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MACKAY MINERALS RESEARCH INSTITUTE
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REPORT SUBMITTED TO
SOUTHLAND ROYALTY COMPANY
1000 FORT WORTH CLUB TOWER
FORT WORTH, TEXAS 76102

WORK PERFORMED AS SUBCONTRACTORS UNDER DOE
CONTRACT No. DE-AC08-79ET27006

STRUCTURAL-TECTONIC ANALYSIS

Third Quarter Report

I. Work Accomplished

The low-sun-angle aerial photography program in Dixie Valley is complete, with all images processed, printed, indexed and interpreted. Structural-tectonic features delineated on this photography were compiled on 1:24,000 scale quadrangle maps that were integrated to produce a 1:62,500 scale base map of the Dixie Valley study area (Plate ST-1). Other imagery formats, including LANDSAT and NASA underflight imagery, have been interpreted and will be evaluated with respect to the composite structural-tectonic features map. The major structural elements defined by geophysical methods (Senturion Sciences, 1977, 1978; Koenig and others, 1976; Keplinger and Associates, 1978) have been compared to the structural-tectonic features delineated by their surface expression on the low-sun-angle aerial photography. Section V. Results presents a summary of the comparison.

II. Problems Encountered

No significant problems concerning the structural-tectonic analysis were encountered during the previous quarter of activity.

III. Projected Activities

The structural-tectonic analysis will proceed in the fourth quarter with field verification of the imagery analysis and resultant composite structural-tectonic features map. Field verification may include: 1) field examination of features of questionable origin, 2) fault scarp morphology as outlined in the second quarter report, 3) minor trenching and 4) limited gravity profiling to define fault limits.

IV. Problems Which May Be Encountered

The program for the next quarter as described above may be hindered by lack of field time and/or inclement weather. The trenching program may be abandoned if safety becomes a controlling factor. The gravity profiling will be dependent on the availability of equipment.

V. Results

Structural-tectonic features delineated in Dixie Valley (Plate ST-1) correspond with subsurface structures delineated by gravity, aeromagnetic and other geophysical surveys (Fig. ST-1) (Meister, 1967; Stewart, 1971; Senturion Sciences, 1977, 1978; Koenig and others, 1976; Keplinger and Assoc., 1978; Trexler and others, 1978). These include:

1. Old Stillwater fault. This zone bounds the Stillwater Range to the southeast and trends approximately N36E from the southern edge of the study area near Dixie Meadows into Pleasant Valley on the northern edge of the study area; this is the range-front fault of the Stillwater Range. The Old Stillwater fault is marked at the surface by scarps in alluvium and in bedrock, as well as by alignments of springs and fumaroles, vegetation and tonal contrasts, grabens, breccia zones, and slumps. Field measurements indicating a general S60E dip along the scarp are consistent with the literature (Slemmons, 1957; Page, 1965; Burke, 1967; Meister, 1967; Stewart, 1971). Site-specific evidence for a S60E dip includes: 1) measurable fault scarps on the surface, 2) surface expression of normal faulting including back-facing grabens and first order Riedel shears indicative of secondary lateral offset, and 3) fault plane solutions for a N55W tensional (sigma 1) direction in northern Dixie Valley (Ryall and Malone, 1971). Regional studies also indicate tension in a general N55W direction (Thompson and Burke, 1973; Wright, 1976). Interpretation of geophysical data by Senturion Sciences (1977, 1978) indicating a N55W dip for this fault is not compatible with this evidence.

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

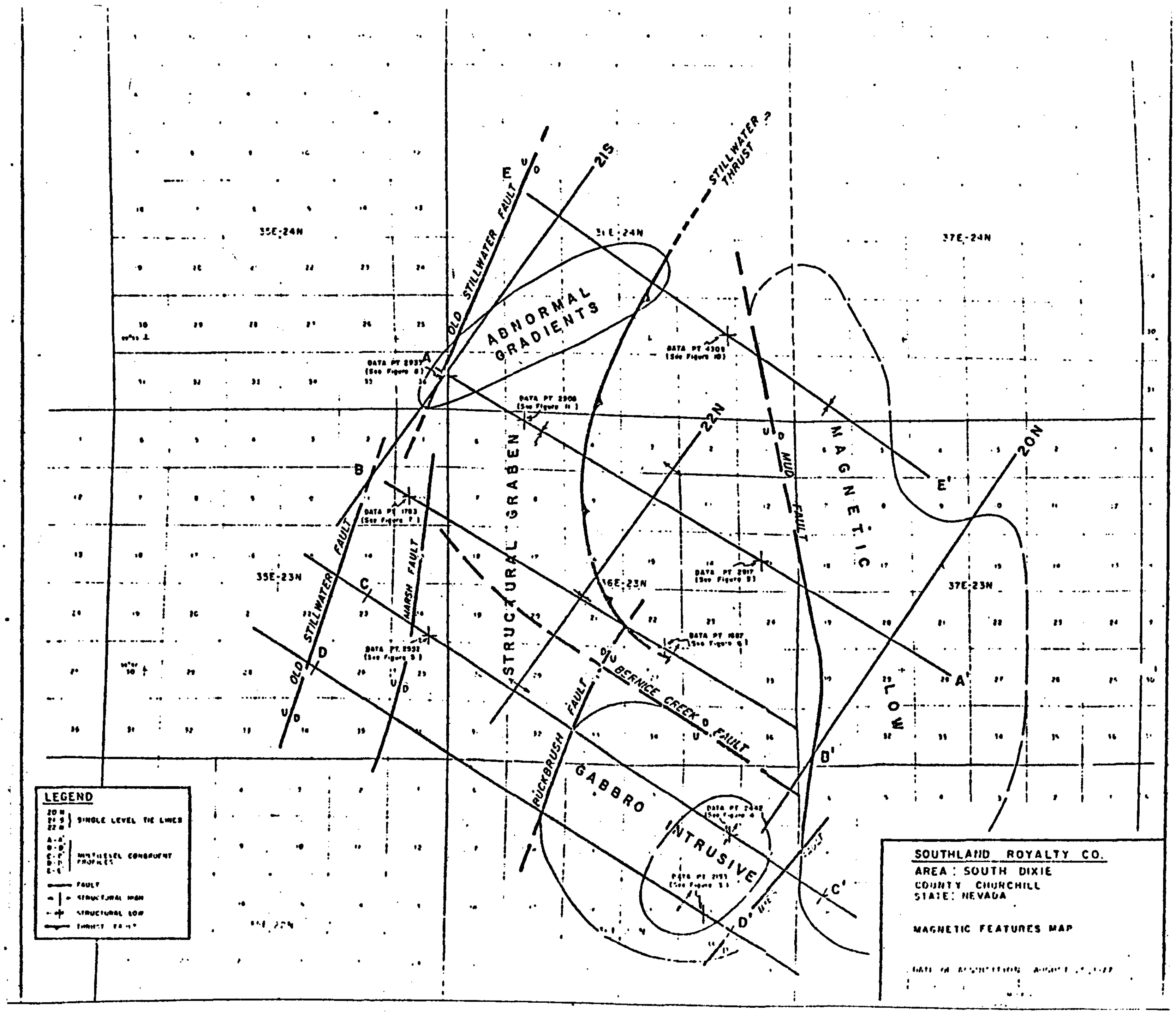

Figure ST-1. Aeromagnetic features of Dixie Valley (from Senturion Sciences,1978)

3. Buckbrush fault. The Buckbrush fault is an arcuate fault trending approximately N30E from the southern edge of the study area to the Bernice fault, where it splays into two branches. The western splay joins the Old Stillwater fault in Section 14, T24N, R36E. The eastern splay intersects the Old Stillwater fault in Sections 28 and 29, T25N, R37E near the Senator fumaroles. Surface expression is marked by scarps, spring alignments and vegetational and tonal contrasts. The northwest block is down-dropped in the southern portion of the zone, but relative movement is not discernible in the northern portion.

4. Mud fault. The Mud fault trends N36E from Dixie Settlement to Section 4, T23N, R35E, then follows a N15E trend to Sou Hills. The surface expression near Sou Hills suggests several possible interpretations: 1) the fault joins the frontal fault system on the east side of Pleasant Valley, 2) the fault joins the Old Stillwater fault system and continues along the west side of Pleasant Valley, or 3) the fault is related to the frontal system on the east side of the Tobin Range.

The southeast splay of this fault mapped by Senturion Sciences (1977, 1978) as the Mud fault shows surface expression near the mouth of Deep Canyon and trends approximately N25E to intersection with the main zone in Section 30, T23N, R37E. Alternatively, this splay could intersect the Shoshone fault, with the Mud fault being a separate and distinct structure extending northeast to Sou Hills. In either interpretation, the blocks to the northwest are down-dropped.

5. Dyer fault. The Dyer fault enters the southern edge of the study area near the mouth of Meadow Spring Canyon and trends N36E along the front of the Clan Alpine Mountains into Jersey Valley; this may be the frontal fault for the Clan Alpines. A splay of this fault appears to cross Dixie Valley just north of Hole-in-the-Wall and joins the frontal fault system on the west side of the Tobin Range in Pleasant Valley. In Dixie Valley the northwest block is down-dropped along this fault.

6. Bernice fault. The Bernice fault is a main crosscutting feature of Dixie Valley. Surface expression of vegetation and tonal contrasts and stream alignments indicate the zone extends along an arcuate trace northwest from Bob

Canyon toward the Dixie Site in the Stillwater Range. A splay of this zone may extend northward and join the Marsh fault near Section 17, T23N, R36E. Relative movement of this zone is uncertain as surface expression is subtle.

7. Stillwater thrust. No surface expression of a postulated thrust fault in the center of the study can be delineated.

8. Unnamed fault at Mississippi Canyon. Another crosscutting fault zone extends from Mississippi Canyon approximately S40E across Dixie Valley. Surface expression includes low scarps, vegetation and stream alignments and tonal contrasts. These features are subtle, but indicate the southwest block is downdropped.

9. Unnamed fault at Dixie Meadows. A major fault zone trends N36E from the southern edge of the study area across Dixie Meadows, parallel to and between the Old Stillwater and Marsh faults. This fault zone corresponds to the valley branch fault of Trexler and others (1978). The zone continues in an arcuate path concave to the northwest and merges with the Old Stillwater fault in Sections 35 and 36, T24N, R35E, then re-emerges from the range-front fault along an arcuate path to a second intersection with the Old Stillwater fault in Section 14, T24N, R36E. The southwest block is downdropped as shown by numerous slumps, assymetric grabens, and eastward-facing scarps. Spring and vegetation alignments are also present.

10. Unnamed fault northwest of Shoshone Creek. An nunamed fault northwest of and subparallel to the Dyer fault northwest of Shoshone Creek (referred to as Shoshone fault in 4 above) has a general strike of N30E and is downdropped to the northwest. Surface expression is marked by scarps and vegetation contrasts. Hyder Hot Springs may be associated with this structure.

11. Pleasant Valley fault. The Pleasant Valley fault is the frontal fault on the west side of the Tobin Range; the fault is marked by a scarp created during the 1915 Pleasant Valley earthquake. This zone crosses the Sou Hills and apparently ends in the northern portion of Dixie Valley.

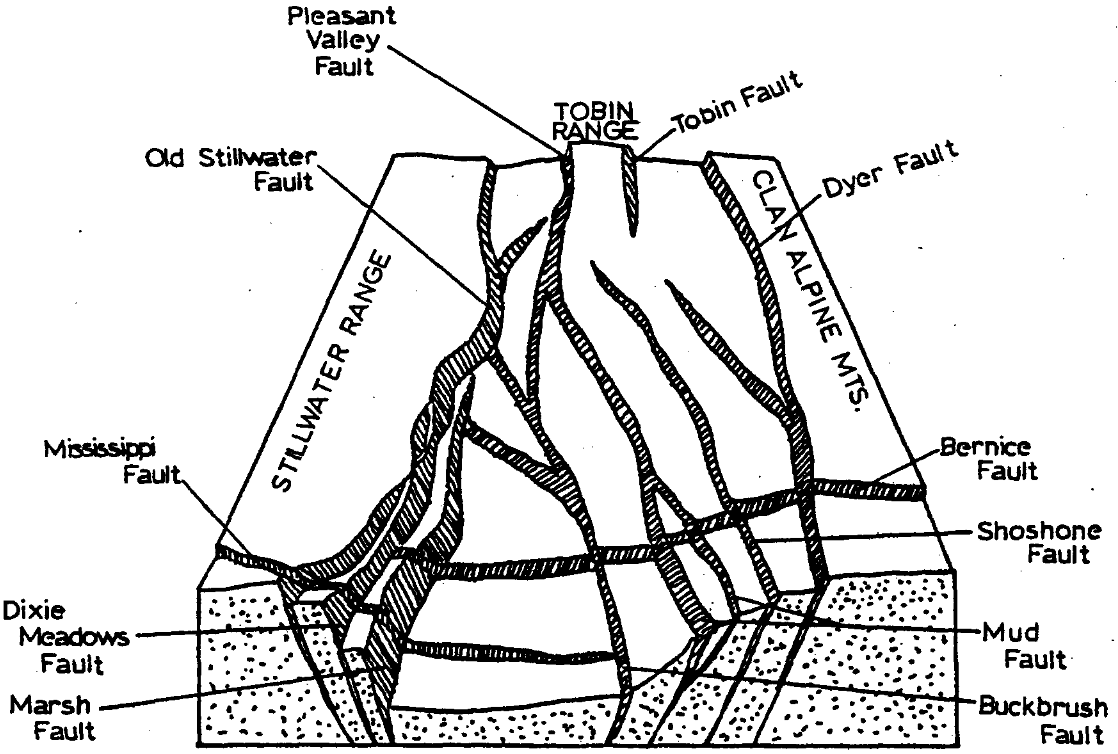

Figure ST-2. Block diagram of bedrock surfaces in central and northern Dixie Valley with alluvium removed and eroded bedrock restored. (Modified by Whitney, this report, after Burke, 1967)

12. Unnamed fault on the east side of the Tobin Range. The frontal fault zone on the east side of the Tobin Range extends southwest into the study area and apparently dies out in the north end of Dixie Valley.

It is stressed that the foregoing fault zone interpretations lack sufficient field verification and are presented only as a working base for future studies in the Dixie Valley area. Figure ST-2 reflects a modification of the model by Burke (1967) for the bedrock surfaces of central and northern Dixie Valley based upon the above listed comparisons of surface and subsurface structural elements. The fourth quarter activity will be focussed on verifying these interpretations and synthesizing these data with field information to develop a model and structural-tectonic history of Dixie Valley that is compatible with all available data derived from the various phases of present contract work.

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HYDROLOGY AND HYDROGEOCHEMISTRY

THIRD QUARTER REPORT

I. WORK ACCOMPLISHED

Approximately 25 man-days were spent in the field collecting water samples from springs and wells. Spring samples were obtained from canyons in the Stillwater Range as well as from the valley bottom. Water samples were also obtained from the Carson Sink, a valley on the west side of the Stillwater Range. The chemistry of these samples will be compared to those of Dixie Valley.

We were also fortunate to obtain fluid samples from both DF 45-14 and DF 66-21. The sample from the latter well was not particularly good. The initial sample from DF 45-14 was also somewhat contaminated; however, just prior to the writing of this report, a good sample was obtained from this well and is currently being analyzed. We also received the oxygen-18 analyses from the University of Arizona and the in-house tritium analyses.

In addition to the collection and chemical analysis of additional samples, some interpretations of the analytical results were performed. These are presented in a later section.

II. PROBLEMS ENCOUNTERED

No particular problems have been encountered although we are still awaiting the results of the deuterium analyses.

III. PROJECTED ACTIVITIES

Most of the fourth quarter activities will consist of analyzing, interpreting and synthesizing the accumulated data. As has been the case throughout the study, heavy reliance will be placed on the chemical and isotopic analyses to delineate the hydrologic-hydrothermal system in the study area. Pre-existing data will also be used in preparing our model of the system. The results of the shallow temperature survey should also prove useful.

Some limited re-sampling will be conducted. We hope to obtain a few samples from the Buckbrush Springs area, since this region is not well represented in our sampling network. We also hope to obtain some samples for carbon-14 and carbon-13/carbon-12 analyses, which will enable us to place some kind of age limits on the water. We also should be able to obtain an uncontaminated sample from DF 66-21.

IV. POTENTIAL PROBLEMS

We do not foresee any insurmountable problems arising during the fourth quarter.

V. DATA ANALYSIS

Dixie Hot Springs and Hyder Hot Springs have been sampled extensively during this quarter. Samples were taken from spring orifices that seemed to represent different groups based on temperature and specific conductance (SpC). At Dixie Hot Springs, SpC values varied from 1800 to 4100 micromhos/cm and temperatures ranged from 48°C to 76°C. An earlier, less extensive survey at Dixie Hot Springs revealed SpC values as low as 1100 micromhos/cm. This would imply that, at least with respect to SpC (i.e., total dissolved solids), there are temporal variations in at least one spring. This could be the result of mixing with recharge water. Variations in SpC and temperature were less drastic in Hyder Hot Springs.

The highest temperature encountered in Sou, Dixie and Hyder Hot Springs was about 76°C, which indicates a common thermal regime. The chemical analyses have not yet been interpreted to their fullest extent. However, the variations in spring chemistry seem to reflect various degrees of mixing.

Additional samples were collected from mountain springs in the Stillwater Range in hopes of finding thermal springs that could be the source of high TDS waters in the mountain streams. We now believe that the most probable cause of high TDS waters in the Stillwater Range are sulfide deposits.

Qualitative estimates of the gas content of DF 45-14 were made. The amount of CO_2 is no more than 0.5 percent, while SO_2 is more than 50 parts per million (ppm) and H_2S is more than 20 ppm. Tests for NH_3 and hydrocarbons were negative.

The results from wells DF 45-14 and DF 66-21 have been integrated into the overall analyses. To a certain extent the samples from these wells are similar to most hot spring waters in the valley (e.g., cation ratios). But on a trilinear diagram the samples from DF 66-21 exhibit a significant shift towards (HCO_3 + CO_3) at the expense of SO_4, whereas the samples from DF 45-14 are shifted towards Cl at the expense of the carbonates.

Most of the thermal waters were plotted on log activity diagrams (Figures 1 and 2). The appropriate phase boundaries were added for a temperature of 260°C. Unfortunately, thermodynamic data are incomplete, so that these diagrams have to suffice for now. All samples seem to indicate a gradual decrease in H^+ activity, reflecting a loss in CO_2. The diagrams seem to suggest a significant role for carbonic acid (i.e., CO_2) in silicate dissolution. Thus, a much lower pH than that encountered at the surface must be present at depth. The pH at depth estimated from Na and K molality is about 6.2 in DF 45-14 and about 6.3 in DF 66-21. This correlates with a pH of 6.05 measured in a strongly degassing well at Sou Hot Springs.

Some calculated log P_{CO_2} (calculated at sampling temperature) were entered into one of the diagrams (Figure 2) in parentheses to indicate the gradual decrease of CO_2. Boiling at depth must not be necessarily inferred from this, since CO_2 separation might be due to sampling techniques at the well heads and the spatial distribution of spring orifices.

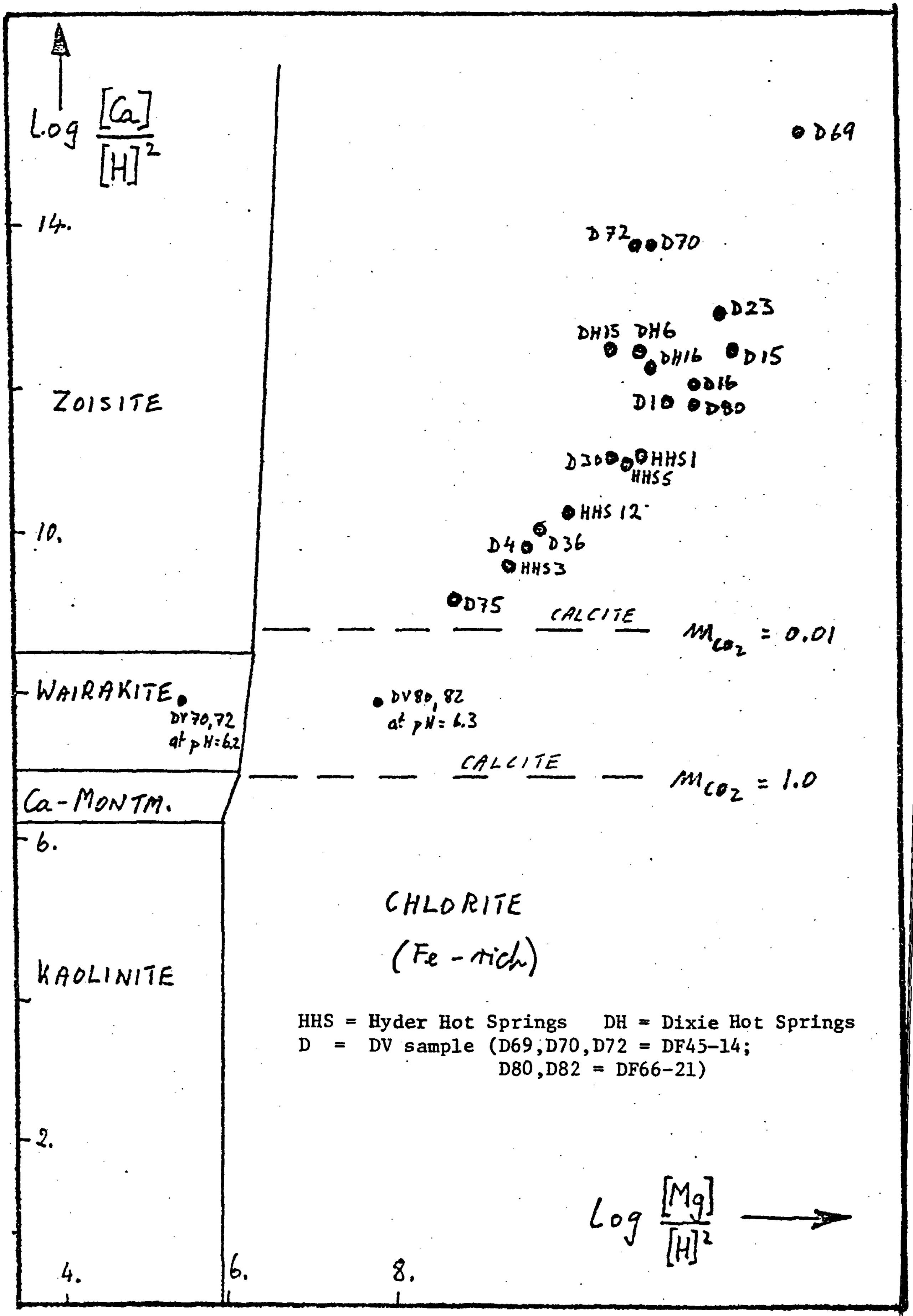


Figure 1. Log $[Ca]/[H]^2$ vs. log $[Mg]/[H]^2$ for selected thermal waters.

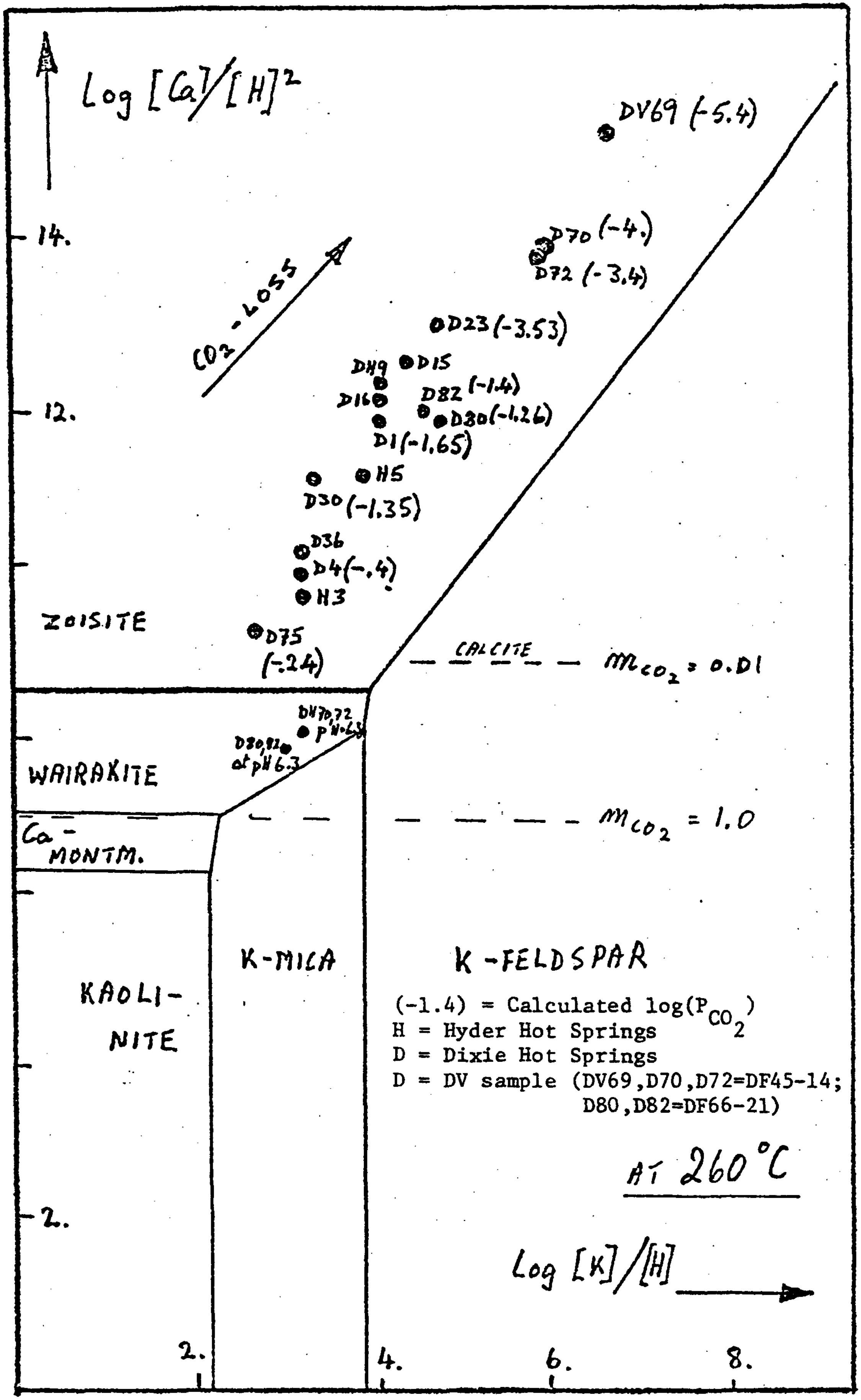


Figure 2. Log $[Ca]/[H]^2$ vs. log [K]/[H] for selected thermal waters.

The phase boundary for calcite precipitation at a CO_2 molality of 0.01 was also entered. According to the diagrams, the waters should be in equilibrium with zoisite and chlorite and calcite should precipitate at the sampling pH. Zoisite and chlorite were found at depth. However, no evidence of extensive calcite deposition was found at depth, which suggests that the CO_2 loss does not occur at depth. The CO_2 molality was also estimated for both deep wells. For well DF 45-14 at a pH of 6.2 and a temperature of 209°C, the log P_{CO_2} is about 0.9, while for well DF 66-21 at a temperature of 190°C the log P_{CO_2} is 1.4. Both of these numbers are much too high, indicating errors in the pH values used. The shift caused by the lower pH was entered into the diagrams.

An attempt will be made to modify the diagrams (and others) for the encountered temperatures of about 200°C. Also, the effect of boric acid and other substances on pH has to be considered.

A plot of log SiO_2 versus log P_{CO_2} (Figure 3) indicates various systems of silicate dissolution by carbonic acid. CO_2 seems to decrease with increased silicate dissolution. The high temperature samples apparently seem to dissolve along a higher SiO_2/CO_2 ratio than the stream waters and shallow cold wells. At this stage the plot is interpreted as follows: there are various distinct ground-water systems, each one of which has a limited supply of CO_2 at a particular temperature regime. Dissolution of silicates occurs until CO_2 is consumed (i.e., pH is too alkaline). The non-thermal

Buckbrush Seeps	B
Sou Hot Springs	+
Dixie Meadows	×
Irrigation wells, surface waters etc.	•
Hyder Hot Springs	H
SR-2A	o
Springs in NE of Dixie Valley, with T no higher than 50°C	N
Sample from neighbouring valley (Carson Sink)	CS6
Mountain Spring in SE	L102
Southland deep wells	SW

SW1 = 45-14; SW2=66-21

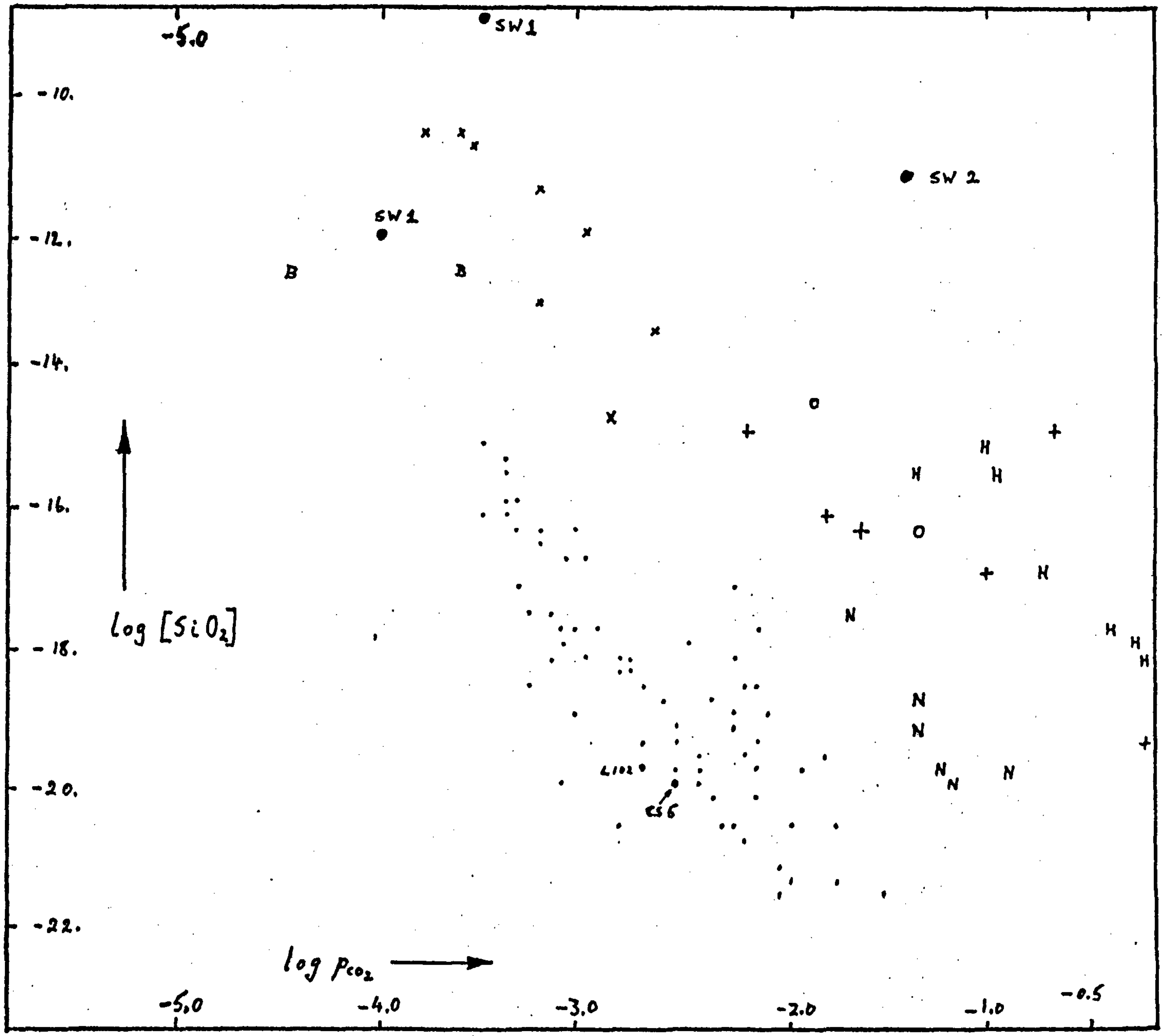


Figure 3. Log $[SiO_2]$ vs. log $[P_{CO_2}]$ for various waters.

systems seem to start with less CO_2. The spatial distribution of the distinctive flow systems is hard to infer. All that can be said is that the level of silica concentrations within a particular system as related to its CO_2 level could indicate its relationship to a deep-seated hot-water reservoir, assuming that CO_2 is added at various depths.

The schematic model of subsurface flow has to be modified somewhat, although we cannot be more specific at this stage. However, the occurrence of a steam dominated reservoir cannot be verified yet, as the results of the deep drilling program have shown. The chloride content apparently will increase with depth, although the analyses from DF 66-21 are not yet available.

Table 1 shows the calculated temperatures for wells DF 45-14 and DF 66-21. The silica content for DF 45-14 is slightly above the solubility of amorphous silica at the sampling temperature, which suggests precipitation of silica within the well. If this is the case, the silica temperature might even be higher.

The oxygen-18 data have provided some information on the hydrologic-hydrothermal system, although in the absence of the deuterium results, a complete interpretation is not possible. Table 2 shows some of the isotope data for selected waters in Dixie Valley. These waters represent both thermal and non-thermal waters in the valley. With exception of well DF 45-14, there seems to be no clear trend, i.e., thermal waters do not seem to exhibit the oxygen shift (enrichment)

Table 1

Calculated temperatures for wells DF 45-14 and DF 66-21

Sample	Na-K-Ca Temperature (°C)	Quartz Temperature (no steam loss) (°C)	Quartz Temperature (with steam loss) (°C)
DV 69	203	-	-
DV 70	201	-	-
DV 72	206	209	191
DV 80	138	-	-
DV 81	132	183	171
DV 82	139	190	177

DV 69, DV 70, DV 72 = DF 45-14

DV 80, DV 81, DV 82 = DF 66-21

Table 2

Oxygen isotope data for selected Dixie Valley waters

Location	Sample #	$\delta^{18}O$ ($^{o}/oo$ SMOW)
Seven Devils Hot Springs (crater)	DV 1	-16.0
Seven Devils Hot Springs (spring feeding ditch)	DV 2	-15.7
Hyder Hot Springs	DV 4	-16.0
McCoy Hot Springs	DV 11	-16.4
Flowing Well	DV 13	-16.8
White Rock Canyon Creek	DV 22	-15.1
Dixie Hot Springs	DV 24	-16.3
Well SR2A	DV 30	-15.3/-15.6
Buckbrush Pond	DV 41	-5.7/-5.9
Buckbrush Spring	DV 42	-17.1
Mountain Spring	DV 45	-15.7
DF 45-14	DV 69	-11.8/-11.6
DF 45-14	DV 70	-13.8
DF 45-14	DV 71	-14.0
DF 45-14	DV 72 A	-13.7

SMOW = Standard Mean Ocean Water

characteristic of some other geothermal systems. The DF 45-14 samples do show enrichment, although there is variability among these samples. Since these well samples were the first ones taken from DF 45-14, sample contamination could be a factor.

The fact that there is little or no oxygen shift is not necessarily disturbing. This could indicate that the hydrothermal system is old, through which there is extensive fluid flow. In such a system, the isotopic composition of the rock has adjusted to equilibrium with the recharge water. However, this is based on the assumption that virtually all the thermal water is local meteoric water which has been heated on its journey through the system. At this stage we are not prepared to say that such a model is valid for Dixie Valley. We will also have to see if the deuterium data show a shift, which is characteristic of some geothermal areas.

One area of note with regard to oxygen-18 is Buckbrush Pond, which shows a very large enrichment. We attribute this to evaporation, which can account for enrichment, since this sample was taken from standing water. We have re-sampled this area just in case there is something unusual in this area, although we are pretty certain evaporation is the culprit.

We at first thought that with regard to oxygen-18, our sampling procedures were suspect. However, after reviewing and checking our procedures, we have found that they are in accord with standard methods.

Tritium results indicate that most of the waters analyzed are probably older than 50 years or so - no great surprise. The analysis for Seven Devils Hot Springs (crater) has a high value of about 300 tritium units (TU). This could be due to evaporation or other effects. The tritium does indicate areas where it might be fruitful to date the water with carbon-14. However, it should be noted that carbon-14 dating of thermal waters is more difficult than for non-thermal waters. Problems arise with estimating the loss of carbon dioxide (and carbon-14 activity) through calcite precipitation or by boiling of the waters. The addition of "dead" carbon can also be difficult to quantify, as can be the input term of carbon-14.

SHALLOW HOLE TEMPERATURE SURVEY

THIRD QUARTER REPORT

I. WORK ACCOMPLISHED

Monthly temperature surveys were conducted in July, August and September. These data have been plotted on base maps which have been supplied to Southland Royalty Company. Six months' worth of data have now been collected; no more will be obtained under the current contract.

Relocation of 20 of the last 72 holes emplaced was accomplished. These holes are now located on 7.5 minute quadrangle maps.

Some interpretations of the data were performed. These interpretations were based solely on visual examination of the data. Although we anticipated performing substantial amounts of data interpretation and analysis during the third quarter, we decided to delay these tasks until all the monthly surveys had been completed. This was done in the interests of saving time and money.

II. PROBLEMS ENCOUNTERED

No significant problems were encountered during the past three months, although it has become increasingly difficult to locate some of the holes. For example, during the September survey only about 200 holes were sampled.

III. PROJECTED ACTIVITIES

During the last quarter, the following will be accomplished:

1. Analysis and interpretation of the monthly temperature data with respect to the delineation of the near-surface hydrologic-hydrothermal system.
2. Comparison of the shallow hole temperature data with those from the temperature gradient holes.
3. Comparison of the entire suite of data with individual monthly surveys to evaluate the reliability of single survey data.
4. A statement on the validity of the shallow-depth temperature survey as a rapid and inexpensive geothermal exploration technique.

IV. POTENTIAL PROBLEMS

We do not foresee any problems arising during the last quarter.

V. DATA ANALYSIS

A visual examination of the data reveals that, to no surprise, the strongest thermal activity lies in the vicinity of Dixie Hot Springs, where a maximum temperature of almost 108°F was measured in August. This was the highest temperature measured during the entire survey.

The single line of holes running across the valley from Shoshone Creek to the west side of the valley also showed an interesting trend. The easternmost end of this line (hole 6) showed higher temperatures than the far western end (hole 5) of the line. This difference was not apparent during the first survey (April), but became evident during later surveys. Temperature differences between these two holes ranged from a low of 1°F in April (Figure 1) to a high of 12.4°F in July (Figure 2). Between these two holes there was no clear trend.

Interesting temperature trends were also noted in the isolated grid of holes located in the Bernice Creek-Dyer Flat area. The April survey revealed a temperature difference of just 5.1°F between the hottest (hole 98) and coldest (hole 91) locations (see Figure 3). In May, June and July the temperature differences between these two holes were greater than 10°F (see Figure 4 for the May temperatures) but by September the difference dropped down to below 6°F. In fact, during April and September the temperatures in this area showed less variation than during May, June, July and August. Whether or not these variations are significant will be determined during the fourth quarter.

The grid encircling Hyder Hot Springs also produced some interesting results. The holes closest to the springs generally had lower temperatures than those farther away. These differences were pronounced in June and July (maximum differences greater than 16°F) than in August and September (maximum differences on the order of

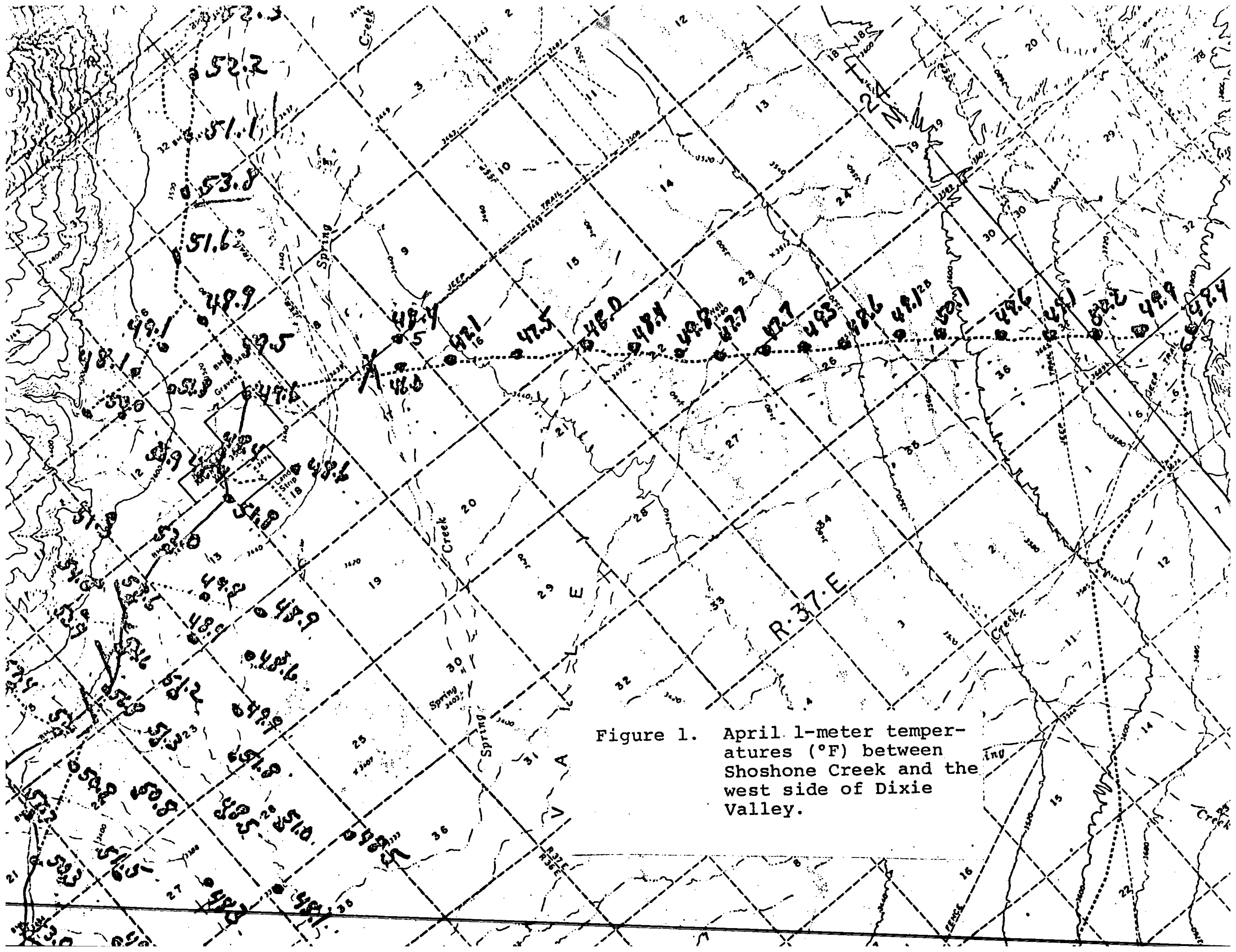


Figure 1. April 1-meter temperatures (°F) between Shoshone Creek and the west side of Dixie Valley.

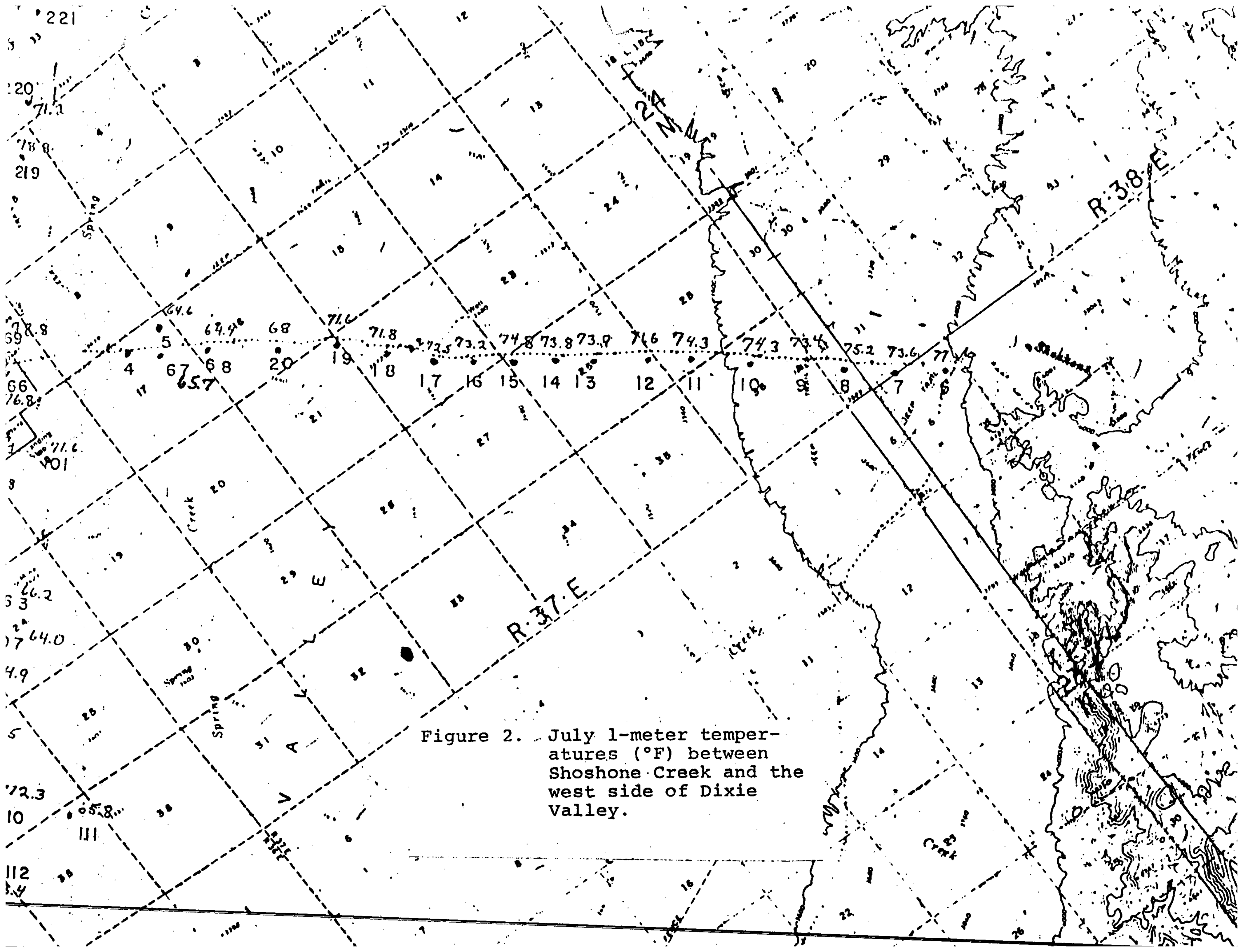


Figure 2. July 1-meter temperatures (°F) between Shoshone Creek and the west side of Dixie Valley.

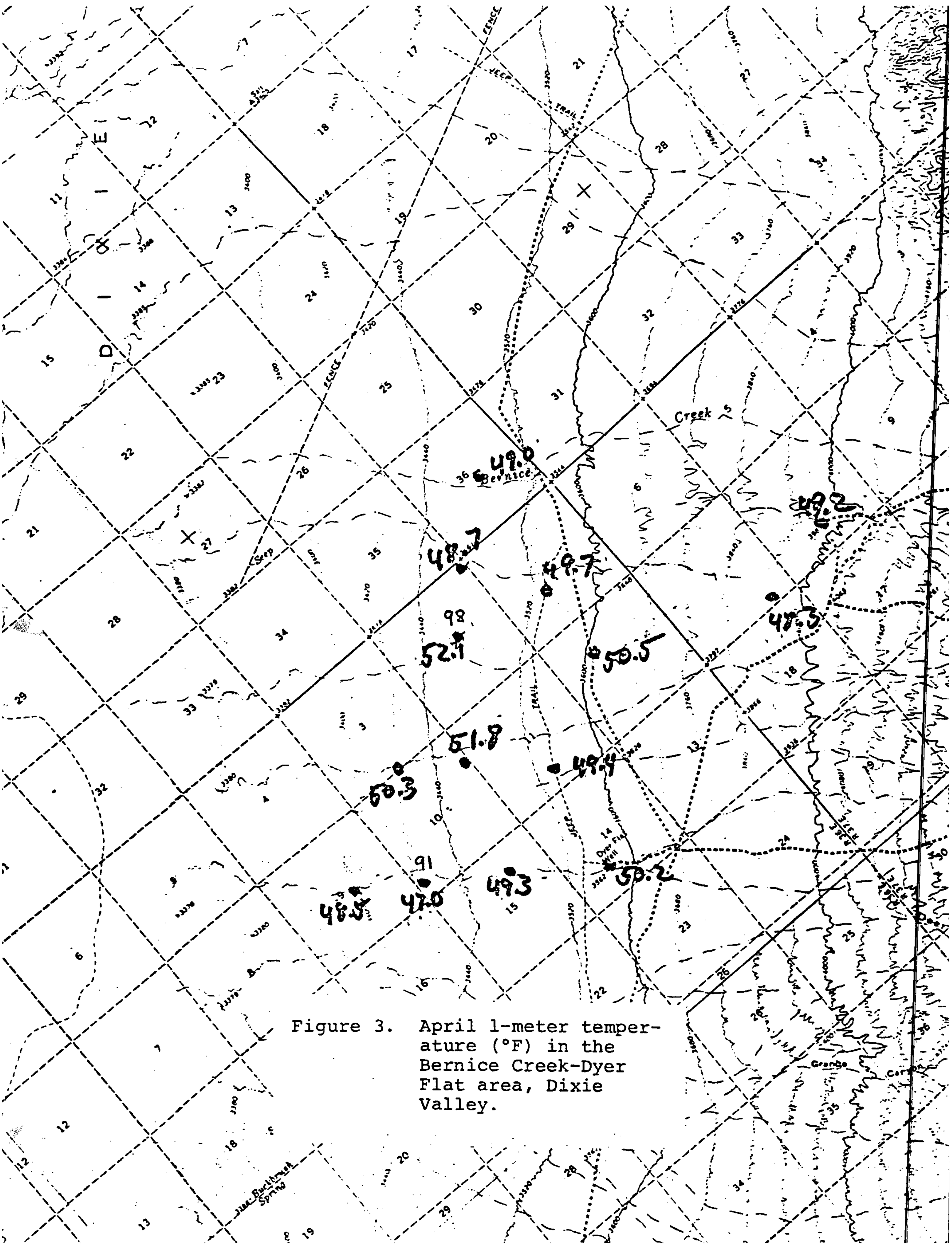


Figure 3. April 1-meter temperature (°F) in the Bernice Creek-Dyer Flat area, Dixie Valley.

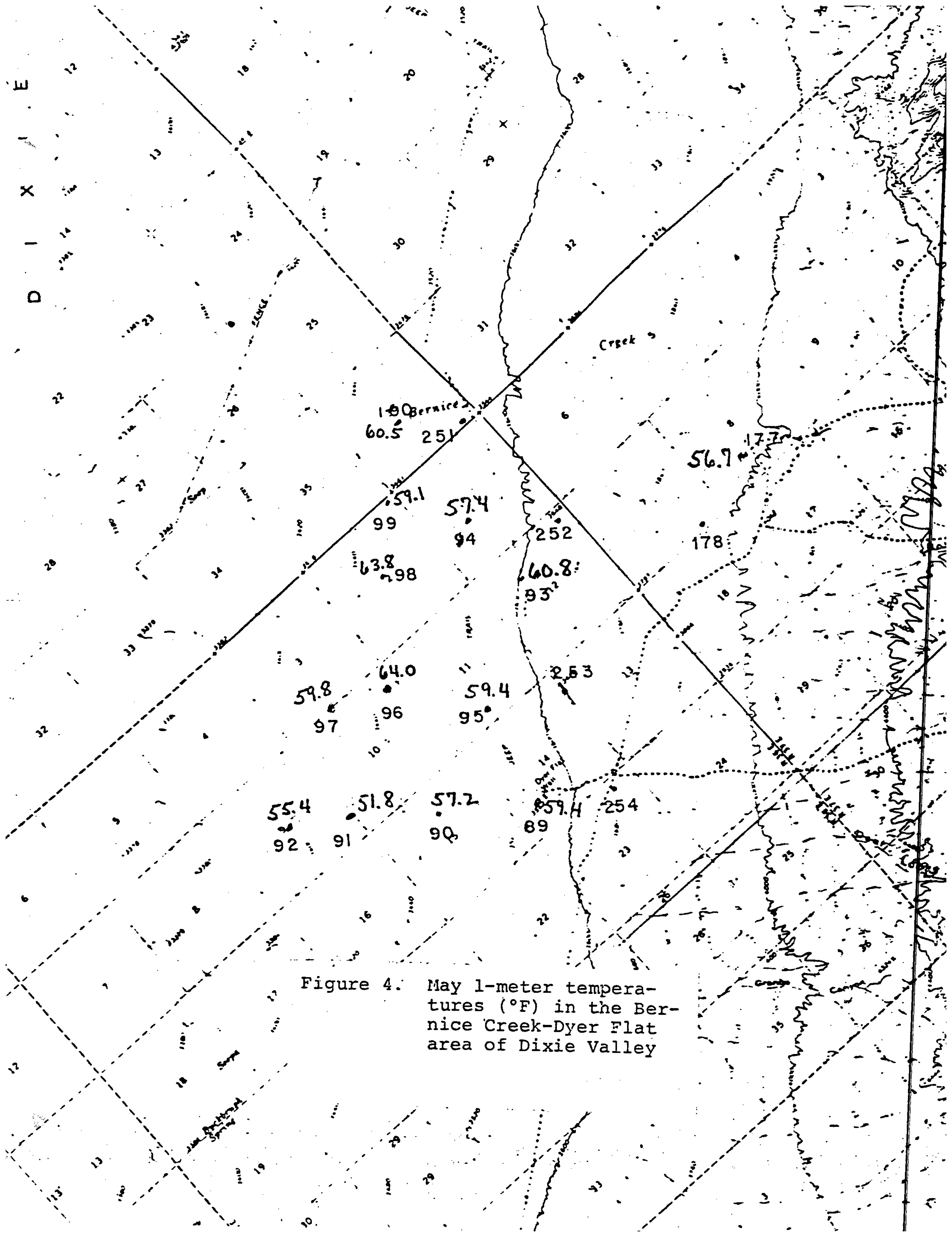


Figure 4. May 1-meter temperatures (°F) in the Bernice Creek-Dyer Flat area of Dixie Valley

of 7°F). This behavior is shown in Figures 5 and 6, which show the results for July and September respectively.

Other than these variations in maximum temperature differences, the temperatures in the Hyder Hot Springs area were not particularly high. The same statement can be made for the Sou Hot Springs-Seven Devils Springs area.

A region of high temperature is hole 61, located in Section 16 about 2.5 miles west of the Boyer Ranch. A maximum temperature of 93.7°F was measured in August. A few holes to the east of hole 61 showed relatively high temperatures, but those to the south and west did not exhibit such high temperatures.

One trend that seems apparent even from a cursory visual examination is the change in the "uniformity" of the temperatures. The April temperature map reveals the most uniformity--most temperatures are in the high 40s and low 50s. The sole exceptions are a few locations in the vicinity of Dixie Hot Springs. This apparent uniformity becomes less obvious as the surveys progress into the summer. Since the data have not been analyzed statistically, this apparent uniformity may not be "real", or if it is real, it may not be as significant as it appears.

Although detailed data analyses will commence after the writing of this quarterly report, it is apparent (and has been all along) that the shallow temperatures are strongly influenced by the local hydrologic system. We were somewhat surprised not to see the effects of the July flash floods on some of the temperatures.

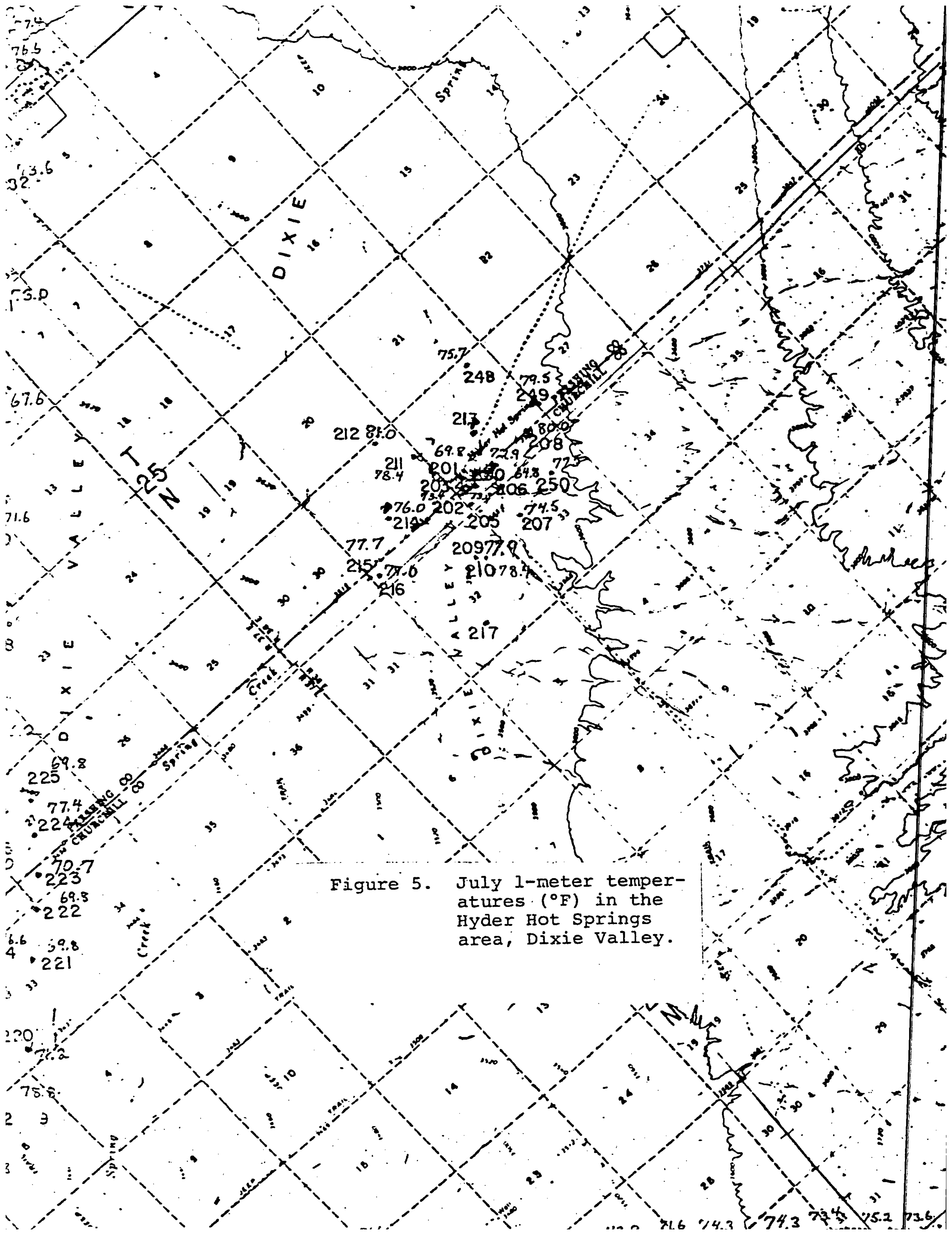


Figure 5. July 1-meter temperatures (°F) in the Hyder Hot Springs area, Dixie Valley.

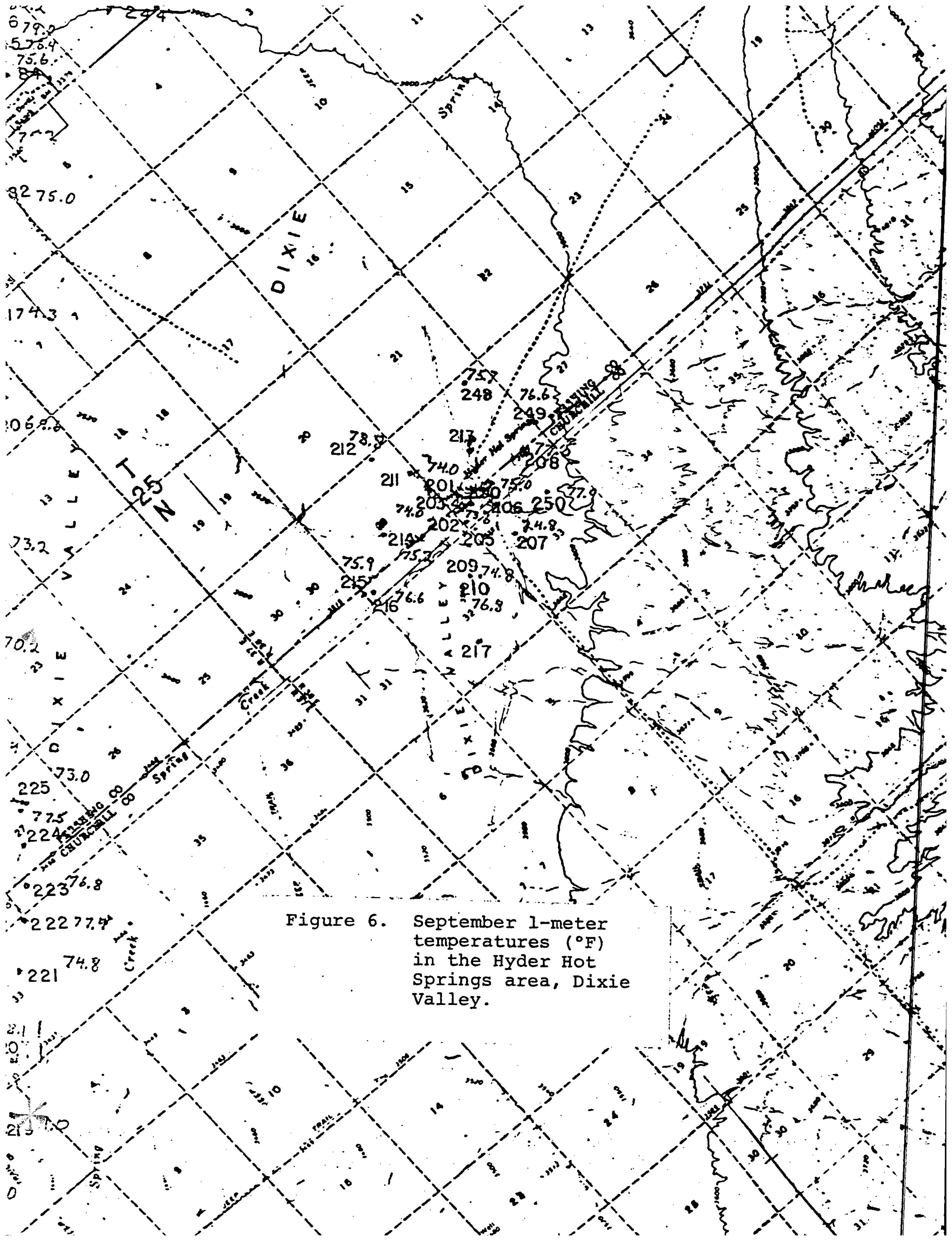


Figure 6. September 1-meter temperatures (°F) in the Hyder Hot Springs area, Dixie Valley.

These floods occurred along portions of the western side of the valley. One would expect infiltrating flood waters to depress the temperatures in certain areas. This belief is not supported by visual examination of the August and September temperature maps. However, it is possible that once the data are processed and analyzed statistically, differences will be apparent. It could also be possible that the effects have not yet manifested themselves.

PETROLOGIC ALTERATION STUDIES

Third Quarter Report

I. Work Accomplished

This analysis and interpretation of the alteration mineralogy from well DF 45-14 is based on currently available data. The purpose of this study is to detect, describe, and interpret the textural, structural, and mineralogic characteristics of the drill cuttings with emphasis on their relationship to the present hydrothermal system in Dixie Valley. This is accomplished by using x-ray diffraction, petrographic and water chemistry data.

All the samples from the second deep exploratory well DF 66-21 have been received for the total depth of the drill hole (9,780 feet). Epoxy thin sections have been prepared for approximately half of the samples, all of which will be composited to represent 100 foot intervals.

To date 80 samples have been x-rayed with ethylene glycol and 12 have been heat treated. Water chemistry data was correlated with the observed mineralogy and is discussed in detail in Section V. Results.

II. Problems Encountered

As noted in the second quarterly report, distinguishing specific authigenic hydrothermal minerals, i.e. in situ alteration, in the thermal gradient holes is difficult to establish from cuttings and the native low grade alteration. The alluvium through which the holes are drilled has pervasive alteration materials and at least part of the alteration predates the present geothermal influence. Thus the contribution to the overall alteration scheme from the present hydrothermal system is difficult to separate.

This same problem exists in interpreting drill cuttings from DF 45-14. The rocks encountered exhibit varying degrees of low-grade regional and low- to medium-grade contact metamorphism. These two events, regional and contact, are superposed in intervals corresponding to diorite/gabbro intrusions (i.e. 4100-4200 feet, 8000-8100 feet, 3700-3800 feet and others). In addition, the overprint of hydrothermal effects related to the present geothermal system further complicates the overall picture. Thus, the problem is to determine the contribution of each of these factors to the observed mineral assemblage. A possible solution to this problem could be obtained by analysis of equivalent rocks in nearby mountain areas that have not been affected by the hydrothermal activity. This could provide a standard to which rocks in the drill holes could be compared.

The small size of the drill cuttings continues to be a problem. The cuttings range in size from 0.1 mm to 15 mm with the majority being about 2 mm. Hence piecing together the overall textural, structural, and mineralogic relationships of a particular rock type is difficult. Moreover, the very fine grain size of the metasediments compounds the problem. Obtaining a representative sample of a particular depth is hindered by sample contamination and mixing caused by variations in circulation rates, settling velocities, and by caving. A representative sample is also a function of the technique used by the well logger to provide a split of the bulk sample. Composite samples of one hundred-foot intervals are used to reduce the effects of some of these problems, yet only a very small percentage of the grains of the composite samples are used in the thin sections. Epoxy is used as the mounting medium, instead of Canada balsam. The uncertain index of refraction for the epoxy, its higher index than the main minerals of the rock fragments and the difficulty of making relief comparisons, makes mineral identification difficult.

The completion of the analysis of DF 45-14 was delayed (2 months) by a malfunction in the x-ray diffractometer. Operation resumed on the 4th of September and consequently, only a limited number of samples are analyzed for this report. Although only partially complete, the data from this analysis is useful in the interpretation of the alteration in the drillhole. A complete interpretation of the clay mineralogy is discussed in Section V.

Certain factors inherent in x-ray analysis tend to bias results, including the type of mineral being analyzed, its degree of crystallinity, and the ability to reproduce an internally consistent sample. All these factors influence the observed peak intensity. Hence the validity of the interpretations derived from these results must be viewed with discretion. Two weeks is insufficient time to run, and interpret 90 x-ray samples. Analysis of clay minerals is especially difficult and time consuming. Only selected samples are heat-treated and/or glycolated within the limited time available since equipment repairs. The preliminary evaluation in this report uses some extrapolations and assumptions; the final report will supplement this data with definitive sheet silicate analysis.

One other problem which inhibited an interpretation of the relationships between the present hydrothermal system and the mineralogy is the uncertainty of the water chemistry. Without adequate chemical data, the equilibrium conditions and stability fields of the observed mineral assemblage(s) can only be estimated.

III. Projected Activities

When pure water chemistry and well log data (e.g. density and temperature logs) become available, the scope of this report will be expanded and the interpretations will be based on more solid information. In addition, information from the down hole elemental distribution analysis, and from surface sampling of altered rocks, and from the second exploratory well will contribute significantly to an understanding of the alteration mineralogy and the potential of the geothermal system.

Petrographic alteration studies and analyses will continue on DF 66-21. This will include detailed examination and interpretation of thin sections and x-ray data. The heavy mineral separations and the subsequent analyses will be conducted under additional funding approved for Geochemical Investigations. In addition, supplemental analyses and interpretations will be made regarding DF 45-14 to provide a thorough interpretation of that well. Much has been learned with regard to interpreting alteration mineralogy as it relates to the evaluation of a geothermal system; this process of learning will continue.

IV. Problems Which May be Encountered

The aforementioned problems will continue to persist, although it is hoped their effects can be minimized by learning and applying new methods and techniques. In addition, the schedule for funding of the Geochemical Investigations (in particular, the down-hole elemental distribution analysis) may delay integration of such data and require a re-evaluation of statements presented in the fourth quarter report.

V. Results

Analytical Methods

Thin section and x-ray diffraction analyses are the chief methods used to study and interpret the textural, structural, and mineralogic characteristics of the drill cuttings in relation to the hydrothermal system in Dixie Valley. Epoxy impregnated thin sections are examined using a polarizing microscope to identify and visually estimate the relative abundance of individual mineral species. Each thin section represents a 100-foot interval of drilled material and as a consequence, only gross changes can be detected. Figure 20 shows the distribution of selected minerals as a function of depth and lithology. The thin sections, in conjunction with the x-ray diffraction data, provide a means of documenting the interrelationships of the various mineral species. Hence, interpretations as to the significance of the occurrence of certain mineral species, as they relate to the geothermal system, can be made.

The x-ray diffraction analysis and interpretation of clay minerals is the most useful method in exhibiting the effects of the hydrothermal alterations in DF 45-14. Studies of the distribution of various types of clay minerals are widely applied in the exploration and evaluation of geothermal systems (Steiner, 1968; Hoagland, 1976). Moreover, the data from the thin section analysis is more meaningful when it is interpreted in light of the x-ray diffraction data (Figures 21 through 24).

Samples selected for x-ray diffraction are pulverized, slurried in distilled water and then centrifuged to isolate the fine fraction which is then pipeted onto glass slides. This procedure tends to orient the platy phyllosilicate minerals and enhances their basal reflections. Samples are then desiccated for 24 hours. Certain samples are heated to 560°C for one-half hour and/or glycolated at 90°C for 24 hours to facilitate identification.

of kaolinite and expandable clays respectively. Data is obtained on a Philip x-ray diffractometer with Cu Ka radiation of an operating voltage of 20 Kv and 15 milliamps and a scanning speed of 2°/min.

Table 1 summarizes the parameters used in the identification of the various clay minerals. The term illite is here used in agreement with Bradley and Grim (1961 p. 775) for micaceous clay minerals (biotite, muscovite, phlogopite, paragonite and sericite) showing little departure from 10A integral series. Micaceous clay minerals with d(001) ranging from 10.28 to 12.45A reflect the presence of interstratified, expandable layers which are sensitive to treatment by ethylene glycol. Thus, micaceous clays with d(001) of 10.16A are classified as randomly interstratified illite-montmorillonites (Steiner, 1968; K. Papke pers. comm., 1979). Na montmorillonite is assumed at a d(001) of 12.25 to 12.4 A. and upon glycolization, a d(001) of 16.5-17.0 A (K. Papke, pers. comm., 1979). One interval, 2300-2400, reveals the presence of a Ca or Mg-montmorillonite with a d(001) of 14.25 A and a shift to a d(001) of 15.5 A upon glycolization. This indicates the presence of interlayered illite (K. Papke, pers. comm., 1979). The chlorites are believed to be Fe-rich as the d(001) of 14.0-14.3 A increases in intensity upon heating to 560° C for one-half hour (Steiner, 1968).

Rock Types in DF 45-14

Wilden and Speed (1974) mapped and described seven rock groups within the Dixie Valley study region. These include: Upper Triassic metasediments, Middle Jurassic quartz arenite, mafic volcanic rocks, and gabbroic rock, Miocene rhyolites and Pliocene and younger sedimentary rocks, basalt and andesite.

DF 45-14 penetrated 6,500 feet of a possible 10,000 (Page, 1965) of Upper Triassic metasediments. These consist chiefly of metasiltstone, metalithic wackes, metasandstone and metarenite. Two facies are present in the drill cuttings based mainly on grain size and the degree of foliation and alteration. The majority of this metasedimentary sequence consists of a very fine-grained, black to medium-dark gray, well sorted, very hard silicic, commonly recrystallized metasiltstone/metashale, often containing abundant

Table 1

X-ray Diffraction Data

Mineral	d(001) Å	Glycolated d(001) Å	Heat Treated, 1/2 hr d(001) Å
Na-montmorillonite	12.3-12.5	16.5-17.2	No data
Ca or Mg-Montmorillonite	14.25	14.55	14.1
Mixed Illite/Mont.	>10.16	no data	unchanged
Illite	9.8-10.16	unchanged	unchanged
Chlorite	14.0-14.3	no change	increase in intensity

carbonaceous matter and varying amounts of crystalline and disseminated pyrite. This unit commonly exhibits well developed foliation that is a result of regional low grade metamorphism during late Jurassic time. The unit is fairly dense and shows only occasional signs of fracturing. The other facies is a coarser-grained light to dark gray to brown, very hard silicic, recrystallized rock. Figure 1 shows some of the representative rock types encountered in the metasedimentary sequence.

Mineralogically, quartz is the dominant mineral in both facies comprising up to 90% of the rock. In the finer grained facies (Figure 2) are fairly abundant occurring as an integranular phase along with lesser amounts of chlorite, calcite, chiastolite, cordierite (?), pyrite, magnetite, carbonaceous matter, kaolinite, and traces of epidote, garnet and tourmaline. Figure 20 depicts the distribution of selected mineral species as a function of depth and lithology.

The mineralogy of the coarser facies (Figure 3) consists of quartz, intergranular micaceous mineral and chlorite and lesser amounts of vermiculite(?), magnetite, pyrite, garnet, calcite, carbonaceous matter, kaolinite, epidote and tourmaline. Widespread recrystallization of this unit may be responsible for the conspicuous lack of a well defined foliation as seen in the metasiltstone/shale. The small variations in grain size, mineralogy, and texture of this coarser-grained facies correspond to the various subfacies, i.e. metarenite, metalithic wacke, etc.

Several intervals of the metasedimentary sequence (3700-3800, 4100-4200, 4600-4700, 7600-7700, and 8000-8100 feet) have been intruded by Middle Jurassic diorite/gabbro dikes or sills. The average thickness of these intrusions is about 30 feet (lithologic well logs). They consist predominantly of a medium to coarse-grained crystalline network of plagioclase and clinopyroxene, with traces of olivine. Secondary minerals include hornblende, chlorite, epidote, calcite, albite, magnetite and pyrite (see Figures 4, 5 and 6).

A total of 1500 feet of volcanic rocks are encountered in the drill hole, in the intervals (700-800 ft and 1100-2500 ft.) and consist primarily of tuffs, andesites and intercalated tuff andesite. The tuffs are gray to green and brown and consist of fine-grained interlocking plagioclase lathes in a very

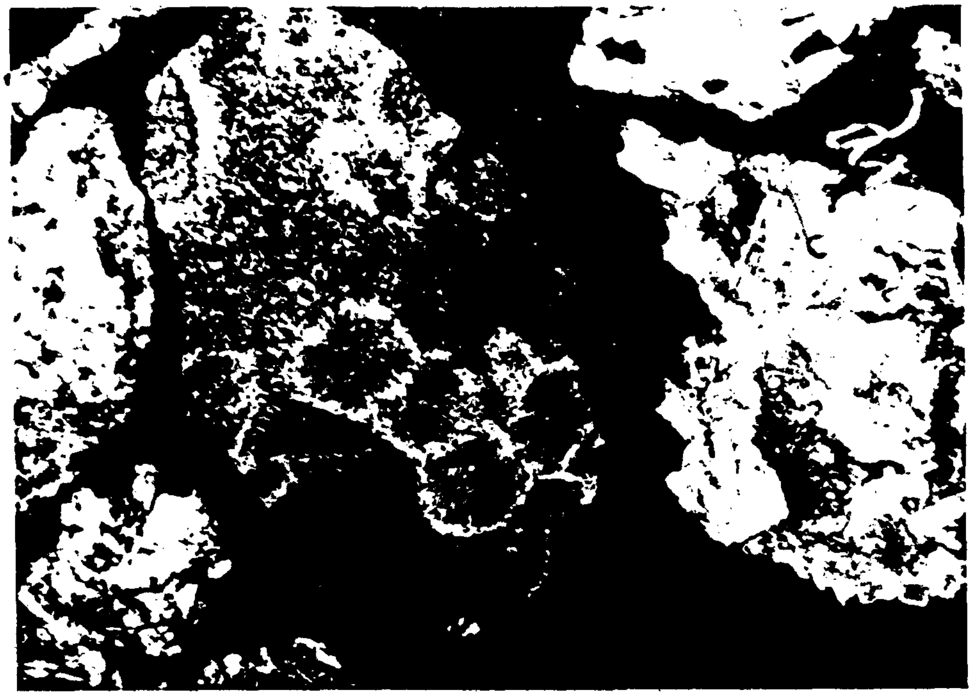


Figure 1 Representative rock types in DF45-14
A, Meta-quartz arenite; B, Meta-siltstone;
C, Diorite/Gabbro.

Figure 2. Foliated chiastolite-bearing meta-siltstone.

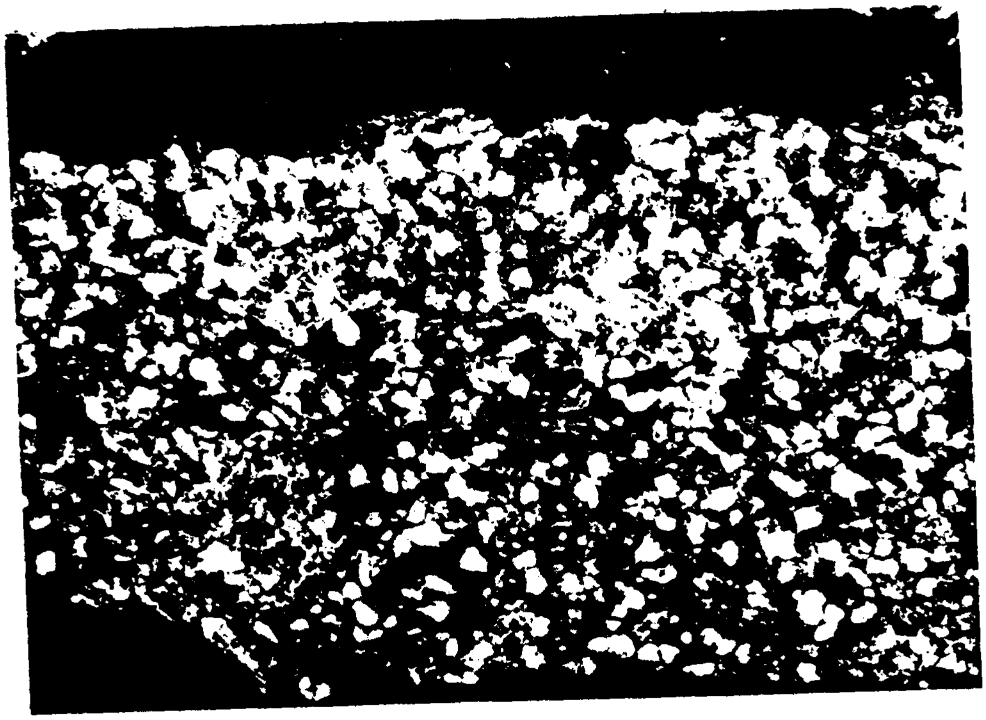


Figure 3. Coarse-grained facies of metasedimentary sequence.

Figure 4. Diorite/gabbro.

Figure 5. Diorite/gabbro (crossed polars)

Figure 6. Diorite/gabbro (uncrossed polars)
A, plagioclase; B, hornblende
C, clinopyroxene; D, magnetite

fine-grained glassy matrix that is altering to kaolinite. These vitric and lithic tuffs, often devitrified, contain varying amounts of disseminated and crystalline pyrite. The andesites are predominantly medium to dark brown, black and green, fine-grained rocks consisting of interlocking plagioclase crystals with minor amounts of pyrite and hornblende in a groundmass of very fine-grained plagioclase.

The approximately 1100 feet of alluvium encountered in DF 45-14 is a mixture of all the rock types described above. The metasediments though are conspicuously rare or absent. The majority of material has been derived from the volcanic terrain in the Stillwater Range. Diorite/gabbro is also fairly abundant. All of these rock types exhibit a greater degree of alteration than their in situ counterparts in the drill hole. Hence, documenting the contributions of the hydrothermal system to the observed mineralogy is difficult.

Effects of Metamorphism

The overall effects of regional low grade metamorphism on the metasediments are minimal; these include crystallization of muscovite, biotite, and quartz and the development of foliation (Figure 2). The pre-metamorphic equilibrium mineral assemblage of quartz, muscovite, and biotite has not been radically altered by the metamorphism. The present assemblage consists of quartz, K-micas (muscovite and biotite) chlorite, kaolinite ± calcite ± laumontite. Chemical reconstitution and textural changes were the chief end products of the regional low grade metamorphism. The present equilibrium mineral assemblage is a result of the composite effects of the metamorphism and the present hydrothermal system.

Superimposed on this regional low grade metamorphic fabric in the metasediments are localized regions of low to medium grade contact metamorphism associated with the diorite/gabbro intrusions; Figure 21 indicates the locations of these intrusions in the stratigraphic column. The contact metamorphism resulted in the growth of chiastolite porphyroblasts (a variety of andalusite (Al_2SiO_5)) and the growth of 'spots' or 'knots' of incipient crystals of andalusite and cordierite that have not completely nucleated (Figures 7-14).

The chiastolite crystals, with their characteristic cross-shaped inclusions of organic material and the spherulites were observed to occur only in the fine grained facies of the metasedimentary unit. Here, the foliation exhibited by orientations of the phyllosilicate minerals, i.e. muscovite, biotite, and chlorite, and to a lesser extent carbonaceous matter, invariably wrap around the 'spots' and in some instances the chiastolite crystals (Figures 2, 15, 15a). Chiastolite though, generally grows across the foliation. The phyllosilicate mineral orientations may have been produced by mimetic recrystallization of the minerals either along foliation planes or along original bedding surfaces (Spry, 1976).

Porphyroblasts of chiastolite and the 'spots' in a fine grained matrix are referred to as maculose structure (Spry, 1976, p. 192). The organic cross or inclusions in the chiastolite crystals are believed to be due to preferred adsorption of carbonaceous particles against the prism faces and particularly at the prism edges (Spry, 1976, p. 194). Chiastolite and particularly its organic inclusions are highly susceptible to sericitic alteration. This feature is pervasive throughout the thin sections (Figures 10, 13, 14). The occurrence of the chiastolite and spotted metasiltstone is spatially related to the location of the diorite/gabbro intrusions (Figure 20).

The sequence of volcanic rocks, the majority of which are tuffs, exhibits a fairly uniform type of alteration. The effect of the regional low grade thermal metamorphic event is difficult to discern from the effects of the present hydrothermal system. The present equilibrium mineral assemblage of albite, chlorite, sericite, clay, calcite and epidote is characteristic of a propylitic alteration assemblage (Coats, 1940). This assemblage is also characteristic of regional low grade or greenschist facies metamorphism (Winkler, 1976). This same mineral assemblage has been documented in other geothermal regions as being indicative of hydrothermal alteration (Browne and Ellis, 1970; Steiner, 1968).

Figure 15. Shows differences in grain size and degree of foliation and alteration between the fine and coarse-grained facies of the metasediments (crossed polars).

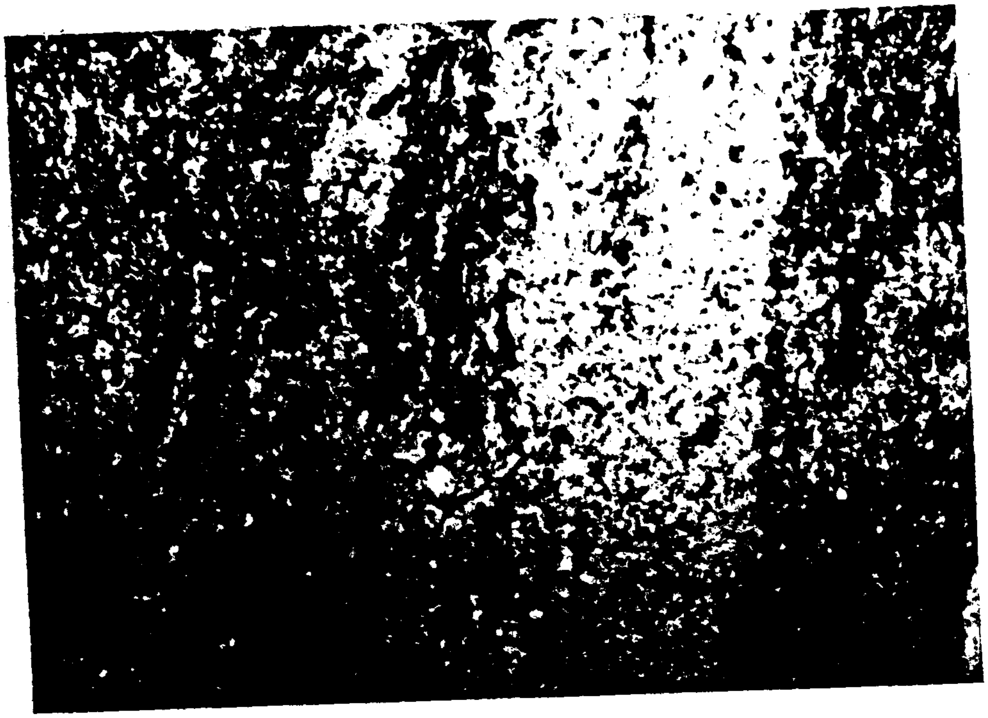


Figure 15a. Same as figure 15 (uncrossed polars). Note 'cleaner' appearance of the coarse-grained facies.

Figure 16. Highly altered volcanic rock; note complete sericitization of the plagioclase lathes.

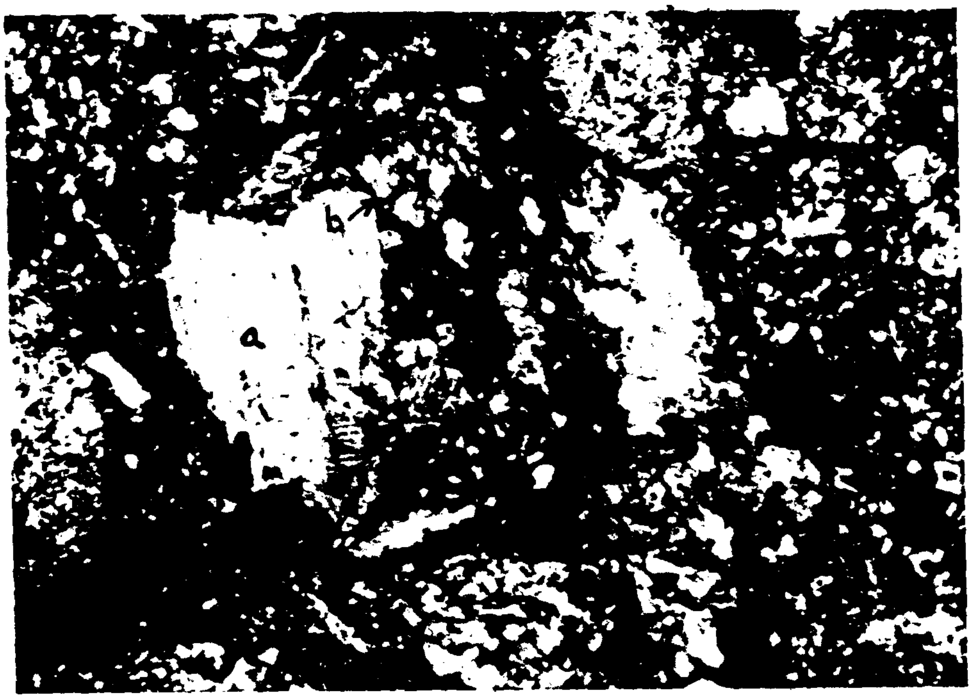


Figure 17. High magnification of drill cuttings from the interval 5400-5500 feet. Note the very small size and the presence of calcite (a) and epidote (b)
(zoisite)

developed in the groundmass of the volcanic rocks, and has largely masked much of the original texture. The presence of montmorillonite clays in the upper 2500 feet of DF 45-14 is indicative of this process. Montmorillonite is characteristic of low temperatures and its formation by alkaline solutions would correspond with the conditions of sericite formation (Turner and Verhoogen, 1960).

The alteration of the ferromagnesian minerals in the alluvial material is pervasive and generally complete. Original pyroxenes of the diorite/gabbro are completely replaced by hornblende in both the diorite/gabbro and the volcanic rocks. Hornblende is altered to chlorite and magnetite ± epidote and calcite. The iron derived from the alteration of these minerals and pyrite is responsible for the pervasive limonite and hematite staining of the cuttings.

The type of alteration described above for the alluvial material also applies to the alteration observed in the volcanic sequence in DF 45-14. The difference is only in degree; this is probably a function of the additional factors of erosion and transport that have affected the alluvium. The observed mineral assemblages of the two regimes are also consistent.

The alteration and observed mineral assemblage in the diorite/gabbro, although limited, bears some resemblance to that described above (Figure 5, 6). Namely, secondary hornblende is replaced by chlorite and magnetite with some pyrite after magnetite. Plagioclase alters to albite, K-mica, calcite and epidote. Despite these phase transitions, the diorite/gabbro maintains a relatively fresh, unaltered appearance especially when compared to equivalent rock fragments in the alluvium. This alteration mineral assemblage appears to be genetically related to that developed in the alluvium and the volcanic sequence. Here again, the difference is one of degree and is probably a function of differences in and permeability and fluid migration relationships in the rock types. Propylitic alterations of a chlorite is described in the Tongonan Geothermal Field, Leyte,Philippines (Kingston, et. al., 1979).

The alteration products of primary minerals in the alluvial material, in the sequence of volcanic rocks, and to a lesser extent, in the diorite/gabbro intrusions are all similar (i.e. albite-chlorite-epidote-K-mica and calcite ± pyrite). This assemblage is described in other geothermal areas as well. Work by Brown and Ellis (1970) at Ohaki Broadlands in New Zealand describes hornblende and biotite replaced by chlorite, illite (K-mica), calcite, quartz, or pyrite, and plagioclase replaced by adularia, albite, calcite, illite, epidote, wairakite, and quartz. Work by Steiner (1968) at Wairakei, New Zealand, describes ferromagnesian minerals altering to chlorite, micaceous clays, quartz, calcite, epidote and pyrite, with plagioclase altering to montmorillonite, micaceous clays, calcite, epidote and quartz and K-feldspar. Work by Schoen and White (1965) at Steamboat Springs, Nevada, identifies hornblende altering to mixed layered illite/montmorillonite clays, calcite, epidote and K-spar.

Evidence to date from DF 45-14 along with the observed alteration in other geothermal systems seems to favor a hydrothermal origin for the observed mineral assemblages in the alluvial material, the sequence of volcanic rocks and in the diorite/gabbro intrusions. The degree of alteration in these rock units is mainly a function of rock texture which determines the porosity and permeability, influences the fluid migration rates, and ultimately the reaction times. For example, the diorite/gabbro intrusions are surrounded by relatively impervious metasediments, which essentially protect it from the more intense alteration occurring in fracture zones. However, permeable passageways may be developed at the contacts, although this has not been verified. In addition, the metasediments have not been significantly affected by hydrothermal solutions except in areas of higher permeability and porosity, i.e. fracture zones.

Alteration in the Metasediments

As mentioned above the response of the metasediments to the hydrothermal activity is rather limited except in localized regions where there is evidence of fracturing. Additionally, the effects of alteration show variations among the various facies (see Figs. 15 and 19). The postmetamorphic mineral assemblage of quartz, muscovite, chlorite, and biotite has not been significantly changed. This is supported by x-ray diffraction on a whole rock sample which indicates the presence of quartz, chlorite, K-mica, and possibly kaolinite.

The presence of kaolinite, though, is not confirmed (heat-treated) for this sample.

Sericitization and/or illitization has been active, particularly in the finer grained facies of the metasiltstone. Alteration of biotite is also observed, predominantly in the coarse-grained facies of the metasedimentary sequence, where it is altered to chlorite and vermiculite. The petrographic evidence for the vermiculite includes: (1) lack of the characteristic 'birds-eye' extinction, (2) a decrease in birefringence, (3) a decrease in pleochroism, (4) a bleached appearance, and (5) pseudomorphs after biotite. One of the major occurrences of vermiculite is as an alteration product of biotite by weathering or by hydrothermal action (Deer and others, 1966). Vermiculite has been documented in the Roosevelt Hot Springs Thermal Area, Utah (Ballantyne, 1978), where it has replaced biotite.

Chiastolite crystals in the fine-grained, organic-rich metasediments are invariably partially or wholly sericitized, particularly the included organic material. However, the stability of chiastolite is such that under normal atmospheric conditions its alteration to sericite is entirely possible. Therefore, the sericitzation of chiastolite is not definitive evidence for hydrothermal alteration (Spry, 1972). A large proportion of the spots in the fine-grained metasiltstone/shale are altered to clay.

Calcite and quartz are the dominant vein and fracture-fill materials along with the Ca-zeolite laumontite and lesser amounts of adularia and gypsum in the metasediments (Figs. 17, 18, 18a, 19). These minerals exhibit a variety of structural and temporal relationships. Calcite, quartz and possibly gypsum veining have been observed to be pre-metamorphic although the majority is post-metamorphic. The distribution of calcite and its relation to fracture zones is depicted in Fig. 20. Laumontite and adularia invariably occur in post-metamorphic veins. Both laumontite and adularia are described in the literature as minerals typically found in geothermal systems (Steiner, 1967; Browne and Ellis, 1970; Ellis and Mahon, 1977). Laumontite is also characteristic of very low grade metamorphism (Winkler, 1967).

The presence of the Ca-rich zeolite laumontite ($CaAl_2Si_4O_{12}4H_2O$) is confirmed by x-ray diffraction. The main diffraction peaks at 9.1-9.4 Å

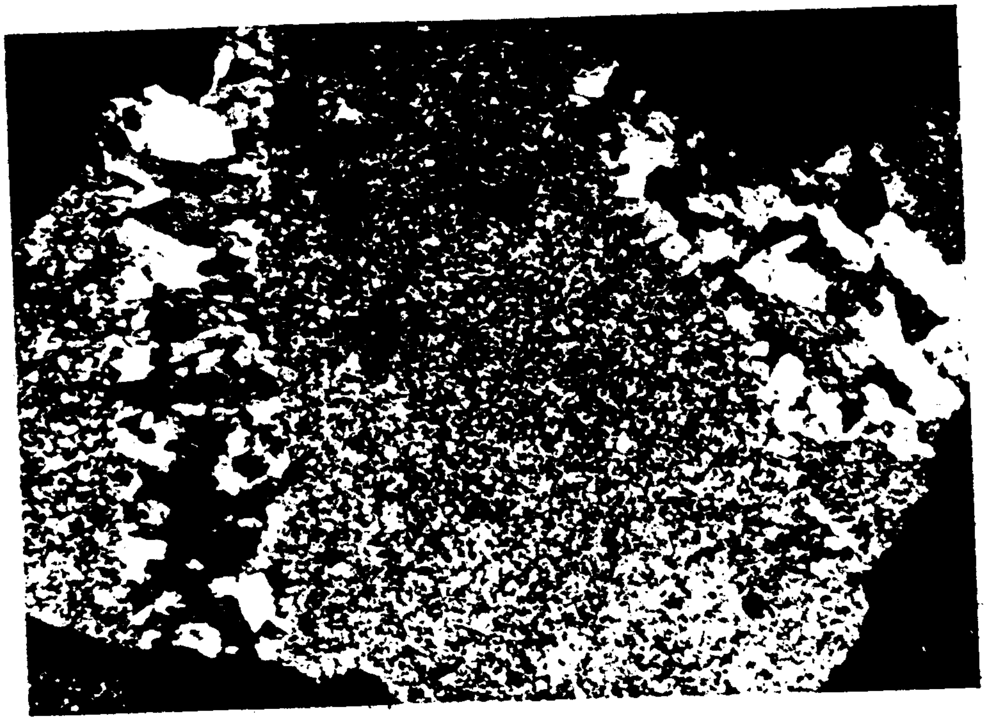


Figure 18. Example of quartz veining in meta-siltstone.
(crossed polars)

Figure 18a. Same as figure 18 (uncrossed polars).

Figure 19. Coarse-grained adularia occurring as vein material.

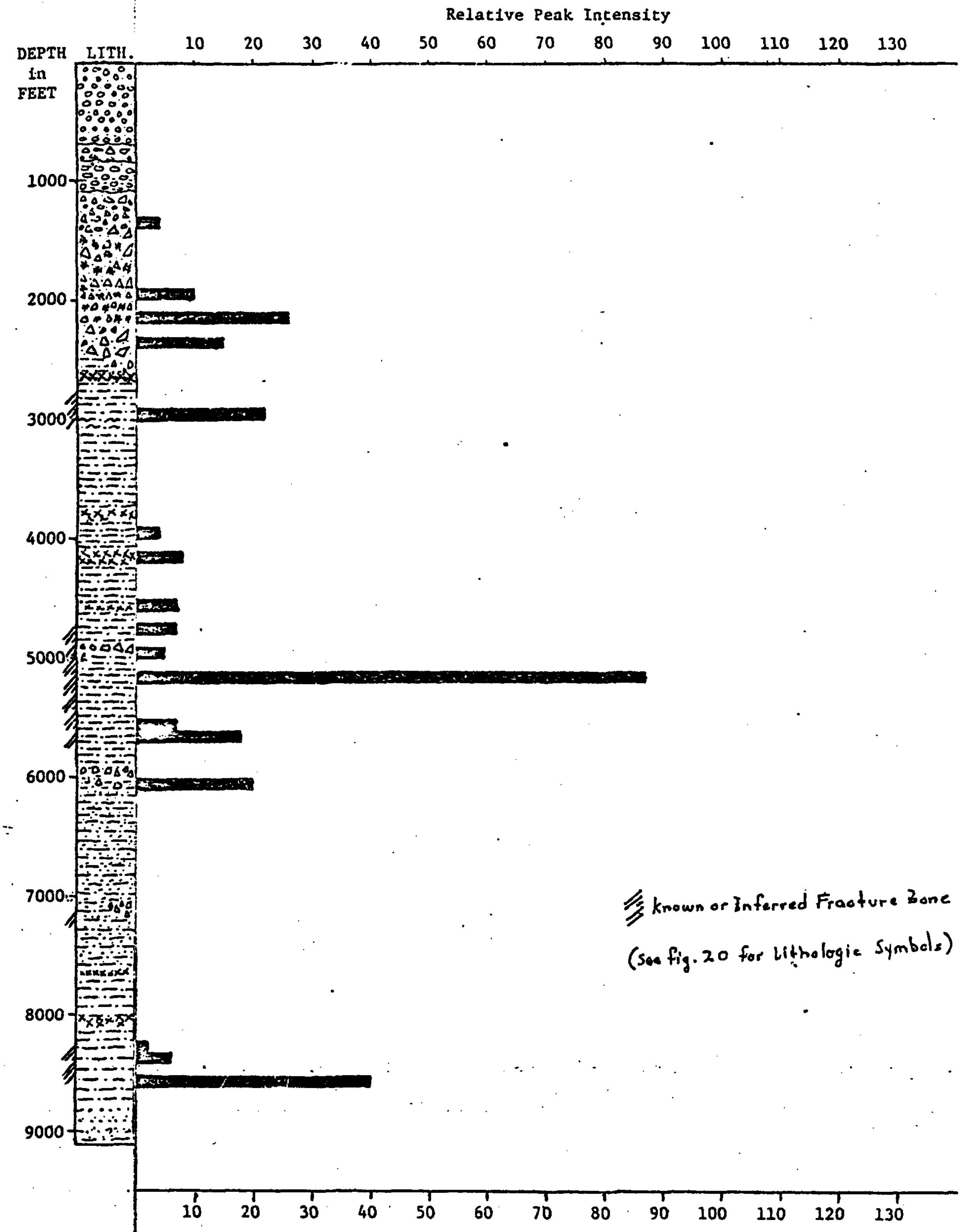
Figure 20 DISTRIBUTION OF MINERAL SPECIES

and 6.7-6.9 Å disappear after heating to 570°C for one-half hour. The occurrence of laumontite in Nevada has not been described in the literature although a more thorough search is necessary to verify this statement. Figure 21 shows the relative abundance and distribution of this mineral with depth and its relation to fracture zones and lithology. In thin section it is coarsely crystalline with a low birefringence and relief and has two directions of observable cleavage, one distinct and the other very obscure. Laumontite occurs with quartz and calcite in veins and fracture fillings. The assemblage quartz-laumontite is typical of the zeolite facies (Deer and others, 1966). This same assemblage is documented in Onikabe, Japan (Ellis and Mahon, 1977). Laumontite was indicative of 150°-200°C temperatures at Wairakei, New Zealand and 170°C in Onikabe, Japan (Ellis and Mahon, 1977). Thus, as can be seen from Figure 21 the distribution of laumontite is also spatially related to fractures. This feature provides additional support for the idea that the permeability and hence alteration effects in DF 45-14 are fracture-controlled.

Clay Mineralogy

Figures 22-24 show the distribution of the various clay minerals with depth, and their relationship to fracture zones and lithologic contacts. These figures exhibit a consistent and highly localized pattern of distribution, particularly for micaceous minerals and kaolinite. The occurrence of the montorillonite and mixed-layer clays can also be related to stratigraphic-structural patterns.

The zones of high clay concentrations are invariably associated with either fracture zones, zones of no drill-cutting returns or changes in lithology. Evidence for fracture zones includes: very fine drill cuttings (Fig. 16), an abundance of vein material (quartz, calcite, and laumontite), or zones of no or limited cutting returns. This pattern of occurrence indicates that the permeability and hence alteration of the rocks is fracture controlled and that rock type is also an important factor. In Wairakei, the montmorillionite zone and the micaceous zones (illite, mixed-layer illite-montmorillonite) are related to temperature and fissures carrying geothermal fluid (Steiner, 1968).



known or Inferred Fracture Zone

(See fig. 20 for Lithologic Symbols)

Figure 21. Distribution of Laumontite

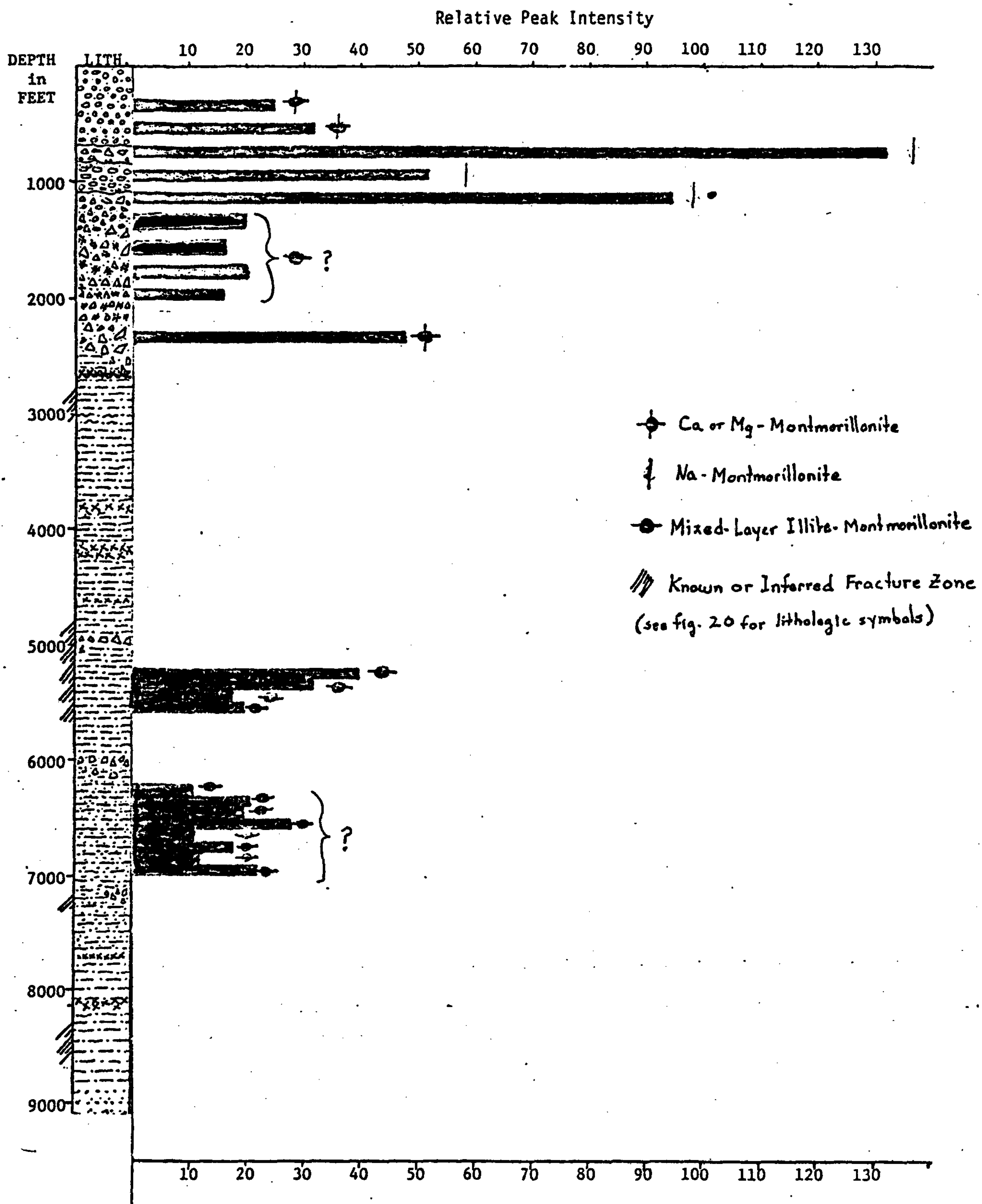
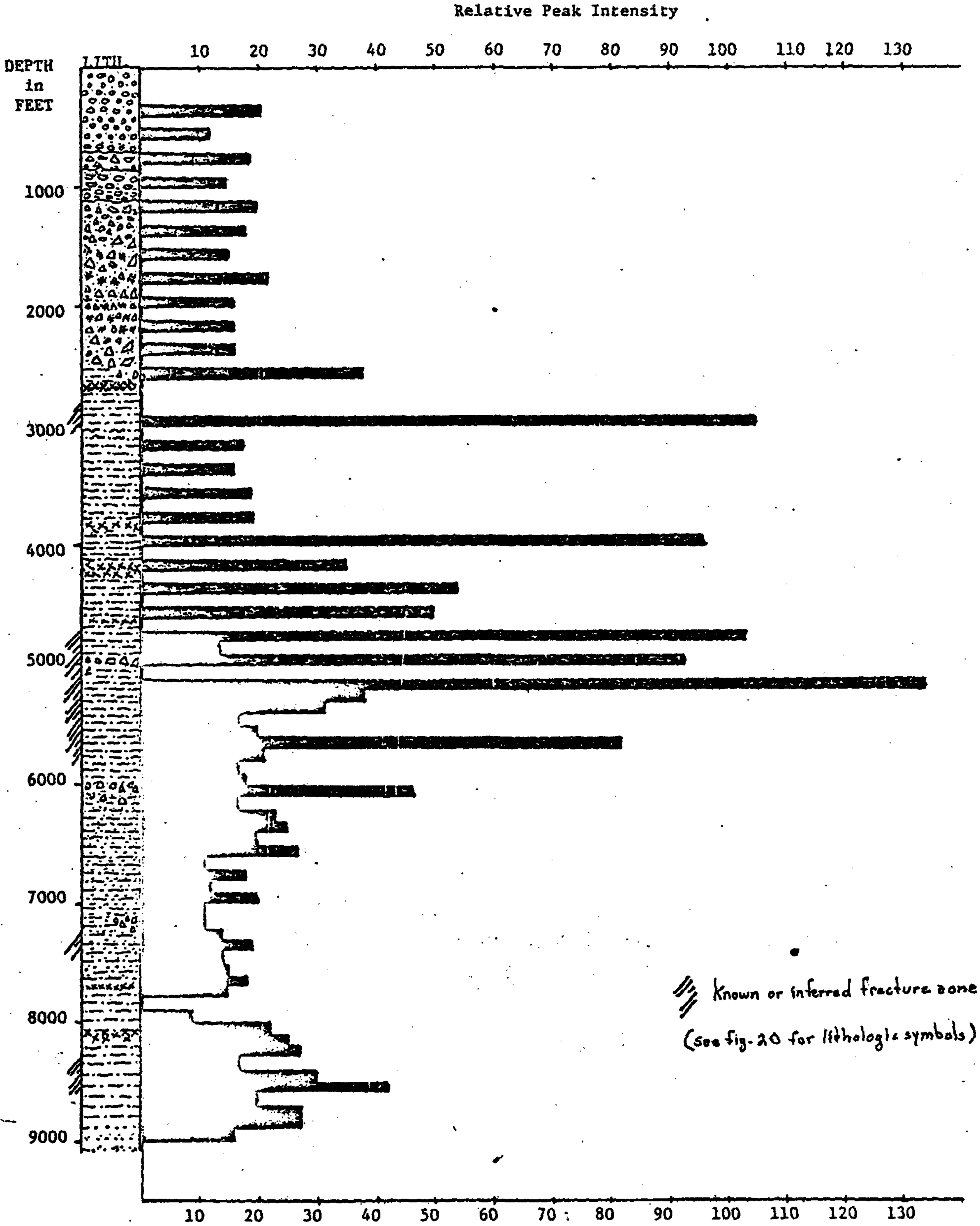


Figure 22. Distribution of Montmorillonite and Mixed-Layer Illite-Montmorillonite.



Known or inferred fracture zone

(see fig. 20 for lithologic symbols)

Figure 23 Distribution of Illite

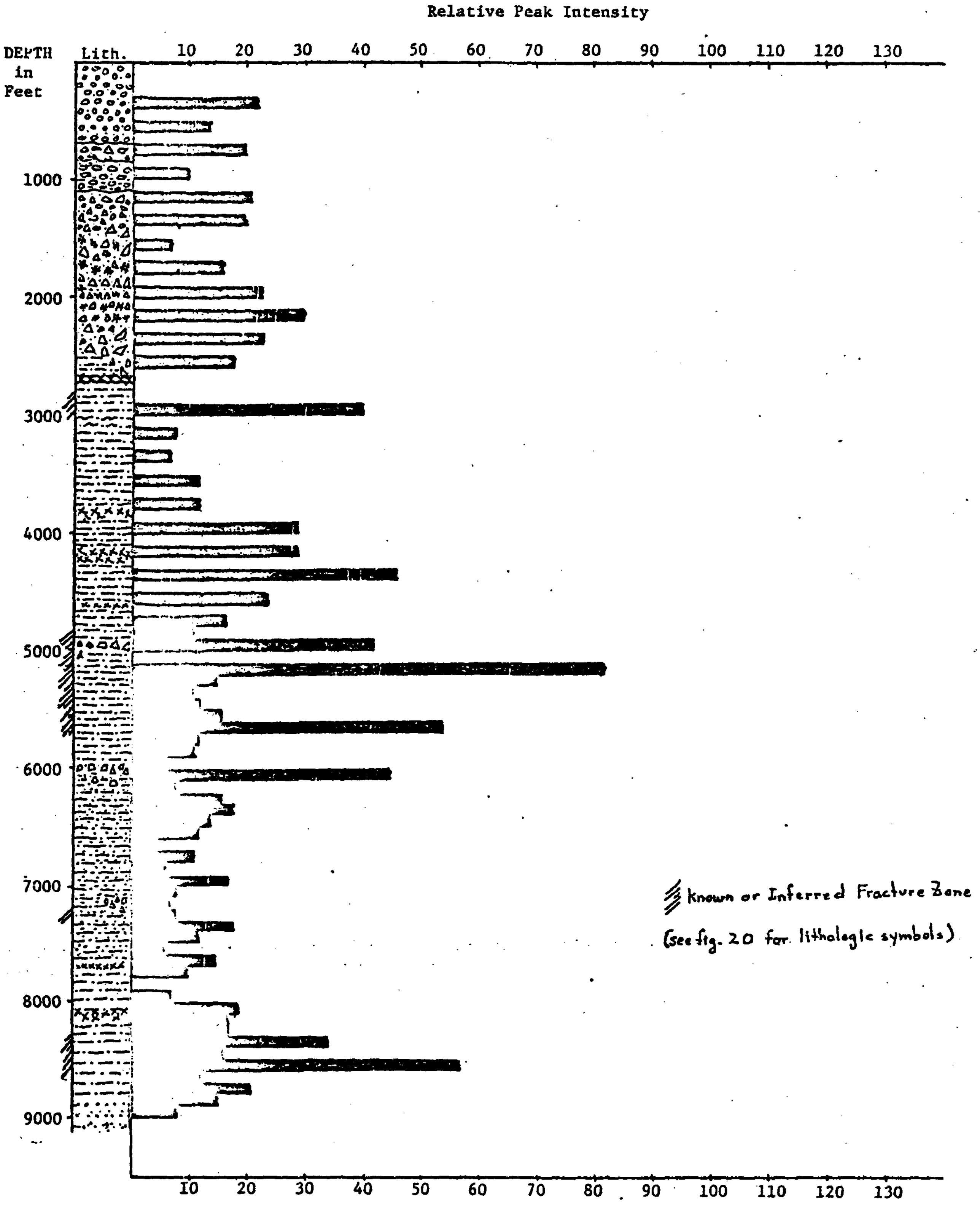


Figure 14 Distribution of Kaolinite

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

The temperature relationships of the chlorite and kaolinite clays are uncertain. Kaolinite forms at temperatures under about 350°C (Krauskopf, 1967), but chlorite has a very wide stability range. Montmorillonite is indicative of temperatures of 100°-200°C in altered rhyolitic tuffs in Wairakei (Steiner, 1968). In the Salton Sea Geothermal Area, a temperature of about 100°C is observed for the termination of mixed-layer illite-montmorillonite while the conversion to illite is complete at water temperatures of about 210°C (Ellis and Mahon, 1977). At Wairakei, Steiner (1968) found that low temperatures between about 120°C and 230°C prevail at shallow depths where mixed-layer minerals are forming, i.e. low lithostatic and water vapor pressures. He found that high pressures and temperatures (above 230°C) are required for the formation of illite and mixed-layer clays deficient in montmorillonite. In Reykjanes, Iceland, montmorillonite is the dominant sheet silicate up to temperatures of about 200°C and for temperatures of 200°-300°C the most common sheet silicates are random mixed-layer minerals of chlorite and montmorillonite. Chlorite in Reykjanes characterizes a higher temperature range of 230°-280°C (Tomasson and Kristmannsdottir, 1972). Thus it can be seen that a fairly wide range of temperatures has been reported for clay minerals. In DF 45-14, the formation of montmorillonite probably occurs at about 80-125°C and slightly higher temperatures, 120-170°C for the mixed-layer clays.

Montmorillonite generally forms under alkaline conditions (Krauskopf, 1967). The occurrence of Na-montmorillonite in the interval 700-800 feet (Fig. 22) corresponds to the location of a tuff unit. The upper contact of this unit is at 700 feet. The interval 900-1000 feet corresponds closely to the lower contact of this same unit. Refer to Fig. 22. Thus, in these two intervals, the abundance of montmorillonite is related either to the rock composition or to the upper and lower contacts of the tuff unit, as these would provide permeable avenues for migrating thermal fluids. From 1100 to 2400 feet, the intercalated tuff-andesite sequence is relatively undisturbed and does not exhibit any appreciable degree of clay mineral formation. The interval 2300-2400 feet, which is interpreted as representing the contact between the metasediments and the overlying volcanics, corresponds to an anomalous occurrence of mixed-layer illite-montmorillonite, with the montmorillonite being Ca or Mg-rich. This is the only observed occurrence of a Ca-Mg montmorillonite in DF 45-14. Its significance is unknown at this time.

Other intervals, 5200-5600 and 6200-7000 feet, although not as strongly developed, indicate the presence of mixed-layer illite montmorillonite. The interval 5200-5700 feet corresponds to the intersection of the range front fault when it is projected to depth at an angle of 60°. The calculated depth is 5500 feet. The interval 6200-7000 feet is less distince as to its significance. It is marked in some places by a lack of cuttings coming to the surface (6900-7000 feet), and by an abundance of "trash" in the cuttings, indicating a loss of circulation in a fracture zone. The cuttings in part of this interval 6300-6600 feet are essentially a 50-50 mixture of metasiltstone and metarenite. If this is an interbedded sequence, then the bedding planes could provide the necessary permeable channels for fluid migration and promote the formation of the observed mixed-layer clays. However, the x-ray diffraction patterns for the mixed-layer clays in this interval are not well defined.

The interval 4700-5700 feet is the most significant interval in the drill hole as it exhibits some of the most intense alteration encountered in the metasedimentary sequence (see Figures 21-24). This zone represents a large fracture zone related to the range front fault. Illite and kaolinite are most abundant in this zone along with lesser amounts of mixed-layer illite-montmorillonite.

In this zone (4700-5700 feet), illite represents the highest temperatures giving way to mixed-layer illite-montmorillonite towards the downhole side of the fault zone (5200-5500 feet). This is interpreted as being an indication of a temperature inversion below the main part of the fault zone. However, downhole temperature data is needed to verify this phenomenon which has been documented in other geothermal areas (i.e. in Wairakei, New Zealand, Steiner, 1968). Hence the distribution of the clay minerals with respect to fractures and faults is chiefly temperature dependent.

The temperature relationships of kaolinite are obscure at this point, however, its distribution (Figure 24) appears to favor factures and higher temperatures. The occurrence of kaolinite is very significant genetically in that it requires acidic conditions. This is entirely possible at depth in DF 45-14 if one assumes that the range front fault communicates with the reservoir at depths where more acidic conditions are postulated to exist. The upward migration of these acidic reservoir fluids along the fault zone promotes the formation of kaolinite. This has probably been a cyclic process relating to the opening and closing of the fractures as a result of tectonic activity, precipitation and/or internal deformation of the metasediments' response to lithostatic pressure.

The genesis and distribution of the chloritic clay is obscure. The hydrothermal alteration of silicic glass to chlorite is a relatively recent discovery, though such alterations of ferromagnesian minerals and plagioclase is well known. The Fe-rich chloritic clay results from stepwise alteration of silicic glass (Steiner, 1968). Steiner (1968) also found that chloritic clay is commonly associated and intergrown with micaceous clays over a wide range of temperatures and pressures. This has been observed in thin sections of cuttings from well DF 45-14. Thus, the significance of chlorite as a means of evaluating the hydrothermal altuation in well DF 45-14 is minimal.

Thus, the analysis of clay minerals has proven to be most useful in defining the distribution and type of alteration in well DF 45-14. The limited and highly localized nature of the distribution of the clay minerals is strong evidence in support of the interpretation that hydrothermal alteration in DF 45-14 is fracture controlled and that the majority of the rocks encountered have not been significantly affected by the present system. Future work along these same lines will be pursued with the hope of refining the methods of analysis and the validity of the interpretations derived.

Water Chemistry

The relationship between the fluid chemistry and the alteration mineralogy in geothermal systems is typically complex and Dixie Valley is no exception. In studying the fluid composition two samples will be considered, DV-72 and DV-30, collected and analyzed by Desert Research Institute personnel. The DV-72 fluid is believed to be representative of the formation fluid below 6000 ft. (depth of casing) while DV-30 was taken from the flow of a shallow (1500') temperature gradient hole SR-1a. Tables 2 and 3 list the chemical analyses for these two samples.

Both samples have low to moderate total dissolved solids (TDS) contents when compared to geothermal fluids in other areas, suggesting the water is of meteoric origin and has achieved its chemical nature through interaction with host rock. This also may indicate a rather short residence time.

The presence of constituents such as lithium, fluoride and boron have been suggested in the past as indicating at least a partial magmatic origin of the fluids but abundant experimental work more recently has shown that concentrations of these elements can occur by rock alteration at moderate temperatures (Ellis and Mahon, 1977). The high boron content in the Dixie Valley water is not particularly uncommon for geothermal waters especially within high organic content rocks.

White (1957) has pointed out that within very dense formations sufficient rock/water contact could not occur to maintain the chemical output of a system over its lifetime. For the Wairakei system Grindley (1967) showed that over its lifetime (500,000 years) its water would have to have been completely replaced 100 times. Flow through formation of limited permeability (as in a tight, fracture-porosity formation) would be expected to rather rapidly (geologically) remove the available constituents from the exposed rock, so that it might be postulated for an area like Dixie Valley

TABLE 2. Distribution of species, sample DV-30(SR1a).

SPECIES		PPM	MOLALITY	LOG MOL	ACTIVITY
CA	2	.12096E+03	.30238E-02	-2.5194	.15347E-02
MG	2	.22001E+02	.90666E-03	-3.0426	.46847E-03
NA	1	.36560E+03	.15933E-01	-1.7977	.13370E-01
K	1	.18356E+02	.47034E-03	-3.3276	.39125E-03
H	1	.14996E-03	.14906E-06	-6.8266	.12882E-06
CL	-1	.57484E+03	.16245E-01	-1.7893	.13514E-01
SO4	-2	.19579E+03	.20420E-02	-2.6899	.10216E-02
HCO3	-1	.19462E+03	.31957E-02	-2.4954	.26977E-02
CO3	-2	.18287E+00	.30532E-05	-5.5152	.15505E-05
H2CO3	0	.43660E+02	.70525E-03	-3.1517	.71001E-03
OH	-1	.22192E-01	.13073E-05	-5.8836	.10863E-05
F	-1	.40840E+01	.21538E-03	-3.6668	.17898E-03
MGOH	1	.14983E-01	.36330E-06	-6.4397	.30913E-06
MGSO4 AQ	0	.14637E+02	.12183E-03	-3.9143	.12270E-03
MGHCO3	1	.20723E+01	.24333E-04	-4.6138	.20309E-04
MGCO3 AQ	0	.13993E+00	.16626E-05	-5.7792	.16746E-05
MGF	1	.71891E+00	.16631E-04	-4.7791	.13940E-04
CAOH	1	.35595E-02	.62471E-07	-7.2043	.52970E-07
CASO4 AQ	0	.56750E+02	.41765E-03	-3.3792	.42065E-03
CAHCO3	1	.17118E+02	.16965E-03	-3.7705	.14385E-03
CACO3 AQ	0	.13494E+01	.13508E-04	-4.8694	.13605E-04
NASO4	-1	.20044E+03	.16868E-02	-2.7729	.14240E-02
NA2SO4	0	.40590E+00	.28631E-05	-5.5432	.28837E-05
NAHCO3	0	.15755E+01	.18794E-04	-4.7260	.18929E-04
NACO3	-1	.11817E+00	.14264E-05	-5.8457	.12041E-05
NA2CO3	0	.13678E-03	.12930E-08	-8.8884	.13023E-08
NACL	0	.26163E+00	.44853E-05	-5.3482	.45175E-05
KSO4	-1	.82496E+00	.61151E-05	-5.2136	.51621E-05
KCL	0	.10157E-01	.13650E-06	-6.8649	.13748E-06
HSO4	-1	.47326E-02	.48848E-07	-7.3112	.40943E-07
H2SO4	0	.16478E-12	.16833E-17	-17.7738	.16954E-17
HCL	0	.19774E-08	.54338E-13	-13.2649	.54729E-13
H4SIO4AQ	0	.16706E+03	.17414E-02	-2.7591	.17539E-02
H3SIO4	-1	.89482E+00	.94265E-05	-5.0257	.78678E-05
H2SIO4	-2	.82756E-03	.88114E-08	-8.0550	.44745E-08
FE	2	.40000E-01	.71761E-06	-6.1441	.37093E-06
AL	3	.16146E-09	.59957E-14	-14.2222	.16130E-14
ALOH	2	.22338E-06	.50877E-11	-11.2935	.25836E-11
AL(OH)2	1	.23584E-03	.38738E-08	-8.4119	.32701E-08
AL(OH)4	-1	.34991E+00	.36899E-05	-5.4330	.30928E-05
ALF	2	.26696E-06	.58172E-11	-11.2353	.29541E-11
ALF2	1	.11578E-02	.17853E-07	-7.7483	.15071E-07
ALF3	0	.13201E-03	.15749E-08	-8.8027	.15862E-08
ALF4	-1	.10650E-04	.10363E-09	-9.9845	.86856E-10
ALSO4	1	.60138E-09	.48969E-14	-14.3101	.41045E-14
AL(SO4)2	-1	.10137E-09	.46353E-15	-15.3339	.38852E-15
H3BO3 AQ	0	.26723E+02	.43301E-03	-3.3635	.43612E-03
H2BO3	-1	.15788E+00	.26006E-05	-5.5849	.21405E-05
LI	1	.97064E+00	.14015E-03	-3.8534	.11983E-03
LIOH	0	.12605E-04	.52738E-09	-9.2779	.53117E-09
LISO4	-1	.64736E-01	.62970E-06	-6.2009	.52993E-06
SR	2	.32000E+01	.36591E-04	-4.4366	.18353E-04
SROH	1	.20513E-04	.19643E-09	-9.7068	.16531E-09
		TDS=1903			

field pH = 6.89
T = 64.5°C

TABLE 3. Disturbution of species, sample DV-72(DF 45-14).

SPECIES		PPM	MOLALITY	LOG MOL	ACTIVITY
CA	2	.95029E+01	.23762E-03	-3.6241	.11692E-03
MG	2	.57719E-02	.23793E-06	-6.6235	.11933E-06
NA	1	.54512E+03	.23764E-01	-1.6241	.19778E-01
K	1	.64499E+02	.16531E-02	-2.7817	.13631E-02
H	1	.93005E-06	.92471E-09	-9.0340	.79433E-09
CL	-1	.69970E+03	.19779E-01	-1.7038	.16309E-01
SO4	-2	.98586E+02	.10285E-02	-2.9878	.49850E-03
HCO3	-1	.18429E+03	.30270E-02	-2.5190	.25357E-02
CO3	-2	.29376E+02	.49061E-03	-3.3093	.24159E-03
H2CO3	0	.28203E+00	.45570E-05	-5.3413	.45893E-05
OH	-1	.71827E+01	.42326E-03	-3.3734	.34861E-03
F	-1	.94998E+01	.50113E-03	-3.3000	.41275E-03
MGOH	1	.13758E-02	.33369E-07	-7.4767	.28192E-07
MGSO4 AQ	0	.19401E-02	.16153E-07	-7.7918	.16275E-07
MGHCO3	1	.57027E-03	.66979E-08	-8.1741	.55429E-08
MGCO3 AQ	0	.88816E-02	.10556E-06	-6.9765	.10636E-06
MGF	1	.54064E-03	.12510E-07	-7.9027	.10400E-07
CAOH	1	.93138E-01	.16351E-05	-5.7865	.13762E-05
CASO4 AQ	0	.22763E+01	.16757E-04	-4.7758	.16894E-04
CAHCO3	1	.17827E+01	.17673E-04	-4.7527	.14875E-04
CACO3 AQ	0	.28855E+02	.28893E-03	-3.5392	.29112E-03
NASO4	-1	.31055E+03	.26143E-02	-2.5826	.21900E-02
NA2SO4	0	.35922E+00	.25345E-05	-5.5961	.25537E-05
NAHCO3	0	.21891E+01	.26121E-04	-4.5830	.26319E-04
NACO3	-1	.43281E+02	.52262E-03	-3.2818	.43779E-03
NA2CO3	0	.46604E-01	.44067E-06	-6.3559	.44401E-06
NACL	0	.46678E+00	.80045E-05	-5.0967	.80652E-05
KSO4	-1	.16541E+01	.12265E-04	-4.9113	.10274E-04
KCL	0	.42679E-01	.57371E-06	-6.2413	.57806E-06
HSO4	-1	.20911E-04	.21590E-09	-9.6657	.17948E-09
H2SO4	0	.30549E-17	.31216E-22	-22.5056	.31453E-22
HCL	0	.38133E-10	.10482E-14	-14.9796	.10561E-14
H4SIO4AQ	0	.18100E+03	.18673E-02	-2.7242	.19016E-02
H3SIO4	-1	.22409E+03	.23614E-02	-2.6268	.19542E-02
H2SIO4	-2	.70918E+02	.75532E-03	-3.1219	.37193E-03
FE	2	.10000E+00	.17946E-05	-5.7460	.90063E-06
H3BO3 AQ	0	.18335E+02	.29717E-03	-3.5270	.29943E-03
H2BO3	-1	.35981E+02	.59285E-03	-3.2271	.48330E-03
LI	1	.15343E+01	.22160E-03	-3.6544	.18651E-03
LIOH	0	.81229E-02	.33996E-06	-6.4686	.34254E-06
LISO4	-1	.49959E-01	.48610E-06	-6.3133	.40586E-06
SR	2	.31943E+00	.36537E-05	-5.4373	.17755E-05
SROH	1	.68053E-03	.65187E-08	-8.1858	.54426E-08
BA	2	.49932E-01	.36436E-06	-6.4385	.17706E-06
BAOH	1	.76786E-04	.49859E-09	-9.3023	.41628E-09
		TDS=2201			

field pH = 9.1
T = 76.5°C

that some volcanic additions are responsible for the fairly high boron, fluoride and arsenic levels reported. However, it must be considered that the water flow and composition has not been constant over time. Periods of appreciable inflow and outflow may occur over relatively short periods of time when tectonic activity creates new flow channels. Flow may decrease drastically when these channels become sealed again by deposition of minerals such as quartz or calcite, or are closed by deformations within the rocks. Waxing and waning of the geothermal flow system at Dixie Valley appears supported by evidence of increased flows of quite different character at many of the springs, such as the tufa domes at Sou Hot Springs. So while the origin of chemicals in the Dixie Valley water is not completely clear, local rock-water reactions may be the most important factor in determining well, spring, and fumarole fluid compositions.

The hydrothermal alteration at Dixie Valley includes a complex interaction of processes which include devitrification, recrystallization, solution and deposition. The end product of these processes is dependent upon numerous factors: temperature, pressure, fluid composition, rate of fluid flow, rock permeability and permeability type (fissure or interstitial porosity). In Dixie Valley the fluid composition, flow rate, and rock compositions and permeabilities vary considerably and contribute to the diversity in observed alteration. Because of these factors, the vertical zoning found in many geothermal fields, which is largely a function of temperature and pressure, is obscured at Dixie Valley. Permeability is a very important factor since many mineralogical changes are not isochemical and any fluid present must be free to move, allowing an influx and outflux of chemical components. In New Zealand it was found that dense impermeable rocks such as fine-grained sediments and ignimbrites were little altered by near-neutral pH solutions, even at high temperatures. Breccias and pumice, however, were altered to an equilibrium assemblage (Ellis and Mahon, 1977). This type of effect may be seen in the drill cuttings of well DF 45-14. Much of the very dense metasiltstone and shale is only weakly altered while tuff and andesite in the upper portions of the well have been extensively altered. Significantly, fracture zones (indicated by lost circulation during drilling, or by brecciation of the cuttings) within the siltstone show extensive alteration, as would be expected for areas of high permeability.

In considering alteration mineralogy and its relationship to fluid chemistry, the question of equilibrium is important. From experimental silicate reaction rates zones of slow water movement at temperatures above 200°C would probably produce an equilibrium mineral assemblage. In Dixie Valley flow rates are unknown and the down-hole temperatures estimated from SiO_2 and Na/K geothermometry are approximately 190-200°C.

In correlating alteration mineralogy to water chemistry the computer program WATEQ (Truesdell and Jones, 1974) has been used to provide activities and saturation index data as determined from the laboratory water analyses.

The following discussion is based upon the method of Helgeson (1969) for deriving phase diagrams expressing mineral stabilities by ion concentration ratios in solution. In the presence of quartz with aluminum assumed fixed, the formation of various silicates depends on the activity of metal oxides in solution. For example, the activity of CaO may be expressed as follows:

$$CaO + 2H^+ = Ca^{2+} + H_2O$$

$$K_{Ca} = a_{Ca} / (a_H)^2 a_{CaO} \; ; \quad a_{H_2O} \text{ assumed} = 1$$

Thus

$$\log a_{CaO} = \log (a_{Ca} / (a_H)^2) - \log K_{Ca}$$

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

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

$$\underset{\text{(K-spar)}}{3KAlSi_3O_8} + 2H^+ = \underset{\text{(K-mica)}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{(Qtz)}}{6SiO_2} + 2K^+$$

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

The diagrams which follow represent the thermodynamically stable phase for a given temperature and pressure. Figure 25 is calculated for a temperature of 100°C and is applicable to sample DV-30. This plot assumes quartz saturation but not amorphous silica saturation. Figures 26 and 27 are calculated for a temperature of 200°C and apply to DV-72. Figure 28 holds for a temperature of 260°C and is shown for comparison. The field shown for wairakite is probably replaced by laumontite at temperatures of 150-200°C (Ellis and Mahon, 1977). This figure also indicates the position of a calcite field which shifts with changing P_{CO_2}. It should be noted that a fluid composition within a stability field for a given mineral does not imply its presence or indicate any quantity which may be present, merely that it should be a stable phase for the given conditions. Other factors such as element availability, ability of a phase to nucleate and the kinetics of grain growth are also factors.

Tables 2 and 3 show the chemical breakdown of water samples DV 30 and DV 72. Only sample DV 72 was used for DF 45-14 as it is believed to be the most reliable (B. Bohm, personal comm.).

The points on the stability diagrams for DV 72 represent the ion ratios at the field pH (9.1) and at successively lower pH values. Because of the effects of degassing (especially CO_2) and/or boiling the pH at the production depth may be considerably lower than field pH. Evidence of CO_2 release has been found at certain springs in the valley.

A mechanism which explains the dramatic pH increase upon boiling and CO_2 loss has recently been proposed by Michels (1979). Waters from the East Mesa geothermal field in Southern California showed pH differences of 2.7 units (6.5-9.2) between flashed waters and those cooled without flashing. He proposes the following reactions:

$$2HCO_3^- \rightleftarrows H_2O + CO_2 + CO_3^{2-}$$

$$H_2O + CO_3^{2-} \rightleftarrows HCO_3^- + OH^-$$

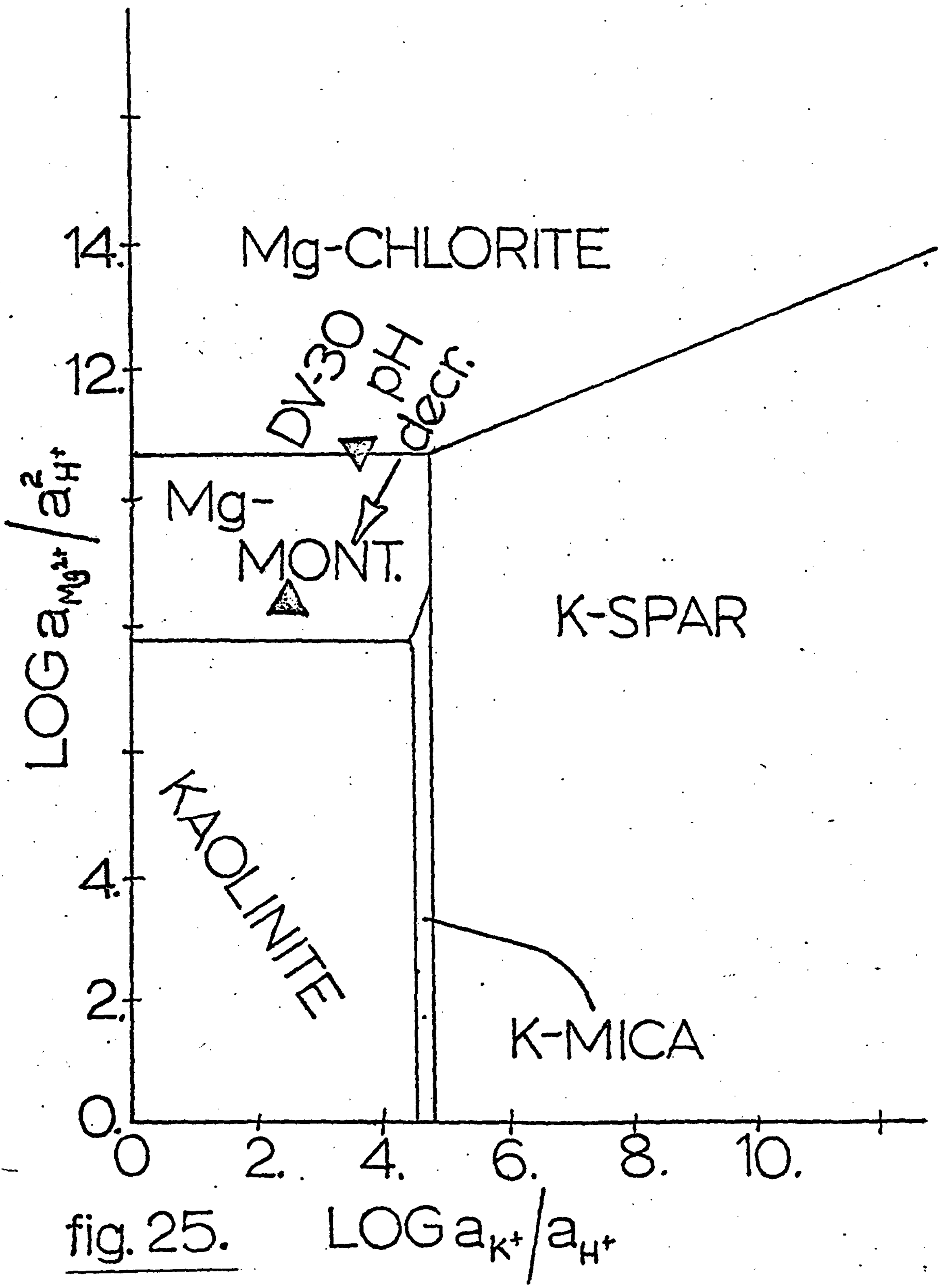


fig. 25.

PHASE STABILITIES AT 100°C.

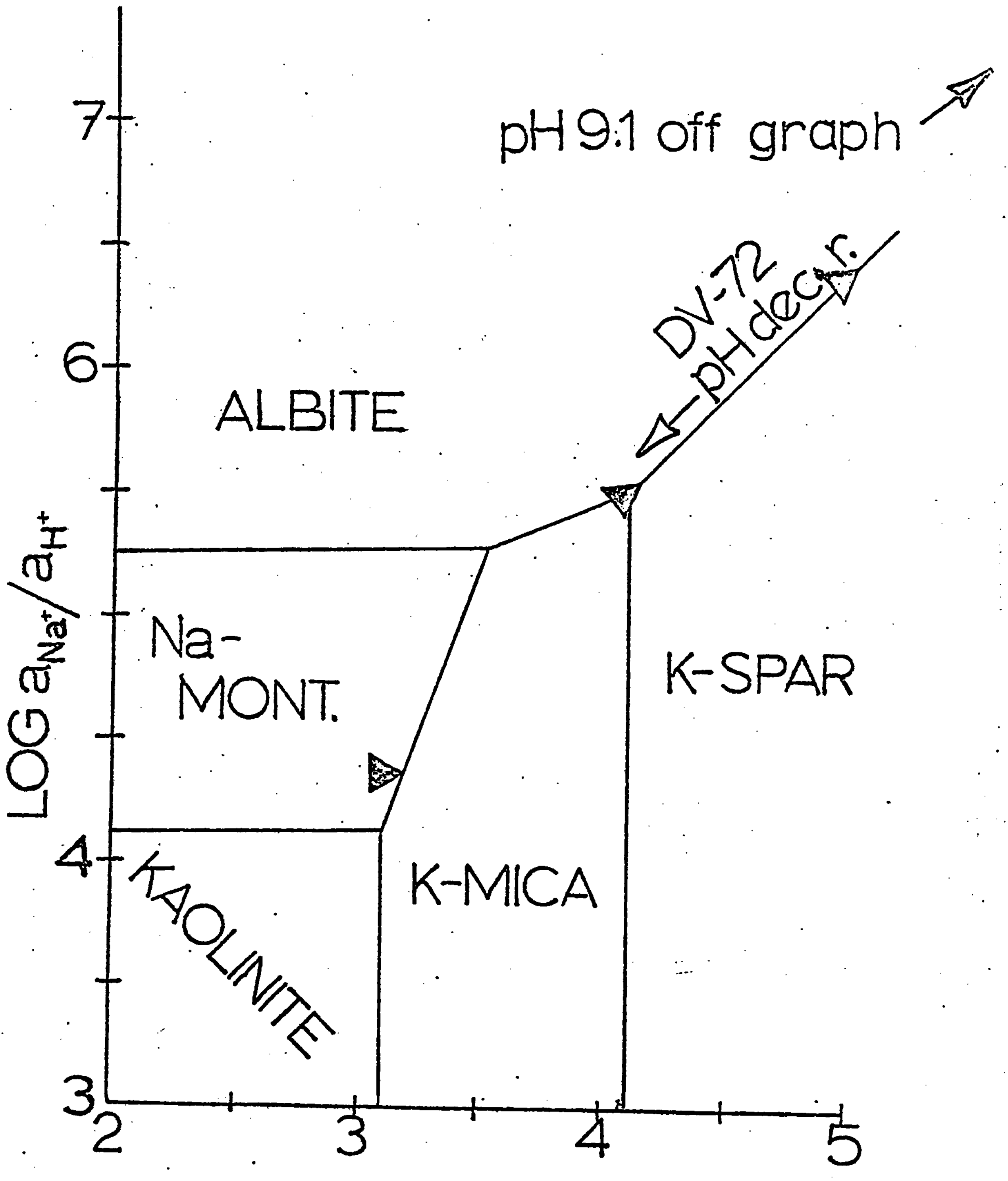


fig. 26. LOG a_{K^+}/a_{H^+}

PHASE STABILITIES AT 200°C.

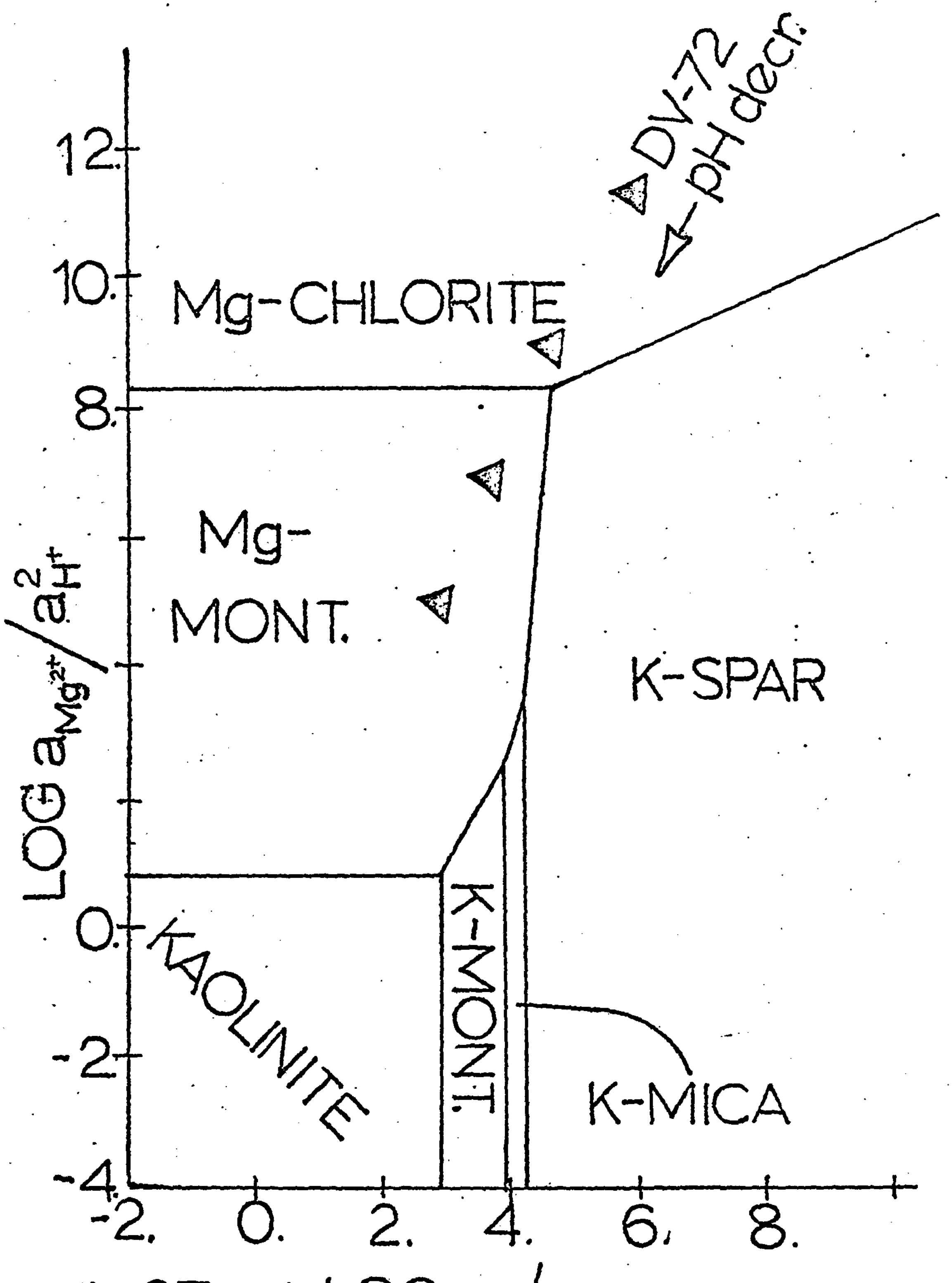


fig.27. LOG a_{K^+}/a_{H^+}

PHASE STABILITIES AT 200°C.

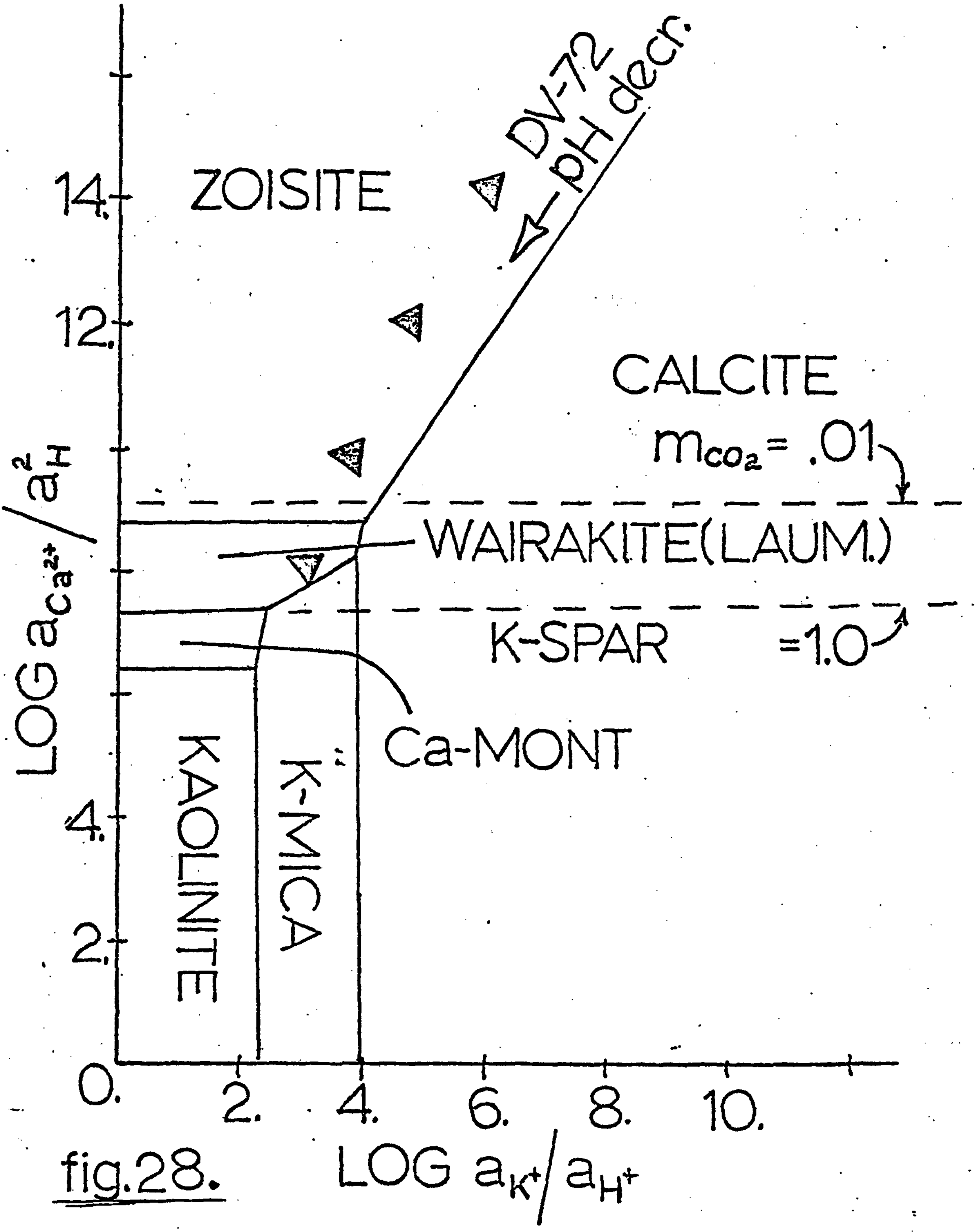


fig.28.

PHASE STABILITIES AT 200°C.

Michels points out that only about 1% of the CO_3^{2-} generated in the first step need be consumed to produce the observed pH values of 9.0+. This is possibly significant in regard to the mCO_3^{2-} values observed in DF 45-14 and supports the hypothesis that the downhole pH may be as low as 6.0

Of particular significance in viewing the diagrams is that neither water sample is near equilibrium with kaolinite which appeared prominantly as an alteration mineral through the lower 7000 feet of DF 45-14 and was clearly associated with postulated fracture zones. Considerably lower than field pH conditions would be required for kaolinite equilibrium.

Because of the uncertainty in the true down-hole pH it is not clear which phases represent an equilibrium assemblage. The observed mineralogy in DF 45-14 below 6000 feet is mainly K-mica, quartz, chlorite with laumontite and/or calcite. Potassium feldspar as adularia occurs in minor amounts in vein material; albite and epidote are extensive within the gabboric intervals. Kaolinite is abundant in two narrow intervals, again both related to fracture zones.

If sample DV 72 (pH = 9.1) truly represents the deep water conditions in DF 45-14, the assemblage quartz-albite-K-spar-chlorite and probably zoisite (epidote) and/or calcite (Figures. 26, 27) would represent an equilibrium assemblage. If a pH of 7 is assumed, the equilibrium assemblage becomes quartz-albite-K-spar-K-mica-chlorite-Mg-montmorillonite and probably epidote-laumontite and/or calcite. This is quite similar to what is observed in the less altered portions of DF 45-14. If a more acid fluid (pH=6) is assumed a more clay-mineral dominated assemblage is in equilibrium, which would include quartz, Na and possibly Ca-montmorillonite, K-mica, and probably laumontite and/or calcite. With a lowering of the pH of less than one unit kaolinite will become stable.

Two zones of alteration appear through the lower 5000 feet of DF 45-14: highly altered intervals with dominant K-mica-kaolinite, spatially associated with presumed fracture zones and a higher pH assemblage generally less altered and negatively correlated with fracture zones. This suggests acid conditions existing in the porous fractures while in less permeable areas more isochemical alteration is occurring.

The variation in observed alteration is also a function of mixing. At various times the water chemistry and flow through rock fractures may fluctuate in response to recharge conditions. These changes are also reflected by depositional periods when fracture permeability in a given area is reduced or destroyed. This latter effect is evidenced by quartz, calcite, adularia and laumontite veining. Rapid CO_2 loss through boiling within fissures is an important mechanism in vein formation.

Figure 25 for water sample DV 30 would indicate an equilibrium assemblage of Mg-chlorite-montmorillonite and possibly K-spar at measured field pH. If its pH were lowered by one unit it would be very near equilibrium with kaolinite-Mg-montmorillonite and K-mica.

The field pH is probably reliable for this sample due to its lower temperature. The assemblage observed in the alluvium tuffs and andesite above 1500 feet seems to be in approximate equilibrium with the waters of DV 30. Chlorite and montmorillonites are most abundant in this interval. The montmorillonites, particularly the Mg type, are found only in this upper portion of the drill hole. Figure 25 illustrates the expected equilibrium assemblage with regard to Mg and K activities. In the deeper portions of the well mixed layers of illite are the predominant clay minerals. An important shift occurs in the water chemistry between the deep and shallow samples. Calcium and magnesium greatly increase while sodium and potassium decrease. This would suggest alteration of Mg and Ca silicates to form K and Na clays which is confirmed (tentatively) by x-ray patterns of the upper portion of DF 45-14. At greater depths Ca and Mg are less abundant due in part to rock type and also because they may be held in chlorite and Ca-zeolites or calcite.

It would appear that the water of DV 30 is more nearly in equilibrium with the assemblage observed in thin section and x-ray scans for that depth than is the water of sample DV 72. The disequilibrium in the deeper portions of DF 45-14 is probably due to mixing between downward flowing groundwater and upward moving water that has been heated at depth. In particular the deposition of quartz veins must occur from solutions moving upward from presumably hotter areas where silica is dissolving. This is not occurring from the SiO_2 saturated DV 72 water. Also more acidic, CO_2-charged waters may be rising from deeper in the system affecting kaolinite alteration in the walls of fissures and depositing calcite as CO_2 is lost. This type of alteration is observed near 4000 feet in DF 45-14 and may represent a relict fracture zone. X-ray patterns show very high occurrences of K-mica and kaolinite. This association of calcite and characteristically acidic alteration (kaolinite) occurs also at approximately 5500 feet (range front fault?) and at approximately 8300-8600 feet in DF 45-14. This hypothesis of a more acid water source at depth could account for the high sulfate contents which are probably due to pyrite oxidation at a greater depth. Much of the pyrite within the shale and metasiltstone to 9000 feet is intact.

The effects and amounts of dissolved CO_2 in the DF 45-14 water is an important point that is not resolved. For a down-hole pH of 6.2 a P_{CO_2} value of .9 may be estimated. This value is consistent with the reported down-hole association of laumontite (Ca-zeolite) and calcite. Much lower values would probably facilitate the silica formation only while much higher values would probably precipitate calcite only (Figure 28). The possible evidence for large CO_2 loss strongly suggests boiling at some point; this hypothesis is further supported by the high carbonate ion content mentioned previously. This and other uncertainities will likely be resolved by more reliable water sampling in the future.

In considering the alteration effects by equilibrium activity plots it should be clear that this is not the whole picture. It is important to realize that only selected components have been considered hence only selected phases. We have considered in part the 7 component system $MgO\text{-}CaO\text{-}K_2O\text{-}Na_2O\text{-}Al_2O_3\text{-}SiO_2\text{-}CO_2\text{-}H_2O$. This implies that for a given temperature and pressure seven stable phases should exist. For example Muffler and White (1969) report for the Salton Sea geothermal field the above components (7 + H_2O) produce 7 phases: quartz, albite, K-spar, chlorite, K-mica, epidote and calcite. This approach, however, ignores other important components such as Fe, Mn and S. Iron is particularly important in Dixie Valley where vermiculite and/or biotite is encountered in much of the altered siltstone. Vermiculite has been described as a "trioctahedral member of the smectite group" by Deer, Howie and Zussman (1966). It seems likely that the lack of montmorillonite in the middle to lower portions of the well where it is predicted from equilibrium considerations is due to the formation of the very similar iron-magnesium mineral vermiculite. We do not have the thermodynamic data to support this at present so this must be regarded as a tentative conclusion. Further evidence of the role of iron in this system is the high iron contents encountered in much of the chlorite.

In considering any model for the alteration observed in Dixie Valley the chemical and structural heterogeneity of the rock must be considered. The extent of fluid mixing, rock/water contact and reservoir(?) surface communication are complexly controlled by tectonic and chemical factors. These effects have been tentatively interpreted here to provide a working hypothesis. More and better data should clarify the picture.

Summary of Evaluation of DF 45-15

The textural, structural and mineralogic relationships of the observed hydrothermal alteration in Df 45-14 reveal that the nature and occurrence of this alteration is strongly influenced by rock permeability and composition. Fluid migration both upward and downward is controlled by the fissure permeability of the metasedimentary unit. The hydrothermal system and the alteration it produces vary temporally, probably due to changing recharge conditions and tectonically induced changes in the fracture permeability. Self-healing may also be important in this regard. The overall hydrothermal alteration has been weak to moderate except in localized areas associated with fractures and lithologic contacts where the alteration is more intense.

The similarity of water chemistry between samples DV 30 and DV 72 suggests strong communication and interaction between the waters below 6000 feet in the metasediments and the overlying regime within the volcanic rocks and alluvium. This communication implies that the metasediments cannot function effectively as a thermal reservoir due to the lack of an overlying cap rock. The apparent limited permeability of the metasedimentary unit further argues against its being a productive reservoir. The fluid of sample DV 72 probably represents a mixture of surface water with much more acidic saline waters below 9000 feet. It may be that the metasediments penetrated by DF 45-14 are capping an underlying reservoir with which it has only limited communication (at present) through fractures. The intensity and distribution of the hydrothermal alteration indicates three main intervals which have at least some communication with a deep fluid source. These occur at approximately 5300, 6100 and 8600 feet. The occurrence of calcite and quartz veining probably represent 'fossils' of earlier fluid conduits.

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