Weather Systems

Most of the storms that produce measurable precipitation in Eagle Valley come from the west, since the prevailing wind direction in northern California is west to east. There is a wide variation of source areas and characteristics of storms. The types of storms that affect northern Nevada and the Sierra Nevada can be classified into five categories based on meteorological synoptic pattern and latitude of origin (Smith, Friedman, Klieforth, and Hardcastle, 1979). These are:

- (1) High (north) latitude front and upper trough (maritime polar air);
- (2) Midlatitude front and upper trough (maritime polar air with some maritime tropical air);
- (3) Low (equator) latitude front with deep upper trough (maritime tropical air);
- (4) Confluent air masses brought together over a region along a frontal surface (with tropical air contributing most of the moisture); and
- (5) Cold cyclone (mostly continental polar or modified maritime polar air, little or no frontal activity)

Midlatitude front, confluent air masses, and cold cyclones are the most common types of storms in the Sierra Nevada. Confluent air mass storms produce substantially more precipitation than midlatitude and cold cyclone type storms. However, cold cyclone and midlatitude fronts

produce lower snow lines than confluent air mass storms, since they are colder storms. The coldest storms pass over the Sierra Nevada Mountains between the winter solstice and the vernal equinox, the time of year with the least amount of solar radiation. As these storms move inland (east) toward the Sierra Nevada, adiabatic cooling produces precipitation. Approximately 90% of the moisture (which is mostly snow) that falls on the Sierra Nevada falls west of the divide, and 10% falls east of the divide. Samples from the Mojave Desert suggest that condensation occurs at about 6500 ft (2000 m) above the ground (Smith, Friedman, Klieforth and Hardcastle, 1979). In contrast, samples from the west slope of the Sierra Nevada suggest condensation occurs on the average, 3000 to 5000 ft (1000 to 1500 m) above the ground, illustrating that condensation over mountains occurs closer to the land surface.

#### 2.2.2 Local Weather Systems

Eagle Valley, located on the eastern edge of the Sierra Nevada, receives winter storms that precipitate on the Sierras as well as summer thunderstorms from the southeast. Locally, two main storm types are recognized. The cold cyclone type produces light surface winds, little snow drifting, and a low amplitude profile of precipitation versus elevation, a ratio of 2:1 in amounts of precipitation measured at high and low elevations, respectively

(Klieforth, 1981). The second major type of storm is the confluent air mass, which produces strong westerly or southwesterly air flow with precipitation and profiles of 20:1 on the leeward (eastern) side of the mountains. A third type of storm, summer thunderstorms, move in from the southeast.

The amount of precipitation change in the Carson Range per elevation change has been calculated at 5.72in/1000ft (Worts and Malmberg, 1966), 5.67in/1000ft (Arteaga and Durbin, 1978), and 5.72in/1000ft (Klieforth, 1981; see Figure 2.1). The Virginia Range (north and east of Carson City) average about 3.0in/1000ft elevation change.

Comparison of meteorological data from the 1980-81 year to the average is important in this study since isotopic content of precipitation is dependent on temperature; precipitation and temperature data are presented in Tables 2.1 and 2.2.

Precipitation data (Table 2.1) indicate the 1980-81 year was about 50 to 60% of average precipitation. There is considerable variation in the data; Carson City and Spooner Summit data both were 60% of the average annual precipitation, while Clear Creek station was 27% of average. Throughout the 1980-81 precipitation year, there was extreme monthly variation and no regular variation between stations. For example, the percent change between the 1980-81 year and

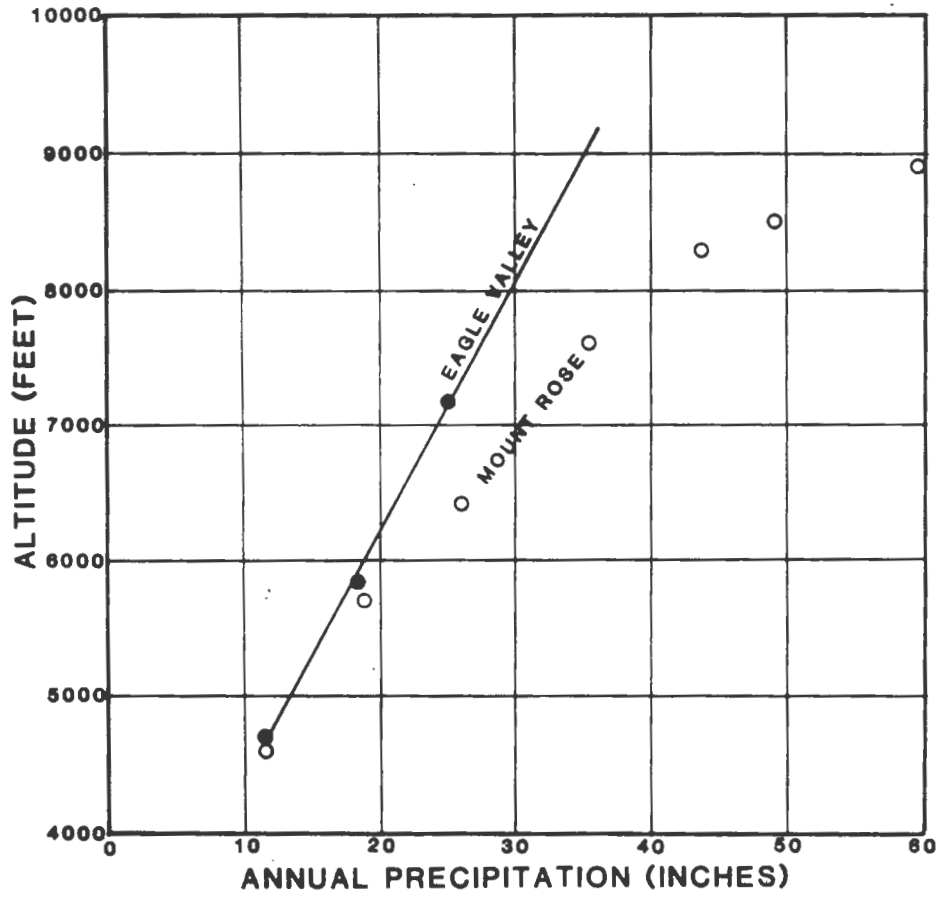


Figure 2.1 Annual Precipitation versus Elevation  
in the Carson Range (Klieforth, 1981)

PRECIPITATION DATA IN EAGLE VALLEY  
AVERAGE MONTHLY PRECIPITATION

Site	Elev. (ft)	Jul	Aug	Sep	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Annual Precip.	Years Averaged
Carson City	4700	.26	.18	.28	.59	1.45	2.23	2.27	1.49	1.11	.54	.66	.45	11.50	1875-1965
Clear Creek	5800	.25	.44	.54	1.55	3.12	3.34	4.25	2.87	2.30	1.02	.61	.11	20.40	1970-1982
Spooner Summit	7146	.44	.39	.40	1.63	3.24	4.56	4.64	4.18	3.39	1.76	.89	.12	25.64	1970-1982

1980-1981 MONTHLY PRECIPITATION

Site	Elev. (ft)	Jul	Aug	Sep	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Precip.	% as Snow
Carson City	4700	.19	.10	.57	.24	.66	.82	2.24	.22	.82	.36	.63	.00	6.85	--
Clear Creek	5800	.00	.00	.00	.00	.00	0.64	2.64	.49	1.44	.55	.41	.00	5.53	34
Spooner Summit	7146	.20	.46	.00	1.04	.92	1.59	5.08	1.62	2.60	1.75	.66	.00	15.92	69

1980-81 YEAR/AVERAGE (%)

Site	Elev. (ft)	Jul	Aug	Sep	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Annual
Carson City	4700	73	56	204	41	46	37	99	15	74	67	95	0	59.6
Clear Creek	5800	0	0	0	0	0	0	62	17	63	54	67	0	27.1
Spooner Summit	7146	45	117	0	64	28	35	109	39	77	99	74	0	62.1

TEMPERATURE DATA IN EAGLE VALLEY AND SURROUNDING AREAS

AVERAGE MONTHLY TEMPERATURE (°F)

Site	Elev. (ft)	Jul	Aug	Sep	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Annual Average	Years Averaged
Carson City	4700	69.7	67.6	61.3	51.5	40.9	34.7	32.8	37.7	41.1	47.4	54.9	62.0	50.0	1941-1970
Minden	4720	68.5	66.7	60.2	50.7	39.7	33.1	30.6	36.0	39.4	46.2	53.8	60.9	48.8	1941-1970
Glenbrook	6400	65.5	64.8	59.2	49.5	39.3	34.0	32.3	33.9	36.0	42.1	49.3	57.2	46.9	1945-1970
Mt Rose Bowl	8250	62.1	--	53.6	43.8	33.6	30.1	27.2	27.2	--	33.2	42.7	54.1	40.7	1973-1981
Virginia City	6400	70.6	69.0	61.7	51.2	41.5	34.1	33.1	35.5	39.0	45.1	52.7	61.8	49.6	1952-1971

1980-1981 MONTHLY TEMPERATURE (°F)

Site	Elev. (ft)	Jul	Aug	Sep	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Annual	Location
Carson City	4700	70.5	67.4	61.5	51.5	42.9	37.2	37.1	39.9	42.7	51.2	56.4	68.7	52.3	
Minden	4720	69.0	66.3	59.9	50.0	41.6	35.1	36.0	37.5	40.4	50.1	55.3	66.1	50.7	12mi. S of CC
Glenbrook	6400	64.1	63.3	58.5	51.5	43.9	38.6	37.8	37.3	38.0	45.7	50.2	61.8	49.2	10mi. SW of CC
Mt Rose Bowl	8250	61.0	--	51.3	46.2	37.2	33.1	33.0	29.3	--	39.9	43.2	57.8	43.2	14mi. NW of CC
Virginia City	6400	70.6	67.1	60.0	52.3	42.7	38.4	37.1	36.2	37.6	46.5	51.1	64.9	50.4	*10mi. NE of CC

1980-81 YEAR - AVERAGE (°F)

Site	Elev. (ft)	Jul	Aug	Sep	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Annual	Oct-May
Carson City	4700	+0.8	-0.2	+0.2	0.0	+2.0	+2.5	+4.3	+2.2	+1.6	+3.8	+1.5	+6.7	+2.2	+2.2
Minden	4720	+0.5	-0.4	-0.3	-0.2	+1.9	+2.0	+5.4	+1.5	+1.0	+3.9	+1.5	+5.2	+1.9	+2.1
Glenbrook	6400	-1.4	-1.5	-0.7	+2.0	+4.6	+4.6	+5.5	+3.4	+2.0	+3.6	+0.9	+4.6	+2.3	+3.3
Mt Rose Bowl	8250	-0.2	--	-2.3	+2.4	+3.6	+3.0	+5.8	+2.1	--	+6.7	+0.5	+3.7	+2.5	+3.0
Virginia City	6400	0.0	-1.9	-1.7	+1.1	+1.2	+4.3	+4.0	+0.7	-1.4	+1.4	-1.6	+3.1	+0.8	+1.2

\* All stations in the Carson Range, except Virginia City

Tables 2.1 and 2.2 Precipitation and Temperature Data (Kiloforth, 1981)

the average for August indicates Spooner Summit (7146 ft) was 117% of normal, Clear Creek station (5800 feet) was 0% of normal, and Carson City (4700 ft) was 56% of normal.

Temperature data (Table 2.2) show that for four stations in or at the foot of the Carson Range, the 1980-81 year averaged  $2.23 \pm 0.1^{\circ}\text{F}$  ( $1.24^{\circ}\text{C}$ ) warmer than the average for the entire year. The temperature variation was regular between stations over a fairly large area. Monthly temperature data indicate that November through June of 1980-81 was substantially warmer than the average, whereas July through October temperatures in 1980-81 were essentially the same as the average. Since the majority of precipitation from the 1980-81 year was during October through May, the average temperature change for these eight months indicates the 1980-81 year was  $2.65 \pm 0.6^{\circ}\text{F}$  ( $1.47^{\circ}\text{C}$ ) warmer than the average. This temperature difference is used with isotopic data from the 1980-81 year to calculate average isotopic composition.

Approximately ten miles west of Carson City is Lake Tahoe, which has a surface area of about 200 square miles. The evaporation from Lake Tahoe has been estimated at 60-65% of inflow (Friedman, Redfield, Schon, and Harris, 1964), which amounts to 30 in/year or 304,000 acre-ft/year. This amount of evaporation probably contributes to moisture in clouds which precipitate over Eagle Valley.



### 2.3 VEGETATION

Eagle Valley has a considerable range in climate from the valley floor to high elevation mountains, which produces different vegetation groups. The valley floor is mostly pasture and urbanized lands where the natural vegetation has been significantly altered. The vegetation in areas in the valley that are not developed can be classified as northern desert shrub, which consists of sagebrush, rabbitbrush, horsebrush, cheat grass, and a variety of other grasses (Bamberg and Friesen, 1972). Areas around stream channels are commonly vegetated with phreatophytes such as cottonwood trees.

Higher elevations in the Virginia Range and Pinyon Hills (up to 7000 ft) are vegetated with similar shrubs with increasing abundance of Utah juniper and Pinyon pine. The Carson Range, which rises to an elevation of 9214 ft, has substantially more precipitation than the lower mountains. Vegetation in this range above 6000 ft consists of a variety of pine, fir, cedar trees, manzanita shrubs, and a large variety of smaller plants.

During the period of 1860-1875, timber was removed from the Carson Range for the Comstock Mines. From 1885 into the twentieth century, uncontrolled migratory sheep herds stripped the land of grasses and shrubs, as in Ash Canyon at 8500 feet. Numerous forest fires, such as those on July 22,

1926, and September 28, 1926, have altered the natural vegetation (Kirkham, 1975). The upper elevations of Ash Canyon are subject to extreme erosion hazard; consequently, from 8500 ft to 9000 ft, the entire slope has been terraced. Vegetation in lower Ash Canyon has been burned away by a recent fire. All of these alterations by man have not only altered vegetation, but have changed the natural infiltration and runoff cycle in recharge areas.

#### 2.4 GEOLOGY

Eagle Valley consists of a sediment-filled valley surrounded by block-faulted mountains. Eagle Valley is located in the western extremity of the Basin and Range Physiographic Province, an area characterized by east-west extensional forces, which produced north-south trending block-faulted mountain ranges and valleys.

Geologic mapping in and around Eagle Valley has been done due to the proximity to the Comstock mining district at Virginia City, Nevada (Trexler, Koenig, Flynn, and Bruce, 1980). The area was first studied by Reid (1911), followed by Zones (1958), Eisinger (1960), and an extensive report by Moore (1969). Geologic maps at a scale of 1:24000 (7 1/2-minute ) have been recently completed by the Nevada Bureau of Mines and Geology: Carson City quadrangle (Trexler, 1977), New Empire quadrangle (Bingler, 1977), and Genoa quadrangle (Pease, 1980). ~~Environmental geology~~ and earthquake

studies have recently been done in Eagle Valley (Kirkham, 1976; McKinney, 1976) due to the potential for moderate seismic hazards.

The entire area of Eagle Valley is underlain by rocks associated with the Mesozoic granodiorite batholith. There are no Precambrian or Paleozoic rocks exposed; the oldest exposures are Jurassic-Triassic metamorphic rocks, which appear as roof pendants in the granodiorite. Post-Mesozoic deposits include a series of Tertiary volcanics that range in composition from andesite to rhyolite. Most of the Quaternary formations consist of fluvial and alluvial deposits with the exception of a very recent basaltic-andesite volcanic vent and its associated flows which are exposed in the Virginia Range (Trexler and others, 1980).

The Carson Range is predominantly Cretaceous, consisting of grayish-white granodiorite with a large roof pendant of Triassic andesite. This andesite formation crops out in a mile-wide, east-west-trending band between Clear Creek and Ash Canyon Creek. The Carson Range is offset from the valley floor by a series of north-northeast-trending normal faults at lower elevations.

The Virginia Range, north of Carson City, is predominantly Cretaceous granodiorite with some Jurassic-Triassic volcanics and Quaternary volcanics. Faults in this range are predominantly northeast-trending high angle normal

faults. The Pine Nut Mountains consist of Cretaceous granodiorite and Quaternary floodplain deposits at lower elevations. At higher elevations, there are outcrops of Jurassic-Triassic brecciated andesite, Triassic calcite marble, and Tertiary andesite. Faults in the Pine Nut Mountains generally trend north-northeast (Bingler, 1977).

Approximately thirty-five known or inferred faults displace alluvial sediments in Eagle Valley and one fault displaces bedrock at the surface. About 80% of all faults trend north to northeast and appear to be high angle (50-75°) normal faults (Kirkham, 1976). Faults appear to control the location of hot springs in Eagle Valley (Trexler and others, 1980).

## 2.5 HYDROLOGY

### 2.5.1 Physical Hydrology

The first extensive hydrologic appraisal of the Eagle Valley groundwater system was the Water Resources-Reconnaissance Series Report 39 (Worts and Malmberg, 1966). Other published information about the Eagle Valley groundwater system include Arteaga and Durbin (1978), the Carson City quadrangle groundwater map (Katzner, 1980), and assessment of geothermal resources (Trexler and others, 1980).

The groundwater system seems to be best defined as a single large unconfined aquifer, since it has been

impossible to define separate aquifers. This is due to the alluvium consisting of noncontinuous interbedded coarse and fine material, with no major beds of coarse material. The alluvial basin in Eagle Valley is about six miles long, three to five miles wide, and has a surface area of about 13,000 acres (Worts and Malmberg, 1966). The alluvium ranges in thickness from less than 100 ft to more than 800 ft.

Natural recharge occurs from the Carson Range to the west, Virginia Range to the north, and the Pine Nut Mountains to the east. The piezometric elevation map (Worts and Malmberg, 1966) indicates flow is generally east across the valley, so the predominant recharge area appears to be the Carson Range. Worts and Malmberg estimate 95% of recharge is from the Carson Range. This recharge occurs via a combination of streamflow infiltration, mountain front recharge, and deep percolation. All previous studies have indicated that streamflow infiltration is by far the predominant mechanism for recharge; direct deep percolation of precipitation is considered negligible. Worts and Malmberg (1966) estimated that 8,700 acre-feet is annually recharged (15% of the total precipitation), whereas Arteaga and Durbin (1978) estimated 1,200 acre-feet/year of natural recharge; their figure is lower due to diversions from streams for municipal and agricultural use.

Other sources of groundwater recharge which have altered the natural recharge system include infiltration from irrigation, septic tank effluent, and land application of sewage effluent. The Carson City Sewage Treatment Plant (STP) currently uses land application as part of its sewage treatment process. Irrigation areas in the valley have been delineated mostly in the Kings Canyon drainage area. Groundwater recharge from municipal and agricultural use has been estimated at 4,400 acre-ft/year (Arteaga and Durbin, 1978). One agricultural area is located between Prison Hill and the Carson River, which receives irrigation water from the Carson River via Mexican ditch. Between agricultural recharge in this area and recharge from the Carson River at certain times in the year, it is possible that an appreciable amount of water infiltrates into the groundwater system in this area, mixing with mountain-recharged water.

There is natural discharge from the aquifer into the Carson River at two locations: near Empire and south of Prison Hill. The combined discharge from the groundwater system at these two areas has been estimated at 2,000 acre-ft/year (Worts and Malmberg, 1966), and 2,700 acre-ft/year (Arteaga and Durbin, 1978), both based on differences in flow from stream gaging surveys.

Total average yield from surface water and groundwater reserves has been estimated at 10,000 acre-ft/year (Worts

and Malmberg, 1966), and 9,000 acre-ft/year (Arteaga and Durbin, 1978). However, this yield comes at the cost of reaching a new piezometric surface equilibrium, or in other words, the depth to groundwater will be much greater. Comparison of 1966 and 1980 depth to groundwater maps (Worts and Malmberg, 1966; Katzer, 1980) indicates a net decline valley-wide of 10 to 20 ft.

The valley alluvium appears to be composed of noncontinuous interbedded gravel, silt, and clay beds, and as a result, average hydraulic conductivities are not very large. Hydraulic conductivities for deep wells have been calculated at  $6 \times 10^{-5}$  ft/sec to  $1.6 \times 10^{-4}$  ft/sec (Worts and Malmberg, 1966) which is approximately the average hydraulic conductivity of silty sand (Freeze and Cherry, 1979). Different hydraulic conductivity zones have been identified in the valley alluvium (Katzer, 1980), with the highest hydraulic conductivity zone located in a square mile area at the mouth of Ash Canyon and Vicee Canyons (north of Kings Canyon). Most of the valley alluvium is classified as having lower hydraulic conductivity, and the northeast portion of the alluvium is classified as having fair hydraulic conductivity.

Faults appear to alter the flow of non-thermal water in the aquifer in the north-central part of the valley, as evidenced by abrupt changes in the depth to groundwater map

(Worts and Malmberg, 1966). High angle large displacement normal faults appear to be the main structural control of thermal water migration and resource localization. Three thermal areas are recognized in Eagle Valley: Carson Hot Springs, in the north-central part of the valley; the sewage treatment plant (STP) well - maximum security prison hot springs (Prison Hot Springs) area at the north end of Prison Hill, and the Pinyon Hills area, at the eastern edge of the valley. Carson Hot Springs has several small faults located near the spring outlet, although none is aligned with the point of discharge. A normal fault northwest of Prison Hill appears to control Prison Hot Springs, and the nearby warm well at the sewage treatment plant. A fault mapped 0.6 miles north of the Pinyon Hills thermal area is inferred to extend south through the thermal area (Trexler and others, 1980). A shallow temperature survey in the Pinyon Hills area (Trexler and others, 1980) indicates a steep temperature gradient and elongated pattern, implying fault-controlled geothermal fluid migration.

#### 2.5.2 Water Chemistry

The first study involving a water chemistry analysis of the Eagle Valley aquifer (Worts and Malmberg, 1966) was done for the purpose of identifying potential water-quality related problems. A total of 22 well waters were sampled and analyzed for major ions; all wells were of good chemical



quality. Data from Worts and Malmberg (1966) and Trexler, and others, (1980) are used in this study. An additional twenty analyses from the Desert Research Institute Water Analysis Data System (WADS) are also used in this study.

When the major ion chemistry of thermal and non-thermal wells in Eagle Valley and Carson Valley are plotted on a Piper diagram (Piper, 1944), non-thermal waters show that carbonate is the predominant anion. The cation composition of non-thermal waters averages 40% calcium, 40% sodium, and 10% magnesium.

Thermal waters have been classified into two groups based on the major ions present: sodium-sulfate enriched waters of the Carson Hot Springs and Prison Hot Springs area; and the calcium-sulfate enriched waters of the Pinyon Hills area. There seems to be evidence of mixing of thermal and non-thermal waters, since there are wells of compositions between the non-thermal water composition endpoint and thermal water composition. Sodium-sulfate rich fluids are the dominant thermal fluid composition in a region that extends north from Eagle Valley south to Tonopah. Calcium-sulfate rich waters occur equally as widespread in other nearby thermal areas. The chemical similarities of these water types suggest that Triassic and Jurassic deposits of gypsum and anhydrite which are present in the Pinyon Hills area may be the source of the calcium and sulfate.

Only one previous study in Eagle Valley has made use of stable isotope chemistry (Trexler and others, 1980). Conclusions reached were: thermal waters are isotopically lighter than non-thermal waters with respect to deuterium; and thermal and non-thermal waters are meteoric.

Geothermometry has been used to estimate maximum fluid temperatures at Carson Hot Springs (Trexler and others, 1980). This spring has a surface temperature of 50°C; chalcedony and  $\text{Na}/\text{K}^2$  geothermometry produces estimates of 79° and 75°C, respectively, for the maximum fluid temperature.

## CHAPTER 3

## THEORY OF HYDROGEOCHEMICAL INVESTIGATIONS

## 3.1 MAJOR CONSTITUENT CHEMISTRY

Analyses of the major cations and anions in water from different parts of the flow system can be used to estimate several aquifer parameters such as flow direction, relative flow velocity, and percent evaporation before recharge.

Estimation of the percent of precipitation that infiltrates (or percent evaporation) can be calculated by collecting samples of precipitation and soil water, and analyzing for a major ion or electrical conductivity. As precipitation slowly evaporates, the ions left behind increase in concentration. Using the ratio of:

$$\frac{\text{Ion Concentration (Precipitation)}}{\text{Ion Concentration (Infiltration)}} = \% \text{ Soil Recharge}$$

one can calculate the percentage of precipitation left that has infiltrated, assuming there is no surface runoff. Note that this only represents recharge into the soil zone at a particular elevation and does not represent recharge to the aquifer. Since many cations such as sodium and calcium can be exchanged for other cations in the soil, anions are considered more accurate for recharge calculations. Both chloride and electrical conductivity analyses were used in

this study. Percent evaporation was calculated at different elevations by sampling at a range of elevations. Evaporation percentages obtained were also used in this study to compare to evaporation percentages obtained from stable isotope analyses.

Chloride and electrical conductivity analyses were used in conjunction with annual precipitation and area data from Eagle Valley for a rough estimate of annual recharge. One assumes with this type of analysis that ion change is not due to other inputs along the flow path.

The Maxey-Eakin method of estimating percent recharge (Watson, Sinclair, and Waggoner, 1976) is similar to the method used because it assumes different percentages of precipitation infiltrate, based on the amount of precipitation. The Maxey - Eakin method (Eakin and others, 1951) was developed by assuming recharge in an undeveloped basin equals the annual discharge. Then, based on a precipitation zone map of Nevada (Hardman, 1936), the percentages of precipitation that recharged were adjusted by trial and error until annual recharge equaled annual discharge. Trial and error adjustment of percent recharge was based on 13 basins in east-central Nevada.

The estimates are as follows: "no significant groundwater recharge is believed to occur in the zones having precipitation of less than 8 inches. In the 8 to 12 inch zone,

the recharge may be about 3 percent of the precipitation; in the 12 to 15 inch zone, about 7 percent; in the 15 to 20 inch zone, about 15 percent; and in the zone having over 20 inches, about 25 percent" (Eakin and others, 1951). The inaccuracy in this method lies in calculating evaporation and transpiration in the valley, the majority of discharge. This method also does not differentiate recharge from precipitation above 20 in; which is most of the Carson Range. In other words the Maxey-Eakin method was developed for arid basins, and the Carson Range of the Sierras is not arid. The Maxey-Eakin method annual recharge estimate for Eagle Valley is compared to the chloride and electrical conductivity recharge estimates from this study in Section 6.3.

Infiltration into the aquifer in Eagle Valley probably occurs mainly via stream channel infiltration. This process can be quantified by stream gaging and sampling at different times of the year. As with infiltration, as the water in a stream evaporates, the flow will decrease along with an increase in conductivity. However, factors such as spring input, groundwater input, or bank storage input will change the conductivity, all which cannot be realistically measured. Practically, one can only make general conclusions based upon flow and conductivity changes.

A decrease in flow over a stream section during summer and winter with a definite increase in conductivity

indicates evaporation, or possibly transpiration. Decrease in flow with no change in conductivity indicates infiltration. Note that a decrease in flow in the summer over a section and no decrease in flow in the winter indicates the flow loss was due to phreatophyte usage in the summer, or the ground was frozen in winter, reducing infiltration. Any groundwater inflow to a stream should have a higher conductivity, since soil water flow will increase ion concentration faster than stream flow.

To some extent, the chemistry of unaltered surface water and groundwater will be characteristic of the geology in a certain area. This is because water will dissolve ions in certain percentages based upon available ions from rocks. The soil zone will supply  $\text{HCO}_3$ , from root respiration, and from atmospheric  $\text{CO}_2$ . Weathering of volcanic rocks produces clays and a certain amount of potassium, sodium, silica, iron, and magnesium. Partial solution of evaporites will supply carbonate, calcium, sodium, sulfate, and chloride. As water travels farther along in the flow system, changes in chemistry will reflect increasing dissolved ions from surrounding rocks.

In his classic paper Chebotarev (1955), based upon the analyses of more than 10,000 chemical analyses of wells, concluded that natural groundwater tended toward the composition of sea water. This is seen in an anion shift from

bicarbonate to sulfate to chloride as the water travels further along the flow path. This shift is accompanied by an increase in electrical conductivity and total dissolved solids. This ion shift reflects: 1) changes in ion availability in different areas; 2) ion solubility differences; and 3) differing reaction rates of ions. Carbonate usually comes from  $\text{CO}_2$  in the soil zone in recharge areas, sulfate and chloride are usually from evaporites, which are found in deep sedimentary basins. From bicarbonate ( $K=10^{-8.3}$ ) to sulfate ( $K=10^{-4}$ ) to chloride ( $K=10^0$ ), solubility increases 10,000-fold for each ion shift. Consequently, as water travels, bicarbonate saturation is reached first, since it is the most prevalent anion at the beginning of the flow path. As the groundwater flow continues, evaporites are encountered, and sulfate becomes the dominant ion, 10,000 times more soluble than bicarbonate. Bicarbonate concentration can still increase, but only to slightly over saturation. Chloride eventually becomes the major anion, once halite beds are reached.

A trilinear diagram of anions, with corners of carbonate, sulfate, and chloride (Piper, 1944), will represent the Chebotarev sequence as a shift from corner to corner, accompanied by a conductivity increase. In a single basin, with groundwater flowing through characteristic rock types and formations, cations will also show a trend.

Cation chemistry of thermal areas can be used empirically to estimate original thermal reservoir temperature. Geothermometry used in this study is based on Na-K-Ca method (Fournier and Truesdell, 1973). Thermal water is less dense than colder water, so equal hydraulic head of a thermal well has a higher pressure head than a cold well with the same hydraulic head. If this head difference is known, one can calculate the depth at which the heads are equal by:

$$[\text{depth}][\text{density}]_{\text{hot}} = [\text{depth} + H][\text{density}]_{\text{cold}}$$

This depth may represent the rock-alluvium contact, where fracture flow thermal waters mix with cold water in the alluvium.

## 3.2 STABLE ISOTOPE CHEMISTRY

### 3.2.1 Introduction

Isotopes can be defined as the same element with differing atomic masses due to a different number of neutrons in the nucleus. There are over 1000 isotopes of the ninety-two naturally occurring elements on earth. Isotopes that have proved useful in various hydrologic, meteorologic and biologic studies include hydrogen ( $^1\text{H}$ ,  $^2\text{H}$ ,  $^3\text{H}$ ), carbon ( $^{12}\text{C}$ ,  $^{13}\text{C}$ ,  $^{14}\text{C}$ ), nitrogen ( $^{14}\text{N}$ ,  $^{15}\text{N}$ ), oxygen ( $^{16}\text{O}$ ,  $^{17}\text{O}$ ,  $^{18}\text{O}$ ), and sulfur ( $^{32}\text{S}$ ,  $^{34}\text{S}$ ). This study makes use of stable isotopes  $^1\text{H}$ ,  $^2\text{H}$ ,  $^{16}\text{O}$ ,  $^{18}\text{O}$ , and  $^{13}\text{C}$ , and radioactive isotopes  $^3\text{H}$ , and  $^{14}\text{C}$ . The present section is devoted entirely to discussion of stable hydrogen and oxygen isotopes; tritium ( $^3\text{H}$ ) will be



discussed in Section 3.3, and carbon isotopes will be discussed in Section 3.4.

Stable hydrogen and oxygen isotopes are particularly useful in hydrologic studies as tracers since the isotopes are part of the water molecule. Therefore, these isotopes are essentially conservative tracers (no hydrodynamic dispersion). Isotopes of hydrogen and oxygen occur naturally in the following proportions:

ISOTOPE		RELATIVE ABUNDANCE		TYPE
		percent	ppm	
<sup>1</sup> H	protium	99.984	999840	stable
<sup>2</sup> H	deuterium	.0154	154	stable
<sup>3</sup> H	tritium	0-10 <sup>-13</sup>	0-10 <sup>-10</sup>	radioactive
<sup>16</sup> O	oxygen	99.76	997600	stable
<sup>17</sup> O	oxygen	.04	400	stable
<sup>18</sup> O	oxygen	.20	2000	stable

(Freeze and Cherry, 1979)

Of the stable isotopes <sup>1</sup>H, <sup>2</sup>H, <sup>16</sup>O, <sup>18</sup>O, there are six combinations of the water molecule:

MOLECULE	ABUNDANCE ppm	ATOMIC WEIGHT
H <sub>2</sub> O <sup>16</sup>	997680	18.
H <sub>2</sub> O <sup>18</sup>	2000	20.
HDO <sup>16</sup>	320	19.
HDO <sup>18</sup>	<1	21.
D <sub>2</sub> O <sup>16</sup>	<1	20.
D <sub>2</sub> O <sup>18</sup>	<1	22.

Only the first three occur naturally in large enough quantities for hydrologic purposes. Which compounds are

present in a sample is inconsequential, since total deuterium and oxygen-18 mass is what is measured.

Concentrations of the isotopes of water are not usually reported in parts per million, but rather as a percent difference of the ratio of heavy to light isotopes compared to a standard. This difference in the isotopic ratios of the sample from the standard is called the  $\delta$ -value and is defined as:

$$\delta^{18}/16 = \frac{^{18}\text{O}_{\text{sam}}/^{16}\text{O}_{\text{sam}}}{^{18}\text{O}_{\text{std}}/^{16}\text{O}_{\text{std}}} - 1 = (R_{\text{ox}} - 1); \quad \delta\text{D}/\text{H} = \frac{\text{D}_{\text{sam}}/\text{H}_{\text{sam}}}{\text{D}_{\text{std}}/\text{H}_{\text{std}}} - 1 = (R_{\text{d}} - 1)$$

where  $R_{\text{ox}}$  and  $R_{\text{d}}$  are referred to as sample isotopic ratio. The  $\delta^{18}/16$  values are most commonly reported in parts per thousand (per mil or ‰) deviation. The  $\delta$ -formula is defined such that the sample with the lighter isotopic content (the most common case) is represented by a negative number, indicating depletion. For example, a sample with  $R = .99$  means  $\delta^{18}/16 = -10\text{‰} = -1\%$ , or 10/1000, or 1/100 less oxygen-18/oxygen-16 than the standard (Muir and Coplen, 1981). The notation that is used in this report,  $\delta^{18}/16\text{‰}$ ,  $\delta\text{D}/\text{H}\text{‰}$ ,  $R_{\text{ox}}$ ,  $R_{\text{d}}$  are defined as:

$$\delta^{18}/16\text{‰} = (R_{\text{ox}} - 1)1000 \quad \delta\text{D}/\text{H}\text{‰} = (R_{\text{d}} - 1)1000$$

Standards used in isotopic studies are defined as follows:

$$\delta^{18}/16 \text{ SMOW} = 0.0\text{‰}, \text{ or } 1993.4 \pm 2.5 \text{ ppm (Craig, 1961a)}$$

$$\delta^{18}/16 \text{ V-SMOW} = 0.0\text{‰} \text{ or } 2005.2 \pm .45 \text{ ppm}$$

$$\delta^{18}/16 \text{ SMOW} = 1.0059 \delta^{18}/16 \text{ V-SMOW}$$

$$\delta D/H \text{ SMOW} = 0.0\% \text{ or } 158 \pm 2 \text{ ppma}$$

$$\delta D/H \text{ V-SMOW} = 0.0\% \text{ or } 155.76 \pm .05 \text{ ppm (Fritz and Fontes, 1980)}$$

$$\delta D/H \text{ SMOW} = 1.0142 \delta D/H \text{ V-SMOW}$$

Other standards used for mass spectrometer calibration are listed in Section 5.2.4.2. The standard SMOW (Standard Mean Ocean Water) has been used in early isotope studies, and V-SMOW (Vienna Standard Mean Ocean Water) is used in more recent studies.

Due to the differing atomic weights of the various isotopic molecules of water, the compounds have slightly different physical properties. The greater the mass of a molecule, the more energy is required to change its phase, and heavier compounds will tend to remain in a lower energy state. In addition, the heavier the molecule, the higher the boiling point (since more energy is required). In liquid form, heavier water molecules ( $\text{HDO}^{16}, \text{H}_2\text{O}^{18}$ ) will have a lower vapor pressure (amount of a substance in gas form) than light water ( $\text{H}_2\text{O}^{16}$ ), since heavier molecules will tend to remain in the liquid state.

Changes in the  $\delta D/H$  or  $\delta^{18}/16$  ratios that occur from phase changes are called isotopic fractionation, and are the basis for use of water isotopes to trace the origin and flow of water. The assumption made in utilizing water isotopes as natural tracers of regional groundwater flow is that the

isotopic content of water once recharged remains relatively constant during the residence time of water in the aquifer. Fractionation occurs along all stages of the hydrologic cycle from evaporation of ocean water, to atmospheric transport of moisture, to precipitation, and flow back to the ocean. Isotopic fractionation and change can occur as the result of:

- (1) Original evaporation from the ocean, where the amount of fractionation is dependent upon the temperature;
- (2) Mixing along the path a cloud takes before precipitation;
- (3) Amount of precipitation along cloud path;
- (4) Temperature of final precipitation in study area;
- (5) Exchange and evaporation between precipitation and moisture in the air;
- (6) Evaporation of precipitation from the ground before infiltration;
- (7) Mixing of different waters in the groundwater reservoir; and
- (8) Exchange and ultrafiltration of groundwater with geologic strata (Simpson, Thorud and Friedman, 1970; Gat, 1971).

A great deal of literature exists discussing fractionation at different stages in the hydrologic cycle. This discussion of isotopic fractionation theory involves only those aspects directly applicable to the study of a groundwater flow cycle. The following sections are divided on the basis of each successive stage of water transport from the ocean

to water in the aquifer. These sections are: Section 3.2.2 - Stable Isotopes during Evaporation and Vapor Transport, Section 3.2.3 - Stable Isotopes during Condensation, Section 3.2.4 - Stable Isotopes during Infiltration, Section 3.2.5 - Stable Isotopes in Trees, Section 3.2.6 - Stable Isotopes in Groundwater Flow.

### 3.2.2 Stable Isotopes during Evaporation and Vapor Transport

Stable isotope analyses are reported as deviation from water samples (SMOW) originally taken from the ocean at a depth of 1600 to 6500 ft (Craig, 1961b). An ocean standard is used since its isotopic content remains fairly constant over time at depth. Studies of cores of ice from the north and south pole indicate that  $\delta^{18}/^{16}$  variation has been less than 1.2‰ during the past 18,000 years (Winograd and Friedman, 1972), and  $\delta D/H$  variation of less than 2‰ in the last 9,000 years (Simpson, Thorud, and Friedman, 1970), so the oceans represent a source of essentially constant isotopic water. However, there is a very regular trend of increasing heavy isotopic content of  $\delta D/H$  in the tropical zone, between 30° north and south latitude (Friedman, Redfield, Schoen, and Harris, 1964), due to increased evaporation which makes the remaining water isotopically heavier (Bostick, 1978).

When water evaporates from the ocean, the vapor is isotopically lighter than the remaining water. This difference

in isotopic content, the amount of fractionation, is dependent on temperature, and is described by the use of  $\alpha$ , the fractionation factor. This is defined by:

$$N_v(H_2O) = N_l P(H_2O) \quad N_v(HDO) = N_l P(HDO) \quad \text{Henry's Law:}$$

$$\frac{N_v(HDO)}{N_v(H_2O)} = \frac{N_l P(HDO)}{N_l P(H_2O)} \quad \begin{array}{l} N_v = \text{mole fraction in vapor} \\ N_l = \text{mole fraction in liquid} \\ P_l = \text{vapor pressure or} \\ \text{partial pressure} \end{array}$$

$$R_v = R_l \cdot \frac{1}{\alpha_d} \quad R = \text{isotopic ratio: } \frac{\text{heavy component}}{\text{light component}}$$

$$\alpha_d = \frac{R_{\text{liquid}}}{R_{\text{vapor}}} = (\text{by definition}) \frac{v_p(H_2O)}{v_p(HDO)} \quad \alpha_d = \frac{P_{H_2O}}{P_{HDO}} = \frac{\text{light component}}{\text{heavy component}}$$

$\alpha_{ox}$  is derived by the same method

Assuming equilibrium evaporation at 20°C,  $\alpha_d = 1.08$ , and  $\alpha_{ox} = 1.009$ , which indicate that the vapor will have 8% more hydrogen and 1% more oxygen-16 (light components) as compared to the source, ocean water. Conversely, the vapor will have  $1/\alpha - 1$  less of the heavy components. At -20°C,  $\alpha_d = 1.15$ ,  $\alpha_{ox} = 1.014$ , which again indicates the vapor has more light components and less of the heavy isotopes (D,  $^{18}\text{O}$ ) than the remaining water. This also shows that the amount of fractionation is greater at lower temperature. In other words, as the temperature drops, the fractionation factor increases, and the lesser amount of heavy isotopes evaporate and condense;  $\delta D/H$  and  $\delta 18/16$  become more negative (depleted). The fractionation factor change with temperature is illustrated in Figure 3.1. The non-continuous fractionation on the graphs of  $\alpha_d$  and  $\alpha_{ox}$  (the break in each curve) is due to the latent heat of fusion at the phase change between liquid and solid; i.e. the energy difference between evaporating vapor from a liquid at 0°C versus evaporating vapor from a solid at 0°C.

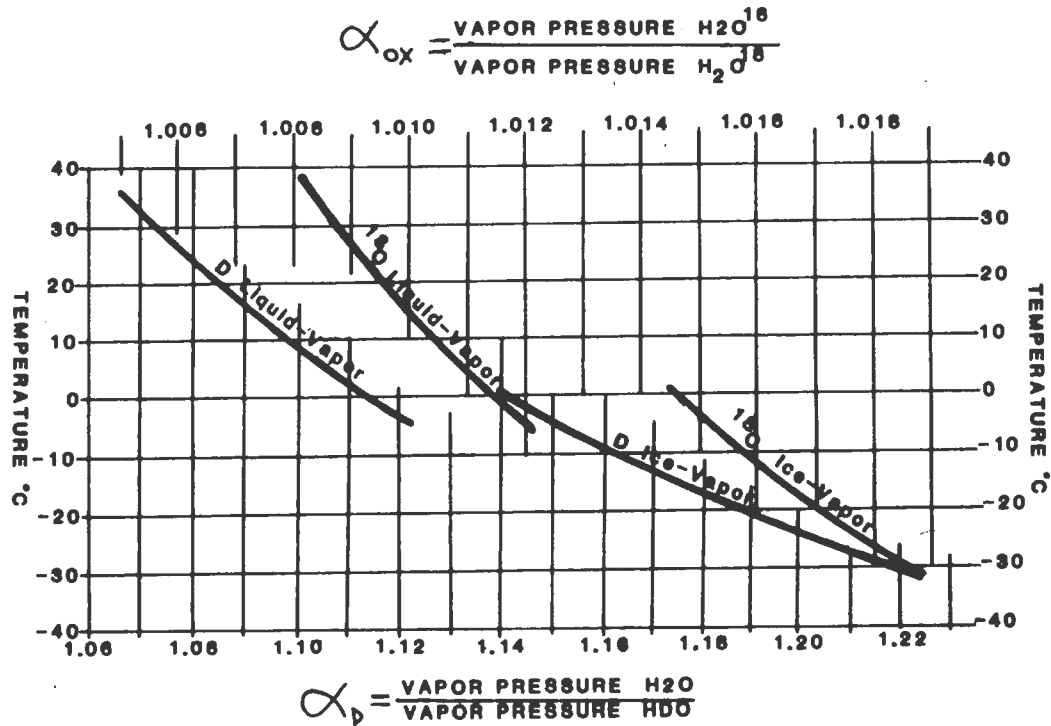


Figure 3.1 Hydrogen and Oxygen Isotope Fractionation under Equilibrium Conditions (modified after Fritz and Fontes, 1980)

This temperature effect, also known as the kinetic effect, is more understandable if one considers the cell cluster model of evaporation. According to this theory, at high temperature, massive amounts of water evaporate as clusters of molecules. For example, if six water molecules evaporate as a cluster, and the first cluster containing all  $\text{H}_2\text{O}$ , the second containing one molecule of  $\text{HDO}$ , and the third containing one molecule of  $\text{H}_2\text{O}^{18}$ , the atomic weights of these clusters would be 108, 109, and 110, respectively. Since the difference between molecule cluster masses is small, fractionation is small (Dansgaard, 1964). At lower temperature, molecules evaporate individually, so the difference in masses of  $\text{H}_2\text{O}$ ,  $\text{HDO}$ , and  $\text{H}_2\text{O}^{18}$  (mass = 18, 19, 20) is much more significant than at high temperature. Put simply, at high temperatures, so much mass evaporates violently, the process is not as selective as at low temperature. In fact,  $\alpha_d = 1.000$  at  $220^\circ\text{C}$ .

The calculation of fractionation factors ( $\alpha_d, \alpha_{18}$ ) is based on the assumption of equilibrium fractionation, which usually exists only when the process (such as evaporation or condensation) is slow. Nonequilibrium fractionation exists when the process is fast (Dansgaard, 1964). Observations and experiments indicate that evaporation in nature is mainly nonequilibrium (Friedman, Redfield, Schoen and Harris, 1964), but fractionation during condensation occurs under equilibrium conditions (Dansgaard, 1964).



Both evaporation and condensation in nature have been approximated by assuming Rayleigh distillation conditions, which assumes a continuous removal of the reaction products. For example, this corresponds to removal of rain from a cloud after condensation takes place. The Rayleigh process leads to higher fractionation than if the two phases were allowed to equilibrate (Dansgaard, 1964). Evaporation under equilibrium conditions, assuming Rayleigh distillation can be described by the following relationship:

$$\frac{R}{R_0} = \theta^{\frac{1}{\alpha} - 1}$$

Condensation under equilibrium conditions can be

described by:

$$\frac{R}{R_0} = \theta^{\alpha - 1}$$

R = Relative Isotopic Ratio of Liquid  
 R<sub>0</sub> = Initial Isotopic Ratio of Liquid  
 θ = Fraction of Liquid Remaining  
 α = Fractionation Factor

R = Relative Isotopic Ratio of Vapor  
 R<sub>0</sub> = Initial Isotopic Ratio of Vapor  
 θ = Fraction of Vapor Remaining  
 α = Fractionation Factor

Once vapor leaves the ocean and moves inland, a change in the isotopic content of the vapor known as the continental effect occurs because of condensation and precipitation. This consists of the cloud vapor becoming progressively depleted isotopically (more negative δD/H and δ18/16) as the cloud moves further from its source. Since the precipitation represents a lower energy state with respect to the vapor, it will remove slightly more heavy isotopes than light isotopes, which leaves the cloud slightly depleted. As this process is repeated, the cloud becomes progressively more depleted. The amount of depletion depends on both the amount of vapor condensing and the temperature of condensation, which changes the fractionation factors.

In conclusion, the mass differences between  $H_2O$ , HDO, and  $H_2O^{18}$  cause fractionation during phase changes which can be generalized as follows: 1) heavier molecules will tend to remain in or be transferred into a lower energy state; 2) a phase change to a higher energy state will cause a depletion in heavy isotopes; and 3) as the temperature decreases,  $\alpha$  increases, precipitating more light isotopes and less heavy isotopes. In other words, condensation at a cooler temperature from a source will be more depleted isotopically than condensation from the same source at a warmer temperature. This is discussed in greater detail in the following section.

### 3.2.3 Stable Isotopes during Condensation

Given that under equilibrium conditions at  $20^\circ C$ ,  $\alpha_w=1.00$ ,  $\alpha_d=1.08$ ,  $\alpha_{18}=1.009$ , one can set up a ratio of fractionation factor changes and mass changes:

$$\frac{\alpha_d - \alpha_w}{\alpha_{18} - \alpha_w} \cdot \frac{\Delta \text{mass (D-H)}/H}{\Delta \text{mass } [^{18-16}]/_E} \quad 8.9:8.0$$

which states that hydrogen isotope mass change is eight times that of oxygen isotope mass  $\Delta$ , and the  $\alpha_d$  change is about nine times the  $\alpha_w$  change, implying that deuterium fractionation should be about eight times greater than oxygen fractionation. This fact is reflected in the meteoric water line:

$$\delta D/H = 8(\delta 18/16) + 10 \quad (\text{Craig, 1961a})$$

which was obtained from over 400 worldwide precipitation

samples. Condensation of water from clouds is brought about by cooling. As the air cools, it reaches saturation and precipitation occurs. Fractionation of isotopes during condensation is not completely understood (Friedman, Redfield, Schoen, and Harris 1964), but if one does assume Rayleigh distillation conditions, condensation occurs by (as defined in Section 3.2.2.):

$$\frac{R}{R_0} = \theta^{\alpha-1}; \text{ substituting } R = \frac{\delta\text{‰} + 1000}{1000}; \frac{\delta\text{‰} + 1000}{\delta_0\text{‰} + 1000} = \theta^{\alpha-1}$$

$\theta$ , the fraction of vapor remaining is determined by assuming saturated vapor at mean sea level (MSL) rising pseudo-adiabatically, for saturated air. This causes cooling at the rate of  $-1.8^\circ\text{C}/1000$  ft which produces supersaturated vapor and some condensation. Thus,  $\theta$  decreases from 1.0 to 0.0 as altitude increases. Using the appropriate fractionation factors  $\alpha_l$  and  $\alpha_{ax}$  in the Rayleigh condensation formula based on an altitude versus temperature relationship developed for the Sierra, and assuming ocean vapor was originally at  $+10^\circ\text{C}$  and  $\delta\text{D}/\text{H} = -100\text{‰}$  (Smith, Friedman, Klieforth and Hardcastle, 1979), Figure 3.2 is calculated. The diagrams are calculated to an elevation of 18,000 ft, although in other mountain ranges radar echos off snow have been reported up to 35,000 ft ( $-50^\circ\text{C}$ ) (Friedman, Redfield, Schoen and Harris, 1964).

These diagrams indicate that enrichment of heavy isotopes occurs when vapor is partially condensed into liquid. For example, using the deuterium diagram, vapor at an

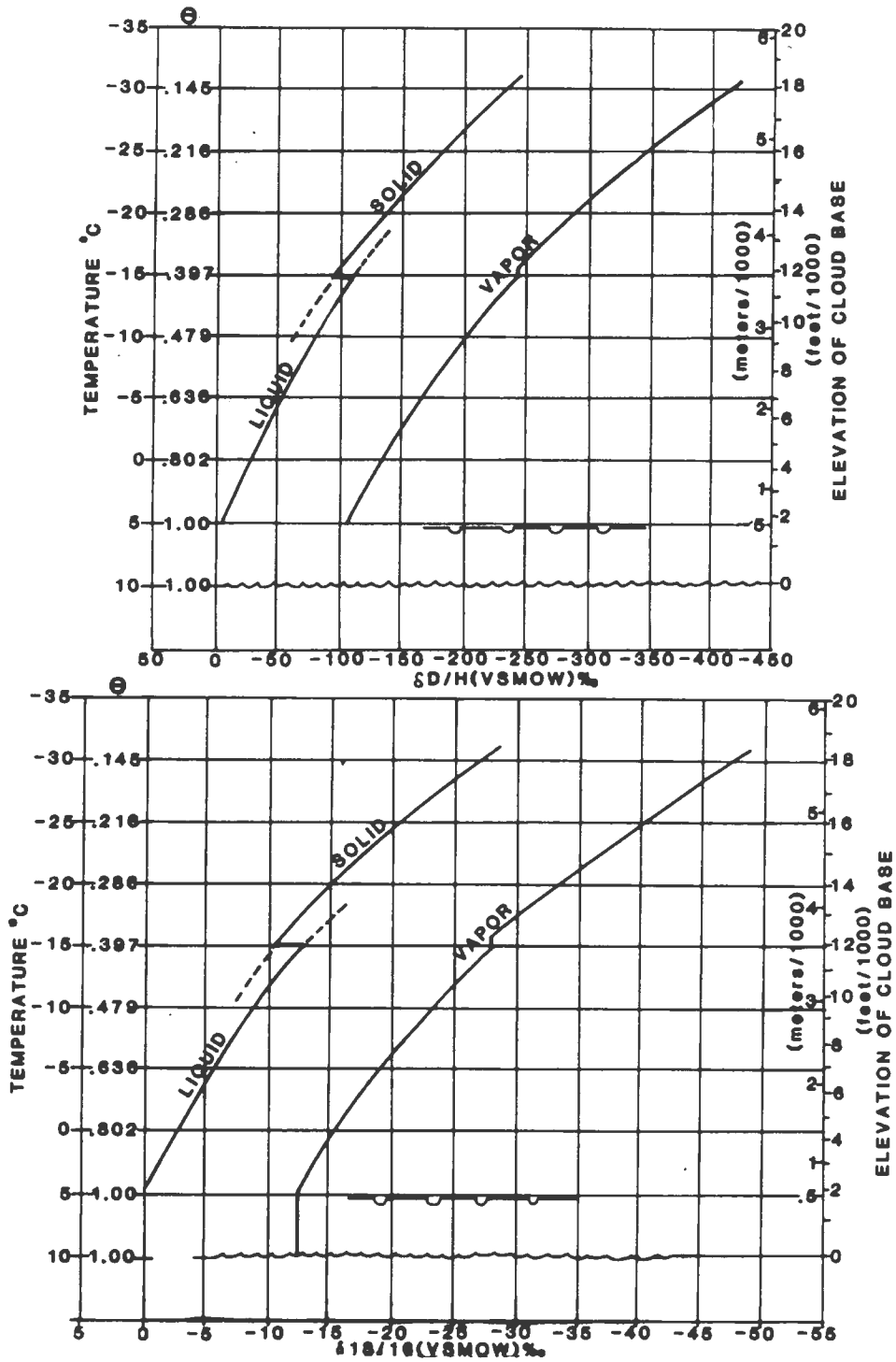


Figure 3.2 Calculated Isotopic Composition of Vapor and Precipitation in the Sierra Nevada Mountains

elevation of 13,000 ft (4000 ft cloud base above a 9000 ft mountain) would have a composition of  $\delta D/H = -270\%$ , but snow condensing from that vapor would have a composition of  $\delta D/H = -120\%$ . This Rayleigh process illustrates the continental effect, since the vapor is getting progressively lighter as the cloud goes higher over mountains, due to enriched condensation that is removed from the system by precipitation. This system, however, does not take into account evaporation off the land back into the cloud. Most observations, however, are close to that predicted by Rayleigh distillation (Dansgaard, 1964).

West Slope of the Sierra Nevada:	$\delta D/H = -12.0\% / 1000 \text{ ft}$
East Slope of the Sierra Nevada:	$\delta D/H = -3.1\% / 1000 \text{ ft}$
Mt. Rose, East Slope:	$\delta D/H = -2.1\% / 1000 \text{ ft}$
Swiss Alps:	$\delta D/H = -15.0\% / 1000 \text{ ft}$
Himalayas (16,000 to 22,000 ft):	$\delta D/H = -40.0\% / 1000 \text{ ft}$
(Smith, Friedman, Klieforth and Hardcastle, 1979)	
Southern Sierra Nevada:	$\delta D/H = -6.0\% / 1000 \text{ ft}$
(Winograd and Friedman, 1972)	
Italy (Fritz and Fontes, 1980):	$\delta^{18}/^{16} = -1.6\% / 1000 \text{ ft}$
Chile (Fritz and others, 1981):	$\delta^{18}/^{16} = -3.0\% / 1000 \text{ ft}$
Cordillera Mtns (Payne, 1972):	$\delta^{18}/^{16} = -0.8\% / 1000 \text{ ft}$

These show greater fractionation (more negative slopes) in cooler areas (higher elevation mountains). Previous studies (Friedman and Smith, 1970; Smith and others, 1979) concluded that there is no regular isotopic trend on the east side of the Sierra Nevada.

The Rayleigh process explains overall gross isotopic variation of precipitation over a large area, but cannot explain local variation on a single mountain slope. On a

local scale, snow falling from a single cloud can have a different isotopic composition at different elevations. Isotopic variances with storm intensity and duration have also been recorded from a single source at the same elevation (Judy and Meiman, 1970). This process is not well understood, but one theory suggests that evaporation of precipitation as it falls could account for this variation. Lower elevations, being warmer and lower, have a longer fall path and greater amount of evaporation potential (due to the temperature) as compared to a higher elevation. Therefore, more precipitation will evaporate falling to a lower elevation and less will evaporate falling to a higher elevation. This generalization of less evaporation at higher elevations is observed in snow falling with greater intensity at higher elevations at the same time.

Due to storm source variation, storm path variation, temperature of condensation, and numerous other factors, there is an extreme variation between storms at a single elevation. One study (Judy and Meiman, 1970) observed variations of  $\delta D/H = -229\%$  to  $-106\%$  from different storms at a single station. This variance is homogenized as the snow melts and eventually infiltrates. To get the most appropriate average isotopic content of precipitation at an elevation, it is best to core snow in late spring.

Although individual isotopic variation between storms

is generally not sampled, summer and winter storms, if sampled separately, show an isotopic difference. Since condensation occurs at a warmer temperature in summer precipitation, they are observed as isotopically heavier (Bostick, 1978) than winter precipitation and can be used in groundwater studies to determine if recharge is mainly from winter or summer precipitation.

### 3.2.4 Stable Isotopes during Infiltration

Once precipitation has fallen, only a fraction infiltrates due to evaporation in the air and on the ground. The amount of evaporation, or conversely, the amount of infiltration can be extremely important in arid regions, and can be estimated with ions changes (see Section 3.1). Evaporation can also be estimated with isotopes using the Rayleigh distillation formula for evaporation as previously defined:

$$(1) \frac{R_i}{R_s} = \theta^{\alpha} - 1$$

$R = R_i$  = isotopic ratio of phase remaining (infiltrated water)  
 $R_s = R_o$  = original isotopic ratio of precipitation or snow  
 $\theta$  = fraction of water remaining from precipitation to infiltration

Combining this with a mass balance equation:

$$(2) R_s Q_s = R_i Q_i + R_v Q_v$$

$R$  = isotopic ratio  
 $Q$  = volume:  
 $s$  = snow or precipitation  
 $i$  = infiltrated water  
 $v$  = vapor

one arrives at an equation containing the isotopic ratio of snow,  $\delta_s$  ‰, isotopic ratio of infiltrated water,  $\delta_i$  ‰ (known), and the fraction infiltrated,  $\theta$ :

$$(3) \frac{\delta_i \text{‰} + 1000}{\delta_s \text{‰} + 1000} = \theta \left[ \frac{1 - \theta \left( \frac{\delta_i \text{‰} + 1000}{\delta_s \text{‰} + 1000} \right)}{1 - \theta} - 1 \right]$$

The fraction infiltrated can only be solved for by an iterative process. The complete derivation of Equation 3 is

found in Appendix 6. This equation is valid only when a phase change in the source does not occur. In other words, evaporation from snow or evaporation from rain can be estimated with this equation, but not evaporation from snow which then melts and continues to evaporate. This is because the phase change from solid to liquid changes the isotopic fractionation and this is not accounted for in the equation. As with chloride ion, this equation can only estimate how much evaporation occurred between two samples (or estimate fraction remaining). This "local recharge" into the soil zone does not necessarily become groundwater recharge.

When considering both  $\delta D/H$  and  $\delta 18/16$  in evaporation, the isotopic composition of the remaining liquid will become heavier as more heavy isotopes will remain in that phase. This corresponds to increasing  $\delta D/H$  and  $\delta 18/16$  content.

Under equilibrium conditions, evaporation will occur with deuterium fractionating eight times faster than oxygen, i.e., on a  $\delta D/H$  versus  $\delta 18/16$  graph, a slope of 8 is produced (Dansgaard, 1964). This is rarely found in nature, indicating that most evaporation is through non-equilibrium conditions.

Non-equilibrium fractionation is not very well understood, but one conclusion based on laboratory experiments and some field data (Dansgaard, 1964) concludes that the faster the evaporation from a limited source, the smaller



the slope ( $\Delta\delta D/H/\Delta\delta^{18}O/16$ ). This is because deuterium seems to be much less sensitive to the kinetic effects (high evaporation) than oxygen isotopes. The oxygen isotope ratio will change quickly in fast evaporation. Therefore, the slope can be used as a rate of evaporation indicator. The slope under fast evaporation at 20 °C equals 3.2 in lab experiments. Field observations of evaporation (Muir and Coplen, 1981; Edmunds and Wright, 1979; Gat, 1971) yielded slopes of 3.0, 3.3, and 4.5. Usually the faster the evaporation, the higher the temperature at which the evaporation occurred.

Several factors must be taken into account when estimating evaporation at different elevations. Low elevations will have fast evaporation or small slopes. Also, since the temperature is higher and there is little precipitation, snow will evaporate quickly. High elevations have slow evaporation, approaching equilibrium evaporation, so the slopes are less than but approaching 8. In addition, evaporation occurs slowly for a long period of time. The result is more fractionation will occur at higher elevations since evaporation occurs over a much longer period of time, and the temperature is cooler (fractionation factor is larger at a lower temperature). This is in agreement with the cell cluster theory discussed in Section 3.2.2 in which it was concluded that lower temperature evaporation produced more fractionation.

Fractionation of precipitation in recharge areas occurs continuously until recharge. Infiltration of part of the precipitation greatly reduces the variance of individual snow storms (Judy and Meiman, 1970). Therefore, measurement of infiltrated water serves two purposes: calculation of percent recharge; and smoothing of input precipitation isotopic composition (Gat, 1971).

### 3.2.5 Stable Isotopes in Trees

In an attempt at collecting a several year average of infiltrated water, samples of water from parts of a tree were sampled in this study. In a study of isotopic fractionation between soil water and tree water, it was concluded that roots do not appear to fractionate water (Ziegler, Osmond, Stichler and Trimborn, 1976). Once in the roots, water can be transferred into several different parts of the tree structure which may or may not cause isotopic fractionation.

The parts of a tree from the bark to the center are: the bark or cork, phloem, new and old xylem, and the pith. The phloem is responsible for making the bark and transporting resins down the tree. The xylem transports water up the tree. At any given time, only a thin rind of young phloem and young xylem is functional. Older xylem near the center of the trunk gradually blocks up with resins and gum, and water conduction is no longer possible (Weisz, 1971). Such central regions are called heartwood. One can analyze the

isotopic ratio of either trapped water or the cellulose water of any part of the tree.

Previous studies of tree ring cellulose (xylem) water (Burk and Stuiver, 1981) have used oxygen isotope changes as an indicator of past climates. The oxygen-18 in organic cellulose molecules appears to come from dissolved  $\text{CO}_2$  in soil water and from atmospheric  $\text{CO}_2$ . There is a high correlation between  $\delta^{18}/16$  in cellulose and mean annual temperature (Gray and Thompson, 1977). Another study (Epstein, Yapp, Crayton and Hall, 1976) concluded that cellulose  $\delta^{18}/16$  in land plants is poorly correlated to  $\delta^{18}/16$  in environmental water. Since some of the oxygen in terrestrial plant cellulose is from atmospheric  $\text{CO}_2$ , and atmospheric  $\text{CO}_2$  has an isotopic composition of  $\delta^{18}/16 = +41\%$  (Epstein, Thompson and Yapp, 1977), mixing of soil water  $\text{CO}_2$  and atmospheric  $\text{CO}_2$  would effectively enrich the  $\delta^{18}/16$  value of the soil water  $\text{CO}_2$ . Therefore, cellulose  $\delta^{18}/16$  cannot be used as an indicator of the isotopic composition of soil water.

In this study, it was decided that sampling water from the water-stabilized portion (old xylem) of a tree would be the best indicator of soil water isotopic composition. In a previous study (Wershaw, Friedman, Heller, and Frank 1966), water from leaves, xylem and phloem was analyzed for deuterium and compared to nearby pond water. Analyses

indicated that for an average of nine samples:  $\delta D/H = -103\%$  for pond water;  $\delta D/H = -94\%$  for xylem;  $\delta D/H = -94\%$  for phloem; and  $\delta D/H = -34\%$  for leaf water. This was interpreted as indicating very little fractionation between recharge water and water in the tree, but considerable enrichment was apparent in the leaves due to evaporation. The trees sampled were phreatophytes (cottonwood and willow) in which transpiration was estimated at about 65 gallons/day. There seemed to be considerable migration of water between phloem and xylem, which accounts for their similarity. The authors concluded that the trees sampled contained no more than a two to three year record of soil water in the xylem due to the water migration.

Water from trees which are not phreatophytes has not been previously studied, nor has this technique been applied on a regional scale as is done in this study.

### 3.2.6 Stable Isotopes in Groundwater Flow

Once water precipitates in recharge areas, it eventually reaches the aquifer through a variety of flow paths: infiltration, then seepage into a stream, more evaporation, and eventual infiltration in the valley alluvium; infiltration then deep underflow to the aquifer; and surface runoff to stream channels and infiltration in the valley aquifer. Recharge into the groundwater system in Eagle Valley probably consists of a combination of all three, which makes it

difficult to quantify evaporation. The process by which water flows into the aquifer tends to homogenize the isotopic variation found in precipitation and shallow soil water. In a study of a small groundwater system in Chile (Fritz, Suzuki, Silva and Salati, 1981), the four-year  $\delta^{18}O/16$  variation from a spring was  $\pm 3\%$ .

Changes which will show up in a regional aquifer are: seasonal (winter and summer precipitation) variations; and any long term change from recharge that occurred in the past when there was a different climate (Gat, 1971), observed in semi-arid and arid areas. The seasonal variance has been observed in runoff from Steamboat Creek (20 mi north of Carson City; White, 1968) where isotope values are enriched in the summer and depleted in the winter primarily due to evaporation. As discussed in Section 3.2.2, precipitation isotope chemistry along the Pacific Coast is considered fairly constant for the past 9000 years, so there is probably little chance of a long term isotopic change from a difference in coastal climates. This may not apply to inland areas such as Eagle Valley, since there are few paleo-climate isotope studies in northern Nevada. Slight variations in temperature will change the climate considerably, as observed in the 1968-69 Sierra snow fall approaching Pleistocene climate (Smith and others, 1979).

Isotopic fractionation can also occur within an

aquifer. Oxygen-18 enrichment of water from carbonate or silicate species has been observed at high temperatures (Back and Hanshaw, 1965). Fractionation has also been observed in deep alluvial basins as water travels through clayey aquitards (Winograd and Friedman, 1972). This process, known as ultrafiltration (Freeze and Cherry, 1979) affects deuterium only and appears to be limited to waters at high pressure, at depths of at least 1500 ft.

### 3.3 TRITIUM DATING OF GROUNDWATER

Tritium, a radioactive isotope of hydrogen, is naturally produced in the atmosphere by cosmic ray bombardment of nitrogen. Natural production of  $0.25 \text{ atoms/cm}^2/\text{sec}$  is equivalent to 1-2 tritium units (TU) worldwide (Payne, 1972). One tritium unit is defined as  $1 \text{ H}^3\text{atom}/10^{18}$  hydrogen atoms or 7.2 decays per minute per liter of water (dpm/l), or 3.2 pCi/l (Fritz and Fontes, 1980). Tritium values are less over oceans due to more precipitation. Since there is usually less precipitation over land, tritium values in precipitation are larger, averaging 5 to 20 TU (Payne, 1972). Tritium in precipitation in fact, increases with overland distance from the coast.

Atmospheric thermonuclear testing in the 1950's and 1960's has increased the atmospheric tritium levels three to four orders of magnitude, as high as 9000 TU in Socorro, New Mexico (Rabinowitz, Gross, and Holmes, 1977).

Unfortunately, tritium levels were not monitored before 1953, so few pre-bomb tritium levels are actually known. Post-1953 water is referred to as modern water or bomb tritium water. Monitoring of tritium levels in Las Vegas (Figure 3.3) represents the closest location where tritium levels were sampled (from Trudeau, 1979, after unpublished data by Kauffman, U.S. EPA, Las Vegas). This graph does not show the seasonal variation of tritium which is less in high precipitation months and greater in low precipitation months. Precipitation records from Ottawa, Canada, do show lower tritium levels in the winter (times of greatest precipitation) and higher tritium levels in the summer (Freeze and Cherry, 1979). In conclusion, tritium in precipitation is a function of available tritium and the amount of precipitation (Rabinowitz, Gross and Holmes, 1977a).

Tritium decays at a fixed rate, so that once removed from the atmosphere, tritium activity in the groundwater progressively decreases. This relationship is described by a decay formula:

$$A = A_0 \cdot 2^{-t/T} \quad \text{or} \quad t = \frac{-T}{\ln 2} \ln \frac{A}{A_0}$$

$A_0$  = initial activity  
 $A$  = activity at time,  $t$   
 $T$  =  $\frac{1}{2}$ life = 12.26 years

Since tritium has such a short half-life (12.26 years), use as an indicator of groundwater age is limited to about 50 years (Campana, 1981).

Calculation of the age of groundwater with tritium can be done by calculating first arrival of tritium peak, or

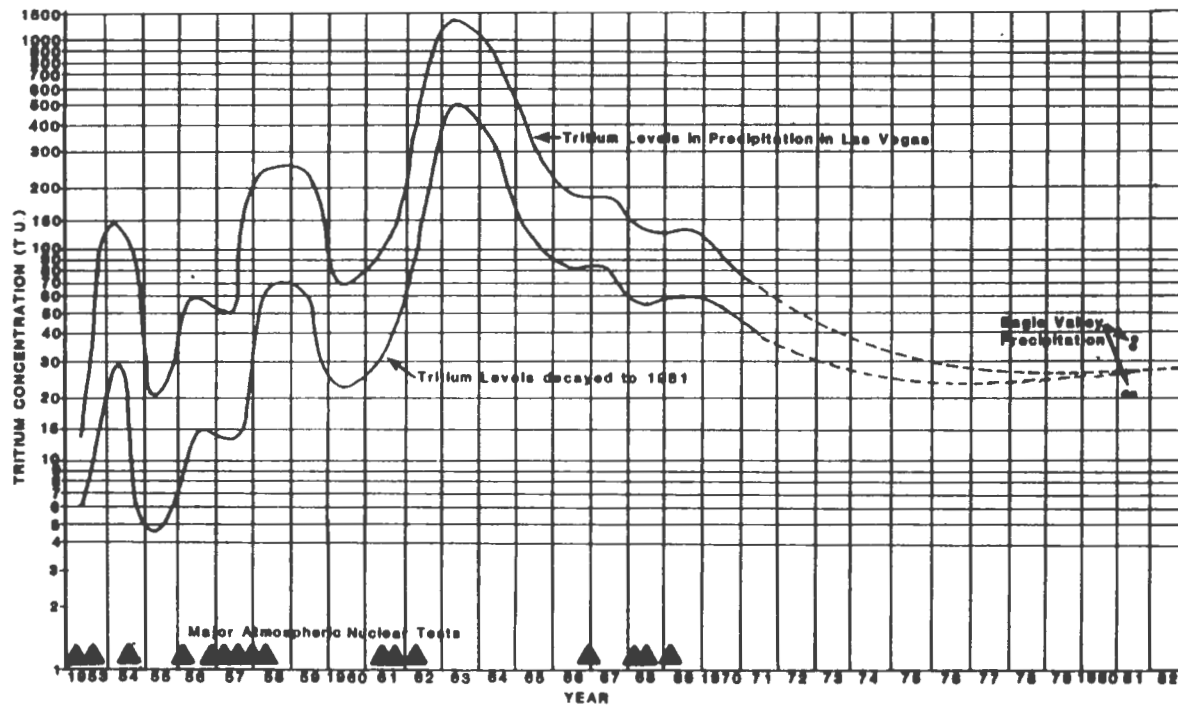


Figure 3.3 Tritium Concentration in Precipitation



timing the travel distance of an identifiable tritium peak between two observation points (Rabinowitz, Gross, and Holmes, 1977b). Other researchers (Rabinowitz, Gross, and Holmes, 1977b; Allison and Hughes, 1975; Nir, 1964; Campana, 1976; Mahin, 1978) have developed computer models for tritium input, decay, dispersion, and groundwater flow to come up with estimates of annual recharge from different sources. In this study, tritium levels in wells were not monitored for several years to identify these peaks, so there is insufficient data to justify modeling the tritium variation in the groundwater system.

A piston flow system (no dispersion or mixing) approximates an arid, unconfined aquifer system with mountain recharge (Nir, 1964). By assuming piston flow, tritium input in precipitation will simply have decayed a certain amount by the present, which is represented by the lower curve on Figure 3.3. Tritium levels in wells can then be compared to this decayed level. For example, a well containing 200 TU probably contains approximately 1963-water, which would make the water sampled in 1981, 18 years old.

#### 3.4 RADIOCARBON DATING OF GROUNDWATER

Carbon-14, like tritium, is a radioactive isotope which is produced in the upper atmosphere by cosmic ray bombardment at the rate of  $2.5 \text{ atoms/cm}^2/\text{sec}$  (Payne, 1972). The carbon-14 content in the atmosphere has been shown, by tree

ring analysis, to have varied naturally during the last 7000 years, but the variation is small. Due to atmospheric thermonuclear testing in the 1950's and 1960's, the carbon-14 concentration in the atmosphere was increased; in 1963, the level was 184% of the natural level (Gayh, 1972) in the northern hemisphere.

Carbon-14 decays at a log-linear rate to nitrogen-14, emitting a beta particle and an antineutrino. The decay can be described by:

$$A = A_0 2^{-t/T} \quad \text{or} \quad t = -8267 \ln \frac{A}{A_0}$$

$A_0$  = initial activity  
 $A$  = activity at time  $t$   
 $T$  =  $\frac{1}{2}$  life = 5730 years  
 $8267 = 5730 / \ln 2$

Because of the long half-life of carbon-14 compared to tritium, radiocarbon dating can be used to determine the age of water which is 1,000 to 50,000 years old (Campana, 1981).

The original carbon-14 activity is now defined as 95% of the activity of NBS oxalic acid in 1950; and equal to 13.56 decays per minute per gram (dpm/g) of carbon. Another standard used is wood grown in 1890 in a fossil  $\text{CO}_2$ -free environment (Fritz and Fontes, 1980). Activity of a sample is frequently reported as deviation in parts per thousand (‰)

from the standard, similar to  $\delta\text{D}/\text{H}$  defined in Section 3.2.1. Carbon-14 activity used in this study is reported as percent modern carbon where:

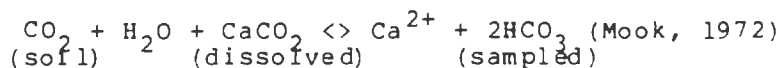
$$A/A_0 \times 100 = \text{Activity (percent modern carbon)} = A(\text{PMC})$$

An activity of 100% PMC is close to "steady state"

activity of tropospheric  $\text{CO}_2$  ( $\text{CO}_2$  in plants; Fritz and Fontes, 1980).

In theory, once carbon-14 in water, either as dissolved  $\text{CO}_2$  gas or carbonate, enters the hydrologic system, there is no further carbon-14 input, and it begins to decay. There is possible subsurface production of carbon-14 near radioactive sources (Zito, Donahue, Davis, Bentley, and Fritz, 1980). Carbon is not an ideal tracer since some carbonate from limestone and carbonate cement will mix with the groundwater as it flows through the aquifer. Since most limestone beds are greater than 50,000 years old, they will have little carbon-14 activity, and as such will effectively dilute the carbon-14 activity of the groundwater, making groundwater samples appear to be older than they actually are.

To calculate the actual age of the groundwater sampled at a point, therefore, one must know what percent of carbon is from the soil zone in recharge areas and the percent from dissolved carbonate along the flow path, as described by:



This is compounded by cation exchange with clays:



which is thought to produce  $\text{NaCO}_3$  waters. There seems to be no theoretical methods by which one can correct for both the

proportion of plant carbon in diluted carbonate and cation exchange.

Dilution can be approximated by use of a stable isotope, carbon-13. There is a difference in  $\delta^{13}\text{C}/12\%$  ( $\delta\%$  as defined in Section 3.2.1) values of soil  $\text{CO}_2$  and marine limestone, which can be used to calculate the percentage of each. Marine limestones have a fairly constant  $\delta^{13}\text{C}/12$  of about  $0.0 \pm 2.5\%$  (also the standard; Smith, Downing, Monkhouse, Otlet and Pearson, 1976). Fresh-water limestones (not as common) have a  $\delta^{13}\text{C}/12$  range of  $+5\%$  to  $-16\%$  (Edmunds and Wright, 1979).

Carbon-13 from the soil zone, derived primarily from root respiration and some from atmospheric  $\delta^{13}\text{C}/12$ , varies with the type of plant:

Calvin photosynthetic plants (C4 type, primarily cold regions):  $\delta^{13}\text{C}/12 = -25. \pm 3\%$  ;

Hatch-Slack photosynthetic plants (C3 type, common in arid regions):  $\delta^{13}\text{C}/12 = -12. \%$  ;

Succulents (C3 and C4 type): use C4 when stressed, C3 when unstressed

Recharge areas with no plants: atmospheric carbon-13 only:  $\delta^{13}\text{C}/12 = -6.4\%$ .

(from Osmond, 1981; Fritz and Fontes, 1980; Lerman, 1972; and Payne, 1972)

The best possible method for correcting carbon-14 samples for dilution is to measure soil  $\text{CO}_2$ - $\delta^{13}\text{C}/12$  in recharge areas and limestone  $\delta^{13}\text{C}/12$  in the study basin. One such

measurement of  $\delta^{13}\text{C}/^{12}\text{C}$  values and carbon-14 activity (Fritz and Fontes, 1980) shows mixing and fractionation between plants, air and soil  $\text{CO}_2$  due to photosynthetic fractionation as illustrated in Figure 3.4. Note that a 1% variation in  $A(\text{PMC})$  is equivalent to about 83 years. During respiration of vegetal carbon to atmospheric sources, it is assumed that carbon-14 enrichment is twice carbon-13 enrichment due to atomic mass ratios. Commonly, soil  $\text{CO}_2$ - $\delta^{13}\text{C}/^{12}\text{C}$  is assumed to be -25‰ and limestone is assumed to have a composition of  $\delta^{13}\text{C}/^{12}\text{C} = 0.0\%$  in order to develop correction equations (Pearson and White, 1967).

In this case, the corrected carbon-14 activity can be calculated by:

$$^{14}\text{A}_{\text{adj}} = ^{14}\text{A}_{\text{measured}}/P, \text{ where } P = p1/\{p1+1s\}$$

Using  $\delta\text{C}_{\text{sample}}$  equal to the some fraction of both plant carbon (p1) and limestone carbon (1s):

$$\begin{aligned} \delta\text{C}_{\text{sample}} &= \frac{p1}{p1+1s}\delta\text{C}_{p1} + \frac{1s}{p1+1s}\delta\text{C}_{1s} & \delta\text{C}_{1s} &= 0.0 \\ P &= (\delta\text{C}_{\text{sample}} - \delta\text{C}_{1s}) / (\delta\text{C}_{p1} - \delta\text{C}_{1s}) & \delta\text{C}_{p1} &= -25. \\ P &= \delta\text{C}_{\text{sample}} / -25. \end{aligned}$$

To take into account both carbonate dilution and cation exchange, the following relationship was developed (Pearson and White, 1967):

$$P' = \frac{C_{\text{initial}}(p1)}{C_{\text{total}}(p1+1s)} = \frac{\delta\text{C}_{\text{sample}}}{-25.}$$

$$\log(C_{\text{total}}) = -\log(\delta\text{C}_{\text{sample}}/-25.) + \log(C_{\text{initial}})$$

If plotted with  $\log[C_{\text{total}}]$  as the abscissa, and  $\log[C_s/-25.]$  as the ordinate, a slope of -1 will give a y-

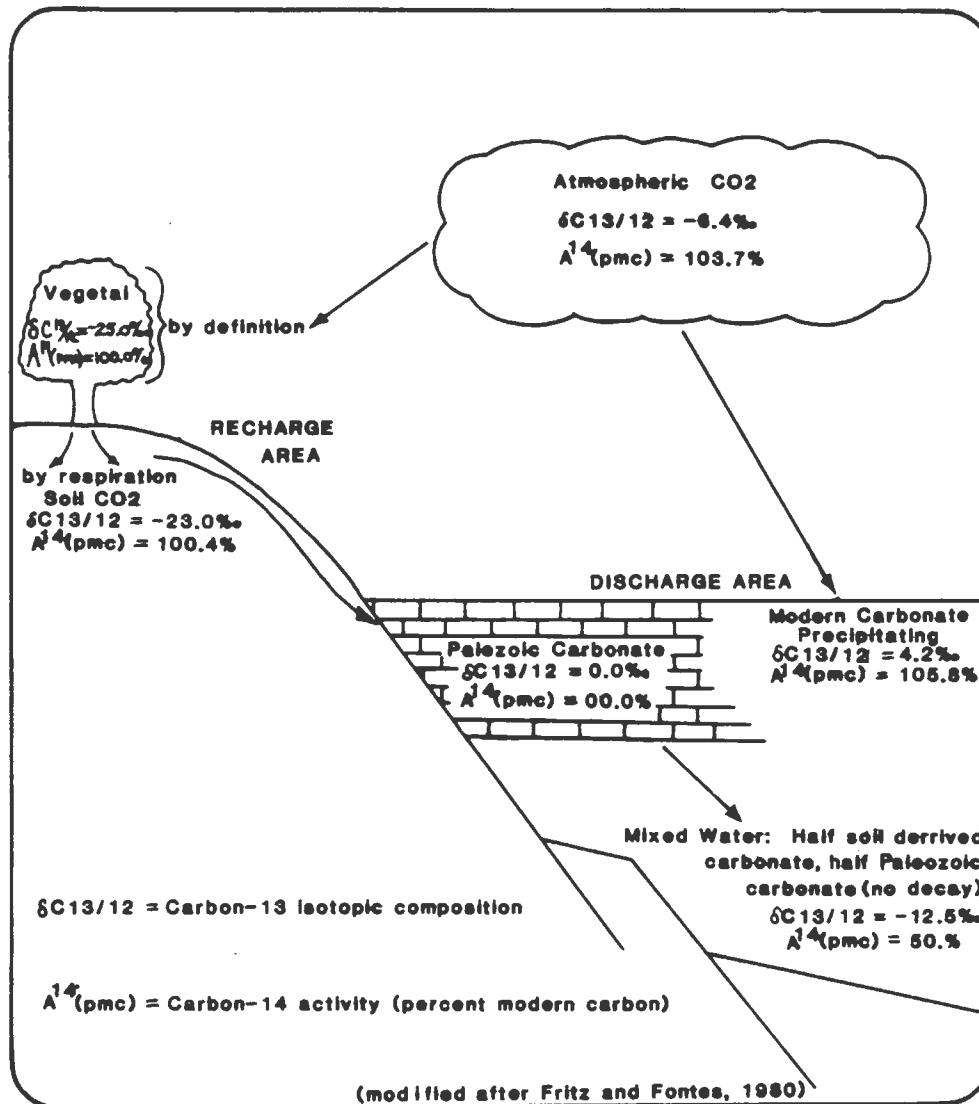


Figure 3.4 The Carbon Isotopic Cycle

intercept equal to  $\log[C_{\text{initial}}]$ . This analysis technique allows one to group samples to calculate  $C_{\text{initial}}$  for the group.  $P'$  is equal to  $P$  if one sample is used.

Other techniques used to calculate initial carbon activity have been attempted (Mook, 1972; Wigley, 1975; Geyh, 1972). One approach uses a U.S.G.S. program, WATEQ-ISOTOP, (Edmunds and Wright, 1979), which assuming a closed system taking into account  $P_{\text{CO}_2}$  and pH changes. This approach was not pursued in this study.

## CHAPTER 4

### FIELD METHODS

#### 4.1 INTRODUCTION

To quantify the hydrologic cycle in Eagle Valley, water from each component of the cycle was sampled. Snow, the predominant form of precipitation in the valley, was sampled at different elevations. Pressure-suction lysimeters were installed at different elevations to collect the portion of water from snow and rain that infiltrated. Tree xylem water was sampled to collect an average of several years of infiltrated water. Since stream channel infiltration is considered the major mechanism for recharge to the aquifer, Ash Canyon Creek was gaged and sampled at different elevations to account for gains and losses in flow and contributions from different elevations. Finally, wells and springs in the unconfined aquifer in the valley alluvium were sampled. An effort was made to sample more wells along the west to east flow path from Ash Canyon Creek to the natural discharge area, the Carson River.

#### 4.2 RECHARGE AREA SAMPLING

##### 4.2.1 Snow Sampling

Eagle Valley's main recharge area (the Carson Range) is located on the leeward side of most storms, and leeward



slopes generally do not show a stable isotopic shift with elevation as observed on most windward slopes. Snow was sampled along west-to-east and north-to-south traverses in an attempt to sample the most regular isotopic shift possible. Snow was sampled along slopes rather than in valleys to insure that the snow sample was not drifted snow from another elevation.

Snow was sampled in the Carson Range using one inch diameter PVC pipe, eight feet long. One end of the PVC pipe was modified with several larger diameter pieces of PVC and a cut out lid of a Cubitainer for the purpose of making sampling snow easier. The sampler was pushed vertically into the snow, and a Cubitainer was screwed onto the top of the sampler tube. The sampler was then withdrawn from the snow, inverted, and a 3/4 inch diameter PVC pipe with a rubber stopper was used to extrude the snow core into the cubitainer (see Figure 4.1). Sampling snow with this method insured an unbiased vertical core of snow. Where the snow was deeper than eight feet, an extension was put on the snow sampler. Air was then pushed out of the cubitainer to minimize exchange of isotopes in the snow with the air. The cubitainers were not inflated before attaching to the sampler in order to prevent any isotopic exchange with the moisture in one's breath. The above process was repeated until approximately a half gallon of snow was obtained.



Figure 4.1 Snow Sampling with Modified PVC Pipe



Figure 4.2 Sampling a Pressure-Suction Lysimeter

A total of 52 snow samples were taken in the spring of 1981 in the Carson Range, Virginia Range, and Pine Nut Mountains. Snow in the Carson Range was sampled on March 14, 1981, along the north-south ridge from Snow Valley Peak to Spooner Summit. Samples were taken at 250-ft elevation intervals, and duplicate samples were taken 50 ft apart at every other elevational station. These samples were not analyzed for stable isotope content.

Snow was again sampled in the Carson Range on April 14, 1981, along an east-west path from Snow Valley Peak through Ash Canyon (east) to Carson City. Samples were taken at 250-ft elevational changes from 9200 ft down to the lowest snow level (6200 ft). Additional snow samples were taken in Ash Canyon on May 28, and June 14, 1981, near lysimeter sites (see Section 4.2.2). Sample locations could not be located accurately at all stations with a map and Brunton compass due to the dense forest; however, samples were taken at correct elevations. A surveying altimeter used to locate stations is accurate to  $\pm 2$  ft.

Snow samples were taken in Ash Canyon in the Carson Range on April 10, 1982, for the purpose of comparing isotopic values with the previous year. These samples were not analyzed.

Snow was sampled in the Virginia Range on McClellan Peak slopes, and in the Pine Nut Mountains on March 28,

1981. Four samples were taken from 7440 ft to 6000 ft in the Virginia Range, and two samples were taken in the Pine Nut Mountains at 6200 ft and 5380 ft.

An additional three snow samples from the Carson Range and one from the Virginia Range were sampled and analyzed for tritium in order to have modern precipitation tritium values in Eagle Valley.

#### 4.2.2 Lysimeter Water Sampling

Pressure-suction lysimeters were installed in different locations in the Carson Range for the purpose of collecting infiltrated snow melt and rain water. Lysimeter water samples were used with snow samples to develop an approximate percent evaporation from stable isotope, chloride, and electrical conductivity shifts from snow to soil water. Lysimeters were also used to collect infiltrated rain water, which, in theory would have a different isotopic composition from the infiltrated snow water; unfortunately, it did not rain in the spring and summer of 1981.

The pressure-suction lysimeters used consisted of a 2.5-inch diameter, three-foot piece of PVC pipe with a porous ceramic cup at the bottom (see Figure 4.2 with Figure 4.1). Two plastic tubes penetrate the top, which is sealed with a rubber stopper. One tube extends the full length of the lysimeter, terminating inside the porous cup. The other

tube just penetrates the top of the lysimeter. To sample soil water, either tube was sealed and a suction (about 80 bars) applied to the other tube. The lysimeters were left in the ground with suction for approximately two weeks. To withdraw the water sample, positive pressure was applied to the upper tube, forcing water out the lower tube. Parizek and Lane (1970) used this type of lysimeter for sampling water up to a depth of 50 ft.

Each lysimeter in this study was installed with approximately eight to twelve inches of fine sand (#50 to #200 sieve size) packed around the porous cup. The sand insures good hydraulic contact and prevents clogging of the lysimeter. All holes were backfilled with native earth to near surface and sealed with bentonite. Dry bentonite was first applied to the surface before wet bentonite to insure that little water used in the bentonite cap would leak down into the porous cup. A hand soil auger was used to dig the holes for the lysimeters.

Three lysimeters were installed in Clear Creek Canyon at Spooner Summit at 7146 ft (L-1), U.S. Highway 50 at 6400 ft (L-2), and at the Clear Creek Youth Center, 6000 ft (L-3).

Five lysimeters were installed near Ash Canyon Creek so they could be easily located during the sampling period. Some lysimeters were placed in springs, so water collected

should not be considered sampled from a specific elevation but from a small range in elevation.

Lysimeter 4, at 8530 ft, is at the upper end of a large pasture at 8200 ft to 8500 ft in Ash Canyon. This is a local discharge area, presumably from water recharged in the fractured rock above (8500 ft to 9200 ft). This pasture area remained saturated to the surface with water from snow melt until mid-September as observed in an open hole drilled near the lysimeter.

Lysimeter 5, at 8010 ft, was installed 20 ft south of Ash Canyon Creek, and generally did not yield water. This area has a low land slope so infiltrating water in the area probably does not flow toward the stream channel.

Lysimeter 6, at 7440 ft, was placed on the north side of Ash Canyon Creek, 50 ft above the creek. The lysimeter was placed at a depth of 2 ft, directly in a small spring. This spring has enough discharge to keep the surface wet, but does not produce any surface flow. The spring drains an area which rises to an elevation of 8350 ft.

Lysimeter 7 was originally placed at 7000 ft, about 20 ft south of Ash Canyon Creek. The soil is very thin here due to outcropping granodiorite, consequently, this lysimeter did not yield any water. This lysimeter was moved on August 13, 1981, to a flowing spring about 150 yards

northeast of the creek. This spring drains an area up to 7800 ft in elevation.

Lysimeter 8 was originally placed at 6400 ft, but since soil moisture was very low (2-4%), the lysimeter did not yield any water. The lysimeter was moved on June 11, 1981, to 6250 ft, near North Ash Canyon Creek, and placed at a depth of four ft. This area consists of a silty coarse sand which is saturated at a depth of three feet; where possible subsurface discharge to the stream may be occurring. The water in the lysimeter is not bank storage from the stream, as evidenced by conductivity measurements. The conductivity of lysimeter water is 145 umhos/cm, and that of the stream is 60 umhos/cm, indicating the lysimeter water probably travelled farther in soil than the stream water.

From the eight lysimeters installed and monitored from May, 1981, until May, 1982, a total of 50 soil water samples were taken.

On May 20, 1982, lysimeters 1, 2, and 3 (in Clear Creek Canyon) were sampled for soil- $\text{CO}_2$ , in an effort to collect initial  $\delta\text{C}13/12$  and  $\text{C}14(\text{pmc})$  values in recharge areas. All three lysimeters were sampled for  $\delta\text{C}13/12$ , and lysimeter 2 was sampled for C-14 activity. Soil  $\text{CO}_2$  was first measured with a Draeger  $\text{CO}_2$ -tube analyzer. Sampling was done by drawing, with a vacuum pump, air from the lysimeters through a series of erlenmeyer flasks filled with NaOH and  $\text{SrCl}_2$

solutions (described in Section 5.4.2). Airstones were used in each bottle to increase CO<sub>2</sub> capture (see Figure 4.3).

#### 4.2.3 Soil Sampling

In order to determine the isotopic composition of water in the soil where lysimeters could not yield water, soil samples were taken, and the soil was distilled in the laboratory to extract water. A three-inch diameter hand soil auger (courtesy of the Nevada Bureau of Mines and Geology) was used to dig out the soil. At each site an attempt was made to sample a zone two to three ft below the surface, the zone where lysimeters sampled water. Soil samples were taken at shallower depths at L-7 site, where the soil is much thinner. About ten pounds of soil were collected at each site, which was placed in double plastic soil sample bags. The procedure used for distilling samples is described in Section 5.2.2.

#### 4.2.4 Tree Water Sampling

Techniques for sampling water in recharge areas need to be developed in order to study remote basins. If tree water sampling can be appropriately developed, it would be possible to sample tree water rather than lysimeter water.

Tree samples were taken in Eagle Valley on July 9, and July 21, 1981, at each lysimeter station for stable isotope analyses. Additional tree samples were taken in Clear Creek





Figure 4.3  $\text{SrCO}_3$  Precipitate from Soil- $\text{CO}_2$



Figure 4.4 Tree Coring

Canyon at 6400 ft on January 25, 1982, (lysimeter 2 site).

The heartwood or water-stabilized part of the tree was sampled in this study by coring. A core was taken through the bark and one inch of the phloem and xylem, then discarded. The tree was then cored through the center of the tree. One core produced about 70 to 90 grams of wood chips, which produced 30 to 40 grams of water when distilled (average moisture content of 45%), enough water for a stable isotope analysis. Twelve cores were taken from a single tree on January 25, 1982, for tritium analyses; eight cores produced 250 milliliters of water.

Coring was accomplished with a one-inch diameter, fifteen-inch long, slow speed wood-boring auger bit and a hand brace (see Figure 4.4, with Figure 4.3). The bit produced shavings of wood suitable in size for use in the distillation apparatus (see Section 5.2.2). Wood shavings were collected in a plastic bag pinned to the tree beneath the area being augered. Air was pushed out of the bags before sealing. The hole in the tree was then filled with a wood dowel and tree tar.

Since different species of trees occupy different ecological niches, it was not possible to sample the same species at all elevations. A total of twenty tree samples were taken, and fifteen were analyzed. Trees sampled are approximately 50 to 70 years old, based on similar sized

trees cored by the U.S. Forest Service in Ash Canyon at 6400 ft.

#### 4.3 STREAM GAGING AND SAMPLING

To confirm the sizable contribution of stream channel infiltration to groundwater recharge, Ash Canyon Creek was gaged at nine locations on August 27, 1981, and at five locations on January 29, 1982.

Stream gaging was performed using the velocity-area method; by measuring the cross-sectional area and velocity at several points across the stream. Each site was gaged at three to six gaging points. The velocity was measured using either a Marsh-McBirney 201 transducer flow meter with digital readout, in August (see Figure 4.5), or a USGS pigmy meter in January.

Ash Canyon Creek and side channels were gaged on August 27, 1981, at 8150 ft, 7870 ft, 7450 ft, 7000 ft, 6620 ft, 6190 ft, and 5290 ft. Temperature, conductivity, and pH readings were taken at every gaging station and at seven springs that flowed into Ash Canyon Creek. Stable isotope (125ml) and tritium samples (1 liter) were taken at each gaging site.

Ash Canyon Creek was gaged again on January 29, 1982, only in the lower section, 6620 ft, 6190 ft, and 5290 ft, where the August gaging indicated channel loss. It was



Figure 4.5 Stream Gaging with Transducer Flow Meter in Ash Canyon Creek at 7800 feet.

hoped that this gaging would help determine if water loss was due to evapotranspiration or stream flow infiltration.

The USGS Parshall flume at 5290 ft was gaged and velocity was calculated from the height to determine gaging error.

#### 4.4 DISCHARGE AREA SAMPLING

##### 4.4.1 Well and Spring Sampling

Wells and springs throughout Eagle Valley were sampled for major ions, stable isotopes, tritium, and carbon isotopes. An east-west zone extending from lower Ash Canyon to the Carson River was more extensively sampled. A total of twenty-seven wells and two springs, throughout Eagle Valley, were sampled and analyzed for major ion chemistry and stable isotopes, and nineteen were sampled for tritium. A total of five wells and two hot springs were sampled for radiocarbon dating.

The procedure used for sampling wells was designed to obtain a fresh groundwater sample. Most wells sampled were privately owned in which owners consented to the sampling, and five wells were Carson Water Company wells. Sampling at a well was done by turning on a faucet, preferably near the well, for about ten minutes to allow the holding tank to drain and the pump to turn on. A five-gallon bucket was then filled, which was used to fill all sample bottles and

perform field water chemistry. Major ion samples were collected in one-gallon plastic cubitainers; tritium samples were collected in one liter glass bottles; and stable isotope samples were collected in two 125-ml glass bottles. Tritium and stable isotope sample bottles were filled and capped underwater so that no air was trapped in the bottles. Both types of bottles have teflon-lined caps, which provide an airtight seal. Electrical tape was then wrapped around sample caps.

Field chemistry data were taken at each well and consisted of a minimum of temperature and electrical conductivity, with some wells tested for pH and dissolved oxygen. All probes were placed in the bucket before samples were taken to allow probes to equilibrate with the well temperature.

Wells that were sampled for carbon isotopes required special treatment. These wells or springs were first sampled and analyzed in the laboratory for bicarbonate present in solution. Field determination of bicarbonate content would have been more accurate, but was not done. The amount of water needed to obtain a minimum of two grams of carbon was then calculated (see Section 5.4). The amount of water necessary for collection was: 100 liters for sample #177, 120 liters for samples #112 and #10, 150 liters for samples #48 and #52, 200 liters for sample #1002, and 400 liters for

sample #178. Fifty liter plastic carboys were used to collect these samples.

Samples #48, #52, and #177 were sampled on October 27, 1981. Samples #48 and #52 are Carson Water Company wells, which were turned on two hours prior to sampling. The Maximum Security Prison Hot Springs (#178) was sampled on November 6, 1981, directly from the spring discharge by use of a sump pump. Carboys were filled directly in the truck. Samples #112, #10, and #1002 (Carson Hot Springs) were collected on May 20, 1982. The procedure used for precipitation of the carbonate from these samples is described in Section 5.4.2.

#### 4.4.2 Stream Gaging

The Carson River was gaged on November 6, 1981 in order to verify the groundwater discharge estimated by Worts and Malmberg (1966) and Arteaga and Durbin (1978). The Carson River was gaged at the USGS gage east of Prison Hill and at the Brunswick Canyon Bridge. In addition, outflow and inflow to Mexican ditch, and the runoff ditch from the sewage treatment plant were also gaged. The Carson River was gaged with a USGS pygmy meter, using approximately 40 stations across the river for velocity and depth; the smaller diversions were gaged with 9 to 11 stations.

## CHAPTER 5

## LABORATORY METHODS

## 5.1 MAJOR ION ANALYSIS

Twenty-nine samples from Eagle Valley were analyzed for pH, electrical conductivity, major anions ( $\text{HCO}_3^-$ ,  $\text{CO}_3^{2-}$ ,  $\text{Cl}^-$ ,  $\text{SO}_4^{2-}$ ), major cations ( $\text{Na}^+$ ,  $\text{K}^+$ ,  $\text{Ca}^{+2}$ ,  $\text{Mg}^+$ ), and silica at the Analytical Laboratory of the Desert Research Institute's Water Resources Center. Cation-anion balance is within  $\pm 3\%$  on all samples (oral communication, Jim Heidker). Overall laboratory error is approximately 5% or less.

The laboratory pH was measured with an Orion Research 901 Ion Analyzer with glass electrode, which has an accuracy of  $\pm 0.05$  pH. The electrical conductivity of all samples was measured in the laboratory at  $25^\circ\text{C}$  with a Beckman Conductivity Bridge which has an accuracy of  $\pm 5\%$  (American Public Health Association, 1980).

All major cations were measured on an Instrumentation Laboratory (IL) Atomic Absorption/Atomic Emission Spectrophotometer 951. Sodium was measured by the flame photometric method with detection limits of 0.1 mg/l, and an accuracy of about 5%. Potassium was also measured by the flame photometric method. Minimum detectable concentrations of potassium are 0.1 mg/l, with an accuracy of 5% by this analytical method.



Calcium concentration in samples was measured by the atomic absorption spectrometric method with detection limits of 0.1 mg/l, and an accuracy of  $\pm 1\%$ . Magnesium was measured by the same method, in which the absorption of light at a specific wavelength is measured. Minimum detectable concentrations are 0.1 mg/l with an accuracy of  $\pm 1\%$ .

The alkalinity or carbonate component of samples was measured by potentiometric titration, in which sulfuric acid is added as the pH change is measured. Accuracy is about  $\pm 3$  mg/l. Chloride content in samples was determined colorimetrically, by addition of mercuric thiocyanate and ferric ammonium sulfate, and measured on a Coulter Automated Analyzer with detection limits of 0.1 mg/l.

Sulfate was measured by the turbidimetric method, in which barium chloride is added to precipitate barium sulfate, and the absorption of light is measured with a photometer. Detection limits are 1 mg/l, accuracy of  $\pm 1\%$ .

Dissolved silica was determined by the colorimetric method by the addition of ammonium molybdate and aminonaphthosulfonic acid to the solution, and measured on a Coulter Automatic Analyzer. Detection limits for silica are 1 mg/l.

## 5.2 STABLE ISOTOPE ANALYSIS

### 5.2.1 Introduction

Water from snow, lysimeters, soil, tree xylem, springs, and wells were analyzed for stable isotopic content. Water had to be extracted from low moisture soil and tree xylem by use of a distillation apparatus described in Section 5.2.2. Most stable isotope samples were stored for a considerable length of time before analysis (several months); the procedure used for preserving samples is described in Section 5.2.3. A discussion of the techniques used for stable isotope analysis is found in Section 5.2.4.

### 5.2.2 Soil and Tree Water Distillation

The objective of the distillation process is to remove all water from the sample, since isotopic fractionation could result with partial distillation. From a previous discussion (Section 3.2.4), it was concluded that fast evaporation results in less isotopic fractionation. Therefore, fast distillation should be incorporated when samples are to be analyzed for isotopic content.

The cold finger distillation apparatus used in this study can heat samples up to 400°C, and uses liquid nitrogen to supercool the vapor (-210°C), which is considerably more efficient than cold water or dry ice. The cold finger apparatus used consists of a flask containing the sample,

which is connected to the condenser by an angled glass tube. Both the flask and the connector tube have heating mantles, to evaporate water. The condenser is double walled so that liquid nitrogen can be poured into the condenser and remain separated from the sample (see Figure 5.1). The system uses a vacuum to draw the vapor from the sample flask into the condenser. The flask and connector tube heating mantles are controlled by a variable voltage supply to regulate temperature. Actual temperature of the sample is difficult to control; temperature is dependent on the number of heating mantles turned on and the mass of the samples. No condensation apparatus is 100% efficient, but when this apparatus was used before for soil distillation for tritium analysis, a spiked sample showed about 95% recovery of water with about 1% fractionation (Hansen, 1978).

The first step in the soil distillation process was to determine the percent moisture of the sample. If possible, 18 ml of water was distilled to insure filling a 13 ml glass vial. Smaller (6.5 ml and 11 ml) vials were also used. A minimum of about 3 ml of water is needed for analysis of oxygen-18 and deuterium (with no rerun). Calculation of percent moisture was done by drying a small quantity of sample for 24 hours at 105°C, and measuring the weight of water lost. Soil moistures sampled ranged from 2 to 20%. Very dry soils required a very large volume of soil to produce enough water for analysis. The 250 ml flasks of the distillation

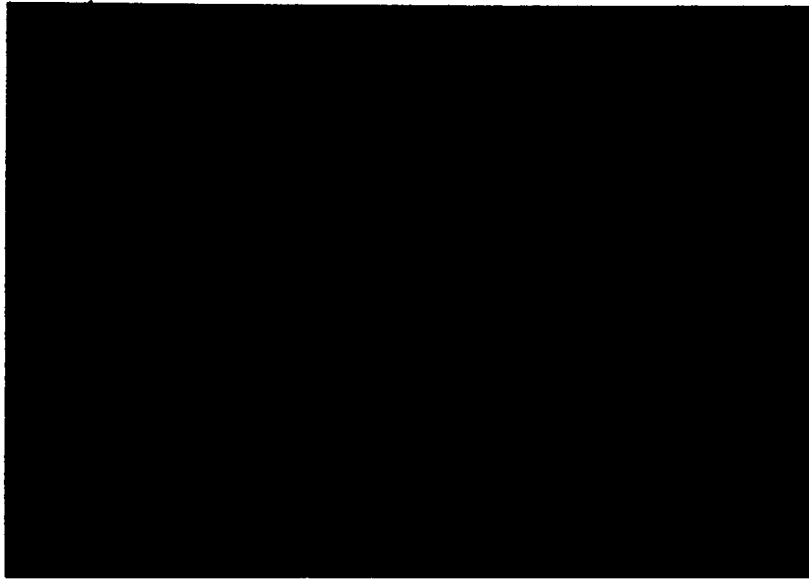


Figure 5.1 Soil and Tree Water Distillation Apparatus

apparatus could hold a maximum of about 450 grams of soil, which sometimes produced only 7 ml of water. Usually about 2-3 ml of the distilled water clung to the walls of the condenser apparatus and were irretrievable. Other problems were encountered with organic soils and silty soils. Organic soil samples usually produced a yellow distillate containing an immiscible organic compound, probably humic acid. Heated rapidly, silty soils had a tendency to burst in the flask when releasing trapped water vapor. This often sent silt up into the condenser, ruining the water sample.

It was determined that soils could be heated to 250°C (55 volts) in most cases. Very wet soils were heated slowly to release vapor slowly, since excess vapor could pass through the condenser and out the vacuum tubing. Very dry soils were also heated slowly to release vapor slowly. Most soil samples were distilled in about two hours.

Tree samples were more difficult to distill correctly than soil samples. Since tree moistures (by weight) ranged from 40% to 55%, it was necessary to heat the samples slowly so that only a limited amount of vapor was released at a time, since vapor can escape through the vacuum tubing. Tree samples taken on July 9, 1981, (samples #127-#133) were heated to about 250°C, which distilled water and a yellow organic compound. Tree samples taken July 21, 1981, (#143-#147) were heated more slowly to a maximum temperature of

160°C, which did not distill any yellow liquid. This yellow liquid has a significant effect on the stable isotopic content of the sample. Tree samples taken January 25, 1982, (#181c and #181y) were distilled at low and high temperatures to document the effect of the yellow compound, as is discussed in Section 6.2.2.3.

#### 5.2.3 Stable Isotope Sample Preservation

In order to insure a minimum of isotopic exchange between samples and the atmosphere, certain precautions were taken. Snow samples were brought back to the laboratory in cubitainers and transferred into glass bottles. This was done in a cold storage room (4°C) to minimize the amount of water vapor present, thus minimizing isotopic exchange. Soil and tree samples that were distilled in the laboratory were transferred into 6.5 ml, 11 ml, or 13 ml glass bottles with teflon-lined caps. All stable isotope samples, once in glass bottles, were sealed with electrical tape and parafilm, or dipped in hot wax.

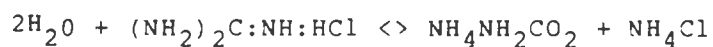
#### 5.2.4 Stable Isotope Analysis

Of the 175 stable isotope samples taken, a total of 71 samples were analyzed for  $\delta D/H$  and  $\delta^{18}O/^{16}O$ ; 27 samples (20 snow, 7 wells) were analyzed by Dr. John Hess at the Scottish University Research and Reactor Centre in East Kilbride, Scotland. The remaining samples were analyzed at the

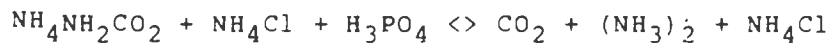
Department of Geosciences, Isotope Geochemistry Laboratory, University of Arizona, Tucson, Arizona. Stable isotope analysis, in general, consists of changing the sample into  $\text{CO}_2$  or  $\text{H}_2$  gas, ionizing the gases, and measuring the different isotopes present by the amount of deflection as the ions pass through the magnetic field of the mass spectrometer. Early determination of stable isotopes was done by very precise measurements of density of water, where it was not possible to differentiate oxygen and hydrogen isotopes.

5.2.4.1 Sample Preparation. The analytical methods of oxygen isotope analysis differed between the two laboratories, although the hydrogen isotope analysis technique was essentially the same.

The Scottish laboratory analyzed oxygen isotopic content by reaction of 8 $\mu\text{l}$  of water with 50 grams of guanidine at 260°C overnight according to the following reaction:



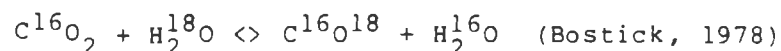
The products are then heated in an oven with  $\text{H}_3\text{PO}_4$  at 80°C for 45 minutes, and the following reaction occurs:



The  $\text{CO}_2$  gas is then analyzed for  $\delta^{18}/^{16}$  (Hess, 1981). This technique has been refined at the S.U.R.R.C. after previous work (Boyer and others, 1961). Six samples were prepared

simultaneously. The oxygen isotopic content of the CO<sub>2</sub> was then measured on the mass spectrometer.

The Tucson laboratory prepared oxygen isotope samples by a method used in most laboratories; equilibrium of the water sample with CO<sub>2</sub> gas. This technique involved placing a small amount of sample water into a small vial, and freezing and melting several times using alcohol and solid CO<sub>2</sub> (-70°C), for the purpose of purging dissolved gases from the water. The sample was then injected with a known amount of carbon dioxide of known isotopic composition. The CO<sub>2</sub> and water were then allowed to equilibrate for 24 hours in a 25.0°C water bath. The isotopic exchange reaction is:



The carbon dioxide gas was then removed from the container by a freezing and thawing process. Samples were prepared individually, and then measured on the mass spectrometer.

Hydrogen isotopic content of water samples was measured by the same method in both laboratories. The δD/H ratio measurement involved liberating hydrogen from a portion of the sample by reaction with hot uranium metal to produce hydrogen gas which was measured on the mass spectrometer.

5.2.4.2 Measurement by Mass Spectrometer. The mass spectrometer partially ionizes the samples and accelerates



the ions produced through a magnetic field. Different ions have different mass and charge ratios, so the magnetic field produces different trajectories for the various ions. The less massive and most abundant ions ( $^{16}\text{O}$  and  $^1\text{H}$ ), are collected over a charged plate, and the more massive and less abundant ions ( $^{18}\text{O}$  and  $^2\text{H}$ ) are collected in a Faraday cage. Measurement of a sample is reported as deviation from a standard. Standards are run before samples to produce a calibration curve. They are run between samples to correct for any drift in the mass spectrometer and to check for any memory developed in the mass spectrometer. Memory occurs when the mass spectrometer does not accurately count a sample because it partially "remembers" a previous sample with very different isotopic composition.

Standards that are used include SMOW (Standard Mean Ocean Water,  $\delta\text{D}/\text{H} = 0.0\%$ ,  $\delta^{18}/16 = 0.0\%$ ), SLAP (Standard Light Arctic Precipitation,  $\delta\text{D}/\text{H} = -428\%$ ,  $\delta^{18}/16 = -55.5\%$ ), and GISP (Greenland Precipitation Sample). Corrections are made for the presence of tritium (in hydrogen isotopic measurement), and for mixing of the sample and standard gases due to valve leakage (Muir and Coplen, 1981).

Oxygen isotope samples analyzed in Scotland were measured on a VG 903 (V.G. Micromass Limited, Winsford, Cheshire, England) Mass Spectrometer, which has a  $90^\circ$  sector magnetic field of 9 cm radius created by a permanent magnet.

Reported accuracy of the guanidine method of oxygen sample preparation on this spectrometer is  $\pm 0.3\%$ . This accuracy band will become smaller as the inaccuracies are worked out in the new spectrometer, the VG903. This method is considerably less time consuming than the water-CO<sub>2</sub> equilibrium method. Deuterium samples analyzed in Scotland were measured on a VG 602 Mass Spectrometer, which has a 6 cm radius, 90° sector created by a permanent magnet. Reported accuracy is  $\pm 0.5\%$ .

Both oxygen and hydrogen isotope samples analyzed in the Tucson laboratory were measured on a VG 502C mass spectrometer. Reported accuracy was  $\pm 0.2\%$  for oxygen, and  $\pm 3.0\%$  for hydrogen. Note that the oxygen confidence band from the Tucson lab is slightly smaller than the Scotland laboratory, since the analytical method in Scotland is still experimental. However, the hydrogen band is considerably larger for the Tucson laboratory.

There appears to be a discrepancy in reported values from the two laboratories which is greater than the stated accuracy. The three samples, (#DR1, #DR1A, #DR1B), from another isotope study (Ingraham, 1981), sent to both labs indicated a regular difference between the laboratories. The Scotland laboratory reported deuterium values averaging  $7.2 \pm 3.0\%$  more depleted (negative) than the Tucson laboratory, and oxygen values  $.88 \pm .82\%$  more depleted than the

Tucson laboratory. The confidence band represents the standard deviation of the three samples. It is unknown why the discrepancy exists. Samples were analyzed within six months at both laboratories, so fractionation during storage is probably not the cause of the difference.

### 5.3 TRITIUM ANALYSIS

#### 5.3.1 Introduction

A total of fifteen samples were analyzed for unenriched tritium (detection limits of 50 TU), and eighteen samples were analyzed for enriched tritium (limits of 5 TU). Of these samples, four were modern precipitation samples, two were stream samples, one was a tree sample, and the remaining were samples from wells and springs. All samples were analyzed at the Desert Research Institute Analytical Laboratory in Reno, Nevada. Most samples were collected in one-liter glass bottles, although some were collected in 125 ml bottles. Since approximately 250 ml of water is needed to analyze for enriched tritium, the 125 ml samples had detection limits of 10 TU instead of 5 TU.

#### 5.3.2 Analysis Method

The procedure used for determination of tritium in water by enrichment consisted of distilling the sample, followed by slow electrolysis at a constant temperature which evaporated the hydrogen and left tritium to be measured by a

scintillation counter. Unenriched tritium samples were counted for tritium activity in the scintillation counter without enrichment.

The procedure for tritium enrichment used was modified after Johns (1975) and has detection limits of about 5 TU. This procedure consists of first distilling 300 to 500 ml of sample water until the distillate has a conductivity below 50 umhos/cm. 100 ml of the distilled sample were then added to a electrolytic cell, and one gram of sodium hydroxide was added to reduce corrosion of electrodes. Electrolytic cells were then put into a constant temperature bath along with one "dead" water (0.0 TU) and two standards (100 TU and 10,000 TU). A current of 3 amps and a potential of 2.7 volts was then applied to each cell. Samples electrolyzed and boiled off hydrogen until about 50 ml of sample remained. Cells were refilled, and this process continued until all 200 ml of each sample had been put into the cells. Electrolysis was then continued for a week until the volume decreased to 25 ml. The current was then decreased to 0.3 amps, and the electrolysis continued until the volume decreased to 7-8 ml. This took approximately four weeks.

Carbon dioxide was bubbled through the remaining fluid to neutralize the sodium hydroxide. Since the solution was somewhat opaque from dissolved solids, samples were then distilled over a liquid nitrogen trap, similar to the cold

finger condenser described in Section 5.2.2. A total of 5 ml of sample water is needed for counting.

Samples, dead water, and standards are now ready for counting. Liquid scintillation cocktail is added to each sample, and several blank samples, which contain scintillation cocktail and dead water. All vials are placed into the scintillation spectrometer and allowed to dark-adapt for 24-36 hours before counting. Blank samples are placed in between each standard and sample.

Vials are now counted for 50 minutes each, and counted 10 to 12 times. The activity of the tritium in each vial causes the scintillation cocktail to fluoresce, and a specific wavelength is counted by the spectrometer. Mean and standard deviation of the counts of each sample, standards, and blanks are then calculated. A calibration curve of counts versus tritium units is calculated from the standards, and sample tritium units are calculated from this curve. The actual tritium counts from a sample are represented by:

$$\text{Sample tritium} = [\text{Counts}(\text{sample}) - \text{Counts}(\text{blank})] \pm$$
$$[\text{Standard deviation}(\text{sample}) - \text{Standard deviation}(\text{blank})]$$

This subtracts out the activity of the scintillation cocktail from the reported tritium level of a sample. The sample standard deviation averaged  $\pm 5$  TU. By enriching more

water and using a different counting method (Nehring, 1980), the standard deviation can be as low as  $\pm .2$  TU. Overall process efficiency is represented by the fractional counting efficiency, which consists of the measured tritium counts of the standards divided by theoretical tritium counts of the standards. Fractional counting efficiency averaged about 20% for all samples.

#### 5.4 CARBON ISOTOPE ANALYSIS

##### 5.4.1 Introduction

Analysis of the age of groundwater by use of carbon isotopes involves collection of enough water to precipitate two grams of carbon as  $\text{SrCO}_3$ , which is the minimum needed to measure carbon-14 activity and  $\delta\text{C}13/12\text{‰}$ .

The volume of water needed to precipitate a minimum of two grams of carbon can be calculated by:

$$\text{Volume(liters)} = [10,000\text{mg CO}_3^{2-}] / [\text{HCO}_3^- + \text{CO}_3^{2-} \text{ in ppm}] \quad (\text{IAEA, 1971})$$

The formula assumes that to produce two grams of carbon, ten grams (10,000mg) of carbonate are needed. In this study, wells were first sampled and analyzed for carbonate content, then resampled for carbon isotope analysis. About 30% more water than the minimum necessary was collected, to insure precipitation of enough carbon.

#### 5.4.2 Carbonate Precipitation

The procedure used for precipitation of carbonate from samples consists of placing the sample water into a container sealed from the atmosphere, raising the pH to convert bicarbonate species to carbonate species, then addition of strontium chloride, to precipitate  $\text{SrCO}_3$ .

The apparatus used to precipitate carbonate consists of a four foot high stainless steel funnel which holds a maximum of 67 gallons (253 liters). The funnel rests on a 55 gallon steel drum. The top of the funnel has a foam seal and a large piece of plexiglass as a lid (see Figure 5.2). The plexiglass has a central hold which through which a shaft and two propellers are attached. Agitation of the solution is done via a hand-crank drill. Other holes in the lid are for supplying nitrogen pressure for evacuating air from above the water, a hole for air escape, and a hole for supplying chemicals.

The bottom of the funnel has a three inch diameter valve modified so that a two liter bottle can be attached below the valve for collection of precipitate. The bottle can be removed by closing the valve. The bottles are made of linear polyethylene, necessary to reduce  $\text{CO}_2$  exchange between the sample and the atmosphere. The precipitator was designed and built by Neil Ingraham (1982) and the author, adapted after Winograd (1971).

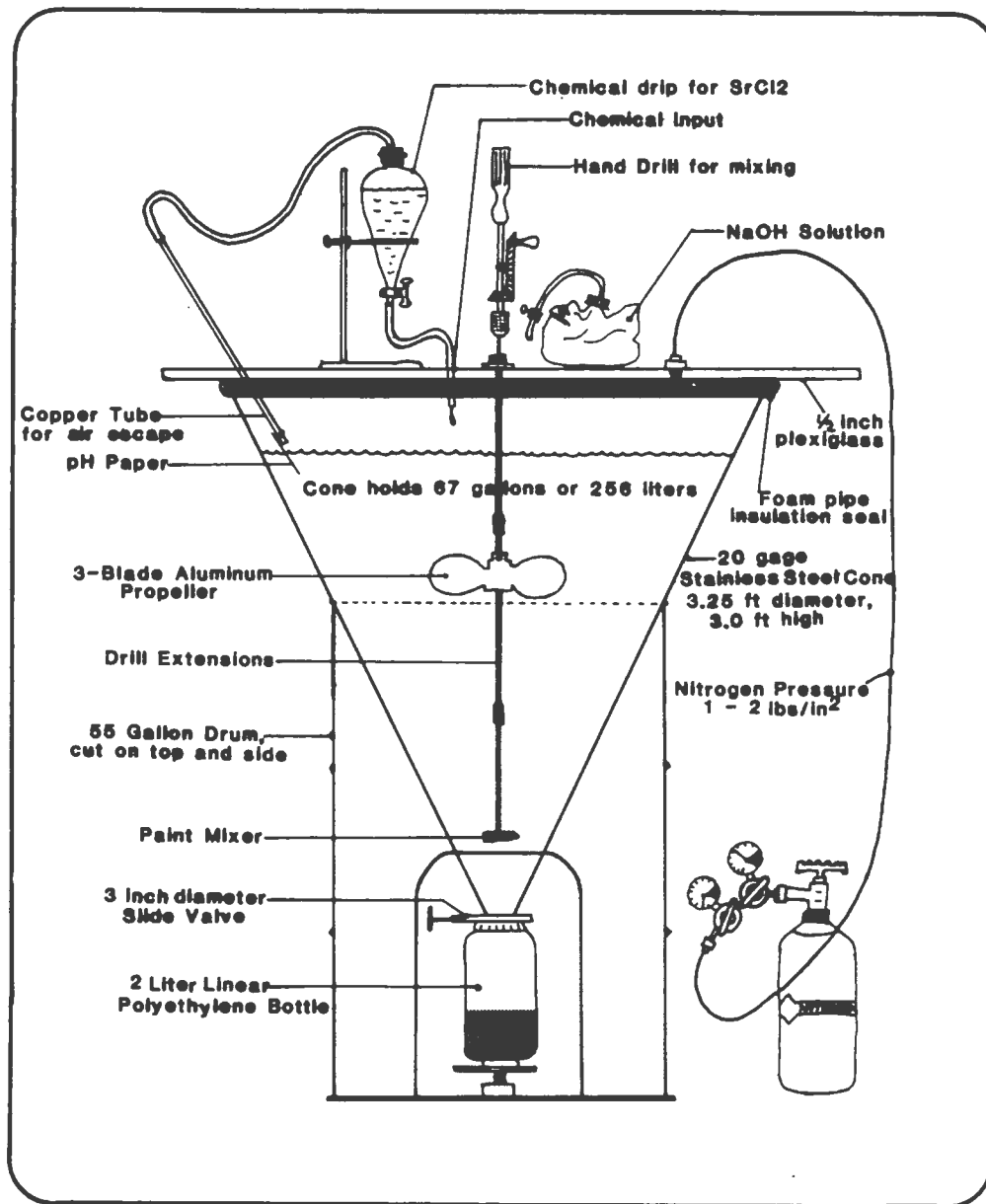


Figure 5.2 Carbonate Precipitation Apparatus

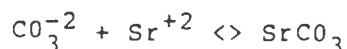


The procedure for precipitation consists of first slowly filling the precipitator with sample water with a minimum of agitation. The plexiglass lid is then installed and nitrogen pressure is applied to remove air from inside the precipitator and minimize  $\text{CO}_2$  exchange of sample with air. Since nitrogen is applied above the sample, it is not necessary to fill the precipitator to the top of the funnel. Two chemicals are needed for precipitation:

(1) A 6N, carbonate-free, sodium hydroxide solution. This is prepared by using boiled, distilled, deionized water to remove carbonate. A small amount of the  $\text{SrCl}_2$  solution is added to precipitate out any carbonate left, and only the clear NaOH solution is siphoned into cubitainers.

(2) A strontium chloride solution. This is prepared by dissolving 454 grams of  $\text{SrCl}_2 \cdot 6\text{H}_2\text{O}$  in one liter of distilled, deionized water. Again, a small amount of NaOH solution is added to precipitate any remaining carbonate, and the clear supernatant solution is siphoned into cubitainers (adapted after Pearson, 1970).

Approximately 500 ml of the NaOH solution is added to raise the pH of the water to 12, converting virtually all carbonate species to carbonate. The pH is checked with pH indicator paper attached to the air escape tube. About 1000 ml of the  $\text{SrCl}_2$  solution is then added and the following reaction takes place:



precipitating strontium carbonate, which falls down the funnel into the sample bottle. If precipitation does not take

place immediately, 5 grams of  $\text{FeSO}_4 \cdot 7\text{H}_2\text{O}$  are added, which causes the formation of coarse agglomerates on which  $\text{SrCO}_3$  crystals can grow, thus speeding the settlement. Another author (Nehring, 1980) suggests adding a few grams of cadmium chloride, which precipitates  $\text{H}_2\text{S}$  and acts as a flocculating agent. This should be added before the  $\text{NaOH}$  and  $\text{SrCl}_2$  solutions.

Samples were allowed to precipitate for 8 to 24 hours, with the funnel being stirred every 2-3 hours. Additional  $\text{SrCl}_2$  solution was added every few hours to check if all  $\text{SrCO}_3$  had precipitated. When precipitation was completed, the valve at the bottom of the funnel was closed, and the precipitate removed. This size precipitator can hold enough water to precipitate two grams of carbon from samples with a bicarbonate content as low as 40 ppm. In addition, by simply closing the valve, and siphoning off the water and reprecipitating more water, much lower carbonate content can be sampled.

#### 5.4.3 Analysis Method

All seven carbon isotope samples were sent to Teledyne Isotopes, Westwood, New Jersey, for analysis of carbon-14 activity, and  $\delta^{13}\text{C}/^{12}\text{C}$ .

Preanalysis treatment of the samples involved filtering off the strontium carbonate and drying. After pretreatment,

CO<sub>2</sub> is generated from the carbonate samples by acidification. The CO<sub>2</sub> is then purified in a vacuum system to remove halides, sulfur compounds (such as FeSO<sub>4</sub>), oxides of nitrogen, and electronegative species which affect the counting characteristics of the gas. Radon is extracted routinely and the purified CO<sub>2</sub> is ready for counting (Teledyne Isotopes, 1981).

Actual radiometric measurement is made by proportional counting of the CO<sub>2</sub> gas housed in a twenty-ton lead and steel shield. Sample purity is monitored by noting the counting characteristics of an external gamma radiation source at low voltage. All samples were counted for about 1000 minutes each time (Currie, 1972), and counted two to three times to eliminate the possibility of error due to traces of residual contamination by radioactive radon gas, and to produce greater precision and sensitivity. Data are reported as deviation in parts per thousand from the oxalic acid standard.

All samples were also analyzed for  $\delta^{13}\text{C}/^{12}\text{C}$ , which is done by use of a mass spectrometer with a technique similar to oxygen and hydrogen stable isotope analysis, as described in Section 5.2.4.2. The  $\delta^{13}\text{C}/^{12}\text{C}$  reproducibility is about  $\pm 0.7\%$  (Fritz, Suzuki, Silva and Salati, 1981).

## CHAPTER 6

## RESULTS AND DISCUSSION

## 6.1 CHEMISTRY OF PRECIPITATION

## 6.1.1 Stable Isotope Chemistry

Snow samples taken in April, 1981, along a west-to-east path down Ash Canyon are plotted as  $\delta D/H$  versus  $\delta 18/16$  in Figure 6.1. The seventeen snow samples produce a straight line equation of:

$$\delta D/H = [6.03 \pm 1.27] \delta 18/16 - [14.02 \pm 20.3], \text{ with } R^2 = .86$$

which indicates correlation at the 5% significance level. The uncertainties of the slope and intercept are based on 95% confidence limits. The 95% confidence bands plotted on the graph show that the samples are not too erratic. Since the slope is statistically less than 8, at the 5% significance level, the precipitation probably occurs under nonequilibrium conditions. A slope of less than 8 could also indicate that some of the water in the clouds is water evaporated from land surfaces. The slope of  $6.03 \pm 1.27$  observed is similar to the slope of 5.80 observed in Dixie Valley, Nevada (Ingraham, 1982), approximately 100 miles east of Eagle Valley.

Plots of elevation versus deuterium (Figure 6.2) and elevation versus  $\delta 18/16$  (Figure 6.3) show a definite

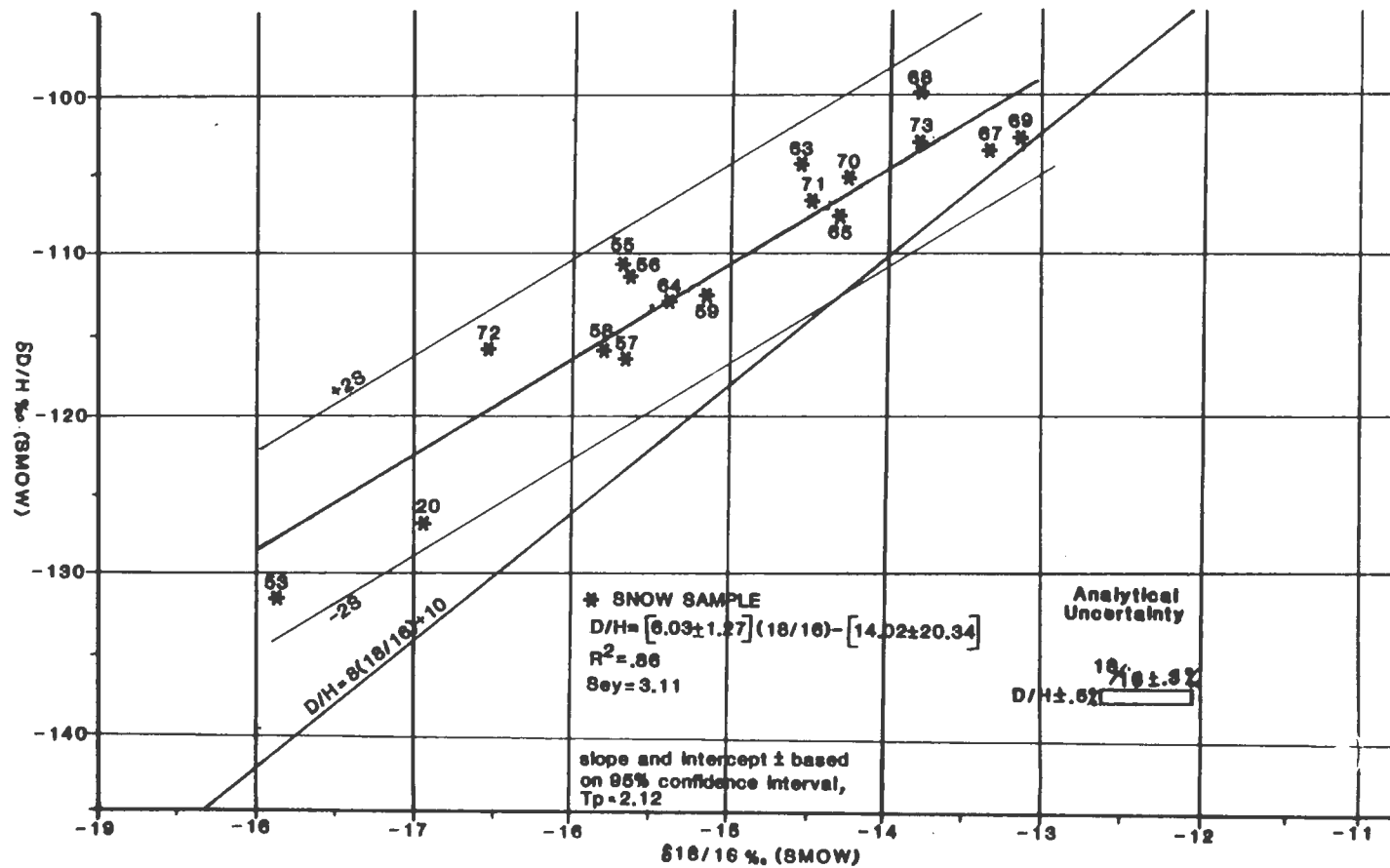


Figure 6.1  $\delta D/H$  versus  $\delta^{18}O/^{16}O$  for Snow Sampled in Spring, 1981

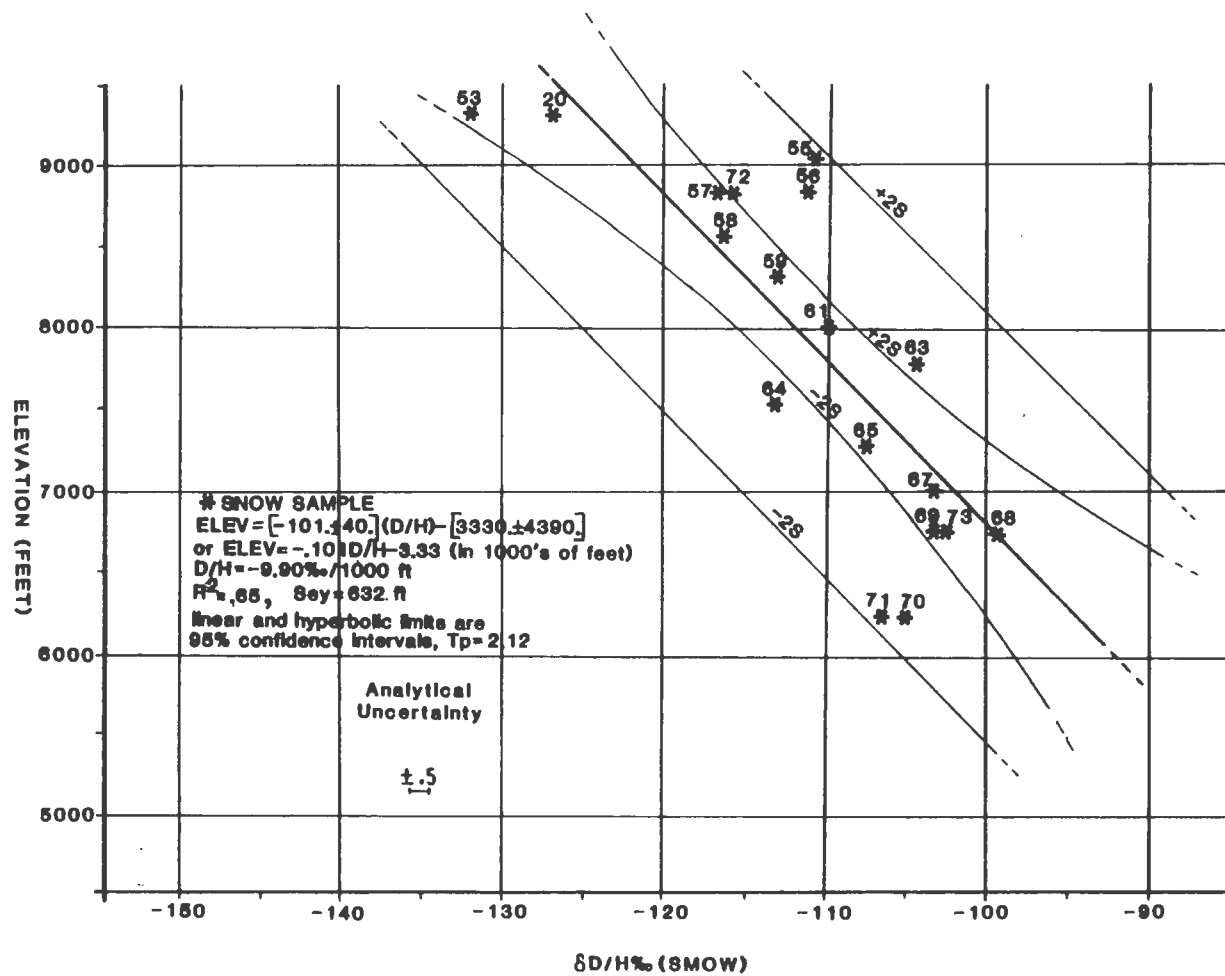


Figure 6.2  $\delta D/H$  versus Elevation for Spring, 1981 Snow Samples

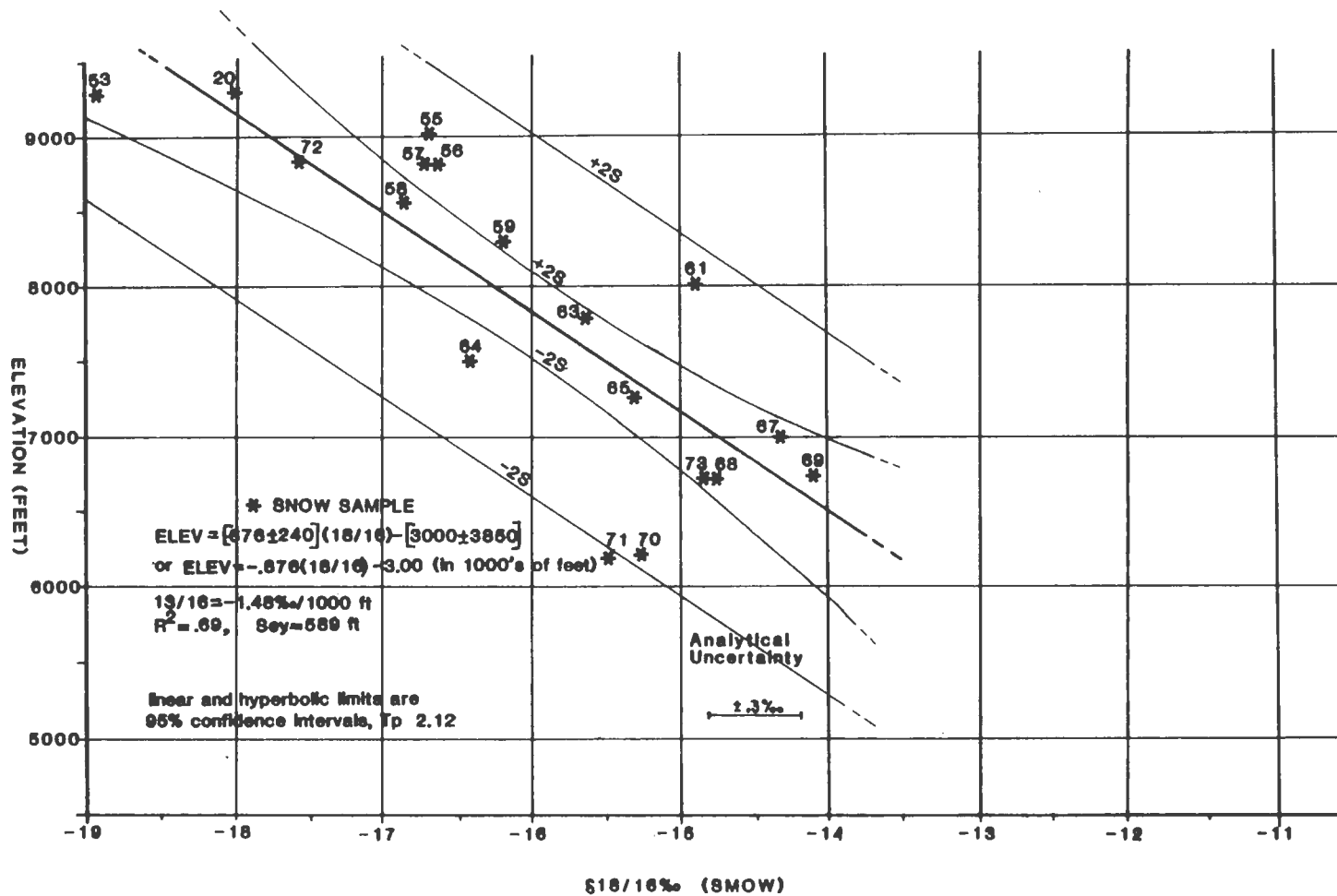


Figure 6.3  $\delta^{18}/16$  versus Elevation for Spring, 1981 Snow Samples

depletion of heavy isotopes with increasing elevation. This is contrary to two previous studies which concluded that there is no elevational control of isotopes on the east slopes of the Sierras (Friedman and Smith, 1970; Smith, Friedman, Klieforth, and Hardcastle, 1979). Slopes obtained are  $\delta D/H = -9.90\% / 1000 \text{ ft}$  ( $R^2 = .65$ ) and  $\delta^{18}/^{16} = -1.48\% / 1000 \text{ ft}$  ( $R^2 = .69$ ), which indicate that regular isotopic fractionation can occur on the leeward side of mountains. Both slopes show correlation at the 5% significance level. These values are reasonable compared to other studies; the west slope of the Sierras have an isotopic slope of  $\delta D/H = -12\% / 1000 \text{ ft}$  (Smith, Friedman, Klieforth, and Hardcastle, 1979), and a study in Italy (Fritz and Fontes, 1980) observed  $\delta^{18}/^{16} = -1.6\% / 1000 \text{ ft}$  (see Section 3.2.3). Also plotted on isotopic elevational shift graphs are 95% standard and hyperbolic confidence bands; the hyperbolic bands are used when the ordinate is used to predict the abscissa (Yevjevich, 1972). The hyperbolic bands show that near the center of mass of the data, the control is better, whereas near the ends of data the confidence band is much wider. Deuterium 95% hyperbolic confidence band is  $\pm 350 \text{ ft}$  at best; the oxygen 95% hyperbolic confidence band is  $\pm 300 \text{ ft}$  at best.

These isotopic data are based on one sampling season, the 1980-81 winter, which was warmer and had less precipitation than average (about 60% of normal). To get some idea of what the average isotopic values should be, the



temperature difference between the year sampled and the average year must be known. Data for the average monthly temperatures from four stations for the last 30 years (in Section 2.2.2) indicate that the 1980-81 winter season (October to May) was 2.65°F (1.47°C) warmer than the average. Referring to Figure 3.2, temperature verses  $\delta D/H$  or  $\delta 18/16$ , using the solid-liquid (precipitation) line, one can estimate how much change 1.5°C would have on isotopic composition. Isotopic values for the average, corrected for the temperature change are:

ELEVATION (feet)	$\delta D/H$ OBSERVED	$\delta D/H$ THEORETICAL	$\delta 18/16$ OBSERVED	$\delta 18/16$ THEORETICAL
9000	-122.	-135.	-17.2	-18.7
8000	-111.	-124.	-16.2	-17.8
7000	-102.	-115.	-14.8	-16.6
6000	-92.	-105.	-13.2	-14.7

In other words, average (theoretical)  $\delta D/H$  values would be about 13% more depleted, and  $\delta 18/16$  would be about 1.6% more depleted than the values observed in the 1980-81 year.

In an attempt to check different sample handling techniques, one sample (#57) was filtered through a small brass screen to remove dirt (analyzed as sample #72), while another sample (#69) was filtered through a .45 $\mu$  millipore filter (analyzed as sample #73). In both cases, filtration resulted in a -0.8% shift for oxygen-18, while deuterium essentially did not shift. This indicates that oxygen

isotopes are probably more sensitive to fractionation by sample handling than deuterium, so one should not filter melted snow samples since it will produce an isotopic change.

#### 6.1.2 Radioisotope Chemistry

Tritium values in precipitation from the closest available location, Las Vegas, are shown in Figure 3.3. This is used in Section 6.5.5.1 to approximate the age of the groundwater in the Eagle Valley. Four samples of precipitation, and two stream samples were taken in the 1980-81 year in Eagle Valley. Tritium levels of precipitation in the Carson Range were  $34 \pm 5$  TU at 9214 ft,  $21 \pm 4$  TU at 8250 ft, and  $21 \pm 4$  TU at 7250 ft. A snow sample from 7330 ft in the Virginia Range north of Carson City contained  $37 \pm 5$  TU (tritium data is in Appendix 3). These precipitation samples suggest that tritium in Eagle Valley is almost down to natural pre-bomb tritium levels of 5-20 TU (Payne, 1972).

### 6.2 CHEMISTRY OF INFILTRATED WATER

#### 6.2.1 Major Ion Chemistry

Chloride and electrical conductivity analyses were performed on snow samples taken in April, 1981, and on soil water samples taken in June and July, 1981. Results are shown in Table 6.1. Percent recharge to the soil zone for samples from each elevation zone was calculated using the

ELEVATION (ft)	SNOW *		LYS. WATER +		% RECHARGE	
	Cl(ppm)	EC	Cl(ppm)	EC	Cl	EC
8500	0.2	18.9	0.4	65.3	60.	30.
			0.3	61.8		
			0.3	63.0		
8000	0.1	17.2	0.9	101.0	11.	17.
7500	0.1	9.8	2.5	160.0	4.0	6.1
7000	0.1	12.6	9.0	253.0	1.1	5.0
6250	0.4	15.6	5.1	393.0	7.8	4.0

\* Snow samples taken 4-14-81

+ Lysimeter water samples taken 6-11,6-25,7-9-81

Table 6.1 Chloride and Electrical Conductivity Data

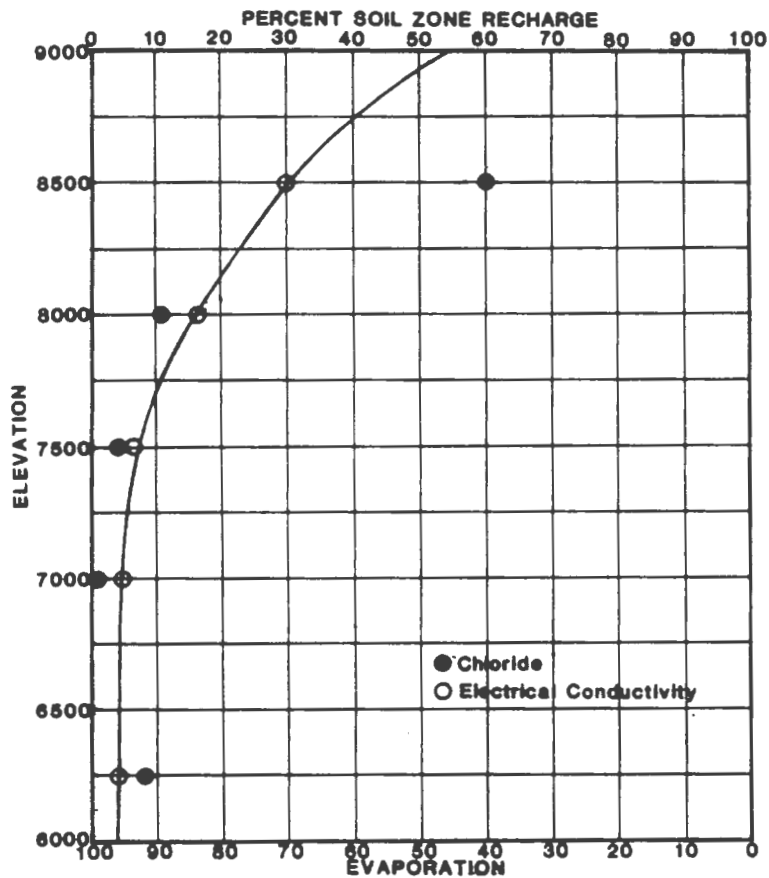


Figure 6.4 Percent Potential Soil Zone Recharge versus Elevation

formula from Section 3.1. The percent recharge versus elevation for the 12 samples is shown graphically in Figure 6.4. This is used in conjunction with planimetered elevation zones and precipitation data (Table 6.2) to produce Figure 6.5, an estimate of the volume of potential soil zone recharge from different elevation zones, assuming no direct runoff. Figure 6.5, shows that the greatest amount of water comes from the 8500 ft elevation zone. Summing up the volumes in each elevation zone gives a total annual recharge in the Carson Range of 4000 acre-ft, or 8.5% of annual precipitation, close to the official estimate of 6500 acre-ft/year for all Eagle Valley (U.S. Geological Survey, 1981). Using the chloride concentration change from snow (0.1 ppm) to groundwater in Eagle Valley (1.2 ppm), the total annual recharge is estimated at 3900 acre-ft, or 8.3% of annual precipitation. These estimates are less than the recharge estimate of 8500 acre-ft/year, or 15% of annual precipitation for the Carson Range calculated by Worts and Malmberg (1966). This discrepancy is discussed further in Section 6.4, Recharge Water to the Aquifer.

The average recharge elevation can be calculated by:

$$\frac{\sum(\text{volume recharged in each zone})(\text{elevation})}{\sum[\text{volume recharged in each zone}]} = \text{weighted average recharge elevation.}$$

The average recharge elevation is approximately 7950 ft. This indicates that high elevations are very important as potential recharge areas, and should not be altered.

ELEVATION ZONE (ft)		AREA (ft <sup>2</sup> )	PRECIP (ft)	%RECHARGE	VOLUME RECHARGED (FT <sup>3</sup> )
9000	8750 - 9250	2.235x10 <sup>7</sup>	2.92	0.53	35.6x10 <sup>6</sup>
8500	8250 - 8750	7.127x10 <sup>7</sup>	2.67	0.30	57.1x10 <sup>6</sup>
8000	7750 - 8250	8.421x10 <sup>7</sup>	2.46	0.15	31.1x10 <sup>6</sup>
7500	7250 - 7750	10.45x10 <sup>7</sup>	2.17	0.05	11.4x10 <sup>6</sup>
7000	6750 - 7250	11.47x10 <sup>7</sup>	2.04	0.05	11.70x10 <sup>6</sup>
6500	6250 - 6750	11.67x10 <sup>7</sup>	1.83	0.03	6.42x10 <sup>6</sup>
6000	5750 - 6250	15.48x10 <sup>7</sup>	1.60	0.02	4.96x10 <sup>6</sup>
5500	5250 - 5750	26.52x10 <sup>7</sup>	1.33	0.02	7.02x10 <sup>6</sup>
5000	4750 - 5250	26.68x10 <sup>7</sup>	1.17	0.02	6.24x10 <sup>6</sup>
					total 171.5x10 <sup>6</sup> ft <sup>3</sup>
					or 4000 acre-ft/year

Table 6.2 Potential Recharge Data

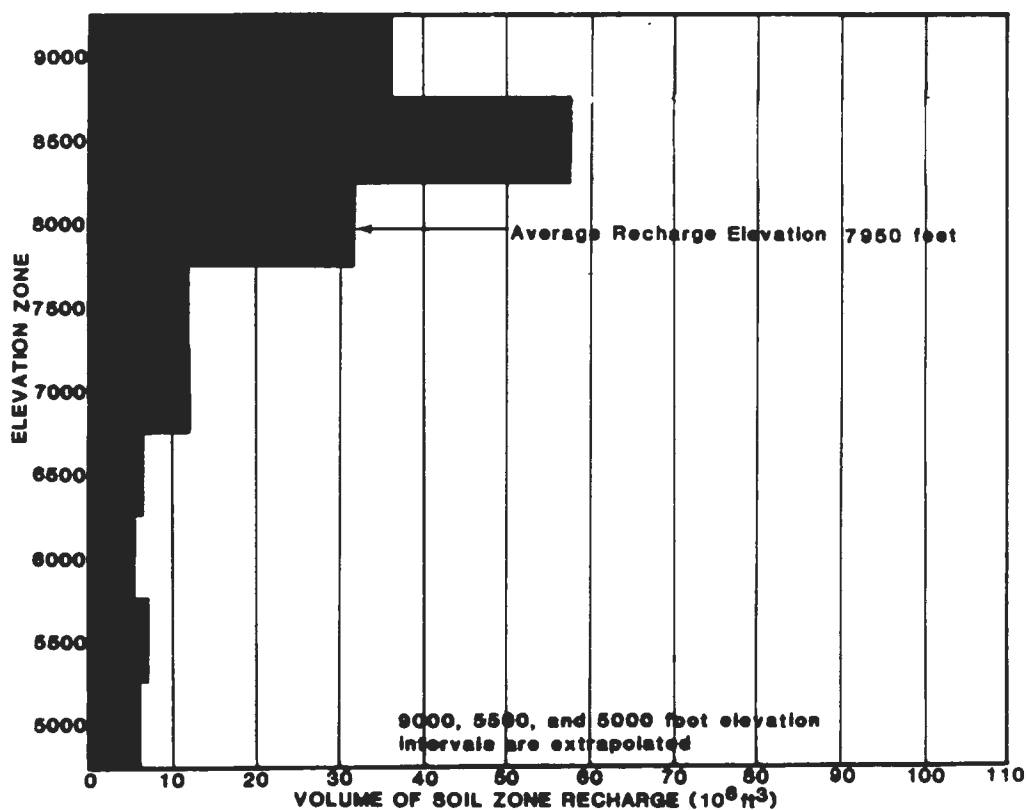


Figure 6.5 Volume of Potential Soil Zone Recharge versus Elevation

A significant amount of recharge to the aquifer is probably from stream channel infiltration at lower elevations, so precipitation at higher elevations eventually discharges to the stream either by direct runoff or soil zone interflow. Therefore, soil zone recharge calculations are somewhat inaccurate due to additional evaporation in the stream channel, and these recharge calculations should be considered potential recharge.

#### 6.2.2 Stable Isotope Chemistry

6.2.2.1 Snow to Lysimeter Water. In an attempt to monitor the amount and timing of recharge, five lysimeters installed from 8500 ft down to 6200 ft in Ash Canyon were monitored from May, 1981, until May, 1982. The lysimeter site at 8500 feet provided a good estimate for the timing of recharge. This site is in a meadow, a local discharge area, presumably from recharge at elevations from 8500 to 9200 ft. An open hole dug to a depth of three feet next to the lysimeter showed the water level at the ground surface from May, 1981, until late August, 1981, indicating that local discharge in the meadow took at least several months after complete melting of the snow, which had occurred by the end of May. The water level in the open hole gradually declined to 18 inches below ground level by late September. By this time, the volume of water obtained in the other lysimeters had also declined.

Monitoring of the lysimeters in late November, January, March, and April showed that, although the soil beneath the snow was not frozen, infiltration was probably very small, since the lysimeters had little or no water, and the ground was not very moist. By March, 1982, the ground at lower elevations (7500 ft and below) was becoming increasingly saturated, with seeps at 7500 ft. By April, the ground was saturated at the lysimeter located at 8000 ft.

In conclusion, it appears that recharge may occur year-round since the ground is not frozen, but it definitely occurs beneath snow pack in the spring and summer. A study in permafrost areas in Canada showed that seepage existed under ice, since tritium was found at depths of several meters below the ground surface (Fritz and Fontes, 1980); therefore, infiltration could occur year-round, albeit at a reduced rate in the winter.

Water infiltrating into the soil zone beneath snow will shift isotopically over time toward a heavier (enriched) composition, as the melting snow evaporates and infiltrates. This is illustrated by Figure 6.6 in which the isotopic enrichment of snow samples over time are plotted against lysimeter samples. Notice that the lysimeter water composition (#96, #119) is in between the original snow isotopic composition (#58) and evaporated snow (#75, #95), indicating that lysimeter water is a mixture of water being

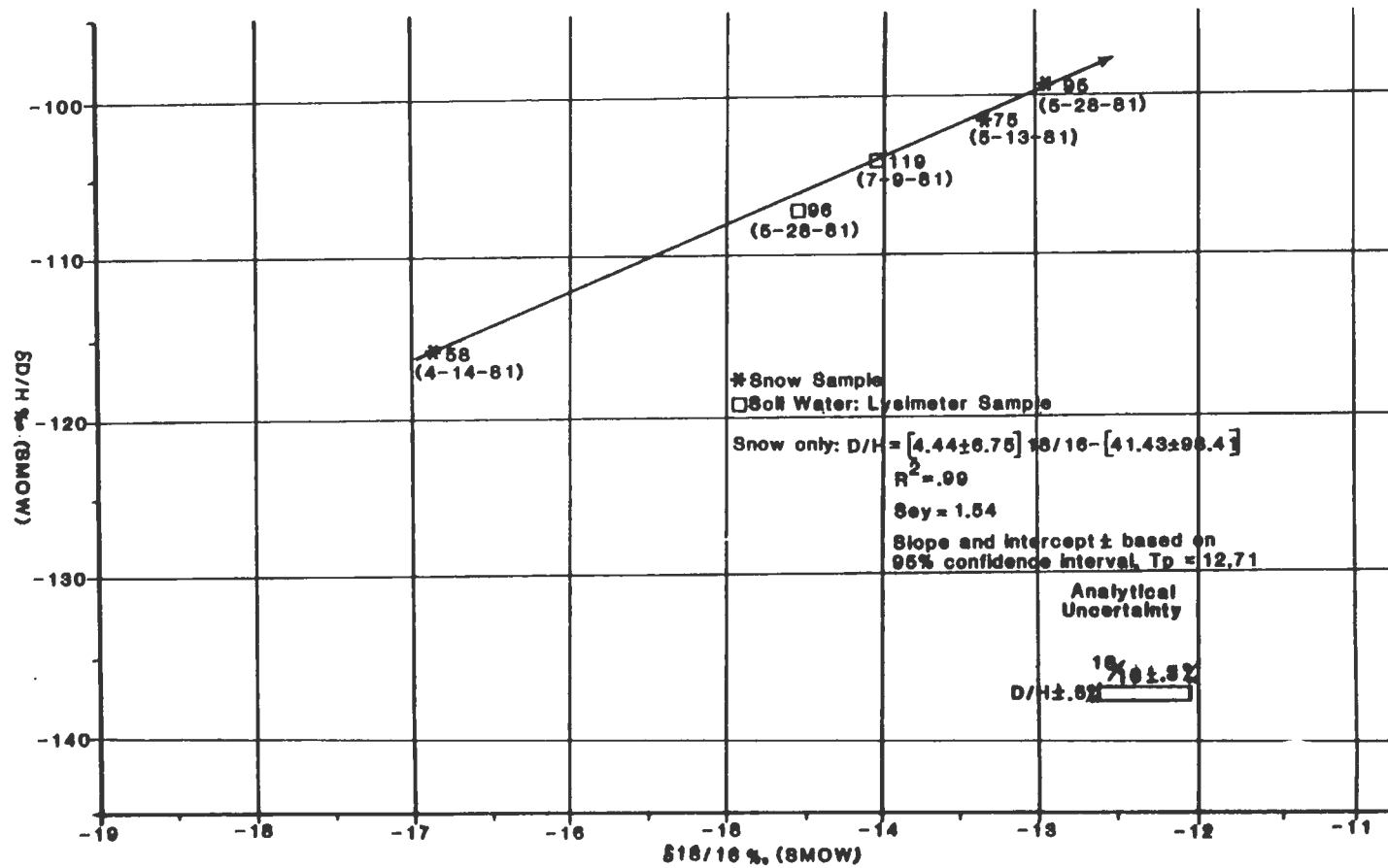


Figure 6.6 Snow to Soil Water Isotopic Fractionation at 8500 ft in Ash Canyon



continuously recharged from the snow. For example, lysimeter sample #96 was sampled at the same time as snow sample #95 and is isotopically lighter, which suggests it was recharged earlier than the overlying snow.

Equation 3, derived in Section 3.4.3, which assumes equilibrium evaporation, was used to calculate the total potential recharge from snow sample #58 (sampled in April, 1981) to lysimeter sample #95,  $\theta_d = 0.98$  for deuterium, and  $\theta_{ox} = 0.996$  for oxygen-18. These values are unusually high compared to values of recharge calculated with electrical conductivity ( $\theta = 0.30$ ).

Using only the Rayleigh distillation formula for evaporation one needs to know  $\alpha$ . Using average monthly temperature data from Mt Rose Bowl (8250 ft, Appendix 5), for May, 1981,  $T = 6.1^\circ\text{C}$ , one can then use Figure 3.1 ( $\alpha_d$  and  $\alpha_{ox}$  versus temperature) to estimate  $\alpha_d = 1.104$ , and  $\alpha_{ox} = 1.1011$ . Using these values in the Rayleigh evaporation equation,  $\theta_d = .814$ , and  $\theta_{ox} = .764$ , which are much more reasonable values; about 80% of the snow remained, 20% evaporated.

Since the combined Rayleigh/mass balance equation did not work, the fault must lie in the mass balance equation. Note that both of the above equations assume Rayleigh conditions with equilibrium evaporation, which is not usually the case in nature. Unfortunately, the Rayleigh equation is the closest approximation available.

A plot showing the isotopic shift from snow to lysimeter water is shown in Figure 6.7. Note that the slopes decrease from 4.1 at 8500 ft to 1.9 at 7500 ft, and then become negative. The negative slopes for 7000 and 6250 ft imply depletion of deuterium, an impossible situation for evaporation. Possible reasons for this include: 1) when snow samples were taken in April, 1981, snow was patchy below 7200 ft, so snow at low elevations had already melted and evaporated; 2) groundwater samples at lower elevations may represent water infiltrated at higher elevations. The decrease in slopes from 4.1 to 1.8 to 1.9 as elevation decreases from 8500 to 7500 ft may be due to faster evaporation at lower elevations, which produces lower slopes (see Section 3.2.4). This is a reasonable conclusion, since lower elevations are warmer.

The isotopic enrichment from snow to lysimeter water can be used to calculate recharge to the soil zone:

ELEV. (ft)	SNOW	LYS	T°C	$\alpha_d$	$\theta_d$	$\alpha_{ox}$	$\theta_{ox}$
8500	58	119	6.1	1.104	.87	1.011	.77
8000	61	103	7.2	1.102	.93	1.0108	.75
7500	64	121	8.3	1.100	.94	1.0106	.79
7000	67	122	9.2	1.098	1.10	1.0105	.65
6250	71	123	10.0	1.096	1.11	1.0104	.97

These data indicate several things. First, lower elevations do not appear to have a lower percent recharge, which is contrary to the usual assumption. Second, at the 7000 ft and 6250 ft elevations, calculations indicate

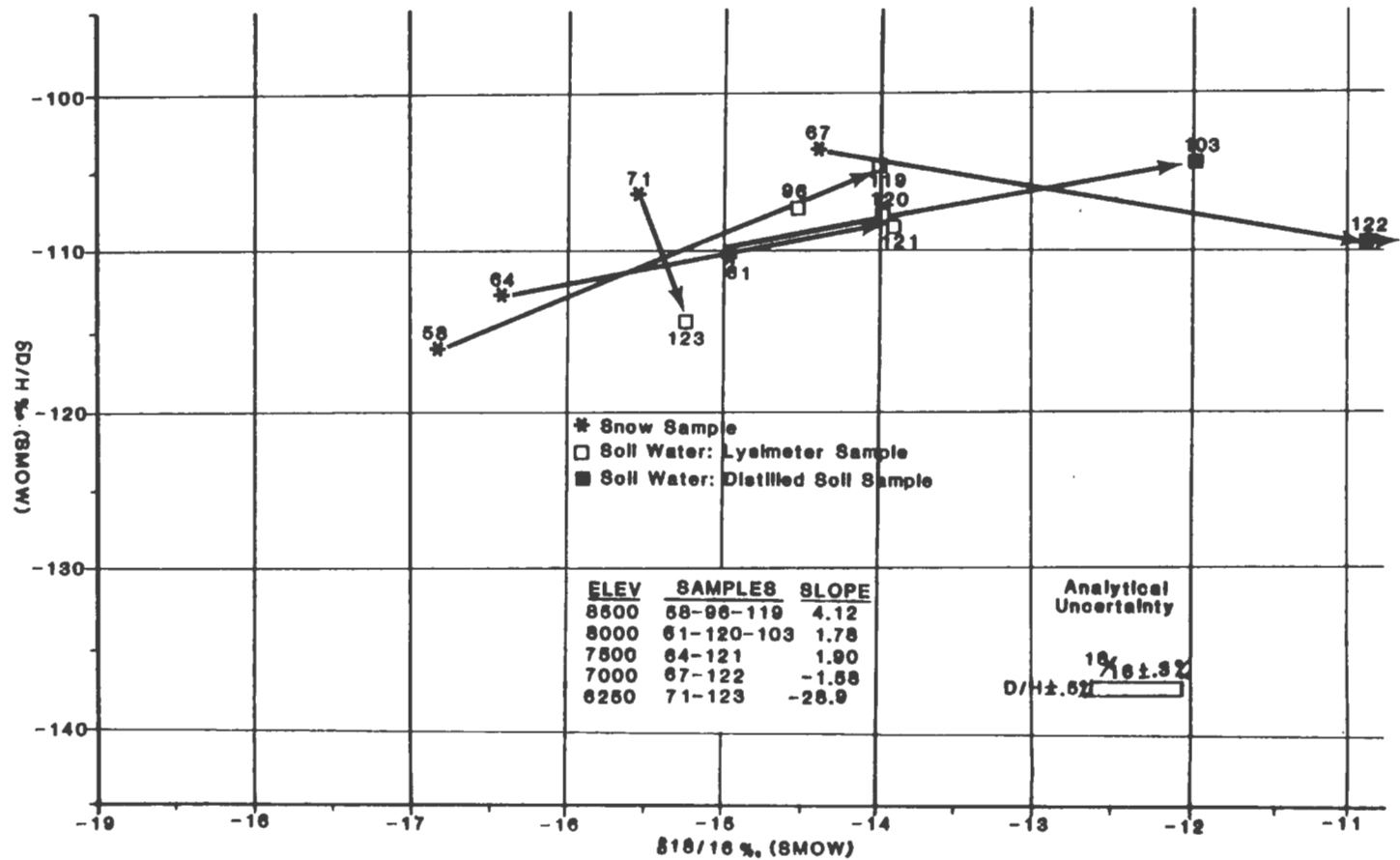


Figure 6.7 Isotopic Shift from Snow to Soil Water in Ash Canyon

inaccuracy, since recharge is over 100%. As discussed earlier, it may be that the snow or lysimeter water is not indicative of that elevation.

6.2.2.2 Lysimeters. Figure 6.8 (Soil Water  $\delta D/H$  versus Elevation) and Figure 6.9 (Soil Water  $\delta 18/16$  versus Elevation) reveal several characteristics. First, the  $\delta D/H$  isotopic shift is very regular ( $R^2 = 0.99$ , samples significantly correlated at the 5% level), and it appears that distilled soil samples are of similar hydrogen isotope composition as lysimeter water samples. However, the lysimeter isotopic trend indicates depletion at lower elevations, the opposite trend indicated by snow and most other isotope studies. The deflection is probably caused by lysimeter locations: they were in a canyon with some lysimeters in springs, so the water could be derived from higher elevation snow-melt. In theory, lysimeters along a ridge would be more likely to sample snow-melt from the same elevation. The soil water trend could also indicate isotopic exchange with water vapor in the air at lower elevations, which would deplete lower elevation soil water compositions.

Figure 6.9 ( $\delta 18/16$  versus Elevation) shows an isotopic trend with elevation which exhibits fair correlation ( $R^2 = 0.78$ , samples are not significantly correlated at the 5% level). In this case, distilled soil samples are highly fractionated compared to lysimeter oxygen isotope samples.

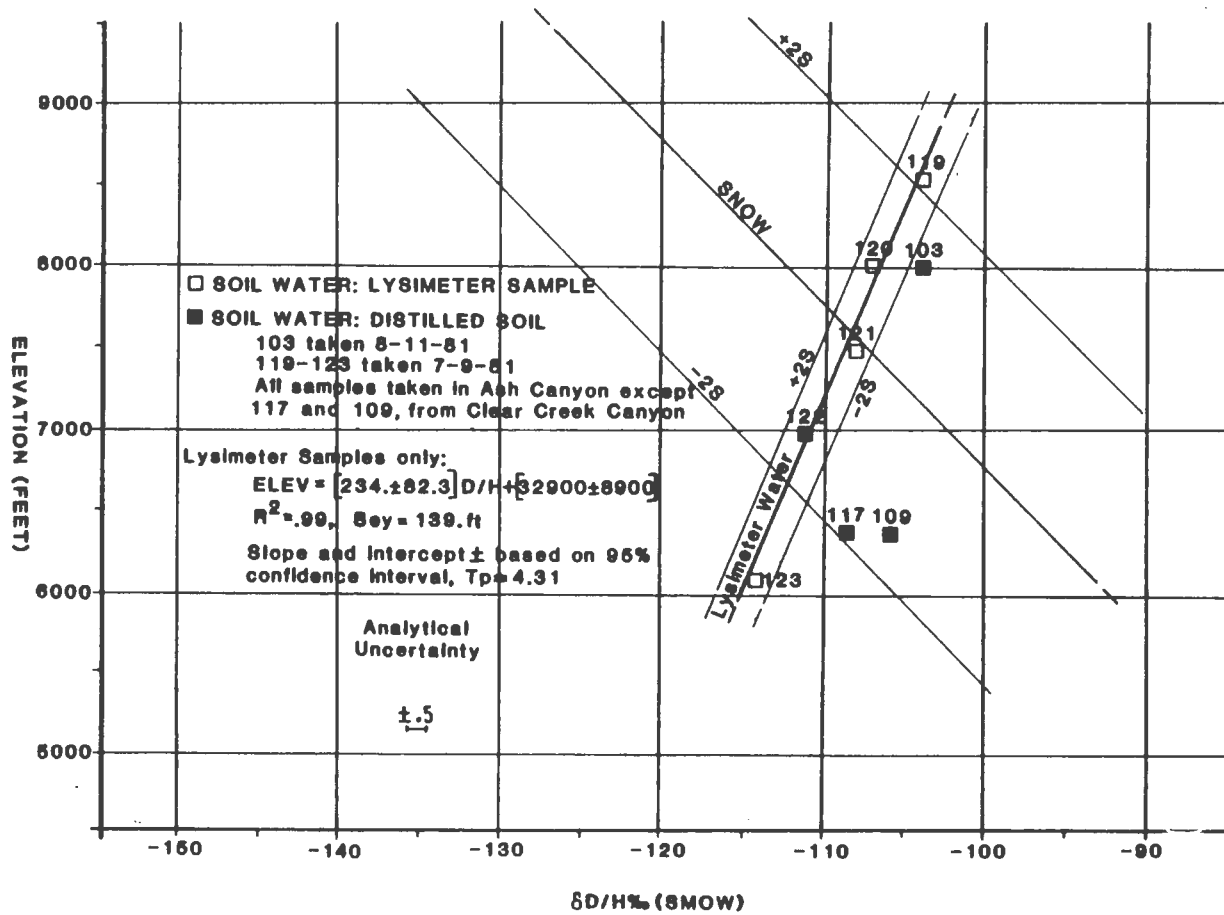


Figure 6.8 Soil Water  $\delta D/H$  Elevational Shift

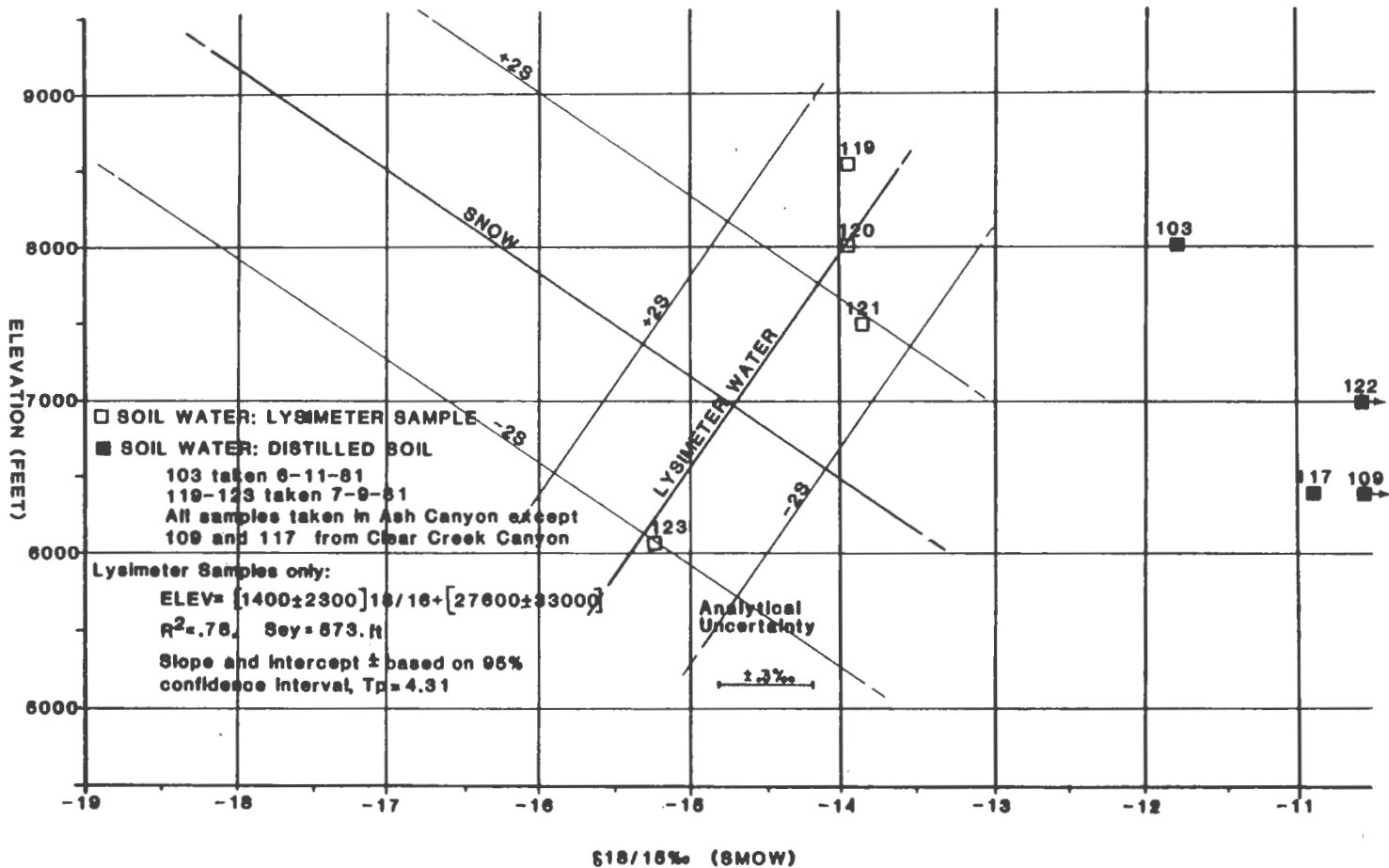


Figure 6.9 Soil Water  $\delta^{18}/16$  Elevational Shift

This means that oxygen isotope fractionation occurs in the unsaturated zone or in the soil distillation apparatus (more so than deuterium). Oxygen isotopes are also depleted with lower elevation, as is deuterium.

The soil samples were distilled at approximately 260°C, a temperature which vaporizes all soil water, clay-lattice water, and possibly some organic compounds known to be part of soil, such as fulvic acid, humin, humus, and humic acid (Bohn, McNeal, and O'Conner, 1979). Organic matter in shallow soils is usually about 0.5 to 5.0% by weight. These organic compounds may assimilate some oxygen from the atmosphere, which has a  $\delta^{18}/^{16}$  of +41‰. Therefore, if organic compounds were distilled with water, the net effect would be enrichment of the oxygen isotopic value of the sample, as was observed. This phenomena was not investigated further in soil water, but was investigated with tree water samples, discussed in the next section.

6.2.2.3 Tree Water. Analyses of the stable isotope composition of tree xylem water are presented in Figure 6.10 ( $\delta D/H$  versus elevation) and Figure 6.11 ( $\delta^{18}/^{16}$  versus elevation). The data and types of trees sampled are listed in Appendix 2. The species of trees sampled were Jeffery pine (*Pinus jeffreyi*), Yellow pine (*Pinus ponderosa*), and Lodgepole pine (*Pinus contorta*). Referring to both graphs, one notes that tree samples are identified as distilled

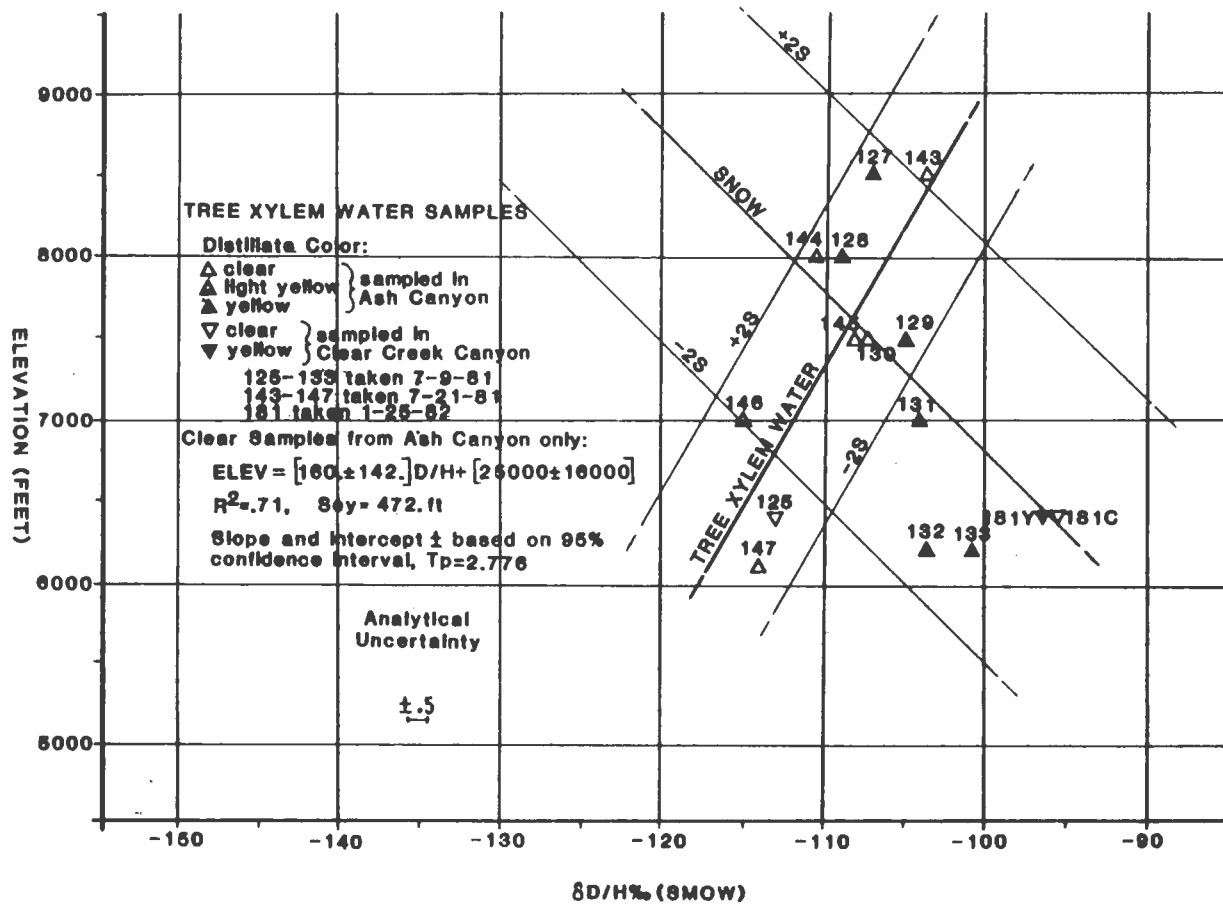


Figure 6.10 Tree Xylem Water  $\delta D/H$  Elevational Shift



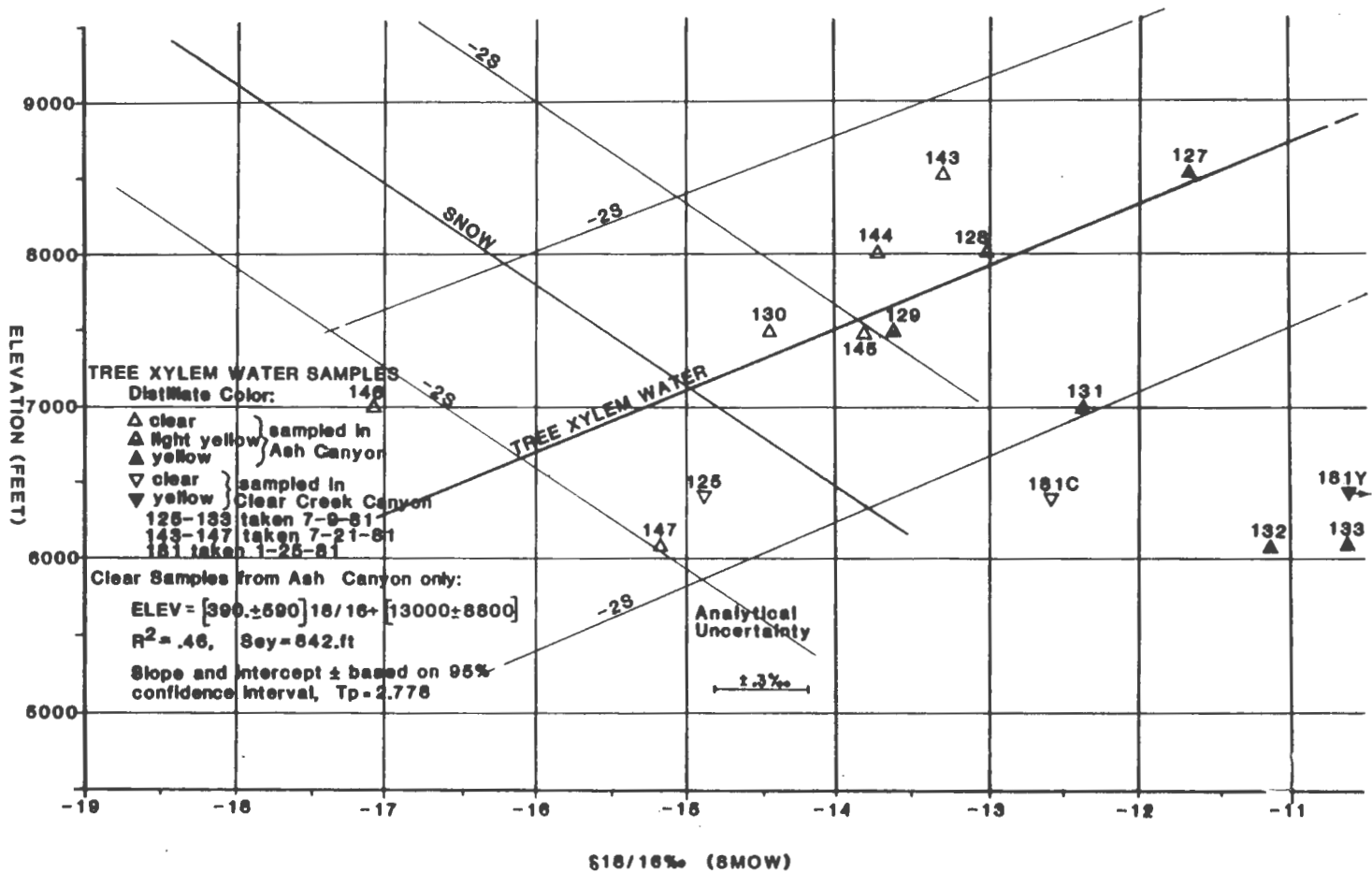


Figure 6.11 Tree Xylem Water  $\delta^{18/16}$  Elevational Shift

clear, light yellow, or yellow. The color of the distilled liquid is directly related to the maximum temperature to which the wood chips were heated in the distillation apparatus. Clear samples were heated slowly to a maximum of 160°C (40 volts), light yellow samples were heated to 190°C (45 volts), and yellow samples were heated to 250°C (55 volts). The length of time taken to distill samples also affected distillate color; samples heated for more than five hours even at 160°C would turn yellow. The yellow immiscible fluid is probably an organic compound(s) which has a higher boiling point than water, which distills to a greater extent at a higher temperature. The amount of yellow fluid in a sample has a dramatic effect on the isotopic ratio of samples.

Referring to Figures 6.10 and 6.11, deuterium analyses show that yellow samples are somewhat enriched isotopically compared to clear samples; whereas oxygen isotope analyses show yellow samples as significantly enriched compared to clear samples. To test the hypothesis of the yellow liquid changing the isotopic composition, one sample was distilled clear (#181C) and yellow (#181Y). Deuterium analyses of both samples indicate the same isotopic ratio within analytical error, but oxygen isotopic analysis revealed that the yellow sample is significantly enriched (+3.3%) compared to the clear sample. Therefore, only clear samples were used to correlate tree xylem water with elevation.

The isotopic shift produced by the yellow liquid is associated with the synthesis of the organic compound in the tree. Carbon-13 (a stable isotope) fractionation during photosynthesis has been observed (Osmond, 1981), so one might hypothesize that oxygen isotope fractionation could also occur during an organic reaction. Another possibility is that some of the oxygen in the organic compound is from plant cellulose, which uses  $\text{CO}_2$  from the air. Atmospheric  $\text{CO}_2$  has a  $\delta^{18}/16$  of +4‰, thus some yellow compound mixed with xylem water would significantly enrich the oxygen isotopic composition as observed.

Referring to clear samples #147 (6250 ft in Ash Canyon, Jeffery Pine), #125 (6400 ft in Clear Creek Canyon, Yellow Pine), and #181C (6400 ft in Clear Creek Canyon, Jeffery Pine), it appears that: the same species of tree in different canyons in the same mountain range (181C, 147) have a different isotopic composition; and trees in the same area (125, 181C) also have different isotopic compositions. This implies either different sources of water or isotopic fractionation. More samples are needed in order to reach a valid conclusion.

To remove as many variables as possible affecting tree water, only clear tree samples in Ash Canyon were correlated with elevation. This results in a reasonably good correlation for deuterium in Figure 6.10 ( $R^2 = 0.71$ , samples were

significantly correlated at the 5% level), and a poor correlation for oxygen isotopes in Figure 6.11 ( $R^2 = 0.46$ , samples were not correlated at the 5% significance level).

There is a very close correlation in isotopic composition between clear tree water and lysimeter water in Ash Canyon (Figure 6.12 for  $\delta D/H$ , and Figure 6.13 for  $\delta^{18}/16$ ). One concludes that trees do not fractionate soil water taken in by the roots. It would be possible, therefore, to use tree water instead of lysimeter water for evaporation calculations. One advantage of sampling trees instead of lysimeters is that with trees one might be sampling a 50-year average of infiltrated water, rather than a single part of one year as sampled with lysimeters.

To estimate the age of tree water, ten tree cores were taken from a Jeffery Pine in Clear Creek Canyon (same tree as sample 181). This tree is approximately fifty years old, based on similar sized trees cored and dated in Ash Canyon by the U.S. Forest Service. The tree water was analyzed at  $33.0 \pm 10.4$  TU, where precipitation in the area is  $21$  to  $34 \pm 5$  TU. This first appears to suggest that the tree contains modern water, which was the same conclusion reached in a study of tree water by Wershaw, Friedman, Heller, and Frank (1966). In this study, investigators only used phreatophytes, and they concluded that the trees contained no more than a two to three year average of water due to

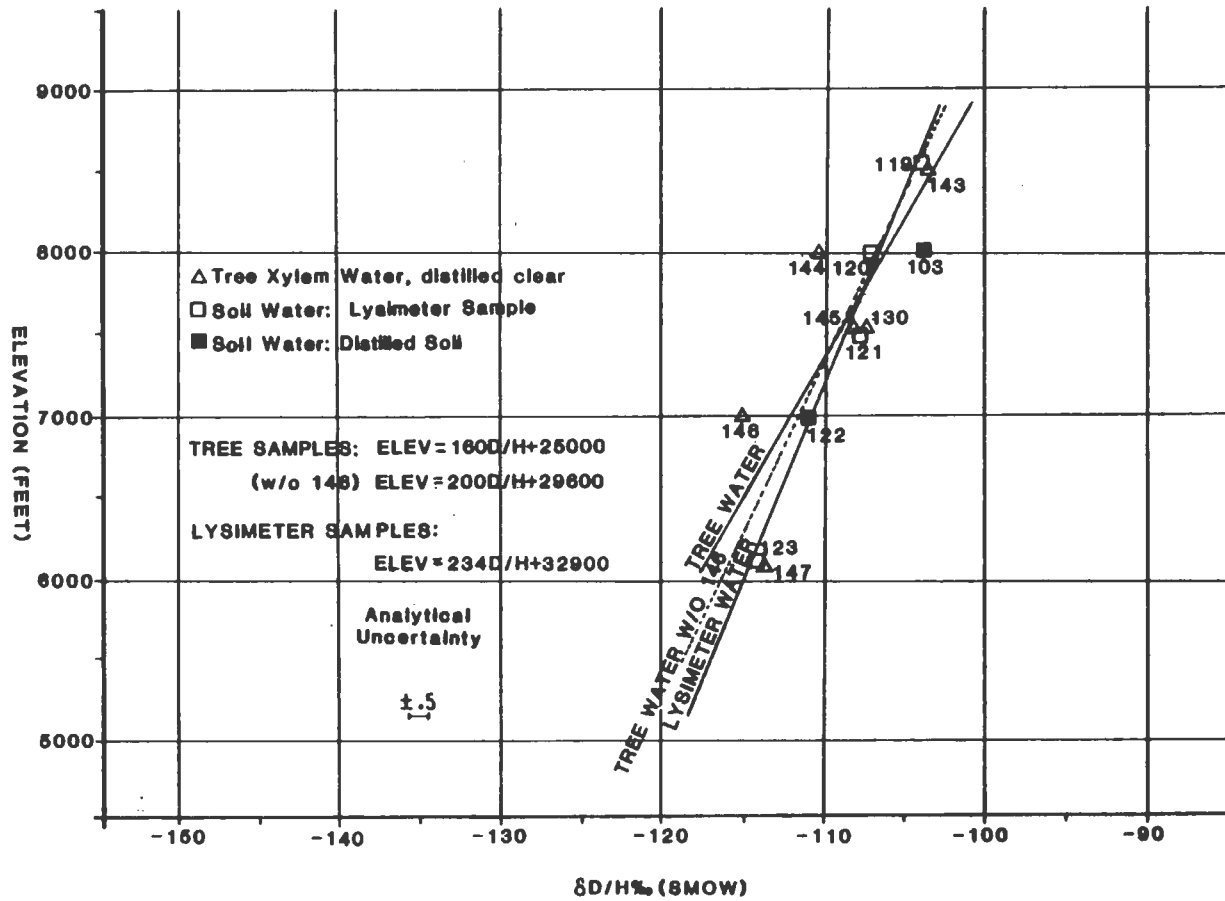


Figure 6.12 Comparison of Soil and Tree Water  $\delta D/H$  Elevational Shift

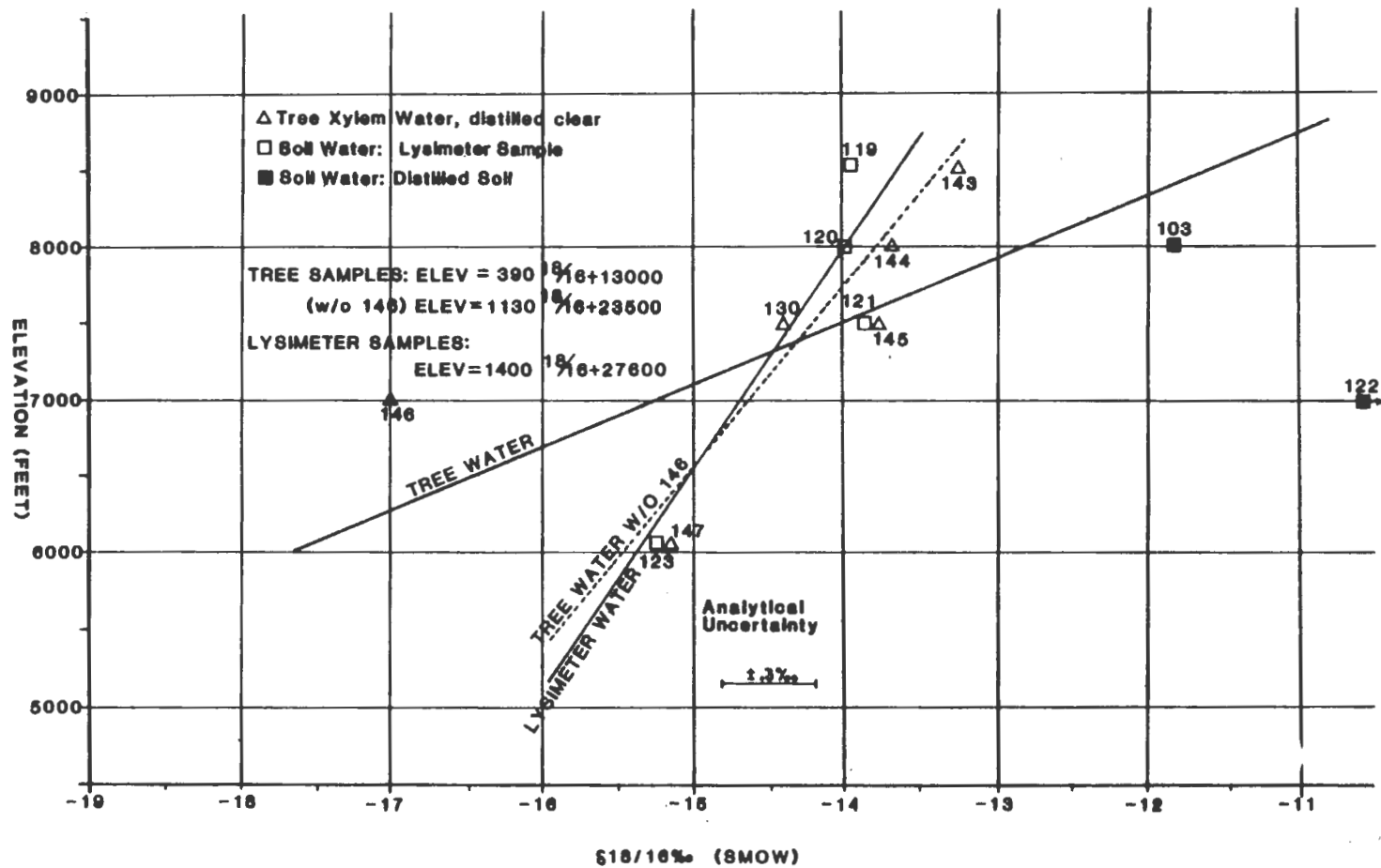


Figure 6.13 Comparison of Soil and Tree Water  $\delta^{18}/^{16}$  Elevational Shift

migration of water between the xylem and phloem.

However, the tritium levels may represent a 50-year average of yearly tritium inputs. Using decayed tritium values in precipitation for the past 28 years (Figure 3.3), and using 3 TU for the previous 22 years before 1953, the average tritium value for fifty years calculates to 40.2 TU. Therefore, 33. TU for the tree could represent modern levels or could represent a 50-year average. A tree would have to be cored vertically down older tree rings to restrict sampling to a 5-10 year core to determine the age of tree water.

By sampling tree xylem water, a number of conclusions have been reached, based on a limited number of samples: 1) tree xylem water is similar isotopically to soil water at the same location (trees do not fractionate soil water); 2) tree or lysimeter samples should be taken along a ridge to avoid sampling soil water that is from a higher elevation; 3) tree and soil water samples must be distilled at 160°C or lower to avoid distilling organics which will alter the distillate isotopic composition; 4) sampling different species of pine trees may affect isotopic composition; 5) tree samples in different canyons at the same elevation have different isotopic composition, possibly indicating different flow paths; and 6) tritium dating of tree xylem water could indicate modern water or a 50-year average tritium composite.

### 6.2.3 Radiocarbon Sampling in Recharge Areas

Three carbon-13 samples were taken from lysimeters in Clear Creek Canyon at 7200 ft, 6400 ft, and 6000 ft. The carbon dioxide level in each lysimeter was measured with Draeger gas indicator tubes. Soil zone CO<sub>2</sub> levels were 0.5%, 0.35%, and 0.6%. These high levels indicate root respiration, since CO<sub>2</sub> in the atmosphere is about 0.033%. The average P<sub>CO2</sub> in agricultural soils is 10 to 100 times atmospheric CO<sub>2</sub>, or 0.3 to 3.0% (Bohn, McNeal, and O'Connor, 1979), indicating the P<sub>CO2</sub> measured at the three lysimeters are reasonable values. The lysimeter at 6400 ft was also sampled for carbon-14, to get an original recharge area activity. Analysis of these samples did not produce modern C-14 and C13/12 since Sr(OH)<sub>2</sub> precipitated instead of SrCO<sub>3</sub>.

### 6.3 STREAMFLOW IN RECHARGE AREAS

The method by which water in recharge areas reaches the groundwater system in the valley is through a combination of infiltration, then interflow, baseflow, and overland flow toward the stream channel, and infiltration to the aquifer at lower elevations. Deep percolation or fracture flow to the aquifer from recharge areas is also a possible contributor to recharge. Computer modeling of water flow on a



convex hillslope (Freeze, 1972) indicates that overland flow and some subsurface flow (interflow) contribute to streamflow. The amount of interflow depends upon the soil hydraulic conductivity. The result of Freeze's work is that subsurface flow other than baseflow can play a major role in generating surface runoff if the conductivity is large.

Interflow of infiltrated water in the source area and discharge to the stream channel by interflow have been observed in Ash Canyon and Clear Creek Canyon. At 8500 ft in Ash Canyon, in the large meadow previously discussed, is a local discharge area. Several small springs flow into the creek until Fall, long after the snow (recharge water) has melted. A similar local discharge system is seen at 5500 ft in Clear Creek Canyon where a large marshy area contributes to the base flow in the summer and fall.

Along Ash Canyon Creek channel, from 8200 ft to 8000 ft, there are four small springs. From 7800 ft to 7000 ft in Ash Canyon, the canyon slope is considerably greater, which would increase the likelihood of springs. Of special interest is a horizontal row of six springs at 7440 ft, all within 1000 feet of each other. These may indicate a fault, although no faults are mapped in the Ash Canyon near this location (Trexler, 1977).

Stream gaging and sampling were done on Ash Canyon Creek in August, 1981, and January, 1982, to try to estimate

evaporation and separate streamflow groundwater input from phreatophyte use. Several trends are indicated by the August, 1981, stream gaging (Figure 6.14). For example, below 7800 ft, the stream loses flow. Total loss along the channel is 0.04 cfs between stations 3 and 4, 0.17 cfs between stations 6 and 7, and 0.27 cfs between stations 8 and 9, or 0.48 cfs in 2.4 miles or 0.2 cfs/mile. This loss could be due to infiltration into the alluvium, evaporation, and/or phreatophyte use.

The electrical conductivity from the August stream gaging shows a very regular trend increasing from 82.8 umhos/cm at 8150 ft to 131.9 umhos/cm at 5290 ft (conductivities normalized to 25.0°C). This could be from evaporation, ion increase from groundwater inflow, or dissolution of rock. Using a mass balance approach, one can calculate if a slug of water loses water but retains the same number of ions. Between stations 8 and 9, the conductivity remained essentially the same (130.0 to 131.9 umhos/cm), while the flow decreased from 0.89 cfs to 0.62 cfs. The conductivity at site 9 would have to be 187 umhos/cm for the flow change caused entirely by evaporation; therefore, conductivity trends suggest stream loss is not by evaporation. In addition, isotope data shows a change from  $\delta D/H = -103.\%$  to  $-102.\%$ , and  $\delta^{18}O/^{16}O = -14.0\%$  to  $-14.9\%$  from 8150 ft to 5290 ft, which again suggests no evaporation since there is no isotopic enrichment. Lack of any stable isotope change into

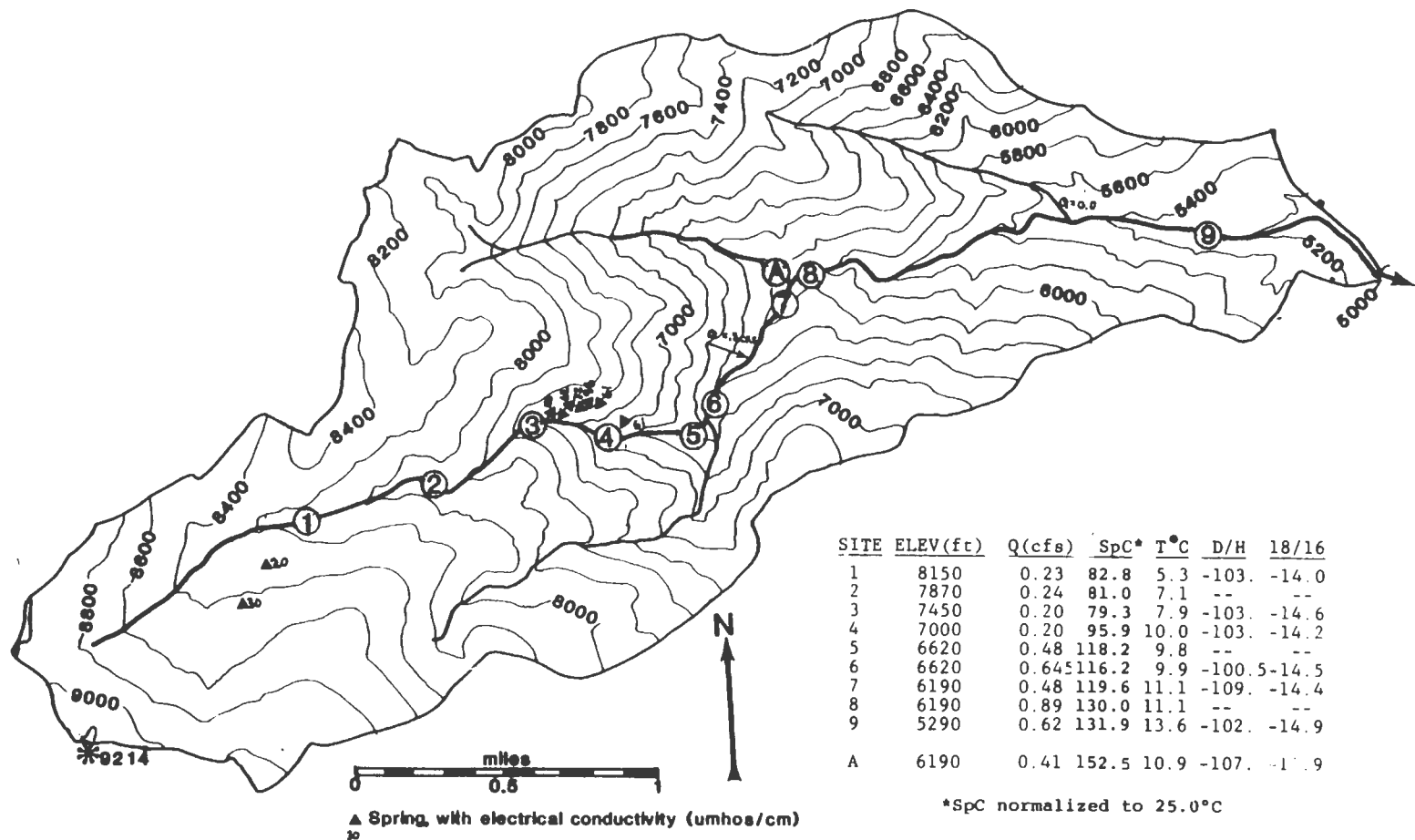


Figure 6.14 Results of August 27, 1981 Stream Gaging on Ash Canyon Creek

the stream indicates that all groundwater and surface water gains into the stream have essentially the same composition, which suggests that groundwater input to the stream at lower elevations must be recharged at higher elevations.

The conductivity and isotope data from the August stream gaging suggest that the 0.48 cfs total loss in Ash Canyon Creek is not due to evaporation. However, this does not indicate whether the loss is due to infiltration or phreatophyte use. Ash Canyon Creek was gaged again in January, 1982, the results are presented in Figure 6.15. Only lower reaches were gaged where August gaging showed a loss. Between 5620 ft and 6190 ft, January gaging showed an increase in flow of 0.12 cfs, versus a loss of 0.17 cfs in August. From 6190 ft to 5290 ft, January gaging showed a loss of 0.13 cfs (including side channel input) versus a loss of 0.27 cfs in August. This means essentially no gain or loss in January. The relatively high conductivity values from the January gaging (200-300 umhos/cm) indicate a substantial amount of baseflow. The fact that most streamflow loss occurs in the summer, rather than in the winter suggests that most of the loss was phreatophyte use, with groundwater recharge below 6000 ft possibly year-round, where both August and January gaging showed a loss.

Conclusions based on stream losses can only be used to suggest trends, since stream gaging in the small creek is

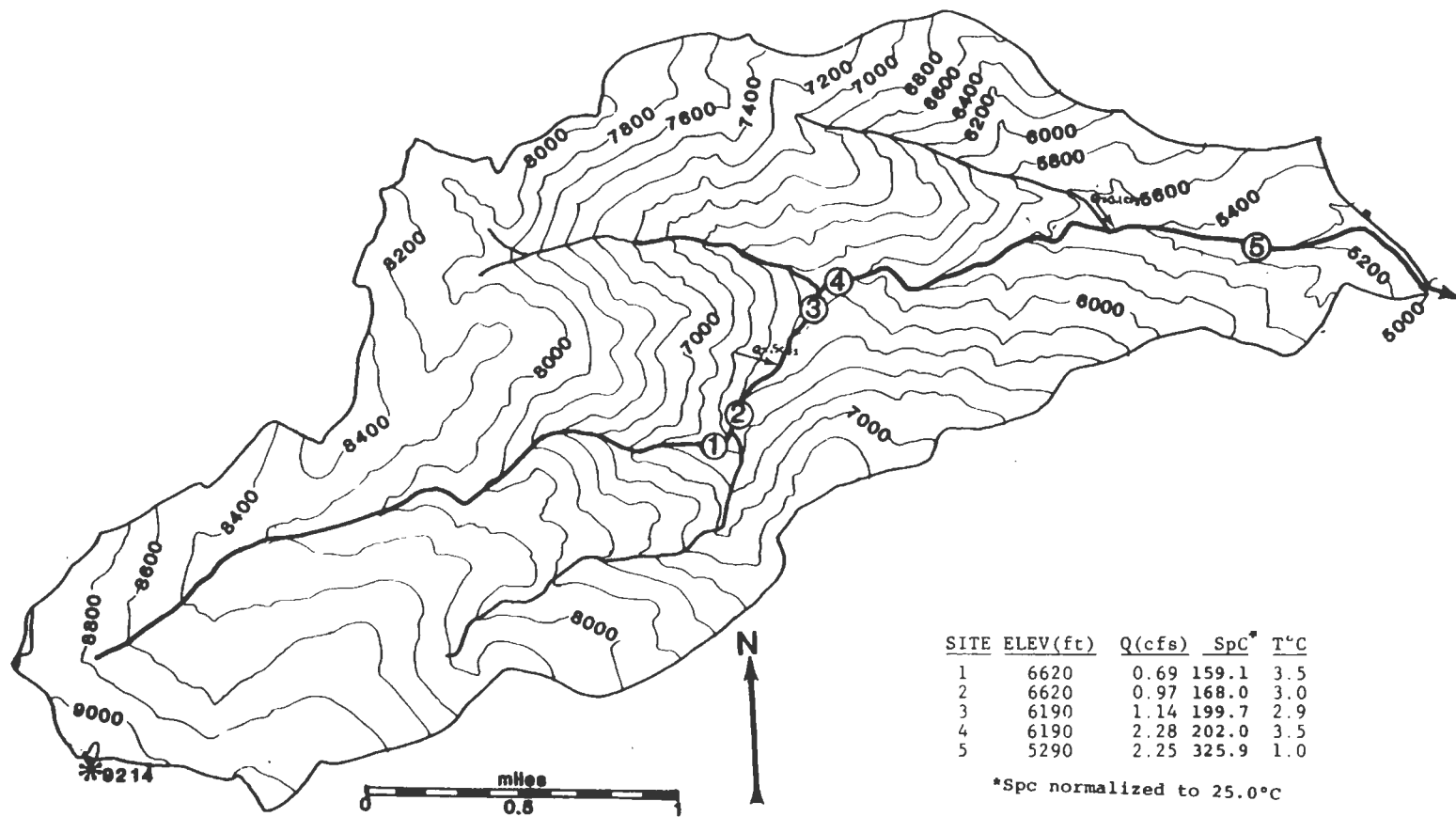


Figure 6.15 Results of January 29, 1981 Stream Gaging on Ash Canyon Creek

notably inaccurate. August stream gaging in the flume at 5290 ft with the transducer flow meter was within 1% of the calculated flow through the flume. January gaging at the same flume with a pygmy meter was 11% off the calculated flow. The pygmy meter does not work well in turbulent water, which may be the reason for the inaccuracy.

Two stream samples taken in August, 1981, were analyzed for tritium. One sample from 6620 ft was  $32.0 \pm 6.8$  TU, and another sample at 5290 ft was  $27.4 \pm 8.7$  TU. These levels have about the same activity as precipitation samples from the Carson Range, which vary from 21 TU to 34 TU.

#### 6.4 AQUIFER RECHARGE WATER CHEMISTRY

Electrical conductivity analysis of snow and lysimeter water in the Carson Range (Section 5.2.1) indicated an annual potential recharge of 4000 acre-ft/year. This is based on the assumptions of: no surface runoff; evaporation values are correct; and that all water infiltrates in recharge areas and becomes part of the aquifer. The latter two assumptions are questionable since infiltration in recharge areas is not recharge to the aquifer, and an insufficient number of samples were taken. In addition, analytical uncertainty is close to measured values.

The Maxey-Eakin method of estimating recharge is very similar; percentage of precipitation from different

elevation zones are used to calculate recharge. However, the Maxey-Eakin method was calibrated with calculation of discharge from basins, by assuming that annual recharge approximates discharge. This approach yields an annual recharge to Eagle Valley Basin of 8500 acre-ft. A comparison of the Maxey-Eakin recharge (Eakin and others, 1951) and results from conductivity recharge estimates from this study are shown below:

Maxey-Eakin Method		This Study	
Precipitation zone (in)	Percent recharge	Elevation zone (ft)	Percent recharge
>20	25	6200-9200	3-50
15-20	15	5400-6200	2
12-15	7	4800-5400	-
8-12	3	-	-
<8	0	-	-

The Maxey-Eakin method estimates 15% of the total precipitation is recharged, versus 8.5% estimated by the conductivity analysis in this study. One notes that the Maxey-Eakin method was calibrated in eastern Nevada, and as such does not adequately resolve percent recharge in high precipitation areas (over 20 inches), which comprise most of the Carson Range. The result is that the Maxey-Eakin method overestimates recharge in the Carson Range. Even though the conductivity analysis is inaccurate, it does show evaporation is inversely proportional to elevation, as does the Maxey-Eakin method.

Electrical conductivity analysis of snow and lysimeter water indicates that the average recharge elevation is 7950 ft. The isotopic composition of the 1980-81 precipitation at that elevation is  $\delta^{18}/^{16} = -16.2\%$ , and  $\delta D/H = -111\%$ . The isotopic enrichment from the snow to samples from Ash Canyon Creek sampled in mid-May, late May, and late August can be used to calculate evaporation with the Rayleigh evaporation formula:

Snow 7950ft ‰	T °C	$\alpha_d$ $\alpha_{ox}$	Stream 5-13-81 1-0	Stream 5-28-81 1-0	Stream 8-27-81 1-0
$\delta D/H = -111. \pm 3.5$	9	1.10	.000	.036	.094
$\delta^{18}/^{16} = -16.2 \pm 4$	9	1.0105	.145	.084	.193

Deuterium analyses show evaporation increasing from 0.0% in mid-May, to 4% in late-May, and to 9% by late August. Oxygen isotopes are more erratic, changing from 15%, to 8%, to 19% evaporation. As discussed in Section 3.4.4, during non-equilibrium evaporation (fast evaporation) the  $\delta^{18}/^{16}$  change becomes larger than the  $\delta D/H$  change. Therefore, the erratic percent evaporation values for oxygen probably indicate non-equilibrium fractionation, and use of the Rayleigh formula is invalid.

To get some idea of what the potential recharge from stream channels is, an average annual isotopic composition for Ash Canyon Creek was calculated. In general, during the winter, the isotopic composition of stream water should be



more depleted than summer streamflow, since winter runoff at the cooler temperature will be subject to less evaporation. Summer runoff will get progressively heavier, as the amount of evaporation increases, shown by the isotope analyses above. The amount of evaporation depends upon availability of precipitation for evaporation and on the temperature. This correspondence between the isotopic cycle and the temperature cycle has been documented over several years (Dansgaard, 1964).

The top graph of Figure 6.16 shows the 1980-81 and average monthly temperature cycle for Carson City. The other graphs are estimates of hydrogen and oxygen isotopic composition, based upon insufficient data. Snow from the average recharge elevation, 7950 ft, probably represents the most depleted stream isotopic composition, assuming no evaporation. The most enriched isotopic composition was estimated from the temperature shift between stream sample #166 and the apex of the temperature graph, using Figure 3.2. The most depleted snow found in the basin (from 9214 ft) is also plotted on the graphs. The average isotopic composition lines are based on the 1.5°C temperature difference between the 1980-81 year and the average. A temperature drop of 1.5°C would make  $\delta D/H$  values 13% more depleted, and  $\delta 18/16$  values 1.6% lighter.

These estimates of monthly isotopic composition were

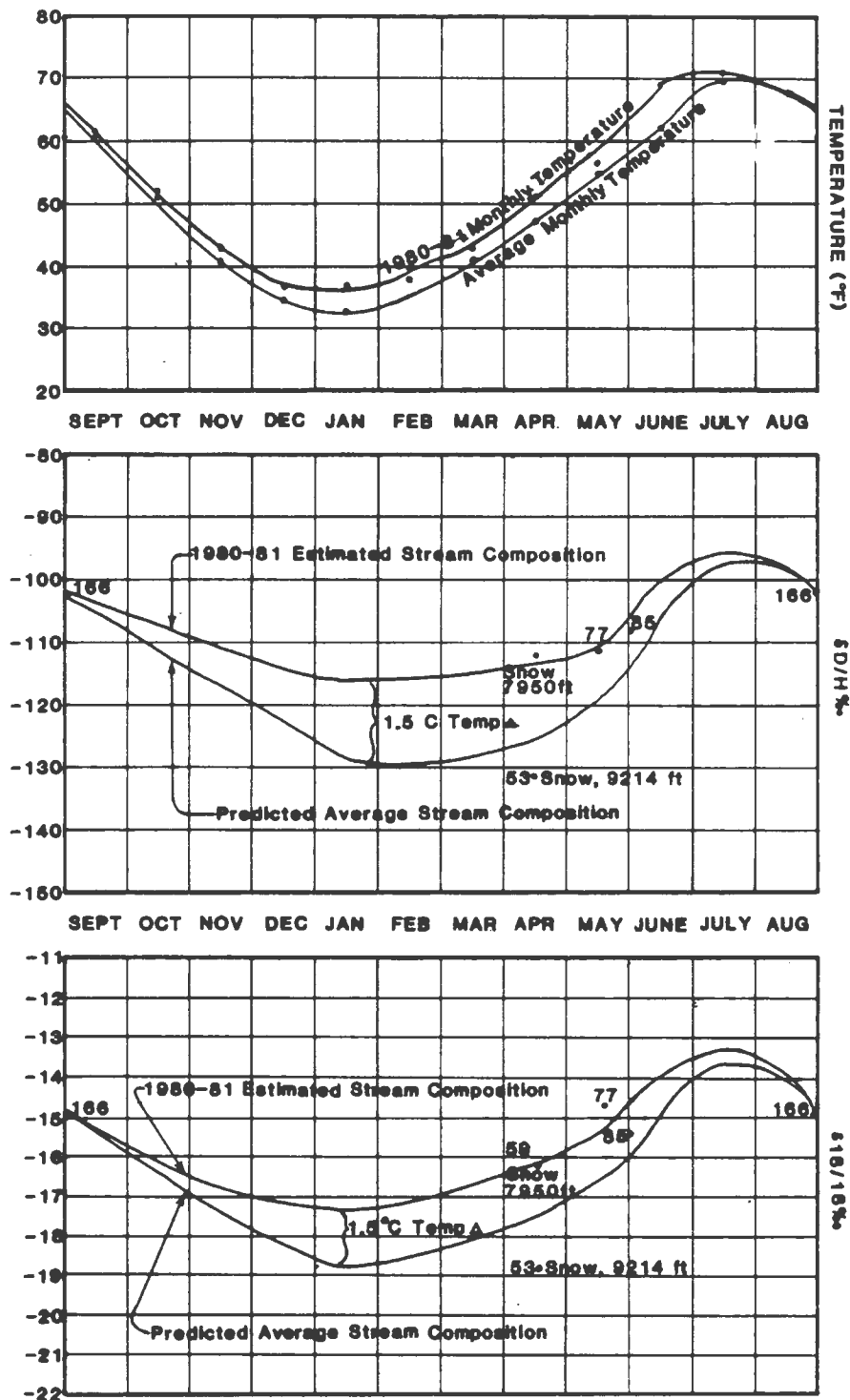


Figure 6.16 Ash Canyon Creek Estimated Isotopic Composition

then multiplied by the monthly flow of Ash Canyon Creek. Streamflow data is a four year average (data in Appendix 4). The weighted average isotopic composition was then calculated as:  $\delta D/H = -109.‰$  and  $\delta 18/16 = -15.9‰$  for the 1980-81 year, and  $\delta D/H = -117.‰$  and  $\delta 18/16 = -16.8‰$  for the average year. This represents potential recharge via stream channel infiltration to the aquifer. If these values are correct, then enriched values in the valley would represent more summer recharge and less winter recharge.

## 6.5 WATER CHEMISTRY OF THE EAGLE VALLEY AQUIFER

### 6.5.1 Introduction

Sampling of Eagle Valley wells and springs in this study along with data from published studies have resulted in about 80 chemical analyses (see Appendix 1). Analyses used in this report are located by Township, Range, and Section and (X,Y) coordinates. Samples 2-180 were sampled in this study and 1000-1049 are data from other sources. The (X,Y) coordinates are based on (0,0) located at T14N, R19E, Sec 10aaaa; this system is used to locate sample points more accurately. Stable isotope data is found in Appendix 2, and tritium data in Appendix 3. A graphical display of sample points in the aquifer is given in Figure 6.17, which was produced by a computer program described in the following text. The irregular outline is the approximate aquifer boundary.

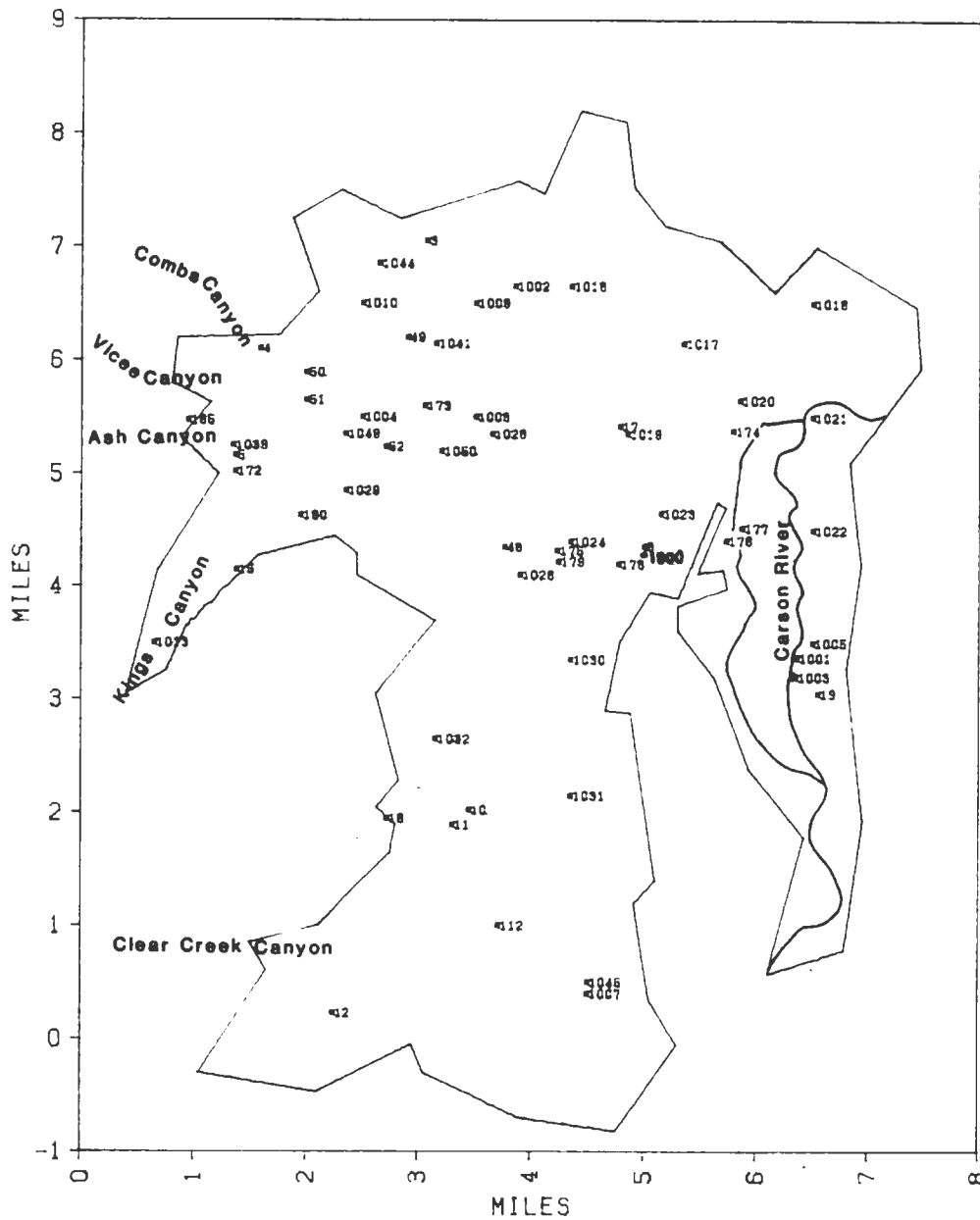


Figure 6.17 Aquifer Sample Locations

Three computer programs were used to show data trends. One is a contour routine (Lukas, 1980) developed at the University of Hawaii and modified to facilities at the University of Nevada, Reno. This program, Contour, plots graphs on a Zeta model 1553 graphics plotter. This program uses both LaPlacian and spline interpolation to produce contour maps. The program initially calculates contour values at node points, then fits contour lines between all node points. The program has a maximum amount of time it can use to calculate node points, within a certain number of iterations. If node points cannot be calculated in the specified time, more widely spaced node points must then be selected. Therefore, some of the graphs have more angular contours due to more widely spaced node points. Since the program was originally developed for a coastal aquifer, the alluvial boundary is considered impermeable, not entirely valid for Eagle Valley, where fracture flow from the recharge area may exist. The proximity of the contour lines to the alluvial boundary is also a function of the node spacing.

The second program used calculates and plots first, second and third order trend surface maps. This program was originally developed by Davis (1973), and was modified to University of Nevada computer facilities by Tom Panian, a Desert Research Institute graduate fellow. The program does not have the capability to reproduce an alluvial boundary, as does contour, so trend map contours overlap the aquifer

boundary. A modified version of WATEQ was used in this study to calculate cation and anion balances, although it is capable of calculating considerably more information. This program has been recently updated by Dr. Burkhard Bohm, a Desert Research Institute graduate fellow.

#### 6.5.2 Aquifer Physical Parameters

To become familiar with the groundwater system in Eagle Valley, temperature and conductivity contour maps are presented. There are three geothermal areas in Eagle Valley, graphically shown on the contour map of temperature (Figure 6.18): Carson Hot Springs in the north-central valley ( $50^{\circ}\text{C}$ ); Pinyon Hills ( $44^{\circ}\text{C}$ ) on the eastern edge of the valley; and the Prison Hot Springs-Sewage Treatment Plant (STP) warm well ( $31^{\circ}\text{C}$ ) area in the central valley. Heavy contour lines are every  $10^{\circ}\text{C}$ , and light lines are  $2^{\circ}\text{C}$ .

A contour map of electrical conductivity in Eagle Valley also delineates the geothermal areas (Figure 6.19). The steep contour gradient at Pinyon Hills is due to a low conductivity nearby. Aside from the thermal areas, there is a trend of increasing conductivity in the northeastern portion of the valley. The conductivity increases from 95 umhos/cm at Ash Canyon Creek to 531 umhos/cm near the Carson River. This could indicate the flow direction, already known to be west to east from the hydraulic head gradient. A plot of third order trend surface maps of all wells and only non-

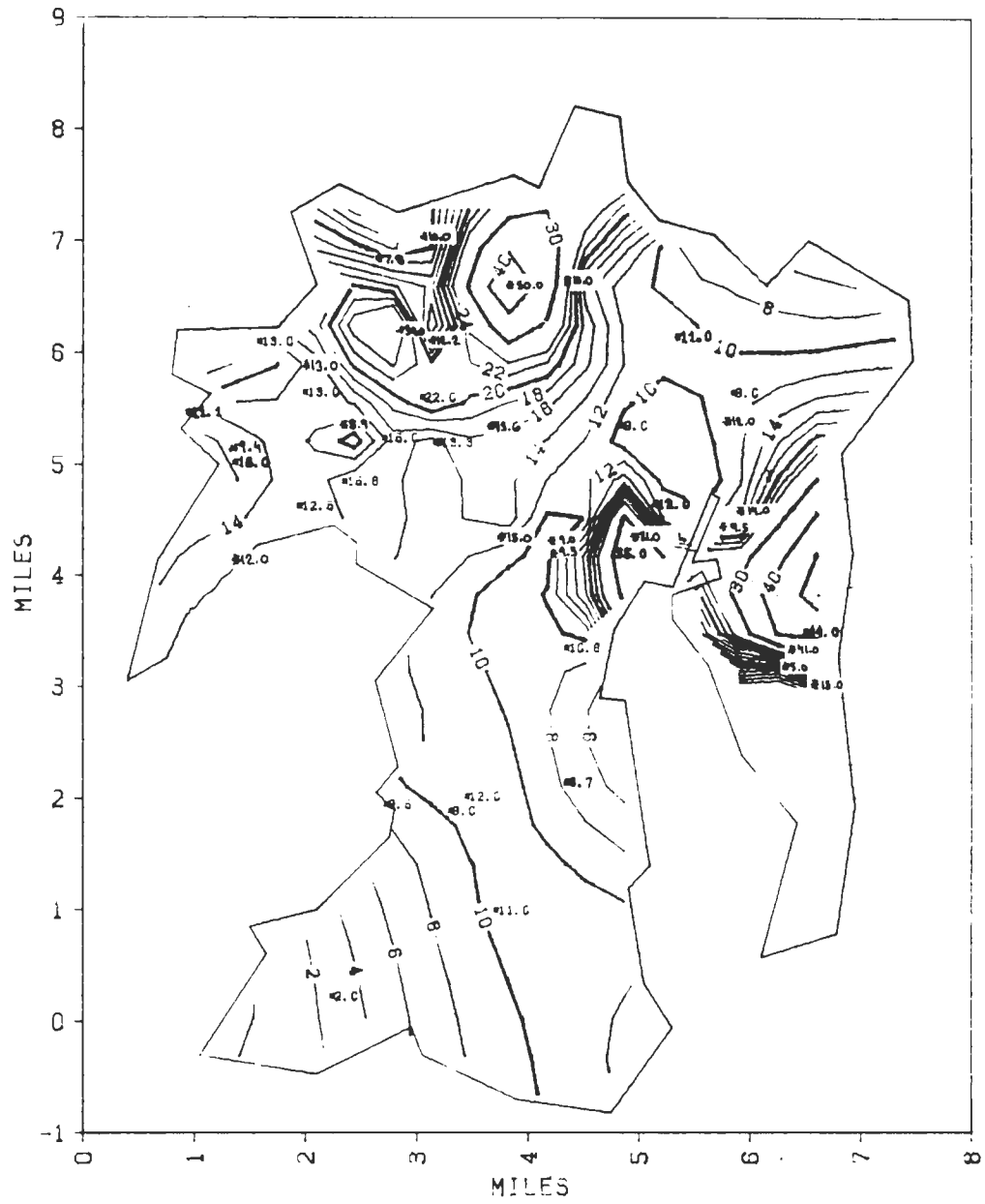


Figure 6.18 Temperature ( $^{\circ}\text{C}$ ), 46 Wells

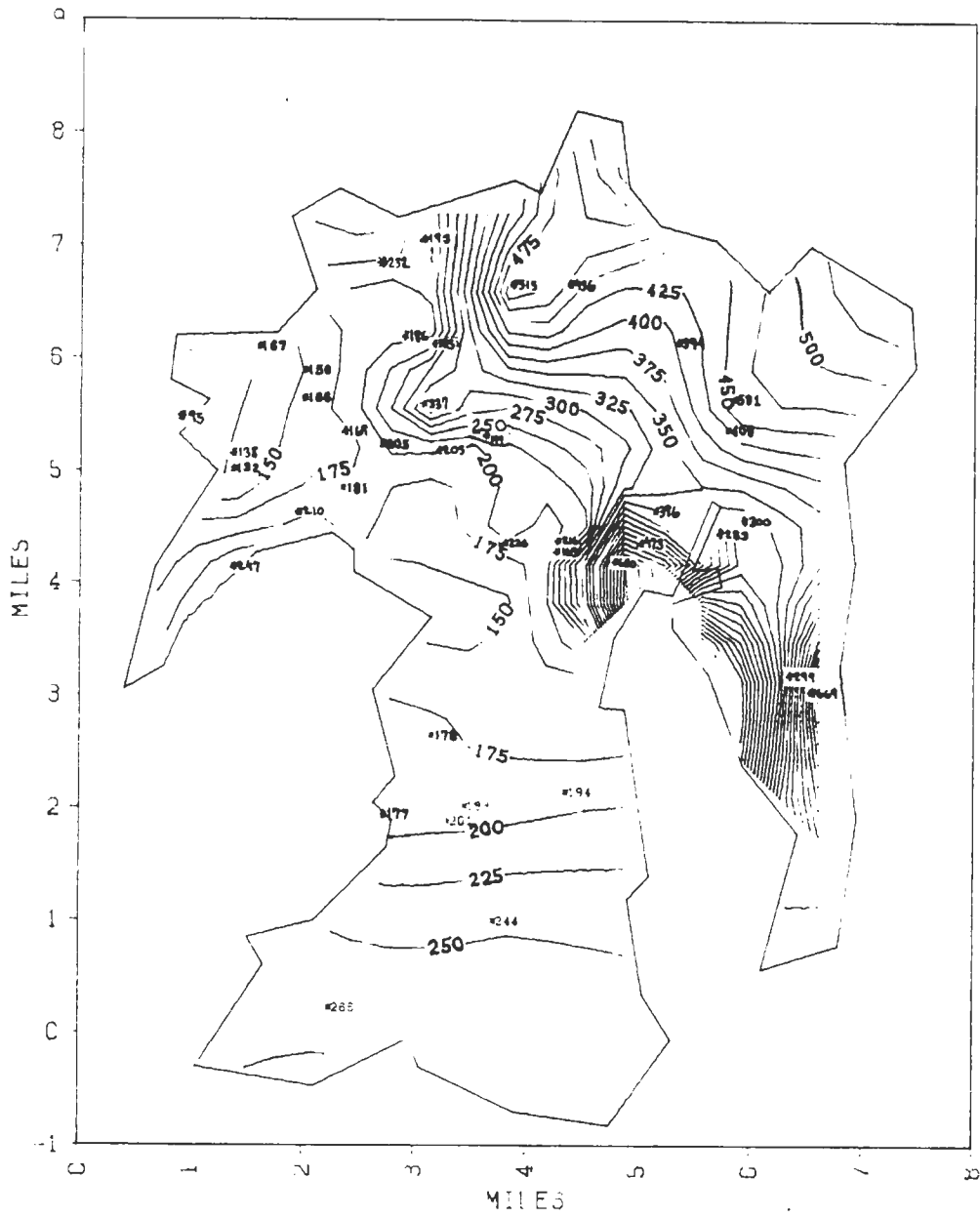


Figure 6.19 Electrical Conductivity (umhos/cm), 47 Wells



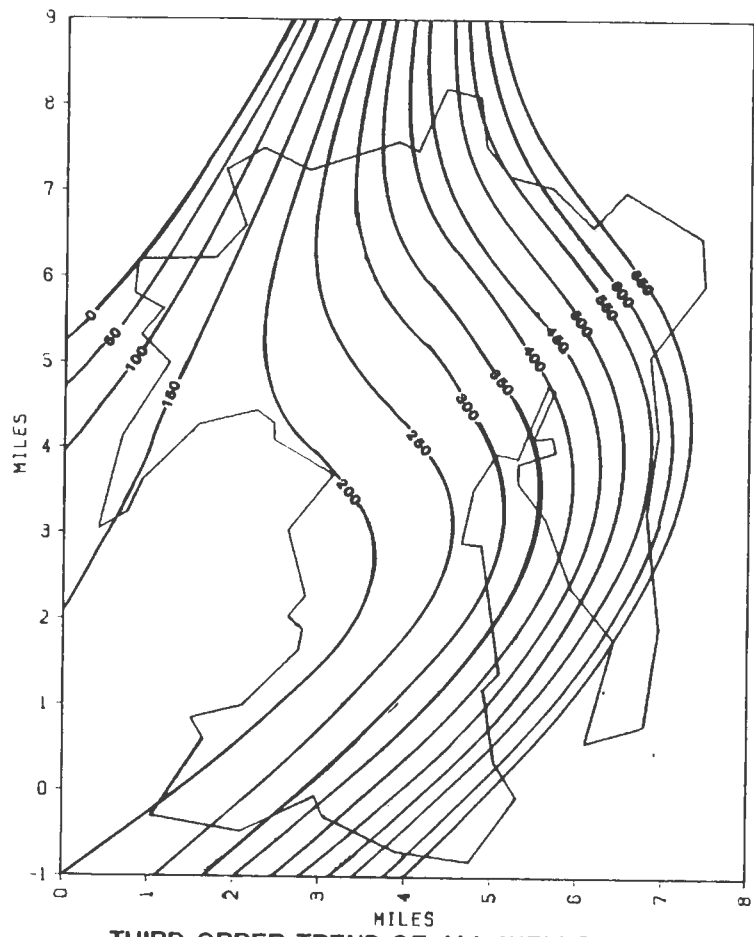
thermal wells (Figure 6.20) shows the trend of increasing conductivity toward the northeast. Also the contours without thermal wells are similar to the contours of the plot with all wells, suggesting that there must be mixing of thermal and non-thermal fluids. If there is no mixing, the plot without thermal waters would be significantly different.

### 6.5.3 Major Ion Chemistry

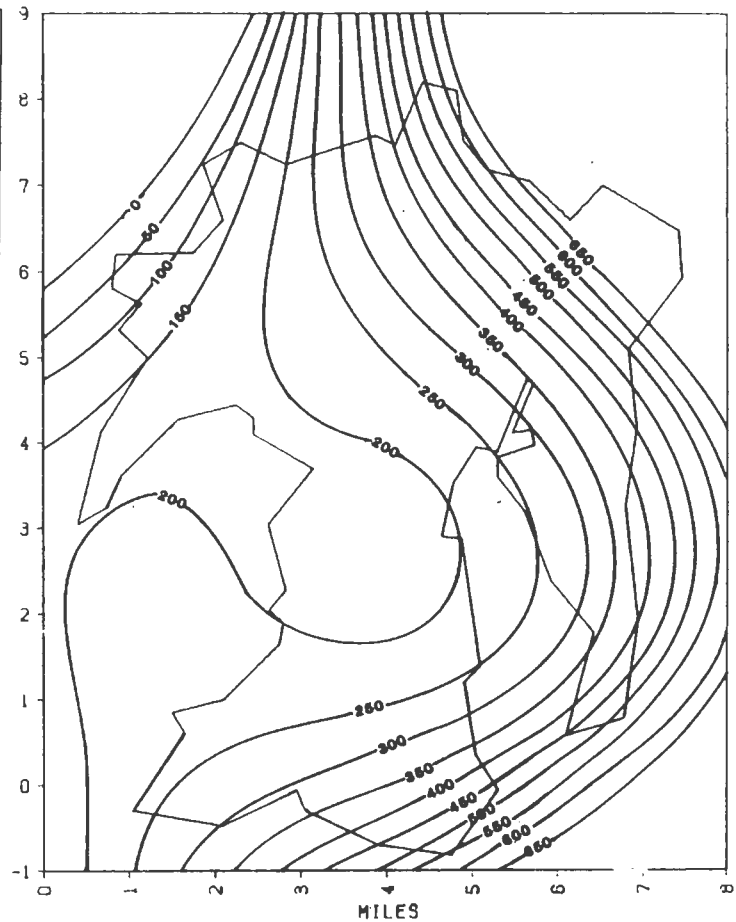
In order to use data sampled over a number of years, one assumes that major changes in water chemistry did not occur. In this study, some data used were collected 16 years ago (Worts and Malmberg, 1966). To check the validity of the assumption of no chemical changes, several wells sampled by Worts and Malmberg were resampled. Data comparison is shown in Table 6.3.

The first group of wells, all non-thermal and sampled in 1965 and 1981, show a significant decrease in chloride, sulfate, and magnesium levels; other parameters are within analytical error. Sample #3 shows a temperature change from 20°C to 10°C, which suggests that the wells sampled may not be the same well. However, in all other wells, chloride and sulfate levels declined, which may represent a definite change in non-thermal water quality.

Perhaps the best indication of water quality changes of



THIRD ORDER TREND OF ALL WELLS



THIRD ORDER TREND OF  
NON-THERMAL WATERS (<20°C)

Figure 6.20 Electrical Conductivity Trend Surface Maps (umhos/cm)

NON-THERMAL WATERS

SAMPLE NUMBER	LOCATION (E,N)	DATE SAMPLED	TEMP. T C	pH	SpC	HCO <sub>3</sub> ppm <sub>3</sub>	Cl ppm	SO <sub>4</sub> ppm	Na ppm	K ppm	Ca ppm	Mg ppm	SiO <sub>2</sub> ppm <sup>2</sup>	Well #
1027	1.59,6.10	3-29-65W	16.1	7.7	181.	92.	3.2	4.8	-12.-	-	15.	9.1	--	7015
4	1.59,6.10	1-31-81	13.0	7.32	167.	103.	1.2	1.5	13.4	2.4	17.	3.96	39.	7015
1033	3.71,1.00	3-26-65W	13.3	7.3	247.	144.	7.0	12.	-23.-	-	27.	5.7	--	Nevada
112	3.71,1.00	4-14-81	11.0	7.84	244.	128.	5.0	9.1	24.7	1.3	21.	3.13	35.	Indian Agency
1034	3.07,7.05	3-26-65W	20.0	8.1	229.	128.	6.3	14.	-24.-	-	23.	4.7	--	8036
3	3.07,7.05	1-31-81	10.0	7.41	198.	105.	3.2	3.9	15.0	3.8	19.	4.24	38.	8036
1039	1.37,5.15	3-29-65W	9.4	7.4	119.	72.	2.6	6.0	-10.-	-	14.	2.7	--	5111
5	1.37,5.15	3-14-81	--	7.00	138.	86.	0.6	1.3	6.70	1.8	16.	3.22	25.	5111
% Average Increase (+) or Decrease (-)			-29%	-2.9%	-1.6%	-.4%	-54%	-61%	-1.9%	-2.4%	-23%			

THERMAL WATERS

SAMPLE NUMBER	LOCATION (E,N)	DATE SAMPLED	TEMP. T C	pH	SpC	HCO <sub>3</sub> ppm <sub>3</sub>	Cl ppm	SO <sub>4</sub> ppm	Na ppm	K ppm	Ca ppm	Mg ppm	SiO <sub>2</sub> ppm <sup>2</sup>	Well #
1035	3.85,6.65	3-30-65W	48.9	9.3	506.	82.	29.	96.	-96.-	-	2.6	0.4	--	Carson Hot
1002	3.85,6.65	1979T	50.	8.84	515.	83.3	27.4	89.4	98.	1.6	2.2	0.01	60.	Springe
% Average Increase (+) or Decrease (-)			+2.2%	-5.2%	+1.7%	+1.6%	-5.8%	-7.4%	+2.5%	-20.7%	-	--		
1000	5.00,4.36	1979T	32.	7.83	501.	47.4	19.7	132.	90.	.63	12.	.01	36.	STP Well,
8	5.00,4.36	1-31-81	31.0	9.20	473.	40.0	17.8	145.	80.	.74	12.	.03	31.	14436
1006	4.77,4.20	1979T		8.8		48.	21.	148.	82.	2.	14.	.33	--	Prison Hot
178	4.77,4.20	10-15-81	35.	8.49	650.	39.3	21.0	152.	84.	2.3	17.	.24	37.	Springe
% Average Increase (+) or Decrease (-)			-3.2%	+5.6%	-5.9%	-20%	-5.3%	+5.8%	-4%	+14%	+10%	-40%	-16%	

W: Worts and Malmberg, 1966  
 T: Trexler, Koenig, Flynn, and Bruce, 1980

Table 6.3 Major Ion Changes over Time

thermal waters is Carson Hot Springs, sampled in 1965, and in 1978. It shows no significant trend above analytical accuracy and resolution. The third set of samples are also thermal, representing sampling three years apart. Again, there is no significant trend above analytical accuracy.

In general, it appears that non-thermal waters might be changing, and thermal waters are not. However, since it is not conclusive that major chemistry changes exist, data from Worts and Malmberg (1966) were used in this study.

A Piper diagram of 29 samples from this study plus 40 other analyses is shown in Figure 6.21. It appears that there are two different thermal waters: one a low TDS Na-SO<sub>4</sub> water of Carson Hot Springs and Prison Hot Springs, and a high TDS Ca-SO<sub>4</sub> of the Pinyon Hills area. This is the same conclusion reached by other investigators (Trexler and others, 1980). The anions of all waters show a very regular trend starting with low TDS alkaline (CO<sub>3</sub><sup>=</sup> + HCO<sub>3</sub><sup>-</sup>) water changing to mostly sulfate water of higher TDS. Cations also show a regular trend starting with 20% Mg, 50% Ca, and 30% Na+K, and then becoming enriched in Na<sup>+</sup>+K<sup>+</sup>. The initial composition is probably due to recharge area rock chemistry. Cation and anion trends are shown graphically in Figure 6.22, a Piper diagram of wells from Ash Canyon Creek east to the Carson River.

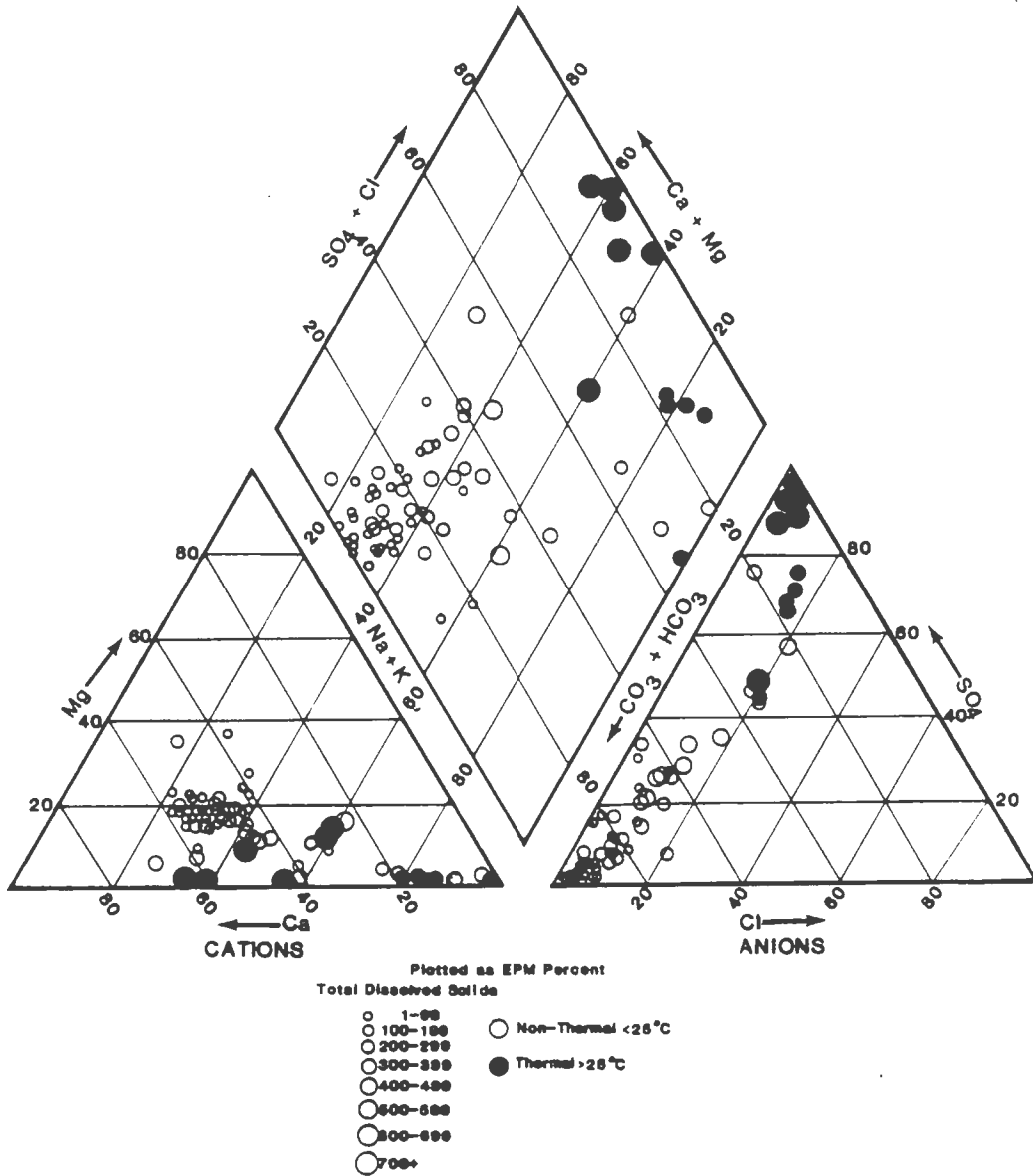


Figure 6.21 Piper Diagram of Waters in Eagle Valley, Nevada

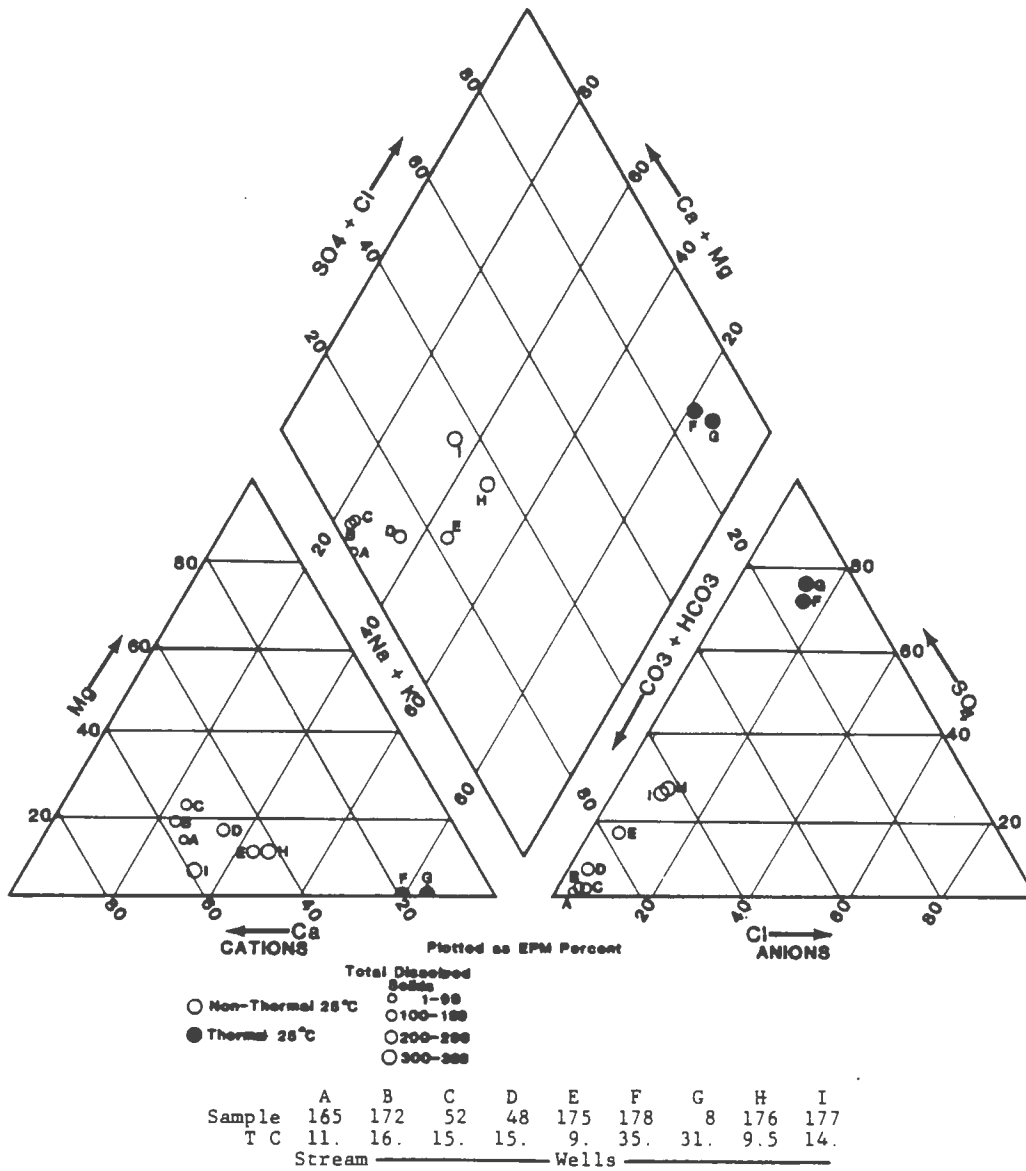


Figure 6.22 Major Ion Changes Eastward across Eagle Valley

Using Figure 6.21, a density plot of well chemistry was constructed using  $\pm 10\%$  hexagons for counting, shown in Figure 6.23. This diagram does show trends and the most common water chemistry composition clearly, but unfortunately, it does not show changes in TDS or indicate which analyses are thermal waters. In both Figure 6.21 and 6.23, sample #17 was omitted since the well was considered contaminated from a septic tank, and the analysis showed anomalously high sodium.

Piper diagrams show general chemistry changes for a group of wells, but do not show the spatial variation of ions over an area. Piper diagrams can also be misleading, since ions are plotted as a percentage of the total. For example, when a trend of decreasing alkalinity percent is illustrated, there may still be an increase in  $\text{HCO}_3^-$  concentration, which is masked by a larger increase in another ion such as sulfate. One method of showing spatial variation of ions is the use of Stiff diagrams, which are symbolic representations of the chemistry of a well. Another method is to present a contour map of each ion, which was done in this study.

6.5.3.1 Major Anions. The Piper diagram (Figure 6.21) indicates that alkalinity percent of the total decreased along the flow path. This does not necessarily mean the concentration of alkalinity decreases; in fact, alkalinity

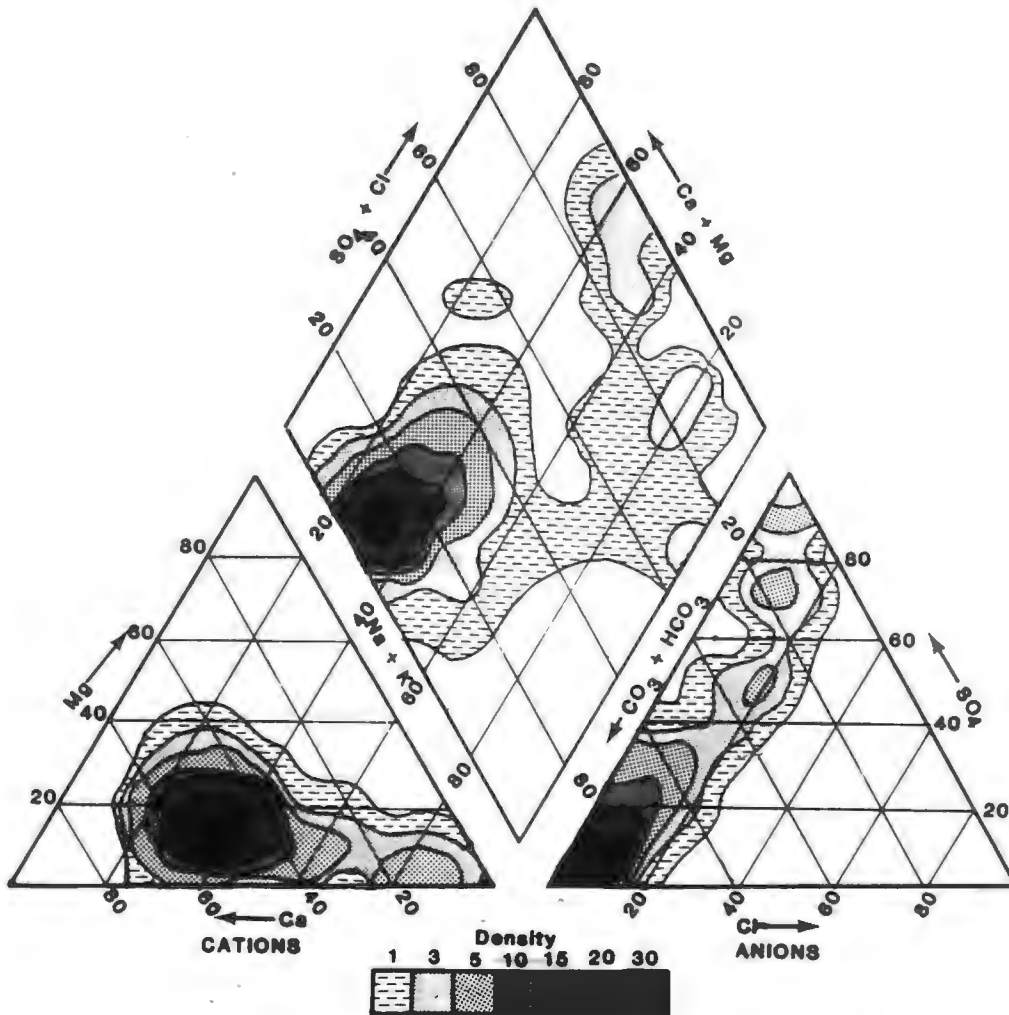


Figure 6.23 Density Plot of Piper Diagram



increases along the flow path (see Figure 6.24). The reason for the decrease in alkalinity percent of total to the east is the greater increase of sulfate. As  $\text{CO}_2$  gas is dissolved into recharging water, it is slowly converted into bicarbonate via the chemical reaction:



This is probably the reason for the alkalinity increase from 51 ppm at Ash Canyon Creek to 250 ppm at the Carson River.

Geothermal areas show up as decreases in alkalinity concentrations. The only other slight trend observed is a small increase in alkalinity concentrations toward the southeast corner of the aquifer. This may represent groundwater flow along the Clear Creek channel.

Sulfate percents in non-thermal wells are shown by the Piper diagram as an increase at a greater rate than alkalinity. In addition, there is a very large increase in sulfate concentrations in thermal areas. A contour map of sulfate concentrations in Eagle Valley (Figure 6.25) shows an increase in sulfate concentrations in non-thermal wells toward the east from 0.0 ppm to about 30.0 ppm at Empire, then a decrease back to 10.0 ppm. The increasing sulfate concentrations appear to represent the flow direction of west to east in the northern half of the basin.

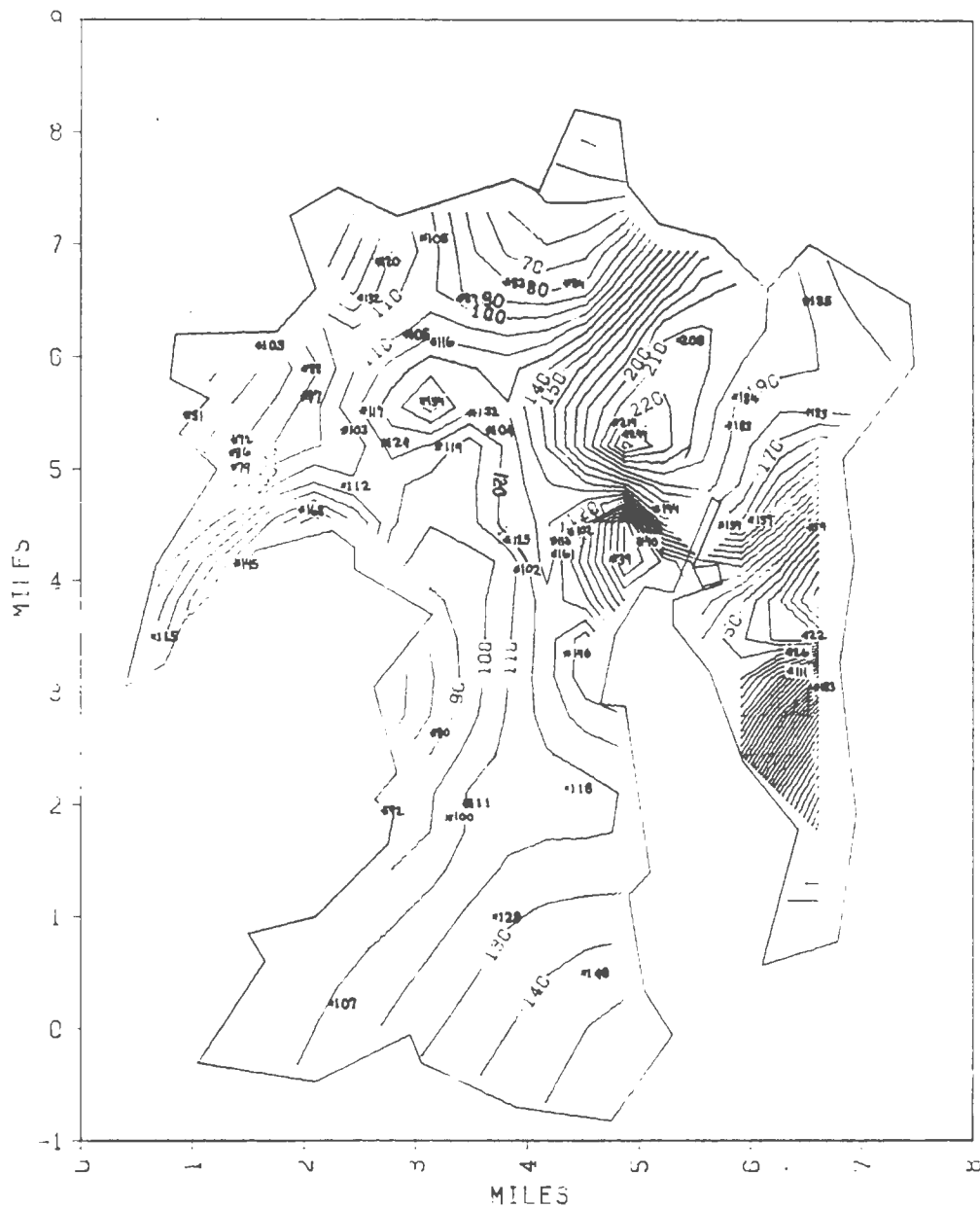


Figure 6.24 Alkalinity ( $\text{HCO}_3 + \text{CO}_3$ , in ppm), 59 wells



The southern half of the basin shows no discernible trend. However, areas with no record of thermal waters, the northwest quarter and southern half of the basin, do not show any trend. Sulfate concentrations only start to increase in the northeastern portion of the basin, near thermal areas. Therefore, the sulfate concentration increase in non-thermal waters is probably due to mixing with thermal waters. The one well near Empire which showed a lower concentration of sulfate may represent groundwater infiltration from the Carson River or irrigation diluting the sulfate levels.

The third major anion, chloride, shows no trends when plotted on the Piper diagram. Shown on a contour map (Figure 6.26), chloride concentrations increase regularly from west to east in the northern half of the basin. There is a slight decrease in chloride concentration near the Carson River, possibly indicating recharge in that area. Geothermal areas have anomalously high chloride concentrations. The southern half and northwest quarter of the basin, where there are no known thermal waters, show no discernible chloride trend, similar to sulfate. This suggests that the easterly increase in concentrations of  $\text{Cl}^-$  and  $\text{SO}_4^{2-}$  is due to mixing of thermal and non-thermal fluids, and not the regional anion shift of the Chebotarev sequence.

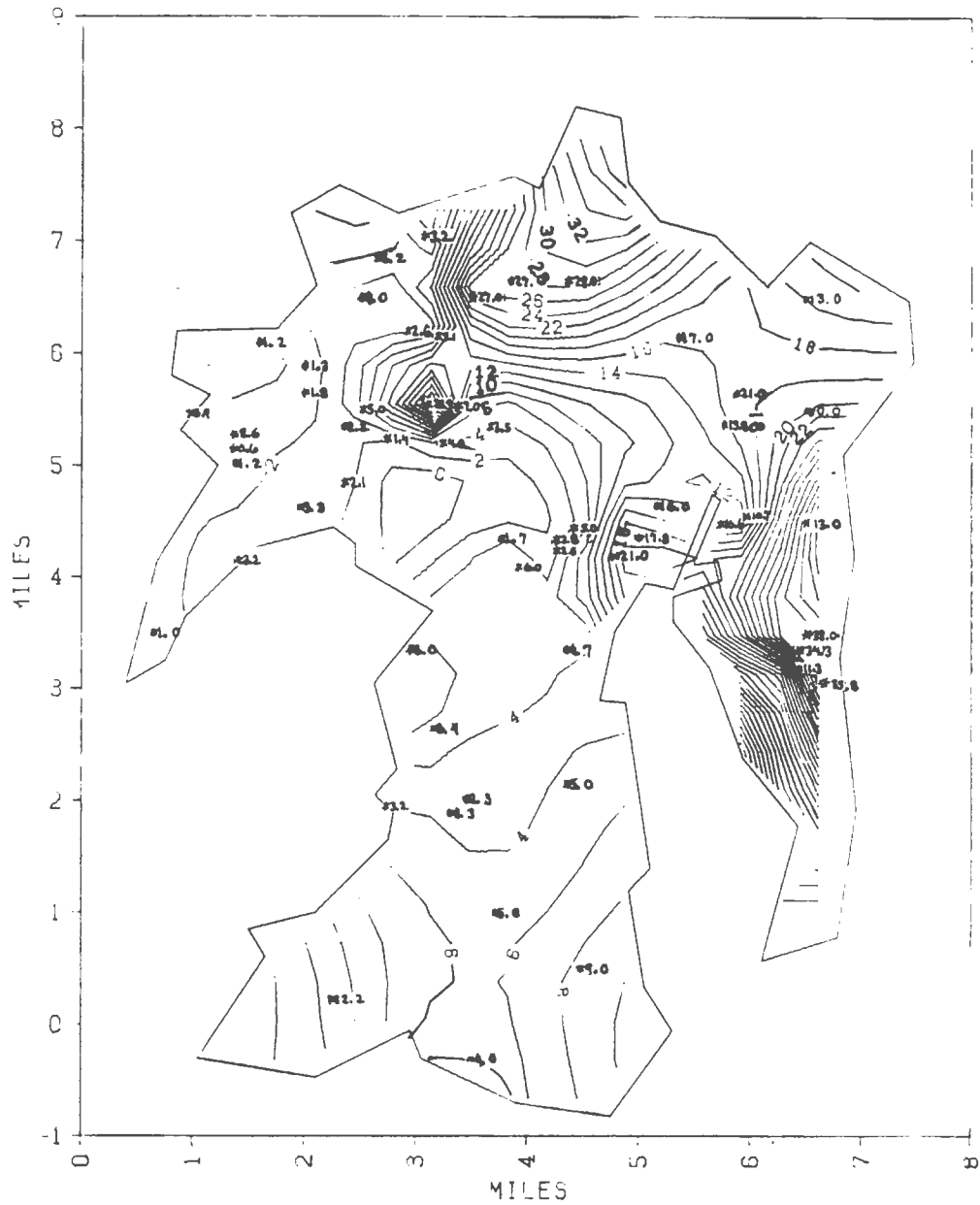


Figure 6.26 Chloride (ppm), 58 Wells

6.5.3.2 Major Cations. Cations are not usually used to imply flow direction, as with anions, due to cation exchange. The Piper diagram of analyses from Eagle Valley does show a general trend of increasing sodium and decreasing calcium percent of total. A contour plot of calcium concentrations (Figure 6.27) shows that calcium levels in the northern half of the basin increase along the flow path (to the east), from 10 ppm at Ash Canyon Creek to 53 ppm at the Carson River. The southern half of the basin, as with alkalinity, shows a small increase toward the east, along Clear Creek drainage. This is based on three data points, and thus may not represent an actual trend.

Magnesium is usually associated with calcium, since substitution in crystals is common due similar ion sizes. A contour plot of magnesium levels in Eagle Valley (not shown) does show the same general trends as the calcium contour map.

Both calcium and magnesium concentrations in geothermal areas show the same trends: Carson Hot Springs and Prison Hot Springs areas have lower concentrations of  $\text{Ca}^{+2}$  and  $\text{Mg}^{+2}$  than surrounding cold wells, and the Pinyon Hills thermal area has much higher  $\text{Ca}^{+2}$  and  $\text{Mg}^{+2}$  levels than surrounding cold wells. A slightly thermal well, sample #49 (30°C), shows up as increased calcium and magnesium concentrations, so probably represents thermal  $\text{Ca-SO}_4$  fluid. Sulfate concentrations of this well are higher than surrounding cold wells.

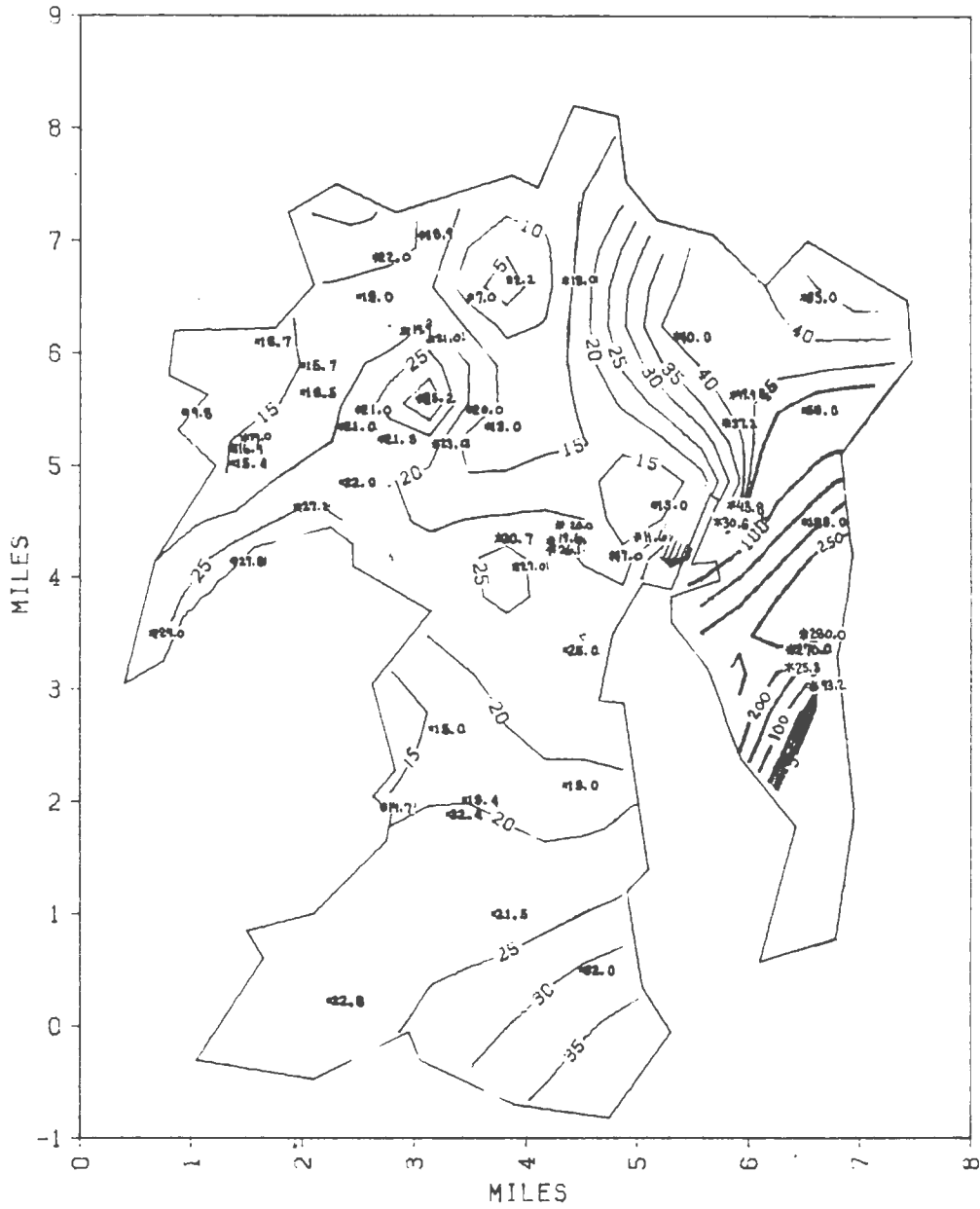


Figure 6.27 Calcium (ppm), 55 Wells

Sodium concentrations should increase across the valley due to mixing of thermal fluids, since Carson Hot Springs and Prison Hot Springs are Na-SO<sub>4</sub> fluids. A contour plot of sodium levels (Figure 6.28) shows that where no geothermal fluids are known, in the northwest quarter and south half of the basin, there is no real trend in sodium concentrations. In the northern half of the basin, once thermal areas are reached, non-thermal waters show an increase in sodium concentrations, again suggesting mixing of thermal and non-thermal fluids. Sodium levels at Carson Hot Springs and Prison Hot Springs are very similar (about 90 ppm), and different from Pinyon Hills area (200 ppm). The potassium contour map did not show any useful trends, so it was not presented.

The Na-K-Ca geothermometer (Fournier and Truesdell, 1973) was applied to Carson Hot Springs, Prison Hot Springs, and the Pinyon Hills area, to calculate maximum fluid temperature and approximate depth of fluid mixing:

Sample	T (°C)	ΔH (ft)	Depth (ft)
1002 Carson Hot Springs	85	10	318.
1001 Pinyon Hot Well	40	-	-
178 Prison Hot Springs	55	10	735.
8 STP Well	35	4	784.

The chalcedony and Na/K<sup>2</sup> geothermometers (Trexler and others, 1980) at Carson Hot Springs indicated 79°C and 75°C respectively, so a temperature of 85°C for the fluid



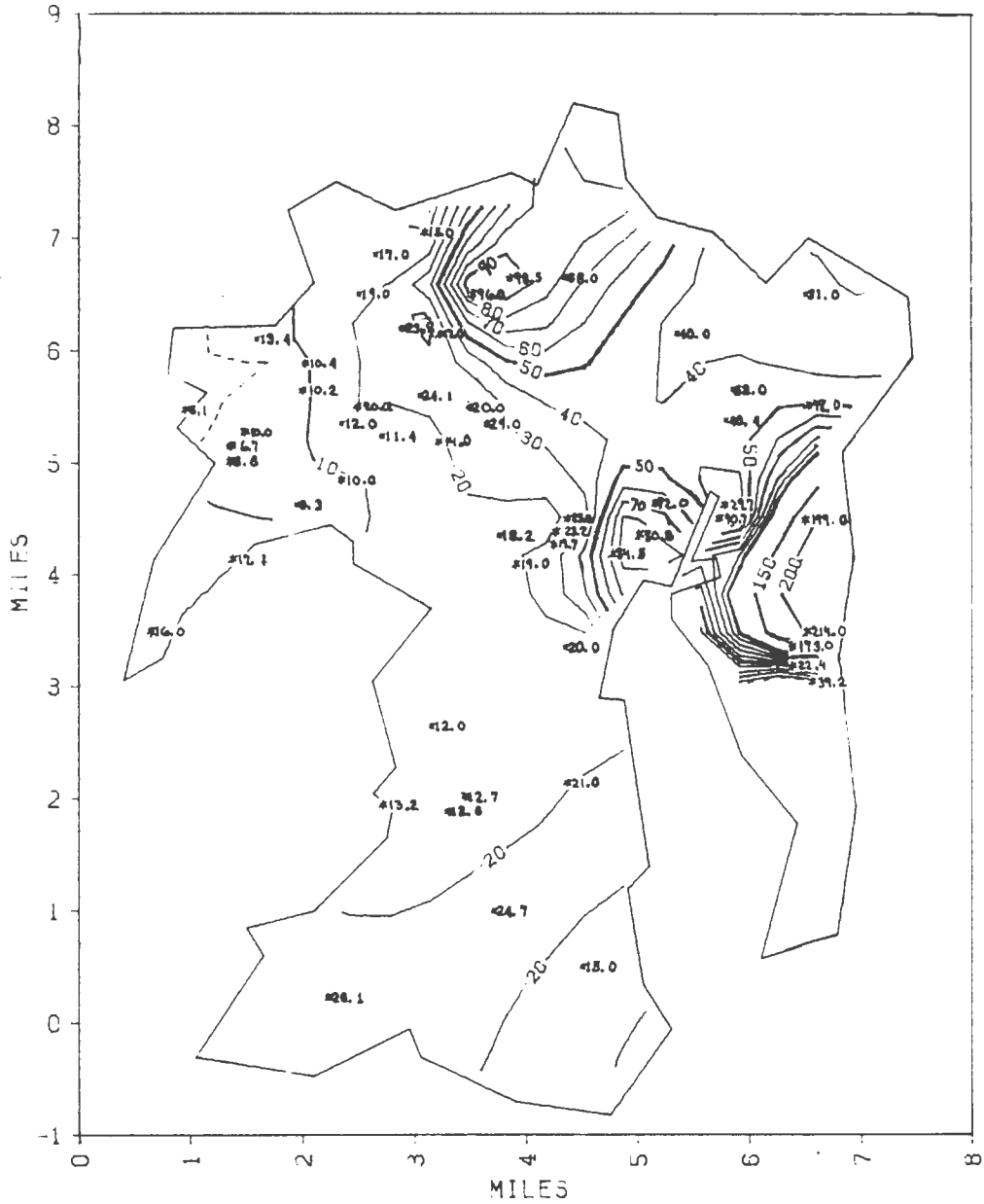


Figure 6.28 Sodium (ppm), 56 Wells

reservoir temperature by Na-K-Ca geothermometry is reasonable. Hydraulic head differences used are by no means accurate or current; they are used simply to demonstrate how the temperature difference in thermal/non-thermal fluids can produce an approximate depth to fluid mixing, based on the density difference between thermal and non-thermal fluids (see Section 3.1). Carson Hot Springs and Prison Hot Springs head differences were 10 ft, based on Worts and Malmberg (1966) groundwater map. A head difference of 4 ft was used for the STP well, based on the well log when drilled. No hydraulic head information was available for Pinyon Hills. Depths of thermal and non-thermal fluid mixing calculated to about 300 ft at Carson Hot Springs, and 700-800 ft at the Prison/STP thermal area, are reasonable values for the depth of the bedrock-alluvium contact.

Another molecule analyzed in wells sampled is dissolved silica. A contour map of silica levels (not shown) shows Carson Hot Springs with very high silica levels. The Prison Hot Springs which was previously considered the same Na-SO<sub>4</sub> water, does not have elevated silica levels. Pinyon Hills area does have elevated silica levels, so it appears that silica does not show trends that other ions show. One well near Clear Creek shows very high (46 ppm) silica, which is unexplainable.

#### 6.5.4 Stable Isotope Chemistry

Throughout the recharge and flow path portions in this study, oxygen-18 has been shown to be more variable than deuterium. Oxygen isotope variability is also greater in the aquifer. There are several geothermal areas in Eagle Valley, and since oxygen isotopic enrichment in geothermal water has been observed (Back and Hanshaw, 1965), oxygen isotope values may not represent the composition of recharge water. Figure 6.29, oxygen versus hydrogen isotopes, shows that non-thermal waters have similar isotopic compositions, and are meteoric waters. Carson Hot Springs, the STP well, and the Pinyon Hills thermal area are also meteoric, although most thermal waters show enrichment of oxygen compared to the local meteoric precipitation (snow). All thermal waters are to be depleted compared to cold waters hydrogen isotopic content, which probably indicates recharge from higher elevations through deeper flow paths.

Two contour maps of oxygen isotopes (Figure 6.30 and Figure 6.31) are presented in this report. Figure 6.30 was made using unaltered analyses from the Tucson and Scotland laboratories. In Figure 6.31, corrections were made for the differences between the two laboratories. Referring to Figure 6.31, non-thermal waters show several trends. First oxygen isotope data indicate that the valley averages about -15.0‰. Comparing this to the potential recharge from Ash

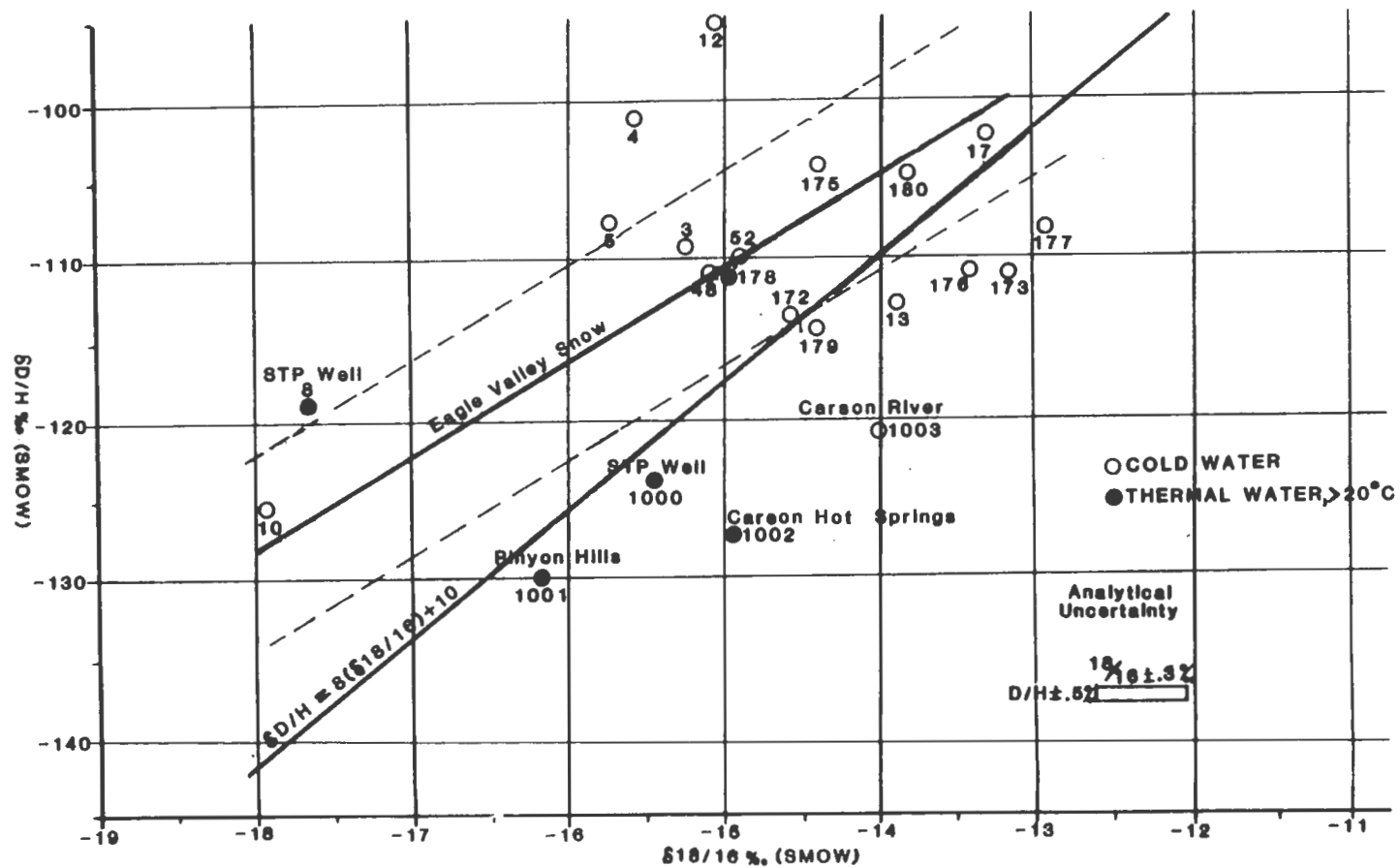


FIGURE 6.29 Oxygen versus Hydrogen Isotopes of Wells in Eagle Valley

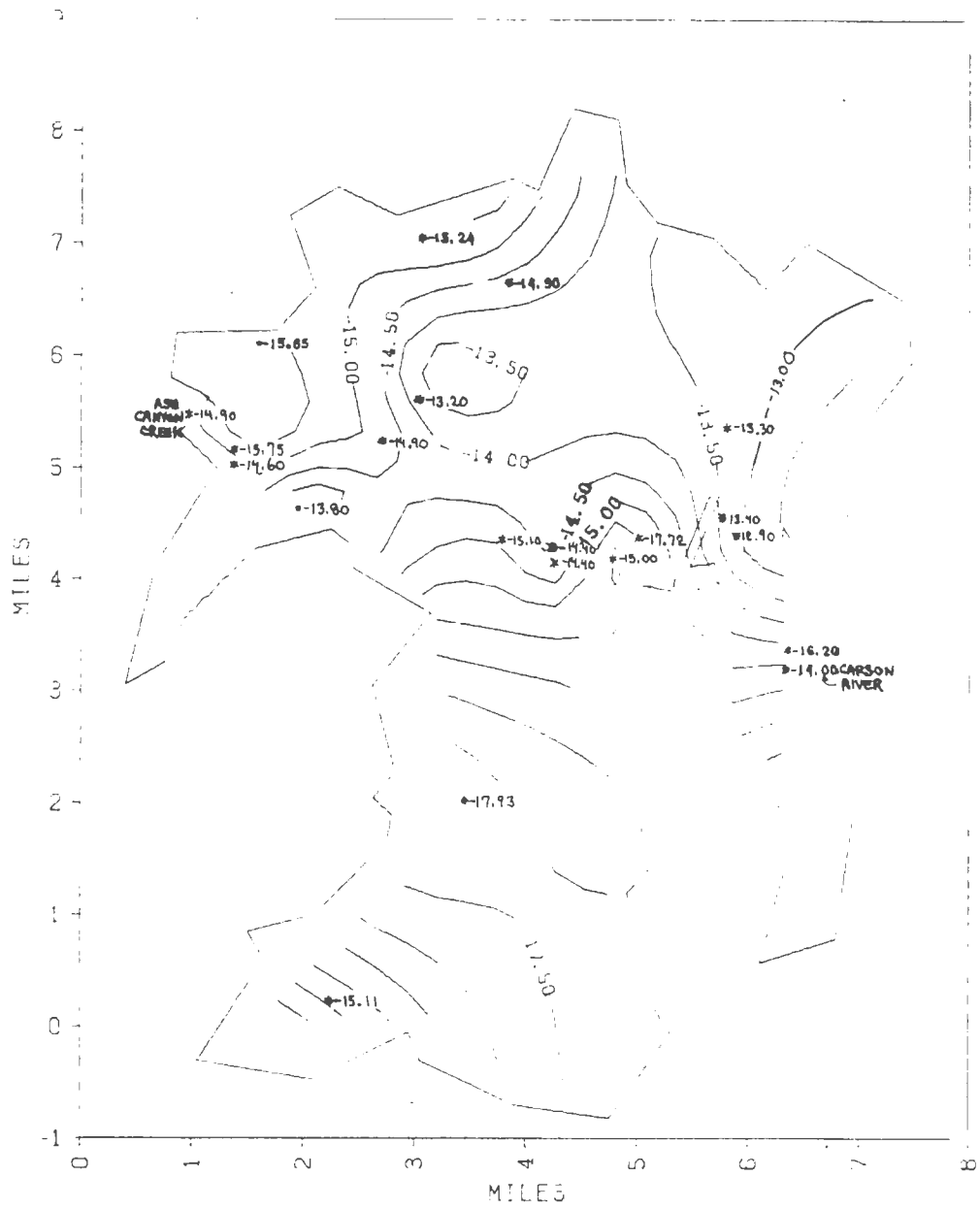


Figure 6.30  $\delta^{18}/^{16}$  (‰), 22 Wells

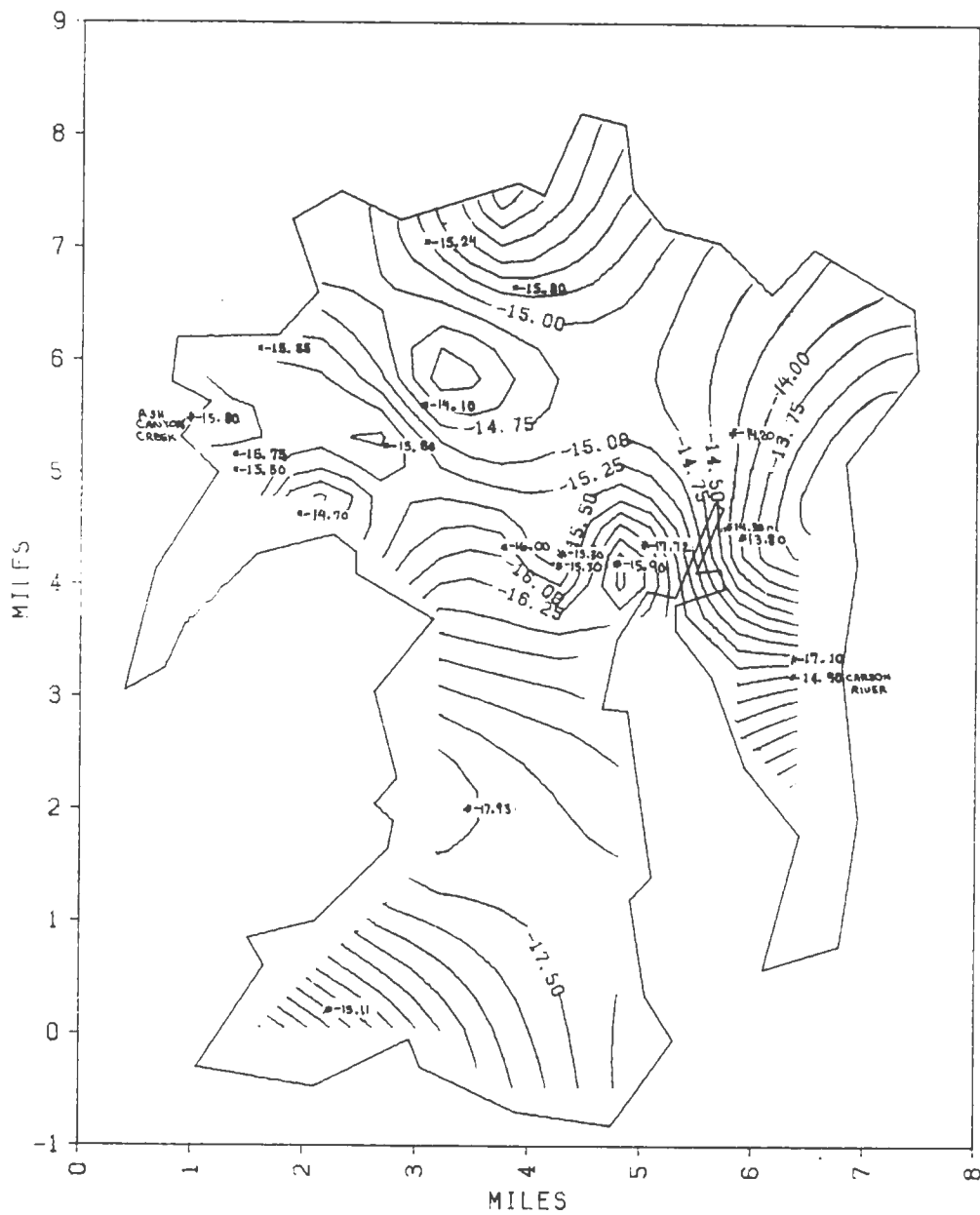


Figure 6.31  $\delta^{18}/16$  (‰), 22 Wells, Corrected for Lab Differences

Canyon Creek of -15.9‰ for the 1980-81 year and -16.8‰ for the average year, one concludes that most recharge must be during the late spring or summer in order for the valley to have such an enriched composition. Snow from the average recharge elevation of 8000 ft exhibits  $\delta^{18}/16 = -17.8‰$  for the average year. This indicates a total evaporation of 24% from the average recharge elevation to the water in the aquifer.

The northeast quarter of the basin appears to be enriched compared to the basin average isotopic composition, and is similar to the isotopic composition of the Carson River,  $\delta^{18}/16 = -14.9‰$ . The south-central portion of the aquifer appears to be much more depleted (-17.9‰), which suggests recharge with little evaporation, possibly deep percolation. However, this observation is based upon one data point.

All three geothermal areas show up as depleted water. The STP well was sampled twice, once in 1980 (Trexler and others, 1980),  $\delta^{18}/16 = -15.4‰$ , and in 1981 with  $\delta^{18}/16 = -17.7‰$ . This is an unusually large discrepancy in values. Correcting for the differences in the samples being analyzed at different laboratories narrows the gap somewhat. This STP well should have the same composition as the Prison Hot Springs ( $\delta^{18}/16 = -15.9‰$ ), since the water is presumably from the same fracture system. Since the compositions are

so different, oxygen isotope data appears to imply that the STP well and Prison Hot Springs are different waters. This conclusion is not supported by hydrogen isotopic data.

Hydrogen isotope contour maps are also presented as uncorrected (Figure 6.32) and corrected for laboratory differences (Figure 6.33). The corrected hydrogen isotope contour map and trend surface maps indicate an average valley composition of  $\delta D/H = -115\%$ . Using the isotopic value of snow from the average recharge elevation,  $\delta D/H = -124\%$  for the average year, a total evaporation of 11% can be calculated from 8000 ft to water in the aquifer.

Referring to Figure 6.33, it appears that the basin has enriched (less negative) areas toward the northwest, northeast, and extreme southwest. The enriched areas represent groundwater input from Clear Creek in the southwest, and Kings, Ash, and Vicee Canyons in the northwest. These inputs are probably summer recharge, which is evaporated and enriched. The center valley being more depleted than the stream channel areas suggest deep recharge to the groundwater system of less evaporated water. This shows that deep recharge is a mechanism of recharge to the aquifer. The northeast enriched area may represent groundwater inflow from the Carson River during the summer. The only available sample of the Carson River was taken in the winter ( $\delta D/H = -128\%$ ), which does not support or disprove



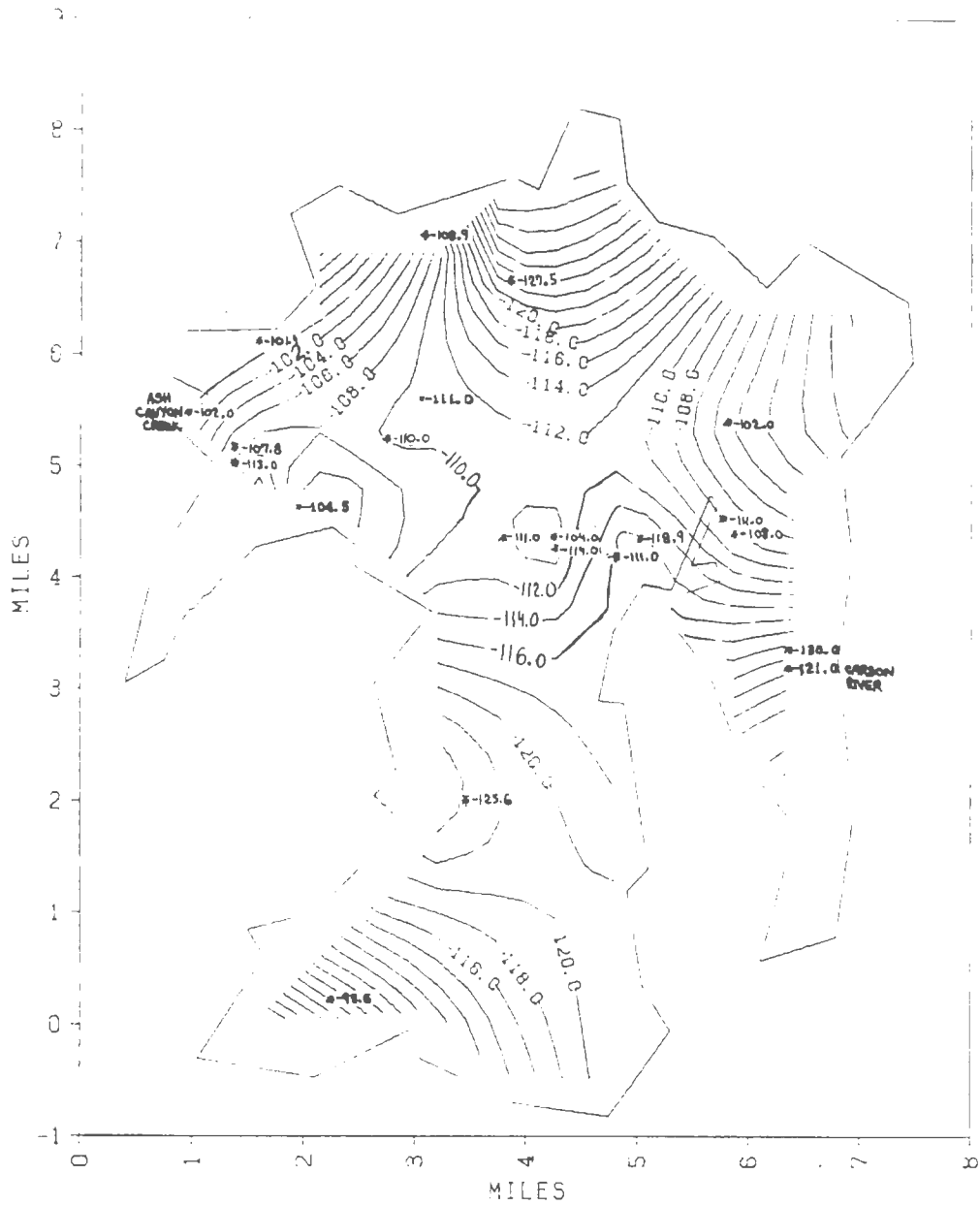


Figure 6.32  $\delta D/H$  (%), 22 Wells

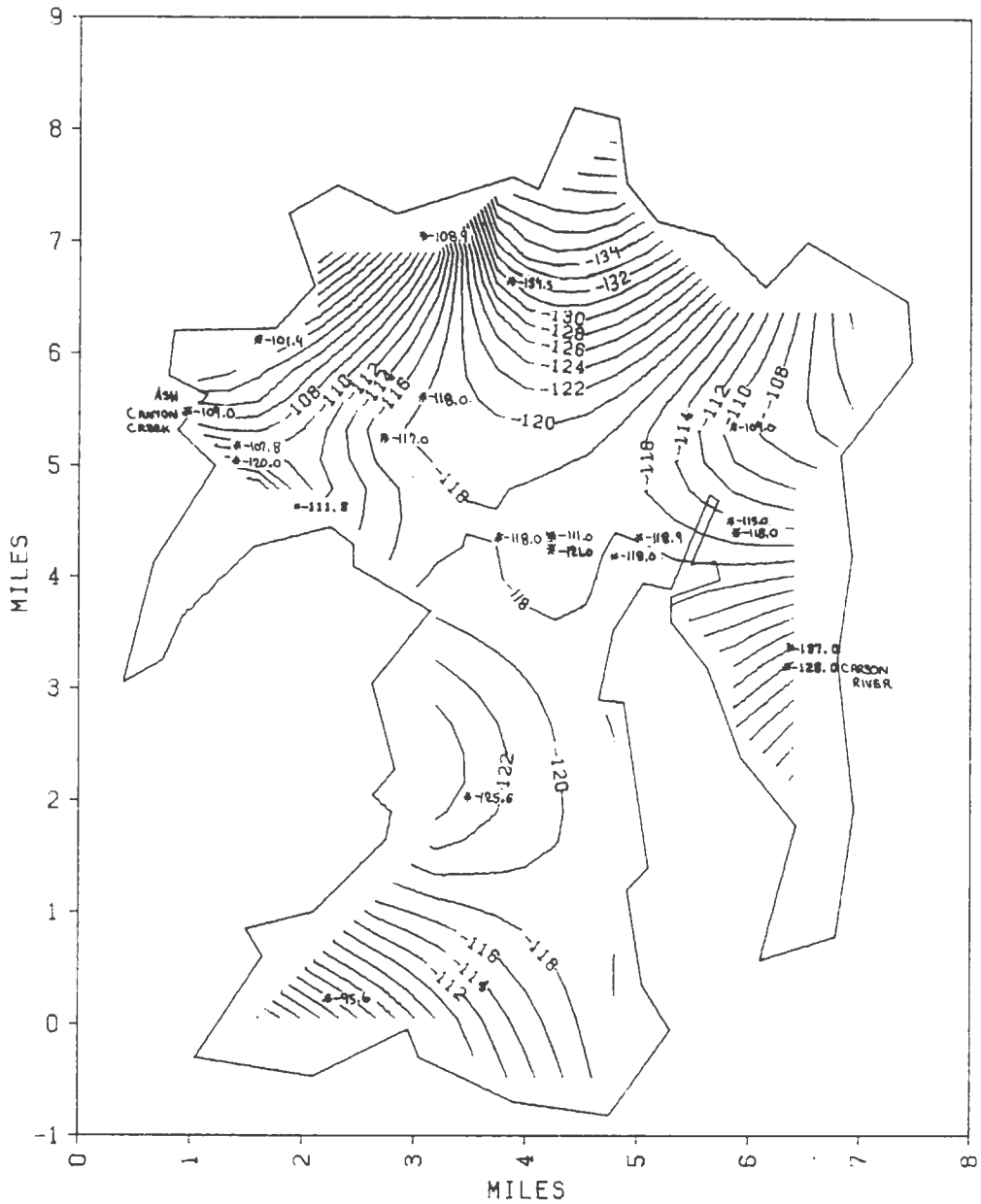


Figure 6.33  $\delta D/H$  (‰), 22 Wells, Corrected for Lab Differences

this hypothesis. The Carson River should have a more enriched composition in the summer.

Hydrogen isotopic values of the Carson Hot Springs and the Pinyon Hills geothermal areas do show up as depleted. This may indicate that these waters are recharged with less evaporated waters, possibly by deep percolation in recharge areas. Since most thermal areas are fracture associated, this is a plausible explanation. Fracture flow from recharge areas is probably much slower, which would make the water older than surrounding cold waters.

Prison Hot Springs and the STP well, when corrected for laboratory differences, have approximately the same hydrogen isotopic composition, suggesting that they are from the same source. This thermal area also has the same composition as surrounding non-thermal waters, implying the same flow system or a considerable amount of mixing of thermal and non-thermal fluids.

Lake Tahoe, 10 miles to the west of Eagle Valley is 1800 ft higher in elevation, and extremely deep, up to 1600 ft. It is possible that there is some regional flow from Lake Tahoe to the Eagle Valley aquifer. The upper Truckee River inflow into Lake Tahoe has a hydrogen composition of  $\delta D/H = -133.‰$ , and the Lake Tahoe surface water has a composition of  $\delta D/H = -65.‰$ . It is difficult to determine what the isotopic composition of water flowing out of Lake Tahoe in fracture systems would be, since no samples have been

taken at depth. Lake Tahoe, because of its great depth, does not mix completely, so the isotopic composition at depth may be different from the surface. Since the Eagle Valley aquifer averages  $\delta D/H = -115.‰$ , this does not exclude the possibility that some water is from Lake Tahoe, but the data are inconclusive. Age dating of some wells in Clear Creek Canyon might provide an answer to this question; if some water in the Clear Creek Youth Center wells are old, then that water may be from Lake Tahoe. Use of biological tracers or other stable isotopes may provide the means to differentiate water recharged in Eagle Valley from the Tahoe basin. The permeability of the fractured granodiorite rocks in this area is small, although the hydrologic head difference between Lake Tahoe and Eagle Valley is 1600 feet, so recharge from Lake Tahoe cannot be ruled out.

#### 6.5.5 Groundwater Age Dating

6.5.5.1 Tritium Age Dating. The tritium activity of water in various wells in Eagle Valley was measured to try to quantify the age and direction of flow. Figure 6.34 is a plot of all tritium analyses in Eagle Valley. A list of the tritium data is found in Appendix 3. All samples but one had less than 50 TU, so the unenriched tritium samples taken were not useful. Almost all enriched tritium analyses were concentrated in the northern half of the basin. In this area, tritium values increase from 27. TU at Ash Canyon

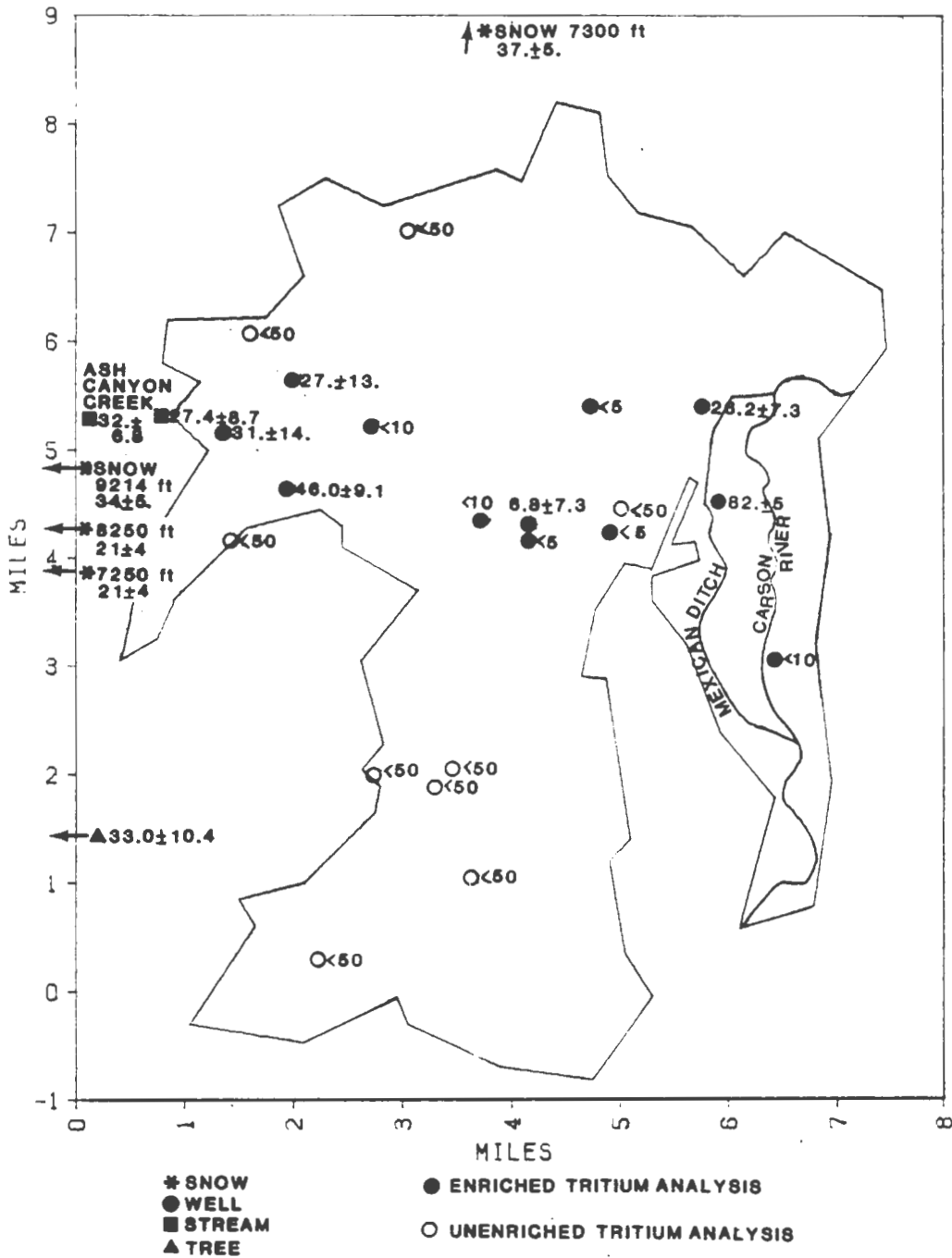


Figure 6.34 Tritium Data (T U)

Creek in the northwest, to 31. TU, 46. TU, then decrease in the central valley to <10 TU in two wells, 6.8 TU in one well, and <5 TU in three wells. Tritium levels then increase from the north-central valley toward the east (toward the Carson River), which is supposed to have the oldest water in the basin. Tritium levels there, near the Carson River, increase to 82. TU and 26. TU.

The tritium data at the mountain front area in the northwest indicate very recent waters suggesting fast infiltration, probably via stream channel infiltration. The oldest water appears to be in the north-central part of the basin, greater than 30 years old, where three wells contain less than 5 TU. The groundwater becomes younger from the central basin toward the Carson River, implying recharge in this area from either irrigation or recharge from the Carson River.

Only one thermal area was analyzed for enriched tritium, Prison Hot Springs. This was <5 TU, indicating water older than 30 years. A cold well in the Pinyon Hills thermal area was analyzed at <10 TU, also implying water older than about 30 years, assuming no dilution. No age estimates were made in the southern half of the basin since no enriched tritium data was obtained there.

In order to further investigate the origin of the younger water in the aquifer near the Carson River, the

river was gaged on November 6, 1981. The results are:

Site	Flow (cfs)
Carson River at USGS Gage	108.7
Carson River at New Brunswick Bridge	103.7
Mexican Ditch inflow	12.3
Mexican Ditch outflow	7.1
STP inflow	7.3

Summing the inflow minus the outflow produces a net loss of 7.5 cfs, with 5.2 cfs lost via Mexican ditch. This difference suggests that the river is losing water in the gaged section, but the difference is within gaging error. However, since the Carson River did not show a measurable gain in flow, it is possible that during certain times of the year, the Carson River loses water to the aquifer. The river was gaged in November, because that seems to be a time of decreased aquifer hydraulic head; the Prison Hot Spring usually dries up during the winter and early spring months.

6.5.5.2 Carbon Isotope Age Dating. A total of seven water samples were dated using carbon isotopes; Carson Hot Springs, Prison Hot Springs, and five non-thermal waters. Three of the non-thermal wells are along an east-west path in the north half of the basin: sample #52 is located 2 miles from the Ash Canyon outlet, and is a 515-ft deep well screened from 270 ft to 458 ft; sample #48 is 3 1/2 miles from Ash Canyon, from a 604 ft deep well, screened from 155 to 585 ft, and sample #177, a well of unknown depth, 5 miles from Ash Canyon and 1/2 mile from the Carson River. Two

other non-thermal samples were taken in the south half of the basin: sample #10 in the center of the basin is a 120-ft deep domestic well; and sample #112 in the south-center portion of the basin is from a 250-ft deep well from the Stewart Indian School. Data and analysis of these wells are presented in Table 6.4.

Corrections of the C-14 activity for dilution were done by three methods, all of which use the stable isotope, carbon-13. The first two methods assume recharge area  $\delta^{13}\text{C}/^{12}\text{C} = -25.0\%$  and carbonate along the flow path has an isotopic composition of  $\delta^{13}\text{C}/^{12}\text{C} = 0.0\%$ . These methods are from Pearson and White (1967), described in Section 3.4. Both of these methods tended to over-correct for dilution; hence, age determinations from these methods represent minimum corrected ages. The third method assumes that sample #177 with modern C-14 activity and  $\delta^{13}\text{C}/^{12}\text{C} = -17.3\%$  represents recharge area  $\delta^{13}\text{C}/^{12}\text{C}$ . Dissolving carbonate along the flow path is still assumed to be  $\delta^{13}\text{C}/^{12}\text{C} = 0.0\%$ . This type of correction method minimizes dilution and produces maximum corrected ages (results shown in Figure 6.35). More accurate carbon-14 age determinations could be made if recharge area  $\delta^{13}\text{C}/^{12}\text{C}$  and flow path  $\delta^{13}\text{C}/^{12}\text{C}$  in Eagle Valley were known. An attempt was made in this study to collect recharge area soil  $\text{CO}_2$  for measurement of  $\delta^{13}\text{C}/^{12}\text{C}$ , this did not prove successful (see Section 6.2.3).



NON-THERMAL WATERS

SAMPLE	ACTIVITY (dpm/g) (1)	ACTIVITY A <sup>h</sup> (pmc) (2)	δC13/12 ‰	ASSUMING MINIMUM DILUTION								
				P (3)	CORRECTED A <sup>h</sup> (pmc)	CORRECTED MIN. AGE (4)	P' (5)	CORRECTED A <sup>h</sup> (pmc)'	P'' (6)	CORRECTED A <sup>h</sup> (pmc)''	CORRECTED MAX. AGE	
52-NWest	12.367±.12	91.2%	-16.0	.640	142.5%	recent	.671	135.9%	.925	98.6%	100+80(recent)	
48-NCent	9.072±.11	66.9%	-15.7	.628	106.5%	recent	.666	100.5%	.908	73.7%	2500+100	
177-NEast	14.631±.14	107.9%	-17.3	.692	172.0%	recent	.524	205.9%	1.000	107.9%	recent	
10-Cent	12.407±.12	91.5%	-15.6	.624	146.6%	recent	.750	122.0%	.902	101.4%	recent	
112-SCent	11.160±.12	82.3%	-16.0	.640	128.6%	recent	.651	126.4%	.925	89.0%	960+80	

THERMAL WATERS

178-Prison Hot Spr.	11.757±.42	86.7%	-12.6	.504	172.0%	recent	.504	172.0%	.728	119.9%	recent
1002-Carson Hot Spr.	2.156±.09	15.9%	-14.7	.588	27.0%	10,800+80	.588	27.0%	.850	18.7%	13,860+30

(1)  $\delta^{13}C = \frac{A-A_0}{A_0} \cdot 1000$  or  $A = \frac{\delta^{13}C \cdot A_0}{1000} + A_0$  (4)  $t = -82671 \ln(A^h(\text{pmc}))$ , using  $T_{1/2} = 5730$  years

(2)  $A^h(\text{pmc}) = A/A_0$   $A_0 = 13.56$  dpm/g (5)  $P' = C_{\text{initial}}/C_{\text{total}}$ , see text

(3)  $p = \delta^{13}C / -25.$  = fraction dateable (6)  $P'' = \delta^{13}C / -17.3$ , see text

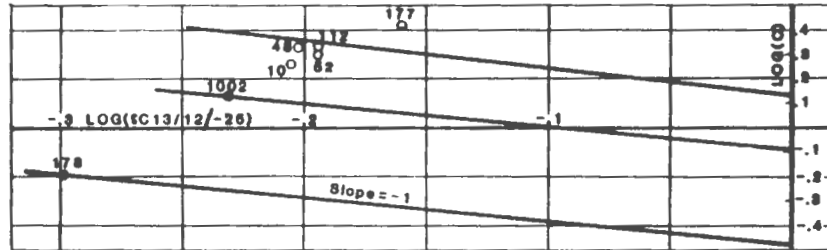


TABLE 6.4 Carbon Isotope Analysis

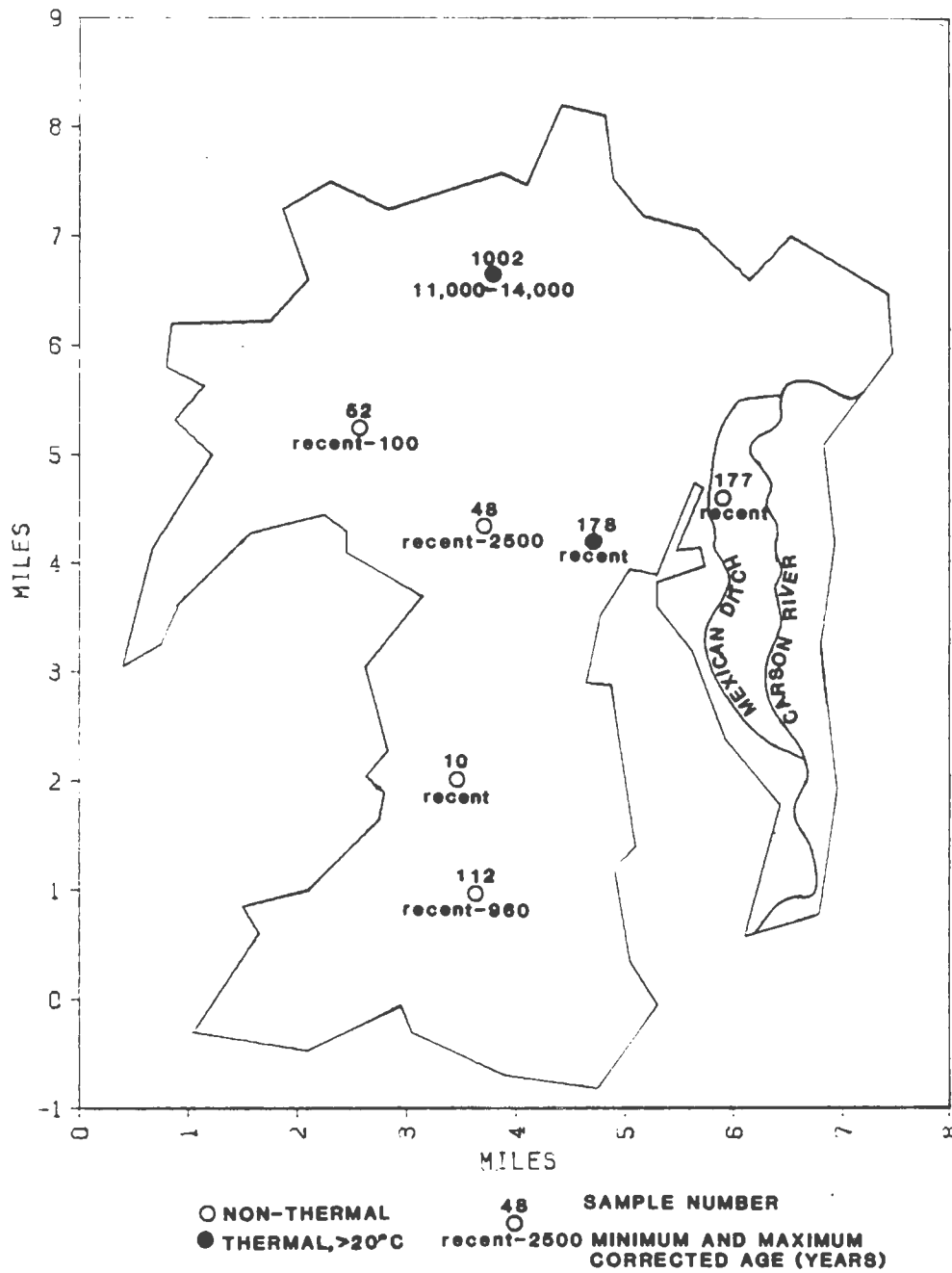


FIGURE 6.35 Apparent Groundwater Age from Carbon Isotope Data (years)

The five non-thermal samples, #52, #48, #177, #10, and #112, had approximately the same amount of contamination from flow path carbonate dissolving, since the  $\delta C_{13/12}$  range was small, -15.6 to -17.3‰. Use of the first two correction methods produced corrected C-14 activities from 100 to over 200 percent modern carbon (pmc), which clearly indicates that the methods are not accurately correcting the activity, since activities should not be substantially greater than modern C-14 activity (100%). Use of the third correction method of assuming minimum dilution produces the following results: sample #52 is recent; sample #48 is recent to 2500 years old; sample #10 is recent; and sample #112 is recent to 960 years old. This indicates that the youngest water in the basin is along the western mountain front and near the Carson River, with the oldest water in the north-central and south-central portions of the basin. These results agree with the age zones indicated by tritium data.

The two thermal waters were sampled at Prison Hot Springs (#178) and Carson Hot Springs (#1002). Major anion chemistry suggests that these two thermal waters are the same Na-SO<sub>4</sub> fluid. In contrast stable isotope data suggests that these are two different waters because Carson Hot Springs is depleted compared to the valley average, and Prison Hot Springs is not. Correction of the carbon-14 activity indicates that Prison Hot Springs contains modern

water. The  $\delta C_{13/12}$  value at Prison Hot Springs indicates a considerable amount of thermal/non-thermal water mixing. In contrast, correction of the carbon-14 activity of Carson Hot Springs water shows that it is between 10,800 and 13,900 years old. Clearly, Carson Hot Springs and Prison Hot Springs are very different waters; Prison Hot Springs is older than 30 years (based on tritium data) and less than 500 years (lower limit of C-14 age dating), whereas Carson Hot Springs is considerably older, between 11,000 and 14,000 years old. Depleted stable isotope content of Carson Hot Springs water could be due to a colder climate at the time of recharge 12,000 years ago.

The third thermal area Pinyon Hills thermal area was, unfortunately, not sampled for C-14 age dating. The only groundwater age data in the area is a non-thermal well which was analyzed for enriched tritium, producing  $<10$  TU, which indicates water older than about 30 years. Pinyon Hills is also depleted isotopically, possibly indicating recharge during a colder climate.

## CHAPTER 7

## CONCLUSIONS AND RECOMMENDATIONS

## 7.1 RECHARGE AREAS

Snow samples taken along an east-to-west traverse down Ash Canyon showed a regular isotopic shift of:

$$\delta D/H\% = 6.0(\delta 18/16\%) - 14 \text{ with } R^2 = 0.86$$

Since the slope is less than 8, this indicates non-equilibrium processes. There is a fair correlation between the isotopic content of the snow and elevation:  $\delta D/H = -9.9\%/1000 \text{ ft}$  ( $R^2 = 0.65$ ),  $\delta 18/16 = -1.48\%/1000 \text{ ft}$  ( $R^2 = 0.69$ ). These relationships indicate that regular isotopic shifts exist on the eastern (leeward) side of the Sierras, contrary to two previous studies. The 95% hyperbolic confidence limits were  $\pm 350 \text{ ft}$  for  $\delta D/H$  and  $\pm 300 \text{ ft}$  for  $\delta 18/16$ . The isotopic shifts obtained are similar to those observed on the western slope of the Sierras. Fractionation in the atmosphere during precipitation is probably the cause of the regular isotopic shift on the leeward side of the Sierras, since clouds are not observed to change altitude during precipitation.

Using electrical conductivity analyses of snow and lysimeter water, a percent recharge was calculated at five different elevations. This was used to estimate a maximum

potential recharge of 4000 acre-ft/year for the Eagle Valley Carson Range, lower than the Maxey-Eakin method annual recharge estimate of 8500 acre-ft. Electrical conductivity analysis produces an estimate of 8.5% of total precipitation is potential recharge versus an estimated 15% by the Maxey-Eakin method. Use of chloride enrichment from snow in the recharge area to mountain front water in the aquifer provides an estimate of 8.3% recharge of total precipitation. Since chloride levels in the groundwater are shown to increase dramatically across the valley, this method should be used with caution. The Maxey-Eakin method probably overestimates recharge in the Carson Range since it does not differentiate different percent evaporation zones in areas with over 20 in of annual precipitation, which includes most of the Carson Range. An average recharge elevation for the Carson Range was calculated at 8000 ft, indicating that high elevations are very important potential recharge areas.

The monitoring of lysimeters from May, 1981, to May, 1982, showed that recharge could occur all year because the ground is not frozen under the snowpack, and that recharge definitely occurs under the snowpack in the spring and summer. Local recharge to a meadow at 8500 feet in Ash Canyon in the spring produces interflow and discharge until the Fall which indicates that soil interflow is a significant component of streamflow. A similar local discharge system in Clear Creek Canyon at 5500 ft provides baseflow for Clear

Creek. Use of the Rayleigh evaporation formula for snow and lysimeter isotopic data from Ash Canyon at 8500 feet produced an estimated 13% and 23% evaporation for hydrogen and oxygen isotopes, respectively. Lower elevations have a smaller calculated evaporation, indicating inaccuracy in the data. A combined Rayleigh formula and mass balance formula developed in this study did not work for evaporation calculations.

Lysimeters in Ash Canyon showed a small isotopic depletion with lower elevation, opposite to the trend for snow. This may indicate that the lysimeter water was derived at higher elevations and traveled to lower elevations by subsurface flow. The lysimeter isotopic shift with elevation was very regular,  $R^2 = 0.99$  for  $\delta D/H$ ,  $R^2 = 0.78$  for  $\delta 18/16$  for 5 samples. Distillation of soil above  $160^\circ C$  fractionates oxygen isotopes, but not hydrogen isotopes; therefore, it is recommended that soils be distilled at less than  $160^\circ C$ .

Isotopic analyses of tree xylem water shows a very close correlation to lysimeter water, which indicates that trees do not fractionate soil water. When tree xylem water was distilled above  $160^\circ C$ , an organic compound was distilled with water, resulting in considerable enrichment of  $\delta 18/16$ . This may be due to the organic compound being partially composed of atmospheric  $CO_2$  which has a  $\delta 18/16$  of +41‰. This

organic distillate did not affect the hydrogen isotopic composition. Use of tree xylem water could prove to be of great importance in isotopic studies. Tritium dating of tree xylem water is inconclusive; it could represent modern water or a fifty-year average.

## 7.2 STREAM CHEMISTRY

Stream gaging of Ash Canyon Creek in August, 1981, and January, 1982, showed that most stream losses in the lower reach were due to phreatophytes with some infiltration in August, and very little infiltration in January. A small conductivity increase along the stream, in addition to a lack of isotopic enrichment, indicates little evaporation, and groundwater input to the stream is probably the cause of the conductivity increase. January stream conductivities were fairly large, indicating a substantial amount of groundwater input. The isotopic composition of Ash Canyon Creek did not change in six locations which suggests that all groundwater and surface inputs have the same average composition, i.e., average recharge composition.

Isotopic enrichment from snow at 8000 ft to stream samples taken in mid-May, late May, and late August indicate 0%, 4%, and 9% evaporation from deuterium analyses and 14%, 8%, and 19% evaporation from oxygen isotope analyses.



### 7.3 AQUIFER CHEMISTRY

A Piper diagram of well analyses from the Eagle Valley aquifer show anions ranging from alkaline waters in the mountain front wells to sulfate waters in thermal areas. Cations show a change from 20%  $Mg^{+2}$ , 50%  $Ca^{+2}$ , and 30%  $Na^{+}+K^{+}$  in mountain front wells to higher sodium in thermal areas. Two separate thermal fluids are defined by the major ion chemistry: a  $Na-SO_4$ , low TDS water of Carson Hot Springs, Prison Hot Springs, and STP well, and a high TDS- $Ca-SO_4$  water of the Pinyon Hills area.

Contour maps of sulfate, chloride, and sodium ion-concentrations of 60 wells show that there is no trend in the northeast quarter and southern half of the basin, where there are no thermal waters. Sulfate, chloride, and sodium increases appear only near thermal areas indicating that  $SO_4^{-2}$ ,  $Cl^{-}$ , and  $Na^{+}$  increases are due to mixing of thermal and non-thermal fluids. Electrical conductivity trend surface maps also indicate mixing of thermal and non-thermal fluids since the map without thermal wells has essentially the same trend as the map with thermal wells. Deuterium concentrations of Prison Hot Springs are the same as nearby non-thermal wells, indicating thermal/non-thermal fluid mixing.  $\delta Cl^{35}/Cl^{37}$  value of Prison Hot Springs also indicates mixing. Therefore, the major ion increase of non-thermal waters in the northeast half of the basin could be from mixing of

thermal/non-thermal fluids.

Although almost all ions increase in concentration to the east (flow direction) in the northern half of the basin, sulfate, chloride, and sodium concentrations decrease near the Carson River in the northeast corner of the basin, possibly indicating some inflow into the aquifer from the Carson River. Alkalinity, calcium, and magnesium concentrations increase regularly toward the northeast, suggesting flow in that direction. Potassium concentrations in the valley did not show any regular trends, nor was it useful in delineating thermal waters.

Four non-thermal wells sampled in 1965 and 1981 showed a significant decrease in sulfate, chloride, and magnesium concentrations, although thermal springs over the same time period showed no major ion changes.

Based on calculated isotopic composition of potential recharge to the basin, areas along the western margin of the basin show recharge from the Carson Range predominantly in warmer months, since water in these areas are enriched. This suggests stream channel infiltration, which is substantiated by tritium dating. The basin center, however, is more depleted isotopically, suggesting recharge via deep percolation. Infiltration in recharge areas and fracture flow from the mountains to the aquifer would contain less evaporated (more depleted) water. Therefore, evidence

suggests that the Eagle Valley aquifer derives recharge from both stream channel infiltration and deep percolation.

Tritium age dating of non-thermal waters in Eagle Valley indicate fast recharge to the aquifer from the Carson Range, probably from stream channel infiltration. Both carbon-14 and tritium age dating indicate that the oldest water is in the north central and south-central parts of the basin. Age data in the north central basin includes several wells containing pre-bomb tritium, and one well containing water between 0 and 2500 years old (by carbon-14 age dating). One well in the south-central basin contains recent to 1000 year old water. Post-1950 tritium in two wells and one modern carbon-14 age all near the Carson River in the north-east portion of the basin suggest that water is flowing into the aquifer, either from the Carson River or from irrigation.

Stable isotope data show that Carson Hot Springs and the Pinyon Hills area are depleted isotopically compared to surrounding cold water possibly indicating recharge during a colder climate. The STP well/Prison Hot Springs area has the same composition as surrounding cold waters. Although Carson Hot Springs and Prison Hot Springs waters are similar chemically, they are isotopically different, which suggests different flow regimes. Prison Hot Springs contains  $<5$  TU, so it is greater than 30 years old, and is of recent age

using carbon-14 dating techniques. Carson Hot Springs is between 11,000 and 14,000 years old. The third thermal area, Pinyon Hills, was not sampled. Non-thermal water in Pinyon Hills area is probably greater than 30 years old, since tritium was counted at <10 TU.

Certain areas in Eagle Valley have a lack of data. More stable isotope samples are needed in the southern half of the valley to confirm possible different recharge mechanisms. Snow, tree, and lysimeter water samples should be taken on a ridge to see if snow isotopic trends are confirmed in soil water trends. A vertical core of a tree should be taken for tritium age dating analysis to confirm if trees contain a several year average of soil water.

Since oxygen isotopes have been shown to behave more erratically, it is recommended that only deuterium be used in future studies. Only isotopic slope data ( $\Delta\delta D/H/\Delta\delta 18/16$ ) information is not lost by analyzing solely for deuterium. This would also save half the cost of analysis. Tree xylem water should be studied further in other areas to confirm if it can be used to replace lysimeter water sampling.

$^{32}\text{Si}$  with a half-life of 330 years and  $^{37}\text{Ar}$  with a half-life of 270 years (Davis and Bentley, 1981) could be used on the thermal areas to estimate groundwater age, since it seems that these areas contain water older than can be determined by tritium, and younger than can be determined by carbon-14 age dating.

## REFERENCES CITED

- Allison, G.B., and Hughs, M.W., 1975, The use of environmental tritium to estimate recharge to a South-Australian aquifer: *Journal of Hydrology*, vol. 26, p. 245-254.
- American Public Health Association, 1980, Standard methods for the examination of water and wastewater, 15th edition: American Public Health Association, New York, 872 p.
- Arteaga, F., and Durbin, T., 1978, Development of a relation for steady-state pumping rate for Eagle Valley groundwater basin, Nevada: U.S. Geological Survey Open-File Report 79-261, 44 p.
- Back, W., and Hanshaw, B., 1965, Chemical geohydrology in *Advances in Hydrosience*, Volume 2, Academic Press, New York, p. 49-109.
- Bamberg, S., and Friesen, H., 1972, A precipitation estimation technique for developing isohyets in an arid area using vegetation and topographic parameters: *Water Resources Bulletin*, vol. 6, no. 8, p. 1108-1119.
- Bingler, E., 1977, New Empire quadrangle geologic map: Nevada Bureau of Mines and Geology, 1:24,000.
- Bohn, H., McNeil, B., and O'Conner, G., 1979, Soil chemistry: John Wiley & Sons, New York, 329 p.
- Bostick, K.A., 1978, A stable isotope investigation of recharge to the Tucson basin aquifer from the Santa Cruz River: unpublished M.S. thesis, Department of Hydrology and Water Resources, University of Arizona, Tucson, Arizona, 51 p.
- Boyer, P.D., Graves, D.J., Suelter, C.H., and Dempsey, M.E., 1961, Simple procedure for conversion of oxygen of orthophosphate or water to carbon dioxide for oxygen-18 determination: *Analytical Chemistry*, vol. 33, no. 13, p. 1906-1909.
- Burk, R.L., and Stuiver, M., 1981, Oxygen isotope ratios in trees reflect mean annual temperature and humidity: *Science*, vol. 211, p. 1417-1419.
- Campana, M.E., 1976, Application of C-14 groundwater ages in calibrating a flow model of the Tucson basin aquifer, Arizona in *Proceedings of the 1976 American Water Resources Association Meeting*, Tucson, Arizona, p. 197-202.

- Campana, M.E., 1981, Environmental isotopes in natural recharge investigations: Water Resources Research Conference, May, 1981.
- Chebotrev, I.I., 1955, Metamorphism of natural waters in the crust of weathering: *Geochemica et Cosmochemica Acta*, vol. 8, p. 22-48, 137-170, 198-212.
- Craig, H., 1961a, Isotopic variation in meteoric waters: *Science*, vol. 133, no. 3465, p. 1702-1703.
- Craig, H., 1961b, Standard for reporting concentrations of deuterium and O-18 in natural waters: *Science*, vol. 133, no. 3467, p. 1833-1834.
- Currie, L.A., 1972, The evaluation of radiocarbon measurements and inherent statistical limitations in age resolution in Proceedings of the 8th International Conference on Radiocarbon Dating, Wellington, New Zealand, Oct., 1972, vol. 1, p. 597-611.
- Dansgaard, W., 1964, Stable isotopes in precipitation: *Tellus XVI*, no. 4, p. 436-468.
- Davis, J.C., 1973, Trend surfaces in Statistics and Data Analysis in Geology: John Wiley & Sons, Inc., New York, p. 322-352.
- Davis, S., and Bentley, H., 1981, Dating groundwater, a short review: American Chemical Society Symposium, Series 176, p. 187-222.
- Eakin, T.E., Maxey, G.B., Robinson, T.W., Fredricks, J.C., and Loeltz, O.J., 1951, Contributions to the hydrology of eastern Nevada: Nevada State Engineer's Office, Water Resources Bulletin 12, 171 p.
- Edmunds, W.M., and Wright, E.P., 1979, Groundwater recharge and paleoclimate in the Sirte and Kufra basin, Libya: *Journal of Hydrology*, vol. 40, no. 3-4, p. 215-241.
- Eisinger, V.J., 1960, Geology of the Prison Hill - Brunswick Canyon area, Ormsby and Douglas Counties, Nevada: unpublished M.S. thesis, University of Nevada, Reno, 69 p.
- Epstein, S., Thompson, P., and Yapp, C., 1977, Oxygen and hydrogen ratios in plant cellulose: *Science*, vol. 198, no. 4323, p. 1209-1215.

- Epstein, S., Yapp, C.J., and Hall, H., 1976, The determination of  $\delta D/H$  ratio of non-exchangeable hydrogen in cellulose extracted from aquatic and land plants: Publications of the Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California, no. 2667, p. 241-251.
- Fournier, R.O., and Truesdell, A.H., 1973, An empirical Na-K-Ca geothermometer for natural waters: *Geochemica et Cosmochemica Acta*, vol. 37, p. 1255-1275.
- Freeze, R.A., 1972, Role of subsurface flow in generating surface runoff, 1. base flow contributions to channel flow: *Water Resources Research*, vol. 8, no. 3, p. 609-623.
- Freeze, R.A., and Cherry, J.A., 1979, *Groundwater*: Prentice-Hill, Inc., Englewood Cliffs, New Jersey, 604 p.
- Friedman, I., Redfield, A.C., Schoen, B., and Harris, J., 1964, The variation of deuterium content of natural waters in the hydrologic cycle: *Reviews of Geophysics*, vol. 2, no. 1, p. 177-224.
- Friedman, I., and Smith, G.I., 1970, Deuterium content of snow cores from Sierra Nevada area: *Science*, vol. 169, p. 467-470.
- Fritz, P., and Fontes, J.Ch., 1980, *Handbook of environmental isotope geochemistry, volume 1, the terrestrial environment, A*: Elsevier Scientific Publishing Company, Amsterdam, Netherlands, 545 p.
- Fritz, P., Suzuki, O., Silva, C., and Salati, E., 1981, Isotope hydrology of groundwaters in the Pampa Del Tamarugal, Chile: *Journal of Hydrology*, vol. 53, p. 161-184.
- Gat, J.R., 1971, Comments on the stable isotope method in regional groundwater investigations: *Water Resources Research*, vol. 7, no. 4, p. 980-993.
- Geyh, M.A., 1972, On the determination of the initial C-14 content in groundwater in *Proceedings of the 8th International Conference on Radiocarbon Dating*, Wellington, New Zealand, Oct., 1972, p. 342-352.
- Gray, J., and Thompson, P., 1977, Climatic information from  $\delta^{18}O/\delta^{16}O$  analysis of cellulose, lignin, and whole wood from tree rings: *Nature*, vol. 270, p. 708-709.
- Hansen, D.S., 1978, Tritium movement in the unsaturated zone, Nevada Test Site: unpublished M.S. thesis, University of Nevada, Reno, 107 p.

- Hardman, G., 1936, Nevada precipitation and acreages of land by rainfall zones: Nevada University Agricultural Experimental Station Mimeo. Rept. and Map, 10p.
- Hess, J.W., 1981, Sample preparation for oxygen-18 analysis: oral communication.
- Ingraham, N., 1982, Environmental isotope hydrology of the Dixie Valley geothermal system, Dixie Valley, Nevada: unpublished M.S. thesis, University of Nevada, Reno, 96 p.
- International Atomic Energy Agency, 1971, Sampling of water for C-14 analysis: IAEA, Vienna, Austria, 3 p.
- Johns, F.B., 1975, Handbook of radiochemical analytical methods, program element 1HA325: National Environmental Research Center, Office of Research and Development, U.S. Environmental Protection Agency, Las Vegas, Nevada, 5 p.
- Judy, C., and Meiman, J.R., 1970, Deuterium variations in an annual snowpack: Water Resources Research, vol. 6, no. 1, p. 125-129.
- Katzer, T., 1980, Carson City quadrangle general groundwater map: Nevada Bureau of Mines and Geology, Map 1Af, 1:24000.
- Kirkham, R.M., 1976, Environmental geology of the western Carson City, Nevada area: unpublished M.S. thesis, University of Nevada, Reno, 97 p.
- Klieforth, H., 1981, Desert Research Institute Atmospheric Sciences Center, published and unpublished data.
- Lerman, J.C., 1972, Soil-CO<sub>2</sub> and groundwater; carbon isotope compositions in Proceedings of the 8th International Conference on Radiocarbon Dating, Wellington, New Zealand, Oct. 1972, vol. 1, p. D93-D105.
- Lukas, R.B., 1980, Contour program: University of Hawaii.
- Mahin, D.A., 1978, Analysis of groundwater flow in the Edwards limestone aquifer, San Antonio area, Texas, unpublished M.S. thesis, University of Nevada, Reno, 49 p.
- McKinney, R.F., 1976, Environmental geology of southeast Carson City, Nevada: unpublished M.S. thesis, University of Nevada, Reno, 135 p.



- Mifflin, M.D., 1968, Delineation of groundwater flow systems in Nevada: Desert Research Institute Technical Report Series H-W, no. 4, 89 p.
- Mook, W.G., 1972, On the reconstruction of the initial C-14 content of groundwater from the chemical and isotopic composition in Proceedings of the 8th International Conference on Radiocarbon Dating, Wellington, New Zealand, Oct., 1972, vol. 1, p. 342-352.
- Moore, J.G., 1969, Geology and mineral resources of Lyon, Douglas, and Ormsby Counties, Nevada: Nevada Bureau of Mines and Geology Bulletin 75, 45 p.
- Muir, K.S., and Coplen, T.B., 1981, Tracing ground-water movement by using the stable isotopes of oxygen and hydrogen, upper Penitancia Creek alluvial fan, Santa Clara Valley, California: U.S. Geological Survey Water-Supply Paper 2075, 18 p.
- Nehring, N.L., 1980, Geochemistry of Steamboat Springs, Nevada: U.S. Geological Survey Open File Report 80-887, 61 p.
- Nir, A., 1964, On the interpretation of tritium age measurements of groundwater: Journal of Geophysical Research, vol. 69, no. 12, p. 2589-2595.
- Osmond, B., 1981, Stable isotopes and plants: Seminar given at the Desert Research Institute, October, 1981.
- Parizek, R.R., and Lane, B.E., 1970, Soil-water sampling using pan and deep pressure-vacuum lysimeters: Journal of Hydrology, vol. 11, p. 1-21.
- Payne, B.R., 1972, Isotope hydrology in, Advances in Hydroscience, Academic Press, New York, p. 95-138.
- Pearson, W., 1970, Collection and preparation of water samples for C-14 analysis: U.S. Geological Survey Pamphlet, 3 p.
- Pearson, W., and White, D.E., 1967, Carbon-14 ages and flow rates in Carrizo sand, Atascosa, Texas: Water Resources Research, vol. 3, no. 1, p. 251-261.
- Pease, R.C., 1980, Genoa quadrangle geologic map: Nevada Bureau of Mines and Geology, 1:24000.
- Piper, A.M., 1944, A graphical procedure in the geochemical interpretation of water analysis: Transactions of the American Geophysical Union, vol. 25, p. 914-923.

- Rabinowitz, D.D., Gross, G.W., and Holmes, C.H., 1977a, Environmental tritium as a hydrometeorologic tool in the Roswell Basin, New Mexico, I. Tritium input function and precipitation-recharge relation: *Journal of Hydrology*, vol. 32, p. 3-17.
- Rabinowitz, D.D., Gross, G.W., and Holmes, C.H., 1977b, Environmental tritium as a hydrometeorologic tool in the Roswell Basin, New Mexico, II. Tritium patterns in groundwater: *Journal of Hydrology*, vol. 32, p. 19-33.
- Reid, J.S., 1911, Geomorphogeny of the Sierra Nevada northeast of Lake Tahoe: California University, Department of Geology, Bulletin 6, p. 89-161.
- Simpson, E.S., Thorud, D.B., and Friedman, I., 1970, Distinguishing seasonal recharge to groundwater by deuterium analysis in southern Arizona: *World Water Balance*, p. 112-121.
- Smith, D.B., Downing, R.A., Monkhouse, R.A., Otlet, R.L., and Pearson, F.J., 1976, The age of groundwater in the chalk of the London Basin: *Water Resources Research*, vol. 12, no. 3, p. 392-404.
- Smith, G.I., Friedman, I., Klieforth, H., and Hardcastle, K., 1979, Areal distribution of deuterium in eastern California precipitation, 1968-1969: *Journal of Applied Meteorology*, vol. 18, no. 2, p. 172-188.
- Teledyne Isotopes, 1981, Sample contamination in radiocarbon age determination: Teledyne Isotopes, New Jersey.
- Trexler, D., 1977, Carson City folio geologic map, Carson City quadrangle: Nevada Bureau of Mines and Geology Map 1Ag, 1:24000.
- Trexler, D.T., Flynn, T., and Koenig, B.A., 1979, Low to moderate temperature geothermal resource assessment of Nevada: unpublished final report under contract to U.S. Department of Energy, No. ET-78-5-08-1556.
- Trexler, D.T., Koenig, B.A., Flynn, T., and Bruce, J.L., 1980, Assessment of the geothermal resources of Carson-Eagle Valleys and Big Smoky Valley, Nevada: Nevada Bureau of Mines and Geology, 232 p.
- Trudeau, D.A., 1979, Hydrogeologic investigation of the Littlefield Springs: unpublished M.S. thesis, University of Nevada, Reno, 136 p.
- U.S. Geological Survey, 1956, Carson City, Nevada; 15-minute quadrangle: U.S. Geological Survey, Denver, Colorado.

- U.S. Geological Survey, 1956, Dayton, Nevada; 15-minute quadrangle: U.S. Geological Survey, Denver, Colorado.
- U.S. Geological Survey, 1979, 1:100,000-scale metric topographic map of Carson City, Nevada: U.S. Geological Survey, Denver, Colorado.
- Watson, P., Sinclair, P., and Waggoner, R., 1976, Quantitative evaluation of a method for estimating recharge to the desert basins of Nevada: *Journal of Hydrology*, vol. 31, p. 335-357.
- Weisz, P.B., 1971, *The science of biology*: McGraw-Hill, Inc., New York, 4th ed., p. 188-191.
- Wershaw, R.L., Friedman, I., Heller, S.J., and Frank, P.A., 1966, Hydrogen isotopic fractionation of water passing through trees in eds. G.D. Hobson and G.C. Speers, *Advances in Organic Geochemistry, Proceedings of the Third International Congress*, Pergamon Press, London, p. 55-67.
- White, D.E., 1968, Hydrology, activity, and heat flow of the Steamboat Springs thermal system, Washoe County, Nevada: U.S. Geological Survey Professional Paper 458C, 109 p.
- Wigley, T.M.L., 1975, Carbon-14 dating of groundwater from closed and open systems: *Water Resources Research*, vol. 11, no. 2, p. 324-328.
- Winograd, I., 1971, Sludge cone for C-14 sampling: U.S. Geological Survey pamphlet, 2 p.
- Winograd, I., and Friedman, I., 1972, Deuterium as a tracer of regional ground-water flow, southern Great Basin, Nevada and California: *Geological Society of America Bulletin*, vol. 83, p. 3691-3708.
- Worts, Jr., G.F., and Malmberg, G.T., 1966, Hydrologic appraisal of Eagle Valley, Ormsby County, Nevada: *Water Resources - Reconnaissance Series Report 39*, State of Nevada Department of Conservation and Natural Resources, 55 p.
- Yevjevich, V., 1972, *Probability and statistics in hydrology*: Water Resources Publications, Fort Collins, Colorado, p. 247-258.
- Ziegler, H., Osmond, C.B., Stichler, W., and Trimborn, P., 1976, Hydrogen isotope discrimination in higher plants; correlations with photosynthetic pathway and environment: *Planta*, vol. 128, p. 85-92.

- Zito, R.D., Davis, S., Bentley, H., and Fritz, P., 1980, Possible subsurface production of carbon-14: Geophysical Research Letters, vol. 7, no. 4, p. 235-238.
- Zones, C.P., 1958, Petrographic and petrofabric study of the metamorphic rocks north of Carson City, Ormsby County, Nevada: unpublished M.S. thesis, University of Nevada, Reno, 80 p.

APPENDIX 1 EAGLE VALLEY - MAJOR ION ANALYSES

SAMPLE NUMBER	LOCATION (T,R,S)	LOCATION (E,N)*	TEMP. °F	pH	S+C (UM/CM)	HCO3 (PPM)	CL (PPM)	SO4 (PPM)	NA (PPM)	K (PPM)	CA (PPM)	MG (PPM)	STNO	SOURCE	WELL OR OWNER	LOG #
2	14N,18E,36	-4.0,+2.0	7.84	70.	41.8	0.3	0.01	3.80	1.28	7.76	1.43	17.		N. CTN CK		
3	14N,20E,32CC	3.07,7.05	10.0	7.41	198.	103.	3.2	3.9	13.	3.8	18.9	4.24	38.		8036	
4	15N,19E,10C	1.59,+6.10	13.0	7.32	167.	103.	1.2	1.3	13.4	2.38	16.7	3.96	39.		7015	
8	15N,19E,15CB	5.00,+4.36	31.0	9.20	473.	40.0	17.8	145.	80.8	.74	11.6	.03	31.		14436	
10	15N,20E,29P	3.45,+2.02	12.	7.18	183.	111.	2.3	2.2	12.7	0.90	19.4	5.22	34.		13315	
11	15N,20E,32	3.30,+1.89	9.0	7.03	201.	100.1	4.3	5.0	12.6	1.21	22.4	4.47	34.		18880	
13	15N,19E,34	-.99,+1.11	6.5	7.00	123.	77.7	0.7	0.01	8.50	1.08	13.1	3.05	34.		9540	
5	15N,19E,12	1.37,+5.16	21.0	7.00	138.	85.6	0.6	1.30	6.70	1.80	16.4	3.22	25.		5111	
12	14N,20E,6	2.24,+0.23	2.0	6.58	264.	107.	12.2	25.1	26.1	1.15	22.8	3.97	46.		11647	
15	15N,19E,13	1.38,+4.15	12.0	6.18	247.	145.	3.6	5.0	17.1	2.79	27.8	4.95	28.		TENNANT	
16	15N,20E,31	2.72,+1.95	9.5	8.06	177.	91.7	3.2	3.00	13.2	1.20	14.7	5.50	24.		19751	
19	15N,20E,23	6.55,3.05	15.0	7.66	669.	183.	35.8	121.	39.2	2.84	93.2	5.16	23.		6452	
112	15N,20E,32	3.71,+1.00	11.0	7.84	244.	128.	5.0	9.1	24.7	1.27	21.5	3.13	35.		10351	
48	15N,20E,17	3.76,+4.35	15.0	7.82	220.	125.	1.7	9.4	18.2	1.22	20.7	4.55	23.		10564	
49	15N,20E,6DD	2.90,+6.20	30.0	8.22	186.	105.	2.6	6.1	23.8	1.29	14.9	0.84	33.		13767	
50	15N,19E,12AA	2.00,+5.89	13.0	7.23	156.	88.	1.3	1.9	10.4	2.61	15.7	2.75	29.		15299	
51	15N,19E,12AD	2.00,+5.65	13.0	7.27	166.	97.	1.8	2.7	10.2	2.68	16.5	3.71	27.		12438	
52	15N,20E,7	2.71,+5.24	15.0	7.44	205.	124.	1.4	1.9	11.4	1.97	21.3	5.74	29.		11262	
17	15N,20E,90A	4.78,+5.42	15.5	7.92	1070.	214.	71.3	262.	138.	6.89	54.4	18.1	53.		6275	
172	15N,19E,12AB	1.37,+5.02	16.	6.73	132.	79.4	1.2	1.4	6.59	2.08	15.4	3.10	26.		16322	
173	15N,20E,8BC	3.05,+5.60	22.	7.67	337.	154.	25.3	12.7	24.1	2.78	35.2	6.43	34.		6188	
174	15N,20E,10DA	5.79,+5.38	14.	7.30	409.	188.	13.8	44.4	40.4	2.77	37.2	7.32	37.		19827	
175	15N,20E,16C	4.23,+4.32	9.	7.67	216.	113.	2.8	19.3	23.2	1.27	19.6	3.19	35.		13689	
176	15N,20E,15	5.73,+4.51	9.5	6.69	283.	154.	10.6	31.4	40.7	2.46	30.6	5.74	39.		1600	
177	15N,20E,15	5.87,+4.52	14.	7.81	300.	159.	10.7	31.2	29.7	2.45	43.8	3.10	32.		GRAHAM	
178	15N,20E,16DB	4.77,+4.20	35.	8.49	450.	39.3	21.0	152.	84.5	2.29	17.0	236	37.		P. HOT SPR.	
179	15N,20E,14C	4.24,+4.27	9.5	6.80	163.	161.	2.5	6.1	19.7	1.47	26.1	7.15	27.		3841	
180	15N,20E,18	1.95,+4.63	12.5	6.49	210.	148.	3.3	2.15	9.30	3.06	27.2	12.0	30.		2680	
165	15N,19E,12	0.95,+5.47	11.1	7.81	95.0	51.0	0.9	0.01	5.07	1.63	9.77	1.38	19.		ASH CN CK	
1000	15N,20E,15CB	5.00,+4.36	32.	7.83	501.	47.4	19.7	132.	89.5	.632	11.6	0.01	36.		T 14436	
1001	15N,20E,23DA	6.36,+3.37	41.	7.26	2470.	26.4	34.3	843.	173.	5.90	270.	0.14	44.		T HURRAY	
1002	15N,20E,5AD	3.85,+6.45	50.	8.84	515.	83.3	27.4	89.4	98.5	1.64	2.16	0.01	60.		T C. HOT SPR.	
1003	15N,20E,23CD	6.35,+3.20	5.	8.09	299.	111.	11.3	37.7	22.4	3.68	25.3	6.43	38.		T C. RIVER	
1004	15N,20E,7	2.50,+5.50		7.89		117.	5.0	1.	20.	2.	21.	5.			T SULLIVAN	
1005	15N,20E,23	6.50,+3.50	44.	8.03		22.	36.	1056.	214.	.01	280.	1.14			T BENNETT	
1006	15N,20E,16DB	4.24,+4.20		8.8		48.	21.	148.	82.	2.	13.5	3.30			T P. HOT SPR.	
1007	14N,20E,4	4.50,+0.50		7.73		146.	9.	7.	15.	1.	32.	7.			T MED. P. WELL	
1008	15N,20E,8	3.50,+5.50		7.25		132.	7.	2.	20.	3.	20.	7.			T DICKERSON	
1009	15N,20E,5	3.50,+6.50		8.87		87.	27.	94.	96.	1.	7.	0.01			T CARSON	
1010	15N,20E,6	2.50,+6.50		7.58		132.	4.	1.	19.	2.	19.	8.			T ODOM	
1050	15N,20E,8CC	3.20,+5.20	13.3	8.4	205.	119.	4.0	4.4	14.	.01	23.	5.7			WM WAGNER	
1013	15N,19E,23	0.65,+3.50		7.		115.	1.	19.	16.	.01	24.	5.			W	
1014	15N,19E,32D	-2.5,+1.50		7.	108.	66.	1.	24.	5.	3.	12.	3.	24.		W	
1016	15N,20E,2	6.50,+6.50		8.		185.	13.	10.	31.	.01	35.	8.			W	
1017	15N,20E,3CD	5.35,+6.15	11.	8.1	394.	208.	17.	28.	40.	.01	40.	9.			WM 3760	
1018	15N,20E,4BD	4.35,+6.65	15.	7.8	434.	84.	28.	87.	68.	.01	18.	2.			WM ORMSBY C.	
1019	15N,20E,9DA	4.85,+5.35	9.	7.8	893.	244.	48.	1760.	122.	.01	55.	12.			WM M. DEWITT	
1020	15N,20E,10AD	5.85,+3.65	9.	7.1	531.	184.	21.	89.	52.	.01	49.	9.			WM 5125	
1021	15N,20E,11	6.50,+5.50		8.		183.	20.	67.	48.	.01	53.	11.			W	
1022	15N,20E,14	6.50,+4.50		8.		54.	43.	840.	199.	.01	186.	21.			W	
1023	15N,20E,15BC	5.15,+4.45	12.	8.5	396.	144.	14.	56.	72.	.01	15.	1.			WM 5589	
1024	15N,20E,16CA	4.35,+4.35	10.	7.7	217.	102.	5.	24.	25.	.01	20.	3.			WM 3841	
1026	15N,20E,16	3.90,+4.10		8.		102.	6.	24.	190.	.01	27.	2.			W	
1025	15N,20E,17DD	4.23,+4.32		8.		104.	59.	93.	28.	6.	61.	10.			W	
1027	15N,19E,10C	1.59,+6.10	16.1	7.7	181.	92.	3.2	4.8	12.	.01	15.	9.1			WM 7105	
1028	15N,20E,8DB	3.65,+3.35	15.6	8.1	194.	104.	3.5	12.	29.	.01	13.	1.8			WM STEINHEIMER	
1029	15N,20E,18BA	2.35,+4.85	15.6	8.2	181.	112.	2.1	4.4	10.	.01	22.	5.4			WM J. WINTERS	
1030	15N,20E,21CA	4.35,+3.35	10.6	7.6	223.	146.	4.7	6.0	20.	.01	25.	6.2			WM 7638	
1031	15N,20E,28CD	4.35,+2.19	6.7	7.2	194.	116.	5.0	12.	21.	.01	19.	5.0			WM S. GRIFFITH	
1032	15N,20E,29BC	3.15,+2.65		8.0	178.	80.	6.4	6.4	12.	.01	15.	4.0			WM 5558	
1033	15N,20E,32DD	3.71,+1.00	13.3	7.3	247.	144.	7.0	12.	23.	.01	27.	5.7			WM	
1034	14N,20E,32CC	3.07,7.05	20.	8.1	229.	128.	6.3	14.	24.	.01	23.	4.7			WM 8036	
1035	15N,20E,5AD	3.85,+6.45	48.9	9.3	506.	82.	29.	96.	96.	.01	2.6	0.4			WM C. HOT SPR.	
1036	15N,20E,19	2.95,+3.35		8.		59.	6.	173.	63.	.01	38.	1.			W	
1037	15N,20E,20CC	3.15,+3.15		7.		293.	23.	65.	102.	.01	34.	13.			W	
1039	15N,19E,12CD	1.35,+5.15	9.4	7.4	119.	72.	2.6	6.0	10.	.01	14.	2.7			WM 5111	
1041	15N,20E,5CD	3.15,+6.15	12.2	7.8	189.	116.	3.1	4.8	17.	.01	21.	3.8			WM 6492	
1046	15N,20E,6AB	2.65,+6.85	7.8	7.4	232.	120.	6.2	7.6	17.	.01	22.	5.6			WM 4045	
1045	14N,20E,4	4.50,+0.50		8.		146.	3.	58.	52.	.01	26.	5.			W	
1047	14N,20E,8	3.50,+0.50		9.		17.	4.	959.	259.	.01	176.	5.			W	
1048	15N,19E,33	-.99,+1.11		8.		112.	1.	14.	20.	.01	24.	2.			W	
1049	15N,20E,7CA	2.35,+5.35	8.9	7.5	169.	103.	2.2	5.2	12.	.01	21.	3.8			WM 4328	

\*\* T: TREXLER, KOENIG, FLYNN, BRUCE, 1980  
 WM: WORTS AND HALBERG, 1966  
 W: WATER ANALYSIS DATA SYSTEM (WADS-DRI)  
 ALL OTHER DATA FROM THIS STUDY

(0.0) LOCATED AT T14N, R19E, SECTION 10AAAA

## APPENDIX 2 EAGLE VALLEY - STABLE ISOTOPE ANALYSES

NUMBER	TYPE	DATE	OXYGEN-18	DEUTERIUM	ELEVATION	LOCATION/OTHER
53	SNOW	4-14-81	-18.82	-131.0	9214	ASH CN
20	SNOW	3-14-81	-17.94	-125.9	9214	ASH CN
55	SNOW	4-14-81	-16.65	-110.2	9000	ASH CN
56	SNOW	4-14-81	-16.59	-110.8	8750	ASH CN
57	SNOW	4-14-81	-16.64	-116.1	8750	ASH CN
72	SNOW	4-14-81	-17.46	-115.5	8750	ASH CN
58	SNOW	4-14-81	-16.78	-115.8	8500	ASH CN
59	SNOW	4-14-81	-16.16	-112.4	8250	ASH CN
61	SNOW	4-14-81	-14.92	-109.7	8000	ASH CN
53	SNOW	4-14-81	-15.62	-104.2	7750	ASH CN
64	SNOW	4-14-81	-16.37	-112.7	7500	ASH CN
65	SNOW	4-14-81	-15.32	-107.3	7250	ASH CN
67	SNOW	4-14-81	-14.37	-103.1	7000	ASH CN
58	SNOW	4-14-81	-14.81	-99.6	6750	ASH CN
59	SNOW	4-14-81	-14.18	-102.7	6750	ASH CN
73	SNOW	4-14-81	-14.82	-102.7	6750	ASH CN
70	SNOW	4-14-81	-15.27	-105.1	6250	ASH CN
71	SNOW	4-14-81	-15.47	-106.2	6250	ASH CN
75	SNOW	5-13-81	-13.4	-102.3	3500	ASH CN
75	SNOW	5-28-81	-13.1	-98.5	8500	ASH CN
71	SOILH2O	5-28-81	-14.8	-107.7	8500	ASH CN, LYSIMETER
103	SOILH2O	6-11-81	-14.8	-104.7	8000	ASH CN, DISTILLED SOIL
119	SOILH2O	7-09-81	-14.0	-104.7	8500	ASH CN, LYSIMETER
120	SOILH2O	7-09-81	-14.0	-107.7	8000	ASH CN, LYSIMETER
121	SOILH2O	7-09-81	-13.9	-108.7	7500	ASH CN, LYSIMETER
122	SOILH2O	7-09-81	-10.0	-111.7	7000	ASH CN, DISTILLED SOIL
123	SOILH2O	7-09-81	-15.2	-114.7	6200	ASH CN, LYSIMETER
109	SOILH2O	6-25-81	-10.1	-106.7	6400	CLEAR CN, DISTILLED SOIL
117	SOILH2O	7-09-81	-10.9	-109.7	6400	CLEAR CN, DISTILLED SOIL
127	TREE	7-09-81	-11.7	-107.7	8500	ASH CN, LODGEPOLE, DIST. YELLOW
128	TREE	7-09-81	-13.0	-108.7	8000	ASH CN, JEFFERY, DIST. SLT YELLOW
127	TREE	7-09-81	-13.6	-105.7	7500	ASH CN, JEFFERY, DIST. SLT YELLOW
130	TREE	7-09-81	-14.4	-108.7	7500	ASH CN, JEFFERY, DIST. CLEAR
131	TREE	7-09-81	-12.4	-104.7	7000	ASH CN, JEFFERY, DIST. SLT YELLOW
132	TREE	7-09-81	-11.3	-103.5	6250	ASH CN, JEFFERY, DIST. YELLOW
133	TREE	7-09-81	-10.7	-100.7	6250	ASH CN, JEFFERY, DIST. YELLOW
143	TREE	7-21-81	-13.3	-104.7	8500	ASH CN, LODGEPOLE, DIST. CLEAR
144	TREE	7-21-81	-13.7	-110.7	8000	ASH CN, JEFFERY, DIST. CLEAR
145	TREE	7-21-81	-13.8	-108.7	7500	ASH CN, JEFFERY, DIST. CLEAR
146	TREE	7-21-81	-17.0	-115.7	7000	ASH CN, JEFFERY, DIST. CLEAR
147	TREE	7-21-81	-15.1	-114.7	6250	ASH CN, JEFFERY, DIST. CLEAR
125	TREE	7-09-81	-14.9	-113.7	6400	CLEAR CN, YELLOW, DIST. CLEAR
181C	TREE	1-25-82	-12.6	-96.7	6400	CLEAR CN, JEFFERY, DIST. CLEAR
181Y	TREE	1-25-82	-9.3	-97.7	6400	CLEAR CN, JEFFERY, DIST. YELLOW
77	STREAM	5-13-81	-14.6	-111.7	7000	ASH CANYON CREEK
95	STREAM	5-28-81	-15.3	-108.7	5200	ASH CANYON CREEK
154	STREAM	8-27-81	-14.0	-103.7	8150	ASH CANYON CREEK
158	STREAM	8-27-81	-14.6	-103.7	7450	ASH CANYON CREEK
159	STREAM	8-27-81	-14.2	-103.7	7000	ASH CANYON CREEK
162	STREAM	8-27-81	-14.5	-100.5	6620	ASH CREEK, BELOW CNFL.
163	STREAM	8-27-81	-14.8	-107.7	6190	N. ASH CREEK, ABOVE CNFL.
164	STREAM	8-27-81	-14.4	-109.7	6190	ASH CREEK, ABOVE CNFL.
166	STREAM	8-27-81	-14.9	-102.7	5290	ASH CANYON CREEK
3	WELL	1-31-81	-15.24	-108.9		
4	WELL	1-31-81	-15.65	-101.4		
5	WELL	1-31-81	-15.75	-107.8		
9	WELL	1-31-81	-17.72	-118.9		
10	WELL	1-31-81	-17.93	-125.6		
12	WELL	3-14-81	-15.11	-95.6		
13	WELL	1-31-81	-13.88	-112.6		
48	WELL	4-14-81	-15.1	-111.7		
52	WELL	4-14-81	-14.9	-110.7		
172	WELL	9-22-81	-14.6,-15.5	-113,-120.		#
173	WELL	9-22-81	-13.2,-14.1	-111,-118.		#
174	WELL	9-22-81	-13.3,-14.2	-102,-109.		#
175	WELL	9-22-81	-14.4,-15.3	-104,-111.		#
176	WELL	10-15-81	-13.4,-14.3	-111,-118.		#
177	WELL	10-15-81	-12.9,-13.8	-108,-115.		#
178	WELL	10-15-81	-15.0,-15.9	-111,-118.		#
179	WELL	10-15-81	-14.4,-15.3	-114,-121.		#
180	WELL	10-15-81	-13.8,-14.7	-104.5,-111.5		#
1000	WELL		-15.4,-16.3	-123,-130.		TREXLER, CW 9, #
1001	WELL		-16.2,-17.1	-130,-137.		TREXLER, CW 10, #
1002	WELL		-14.9,-15.8	-127.5,-134.5		TREXLER, CW 11, #
1003	RIVER		-14.0,-14.9	-121,-128.		TREXLER, CW 12, #
DR-1	STREAM		-16.85	-117.9		SCOTLAND LAB
DR-1	STREAM		-15.7	-113.7		UNIVERSITY OF ARIZONA LAB
DR-1A	SNOW		-22.03	-154.0		SCOTLAND LAB
DR-1A	SNOW		-20.5	-148.7		UNIVERSITY OF ARIZONA LAB
DR-1B	SNOW		-20.46	-155.7		SCOTLAND LAB
DR-1B	SNOW		-20.5	-145.7		UNIVERSITY OF ARIZONA LAB

SAMPLES ANALYZED IN SCOTLAND LAB: 53-75, 3-52, DR-1A, 1B.

SAMPLES ANALYZED IN TUCSON LAB: 95-166, 172-1003, DR-1A, 1B.

# SECOND NUMBER ON SAMPLES ANALYZED IN TUCSON LAB ARE CORRECTED FOR DIFFERENCE BETWEEN LABS (SEE TEXT).

## APPENDIX 3 TRITIUM ANALYSES

NUMBER	TYPE	DATE	TRITIUM (TU)	INTERVAL SCREENED (FT) (DTL)	DESCRIPTION
1	WELL	1-31-81	50	42-45-70-114	
4	WELL	1-31-81	50	310-350	
5	WELL	1-31-81	31.1±14	159-180	
8	WELL	1-31-81	50	105-115, 170-175, 210-215	
10	WELL	1-31-81	50	86-116	
11	WELL	1-31-81	50	180-200	
112	WELL	4-14-81	50	150-185, 200-241	
12	WELL	3-14-81	50	5-31	
13	WELL	1-31-81	50	55-76, 106-124, 167-224	
15	WELL	3-14-81	50		
16	WELL	3-14-81	50	272-292	
17	WELL	3-13-81	10	102-125	
43	WELL	4-14-81	10	336- -589 (MULTIPLE SCREENED INTERVALS)	
51	WELL	4-14-81	27.1±13.	295- -494	
52	WELL	4-14-81	10	270- -458	
174	WELL	9-22-81	26.2±7.3	170-190	
175	WELL	9-22-81	6.0±7.7	100-120	
179	WELL	9-22-81	<5.		
17	WELL	7-09-81	5.	80-100	
180	WELL	10-15-81	46.0±9.1	30-40, 70-80	
177	WELL	10-15-81	30.1±5.		
178	WELL	10-15-81	5.		FRISON HOT SPRINGS
20-22	SNOW	3-14-81	34.1±5.		SNOW VALLEY PEAK, 9214'
11-42	SNOW	3-28-81	37.1±5.		MCCLELLAND PEAK, 7330'
59-60	SNOW	4-14-81	21.1±4.		ASH CANYON, 8250'
65-66	SNOW	4-14-81	21.1±4.		ASH CANYON, 7250'
181	STREAM	8-27-81	32.1±6.8		ASH CANYON CREEK, 6620'
185	STREAM	8-27-81	27.4±8.7		ASH CANYON CREEK, 5290'
181	TREE	1-25-82	33.0±10.4		WATER DISTILLED FROM JEFFERY PINE, CLEAR CREEK CANYON, 6400'

APPENDIX 4 ASH CANYON CREEK - STABLE ISOTOPE ANALYSIS  
FLOW (CFS)

YEAR	SEPT	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG
1976-77	1.84	1.75	1.73	1.77	1.72	1.91	2.51	2.25	1.60	1.10	0.87	1.06
1977-78	1.34	1.52	1.85	2.07	1.89	2.67	3.10	3.94	3.51	2.40	2.02	1.83
1978-79	1.79	1.98	2.04	2.60	2.22	2.91	3.09	4.37	2.83	2.30	2.07	1.62
1979-80	1.73	1.83	1.95	3.86	2.77	2.39	4.01	6.11	5.94	3.83	2.61	2.40
AVERAGE	1.68	1.77	1.89	2.58	2.15	2.47	3.18	4.17	3.47	2.38	1.89	1.73

E=29.36

## ESTIMATED ISOTOPIC COMPOSITION (%)

YEAR	SEPT	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG
80-81δD/H-103.	-107.	-111.	-114.	-114.	-116.	-115.	-114.	-113.	-111.	-100.	-96.	-98.
80-81δ <sup>18</sup> / <sub>16</sub>	15.4	-16.2	-16.8	-17.1	-17.3	-17.1	-16.6	-16.2	-15.4	-13.9	-13.3	-14.0
AVEδD/H	-106.	-112.	-117.	-124.	-128.	-129.	-127.	-125.	-118.	-105.	-97.	-98.
AVEδ <sup>18</sup> / <sub>16</sub>	-15.4	-16.5	-17.5	-18.3	-18.8	-18.5	-18.0	-17.5	-16.6	-14.9	-13.6	-14.0

## APPENDIX 5 PRECIPITATION DATA FROM THE CARSON RANGE

IN EAGLE VALLEY	ELEV (ft)	Precip (in)	Year
Carson City Airport	4700	11.0	1875-1965
Clear Creek	5800	17.9	1971-1979
Spooner Summit	7146	25.0	1971-1979
ON THE LEEWARD SIDE OF MOUNT ROSE			
Jct 395 & 27	4590	10.9	1969-1979
Callahan Ranch	5700	19.1	1969-1979
Galena Creek	6400	26.1	1969-1979
Sky Tavern	7620	35.6	1969-1979
Mt. Rose Resort	8280	43.9	1969-1979
Tahoe Meadows	8540	49.6	1969-1979
Summit-Mt. Rose Road	8900	60.0	1969-1979

(Klieforth, 1981)

## APPENDIX 6 DERIVATION OF THE INFILTRATION EQUATION

$$\frac{R}{R_o} = \theta^{\frac{1}{\alpha}} - 1$$

Rayleigh Distillation Formula  
for evaporation

$$\frac{R_i}{R_s} = \theta^{\frac{1}{\alpha}} - 1$$

$R_o = R_s$  = original isotopic ratio of  
precipitation or snow

$R = R_i$  = isotopic ratio of phase  
remaining (infiltrated water)

$\theta$  = fraction of water remaining  
from precipitation to infiltration

$\alpha$  = fractionation factor

substituting  $\delta\% = (R-1)1000$ , or  $\frac{\delta\% + 1000}{1000} = R$

$$(1) \frac{\delta_i\% + 1000}{\delta_s\% + 1000} = \theta^{\frac{1}{\alpha}} - 1$$

This equation is only valid for  
evaporation from a single phase  
of water, either evaporation from  
ice or evaporation from water.

combining this with the mass balance equation:

in = out

$$R_s Q_s = R_i Q_i + R_v Q_v$$

$R$  = isotopic ratio

$Q$  = volume:

$s$  = snow or precipitation

$i$  = infiltrated water

$v$  = vapor

(assumes no surface runoff)

$$1 = \frac{R_i}{R_s} \frac{Q_i}{Q_s} + \frac{R_v}{R_s} \frac{Q_v}{Q_s}$$

$$1 = \frac{R_i}{R_s} \theta + \frac{1}{\alpha} (1 - \theta)$$

$\theta = \frac{Q_i}{Q_s}$  = fraction of water remaining

$1 - \theta = \frac{Q_v}{Q_s}$  = fraction of water escaping  
as vapor

$$\frac{1}{\alpha} = \frac{1 - \theta \frac{R_i}{R_s}}{1 - \theta}$$

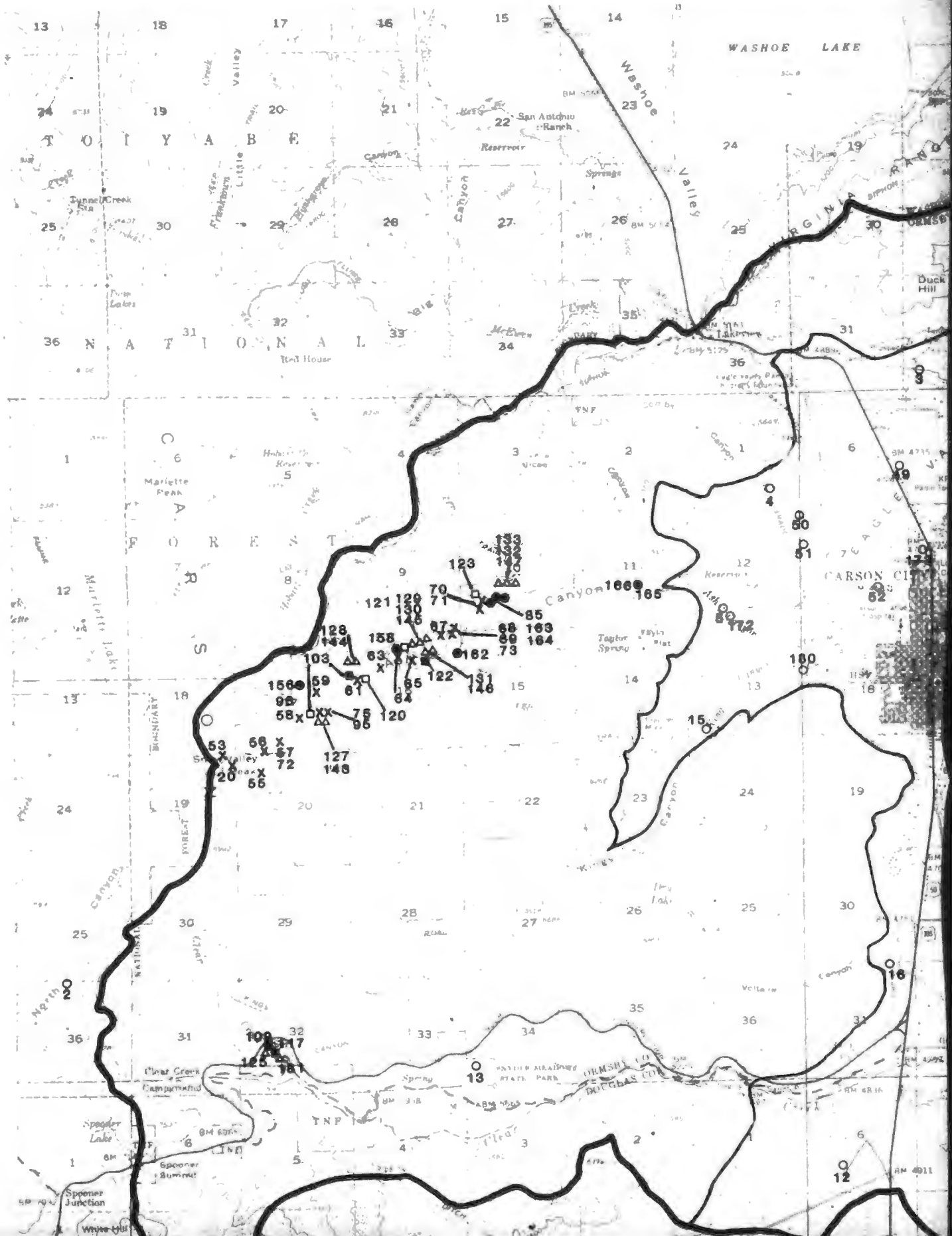
$\alpha = \frac{R_s}{R_v}$  = ratio of vapor pressures of  
light to heavy isotopes

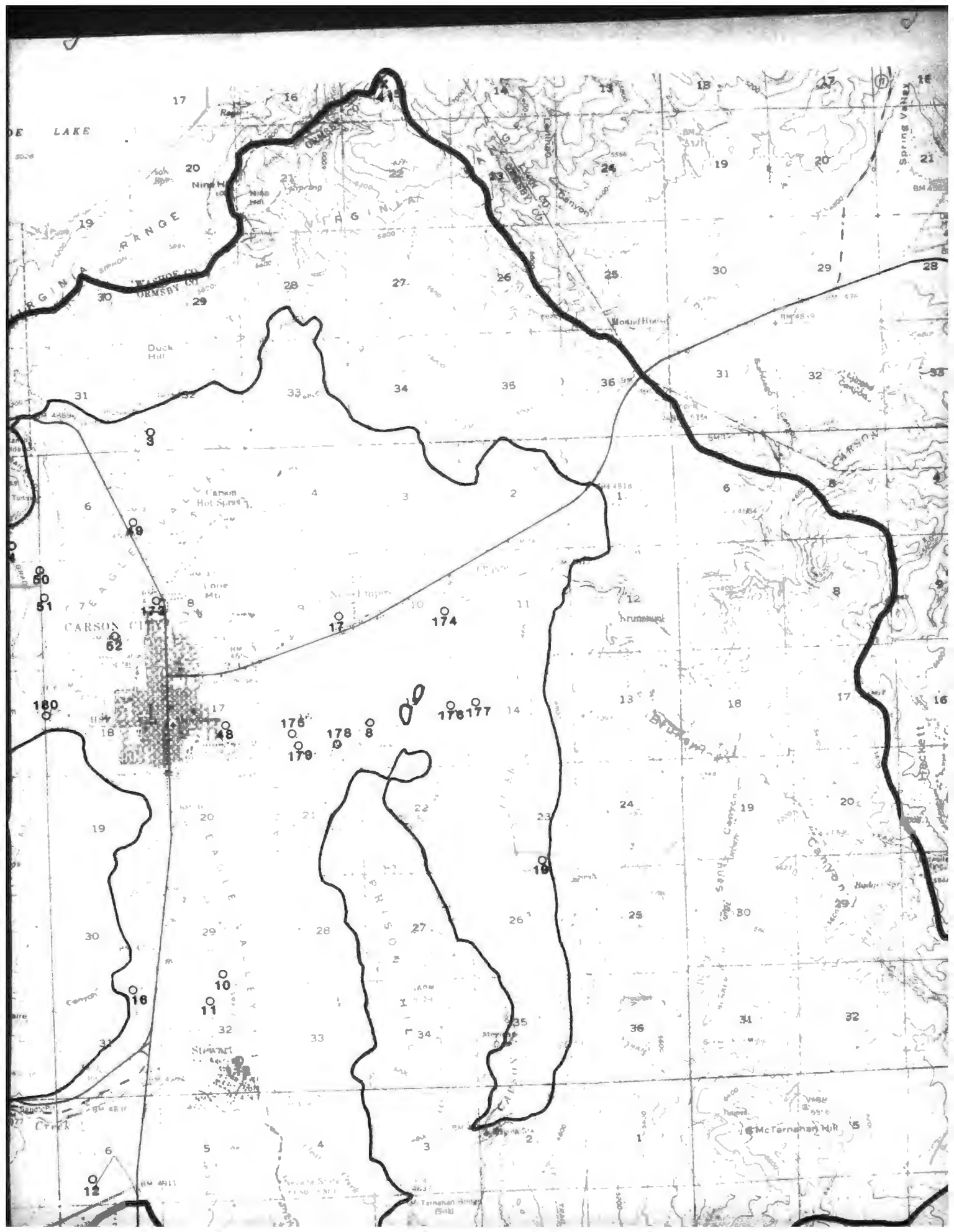
$$(2) \frac{1}{\alpha} = \frac{1 - \theta \left( \frac{\delta_i\% + 1000}{\delta_s\% + 1000} \right)}{1 - \theta} \quad \text{by substituting in } R = \frac{\delta\% + 1000}{1000}$$

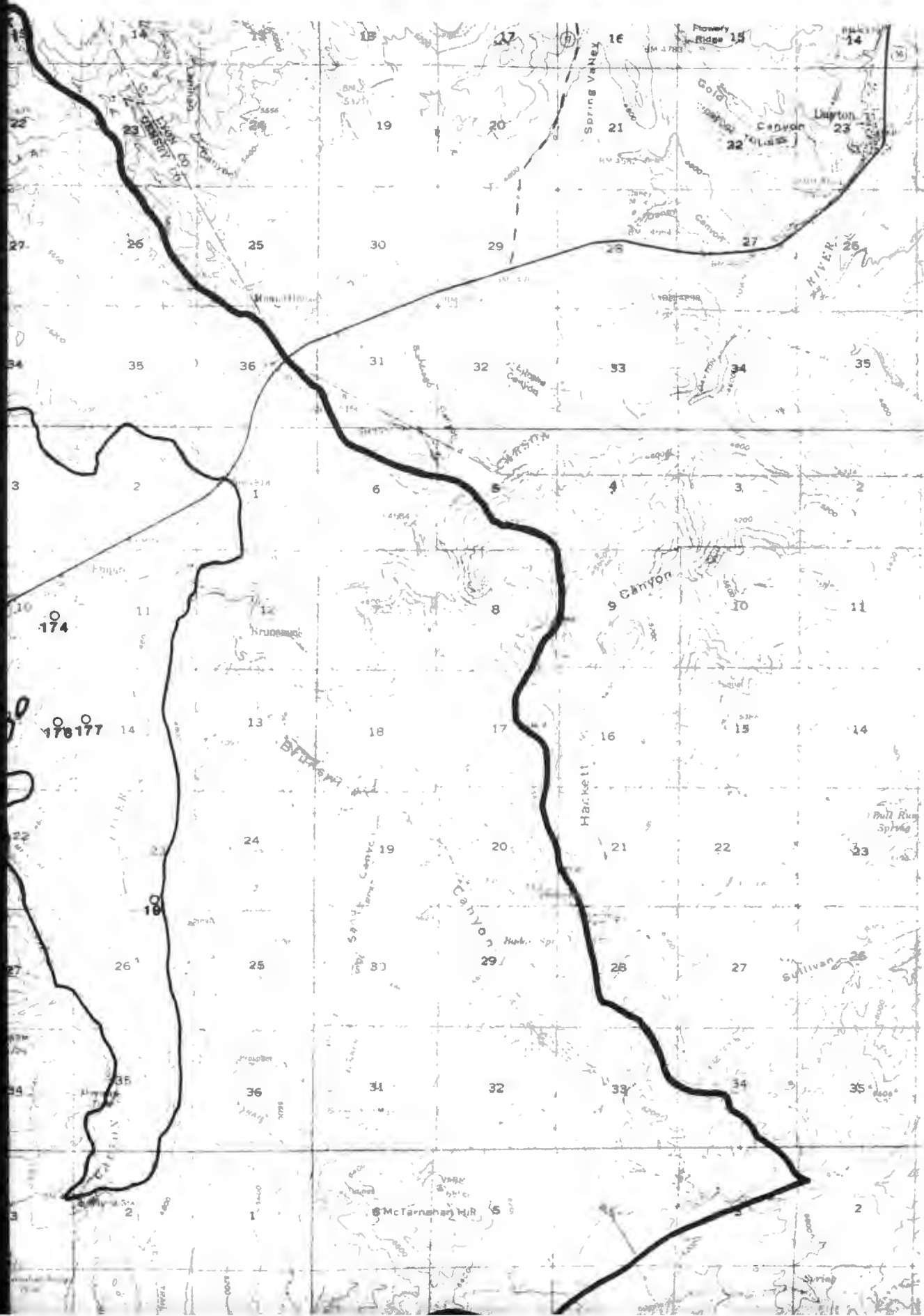
substituting equation(2) into equation (1):

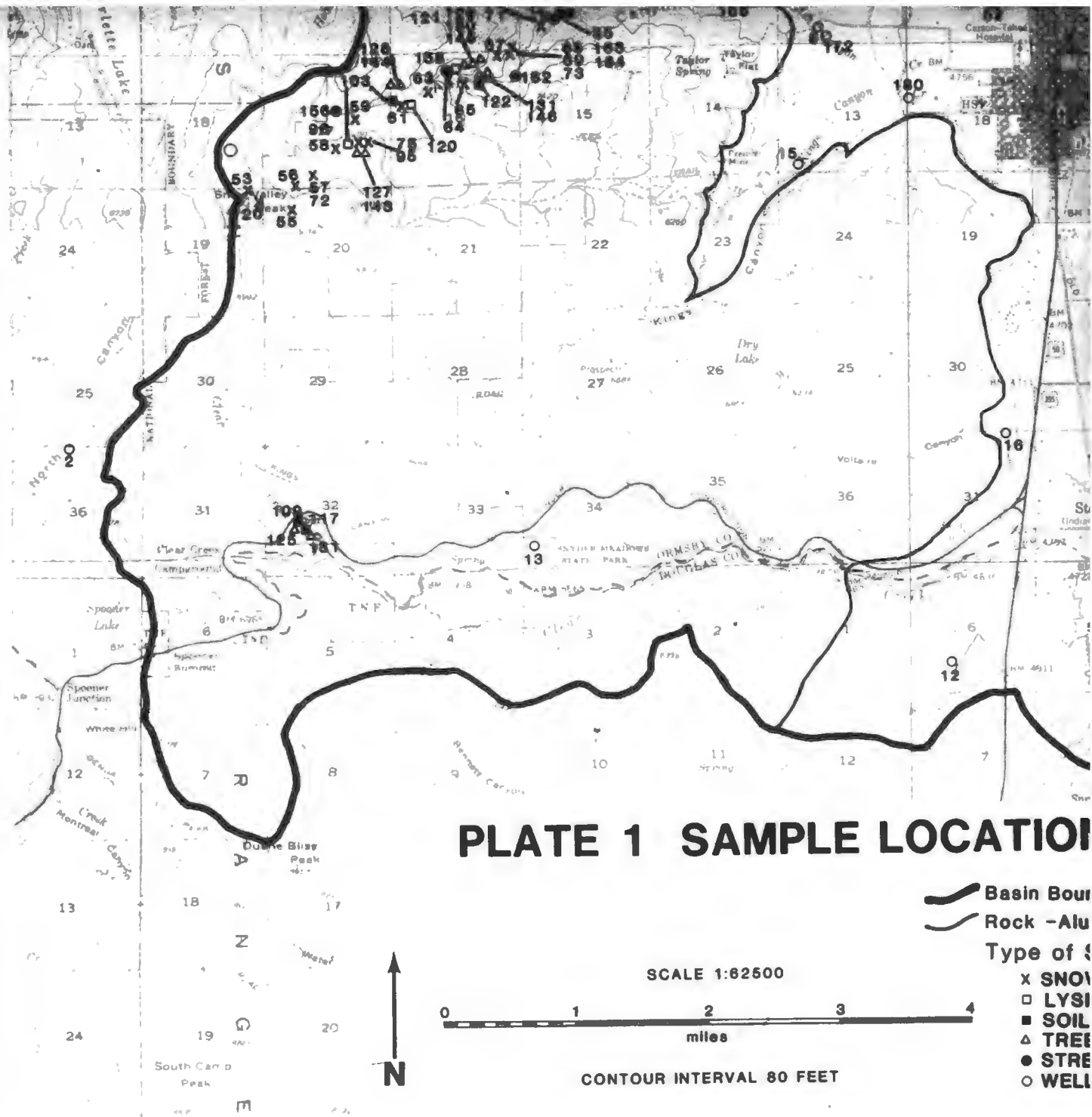
$$(3) \frac{\delta_i\% + 1000}{\delta_s\% + 1000} = \theta \left[ \frac{1 - \theta \left( \frac{\delta_i\% + 1000}{\delta_s\% + 1000} \right)}{1 - \theta} - 1 \right]$$



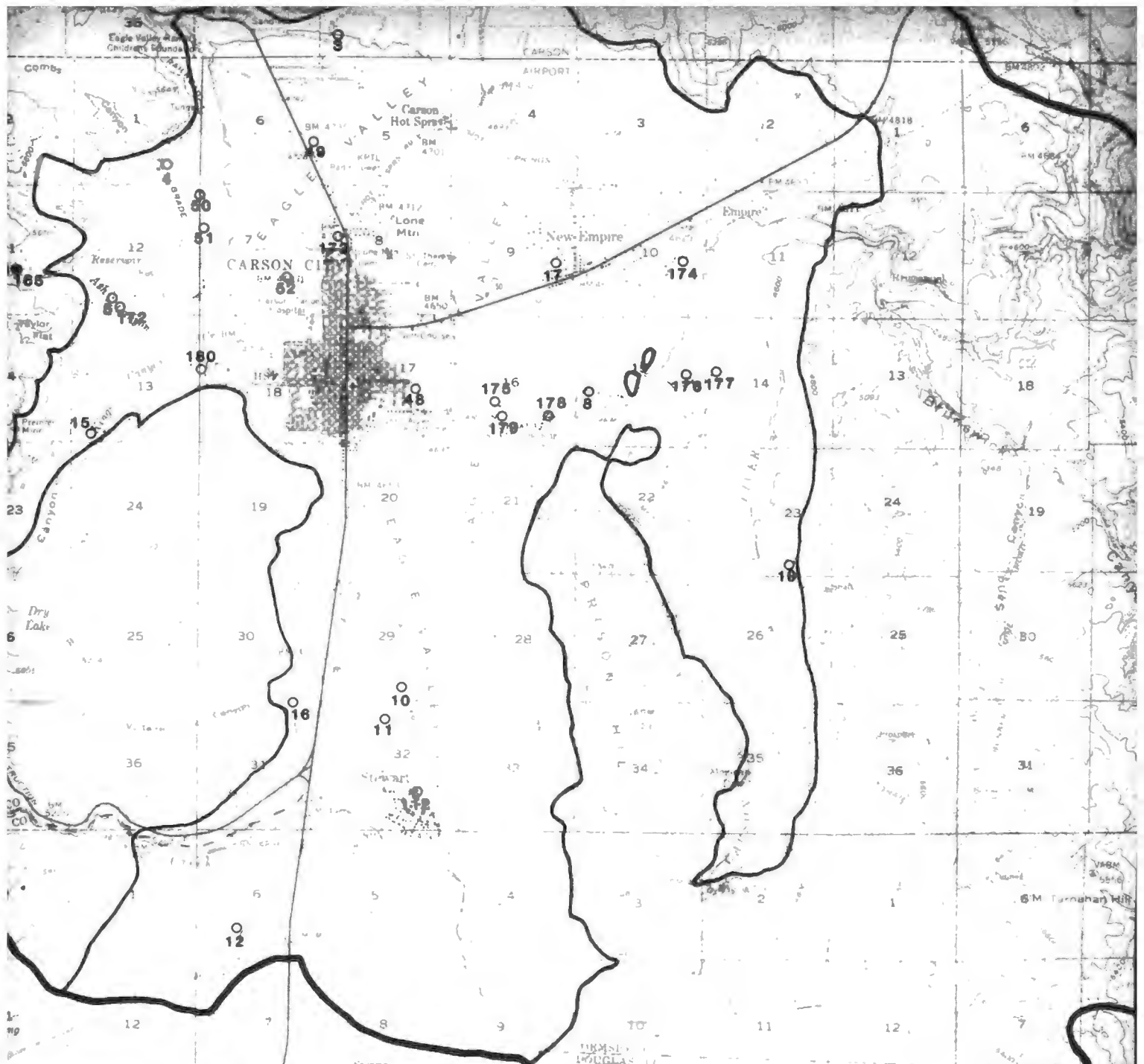








# PLATE 1 SAMPLE LOCATION



# SAMPLE LOCATION

-  Basin Boundary
-  Rock - Aluvium Contact

## Type of Sample

- x SNOW
- LYSIMETER
- SOIL
- ▲ TREE
- STREAM
- WELL

Scale: 1:62500



VAL 80 FEET

