

ISOTOPE HYDROLOGY OF SOUTHERN HONEY LAKE VALLEY  
NEVADA AND CALIFORNIA

prepared for

Washoe County  
Department of Public Works  
Utility Division  
P. O. Box 11130  
Reno, Nevada 89520

Burkhard Bohm, Ph.D.

Hydrogeologist

August 30, 1990

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# DISCLAIMER

The contents of this report are subject to review and are intended for discussion only within Washoe County Utility Division, and it's consultants working on the Fish Springs Ranch Project. Findings and statements made in this draft report should be used only within the context implied within this text.



## 1.0 EXECUTIVE SUMMARY

The hydrology of southeastern Honey Lake Valley was studied in detail, using environmental isotope and geochemical data, from at least 100 sampling sites, with the main emphasis on analyzing the Fish Springs aquifer system.

### Results:

1. Isotope and geochemical data suggest that the ground water flow systems in southern Honey Lake Valley discharge into an area of evapotranspiration northwest of the playa.
2. The recharge area for the Fish Springs aquifer system covers at least the northern Virginia Mountains. With the available data, and based on the present interpretation it is not absolutely clear how far south the recharge area extends, though isotope data suggest that it may reach as far south as Tule Peak, provided the necessary geological structures are present. Under natural conditions no recharge occurs into the wellfield from the northeastern Virginia Mountains.
3. The central and eastern Fish Springs Aquifer System is very homogeneous in composition, suggesting good communication between the highlands in the southeast and the central and eastern wellfield, including the area from the Hodges to the Ferrel wells.
4. The far western wellfield (Wilson, Nork and Ford wells), produces water from an entirely different flow system, with different composition and temperature, the source of which is associated with the Warm Springs Fault Zone. These wells are probably affected by geothermal waters, and have higher dissolved mineral concentrations.
5. Most of the water pumped was been recharged at least 35 years ago. Waters in the central and eastern wellfield are at the most 10,000 years old. Based on ground water temperatures, depth of ground water circulation is between 1000 and 1500 feet deep.
6. The high TDS waters north and northwest of the playa are possibly derived from water flowing through the far western wellfield, and to a lesser extent from the wellfield to the east. This requires, at least in part, an alternate discharge area for the central and eastern wellfield.
7. The interpretation of isotope and hydrochemical data suggests that no ground water flows from Honey Lake Valley through Astor Pass into Pyramid Lake Valley. Flow from Honey Lake Valley through northern Dry Valley into Smoke Creek Desert remains unclarified with the data at hand.

## 2.0 INTRODUCTION

In July 1989 Plumas Geo-Hydrology was requested by Washoe County Utility Division (WCUD) to develop a proposal for analyzing the ground water hydrology of southeastern Honey Lake Valley by means of environmental isotopes. A field sampling program was submitted to WCUD in August 1989, proposing a limited number of sampling sites, to obtain a preliminary picture of the regional hydrology of the basin. To augment the isotope data, the program also proposed collection of hydrochemical data. The samples were collected by WCUD personnel in September and October 1989, and the first lab results were available by November. Though the initial proposal was rather limited in extent, WCUD ended up collecting almost three times as many samples as originally proposed.

The objectives of this study are:

1. Analysis of ground water recharge areas, their location and areal extent.
2. Analysis of potential ground water flow paths.
3. Analysis of ground water discharge areas.
4. Analysis of subsurface residence times.

In December 1989 Plumas Geo-Hydrology was retained by Western Water Development Corporation, Inc. (WWDC) in Reno to analyze the water quality and model the effects of pumping on ground water quality. In the context of this study a great deal of new data needed to be collected, particularly to outline the extent of high TDS ground waters in the area. As both studies proceeded, several unforeseen problems were realized during the ongoing discussion between the County, WWDC and their consultants, including William E. Nork, Inc. and JBR Consulting Group. The points of concern included:

1. The question of interbasin flow from Honey Lake Valley to Smoke Creek Desert and Pyramid Lake Valley.
2. The need to determine the ultimate ground water discharge area of the waters flowing through the wellfield.

The newly available data could not be ignored, and the amount of information available increased the size of the study, though the original objectives remained the same. Therefore, since it's original inception, the size of the study has increased significantly, both in scope and in the amount of data required to be analyzed. As it turned out, both studies complement each other, allowing use of a common data base for both projects, and preventing duplicate efforts.



## 2.2 METHODS OF INVESTIGATION

For the benefit of the reader who is not well versed in hydrology, the basic methodology applied in this study is outlined briefly. Further details are given in Appendix B.

Ground water is recharged in the mountains and is discharged at lower elevations. Recharge areas have the highest elevation ground water tables, discharge areas the lowest elevation water tables. A continuous flow of ground water from a particular section of the highlands to a discharge area constitutes a ground water flow system. The purpose of wellfield development is to intercept an optimal amount of ground water before it flows into a ground water sink.

Since the basin is surrounded by mountains, depending on where the ground water flow system is to be intercepted by means of wells, one or several areas could serve as recharge areas for the area of interest.

To determine the source of water pumped in the proposed well field, the stable isotope signatures of all potential recharge waters are determined. Since it is difficult and time consuming to sample soil waters after infiltration in the mountains, samples from high elevation springs, constituting local flow systems (i.e. localized recharge) are sampled. The springs are assumed to represent the isotope signature of the recharge waters in that part of the basin. These samples are then compared with those collected from the area of wellfield development in the Valley.

An additional tool used is the geochemical evolution of ground waters. Geochemical trends provide hints about flow systems (verifying the isotope methods) and provide qualitative information about how "easy" ground water migrates through the system. For the purpose of this report the terms "geochemistry", and "hydrochemistry" are used for anion and cation geochemistry, excluding environmental isotopes. The term "isotopes" refers to environmental isotope data.

Once the primary source areas have been determined the amount of recharge can be determined by using the areal extent of recharge and precipitation data.

The geology of the Nevada side of Honey Lake Valley and the adjacent highlands was described by Bonham and Papke (1969). The hydrology of southeastern Honey Lake Valley has been outlined very briefly in a reconnaissance study by Rush and Glancy (1967). The remaining basin hydrology was analyzed by California Department of Water Resources (Ford et al., 1963). Geothermal resources in Honey Lake Valley were addressed by US Bureau of Reclamation (1976). The regional aspects of geothermal aquifers in Honey Lake Valley were discussed by Juncal and Bohm (1987). Therefore a fair amount of information is available from the basin, yet most of the work is rather preliminary and little



detail is provided about the Fish Springs Aquifer System.

The most important geographical features referred to in this report are shown in Plate 1. In southeastern Honey Lake Basin the potential recharge areas are located in the Skedaddle Mountains to the east, the Virginia Mountains in the south and east, the Dogskin Mountains in the south and the Fort Sage Mountains in the west. The Diamond Range in the west has the highest precipitation (Ford et al., 1963) and is probably the most important recharge area for the entire basin. However, the Diamond Range is relatively far away and separated from the well field by a flow system entering from Long Valley (Ford et al. 1963) and the apparent ground water sink for this part of the basin, an area that is characterized by high TDS ground waters, presumably by evaporation or transpiration (or both).

For the purpose of this study the Virginia Mountains are divided into two distinct areas (see Plate 1). The central Virginia Mountains are referred to as that part of the Virginia Mountains south of the mouth of Cottonwood Canyon, i.e. south of the Hodges Well. The northeastern Virginia Mountains are referred to as that part east of the Hodges Well, i.e. the area north of Telephone Pole Canyon.

### 2.3 DATA USED FOR THIS STUDY

A total of 43 samples were collected by WCUD for the purpose of this study. An additional eleven samples were collected by Plumas Geo-Hydrology (in a co-operative effort with William E. Nork, Inc.) for the purpose of analyzing the water quality in the area northwest of the playa. Wells and springs were sampled for deuterium and oxygen-18 (stable isotopes) and major ions (hydrogeochemistry). Whenever possible, pH, electric conductivity and temperature were measured in the field and the spring discharge estimated. In combination with a large number of literature data, these data provided valuable information about potential recharge areas, flow systems and ground water sinks.

Also 15 springs and wells were sampled for tritium to permit conclusions about subsurface residence times.

Though the area of interest is of limited areal extent, the main objective was to put the wellfield into its basin-wide hydrologic context. Therefore none of the available literature data could be ignored and had to be integrated into the data base, to assure an optimal amount of information and interpretation.

Doug Maurer of the U.S.G.S. in Carson City, Nevada provided 17 sets of deuterium/oxygen-18. The California Department of Water Resources (DWR) in Red Bluff, California provided 34 chemical data sets for the Doyle and Herlong area. All data used in this study are listed in Appendix C.

All sample locations are shown in Plate 1. The entire data base

covers a limited area in the Fort Sage Mountains, an area in the north-central and northeastern Virginia Mountains (between the Warm Springs Fault Zone, Fish Springs Ranch and Tule Peak), two springs in the Skeddaddle Mountains, and limited number of wells in the Astor Pass and Bonham Ranch area (Plate 1). The waters discharged in the valley proper are represented by about fifty irrigation and domestic wells and more than twelve test wells (drilled by WCUD) from the Herlong area in the west to the Neversweat Hills in the east.

The error margins for the stable isotope data are  $\pm 0.3$  and  $\pm 1.0$  per mil for oxygen-18 and deuterium, respectively. Milliequivalent balances for the major ions are well within  $\pm 10\%$  for more than 90% of the data used.

In order to draw contour maps, an arbitrary co-ordinate system was applied. The x-axis was drawn along the line between townships 25 and 26 and the y-axis along the California-Nevada state line. Units of measurement are in miles.

For easier identification of the large amount of data plotted on the diagrams produced for this study, each data point was assigned a two to four letter code. A capital letter was used to denote the general area (e.g. "V" for Virginia Mountains). Lower case letters were used as abbreviations of the sampling point name. The codes were assigned not only on the basis of geographical location but also on data interpretation results as to which flow system or geochemical group they belong to.

The following letters were assigned to identify geographical locations and/or ground water types in southeastern Honey Lake valley:

AP	-	Astor Pass (numbered), 2 wells
B	-	Bonham Ranch area, Smoke Creek Desert, 3 wells
DL	-	Duck Lake area (numbered), only 3 wells
Fl	-	Flanigan area, 2 wells
FS	-	Fort Sage Mountains ("f" stands for alluvial fan, or NE mountain front), only 2 springs
G	-	Geothermal waters, <u>A</u> medee and <u>L</u> ong Valley H.S.
H	-	Herlong area, 19 wells:
		b - Herlong <u>B</u> ase
		d - <u>D</u> oyle area
		j - Herlong <u>J</u> unction
		s - immediately <u>S</u> outh of Herlong Base
NS	-	Neversweat Well
P	-	Phreatophyte area, NW of playa
Smf	-	Skeddaddle Mountains, SW mountain front, 7 wells
S	-	Skeddaddle Mountains, 2 springs
St	-	Stacey area, 10 wells
T	-	TMP, 12 samples from a total of 8 wells and Fish Spring
V	-	Virginia Mountains, 18 samples from 15 springs, 2 creeks and 1 well
W	-	Warm Springs and Fort Sage Fault Zones, 6 springs,

distinction based on isotopes and geochemistry.

When studying the diagrams in this report, it is advisable to refer to the table of sampling point codes in the beginning of Appendix C. Other codes used in the diagrams are explained in the respective legends.

### 3.0. RECHARGE AREAS

#### 3.1. GENERAL OBSERVATIONS

Before looking in detail at the various hydrologic components of interest it will be useful to look at the overall basin-wide isotopic and geochemical trends.

Figure 1 is a contour map of deuterium levels in wells and springs sampled in southern Honey Lake Valley. Though the contour map gives the impression of continual trends/changes from one area to another, certain provenances of deuterium are evident. Most obvious is the area north and northwest of the Wilson Well, an area with deuterium enriched ground waters, suggesting the effect of evaporation in a ground water sink. (The extension of the enriched deuterium zone around Duck Lake is somewhat misleading, since one of the three Duck Lake wells is probably influenced by evaporating surface water).

Due to the small scale the area cannot be outlined exactly. However, it appears as if the ground water flow systems of the entire southern Honey Lake Basin flow into the area north and northwest of the Wilson Well, where deuterium is enriched by evaporation. Also noticeable are the lower deuterium levels (less than -114 per mil) in the Wilson Well area and the adjoining Warm Springs Fault Zone, extending to the southeast into Warm Springs Valley.

The concentration of total dissolved solids (TDS) is a qualitative indicator of geochemical evolution, increasing with distance along a ground water flow path. Low concentrations indicate little chemical interaction with aquifer material (recharge areas), whereas higher concentrations reflect extended subsurface residence time, long distance of flow or evaporation. The map in figure 2 shows high concentrations in the area northwest of the playa in an elongated strip from about 3 miles northwest of the Wilson Well to Flanigan. Low levels of TDS are found in the Virginia/Fort Sage Mountains, the Herlong/Doyle area and the Skeddaddle Mountains.

Similar, though more explicit trends were discerned by applying geostatistical models to ion ratios (e.g.  $\text{Cl}/\text{HCO}_3$  and  $(\text{Na}+\text{K})/(\text{Ca}+\text{Mg})$ ). The overall trends seem to suggest ground water flow from the southeast and southwest into the area northwest of the playa. A trend from the high TDS area to the northeast seems to suggest ground water flow from or to the northeast (Smoke Creek Desert). However, this trend is somewhat misleading since

the isotope trends in figure 1 do not support this trend.

### 3.2. THE FISH SPRINGS AQUIFER SYSTEM

The term "Fish Springs Aquifer System" is used to address the hydrologic unit providing ground water to the existing irrigation wells in the Fish Springs Ranch area, i.e. the area from the Ford and Wilson Wells just to the east of the stateline to the Hodges Well south of the Neversweat Hills. In this subsection it will be shown that the Wilson, Ford and Nork wells produce water that is of a different isotopic and chemical composition (derived from a different source), than the remaining wellfield to the east. The hydrogeochemistry and isotope data from the Virginia Mountains and the wellfield are discussed together, due to a great deal of similarities between the two.

#### 3.2.1. Stable isotopes in the Fish Springs Aquifer System

Deuterium and oxygen-18 for all irrigation wells and all high elevation springs are plotted in figure 3, together with the world meteoric water line (WMWL). It is important to notice that these waters plot in two distinct groups, with respect to both deuterium and oxygen-18.

In the central and eastern wellfield, represented by the Hodges, Headquarters, Jarboe and Ferrel wells (JFH wells, for brevity) deuterium levels range between -114 and -112 per mil. The homogeneous composition over a large distance is remarkable, suggesting a single source area. With a few exceptions, these same deuterium levels are limited to springs in the central Virginia Mountains, discharging from a wide range of elevations between 4650 and 7800 feet. Exceptions are the high discharge rate springs at low elevations, i.e. Sheep and Willow Springs. Willow Springs (elev. 4600 ft) discharges 17 gpm (JBR Consultants, personal communication) and Sheep Springs (elev. 4840 ft) constitutes a significant discharge area including a group of springs.

In contrast, the far western section of the well field represented by the Wilson, Nork and Ford Wells (WNF wells, for brevity), have a distinctly different deuterium signature, ranging between -118 and -116 per mil deuterium. These deuterium levels are not observed in any of the springs located in the north-central Virginia Mountains, but are limited to an area associated with the Warm Springs Fault Zone (or Fort Sage Fault Zone) between the Virginia and the Fort Sage Mountains. These springs discharge between the elevations of 5000 and 5900 feet.

The test wells drilled in early 1990 by WCUD in Cottonwood Canyon (Tcwt) and the "First Fault Well" (Tflt1) plot in line with the surface water samples (Mullen, Cottonwood and Fish Springs Creeks) in figure 10. It appears as if these waters are part of a shallower flow system, recharged by surface waters, and flow



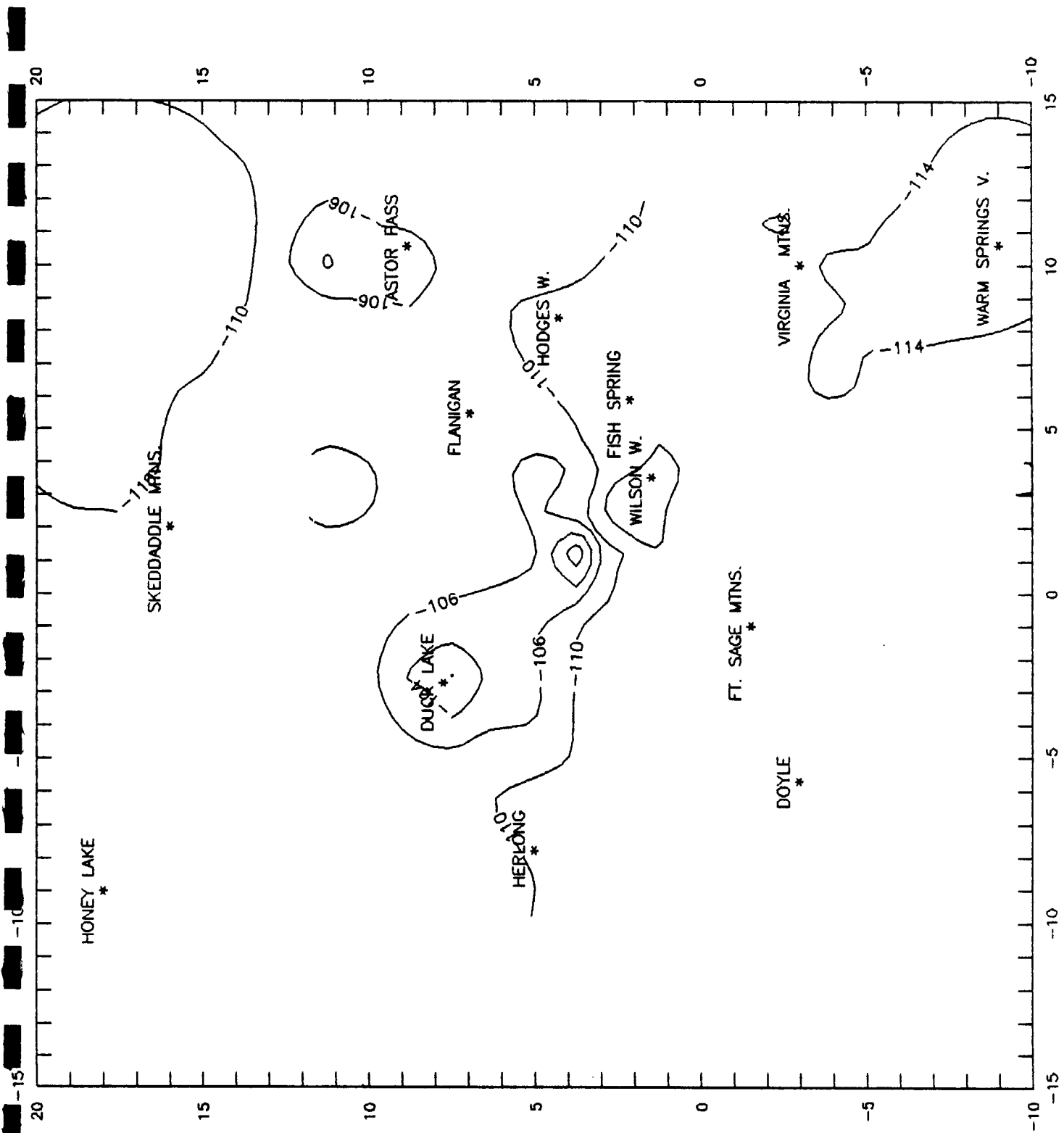


Figure 1: Map of deuterium levels in ground waters sampled in southern Honey Lake Valley.

system(s) different from those at the eastern wellfield. A similar situation may occur in the Ferrel Wells. The temperature of 27 C (the highest in southeastern Honey Lake Valley) and TDS of less than 200 ppm in the "First Fault Well" suggests that circulation can reach significant depths within relatively short subsurface residence times.

In summary, the isotope data suggest that the wellfield is recharged at least from the north-central Virginia Mountains. Similar deuterium levels in the wellfield and high mountain spring waters suggest that the recharge area may reach as far south as Tule Peak. However, that would by far exceed the topographical basin draining surface water into Honey Lake Valley. If the recharge area indeed reaches as far south as Tule Peak, structural arguments would have to be found to support this interpretation. Bonham and Papke (1969, p. 50) in their discussion of the structural geology of the Virginia Mountains discuss the presence of north to northwesterly trending faults related to the Walker Lane.

### 3.2.2 Hydrogeochemistry of the Fish Springs Aquifer System

Hydrogeochemical data corroborate the interpretation of the stable isotope data, and show some features that are useful for further analysis of regional and intrabasin ground water flow.

The logarithmic ratio of  $\text{Cl}/\text{HCO}_3$  is a useful indicator of ground water facies using the concept developed by Back (1960). For a detailed analysis the reader is referred to Appendix A. Four levels of  $\text{Cl}/\text{HCO}_3$  (facies A, B, C and D) are observed in this basin. Low ratios suggest little dissolution of aquifer material, higher ratios indicate increasing TDS levels and removal of  $\text{HCO}_3$  (bicarbonate) due to carbonate mineral precipitation.

The  $\log\text{-Cl}/\text{HCO}_3$  facies groups (figure 4) suggest that the aquifer formation from the Hodges to the Jarboe wells is chemically homogeneous (facies B), essentially the same as the spring waters in the central Virginia Mountains. In contrast the wells in the far west (Wilson, Nork and Ford) produce a different type water (facies C).

The milliequivalent percentages of the major anions and cations were plotted in figure 5 (which is essentially the same as the "diamond" in a Piper or trilinear diagram). The diagram shows trends of geochemical evolution typically expected for waters evolving from low TDS recharge waters to high TDS waters concentrated by evaporation. The evolution is from calcium-bicarbonate (low TDS) to sodium chloride waters (high TDS). In this scheme the waters from the western wellfield (as referred to in Plate 1) are different from those in the central and eastern wellfield.

There are several possibilities why the far western wellfield is chemically so different from the central and eastern wellfield:

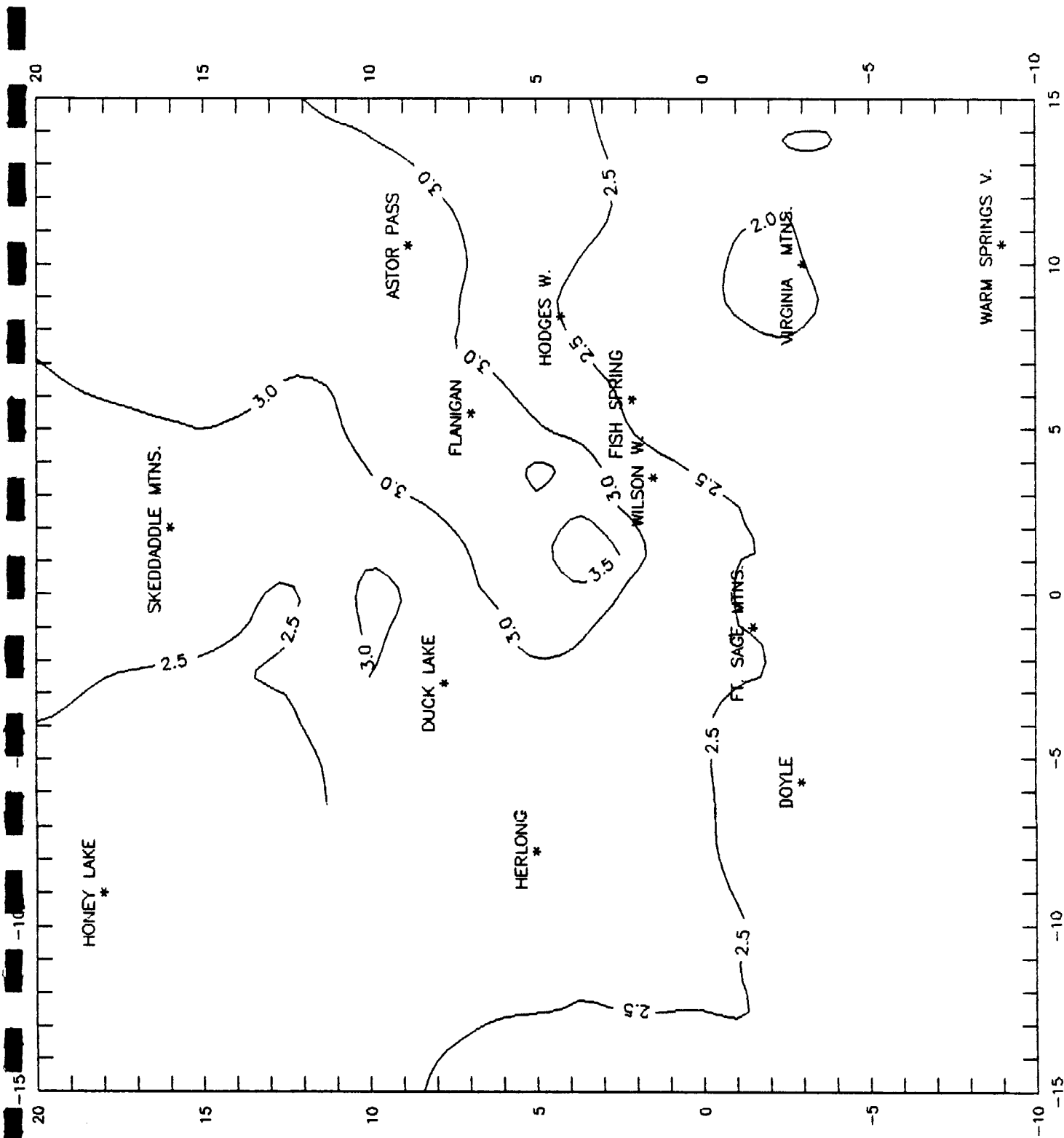


Figure 2: TDS concentrations in ground waters in southern Honey Lake Valley

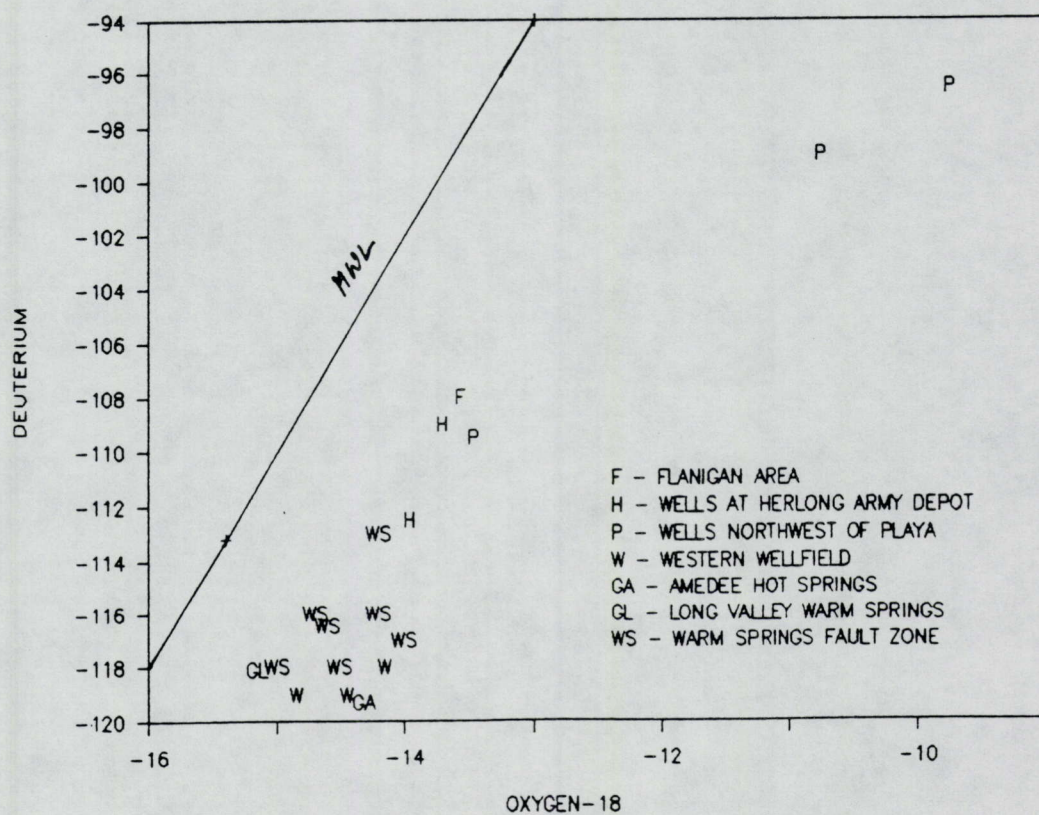
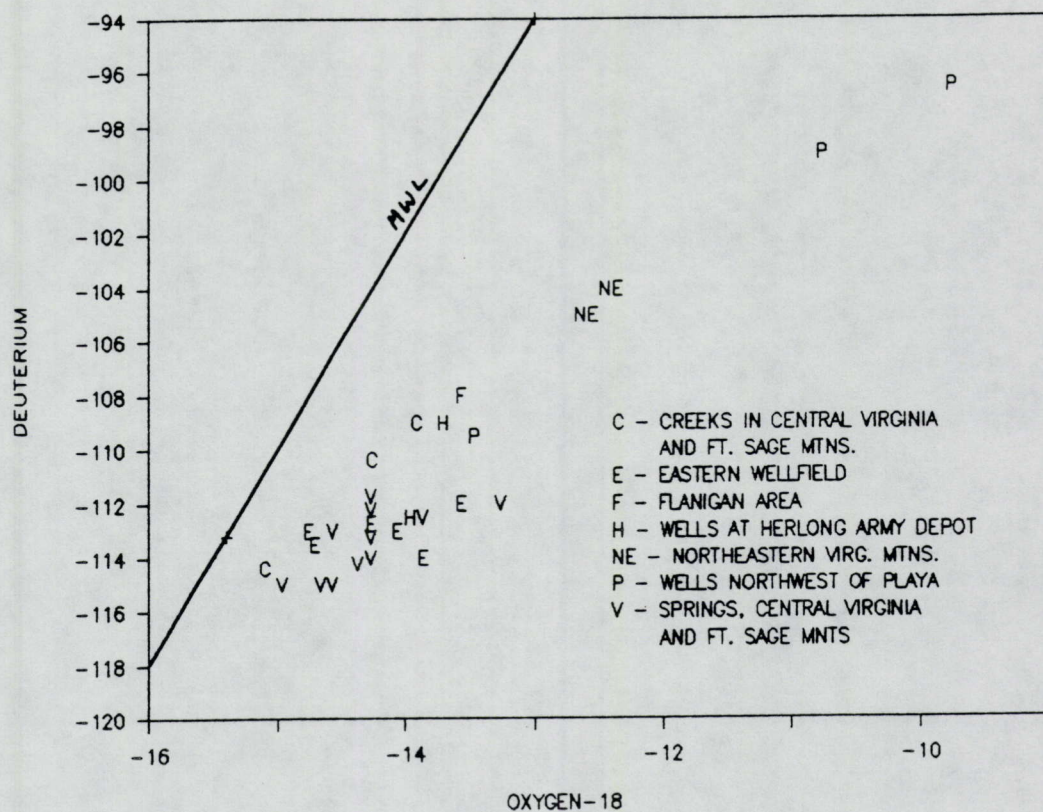


Figure 3: Deuterium and oxygen-18 in ground waters in southern Honey Lake Valley



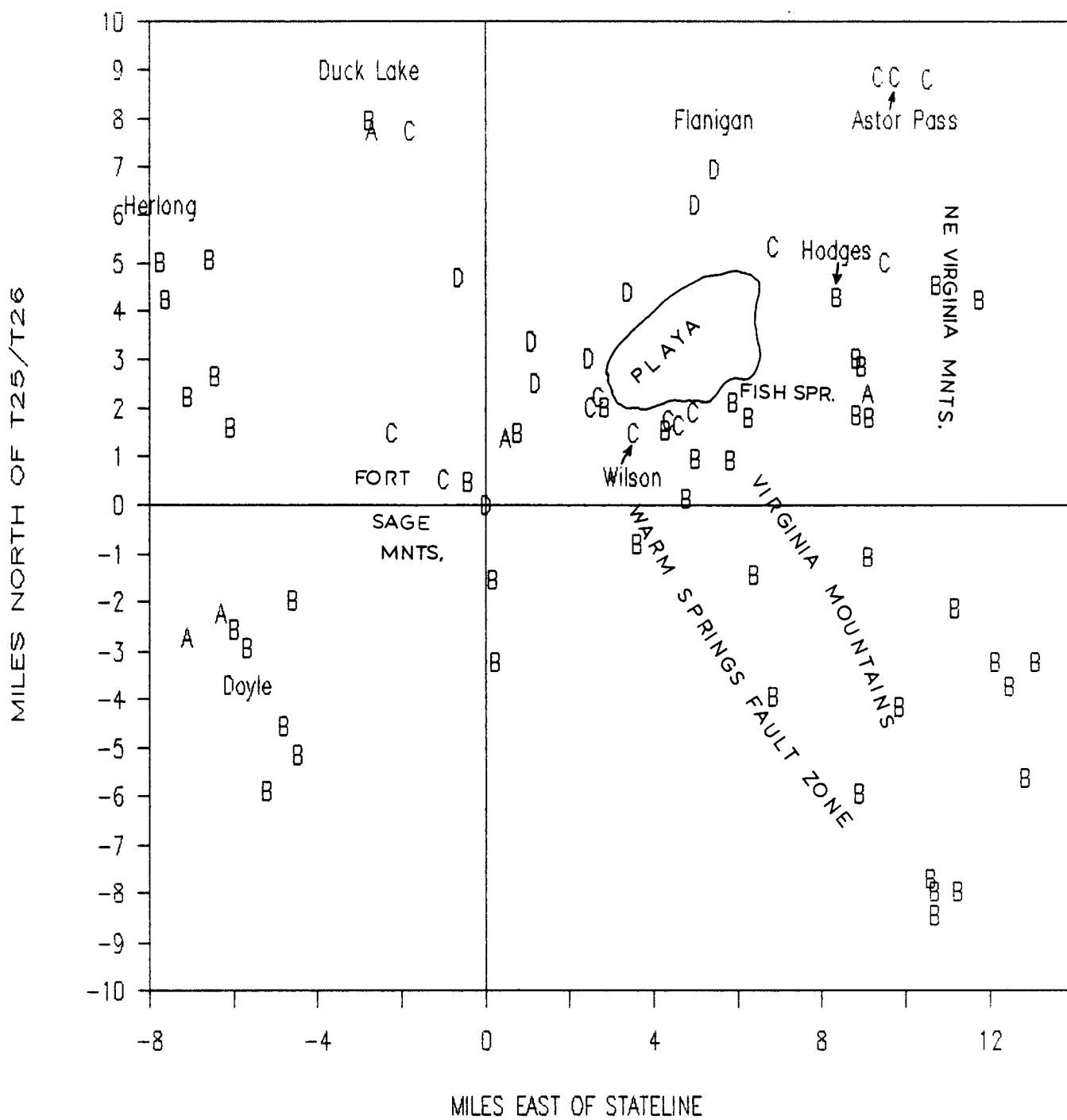


Figure 4: Hydrogeochemical facies of  $\text{Cl}/\text{HCO}_3$  S. Honey Lake Valley (Facies A, B, C and D, as determined in Appendix A.1).

1. As the stable isotope data suggest, the C waters pumped in the Wilson, Ford and Nork wells are part of a separate flow system.
2. The chemical differences may also be due to different aquifer materials or aquifer types. The Wilson, Ford and Nork wells all discharge from alluvial aquifers, in contrast to the eastern wells which produce almost entirely from bedrock aquifers (the Hodges well is an exception). Since transmissivity in the Wilson Well is substantially lower than in the bedrock wells, ground water migration is slower, allowing for more rock-water interaction and subsequently higher levels of TDS, etc. However, the effect of lake beds rich in organics and possibly evaporite deposits may also play a role. The latter aspects will be discussed in detail in a concurrent study prepared by PGH for WWDC.

Therefore, for waters produced from the wellfield, two distinct source areas are identified. The first one is associated with the central Virginia Mountains, providing water for the central and eastern wellfield. The second one is associated with the Warm Springs Fault Zone, providing water for the far western wellfield. The implications for regional flow systems will be discussed in subsection 3.2.

A few more valuable observations can be made by looking at the hydrochemical data from the high elevation springs in the Virginia Mountains and Fort Sage Mountains (denoted V, FS, W):

1. As expected for a recharge area, all springs in the Virginia Mountains and Fort Sage Mountains are calcium-bicarbonate waters, with TDS ranging from less than 50 to 150 ppm. The waters from the central and eastern wellfield are sodium-bicarbonate waters, with TDS ranging from 150 to 280 ppm. In geochemical terms, this is a logical consequence of evolution from the central Virginia Mountains to the eastern wellfield waters.
2. Chloride levels in the mountain springs range from 2 to 7 ppm. The chloride levels in the eastern wellfield are only slightly higher, ranging between 7 and 22 ppm. TDS in the springs range from less than 50 to 150 ppm. In the eastern wellfield TDS ranges from 150 to 280 ppm.
3. The spring waters from the Warm Springs Fault Zone are chemically similar to those discharged in the Virginia and Fort Sage Mountains, with the exception of slightly higher bicarbonate levels and TDS (150 to 250 ppm). This is in contrast to their differences in isotope composition, suggesting different sources but similar geochemical environments.

Based on the hydrochemical data it can be argued that the rocks making up the central Virginia Mountains are transmissive enough to allow relatively rapid ground water migration from the north-

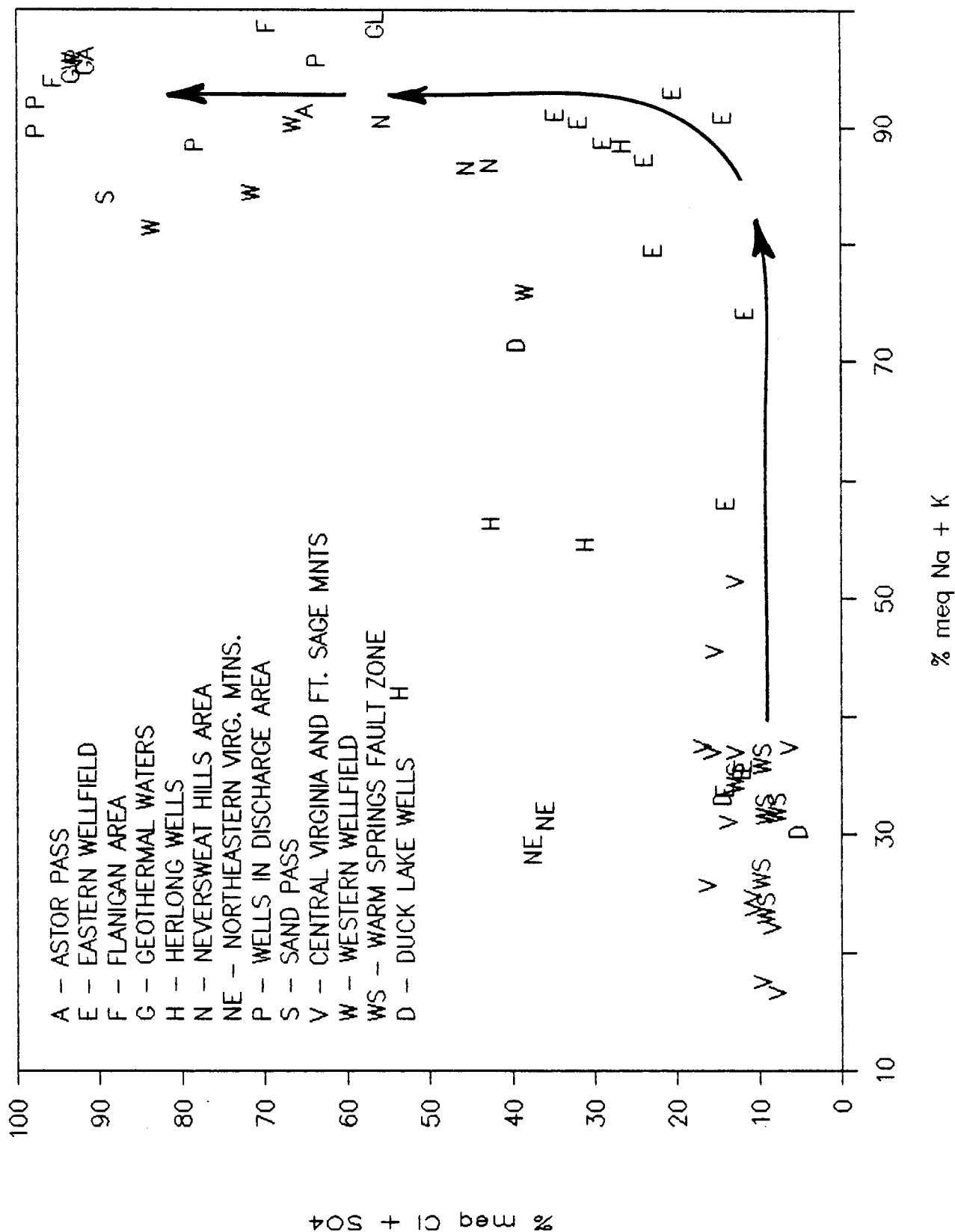


Figure 5: Anion and cation percentages in ground waters from southeastern Honey Lake Valley.

central Virginia Mountains into the eastern wellfield.

The spring waters associated with the Warm Springs Fault Zone are different from those in the central and eastern wellfield. They differ mostly in their isotopic signatures and less in their chemical composition. These waters stand out as an anomaly in the hydrochemistry of southeastern Honey Lake Valley. Ground waters pumped in the shallow domestic wells west of the Wilson Well are low TDS waters with isotope signatures and chemistry similar to those observed in the central Virginia and Fort Sage Mountains (e.g. FS2tp).

The effects of the Warm Springs Fault Zone are noticeable not only in the wells on the valley floor (below 4000 ft), but also in mountain springs at elevations up to 6000 ft (e.g. Rat Spring, Wrs, in the Fort Sage Mtns.). This supports the presumption of a large regional geothermal flow system at great depth (Juncal and Bohm, 1987), manifested locally where deep reaching faults allow upward migration, be it in the mountains or on the valley floor. This system is apparently diluted by the shallower flow system recharged in the mountains surrounding the valley proper.

In contrast, the isotope data and geochemical compositions of the Neversweat Well and the high elevation springs in the northeastern Virginia Mountains (Vua and Vuj, with TDS levels of 400 and 550 ppm, respectively, see figure 5), are entirely different from those observed in the wellfield. This suggests that little, if any, recharge occurs from the northeastern Virginia Mountains.

### 3.2.3 Ground Water Temperatures in the Fish Springs Aquifer System

Further support of the stable isotope data interpretation can be found made when comparing temperatures with sodium levels and discharge rates. In the data available, discharge rates generally correlate with ground water temperatures and TDS levels, suggesting that the larger springs are associated with more extensive flow systems, i.e. deeper circulating waters. The temperature correlations in figures 6, 7 and 8 are impressive. Among the springs those associated with the Warm Springs Fault Zone (all at relatively low elevations between 5000 and 5900 feet), have among some of the higher temperatures and discharge rates (figure 6).

Spring temperatures also correlate reasonably well with sodium (figure 7). Therefore sodium and temperature can be used as qualitative indicators of subsurface residence time (if not as indicators of depth of circulation) in these dilute waters (temperature of Sheep Springs is assumed to be 19 C, as measured by the writer on 7-14-86). Incidentally, temperature does not correlate well with chloride, suggesting that for low TDS recharge waters chloride is a poor indicator of subsurface residence time.



In contrast, discharge temperatures of the two springs sampled in the northeastern Virginia Mountains (Vua and Vuj, see figure 8) are less than 8 C, much lower than the spring temperatures measured in the central Virginia Mountains.

Based on their low temperature it can be argued that these waters did not penetrate very deep. If these waters are assumed not to have penetrated very deep, they should have low TDS levels. However, their comparatively high TDS levels and their stable isotope compositions suggest that these waters were much more subject to evaporation before infiltration. Hence the process of recharge may be less efficient in the northeastern Virginia Mountains. This will be discussed in more detail in subsection 3.2.5.

The correlation between temperature and flowpath (residence time) also applies to the irrigation wells, Fish Spring and the test wells drilled in early 1990 (figure 8). Based on the linear trend continuing from the springs into the wellfield, it can again be argued that the wells are producing from the distal part of a flow system with a source in the north central Virginia Mountains.

The temperature differences between alluvial and bedrock aquifers are noteworthy. In figure 8 the waters from the Hodges, Headquarters and Ferrel wells plot in a cluster, again suggesting the physical homogeneity of the aquifer in the eastern wellfield. Generally the temperatures in the well field are anomalously high for ground water, with a maximum temperature of 27 C in the First Fault Well (Tflt1). The temperature in the alluvial Hodges Well is probably affected by a bedrock aquifer in close proximity. Similarly the temperature in the Neversweat #2 well (completed in bedrock) is 22 C.

In contrast, temperatures from the Ford and Wilson Wells are consistently lower, 15 and 16 C, respectively (the Nork Well with a temperature of 21 C is probably also influenced by bedrock in the close proximity). The temperatures observed in the alluvial wells further north are even lower (between 13 and 15 C). Similarly the temperatures in the three BB-2 piezometers northwest of the Hodges Well range between 14 and 17 C (as expected the temperature increases with depth).

A general pattern of higher temperatures in the bedrock wells (by about 4 C) can not be ignored. Considering that bedrock has a higher thermal conductivity than alluvium, these temperature differences come of no surprise. Clearly the ground water temperatures observed in the bedrock aquifers constitute an anomaly and are attributed to the higher than average heat flow observed in this region (e.g. Blackwell, 1982).

As a fringe benefit, the temperature data can also be used to estimate part of the Fish Springs Aquifer System geometry. Using

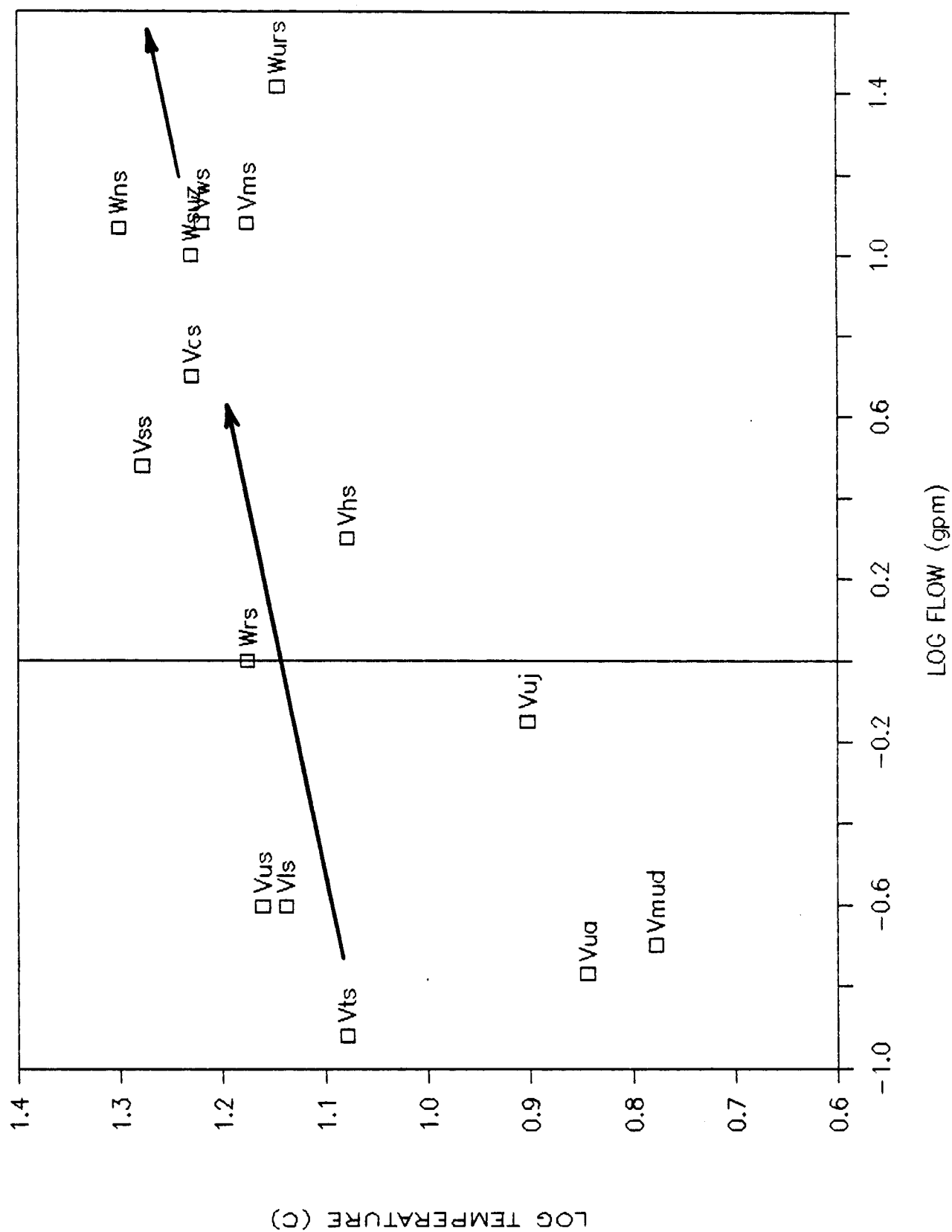


Figure 6: Temperature and spring discharge in the Fort Sage and Virginia Mountains

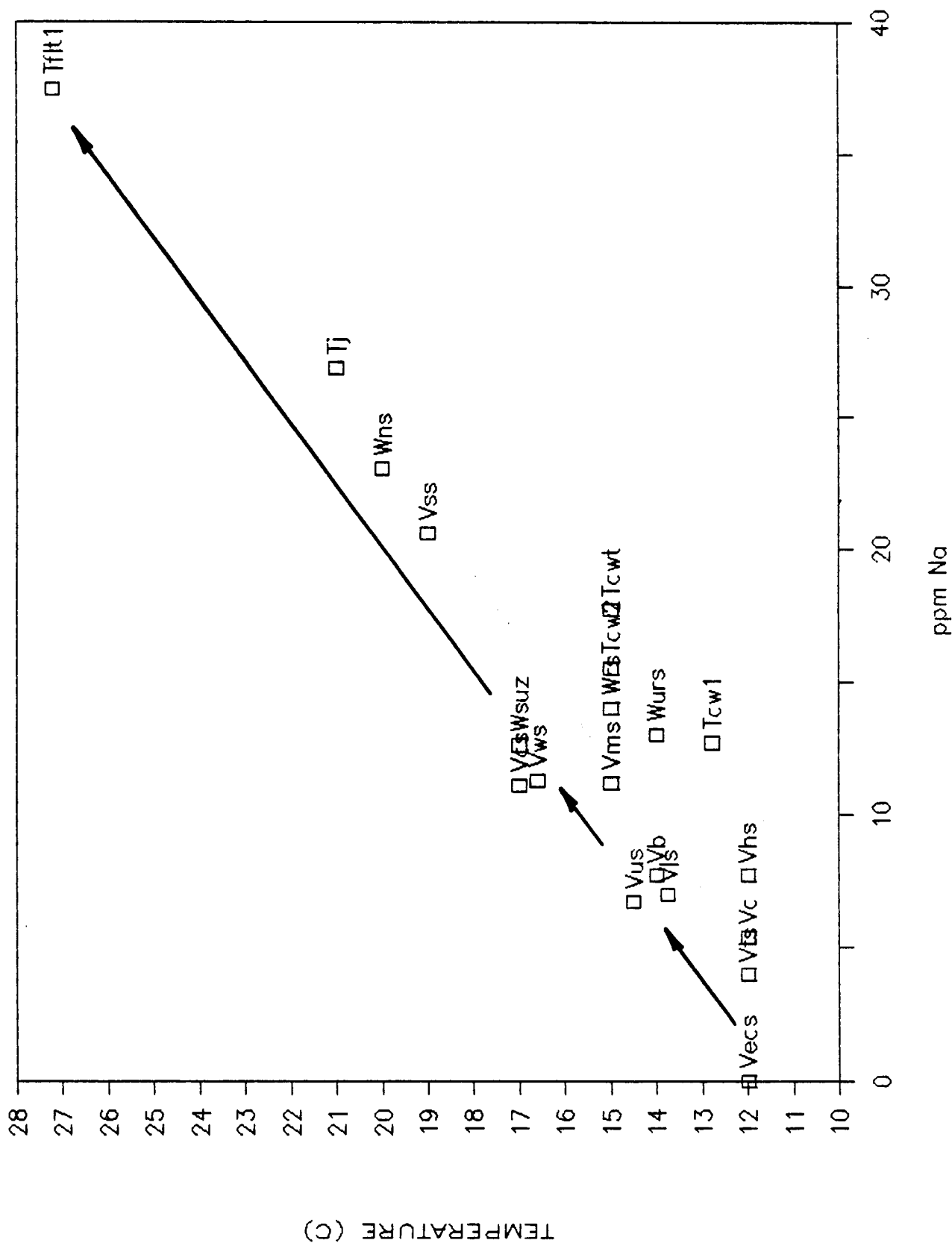


Figure 7: Temperature and sodium concentrations in springs in the Fort Sage and Virginia Mountains

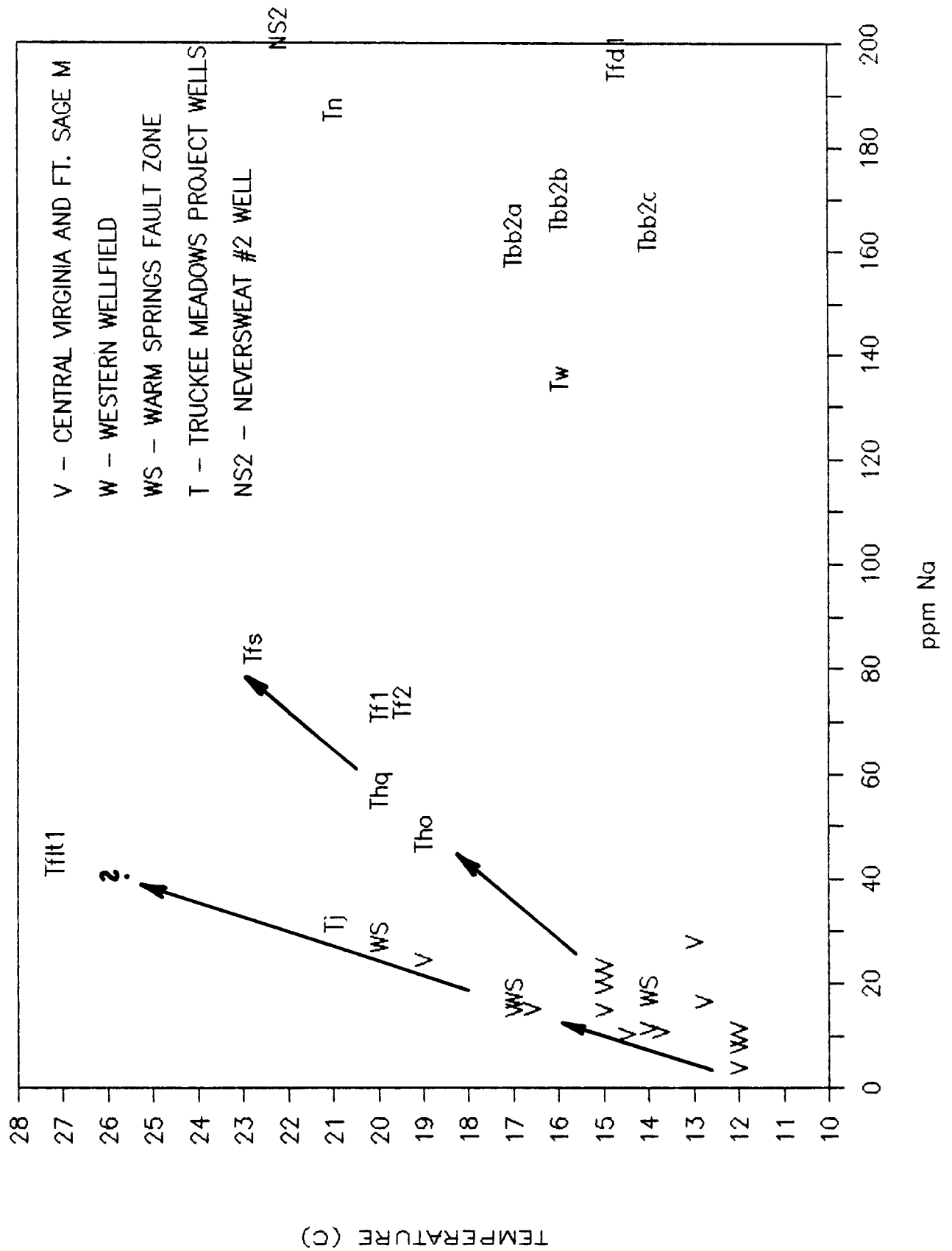


Figure 8: Temperature and sodium concentrations in the Virginia Mountains and irrigation wells.



the geothermal gradient of 1.77 C/100' determined for the Wendel KGRA (Juncal and Bohm, 1987), a depth of circulation between 1000 and 1500 feet can be estimated. Possibly the temperature data could also be used to determine the center of the ground water recharge area (a large batch of temperature data is available from the field study by JBR Consultants). However this would exceed the scope of the present study. Further implications derived from ground water temperatures are discussed in section 3.3.

In summary, the ion ratios and temperature-TDS-discharge relations corroborate the stable isotope data interpretation. The waters produced in the eastern and central wellfield are part of a larger flow system with its source in the Virginia Mountains. In contrast, the Wilson, Nork and Ford wells in the far western wellfield produce waters from an entirely different isotope, chemical and temperature regime.

#### 3.2.4. EFFECTIVE ELEVATION OF RECHARGE

One objective of this study is to estimate the area of recharge by determining mean "effective" elevation of recharge, using stable isotope data. The following discussion shows that it is not possible to quantify the area of recharge by means of an elevation correlation.

A tentative correlation between elevation of recharge and deuterium was established. The approach (including a diagram) is explained in detail in Appendix A.2. However, the relationship suggests that the waters in the wellfield originate from elevations above 7500 feet. Only very small areas occur above 7500 in the Virginia and Fort Sage Mountains, and in the mountain ranges to the south or west, resulting in unreasonably small recharge areas.

It is thus concluded that the isotope signature in this area is not dependent on elevation. Discussions in the literature are somewhat inconclusive. Smith et al. (1979) believe that deuterium content east of the Sierra Nevada is independent of elevation, though Szecsody (1982) found evidence to the contrary. Considering the unreasonably small recharge areas estimated with tentative deuterium-elevation correlations in this study, it must be assumed that the isotope signature is independent of elevation in this area. Most likely the limited range of deuterium found in the wellfield is representative of the area covered by sampling the high elevation springs in the Virginia Mountains.

#### 3.2.5. RECHARGE MECHANISMS BASED ON STABLE ISOTOPE DATA

Since the isotopic composition of precipitation entering the system is difficult to determine, high elevation spring waters,

representing local recharge, are used to obtain the isotope signature of regional recharge water. The isotope data provide indications of recharge mechanisms, and it is worthwhile to take a closer look.

In figure 3, the data from the Virginia, Fort Sage, and southern Skedaddle Mountains do not even plot close to the world meteoric water line (WMWL) of Craig (1969). There are two possible explanations:

1. The Local Meteoric Water Line (LMWL) is different from the World Meteoric Water Line (WMWL).
2. The spring waters (and the wellfield waters) are all affected by evaporation before, during or after infiltration.

The first possibility is highly unlikely, since previous studies done in western Nevada (Ingraham, 1982; Szecsody, 1982; Jacobson et al., 1983; Kolterman, 1984) and in southern Nevada (Benson and Klieforth, 1989) found LMWL's that are, for all practical purposes, identical with the WMWL. Hence it is assumed that the LMWL for this area is also identical with the WMWL. A final clarification can be made once isotope analysis from snow samples are available.

The high elevation springs in the central Virginia Mountains plot on an apparent trend (figure 9). Regression analysis yields the following equation:

$$\text{del-deuterium} = 3.8 \times \text{del-O-18} - 57$$

However, the correlation coefficient of 0.47 is not adequate. A better correlation coefficient of 0.93 is obtained, when the data from the northeastern Virginia Mountains (Vua and Vuj) are included. Then the relation is:

$$\text{del-deuterium} = 4.6 \times \text{del-O-18} - 46.4$$

An almost identical result, with a slightly lower correlation coefficient of 0.87 is obtained when the springs from the southern Skedaddle Mountains (Ssp and Sps) are included. It is encouraging that data from a large region yield such high correlation coefficients, suggesting that stable isotope signatures from the high elevation spring waters are indeed indicative of the waters recharged in the area.

The data are plotted in figure 9 together with the WMWL and the estimated Local Ground Water Line (LGWL), using the second equation above. According to Allison (1982) a slope of about 5.0 indicates evaporation from an unsaturated medium. Hence, it can be argued that water infiltrating into the soil is not so much affected by evaporation from snow, but rather by evaporation from the soil zone, or both. The farther away a water plots from the WMWL, the more it has been subjected to evaporation before entering the local ground water flow system, to emerge in one of

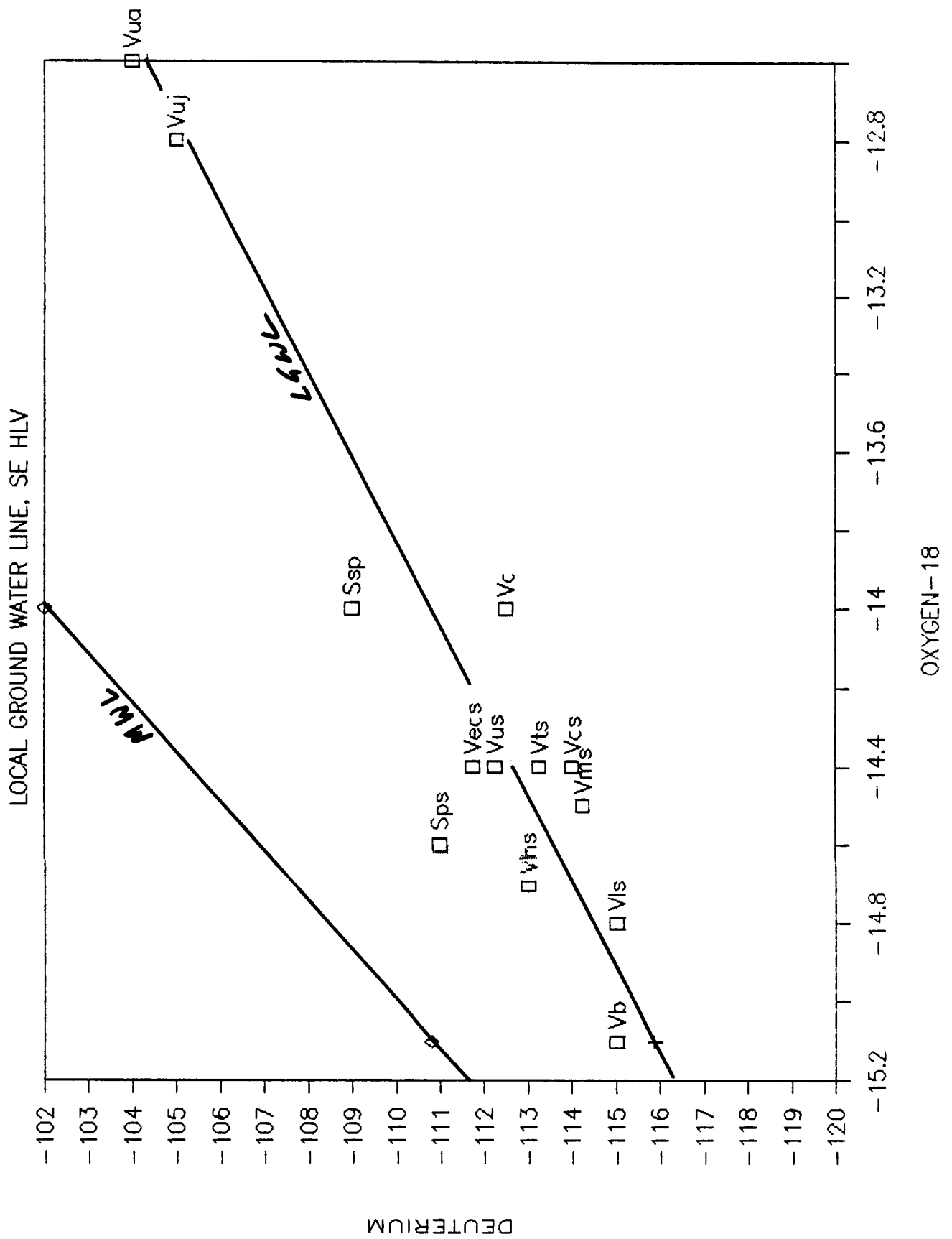


Figure 9: Deuterium and oxygen-18 in springs, Virginia Mountains.

the high elevation springs (or in the wellfield).

Since the isotopic composition of the source water (snow and rain) is not known, the original composition of precipitation in the area must be estimated from the point of intersection between WMWL and LGWL. Assuming that the LGWL came about by site specific evaporation processes, then the composition of the original meteoric source water for the area is indicated by the point of intersection with the WMWL in figure 10:

oxygen-18 = -16.6,            deuterium = -122.7

This source water composition is, for all practical purposes, identical with what Kolterman (1984) estimated for the Virginia City Highlands, even though his LGWL has a slope of only 3.0. Therefore, in the high elevation springs the increase in oxygen-18 and deuterium is between 1.4 and 2.6 per mil and 6.7 to 10 per mil, respectively. Szecsody (1984) observed a deuterium increase of up to 10 per mil from snow to soil water and about three per mil increase in oxygen-18, in the Carson Range. The increase in the northeastern Virginia Mountains is even higher than that. (The "source" should be understood as the meteoric composition, under cumulative time effects, the average).

These results are encouraging, suggesting that the observed trends are indeed meaningful and can be used to determine within limits how much infiltrating precipitation is evaporated before recharge, by comparing snow chloride levels with those in the high elevation springs. It is noteworthy that the low TDS spring waters have chloride levels between 2 and 7 ppm, whereas bicarbonate ranges between 87 and 160 ppm. For a volcanic rock terrain such bicarbonate levels are high (together with low chloride levels). Yet, they are no anomaly in the Great basin spring waters (Bohm and Jacobson, 1983), being indicators of high CO<sub>2</sub> pressures due to evaporation and/or biological activity in the soil. Quantification of recharge under these considerations exceeds the scope of this study.

An unknown at this time is the amount of water lost from evapotranspiration. Since plant evapotranspiration does not affect isotope compositions, the extent of evapotranspiration in the high mountain watersheds has to be determined by other means.

### 3.3 FLOW THROUGH THE WARM SPRINGS FAULT ZONE

As observed in sub-section 3.2.1, spring waters with deuterium levels in the range of the Wilson, Nork and Ford wells are all associated with the Warm Springs Fault Zone. It is thus concluded that the Warm Springs Fault Zone provides an important avenue of ground water migration from the southeast, from an area that receives recharge with depleted deuterium levels (and elevated oxygen-18), compared to those recharged in the Virginia Mountains.

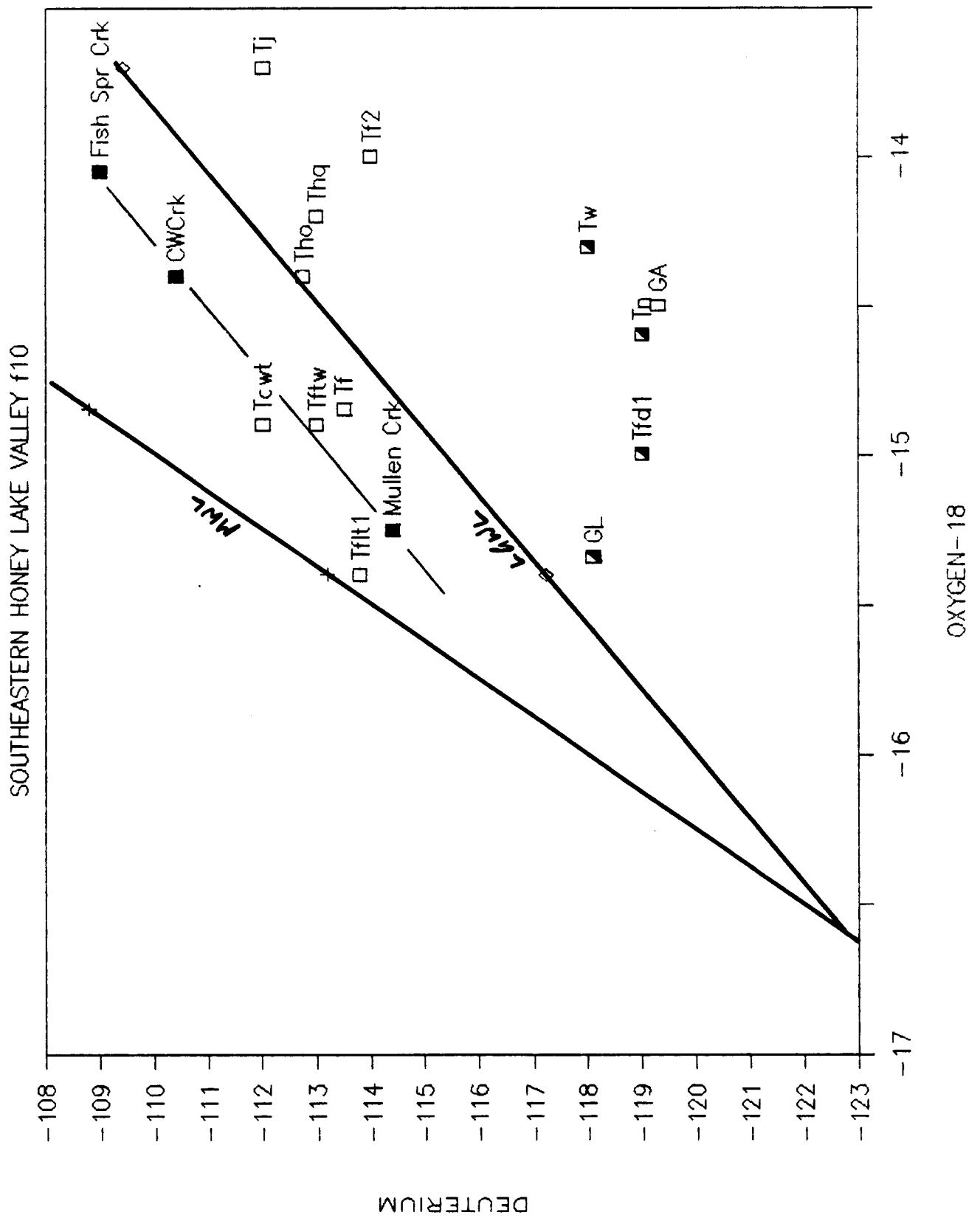


Figure 10: Deuterium and oxygen-18 in the irrigation wells, FSR.



Generally, regional trends in stable isotope composition are believed to be absent in the Great Basin (Ingraham and Taylor, 1986), which makes it difficult to determine the source of the waters migrating through the Warm Springs Fault Zone. However, there seems to be little disagreement that ground waters with lighter deuterium (and O-18 compositions) are associated with waters older than 10,000 years, recharged during the colder Pleistocene (Mariner et al., 1983; Taylor, 1974). These are usually associated with regional flow systems (including geothermal waters). It is therefore concluded that the source of the waters produced from the far western wellfield are the distal parts of a flow system more extensive than that providing recharge further to the east, probably from outside the basin.

It can not be ignored that deuterium and O-18 levels in the western irrigation wells are similar to those observed in Long Valley, Amedee and Wendel Hot Springs (figure 12). There are also similarities in chemical composition, i.e. the WNF wells are sodium-chloride waters, like the geothermal waters. Evidence of a regional geothermal flow system in the granitic basement of Honey Lake Basin, with homogeneous chemical composition, has been discussed by Juncal and Bohm (1987). Therefore the possibility of geothermal waters affecting the wellfield can not be ignored. The anomalous high ground water temperatures throughout most of the irrigation wells are probably an indication of high heat flow (geothermal at depth, see section 3.2.3), masked by the relatively shallow and cold Fish Springs Aquifer System.

Also noticeable are the apparent oxygen-18 shifts in the Wilson, Ford and Nork wells (figure 11). Since a simple oxygen-18 shift would imply deuterium heavier than in present precipitation as deducted in figure 10, the present isotope signature probably came about by a combination of partial evaporation of infiltrated soil water and oxygen-18 shifts.

### 3.4 UNDERFLOW FROM THE WEST

Ground water table maps prepared by Ford et al., (1963) suggest a major flow system entering the southern Honey Lake Valley from Long Valley, between the Fort Sage Mountains and the Diamond Range. At least a part of this flow system may discharge into the high TDS area northwest of the playa. The closed contours on the deuterium and TDS maps also seem to suggest a significant contribution from the west into this area (figures 1 and 2). Obviously, the presence of the high TDS ground waters northwest and west of the playa (assumed to be the ground water sink for southern Honey Lake Valley) preclude recharge from the west into the wellfield area under natural conditions.

### 3.5 GROUND WATER FLOW FROM THE SKEDDADDLE MOUNTAINS

It is unlikely that the Skeddaddle Mountains contribute recharge to the wellfield since an area containing high TDS ground water

SOUTHEASTERN HONEY LAKE VALLEY fig11

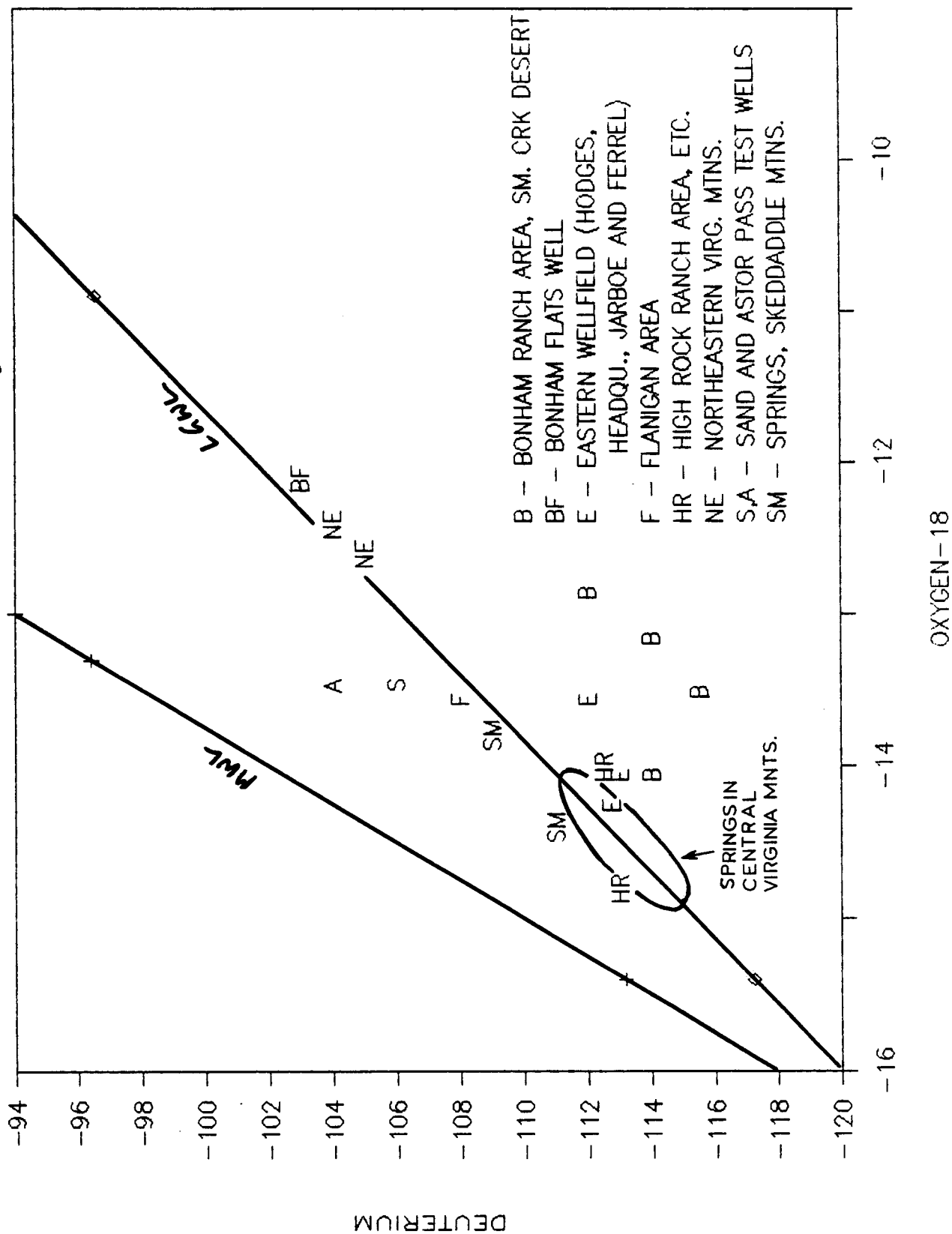


Figure 11: Deuterium and oxygen-18 in selected wells and springs.

is located in-between, in the Flanigan area. Most likely ground water flowing into the basin from the north discharges into this area.

The diagram in Figure 11 essentially summarizes the isotope hydrology of southeastern Honey Lake Valley. The Headquarters, Jarboe, Ferrel and Hodges Wells in the eastern well field have the same deuterium levels as the high elevation springs in the central Virginia Mountains (indicated by the ellipsoid on the LGWL), suggesting a common source area. Oxygen-18 levels in some of the irrigation wells are slightly increased (shifted to the right) due to extended subsurface residence time. Isotope composition of the two Skeddaddle Mountain Springs is similar to that in the springs in the central Virginia Mountain (though deuterium levels are slightly higher).

The isotope composition of the three waters sampled in the High Rock Ranch Area ("HR" on figure 11) north of Flanigan plot in the same range as the Bonham Ranch Well and the two flowing wells located to the south (though some of them have been affected by a geothermal oxygen-18 shift at temperatures ranging between 20 and 30 degrees C). Considering the similarity in isotope signature between these two areas (separated by a mountain range), the recharge area for these waters is best sought in the Skeddaddle Mountains. Indeed, the isotope signature for recharge waters in the Skeddaddle Mountains ("Sps" and "Ssp") is similar, though probably slightly affected by evaporation.

However, the situation is somewhat complicated by the similarity in isotope signatures between the eastern well field, southwestern Smoke Creek Desert (Bonham Ranch area) and the High Rock Ranch area waters (for locations see Plate 1). Hence, it can also be argued that ground water flows from southeastern Honey Lake Valley into Smoke Creek Desert and/or the High Rock Ranch area. The latter possibility is precluded by the water table configuration. Flow into Smoke Creek Desert is of particular interest since ground water modeling of the Fish Springs Aquifer System (Handman et al., 1990) suggests flow north- and eastward through Dry Valley. This hypothesis should be tested with isotope data from the only well located in Dry Valley, and by comparing water level data.

### 3.6 THE SAND AND ASTOR PASS AREAS

This area has been subject to close scrutiny due to the results of ground water modeling by the United States Geological Survey (1990), implying ground water flow from west to east through Sand and Astor Pass. Ground water chemistry and isotope data provide valuable insight into the hydrology of this area.

Ground water isotope signatures in the test wells at Sand Pass and Astor Pass are entirely different when compared with the waters sampled elsewhere in southern Honey Lake Valley (in particular those sampled in the wellfield). In figure 11 they

plot above the LGWL, but below the MWL, clearly suggesting a different source area than for the eastern irrigation wells. Since the isotope compositions in the springs sampled in the Skeddaddle and northern Virginia Mountains (Sps, Ssp, Vuj, Vua) and in the Bonham Flats area (BF) are different, it is unlikely that these areas or the Terraced Hills to the east provide recharge for the Sand and Astor Pass ground waters.

In the absence of organic rich sediments that would cause a deuterium shift (deuterium increase), the plotting position of the Sand and Astor Pass test wells is best interpreted as an oxygen shift to the right from the MWL. Deuterium levels are similar to those in waters from the Needles Well (plotting outside Figure 11) at the northern shore of Pyramid Lake, however, the oxygen-18 shift is by far not as drastic. Low temperatures observed in these wells do not necessarily preclude geothermal effects (similar as in the western irrigation wells), considering the presence of hydrothermally altered pyroclastic rocks (halloysite clay deposits) mapped by Bonham and Papke (1969, page 110) along the western side of the Terraced Hills.

Chemical compositions of the Sand and Astor Pass Wells are compared in Figure 13. The chloride waters in the Sand Pass area are very homogeneous, and similar to the Bonham Wells. Two explanations are possible:

1. Two different ground water flow systems merge between Sand Pass and Bonham Ranch, where their chemical composition is determined by the same geochemical regime.
2. The Sand Pass ground waters were recharged in the Skeddaddle Mountains (affected by evaporation before infiltration), i.e. a source similar to that of the Bonham Wells. Considering the margin of error in lab analysis it is possible that they differ from the Astor Pass waters.

In contrast, the Astor Pass waters are chloride-bicarbonate waters, with bicarbonate levels (greater than 900 mg/l) exceeding that of any other waters sampled in southern Honey Lake Valley. The source of the high bicarbonate levels is not clear at this time.

Geochemical and isotope evidence suggests that ground waters sampled in the Sand and Astor Pass areas originate from a source dissimilar to the eastern wellfield. Probably they originate from a similar source as The Needles geothermal waters (though the chemical composition is different). These observations suggest that little, if any ground water migrates from southeastern Honey Lake Valley through Sand and Astor Pass.

### 3.7 GROUND WATER FLOW FROM THE NORTHEASTERN VIRGINIA MOUNTAINS

The stable isotope data suggest that under natural conditions no contributions occur from the northeastern Virginia Mountains into

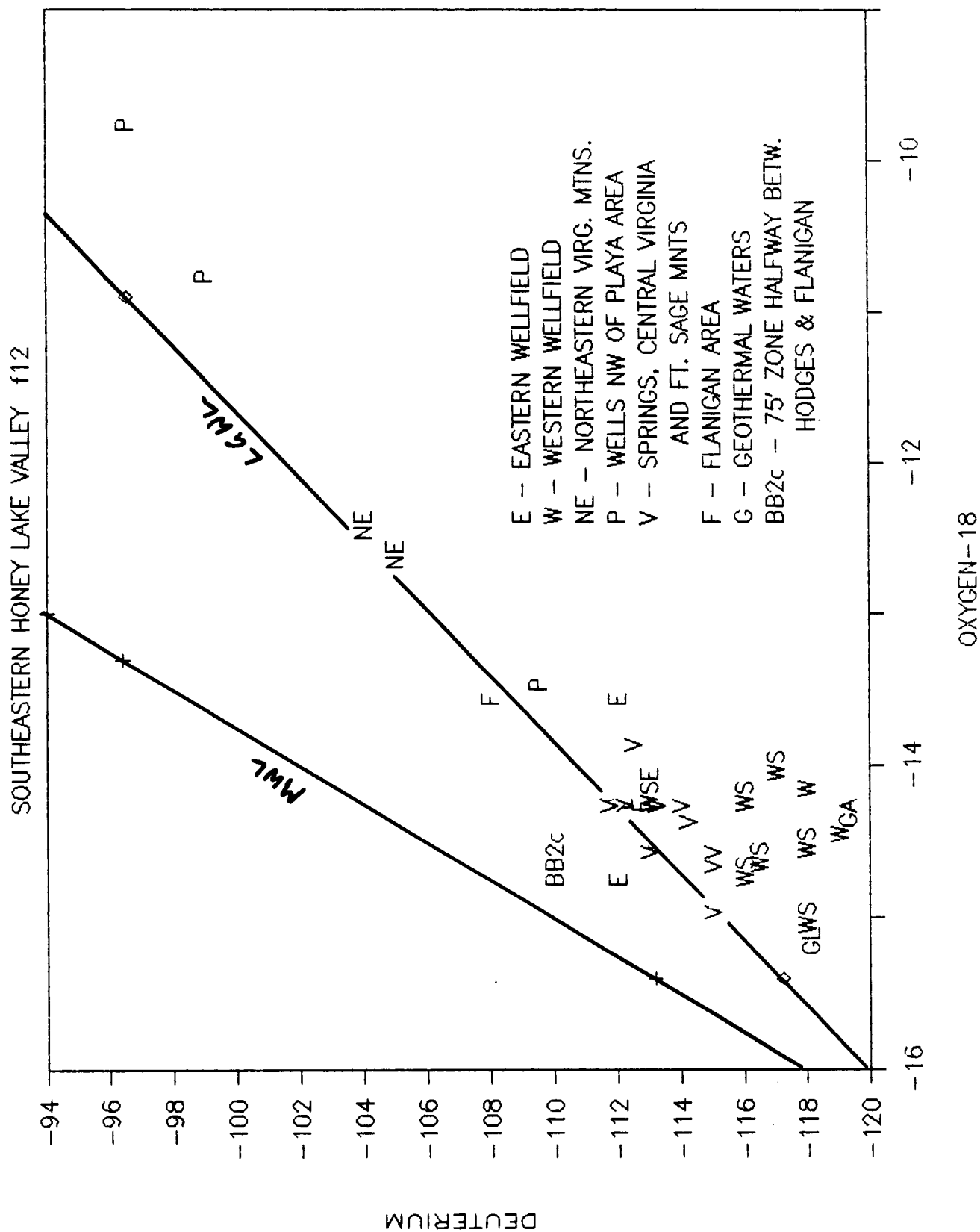


Figure 12: Deuterium and oxygen-18 in selected wells and evaporation trends.

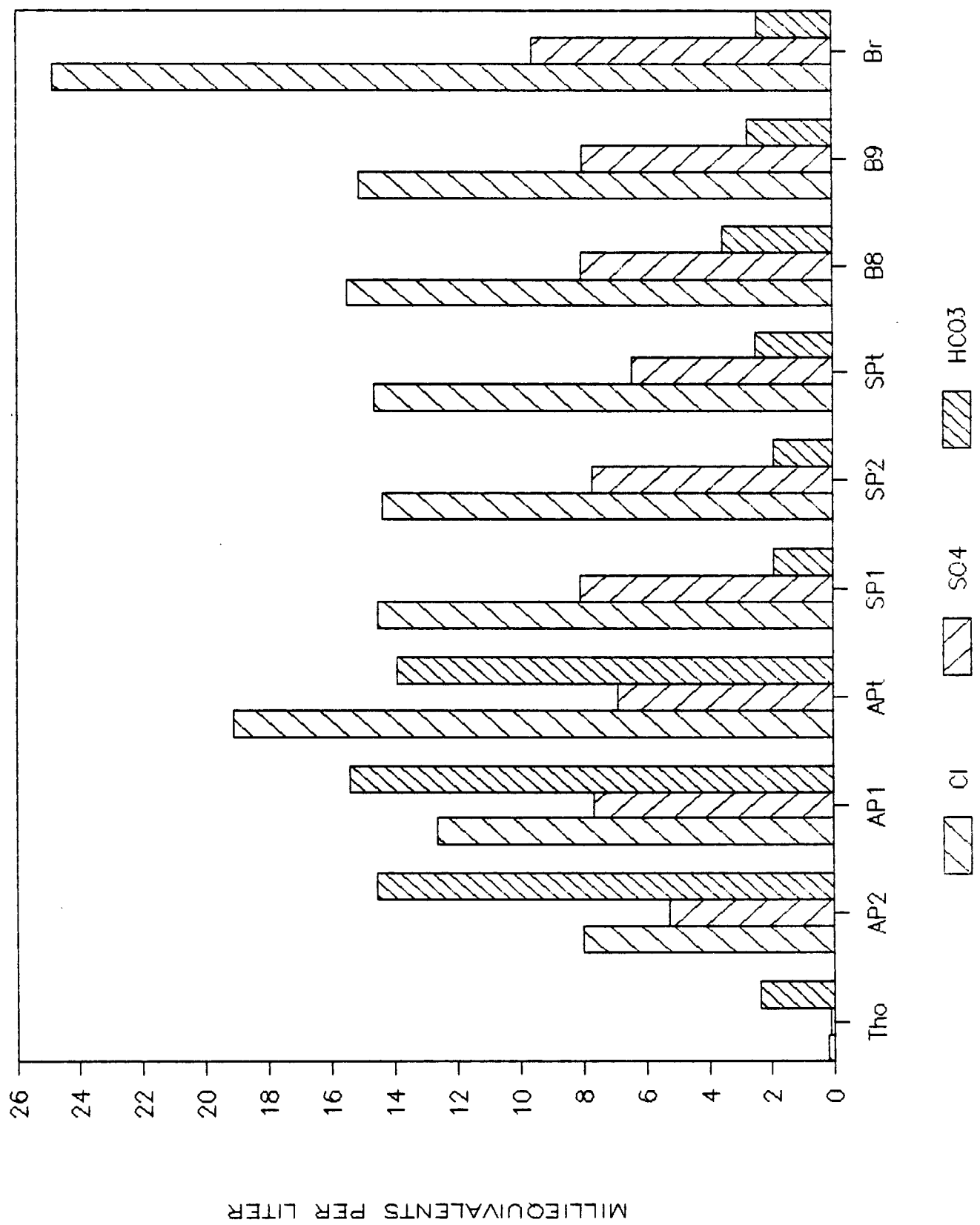


Figure 13: Ground water chemistry in Sand and Astor Pass.



the eastern wellfield (section 3.2.2). The large isotope shift along the LGWL, and the anomalous high TDS levels in combination with low chloride (less than 20 ppm) and high bicarbonate (greater than 270 ppm) are typical for waters strongly affected by evaporation in the soil zone. This pattern suggests that little recharge occurs in the northeastern Virginia Mountains. Nevertheless, the hydrochemical data from the BB-2 piezometers seem to suggest that some ground water migrates northwest from the Neversweat Hills, maybe towards the Flanigan area (June 30, 1990 Memorandum prepared by PGH for Mike Widmer, Hydrologist, WCUD).

The waters sampled in the BB-2 piezometers may be a mixture between waters similar to the Hodges Well and waters from the Neversweat Hills area. Hence, waters originating from the Hodges area may after all not discharge into the Flanigan area, but probably to the west and/or to the north into Dry Valley (after mixing with Neversweat waters). This is supported by the isotope data discussed in section 3.2.2.

#### 4.0 DISCHARGE IN THE AREAS NORTH AND NORTHWEST OF THE PLAYA

As was discussed in subsection 3.1, due to the high levels of TDS in the ground waters sampled from the wells north, west and northwest of the playa this area is considered to be the ultimate discharge area for the ground water flow systems in southeastern Honey Lake Valley. The May and Austin wells (P3 and P2) located in this area, have among the highest levels of TDS in the basin (43,000 and 46,000 ppm, respectively). In this section the ground water discharge areas are evaluated using the stable isotope and hydrogeochemical data.

It is noteworthy that all the high TDS waters (denoted "P" and "F1") sampled in this area plot close to the LGWL in figure 12. Since the LGWL seems to indicate evaporation from an unsaturated medium (subsection 3.2.4), significant evaporation seems to occur through the unsaturated zone. To what extent evaporation can occur through 20 to 30 feet of unsaturated alluvium remains to be clarified, though the process of "evaporative pumping" postulated by Hsu and Siegenthaler (1969) could provide an explanation. Quantifying the amount of evaporation based on chloride and isotopes exceeds the scope of this study.

Unclear remains the role of plants in removing water from the subsurface and raising the chemical concentration of ground water in this area. According to Handman et al. (1990, Plate 4), the entire valley floor of the basin east of the stateline is covered mostly by "bare playa and sparse phreatophytes" and to a lesser extent by "mixed phreatophytes". Ground water uptake by plants results only in insignificant isotope fractionation, and ground water removal by "phreatophytes" can not be discerned with the data at hand.

However, it is possible that some of the high TDS waters (e.g. P1

and F11) which show less of an evaporation isotope shift have been affected by transpiration (it is unlikely that the TDS of about 5000 ppm in P1 is the result of geothermal processes since none of the geothermal waters in the valley exceed 2000 ppm TDS). On the other hand, in an area where "phreatophytes" have been mapped such as the water sampled at 75 feet in BB2c, evaporation or transpiration effects can not be implied from the chemical and isotope data (see discussion below), though it is possible that plants in some areas rely on perched precipitation. This could be clarified by collecting isotope data from plant tissue fluids and comparing them with ground and precipitation waters. Therefore the relative importance of phreatophytes in removing ground water remains unclarified and the word "evapotranspiration" should be understood as a process whereby ground water is removed by means of evaporation or transpiration, or both, not implying the relative importance of either process.

Chloride and the stable isotopes are conservative components (i.e. not involved in chemical reactions) and are useful as tracers. Evaporation trends can be observed, when plotting chloride against deuterium or oxygen allowing conclusions about the source of the high TDS waters (discussed in more detail in the concurrent study for WWDC). However, lines of evaporative concentration are ambiguous due to the large differences in chloride concentrations between low TDS water and the evaporative brines.

In figure 12 the brines sampled in the May and Austin wells (P2 and P3), the waters from the Wilson, Nork and Ford Wells (Tw, Tn, Tfd1) and the geothermal waters (GL, GA) are offset from the LGWL by roughly equal amounts (up to +1.5 per mil O-18). This could suggest an evaporation line connecting the western wellfield waters with the high TDS waters, parallel to the LGWL. However, in the absence of more isotope data from the high TDS ground waters in this area, evaporation trends do not necessarily have to parallel those observed in the Local Ground Water Line (LGWL).

An anomaly among the isotope data is the water sampled between 75 and 85 feet in the BB-2c piezometer, halfway between the Hodges Well and Flanigan. Among all stable isotope data used in this study, this is the only sample that, for all practical purposes, plots on the meteoric water line (MWL, see figure 12). The comparatively high TDS of 513 ppm may be due to salts (accumulated in the soil during the dry season) flushed out by precipitation on the valley floor. This process could be verified by taking seasonal samples from BB-2c. According to Smith and Drever (1976) this may be a common process affecting shallow ground waters at low elevations.

Ground water temperatures in the ground water discharge area, which are comparable with those in the Wilson and Ford wells, could also point to a source in the western wellfield. However, this interpretation remains ambiguous when considering the lower thermal conductivity (and lower geothermal gradients) in the alluvium and the uncertainty of evaporation effects on ground

water temperature (see section 3.2.3).

The question about the source of water in the high ground water TDS area may eventually be finalized by means of geochemical modeling (concurrent study for WWDC). However, modeling evaporation trends has been found to be difficult with the computer codes at hand. Complications arise due to the uncertainty about chemical composition of the vertical ground water flow component and residual salts from evaporation of paleolake waters. Nevertheless, the high TDS waters are variable in composition (figure 14), with TDS ranging from 2,400 to 46,000 ppm. All high TDS waters are of the sodium-chloride type. Peculiar are the extremely low sulfate values in some of these waters, suggesting sulfate reduction due to decomposition of organics in lakebeds or decreased oxygen solubility inherent in evaporation processes. Hydrogen sulfate has been noticed in most of these samples.

A preliminary interpretation using the rationale of "chemical divides" (Hardie and Eugster, 1970) does not preclude either water from the eastern and western wellfield as well as waters from the Duck Lake, High Rock Ranch and Herlong areas as source waters. This is mostly because some of the waters assumed to discharge from the west and north are of similar chemical compositions as those discharged in the wellfield.

An unknown component of flow discharged upward from a bedrock aquifer at depth can not be ignored. This would be the distal part of a regional flow system, using the schematic depicted in Mifflin (1968, p.12). It is interesting to note that when comparing figure 2 with plate 1 the high TDS zone has a northeastern trend, apparently parallel to two implied faults (one implied by Harding Lawson Associates (1989) and one by Guyton Assoc. (1986)). It is therefore possible that the location of the ground water discharge area is structurally controlled.

Depth to bedrock could be estimated by obtaining temperature gradients from the wells in this area. A rough estimate of depth to bedrock can be obtained by using the temperature gradient estimated from temperatures measured in waters from the BB-2a and BB2-b piezometers. Assuming a ground water temperature of 23 C (discharge temperature of the former Fish Spring) at the bedrock-alluvium interface, depth to bedrock underneath the alluvium between Flanigan and the Hodges Well would be about 700 feet. Unfortunately no temperature data are available from the BB-1 test hole.

A test hole (BB-1) was drilled in the spring of 1990 amidst the playa, midway between the May and Headquarters Wells. The hole intersected 700 feet of clays and silts. Though no water sample was obtained, downhole logging data suggest TDS levels between 400 and 1000 ppm (Steve Roberti, WELENCO, Reno). This suggests the presence of low TDS waters underneath the Playa, even though a highly transmissive zone was not encountered to that depth.

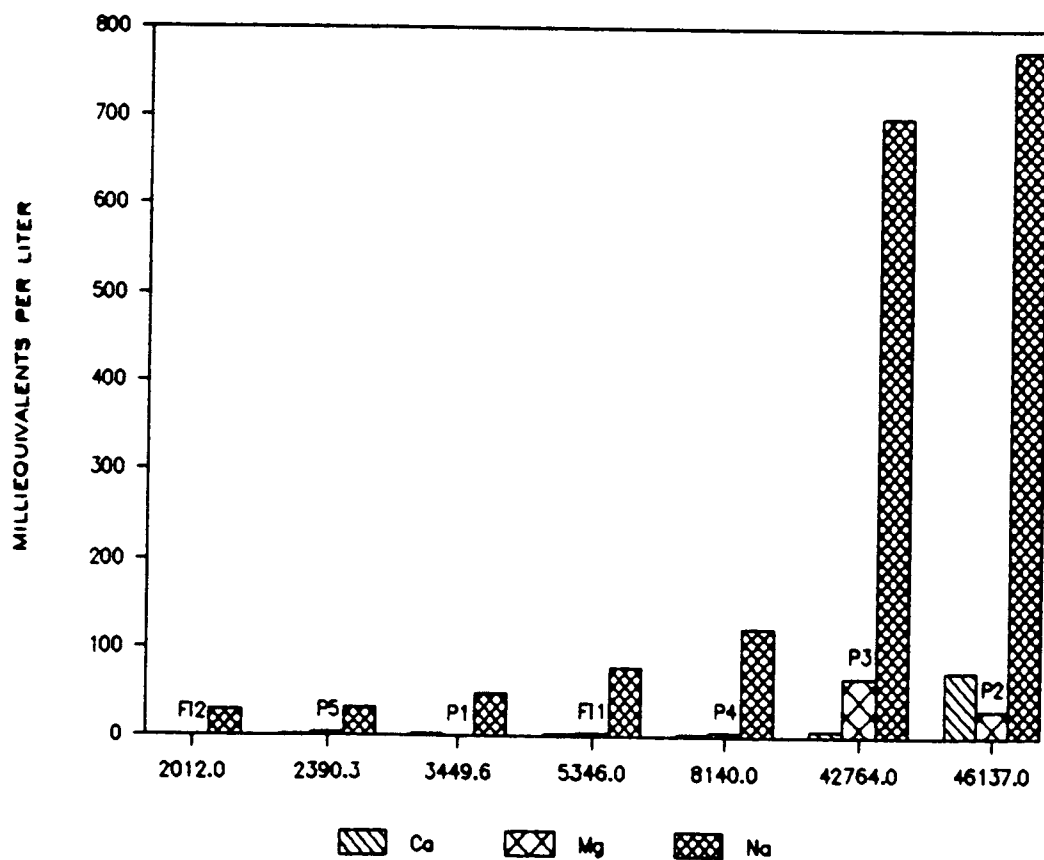
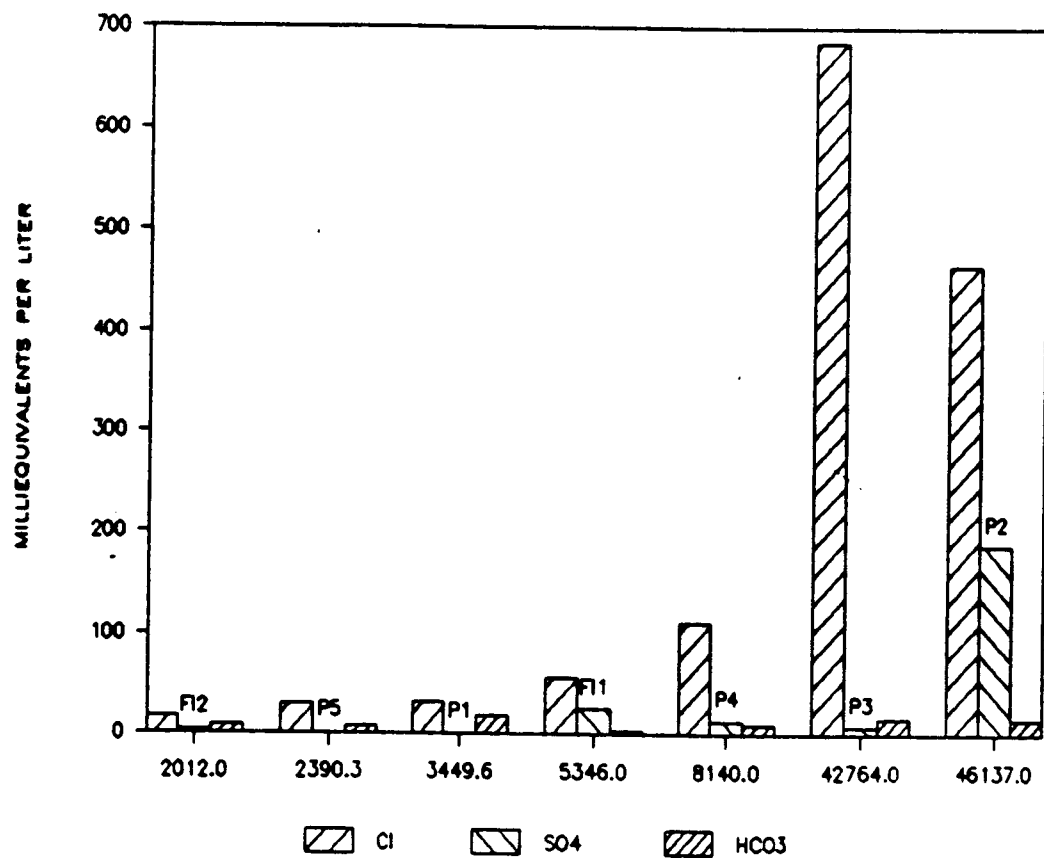


Figure 14: Chemistry of high TDS ground waters in southeastern Honey Lake Valley.

Fluids sampled from a 15 foot auger hole about 1 mile south of BB-1 had a conductivity of 8,900 micromhos (Rush and Glancy, 1967). This suggests that high TDS pore fluids accumulate at least in a shallow zone due to evaporation of either surface or ground waters (or both) from the Playa.

## 5.0 SUBSURFACE RESIDENCE TIME BASED ON TRITIUM DATA

For the purpose of estimating subsurface residence time in the Fish Springs aquifer system a total of fifteen samples were analyzed for tritium (enriched). The samples include five wells from the wellfield, the three Bonham Ranch wells and seven springs from the Virginia Mountains. All samples had tritium below detection limit (3 TU), except for Cold Spring ("Vc"), located at 5900 feet in the Virginia Mountains. A brief rationale for the methodology is given in Appendix B.

There are several ways of interpreting the available data. Precipitation samples taken in 1982 in the Carson Range had significant tritium concentrations (Szecsody, 1982). Since all but one sample collected for this study had non-detectable tritium levels, there is probably little question that most of the water produced in the wellfield was recharged more than 35 years ago. The same must apply for most high elevation springs.

Ground water flow does not occur like "piston flow", and most likely recharge occurs all along the flow path in the highlands. Therefore it is possible that the waters sampled contain some modern tritium, though highly diluted (i.e. in concentrations below detection limit). In this sense, the concentration of 8 TU observed in Cold Spring is best interpreted as a result of mixing between old and "post-bomb" waters (less than 35 years). A similar situation may occur in all other waters where tritium levels are below detection limit.

It is interesting to know that for most waters collected in the Great Basin no tritium has been detected so far (Roger Jacobson, Desert Research Institute, personal communication), which suggests that most waters discharged are older than 35 years. In this respect the Fish Springs aquifer system is no exception.

Still, assuming the waters have been recharged at least 35 years ago, their effective age remains open. A maximum age can be established realizing the occurrence of the Warm Springs Fault Zone flow system, with an isotope composition that suggests recharge during a climate cooler than that during recharge of the eastern wellfield waters. Assuming that the eastern waters were recharged in postglacial times, a maximum ground water age of 8,000 to 10,000 years can be established. This could be of importance when estimating ground water flow rates based on average subsurface residence time. The stable isotope data show very little variance (section 3.2), suggesting a well mixed aquifer system. The same can be said about the major ion compositions. Also the low TDS levels in both high mountain

spring waters and the wells in the eastern wellfield seem to suggest that the waters have not changed significantly in their chemical composition from recharge area to discharge area. Though no quantitative estimate can be made, this suggests relatively rapid flow across the distance of the area sampled (Tule Peak to the wellfield). This should come of no surprise in light of the high transmissivities observed in the bedrock aquifers at Fish Springs Ranch.

Nevertheless, assuming a range of age (residence time), one could calculate a minimum and maximum flow rate (recharge rates), assuming certain ranges of porosity, thickness and areal extent of the Fish Springs Aquifer System (the temperature data may be helpful in determining aquifer geometry, section 3.2.3). However, the procedure and constants entered into such a calculation should be carefully scrutinized. Furthermore the southward extent of the recharge area should also be confirmed.

## 6.0 SUMMARY AND CONCLUSIONS

The hydrogeochemistry of southeastern HLV was studied in detail, using environmental isotope and major ion chemistry data, from more than 100 sampling sites. Though the data covered the entire southern basin from Herlong to Astor Pass and the Bonham Ranch area in Smoke Creek Desert, the main emphasis was placed on analyzing the Fish Springs aquifer system, including its recharge and discharge areas. The analysis provided the following results:

Areal distribution of deuterium and anion concentrations and anion-ratios indicate that the entire hydrologic system of southern HLV discharges into the area northwest of the playa as the ultimate ground water and geochemical sink.

The recharge area for the Fish Springs aquifer system is located at least in the northern Virginia Mountains. The stable isotope data suggest that it may reach as far south as Tule Peak. However, this needs to be supported with other data, such as geological structure. The volcanic bedrock apparently provides reasonably good hydraulic communication between the highlands and the pediment aquifers tapped in the central and eastern wellfield, i.e. the area from the Hodges to the Ferrel wells.

The Fish Springs Aquifer System is chemically and isotopically very homogeneous, suggesting a system that is well mixed with good hydraulic communication and relatively rapid mass transfer. Ground water temperatures in springs and wells discharging from fractured bedrock suggest high heat flow or significant depths of circulation. Assuming geothermal gradients as in the Wendel KGRA a minimum depth of circulation of 1500 ft is estimated. Temperatures gradually increase along a presumed flow path from the southern Virginia Mountains, towards the irrigation wells and Fish Spring.

Tritium data suggest that most of the water pumped from the eastern wellfield has been recharged at least 35 years ago, though the possibility of mixing and dilution with more recent waters can not be excluded. However, the stable isotope data interpretation suggests that the waters in the eastern wellfield may be younger than 10,000 years and those in the western wellfield may be older than 10,000 years.

The far western wellfield represented by the Wilson, Nork and Ford wells, produces water from an entirely different flow system, with different chemical and isotopic compositions and different temperatures. This flow system has its origin in the Warm Springs Fault Zone, where ground water is discharged from fractured bedrock into the alluvial aquifers. These waters are strongly affected by geothermal waters, if not conductively cooled geothermal water themselves.

In contrast, the source of the waters discharged from the western wells and the Warm Springs Fault Zone must be further south or southeast along the trend of the Warm Springs Fault Zone.

The geochemical and isotope data suggest that under natural conditions no recharge occurs into the wellfield from the northeastern Virginia Mountains.

Evidence derived from stable isotope, geochemical and temperature data remains ambiguous as to the source of the high TDS waters north and northwest of the Playa. Sources may include waters from the eastern and western wellfield, the area to the west and the Skeddaddle Mountains.

The stable isotope data allow for the possibility of ground water being discharged from the eastern wellfield to the north through Dry Valley. Yet, the presence of water migrating from the Neversweat Hills to the Flanigan Area (based on major ion chemistry arguments) seems to be in contradiction with this observation.

Contrary to the hydrologic interpretation based on previous modeling efforts by the Handman et al. (1990), the isotope data suggest that no ground water flows from Honey Lake Valley through the Astor Pass area into Pyramid Lake Valley.

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## APPENDIX A: CALCULATIONS

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### APPENDIX A.1: HYDROCHEMICAL FACIES

The concept of hydrochemical facies was developed by Back (1960), distinguishing certain types of ground water chemistry. The chemical types indicate certain chemical processes in the aquifer material, useful for distinguishing ground water flow systems and source rocks, as well as discharge areas.

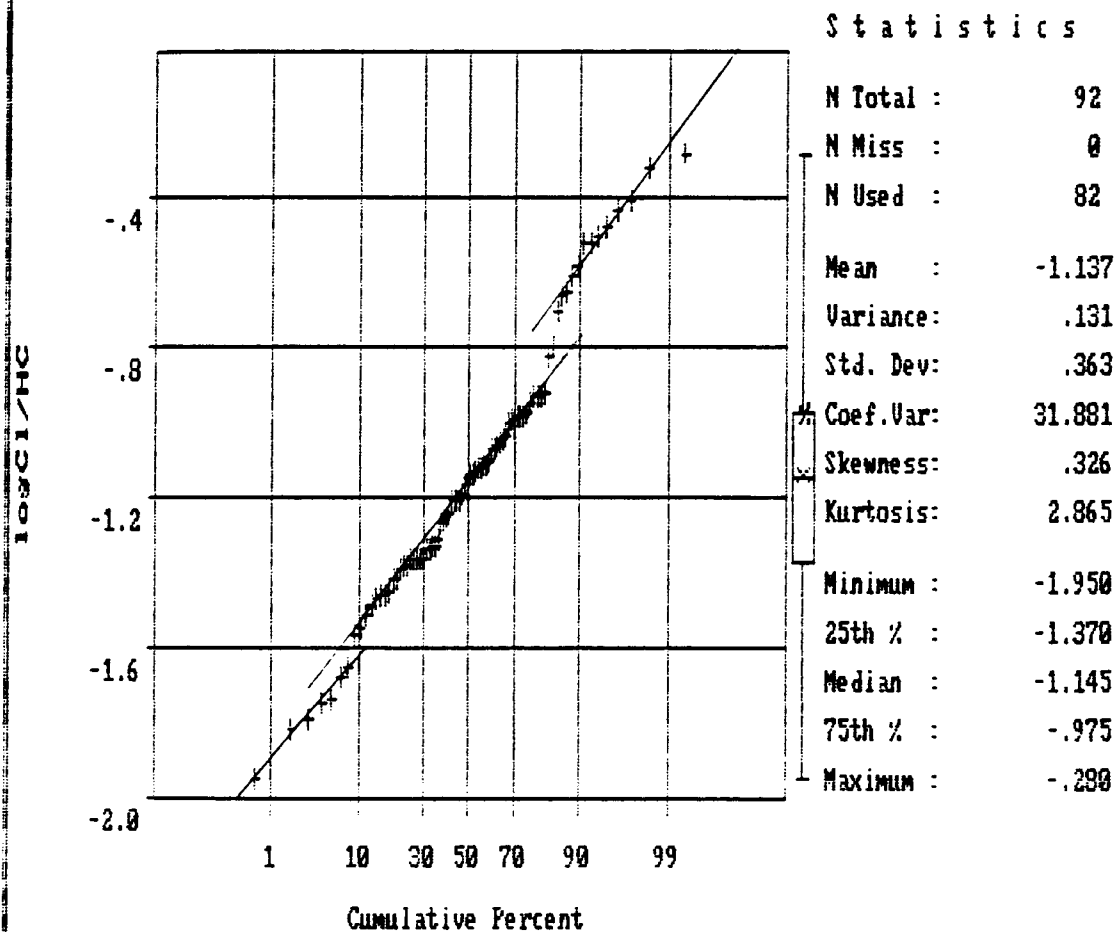
The log-ratio of  $Cl/HCO_3$  was found to be a useful distinction of ground water facies. Low ratios indicate little chemical interaction with aquifer materials, whereas higher ratios reflect increasing TDS levels and subsequent removal of bicarbonate due to calcite precipitation. Therefore an increasing ratio is a qualitative indicator of subsurface residence time (if not geochemical evolution) in a particular hydrochemical environment.

The log-ratios were plotted on a cumulative frequency diagram on probability paper, using the GeoEAS geostatistical software package. Using the methodology developed by Sinclair (1973) four straight line segments were used to identify four subpopulations, representing facies A, B, C, and D:

Facies A:	less than -1.6
Facies B:	-1.6 to -0.9
Facies C:	-0.9 to 0.0
Facies D:	0.0 to 1.6

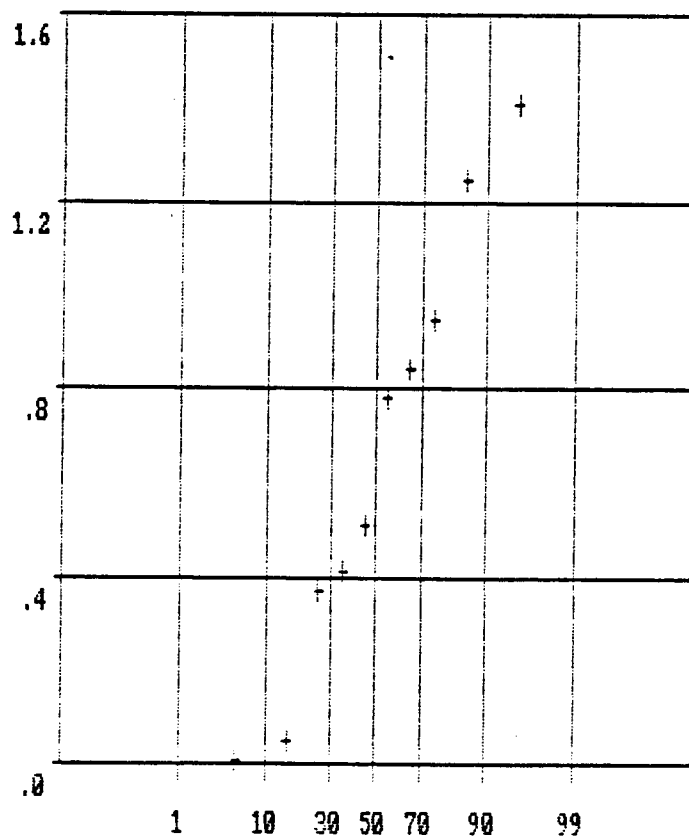
The facies denomination for each water sampled in the basin was plotted in figure 4.

Normal Probability Plot for logCl/HC  
Data file:



FACIES :    A    < -1.6  
               B    -1.6 < B < -.9  
               C    -.9 < C < 0.0  
               D    0.0 < D < 1.6

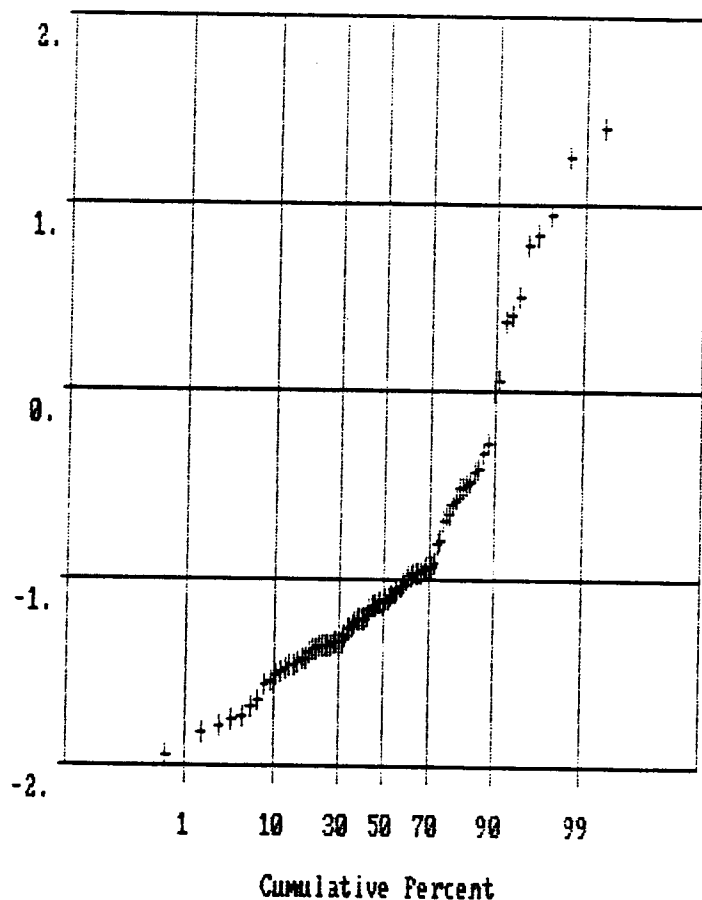
Normal Probability Plot for logCI/HC  
Data file:



Statistics

N Total :	92
N Miss :	0
N Used :	10
Mean :	.658
Variance:	.223
Std. Dev:	.472
% Coef.Var:	71.734
Skewness:	.141
Kurtosis:	1.928
Minimum :	.010
25th % :	.210
Median :	-1.115
75th % :	.895
Maximum :	1.410

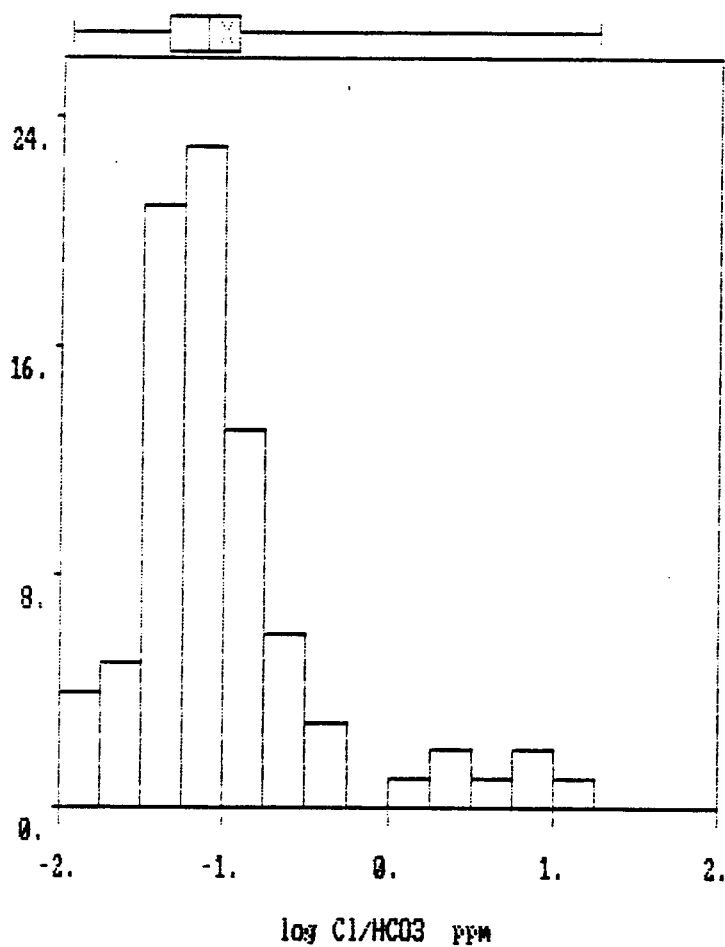
Normal Probability Plot for logCl/HC  
Data file:



Statistics

N Total :	92
N Miss :	0
N Used :	92
Mean :	-.942
Variance:	.455
Std. Dev:	.674
% Coef.Var:	71.580
Skewness:	1.626
Kurtosis:	5.561
Minimum :	-1.950
25th % :	-1.360
Median :	-1.115
75th % :	-.800
Maximum :	1.410

TRUCKEE MEADOWS PROJECT  
Data file FSR-WRK.wk1



Statistics

N Total :	82
N Miss :	0
N Used :	82
Mean :	-1.013
Variance:	.362
Std. Dev:	.602
% Coef.Var:	59.422
Skewness:	1.790
Kurtosis:	6.593
Minimum :	-1.950
25th % :	-1.360
Median :	-1.125
75th % :	-.930
Maximum :	1.250



## APPENDIX A.2: ESTIMATES OF EFFECTIVE RECHARGE ELEVATION

A considerable amount of time was spent determining effective elevation of recharge by estimating the relationship between deuterium and elevation.

One approach used was by entering various chemical variables that are an indirect indicator of distance of flow residence time into multiple regression analysis. However, the regression coefficients always remained less than .5, suggesting that this approach is not very useful for estimating the original elevation of recharge, probably mostly because the horizontal distance of flow is uncertain.

Better results were attained by looking at the local hydrology of high and intermediate elevation springs. The following criteria were used to estimate recharge elevation of local ground water flow systems in high mountain watersheds:

1. To assure elevation dependent recharge the watershed should be located at high elevations, i.e. above 6000 feet.
2. The waters should be low TDS waters and have a low  $Cl/HCO_3$ -ratio.
3. Preferably the springs should be located on slopes of west facing aspect, to eliminate potential errors due to turbulence on leeward mountain sides.

The deuterium levels of the low TDS wells in the project area are within the range of the deuterium levels observed in most of the springs sampled in the Virginia Mountains. Most sampling sites were selected such that their subsurface hydrology could be defined reasonably well, including the most likely elevation of recharge. High elevation springs located in small, well defined basins were assumed to be locally recharged ground water. The recharge elevation was based on the elevation of the highest ridge in the local basin, estimated to the highest one hundred foot interval. Springs with an ambiguous hydrologic situation were not used to develop a deuterium-elevation line.

The springs found to be most suitable for this kind of analysis are listed in table A.1.3 and results of regression analysis are given on the same table. The springs are plotted at their discharge elevations (fig. ) and at the estimated recharge elevation, together with the regression line (fig. ). Based on the range in deuterium levels observed in the irrigation wells, two major recharge elevations are defined:

- 1: Areas between 7500 and 8000 feet, providing recharge for most of the wellfield, from the Ferrel to the Hodges Wells. Mountains ranges with such elevations are the Virginia and Fort Sage Mountains.
- 2: Areas located between 9000 and 9200 feet provides recharge

for the Ford, Wilson and Nork wells. None of the mountain ranges in this area are higher than 8000 feet and no recharge area higher than 9000 feet within a reasonable distance can be identified. The implications of this observation are discussed below.

Deducting a tentative recharge elevation, recharge for most of the wellfield was assumed to occur above 7500 feet, and the area was estimated. The estimated recharge area appears unrealistically low. It therefore seems that the homogeneity in deuterium signature suggests that recharge occurs over a wider range of elevation.

With a few exceptions, the deuterium levels in the Virginia and Fort Sage Mountains show a general negative correlation with the TDS levels (see diagram, TDS vs. Deuterium). Exceptions are Rat Spring (FSrs) discharging from granitic rocks and Suzie Spring (Wsuz). Generally, this diagram suggests that the waters recharged at colder temperatures (implying higher elevations) where discharged along longer flow paths, i.e. longer subsurface residence time (higher TDS). This observation suggests that the method of estimating average recharge elevation from deuterium levels maybe applicable in this limited area. Hence a higher elevation recharge area has to be sought for the waters with deuterium levels between -118 and -116 per mil, including the Nork, Wilson and Ford Wells. However, in the absence of such areas in the immediate area, the possibility of ice-age waters is more likely (see section 3.3).

### A.3. THE LOCAL GROUND WATER LINE (LGWL):

The following springs were used to calculate the local ground water line (LGWL):

Code	Oxygen-18	Deuterium	WMWL	LGWL	Elevation
Vb	-15.1	-115	-110.8	-115.86	6420
Vc	-14	-112.5	-102	-110.8	5900
Vcs	-14.4	-114	-105.2	-112.64	4650
Vecs	-14.4	-112	-105.2	-112.64	6600
Vhs	-14.7	-113	-107.6	-114.02	7750
Vls	-14.8	-115	-108.4	-114.48	7230
Vms	-14.4	-114	-105.2	-112.64	5350
Vts	-14.4	-113	-105.2	-112.64	7800
Vua	-12.6	-104	-90.8	-104.36	4700
Vuj	-12.8	-105	-92.4	-105.28	5100
Vus	-14.4	-112	-105.2	-112.64	7550

#### Regression Output:

Constant	-46.4370
Std Err of Y Est	1.032340
R Squared	0.931820
No. of Observations	11
Degrees of Freedom	9
X Coefficient(s)	4.607006
Std Err of Coef.	0.415390

## APPENDIX B: BASIC PRINCIPLES OF ISOTOPE HYDROLOGY

The following is a brief introduction into the principles of isotope hydrology as applied to ground water resource evaluation. A great deal of this material has been adapted from reports prepared by the author for geothermal and ground water exploration in other areas. For a detailed treatment the reader is referred to Faure (1977) and Fritz and Fontes (1980).

The environmental isotope composition of ground waters can provide valuable information on recharge areas, flow paths and subsurface residence times and recharge rates. The most commonly used environmental isotopes are deuterium, oxygen-18 and tritium. Others used in more sophisticated studies are isotopes of chlorine, sulfur and carbon.

### **Deuterium and Oxygen-18 (stable isotopes):**

Deuterium (D) together with oxygen-18 (O-18), referred to as stable isotopes, used most frequently to delineate ground water flow systems, are heavier isotopes of hydrogen and oxygen. They are useful since the heavier isotopes are fractionated during evaporation and precipitation. Because any water (rain or snow) condensed from atmospheric vapor (clouds) will be richer in the heavier O-18 and D than the original vapor, any subsequent precipitation from the same air mass must be lower in these isotopes than the earlier condensate. For that reason O-18 and D contents of rain and snow are dependent on both elevation of precipitation and its distance from the point of evaporation (ocean). The exact composition of meteoric (atmospheric) water is also dependent upon the temperature of water condensation and, as is particularly the case with rain in arid environments, how much evaporation the precipitation is subjected to during its descent. The importance of this effect on ground water composition in the Great Basin is minimal since most recharge occurs from snow melt. Precipitation under cold conditions (high altitudes and/or high latitude snow) is more depleted with respect to both D and O-18 and plots on the lower portion of the meteoric water line (MWL) which is the lower left hand corner in figure 1. On the other hand, precipitation under warmer conditions (low elevation) is more enriched with respect to D and O-18 on the MWL, i.e. in the upper left hand corner of figure 1.

The D and O-18 composition of ground water can be altered by several processes. The most common one is evaporation. In figure 1 the trend of evaporation of a water composition at point A is indicated as a range of trends, showing various ratios of enrichment of O-18 and D. The direction on the diagram depends on various factors, the most important one of which is the temperature of evaporation.

Changes due to evaporation can occur in various parts of a ground water flow system. First, partial evaporation can occur in the soil or near soil zone in a recharge area, essentially distorting

the input signal provided by the meteoric water at that particular elevation. Second, evaporation can occur in the discharge area. This is particularly important during sampling work where the effects of phreatophytes, steam separation in geothermal springs and general evaporation in playa environments can have an important impact in low discharge springs. Data from the Great Basin suggest that quite frequently shallow water table aquifers can be affected by evaporation. However, this can also be due to recharge of shallow valley fill aquifers from surface runoff, which is subject to partial evaporation, percolating into the alluvial fans. Overall the effects of evaporation tend to complicate the interpretation of D and O-18 data.

Once the water is incorporated into the deeper ground water flow system the D and O-18 composition can be changed by two mechanisms. The most common process is that of rock-water interaction which increases the O-18 content since the minerals that are weathered by hydrolysis contain "heavier" oxygen. The amount of change is a function of residence time and most importantly temperature, resulting in the "oxygen shift" commonly observed in geothermal waters (see figure 1). A less commonly observed water-rock reaction is the so-called "deuterium shift" when water reacts with hydrogen bearing rocks (rich in clay minerals, hydrogen sulfide or organic substances). The process of stable isotope exchange between rock (heavy) and water (light) increases with temperature. Hence a large oxygen shift can be a qualitative indicator of subsurface temperature.

A second factor determining the magnitude of the oxygen shift is the rock-to-water ratio. Hence the effects of low permeability rocks (longer residence time and high clay content) should not be ignored. However, for the low temperature and low TDS waters observed in the pumping wells no significant oxygen shift is anticipated due to the presumed short subsurface residence time.

Alteration of D and O-18 composition can also occur by mixing of waters of various compositions. Where mixing occurs in varying proportions, mixing lines can be developed on a D-O-18 diagram. Only seldom are data suitable for this kind of analysis.

#### Tritium:

The unstable isotopes tritium (T) and carbon-14 (C-14) are used to determine subsurface residence time of ground waters (age-dating). Tritium is an unstable isotope of hydrogen introduced into the atmosphere during nuclear bomb tests. Testing on a large scale did not start until 1953. Hence the tritium content is useful to provide an approximate minimum or maximum age of ground water. For example a water with a significant amount of tritium suggests a ground water recharged within the last 35 years. Since atmospheric testing does not occur on a large scale any more, no more significant amounts of "bomb tritium" is introduced into the hydrologic cycle, and most water analysis for tritium require time consuming enrichment processes in the laboratory. So far no bomb tritium has been detected in ground waters of the western

Great Basin, suggesting that the residence time of most ground waters in this area is more than 35 years.

Theoretically it is possible to calculate the actual age of ground water with detectable tritium, using the tritium decay curve and historical levels of atmospheric tritium. However, this is not very useful in practice since piston flow in most aquifers is an unrealistic assumption. A further complication are uncertainties in geographical and seasonal variations of historical tritium in the atmosphere. At best tritium can provide an indication whether any "modern" water (recharged less than 35 years ago) is present in a ground water of mixed age.

#### Carbon-14:

A more elaborate way of determining subsurface residence time is by means of carbon-14 dating. This method is more expensive and is applied usually only when the tritium method yields no results, and the demand of a particular project makes the effort worthwhile. Radioactive carbon-14 is generated in the upper atmosphere by the interaction of solar radiation with nitrogen molecules. Natural carbon-14 introduced into ground water via the soil zone is isolated from "younger" carbon in the atmosphere. Since it decays at a predictable rate it can provide an indication of how long ago the water was recharged. The method yields age data in the range 2,500 to 50,000 years.

## APPENDIX C: DATA

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### Tritium Data:

SAMPLE	Code	Tritium
Well 3891	B8	-3.13
Well 3902	B9	-3.13
Bonham Ranch	Br	-3.13
Ferrel W. #2	Tf2	-3.13
Hodges Well	Tho	-3.13
Headquarters Well	Thq	-3.13
Jarboe Well	Tj	-3.13
Wilson Well	Tw	-3.13
Cold Spr.	Vc	8.44
Cottonwood Spring	Vcs	-3.13
Willow Spr.	Vws	-3.13
APD	Wa	-3.13
Sugarcane	Ws	-3.13
Spring de Casa	Wsc	-3.13
Suzie	Wsuz	-3.13



SAMPLE	Code	x-miles	y-miles	Location
Astor Pass 1	AP1	10.53	8.82	27/19-24bb
Astor Pass 2	AP2	9.36	8.88	27/19-23ba
Astor Pass Test Well	APt	9.73	8.88	27/19-14D
Well 3891	B8	11.13	14.45	T28R20-19SW, S
Well 3902	B9	10.18	15.14	T28R19-24NW, S
Bonham Flats Well	BF	10.43	11.50	27/19-1
Bonham Ranch	Br	10.48	16.92	T28R19-12NE
Bonham Ranch	Br	10.48	16.92	T28R19-12NE, S
27/17-15p2	DL1	-2.78	7.97	27/17-15p2
Duck Lake Irrig. W.	DL2	-2.73	7.75	T27R17-22NENW
27/17-23d1	DL3	-1.82	7.75	27/17-23d1
27/19-31ba	F11	5.45	6.95	27/19-31ba
Tipton Well	F12	4.97	6.20	18/27-36dd
2TP Spr.	FS2tp	0.21	-3.21	T25R18-20NW, F
26/17-22g1	FSf1	-2.25	1.50	26/17-22g1
26/17-25g1	FSf2	-0.43	0.48	26/17-25g1
Mullen Creek	FSf3	0.64	-0.48	T25R18-5
Hail Domestic W.	FSf4	-1.02	0.53	26/17-25SEnw
Austin Dom. W.	FSf5	0.48	1.39	26/18-29baa
McCorkle Dom. W.	FSf6	0.75	1.50	26/18-29ad
Amedee Hot Spr. #3	GA			T28R16-8SWNE
Zamboni H.S.	GL			T24R17-24NENW
Wendel Hot Spring	GW			T29R15-23NESW
26/16-3d2	Hb1	-9.09	4.81	26/16-3d2
Well #2, Army Depot	Hb2	-6.58	5.08	27/16-36q4
Well #5, Army Depot	Hb5	-7.75	5.03	27/16-35p1
Well #9, Army Depot	Hb6	-7.65	4.22	26/16-2g1
25/16-12j1	Hd1	-7.11	-2.73	25/16-12j1
25/17-7j1	Hd2	-6.31	-2.25	25/17-7j1
25/17-7j2	Hd3	-5.99	-2.57	25/17-7j2
25/17-8n3	Hd4	-5.70	-2.94	25/17-8n3
25/17-4n1	Hd5	-4.60	-1.98	25/17-4n1
25/17-20b1	Hd6	-4.81	-4.55	25/17-20b1
25/17-21n1	Hd7	-4.49	-5.13	25/17-21n1
25/17-29h1	Hd8	-5.24	-5.88	25/17-29h1
26/15-2m1	Hj1	-13.90	4.28	26/15-2m1
26/16-6d1	Hj2	-12.03	4.65	26/16-6d1
26/16-16q1	Hs1	-9.52	1.98	26/16-16q1
26/16-15e3	Hs2	-9.09	2.41	26/16-15e3
26/16-13m1	Hs3	-7.11	2.25	26/16-13m1
26/16-13a1	Hs4	-6.47	2.67	26/16-13a1
26/17-19e1	Hs5	-6.10	1.60	26/17-19e1
Never Sweat #2	NS2	9.52	5.03	26/19-02DBC
Perkle Dom. W.	P1	1.18	2.51	26/18-21bc
Austin North Dom. W.	P2	1.07	3.37	26/18-16cb
May Well	P3	3.37	4.39	18/26-11ca
Stanley Well	P4	2.46	3.05	18/26-15dc1
26/17-1c1	P5	-0.64	4.71	26/17-1c1
28/17-18k1	Smf1	-5.24	14.17	28/17-18k1
Hi-Rock Ranch	Smf10	-0.80	12.03	28/17-36
28/17-2011	Smf2	-4.55	13.37	28/17-2011
28/17-2013	Smf3	-4.44	12.99	28/17-2013
Artesian Well #1	Smf4	-4.22	14.39	T28R17-17NWSE

SAMPLE	Code	x-miles	y-miles	Location
Artesian Well #2	Smf5	-5.61	13.64	T28R17-19NWNNE
Artesian Well #3	Smf6	-5.51	13.37	T28R17-19S1/2N
CV Ranch Irrig. W.	Smf7	-4.76	13.26	T28R17-20SESW
Flaniken Ranch Well	Smf8			
Skedaddle Mtn Ranch	Smf9			
Sand Pass #1	SP1	11.23	12.83	27/20-01
Sand Pass #2	SP2	12.30	13.90	27/20-01
Pipe Spring	Sps	4.39	18.09	29/18-36dbb
Sand Pass Test Well	SPt	11.76	13.37	27/20-01
Spencer Spring	Ssp	-5.21	18.34	29/17-32NWNW
28/17-26c1	St1	-1.44	12.83	28/17-26c1
27/17-3ad	St10	-2.35	10.21	27/17-3ad
28/17-25p3	St2	-0.53	12.14	28/17-25p3
28/17-34r1	St3	-2.19	11.02	28/17-34r1
27/17-3a1	St4	-2.19	10.80	27/17-3a1
27/17-4a3	St5	-3.21	10.80	27/17-4a3
27/17-4a2	St6	-3.16	10.59	27/17-4a2
27/17-1p1	St7	-0.64	10.05	27/17-1p1
28/17-25cd	St8	-0.53	11.98	28/17-25cd
Windmill Well	St9	-1.44	12.03	T28R17-26SESW
BB2-A 465-475	Tbb2a	6.84	5.35	26/19-05DAD
BB2-B 250-260	Tbb2b	6.84	5.35	26/19-05DAD
BB2-C 75-85	Tbb2c	6.84	5.35	26/19-05DAD
Cottonwood #1	Tcw1	8.82	3.05	26/19-26b
Cottonwood #2	Tcw2	8.93	2.89	26/19-14CCC
Cottonwood Test Well	Tcwt	8.82	1.87	26/19-27AA
Ferrel Well/USGS	Tf	4.28	1.55	T26R18-25bd
Ferrel W. #1	Tf1	4.57	1.66	26/18-25ac
Ferrel W. #2	Tf2	4.28	1.55	T26R18-25bd
Ford Well #1	Tfd1	2.67	2.25	26/18-22da
Ford Well #2	Tfd2	2.83	2.03	T26R18-22dd
First Fault Well	Tflt1	4.97	0.96	26/19-31bbb
Ferrel Playa Well	Tfpl	4.92	1.93	26/19-30BBC
Fish Springs	Tfs	5.88	2.14	26/19-19dd
Ferrel Test Well	Tftw	4.33	1.76	26/18-25BD
Hodges Well	Tho	8.36	4.28	T26R19-10cd
Headquarters Well	Thq	6.26	1.82	T26R19-29ba,
Headquarters 1966	Thq	6.26	1.82	26/19-29ba
Jarboe Well	Tj	5.83	0.94	T26R19-30dd,
Nork Well	Tn	2.49	2.03	T26R18-26db
Wilson Well	Tw	3.50	1.50	T26R18-26db
Black Canyon	Vb	9.84	-4.17	T25R19-26NE
Cold Spr.	Vc	9.09	-1.07	T25R19-11NW
Cottonwood Creek ??	Vcc	9.09	2.33	T26R19-26SWNW
Cottonwood Spring ??	Vcs	9.14	1.82	T26R19-26SWNW
E. Cottonwood Spr.	Vecs	9.06	-1.87	T25R19-11SW
Fish Springs Creek	Vfsc	2.67	0.00	T25R18-3NWNNE
Horse Spring	Vhs	12.83	-5.61	T25R19-32
Lower Scott	Vls	12.14	-3.21	T25R20-20NW
Mustang Spr.	Vms	6.36	-1.44	T25R19-8NW
Mud Spr.	Vmud	11.18	-2.14	T25R20-18NW
Sheep Spr.	Vss	3.58	-0.80	T25R18-2SE, N
Sheep Trough Spring	Vst	3.58	-0.80	T25R18-2SE, N

SAMPLE	Code	x-miles	y-miles	Location
Tule Spr.	Vts	13.10	-3.21	T25R20-21NW
Upper Adobe Spr.	Vua	11.76	4.22	26/20-7
Upper Juniper Spr.	Vuj	10.75	4.55	26/19-12
Upper Scott	Vus	12.46	-3.74	T25R20-20SW
Willow Spr.	Vws	4.76	0.13	T26R18-36SE
APD	Wa	10.70	-7.97	T24R19-12SE
Newcombe Spring	Wns	6.84	-3.96	25/19-20dcd
Rat Spr.	Wrs	0.16	-1.55	T25R18-9NW
Sugarcane	Ws	10.70	-8.45	T24R19-13NE
Spring de Casa	Wsc	10.59	-7.70	T24R19-12
Suzie	Wsuz	11.23	-7.97	T24R20-7SW
Upper Red Spring	Wurs	8.88	-5.94	25/19-34dd

Code	date	depth	source	temp	gpm	f-ec	f-pH	l-pH
AP1	2-9-90		WC/SEM					8.3
AP2	2-9-90		WC/SEM					8.5
APt	4-03-90	440	WC/SEM	21.0				7.9
B8	10-3-89		WC/SEM	22.2	6	2200		8.4
B9	10-3-89		WC/SEM	20.0	1	2300		8.1
BF			WC/DRI					
Br	1989		USGS/89					
Br	10-3-89		WC/SEM	30.0	40	3500		8.2
DL1	4-16-88	53	DWR	12.2		585	8.9	7.9
DL2	1976		USBR/76	18.0				7.7
DL3	11-7-88	44	DWR	15.0		1020	8.4	8
F11	8-23-66		USGS/66			7720		7.8
F12	2-6-90	180	PGH	13.3		4800	8.46	8.1
FS2tp	9-28-89		WC/SEM		0.02			7.5
FSf1	4-13-88	60	DWR	28.3		460	8.6	8.1
FSf2	4-13-88	80	DWR	17.8		263	8.6	7.9
FSf3	1989		USGS/89					
FSf4	1-21-90	35	PGH	5.0		260		8.6
FSf5	1-21-90		PGH	16.0		360		8.1
FSf6	1-24-90		PGH	13.0		350	7.36	7.8
GA	1976		USBR/76	96.0				8.4
GL	1976		USGS/83	41.0				9.37
GW	1976		USBR/76	95.5				8.3
Hb1	6-9-77	80	DWR	16.7		520	8.3	8.4
Hb2	7-16-62	700	DWR/US89	15.6		1100		7.2
Hb5	10-28-59	550	DWR/USGS	15.6		543		7.5
Hb6	7-24-80	530	DWR/US89	21.1		455	8.3	7.6
Hd1	10-12-88	158	DWR	20.0		228	8.4	7.4
Hd2	10-12-88	130	DWR				8.6	7.6
Hd3	10-12-88	160	DWR	17.8		281	8.5	7.6
Hd4	9-29-80	513	DWR/USGS	14.4		228		6.4
Hd5	9-29-80	350	DWR	22.8		227		7
Hd6		260	DWR	16.1		470	6.6	7.1
Hd7	6-13-85	120	DWR	16.1		198		7
Hd8	6-9-77	80	DWR	17.2		240	7.8	6.8
Hj1	9-29-80	564	DWR	19.4		199		7.3
Hj2	7-31-74	75	DWR	18.9		466	7.1	7.3
Hs1	6-25-85	650	DWR	15.6		410	7.9	7
Hs2	6-12-85	60	DWR	15.6		700	8	7
Hs3	4-13-88	250	DWR	16.1		430	8.7	7.1
Hs4	4-13-88	145	DWR	18.9		1100	8.6	7.8
Hs5	6-12-85	160	DWR	19.4		950	8.1	7.4
NS2	1-30-90		WC/SEM	22.2				8.2
P1	1-21-90		PGH			5360		8.2
P2	1-24-90		PGH	13.6			7.9	7.8
P3	2-6-90	250	PGH	12.4			7.09	7.5
P4	2-6-90	490	PGH					8.3
P5	11-7-88	300	DWR	15.0		4100	7.4	8.3
Smf1	6-11-85	100	DWR	15.6		320	8.1	8
Smf10	8-86		USGS	30.0		240		8.25
Smf2	10-13-88	85	DWR	16.7		230	8.4	7.9
Smf3	10-13-88	520	DWR	25.6		220	8.4	8.1
Smf4	1976		USBR/76	15.0				9

Code	date	depth	source	temp	gpm	f-ec	f-pH	l-pH
Smf5	1976		USBR/76	23.0				8.4
Smf6	1976		USBR/76	26.5				8.2
Smf7	1976		USBR/76	26.0				8.1
Smf8	8-86		USGS	22.0		226		8
Smf9	8-86		USGS	27.0		208		8.18
SP1	3-27-90	712	WC/SEM	15.5				8.7
SP2	3-27-90	480	WC/SEM	16.3				8.2
Sps			PGH					
SPT	4-08-90	580	WC/SEM					8.1
Ssp			PGH					
St1	10-11-88	100	DWR	18.3		257	8.5	8.2
St10	8-23-66		USGS/66	19.4		1560		8.4
St2	10-13-88	100	DWR	20.6		245	8.6	8.3
St3	10-11-88	152	DWR	15.6		383	8.5	7.7
St4	11-7-88		DWR	9.4		765	7.5	7.7
St5	5-4-84	500	DWR			962		8.1
St6		120	DWR	13.9		966		8.9
St7	11-7-88		DWR	15.0		5250	9	8.3
St8	8-23-66		USGS/66	30.0		253		8.1
St9	1976		USBR/76	14.0				8.2
Tbb2a	5-14-90	475	WC/SEM	17.0				8.2
Tbb2b	5-14-90	260	WC/SEM	16.0				8.1
Tbb2c	5-14-90	85	WC/SEM	14.0				8.5
Tcw1			WC/SEM	12.8				7
Tcw2	5-08-90	495	WC/SEM	15.0				7.6
Tcwt	5-10-90	665	WC/SEM	15.0				7.1
Tf	1989		USGS/89	20.0				
Tf1	7-27-66		USGS/66	20.0		303		7.9
Tf2	8-29-89		WC/SEM	19.5	2000			8.3
Tfd1	1-24-90		PGH	14.7		1046	8.37	8.5
Tfd2	8-21-84		Nork/SEM				7.8	7.8
Tflt1	5-27-90	500	WC/SEM	27.2				
Tfpl	3-30-90	477	WC/SEM					8.3
Tfs	7-27-66		USGS/66	22.8		328		8
Tftw	4-23-90	470	WC/SEM					8.2
Tho	9-12-89		WC/SEM	19.0	1500			7.6
Thq	8-29-89		WC/SEM	20.0	1800			8.2
Thq	8-23-66		USGS/66	20.0		287		8
Tj	9-6-89		WC/SEM	21.0	1050	110		8.2
Tn	9-15-89		WC/SEM	21.0	400	1000		8.1
Tw	8-29-89		WC/SEM	16.0	2000			8.2
Vb	9-12-89		WC/SEM	14.0	3	140		7.4
Vc	9-12-89		WC/SEM	12.0	5	110		7.3
Vcc	7-3-86		NSDH/US89	15.0				8.4
Vcs	9-7-89		WC/SEM	17.0	5	110		8
Vecs	9-12-89		WC/SEM	12.0	0.00	110		
Vfsc	1989		USGS/89					
Vhs	9-13-89		WC/SEM	12.0	2	160		7.3
Vls	9-13-89		WC/SEM	13.8	0.25	190		7.8
Vms	9-15-89		WC/SEM	15.0	12	160		7.3
Vmud	9-13-89		WC/SEM	6.0	0.2	110		7.2
Vss	9-28-89		WC/SEM	19.0	3	140		7.5
Vst	1989		JBR	13.0		390	6.9	7.63

Code	date	depth	source	temp	gpm	f-ec	f-pH	l-pH
Vts	9-13-89		WC/SEM	12.0	0.12	65		7.2
Vua	1989		JBR	7.0	0.17	675	7.5	7.81
Vuj	1989		JBR	8.0	0.71	545	7.5	8.22
Vus	9-13-89		WC/SEM	14.5	0.25	110		6.8
Vws	9-15-89		WC/SEM	16.6	12	140		7.7
Wa	8-31-89		WC/SEM		25	130		8.1
Wns	1989		JBR	20.0	11.7		7.3	7.2
Wrs	9-28-89		WC/SEM	15.0	1	125		7.3
Ws	8-31-89		WC/SEM		30	140		7.9
Wsc	8-31-89		WC/SEM		10	150		7.9
Wsuz	9-12-89		WC/SEM	17.0	10	130		7.7
Wurs	1989		JBR	14.0	26			7.9

Code	Ca	Mg	Na	K	Cl	SO4	HCO3
AP1	13	12	828	12.2	447	367	940.
AP2	9.8	7.9	668	9.2	284	253	888.
APt	39.6	27.8	849	13.4	677	330	846.
B8	27.2	9.6	544	42	549	386	215.
B9	8.6	1.5	611	31	535	383	164.
BF							
Br							
Br	40	0.8	810	53	880	460	146.
DL1	56	19	40	9.1	14	25	323.
DL2	44	13	25	8	3	8	267
DL3	34	20	169	11	95	72	389.
F11	53	49	1810		2010	1200	224
F12	5	9	680	18	590	180	530
FS2tp	18.6	10.2	7.7	1.9	7	4	126.
FSf1	11	4	70	7.5	29	70	109.
FSf2	9	3	36	5.7	10	34	85.4
FSf3							
FSf4	12	1	33	5.5	12.2	35	76.3
FSf5	39.3	12.9	18	0.4	3.9	26	187
FSf6	18.7	22.4	31	8.1	5	49	181
GA	15	0.1	235	6	155	280	57
GL	2.2	0.02	66	0.57	14	57	74
GW	20	0.1	280	8	185	340	53
Hb1	12	1	92	16	20	40	228
Hb2	95	31	109	5.6	32	348	419.
Hb5	32	12	66	8.2	20	85	192
Hb6	29	9	53	5	14	50	193.
Hd1	15	9	13	1	2	2	122
Hd2	21	8	16	2.7	2	8	132.
Hd3	14	4	39	1.3	5	23	131.
Hd4	25	7	18	2.3	3	7	134.
Hd5	21	7	16	2.9	7	8	112.
Hd6	42	14	32	1.4	8	12	234.
Hd7	18	5	15	1.1	4	8	96.3
Hd8	25	4	14	0.9	6	6	126.
Hj1	20	5	16	4.3	3	15	106.
Hj2	30	12	47	4	14	58	184
Hs1	41	12	33	3	11	36	179.
Hs2	54	13	82	4.3	20	122	235.
Hs3	28	10	43	4.1	13	41	151.
Hs4	88	46	107	5.2	25	280	337.
Hs5	50	38	76	6.1	10	259	186.
NS2	16	4	196	13.5	84	173	278.
P1	59.3	2.7	1130	30	1122.	2	1103
P2	1516	382	17800	42	16476	9000	921
P3	150	820	16100	94	24300	350	950
P4	52	60	2830	51	3960	620	567
P5	35	41	752	33	1050	34	445.
Smf1	6	1	64	2.9	14	26	129.
Smf10							
Smf2	4	2	42	3	9	13	101.
Smf3	3	2	41	3.3	9	16	97.6
Smf4	4	0.9	57	3	12	22	118

Code	Ca	Mg	Na	K	Cl	SO4	HCO3
Smf5	5	1	47	2	11	20	99
Smf6	5	2	39	5	9	16	97
Smf7	3	2	43	3	10	17	97
Smf8							
Smf9							
SP1	49.4	14.8	481	35	514	387	115.
SP2	40.2	17	480	27	509	370	115.
Sps							
SPt	62	17.8	468	32	519	307	152.
Ssp							
St1	6	2	47	1.3	12	18	114.
St10	5.6	10	393		233	74	593
St2	4	1	49	2	9	14	112.
St3	31	10	32	4	4	9	224.
St4	11	9	144	6.8	93	11	250.
St5	34	42	100		154	5	295
St6	2	1	230		79	42	365
St7	16	14	1300	27	465	464	2086
St8	5.6	0.5	55		13	16	124
St9	27	14	120	10	43	28	369
Tbb2a	17.5	4.3	154	10.6	82	50	274.
Tbb2b	16.6	5.9	161	12.1	92	59	276.
Tbb2c	7.5	3.6	157	8.7	46	9	350.
Tcw1	16	6.7	12.7	4.8	5	5	106.
Tcw2	22.2	7.8	15.5	5.6	7	11	154.
Tcwt	22.2	8.6	17.7	6.5	7	7	148.
Tf							
Tf1	4	2.9	67		22	18	144
Tf2	7.4	1.9	67.8	6.4	17	14	143.
Tfd1	34.9	16.5	190	23.7	110.6	32	366
Tfd2	15.1	4.9	195	12.5	24	280	199
Tflt1	8.9	0.6	37.5	3.5	9	11	98.8
Tfpl	7.1	1.1	77	5.8	27	22	154.
Tfs	3	3	78		18	17	179
Tftw	5.6	1.3	71.4	5.3	24	25	136.
Tho	9.4	3.8	42.1	8.2	7	6	143.
Thq	3.5	1.4	50.3	7.8	8	7	129.
Thq	6.4	3.4	60		12	16	155
Tj	13.4	4.6	26.8	6.8	7	6	114.
Tn	36.9	2.8	182	4.5	32	364	97.6
Tw	21.6	1.7	130	4.8	22	192	111.
Vb	19.4	8.4	7.7	3.5	4	3	115.
Vc	14.3	5.8	5.4	4.7	4	3	87.8
Vcc	16.8	8.5	19.58		2.5	4	136.
Vcs	14.4	5.6	11.1	6.2	4	6	97.6
Vecs	ND	ND	ND	ND	2	4	ND
Vfsc							
Vhs	25.6	7	7.7	0.4	6	3	131.
Vls	22.9	11.7	7	2.4	5	3	142.
Vms	21.3	7.7	11.2	1.7	8	6	109.
Vmud	9.1	3.4	8.6	1	4	3	57.3
Vss	19	3.2	20.6	1.8	7	6	107.
Vst	16	4	24	2.6	4.7	6.2	107



Code	Ca	Mg	Na	K	Cl	SO4	HCO3
Vts	5.4	2.1	4	2.8	2	3	35.3
Vua	86	19	47	1	12	113	273
Vuj	66	10	38	1.2	16	69	204
Vus	14.2	6.3	6.7	2.1	4	3	89.0
Vws	18.2	5.7	11.3	3	6	5	103.
Wa	19.8	6.9	11.9	5.3	5	3	120.
Wns	30	9	23	5.7	6	6	168
Wrs	19.2	1.8	14	3.1	6	8	84.1
Ws	18.8	7.4	11.7	5.6	4	3	124.
Wsc	16.9	6	11.6	5.2	4	7	106.
Wsuz	26.6	9.7	12.6	4.9	7	4	158.
Wurs	33	11	13	4.7	5	6	160

Code	SiO2	Arsen.	Boron	F	Iron	Mn	NO3
AP1		0.029	3.8	3.1	0.11	0.07	0.5
AP2		0.052	3.4	3.2	0.15	0.05	0.5
APt		0.012	4.6	2.2	0.04	0.14	-0.1
B8							
B9							
BF							
Br							
Br							
DL1			0.3	0.4			4.2
DL2	54		1				
DL3			0.7	1.9			2.6
F11							
F12	14		1.8	1.7	0.1	0.39	12
FS2tp							
FSf1			1	2.4			5.8
FSf2			0.2	0.5			2.1
FSf3							
FSf4	8		0.4	0.5			-1
FSf5	8		0.2	0.3	-0.1	0.04	2.7
FSf6	12		0.2	0.3	-0.1	0.01	3.5
GA	98		3.8				
GL	36			2			
GW	125		5.6				
Hb1			0.2				5.8
Hb2		0.05	0	0.5	0	1.7	10
Hb5	59		0.2	0.4			0.3
Hb6		0.03	0.2		0.18	0.26	0
Hd1			0				2.9
Hd2			0				7
Hd3				0.3			5.8
Hd4			0.02	0.1			
Hd5			0.06	0.2			
Hd6			0	0.2			14
Hd7			0	0.2			6.6
Hd8			0				13
Hj1			0.05	0.1			
Hj2			0.2				1.4
Hs1			0.1	0.1			14
Hs2			0.4	1			8.8
Hs3		0.009	0.1	0.5	-0.1	-0.0	31
Hs4			0.3	0.4			1.8
Hs5			0.2	1			1.6
NS2		0.01	1.4	0.3	0.47	0.03	10.1
P1	10.5		1.6	0.7	0.5	0.03	2.7
P2	5.3		10	1.7	0.2	0.22	3.1
P3	18		16	0.4	9	0.5	1
P4	20		1.9	0.2	-0.1	-0.0	8
P5			1.7	0.5			63
Smf1		0	0.1	0.2	0	0	2.2
Smf10							
Smf2			0.1	0			0.4
Smf3			0.1	0.4			0.2
Smf4	29		1				

Code	SiO2	Arsen.	Boron	F	Iron	Mn	NO3
Smf5	32		1				
Smf6	38		1				
Smf7	34		1				
Smf8							
Smf9							
SP1		0.002	2.6	0.7	0.12	-0.0	0.3
SP2		0.002	2.1	0.7	0.05	0.03	1
Sps							
SPt		0.012	2.5	0.6	0.1	-0.0	1.6
Ssp							
St1			0	0			0.4
St10							
St2			0.2	0.5			4.4
St3			0.2	0.3			4.2
St4			0.3	0.3			26
St5		0.168	0.4	0.3	0.1	0.00	0.4
St6		0.168	1.4	1.6	0.1	0.00	0.7
St7		0.394	4.7	6.2	0.7	0.02	11
St8							
St9	63		1				
Tbb2a		0.003	1.2	0.1	-0.0	-0.0	-0.1
Tbb2b		0.002	1.1	0.1	0.19	-0.0	1
Tbb2c		0.003	0.5	1	0.02	-0.0	-0.1
Tcw1		0.002	0.1	0.2	-0.0	-0.0	2.4
Tcw2		0.002	-0.1	0.1	0.28	0.05	1.6
Tcwt		0.001	-0.1	0.1	0.12	0.02	4.8
Tf							
Tf1							5.2
Tf2							5
Tfd1	5.3		0.6	1.7	0.2	0.03	3.1
Tfd2			0.9				
Tfl1t1		0.003	0.1	0.2	0.13	-0.0	1
Tfpl		0.01	0.4		-0.0	-0.0	4
Tfs							
Tftw		0.025	0.4	0.4	0.02	-0.0	0.9
Tho							3
Thq							4.4
Thq							
Tj							4.5
Tn							1.8
Tw							0.3
Vb							
Vc							
Vcc		0		0.21	0.11	0.03	0.2
Vcs							
Vecs							
Vfsc							
Vhs							
Vls							
Vms							
Vmud							
Vss							
Vst							-0.1

Code	SiO2	Arsen.	Boron	F	Iron	Mn	NO3
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Vts							
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Vua					-0.1		
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Vuj					-0.1		
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Vus							
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Vws							
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Wa							
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Wns							
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Wrs							
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Ws							
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Wsc							
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Wsuz							
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Wurs							
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Code O-18 Deuterium sum TDS epm-balance

AP1			2619.8	3.1
AP2			2120.1	4.5
APt	-13.6	-104.0	2783.5	2.1
B8	-13.0	-112.0	1773.7	-0.3
B9	-13.3	-114.0	1734.8	4.0
BF	-12.3	-103		
Br	-13.7	-115.5	0.0	ERR
Br	-14.2	-114.0	2390.2	2.5
DL1			486.4	0.9
DL2		-101.6	368.0	-0.8
DL3			790.2	1.9
F11			5346.0	0.0
F12	-13.7	-108.0	2012.0	3.2
FS2tp	-14.7	-115.0	176.3	-4.6
FSf1			301.3	0.5
FSf2			183.1	0.4
FSf3	-15.3	-114.0	0.0	ERR
FSf4	-15.3	-116.0	175.0	-1.5
FSf5	-14.9	-114.0	287.5	1.3
FSf6	-15.2	-116.0	315.2	2.4
GA	-14.6	-119	748.1	-0.0
GL	-15.3	-118.1	213.8	3.5
GW		-121.2	886.1	0.8
Hb1			409.0	-0.4
Hb2	-14.1	-112.5	1039.9	
Hb5			415.2	1.6
Hb6	-13.9	-109.0	354.0	0.1
Hd1			164.0	-0.4
Hd2			190.7	1.4
Hd3			218.1	-0.4
Hd4			196.5	4.6
Hd5			174.1	4.1
Hd6			343.6	4.0
Hd7			147.5	3.4
Hd8			182.8	-3.6
Hj1			169.4	1.8
Hj2			349.0	0.1
Hs1			315.3	6.4
Hs2			530.8	3.3
Hs3			290.4	6.3
Hs4			889.1	3.6
Hs5			625.8	2.0
NS2			764.7	-2.6
P1	-13.6	-109.5	3449.6	3.2
P2	-9.9	-96.5	46137.0	13.9
P3	-10.9	-49.0	42764.0	4.7
P4			8140.0	-0.7
P5			2390.3	1.4
Smf1			243.2	2.9
Smf10	-14.1	-111		
Smf2			174.3	1.9
Smf3			171.9	-0.1
Smf4			216.9	1.8

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Smf5	-113.3	185.0	1.6
Smf6	-113.3	173.0	1.4
Smf7		175.0	0.8
Smf8	-14.2 -112.5		
Smf9	-15 -113		
SP1		1597.1	2.1
SP2		1559.1	2.1
Sps	-14.6 -111.0	0.0	ERR
SPT	-13.6 -106.0	1558.3	4.5
Ssp	-14.0 -109.0	0.0	ERR
St1		201.0	-1.0
St10		1308.6	1.0
St2		191.2	1.6
St3		314.5	-1.5
St4		524.9	5.3
St5		630.0	1.2
St6		719.0	5.7
St7		4372.2	1.9
St8		214.1	-0.4
St9		611.0	0.8
Tbb2a		592.9	2.1
Tbb2b		623.5	1.6
Tbb2c	-14.9 -110.0	581.9	3.3
Tcw1		156.3	1.0
Tcw2		224.0	-7.2
Tcwt	-14.9 -112.0	217.8	-0.6
Tf	-14.9 -113.5	0.0	ERR
Tf1		257.9	-0.0
Tf2	-14.0 -114.0	258.5	7.5
Tfd1	-15.0 -119.0	773.7	10.0
Tfd2		730.5	1.0
Tflt1	-15.4 -114	169.3	2.6
Tfp1		294.9	2.4
Tfs		298.0	-0.1
Tftw	-14.9 -113.0	269.2	2.7
Tho	-14.4 -112.8	220.5	2.6
Thq	-14.2 -113.0	207.3	3.6
Thq		252.8	-0.0
Tj	-13.7 -112.0	179.3	4.0
Tn	-14.6 -119.0	719.8	0.1
Tw	-14.3 -118.0	483.1	4.2
Vb	-15.1 -115.0	161.9	0.2
Vc	-14.0 -112.5	125.0	-2.2
Vcc	-14.4 -110.4	187.8	-0.0
Vcs	-14.4 -114.0	144.9	-0.5
Vecs	-14.4 -111.8	6.0	
Vfsc	-14.1 -109.0	0.0	ERR
Vhs	-14.7 -113.0	181.5	-4.2
Vls	-14.8 -115.0	194.7	-1.4
Vms	-14.4 -114.0	165.7	1.8
Vmud	-12.2 -108.0	86.4	0.8
Vss	-13.4 -112.0	165.0	1.7
Vst		164.5	5.2

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Vts	-14.4	-113.3	54.7	-0.8
Vua	-12.6	-104.0	551.0	5.0
Vuj	-12.8	-105.0	404.2	5.2
Vus	-14.4	-112.3	125.4	-2.0
Vws	-14.4	-114.0	152.9	-0.7
Wa	-14.4	-116.0	172.7	0.6
Wns	-14.9	-116.0	247.7	5.2
Wrs	-14.7	-118.0	136.3	2.2
Ws	-14.4	-113.0	174.9	-0.4
Wsc	-14.2	-117.0	156.8	-0.6
Wsuz	-15.2	-118.0	223.4	-1.4
Wurs	-14.8	-116.5	232.7	5.7

