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MACKAY MINERALS RESEARCH INSTITUTE  
ANALYSIS OF SHALLOW GRADIENT HOLES  
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## Foreword

Petrologic alteration studies of the six shallow thermal gradient holes (S-8, SR2, SR2A, DD-9, H-2, and H-1; Plate I) in Dixie Valley were undertaken as part of subcontract DE-AC08-79ET27006. The completion of this work by Mackay Minerals Research Institute personnel was accomplished primarily under the summer funding program sponsored by Southland Royalty Company. The data and conclusions in this report summarize the final evaluation of the shallow gradient holes with respect to the hydrothermal alteration in Dixie Valley. The results will be appended to the final report on the Geothermal Reservoir Assessment of Dixie Valley to enlarge and enhance the data base for developing a model of the Dixie Valley Geothermal System.

Mackay Minerals Research Institute personnel responsible for this report include: Graduate Research Fellows Tom Bard and Rus Juncal, and Principal Investigators D. Burton Slemmons, Elaine J. Bell and Lawrence T. Larson. Tom Bard was primarily responsible for the petrologic alteration analysis, with assistance provided by Rus Juncal. Technical guidance was provided by D. B. Slemmons. E. J. Bell and L. T. Larson supervised preparation of the following report and provided technical and editorial review. We acknowledge the valuable assistance of Keith Papke, Nevada Bureau of Mines and Geology, for his aid in developing techniques for clay mineral analysis and identification.

The following report provides a concise text supplemented by the necessary graphic display of data. To facilitate textual reading and the necessary referencing and cross-referencing of data, all figures are presented at the end of the report.

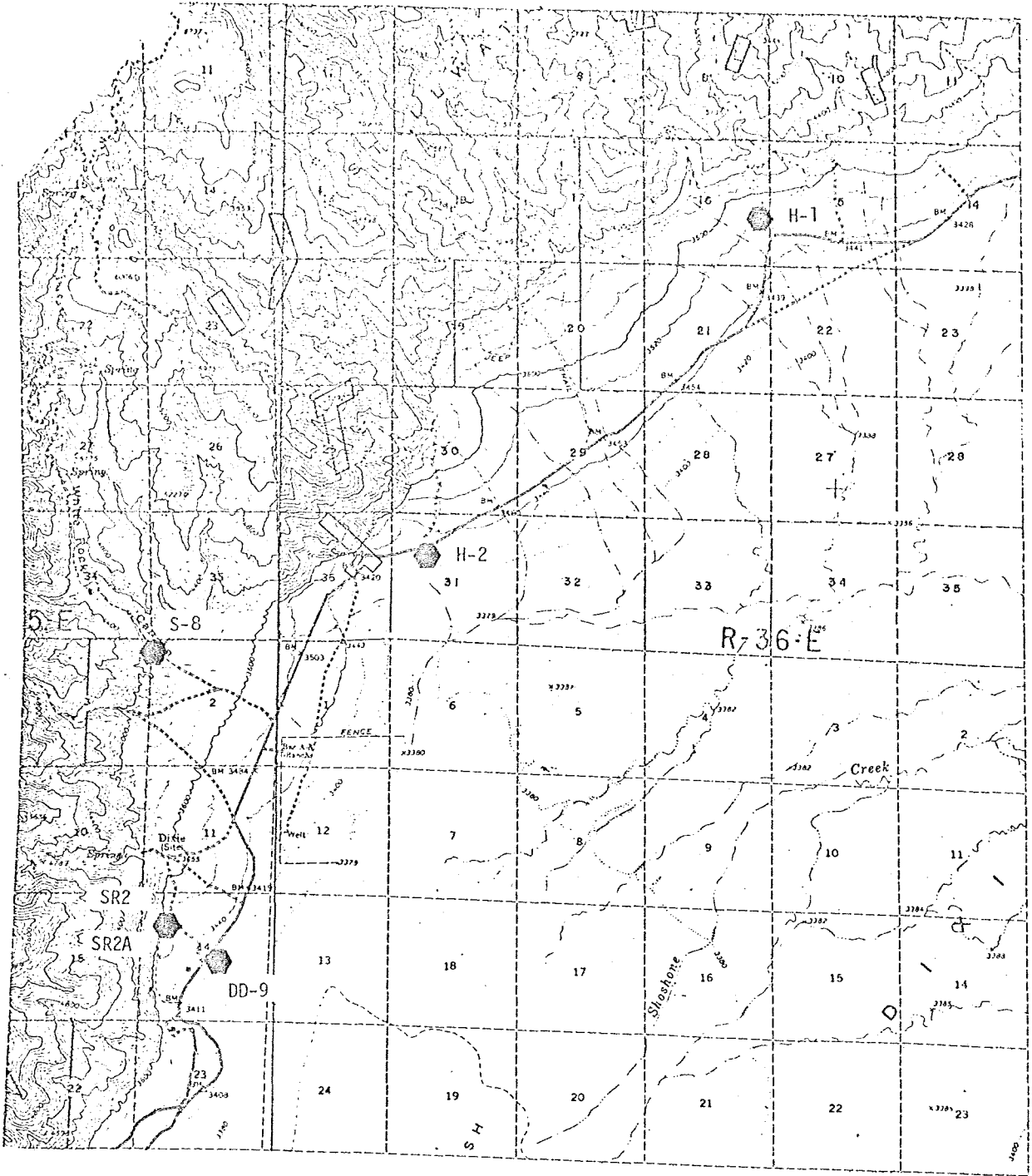


PLATE I. LOCATION OF SHALLOW GRADIENT HOLES IN DIXIE VALLEY, NEVADA



## I. Work Accomplished

The analysis and interpretation of drill cuttings from the six thermal gradient holes (S-8, SR2, SR2A, H-1, H-2, DD-9) are complete. The purpose of this investigation was to detect, describe and interpret the observed alteration effects as they relate to the present hydrothermal system in Dixie Valley. Interpretations regarding correlations between the thermal gradient holes are also presented. The results of this investigation are presented in Section III. Results. The locations of these thermal gradient holes are shown on Plate I.

Following an initial binocular examination of the drill cuttings to determine gross lithology from each of the wells, thin sections were made for each one hundred foot interval. These sections were then examined to determine specific lithologies, mineral relationships and distributions of the relative amount of alteration.

Selected samples analyzed by x-ray diffraction included 45 whole-rock samples and 36 oriented samples of the fine fraction (less than 8.0 phi; Folk, 1974). Twenty-two of the oriented samples were treated with ethylene glycol and heat, then rerun for confirmation of specific minerals. The diffraction patterns were interpreted to identify minerals and establish their distribution with respect to depth, temperature, and lithology.

An examination of available water chemistry data of the six thermal gradient holes, chiefly sample DV-30 from SR2A, is incorporated into the interpretation of the alteration mineralogy. Additionally, spring and other water samples near the gradient holes were assumed to approximate the water chemistry where specific samples were not available.

## II. Problems Encountered

Problems and difficulties in interpretation include the following:

- 1) distinguishing 'in situ' alteration from weathering and alteration that occurred prior to deposition of the alluvial sediments;
- 2) the heterogeneous nature of the alluvium and the differential responses of the various lithologies to the processes of weathering, transportation, and alteration by hydrothermal fluids;
- 3) lack of specific water chemistry data;

- 4) difficulties and complexities of the clay mineral analysis; and
- 5) the uncertainty of obtaining representative samples.

### III. Results

#### General Stratigraphy

Wilden and Speed (1974) described seven rock groups within the Dixie Valley region. These include: Upper Triassic metasediments; Middle Jurassic quartz arenite, mafic volcanic rocks, and gabbroic rocks; Miocene rhyolites; and Pliocene and younger sedimentary rocks, basalt, and andesite. A few lithologies present in the alluvial samples such as granite and aplite are distinctly atypical of the local Stillwater Range source area. The most abundant rock types are the silicic to intermediate volcanic rocks and the diorite/gabbro. Although outcrops of the Upper Triassic metasediments are present in the Stillwater Range, they are conspicuously rare or absent in the alluvial sediments encountered in the six thermal gradient holes.

Figures 1 through 6 depict the lithologic logs for the six thermal gradient holes, and Figures 7 through 12 show the distribution of various mineral phases and alteration as a function of depth, lithology and temperature. None of the six thermal gradient holes penetrated bedrock, although all of the wells, except H-2, encountered intervals of relatively homogeneous material, suggesting distinct lithologic units. Thermal gradient hole S-8 penetrated approximately 300 feet of intercalated lithic and vitric tuff (Figure 13-14) and andesite. This material is relatively fresh although alteration is apparent in much of the rock as evidenced by devitrification and clay mineral development. The x-ray diffraction data for S-8 are anomalous and will be presented in the discussion of the clay mineralogy. Thermal gradient hole H-1 encountered a homogeneous diorite/gabbro unit at approximately 1125 feet, continuing to a depth of about 1430 feet, where the homogeneity decreases sharply. The upper portion of the unit is more altered than the central portion. A gabbroic unit with an unusually high magnetite content was reported at approximately 1145 feet; and a metasedimentary unit was encountered from 1470 feet to total depth of the well at 1500 feet (Keplinger and Associates, 1978). The existence of this metasedimentary unit is not substantiated by analysis of the drill cuttings. Rather, alluvial sediments appear to increase in abundance below 1450 feet (Figure 6). Thermal gradient holes SR2, SR2A and DD-9 also penetrated intervals of increased homogeneity (Figures 2-4), although the amount of admixed

alluvium prohibits a definite conclusion that these intervals represent distinct lithologic units. Lack of contact metamorphic effects in the alluvial materials bounding the diorite/gabbro suggest that the unit was not intruded into the alluvium. The structural and stratigraphic relationship of this diorite/gabbro unit suggests that it may represent a landslide block although conclusive evidence is lacking.

#### Rock Types and Alteration Effects

The drill cuttings from the six thermal gradient holes contain various lithologies but exhibit a similar alteration assemblage. The mineral assemblage consisting of albite, sericite, chlorite, calcite, quartz and clay minerals along with iron oxides and hydroxides was present in nearly every sample examined. This particular assemblage is characteristic of the propylitic alteration described by Coats (1940) for altered volcanic rocks of the Comstock Lode. Coats attributes the origin of this propylitic alteration to hydrothermal alteration at relatively high temperatures.

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

Thus, the observed alteration mineral assemblage in the six thermal gradient holes is consistent with hydrothermal alteration products characteristic of other geothermal areas. However, this does not appear to be a result of activity presently occurring in the shallow alluvium within Dixie Valley. Rather, the alluvium appears to have acquired the mineral assemblage from a source area affected by an episode of hydrothermal alteration prior to erosion and deposition. Evidence for a previous period of intense hydrothermal alteration is found along the lower portion of the eastern front of the Stillwater Range in the immediate area of the thermal gradient holes. This zone of alteration is the source area for much of the altered alluvial sediments.

Albitization (determined by the Michel-Levy method) affects the plagioclase in the volcanic rocks and to a lesser degree in the diorite/gabbro. This process renders the plagioclase phenocrysts less translucent and the twinning becomes diffuse (figures 15-22). The existence of abundant albite is also supported by x-ray diffraction data from 45 whole rock samples of the alluvial material. Albitization is the process whereby the original Ca-rich plagioclase is converted to the Na-rich end member of the series: albite. This process releases Ca ions which are probably incorporated into the formation of epidote, calcite, and/or Ca montmorillonite, all of which were observed in the drill cuttings.

Figures 7 through 12 show the percent plagioclase alteration as a function of depth, lithology, and temperature. The alteration of plagioclase to sericite (illite or K-mica) + clay is pervasive and in some cases complete as it consumes entire crystals leaving only a ghost of the original crystal (Figures 15-22). The process of sericitization is thought to indicate alteration by alkaline solutions at higher concentrations of potash ( $K_2O$ ) (Turner and Verhoogen, 1970). Water samples from gradient hole SR2A (sample DV-30) indicate a relatively high concentration of this element; however, the measured pH is 6.9 indicating near-neutral conditions. On the other hand, the various surface and spring water samples have pH values ranging from 8.0 to 8.5 indicating more alkaline conditions. Sample DV-30 probably represents a mixture of deeper, more acid waters migrating upward with downward percolating alkaline near-surface waters. Hydrothermal alteration of plagioclase to sericite has been described at Santa Rita, New Mexico (Kerr and others, 1950).

Plagioclase is also altered to an epidote mineral (Figures 23-38). Some of the diorite/gabbro is saussuritized, although in many instances the epidote and/or clinozoisite/zoisite is coarsely crystalline. This replacement more often takes place in the cores of the plagioclase crystals. Replacement of plagioclase by clay and lesser calcite is commonly developed in the groundmass of the finer-grained intermediate volcanic rocks and has largely masked much of the original texture. Small phenocrysts of plagioclase often merge with the ground mass. The glassy matrix of some of the more acidic volcanic rocks now consists of chlorite, calcite, quartz, albite, and clay. Fine opaque granules disseminated throughout many of the rocks are commonly altered to limonite and hematite and probably represent magnetite derived from an earlier alteration episode.

The significance of the distribution of the percent plagioclase alteration as shown in Figures 7 through 12 is uncertain. No distinct pattern occurs that is related to depth, temperature or clay mineral distribution. In H-1 the plagioclase appears to be related to rock type (diorite/gabbro, Figure 12). In the interval 900 to 1200 feet, the percent plagioclase alteration steadily increases (Figures 29-30). The interval 1200 to 1400 feet shows a sharp decrease in plagioclase alteration indicating the hole penetrated fresh rock (Figures 31-32). From this evidence, it appears that the upper surface of this possible diorite/gabbro unit is highly altered whereas the internal portions of the unit are considerably less altered. None of the other wells however exhibit a distinct relationship between the percent plagioclase alteration and lithology -- due to the heterogeneity of the alluvial sediments. As mentioned above, the various rock types represented in the alluvial sediments respond differently to alteration and weathering. The volcanic rocks invariably show a greater degree of plagioclase alteration to sericite  $\pm$  clay and epidote than the diorite/gabbro. Any quantitative interpretation and/or correlation of the alteration mineralogy of the alluvial sediments between the thermal gradient holes or in any one of the holes would require extensive knowledge of the source rock and sediment composition -- such data is not presently available.

The alteration of the ferromagnesian minerals, including clinopyroxene, hornblende and biotite, is pervasive and generally complete except in intervals where a homogeneous rock unit is encountered (e.g., H-1, 1200 - 1400 feet). Original clinopyroxene in the diorite/gabbro is completely replaced by hornblende which sometimes occurs as large poikilitic crystals (Figures 33-34) and rarely as phenocrysts in the volcanic rocks. More often, hornblende and biotite are completely replaced by chlorite. Magnetite and lesser amounts of epidote and calcite are also derived from the alteration of hornblende. In the few samples in which hornblende does occur, it is usually the green pleochroic variety, although brown and integrated green and brown varieties are not uncommon. Much of the green hornblende appears bleached. This may be due to a loss of iron to the formation of disseminated and crystalline magnetite with which it is associated. The iron derived from the alteration of these minerals and pyrite is responsible for the pervasive limonite and hematite. The presence of these minerals indicates that a large amount of oxidation took place prior to deposition of the alluvium. Post-depositional oxidation, however, cannot be ruled out.

The distribution and amount of pyrite varies with regard to depth and temperature, as well as rock type. Where it does occur, it is commonly in the silicic volcanic rocks. In some of the holes (e.g., SR2A, 280 - 300 feet; DD-9, 500 - 560 feet) it is quite abundant, but may also be present in only minute amounts, or completely absent. Conclusive evidence for secondary pyrite (i.e., pyrite on limonite) has not been found in the alluvial sediments.

Vein and fracture fill material in the alluvial sediments consist predominantly of calcite and quartz with lesser amounts of epidote and possibly laumontite, although this has not been confirmed by x-ray diffraction. The Ca-zeolite laumontite is fairly common in other geothermal areas (Steiner, 1968; Browne and Ellis, 1970; Ellis and Mahon, 1977). The Ca for the laumontite may have been derived from the alteration of hornblende or Ca-rich plagioclase. The volcanic fragments generally exhibit more veining, although veins in the diorite/gabbro are not uncommon (Figures 35-40). The formation of these veins occurred prior to deposition. The alluvium shows no sign of cementation by crystalline and/or amorphous quartz or calcite as has been described elsewhere, such as in Roosevelt Hot Springs, Utah (Ballantyne, 1978). This is further evidence that the effects of the present hydrothermal system on the alluvial sediments in Dixie Valley are minimal and that most of the observed alteration is a product of a prior period of hydrothermal alteration before deposition of the sediments.

A phenomenon that is observed in certain intervals of gradient holes SR2, SR2A, S-8 and DD-9 (Figures 1-4) and which may be an indication of present or recent hydrothermal activity is the occurrence of aggregates of felsic and lithic fragments loosely cemented by  $\text{CaCO}_3$ . These completely disaggregate with the application of dilute HCl. The available water chemistry data indicate that calcite is essentially in equilibrium, or only slightly saturated in sample DV-30. The data also show that Ca and  $\text{HCO}_3^-$  ions are quite abundant (Table 2, Water Chemistry). Data for the spring and surface water samples show them to also be slightly saturated with respect to these elements. Precipitation of calcite may be governed by the reaction:  $\text{Ca}^{++} + \text{HCO}_3^- = \text{CaCO}_3 + \text{H}^+$ . Thus, given a high concentration of  $\text{HCO}_3^-$  and Ca, the precipitation of calcite will be favored. In sample DV-30 for SR2A, this appears to be the case and the  $\text{CaCO}_3$  cemented aggregates do occur. However, there is no conclusive evidence for post-depositional hydrothermal activity with the possible exception of the development of certain clay minerals, particularly the montmorillonite clays.

## Clay Mineralogy

Figures 7 through 12 show the distribution and relative amounts of the various clay minerals as a function of depth, lithology and temperature, as well as percent plagioclase alteration for each of the thermal gradient holes. Relative clay abundance is here interpreted as being proportional to the peak intensity; this can also be influenced by a number of other factors, including: orientation, crystallinity, type of clay, and sample reproducibility. Thus the conclusions and interpretations presented below must be viewed with discretion.

Overall, a conspicuous correlation between the six thermal gradient holes is lacking with respect to clay mineral zonation. This is probably due to the nature of the alluvial sediments and the lack of correlative horizons, and the lateral and vertical variation within the sediments. However, in SR2 (490 feet), in SR2A (700 feet), in S-8 (450 feet), and possibly in H-1 (800-900 feet) there appears to be a rough correlation with specific clay mineral abundance and distribution. These zones of lost circulation (Figures 7-9, 12) are marked by inversions in the temperature gradients which in other geothermal areas, such as Wairakei (Steiner, 1968), indicate fault or fracture zones.

The clay minerals which were identified by x-ray diffraction are kaolinite, illite, Na and Ca and/or Mg montmorillonites, mixed-layer illite/montmorillonite, chlorite, and an anomalous occurrence of what is referred to as 'vermiculite-like' and mixed illite/vermiculite clays (K. Papke, pers. commun., 1979; Table 1). As mentioned above, the variable chemical and structural composition of clay minerals makes identification very difficult. X-ray diffraction analyses must be supplemented with other techniques such as heat treatment and the use of organic liquids. However, even these confirming tests are not conclusive. For instance, heat treatment of a sample at 560°C for one-half hour destroys the structure of kaolinite, but it has been reported that the structure of chlorite may also collapse. Because both have a reflection at 7.1Å, their identification is uncertain, especially if the 14Å peak for chlorite is present.

The utility of argillaceous alteration products for describing zonation due to hydrothermal action has long been recognized. Many ore deposits have been discovered by the argillaceous alteration halo around them. Although complex, the mechanism of clay mineral alteration is similar for hydrothermal action associated with a geothermal system. The study of the alteration products

Table 1. X-ray Diffraction Data

<u>Mineral</u>	d (001) Spacing in Angstroms		
	Untreated	Glycolated	Heat Treated
Na-montmorillonite	12.0-12.5	16.7-17.5	9.8-10.1
Ca and/or Mg montmorillonite	14.0-14.5	17.0-17.4	9.8-10.1
Mixed Na-Ca-Mg Montmorillonite	12.6-13.5	16.4-17.4	9.8-10.1
Mixed illite-montmorillonite	10.16 +	variable	9.8-10.1
Illite	9.8-10.16	no change	no change
Kaolinite	7.0-7.25	no change	no peak
Chlorite	13.75-14.3	no change	intensified
Mixed Illite/Vermiculite	10.2-14.2	slight increase	collapse to 9.9-10.1



in the alluvial sediments of the thermal gradient holes is complicated by the difficulty of distinguishing between hypogene and supergene products. It is quite probable that the two are superimposed upon one another. Kaolinite, for example, in some instances may be formed as a hypogene product and later by supergene alteration of some other clay minerals (Grim, 1968). Steiner (1953, 1968) studied in detail the hydrothermal rock alteration of Wairakei in New Zealand. He found a surficial zone of kaolinite and alunite which grades into a zone dominated by smectite clay which in turn gives way to zeolite and adularia. At greater depths, sericite is formed. Steiner concluded that the initial hydrothermal agents are acidic and become neutralized and then alkaline as they proceed to the surface. The surficial kaolinite is believed to be in part produced by supergene alteration. These zones are temperature dependent. Thus, an understanding of the clay minerals can be an important factor in determining the character of a hydrothermal system.

#### S-8

The clay mineral content of gradient hole S-8 (Figure 7) is unique compared to the others because of the occurrence of what is believed to be vermiculite and mixed-layer chlorite/vermiculite and/or illite/vermiculite. Hayashi and others (1961) identified smectite, vermiculite, and randomly interstratified combinations of chlorite-vermiculite and regular interstratified combinations of smectite-vermiculite in basaltic rocks from Japan. S-8 is also unique with regard to the rocks encountered as the hole only penetrates about 100 feet of alluvium and then remains in intercalated silicic to intermediate volcanic rocks to total depth (Figure 1). The location of S-8 is very close to the range front fault and the visible zone of intensely altered rocks along the range front (Plate 1).

Figure 7 shows the sharp increase in vermiculite-like clay (100-200 feet) which is directly related to the transition from alluvial sediments to an intercalated tuff-andesite sequence. The vermiculite is probably derived from the alteration of mafic minerals (biotite, hornblende) as evidenced by the absence of these minerals in the rocks. It is quite likely that the vermiculite-like clay is actually randomly interstratified mixed-layer chlorite/vermiculite or possibly randomly interstratified illite/vermiculite.

Illite, as well as kaolinite and chlorite, is more abundant in the upper 100 feet and probably represents the influence of the alluvial sediments. The relative amounts of these clay minerals drop off where the volcanic rocks are

encountered, indicating the strong influence of lithology on the distribution and abundance of the clay minerals. From 100 to 400 feet, the relative amount of the vermiculite-like clay mineral gradually decreases. Kaolinite and illite roughly follow a similar pattern whereas the distribution of chlorite is relatively uniform throughout this interval (100-400; Figure 7).

The interval 400 to 500 feet in S-8 is significant because a zone of lost circulation has been documented (450-475 feet). The evidence for a rapid change of conditions in the hole is reflected in the clay mineralogy. This zone of lost circulation is interpreted here as representing the intersection of the hole with the eastward dipping range front fault. By assuming a 60 degree dip on the fault, the calculated position of the well is approximately 250 to 275 feet from the fault (Plate 1). This fault may be a conduit for thermal fluid migration. The dramatic increase in the relative amount of a mixed-layer illite-vermiculite (illite dominant) in this 400 to 500 feet interval is most likely the result of the higher temperatures encountered during active hydrothermal alteration. Although the bottom hole temperature at the present time is relatively low (48°C), the presence of abundant clay in the zone indicates that the intensity of hydrothermal activity has varied through time. Illite and mixed-layer illite-montmorillonite in Imperial Valley (Hoagland, 1976), Wairakei (Steiner, 1968), Ohaki-Broadlands (Browne and Ellis, 1970) and Reykjanes, Iceland (Tomasson and Kristmannsdottir, 1972) indicate higher temperatures than does montmorillonite. Thus, the occurrence of illite or illite in mixed-layer minerals is indicative of higher temperatures in thermal gradient hole S-8.

The distribution of kaolinite and chlorite in S-8 does not reveal any distinct pattern aside from those noted above. Kaolinite does show a gradual decrease with depth and its relative abundance is antithetical to that of illite and mixed-layer illite-vermiculite. The significance of this observation is uncertain, however it is probably related to fluid migration and temperature regimes and possibly rock type.

Thus, in thermal gradient hole S-8, the distribution and intensity of the observed alteration effects, particularly the clay mineralogy, is related primarily to structural controls, i.e., faulting. Lithologic variation is a secondary factor influencing the clay mineral distribution.

#### SR2

The clay mineral distribution in gradient hole SR2 (Figure 8) is quite

similar to that observed in S-8. Both holes are located very close to the range front fault, approximately 280 and 250 feet, respectively (Plate I). Both drill holes encountered zones of lost circulation: 450 feet in S-8; 490 feet in SR2. Lithologically and mineralogically, however, the holes are different. As Figure 2 shows, SR2 encountered predominantly alluvium which in certain intervals (150-275 feet and 450-500 feet) contains abundant volcanic rock suggestive of distinct lithologic units or volcanic gravels. None of the vermiculite-like clays were identified in SR2. The Na-montmorillonite and chlorite both show a sharp decrease in relative abundance from 300 to 400 feet and from 400 to 500 feet. The interval containing the zone of lost circulation (400-500 feet) is not marked by an increase in any of the clay minerals. This may be due to experimental error, lithologic differences, or more likely, to differences in fluid migration activity along the fault zone. The distribution of montmorillonite is probably a function of temperature. Montmorillonite is indicative of lower temperatures in other geothermal areas as noted above, and this appears to be the case here in SR2. Additionally, the interval 200 to 300 feet in SR2 roughly corresponds to a transition from alluvial sediments to predominantly silicic to intermediate volcanic rocks, possibly a lithologic unit. This transition is marked by a dramatic increase in the relative amount of montmorillonite, and to a lesser extent kaolinite, illite, and chlorite. The clay mineral distribution appears to be controlled largely by lithology in SR2.

Kaolinite decreases with depth from 200 to 500 feet in SR2 (Figure 8). This is suggestive of supergene conditions since upward migrating fluids leach pyrite, and as  $H_2S$  is oxidized, the fluids become more acidic and more conducive to the formation of kaolinite. This process has not been substantiated. Rather the presence of kaolinite may be a relict from a previous episode of more acidic hydrothermal alteration. Alternatively, lithology could be the controlling factor; that is, as the amount of volcanic material decreases, so does the amount of argillaceous alteration products.

#### SR2A

The pattern of clay mineral distribution in thermal gradient hole SR2A is also difficult to interpret. The distribution of the clay mineral species for SR2A is shown in Figure 9. The interval 700 to 800 feet appears to be significant, as drilling encountered a zone of lost circulation. This zone is interpreted as the intersection with the range front fault; SR2A is approximately 450 feet from the fault zone (Plate I). Lithologically, this 700

to 800 foot zone is marked by an increase in diorite/gabbro fragments (Figure 3). The amount of intermixed alluvial sediments and the altered appearance of all the diorite/gabbro fragments suggests that this is not a bedrock zone. However, the observed alteration may be related to the permeability associated with the contact surface between alluvium and a possible diorite/gabbro intrusive. Thus, the alteration observed in the interval 700-800 feet is influenced by lithologic and/or structural controls. Elsewhere in SR2A the clay mineral abundance is largely controlled by lithology.

The distribution of mixed-layer illite-montmorillonite increases toward the zone of lost circulation (Figure 9) suggesting an increase in temperature. The decrease in the relative amount of mixed-layer illite-montmorillonite below 800 feet may be related to a temperature inversion associated with the zone of lost circulation. Lithologic influences cannot be ruled out. Nevertheless, the temperatures and degree of hydrothermal activity have probably been much higher in the past in SR2A.

The large concentration of Ca and/or Mg-montmorillonite in the interval 200 to 400 feet (Figure 9) corresponds to the occurrence of abundant aggregates of lithic and felsic fragments loosely cemented by  $\text{CaCO}_3$  and clay. From 500 to 800 feet, all the clay minerals exhibit relative increases in abundance. In the interval 700 to 800 feet, increases in Na-Ca-Mg montmorillonite occur with mixed-layer illite/montmorillonite showing a less pronounced increase. This observation is consistent with the interpretation that this interval in SR2A represents the intersection with the range front fault. The presence of the fault, acting as a conduit for migrating thermal fluids, would explain why this zone exhibits a greater degree of alteration.

#### DD-9

The alteration effects observed in drill hole DD-9 are related primarily to lithology, and possibly to lithologic contacts which serve as fluid conduits. The variations of the alteration effects as they relate to lithology are best exhibited by montmorillonite (Figure 4, 10). The interval 700 to 800 feet and 1100 to 1200 feet which are marked by large increases in volcanic material (possibly distinct units) clearly show this relationship. The response of the other clay minerals to lithologic changes is not as distinct. Ca and/or Mg montmorillonite predominate in the upper portions of the hole and give way at depth to Na-montmorillonite. Mixed-layer illite-montmorillonite occurs at the bottom of the hole (1300-1400 feet) probably reflecting the higher pressures and temperatures of its stability range. The reason for the change from Ca and/or Mg-montmorillonite to Na-montmorillonite is unclear,

although it must be related to water-rock interaction.

No zones of lost circulation have been documented in DD-9. The range front fault quite likely is much deeper corresponding to an increased distance from the Stillwater Range (1/2 mile, Plate 1). Such an intersection is projected to occur at about 4700 feet depth in the vicinity of DD-9.

#### H-2

The clay mineralogy in H-2 is also basically controlled by lithology. The rocks in H-2 consist primarily of alluvial sediments (Figure 5) and the clay minerals exhibit a fairly uniform but nondiagnostic pattern in their distribution (Figure 11). The clay minerals present cannot be correlated with specific parameters, such as changes in lithology or lost circulation. Only one interval, from 900 to 1000 feet, appears anomalous due to the appearance of a mixed-layer illite/montmorillonite mineral and a relative increase of all clay minerals, although this may be due to experimental error. The significance of this interval is uncertain. The observed clay mineral distribution appears to confirm that H-2 penetrated alluvial sediments throughout its total depth of 1500 feet.

#### H-1

The distribution and relative abundances of the clay minerals in thermal gradient hole H-1 are illustrated in Figure 12. The clay mineralogy in the upper half of the hole is not very diagnostic. As in DD-9 and SR2A, Ca and/or Mg-montmorillonite predominate in the upper portions and then give way to Na-montmorillonite at depth. The increase in all the clay minerals in the interval 800 to 900 feet coincides with the postulated intersection of the drill hole with the range front fault (Plate 1). The large increase in kaolinite in this interval may be related to upward migrating acidic thermal waters. However, the conditions for the formation of montmorillonite and kaolinite are quite different suggesting that one of these minerals is out of equilibrium with the present system, being a relict of a previous episode of hydrothermal activity.

The interval 1200 to 1300 feet in H-1 corresponds to the top of a diorite/gabbro unit (Figures 6, 12). A relative increase in Na-montmorillonite is probably related to the contact surface between the diorite/gabbro and the overlying alluvium. Figures 29-32 support the existence of a homogeneous unit from 1200 to 1450 feet, the outer portions of which are highly altered. Additionally, the temperature gradient for H-1 shows a slight inflection at the supposed contact. This may be due to the differences in conductive proper-

ties between the alluvium and the diorite/gabbro, or possibly a change from convective heat flow in the diorite/gabbro to conductive heat flow in the alluvium.

#### Water Chemistry

Specific water chemistry data is available only from thermal gradient hole SR2A; Table 2 lists the chemical composition of this water sample (DV-30). However, some assumptions are made based on chemical data from spring and surface waters collected at or in the vicinity of the six thermal gradient holes.

Sample DV-30 is relatively low in total dissolved solids (TDS = 1903; Table 2) as compared to other geothermal areas: 21,380 to 32,250 in Imperial Valley (Hoagland, 1976). This suggests that the water is largely of meteoric origin and has arrived at its present chemical composition through interaction with the host rock. The low TDS is also suggestive of a rather short residence time for the water. The presence of constituents such as lithium, fluorine, and boron, however, have been suggested as indicating at least partial magmatic origin of fluids. Continuous rock-water interaction may eventually deplete the available constituents in a formation. On the other hand, these elements (Li, F, B) may be derived from rock alteration at moderate temperatures (Ellis and Mahon, 1977). Experimental evidence suggests that these elements may also be released by alteration of organic-rich materials, such as the meta-sediments in Dixie Valley.

The origin of chemicals in the water is obscure, but local water-rock reactions are probably the most important factor determining the fluid compositions. However, these in turn are influenced by the water flow and compositional variations over time, probably as a result of tectonic activity which may create or destroy the fluid migration channels. Deposition, solutions, and tectonic adjustments with the rock are all factors which influence the flow channels. In the case of thermal gradient holes S-8, SR2, SR2A and H-1, the range front fault has served as a conduit for migrating thermal fluids. The unconsolidated nature of the alluvial sediments suggests that fluids could also ascend through the porous materials with the rate dependent on permeability.

The dynamic character of the geothermal flow system in Dixie Valley is evident in the variety of flow rates and fluid compositions of the many springs, surface waters, and wells in the area. Additionally, the alteration effects in the thermal gradient holes are supportive of a changing system as different periods of activity, characterized by a certain alteration assemblage, may be

TABLE 2. Distribution of species, sample DV-30

SPCIES		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	.16622E-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	.18029E-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	-5.3488	.45175E-05
KSO4	-1	.82496E+00	.61151E-05	-5.2135	.51621E-05
KCl	-0	.10157E-01	.13650E-06	-6.8649	.13748E-06
HSO4	-1	.47322E-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
H4SiO4 AQ	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
H3PO3 AQ	0	.26723E+02	.43301E-03	-3.3635	.43612E-03
H2PO3	-1	.15788E+00	.26006E-05	-5.5849	.21405E-05
LI	1	.97064E+00	.14015E-03	-3.8534	.11883E-03
LiOH	0	.12605E-04	.52738E-09	-9.2779	.53117E-09
LISO4	-1	.64736E-01	.62970E-06	-6.2009	.52993E-06
SO	2	.32000E+01	.36591E-04	-4.4366	.18353E-04
SDOH	1	.20513E-04	.19643E-09	-9.7068	.16531E-09

TDS=1903

field pH = 6.89  
T = 64.5°C

superimposed upon one another. Thus the relationship between the fluid chemistry and the alteration mineralogy in Dixie Valley is quite complex.

Various processes of hydrothermal alteration are active in Dixie Valley, including: solution, deposition, recrystallization, devitrification and ion-exchange reactions. The end product of these interactive processes is a function of temperature, fluid composition, rate of fluid flow, and permeability. Variations in these factors contribute to the diversity in observed alteration. In the heterogeneous unconsolidated alluvial sediments, the influence of each of these factors is difficult to determine and vertical zoning, common to other geothermal areas, is obscure. Permeability is especially important as evidenced by the areas of intense alteration associated with the fracture or fault zones marked by zones of lost circulation.

In dealing with the relationship between the alteration mineralogy and fluid chemistry, the concept of equilibrium must be incorporated. The attainment of equilibrium is dependent on a number of factors: the composition and properties of the fluid and rock, the rate of fluid flow, and temperature and pressure. Temperatures in the thermal gradient holes range from a low of 24°C in S-8 to 97°C in H-1. Figure 41 represents the thermodynamically stable phases for DV-30 for a temperature of 100°C. Although the bottom hole temperature in SR2A is only 92°C, the diagram is a good approximation of the existing conditions, with regard to Mg and K activities. The plot assumes quartz saturation.

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. Element availability, diffusivity, ability of a phase to nucleate, and the kinetics of grain growth are also factors influencing the existence of a particular phase.

Figure 41 indicates an equilibrium assemblage of chlorite Mg-montmorillonite and possibly K-feldspar at the measured field pH. A decrease in pH would bring the assemblage kaolinite-Mg-montmorillonite and K-mica into equilibrium. The effect of a decrease in temperature on the stability fields is uncertain. This plot (Figure 41) has only used selected components of a multi-component system ( $\text{MgO-CaO-K}_2\text{O-Na}_2\text{O-Al}_2\text{O}_3\text{-SiO}_2\text{-CO}_2\text{-H}_2\text{O}$ ), thus limiting the number of possible phases according to the phase rule. Iron should probably be included to allow for vermiculite as a possible phase. Given the above components,



this implies that for a given temperature and pressure, seven stable phases should exist. Muffler and White (1969) in the Salton Sea geothermal field found the above components producing the following phases: quartz, albite, K-spar, chlorite, K-mica, epidote and calcite. This same assemblage, minus the K-spar, is found in the rocks of all the thermal gradient holes suggesting a state of relative equilibrium.

The mineral assemblage observed in SR2A appears to be in approximate equilibrium with the water of sample DV-30 according to Figure 41. The x-ray diffraction data indicate that Ca-Mg montmorillonite and chlorite are most abundant, along with lesser kaolinite and illite. The Ca-Mg montmorillonites give way at greater depths to Na-montmorillonite. This change was generally observed in all the thermal gradient holes and is likely related to the fluid temperature and the availability of ions. The formation of montmorillonite is favored under alkaline conditions which do exist at shallower depths in the thermal gradient holes. This is generally compatible with the observed distribution of this mineral being more abundant in shallower depths. Water chemistry data of DV-30 reveal that Na, Ca, and Mg-montmorillonite are all highly saturated, as is kaolinite, K-mica, and chlorite. Thus the conditions for the precipitation of these minerals is favored, but not necessitated. Although the field of kaolinite is slightly out of equilibrium with the present conditions (Figure 41), the amount of kaolinite observed may be a relict feature related to periods of increased hydrothermal activity when conditions were more acidic. This supposition may be applied to the K-mica as well.

The pH values for the various spring and surface water samples are invariably higher (7.5 - 8.5) than that recorded for DV-30 (6.89). This is suggestive, although not conclusive, that mixing of the alkaline surface waters with deeper, more acidic waters is occurring. The mineralogy of SR2A, compared with the other drillholes is not significantly different to substantiate this process in the upper 1500 feet of alluvium. If water sample DV-30 can be interpreted as being representative of water in the other thermal gradient holes, with the possible exception of S-8, it would be compatible with the similar mineral assemblage found in the drill holes.

Possible indications of the conditions necessary to form the various minerals observed is found in an examination of the geochemistry of argillic and related types of rock alteration in some ore deposits. For instance, in acid solutions containing potassium and aluminum, the feldspars alter to sericite

at high temperatures and to kaolinite at lower temperatures, but change to zeolites at temperatures from 60°C to 550°C in alkaline solutions undersaturated with respect to silica. Magnesium and ferromagnesian silicates of the pyroxene, amphibole, and chlorite groups are stable in alkaline solutions, but are readily decomposed by hot acid solutions (Lovering, 1950). On the other hand, mildly alkaline solutions, such as those found in the spring and surface waters of Dixie Valley, are relatively ineffectual in causing decomposition of the common rock silicate minerals (Lovering, 1950).

Following the above line of reasoning, the alteration assemblage in the six thermal gradient holes has been derived from acidic solutions. However, these acidic solutions presently do not exist, except at much greater depths. Hence, the alkaline shallow surface and spring water is dominated by water of meteoric origin which is not affecting a significant amount of alteration in the drill holes. The low pH of the water in DV-30 is probably derived by mixing of meteoric water with deeper, more acidic waters.

The relationship of the water chemistry to the observed alteration is complicated by the heterogeneity of the rocks encountered. The amount of fluid mixing, the rates of fluid flow, and the water and rock compositions also make conclusive interpretations difficult. The textural and structural relationships of the observed mineral assemblage and the available water chemistry data indicate a dynamic hydrothermal system characterized by alteration by acidic solutions. However, evidence of the adjustment of the mineral assemblage to changing conditions is lacking.

#### Summary of Evaluation of Thermal Gradient Holes

Only generalized interpretations can be made regarding the temperature-depth relationships of the clay mineralogy. Montmorillonite clays are indicative of near-surface alteration by alkaline solutions at lower temperatures. In zones of higher temperatures, these clays grade into interstratified mixed-layer illite-montmorillonite clays. The relationship of the distribution of Ca and/or Mg montmorillonite and Na-montmorillonite is uncertain, but is probably locally controlled by changes in rock and fluid compositions. The distribution pattern of kaolinite is obscure; its formation, however, indicates alteration by acid solutions of relatively low temperatures. Chlorite distribution is nondiagnostic and is probably related to local lithologic variations, independent of temperature. Based on the pH of the spring and

surface water samples in the vicinity of the thermal gradient holes, the kaolinite occurrences appear to represent relict phases of an earlier, more acidic period of hydrothermal alteration, whereas the montmorillonite-type clays are more indicative of the present conditions.

Continued examination and study of the geochemical nature and origin of the observed alteration in drill holes should provide viable evidence to further an understanding of the character and potential of the geothermal system in Dixie Valley.

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Depth	Lithology	Description
100		<p>Alluvial sediments: unconsolidated, light brown weathered fragments, abundant silicic-intermed. volcanics, some propylitized, diorite, chloritized biotite and hornblende, limonite-hematite, minor calcite, epidote and clay.</p> <p>Alluvial sediments: cont'd as above, with calcite cemented lithic aggregates, abundant highly altered chloritized diorite and clay.</p>
200		<p>Andesite: predominantly gray to reddish brown, some probably oxidized, silicic, aphanitic crystalline, no fresh mafics (chloritized) and euhedral plagioclase.</p>
300		<p>Tuff: medium light gray silicic welded lithic and vitric tuff, altered to clay, partially devitrified with doubly terminated quartz crystals, minor biotite weathering to chlorite.</p>
400		<p>Andesite: brownish-gray silicic, some oxidized, intercalated with silicic tuff.</p> <p>Tuff: light brown-gray fresh, unaltered silicic welded tuff with minor biotite and Fe-oxidation.</p>
500		<p>Tuff: cont'd as above, with varying lithic fragments, diorite/gabbro, andesite.</p> <p>Alluvial sediments: predominantly andesite, propylitized, minor diorite/gabbro, tuff, quartzite.</p> <p>Tuff: silicic light gray, altering to clay.</p>

Figure 1. Lithologic log for thermal gradient hole S-8.

Depth	Lithology	Description
100		Alluvial sediments: unconsolidated predominantly light brown to gray weathered sediments, abundant volcanics, andesite (some propylitized), abundant felsic fragments, varying amounts silicic tuff, aplite, diorite/gabbro, quartzite, abundant limonite-hematite staining, chloritized biotite, hornblende, trace epidote, magnetite after pyrite, calcite, CaCO <sub>3</sub> cement.
200		Alluvial sediments: cont'd as above, with abundant crystalline pyrite on fine-grained volcanics. Alluvial sediments: cont'd as above, with predom. silicic tuff, decreased pyrite, abundant felsic fragments, CaCO <sub>3</sub> cemented aggregates of lithic-felsic fragments, abundant altered diorite/gabbro, chloritized ferromagnesian, minor sulfur, epidote, pyrite, clay, hematite-limonite.
300		Alluvial sediments: cont'd as above, with abundant lithic and felsic fragments, altered diorite/gabbro, Fe-oxides, trace sulfur, epidote.
400		Alluvial sediments: cont'd as above, unconsolidated, predominantly white to light green, coarsely crystalline quartzite, with some secondary quartz, felsic fragments, aplite, silicic volcanics, diorite/gabbro, altered granite, hematite-limonite, minor sulfur, epidote, chlorite after biotite, hornblende, magnetite, trace crystalline calcite, CaCO <sub>3</sub> cement, clay.
500		Alluvial sediments: cont'd as above, with abundant silicic tuff, devitrified, altering to clay and calcite (?), minor pyrite; increased amounts of pink aplite and diorite/gabbro with plagioclase altering to clay and chloritization of ferromagnesian.

Figure 2. Lithologic log for thermal gradient hole SR2.

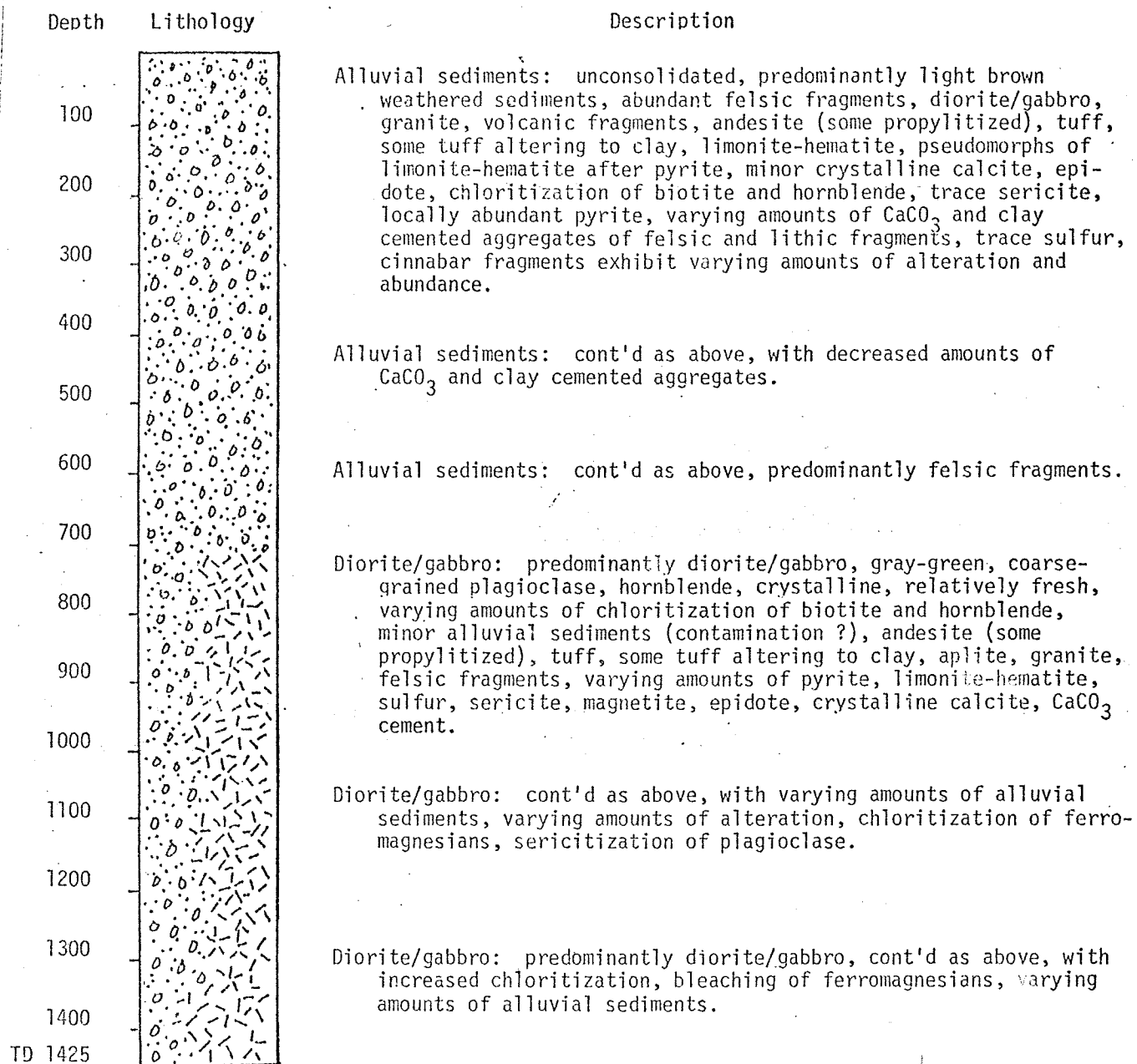


Figure 3. Lithologic log for thermal gradient hole SR2A.



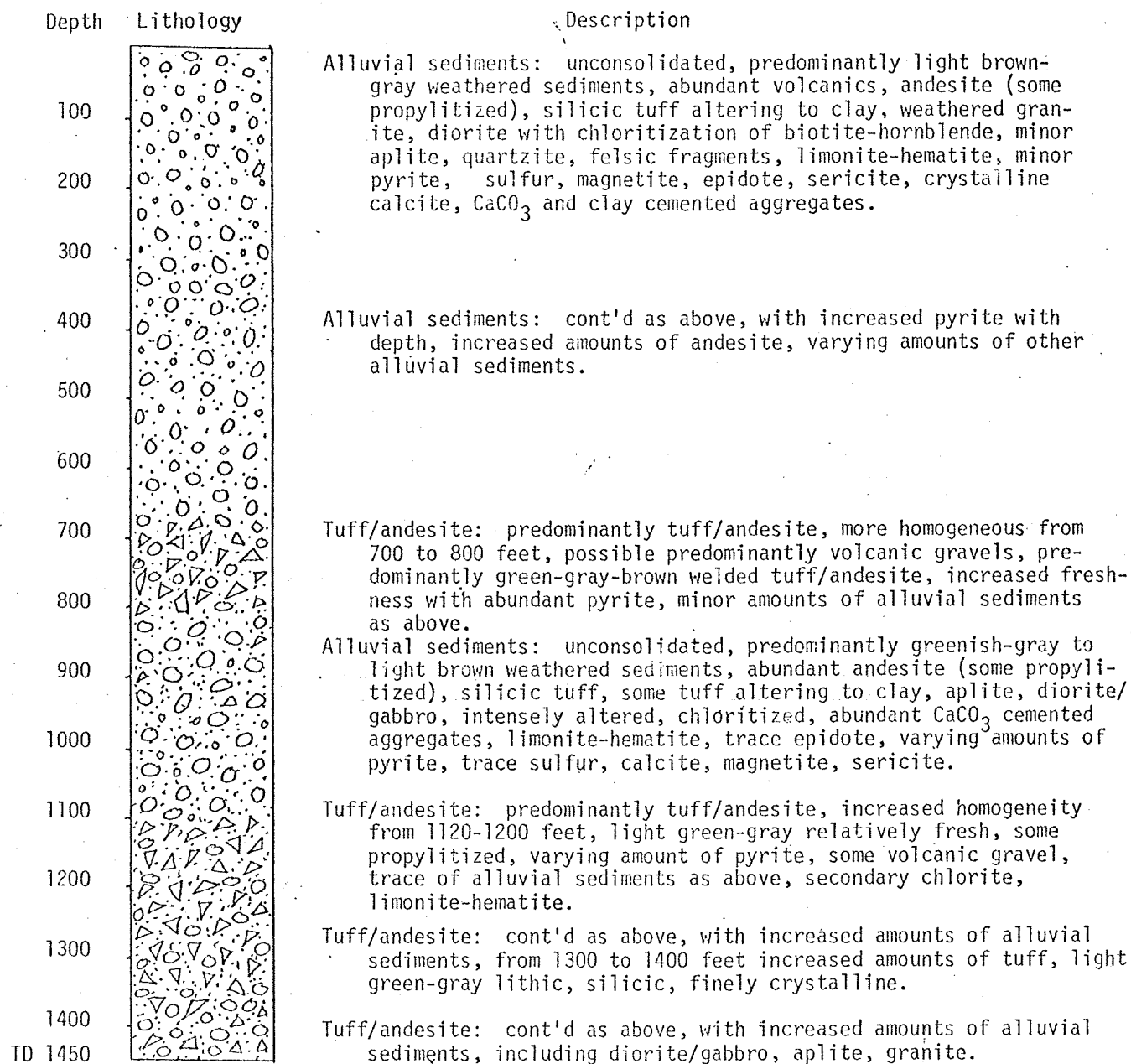


Figure 4. Lithologic log for thermal gradient hole DD-9.

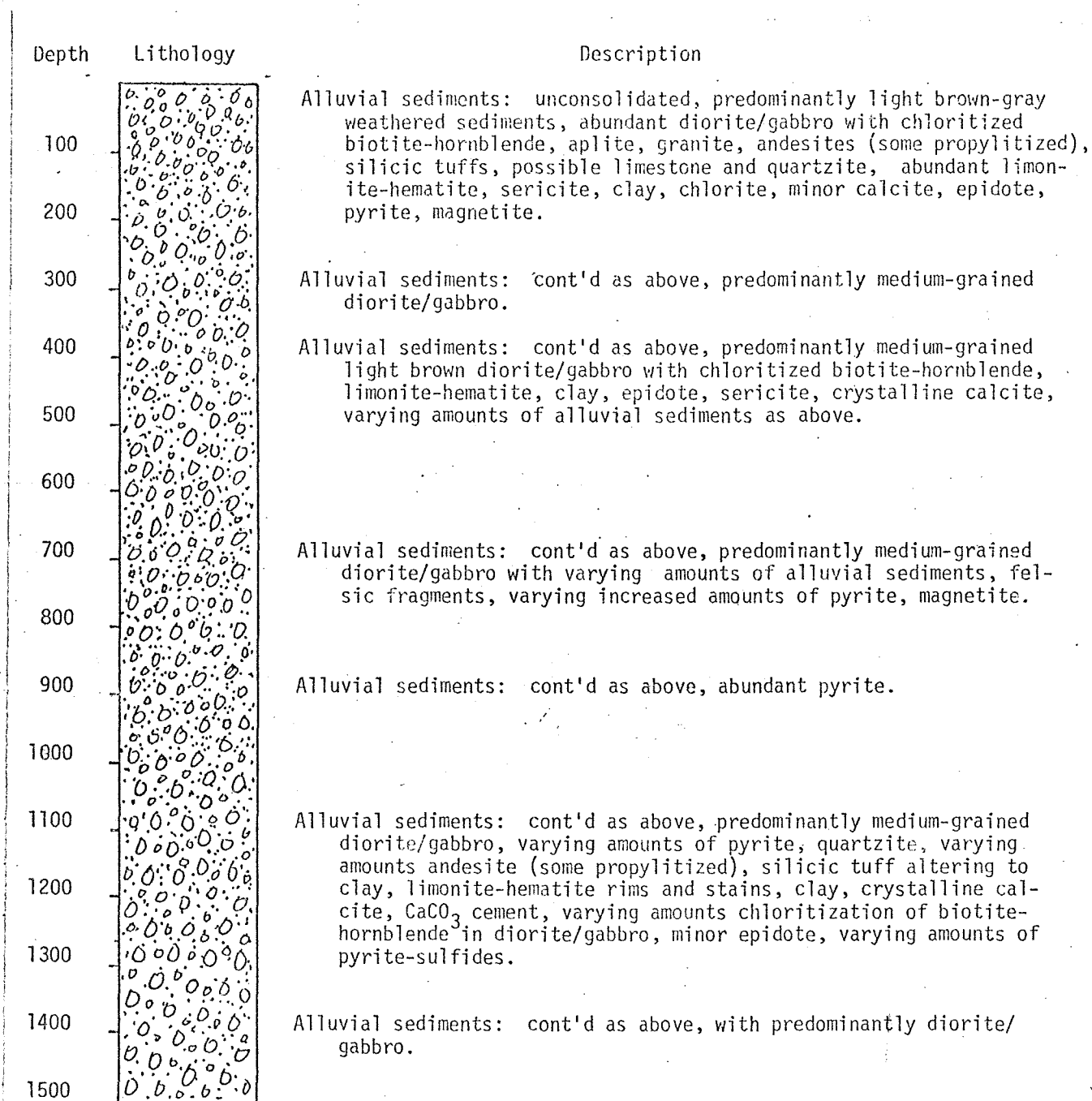


Figure 5. Lithologic log for thermal gradient hole H-2.

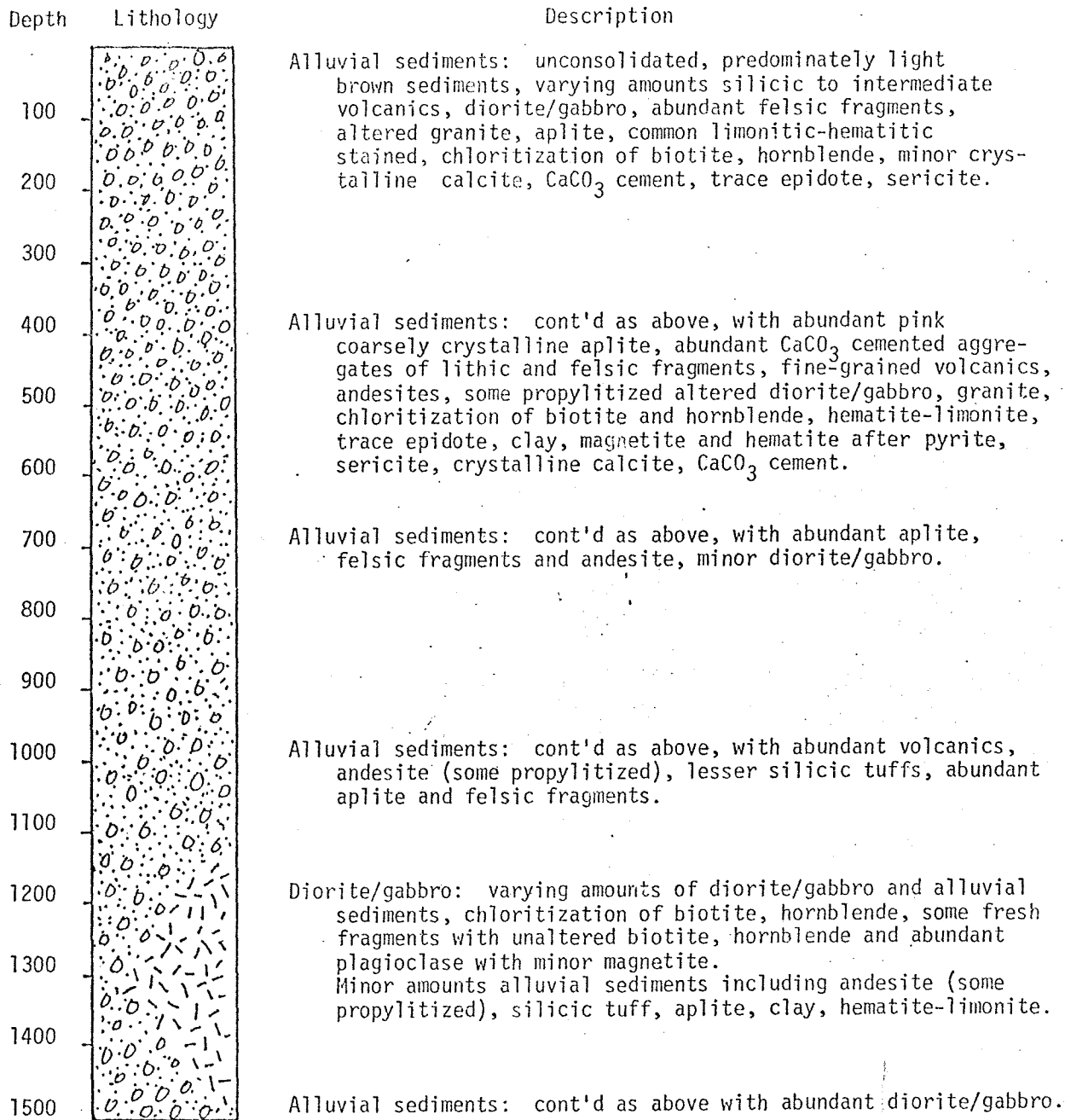
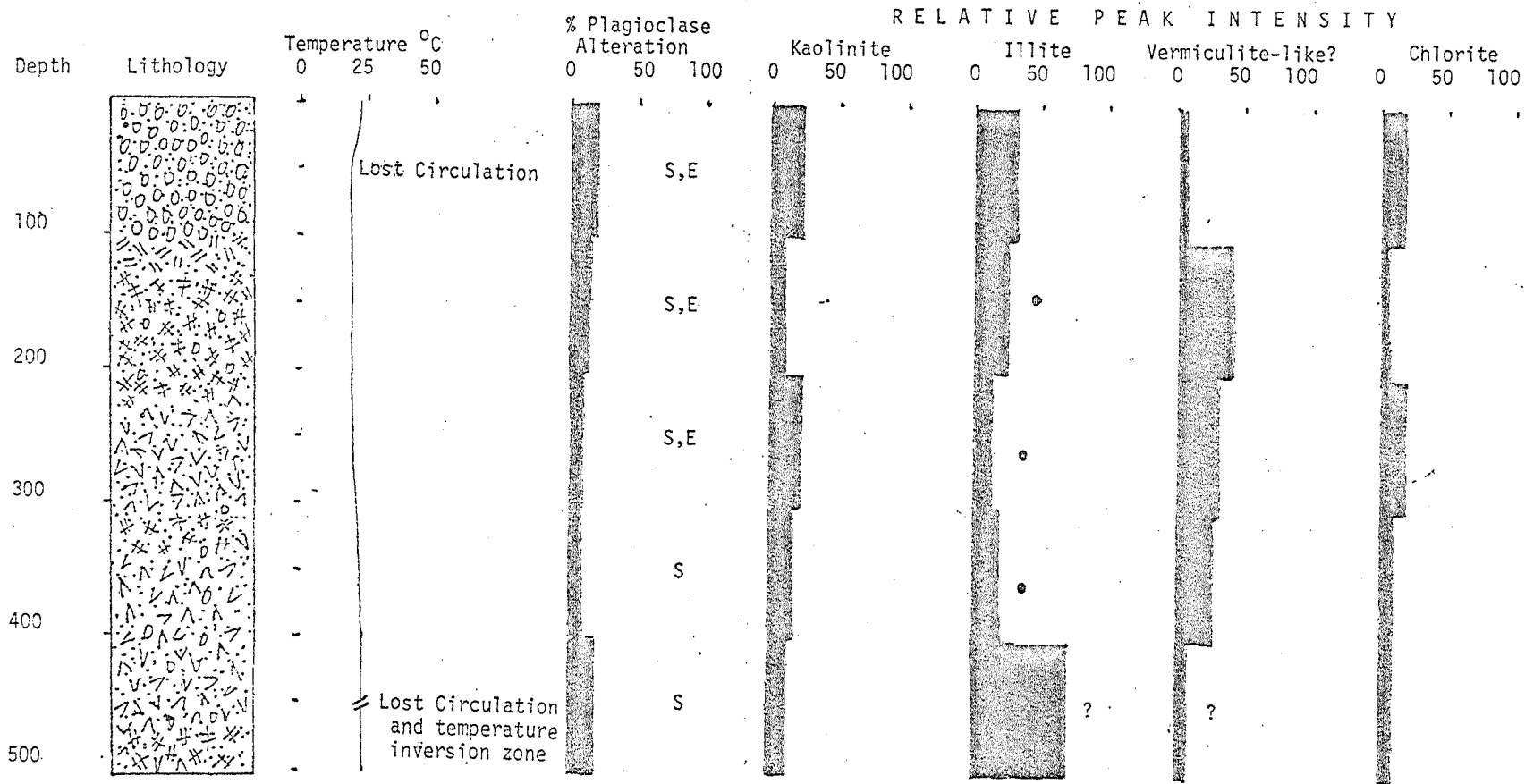
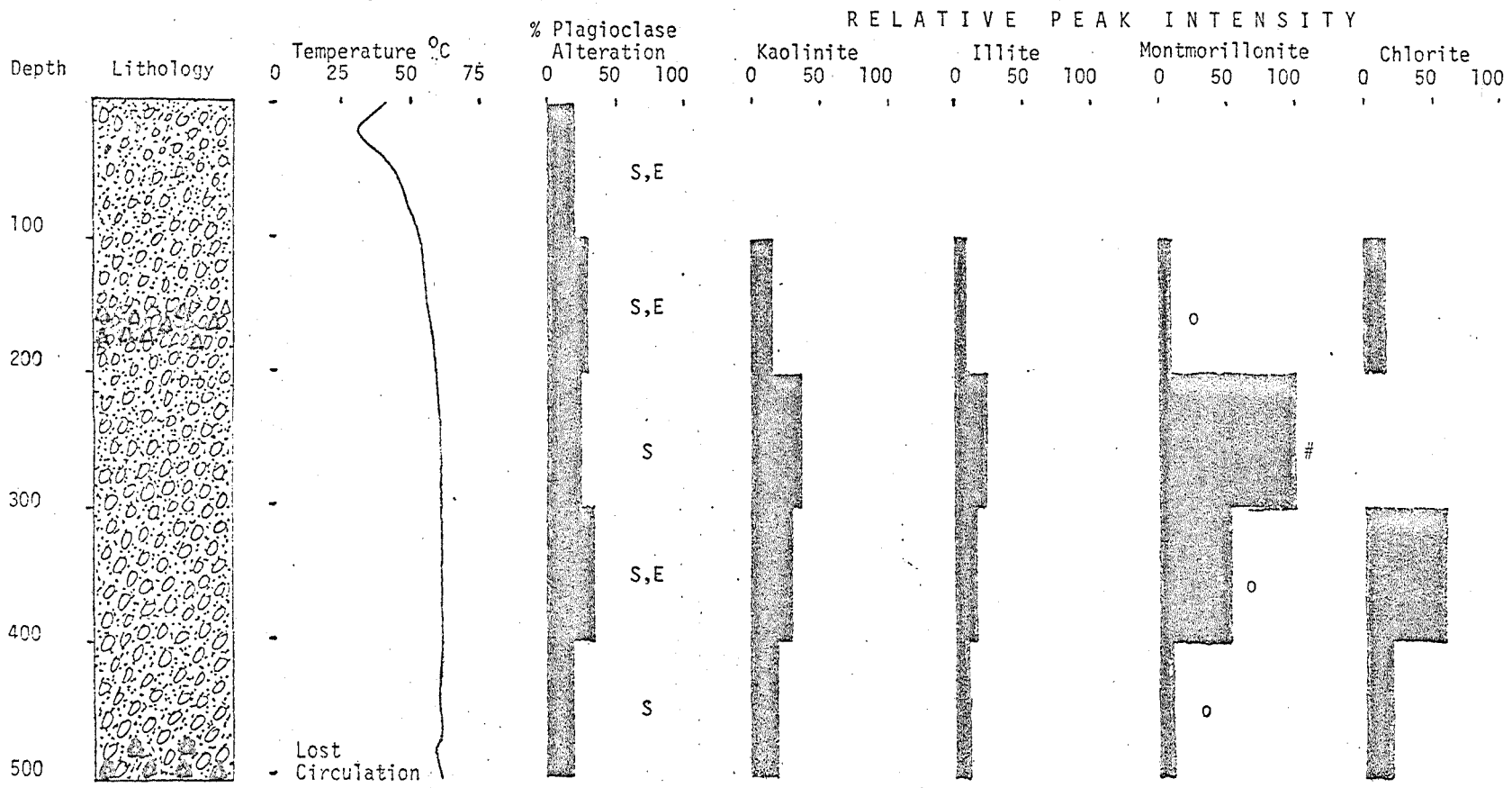


Figure 6. Lithologic log for thermal gradient hole H-1.



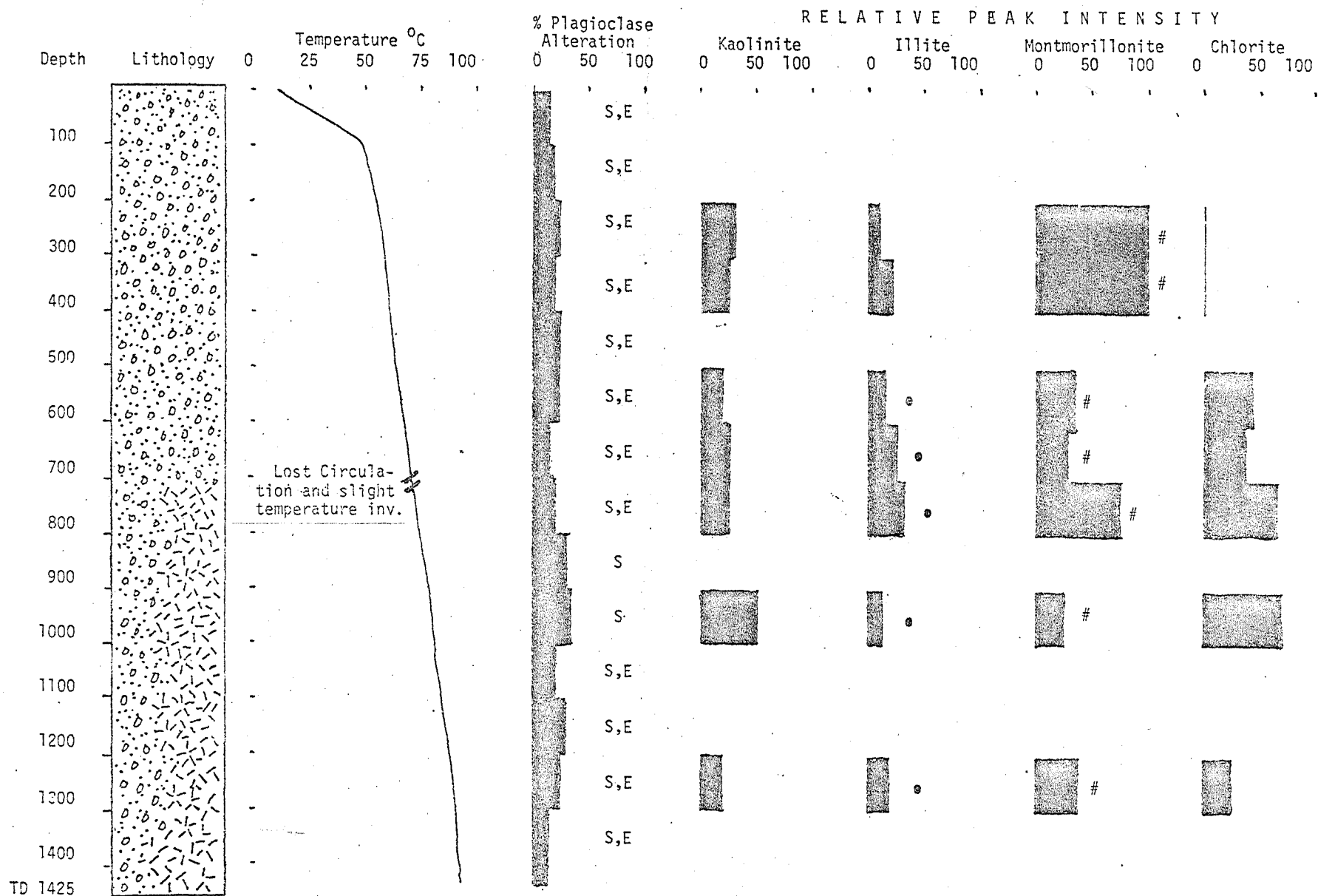
Note: # - Interlayered Na-Ca-Mg-Montmorillonite; E - Epidote; S - Sercite; \* - Ca and/or Mg Montmorillonite; o - Na-Montmorillonite; • - Mixed-layer Illite/Vermiculite(?)

Figure 7. Distribution of clay mineral species for thermal gradient hole S-8.



Note: # - Interlayered Na-Ca-Mg-Montmorillonite; E - Epidote; S - Sericite; \* - Ca and/or Mg Montmorillonite; o - Na-Montmorillonite; • - Mixed-layer Illite/Montmorillonite.

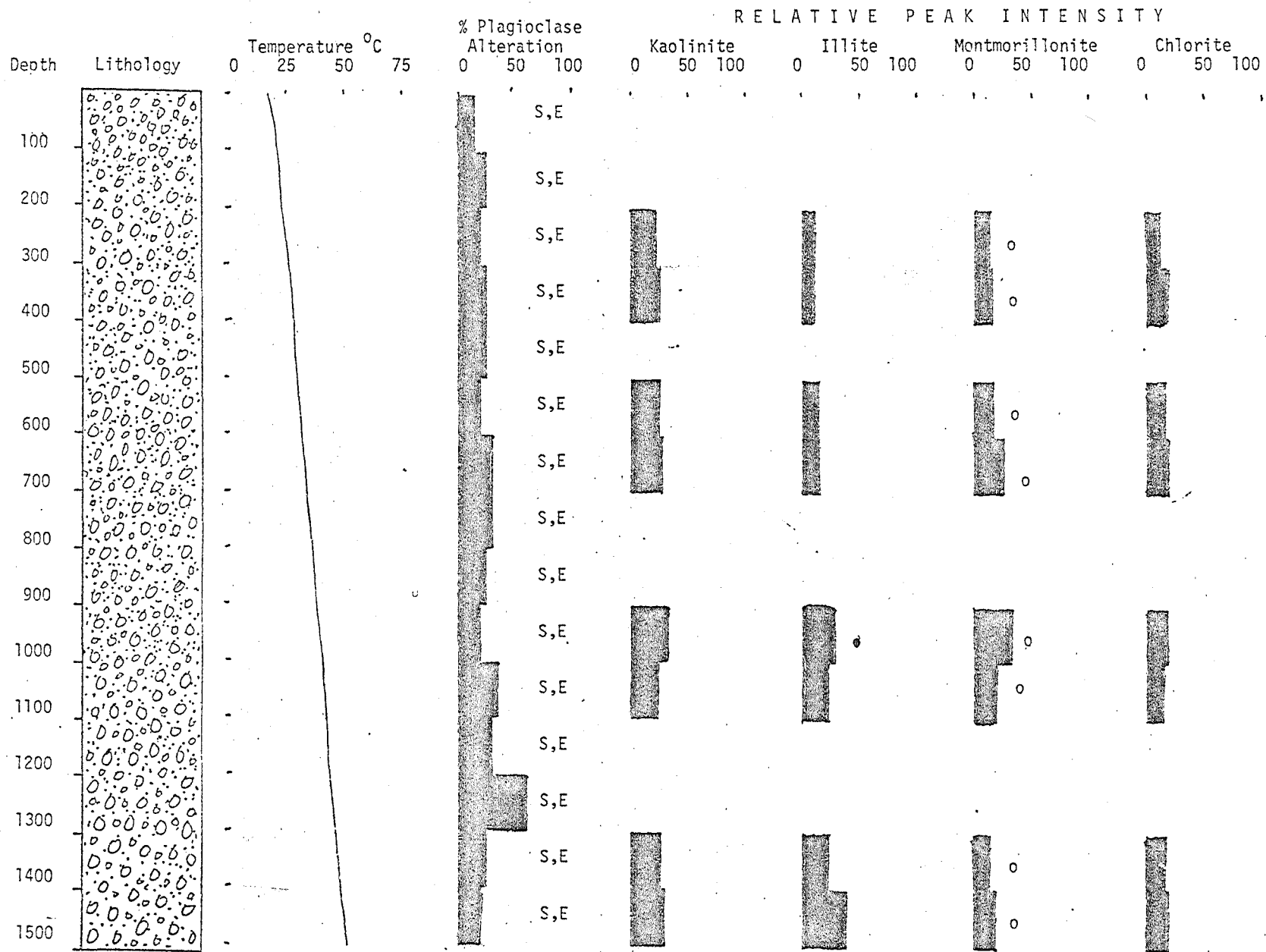
Figure 8. Distribution of clay mineral species for thermal gradient hole SR2.



Note: # - Interlayered Na-Ca-Mg-Montmorillonite; E - Epidote; S - Sericite; \* - Ca and/or Mg Montmorillonite; o - Na-Montmorillonite; • - Mixed-layer Illite/Montmorillonite.

Figure 9. Distribution of clay mineral species for thermal gradient hole SR2A.

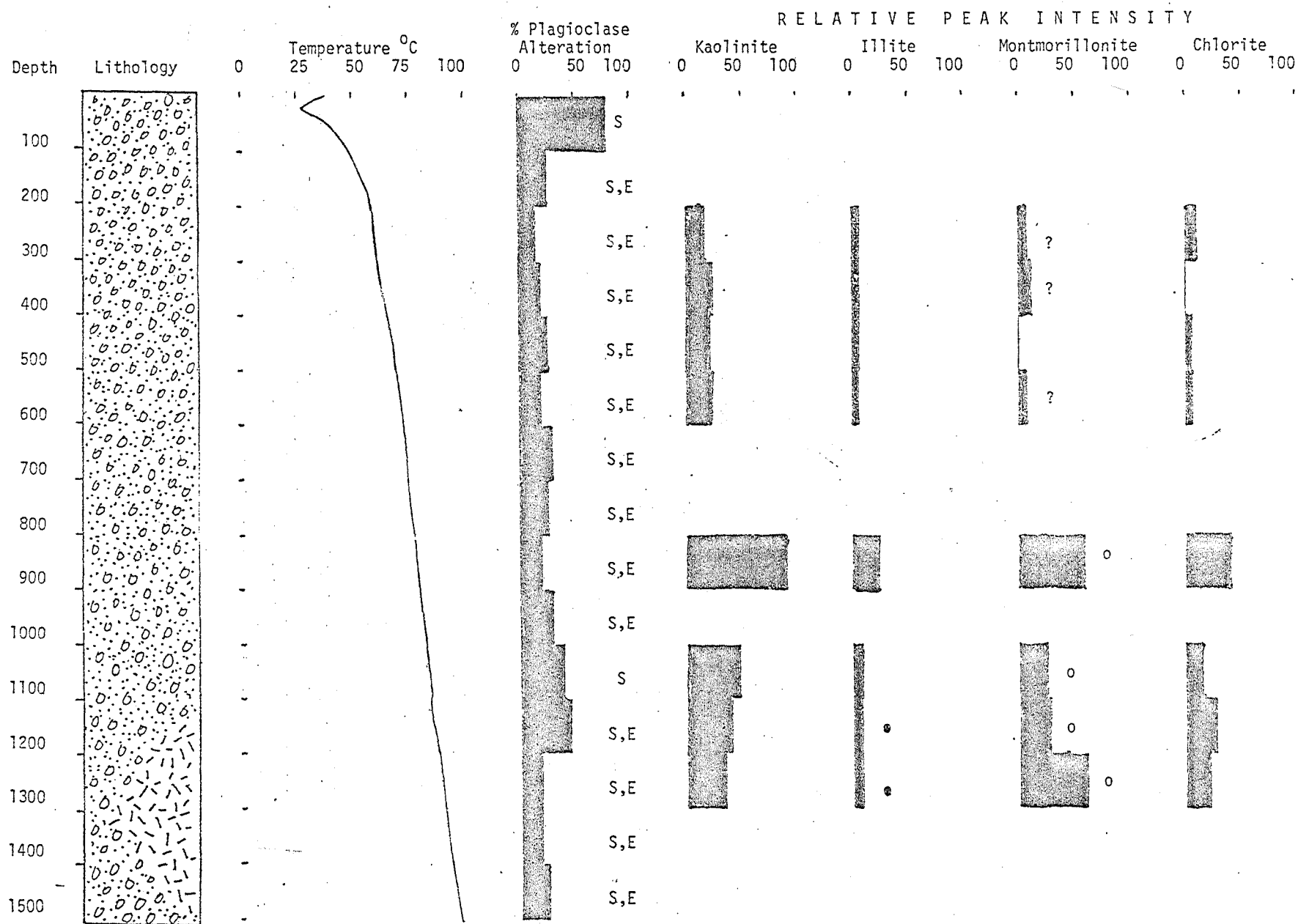




Note: # - Interlayered Na-Ca-Mg-Montmorillonite; E - Epidote; S - Sericite; \* - Ca and/or Mg Montmorillonite; ○ - Na-Montmorillonite; ● - Mixed-layer Illite/Montmorillonite.

Figure 11. Distribution of clay mineral species for thermal gradient hole H-2.





Note: # - Interlayered Na-Ca-Mg Montmorillonite; E - Epidote; S - Sericite; \* - Ca and/or Mg Montmorillonite; o - Na-Montmorillonite; • - Mixed-layer Illite/Montmorillonite.

Figure 12. Distribution of clay mineral species for thermal gradient hole H-1.

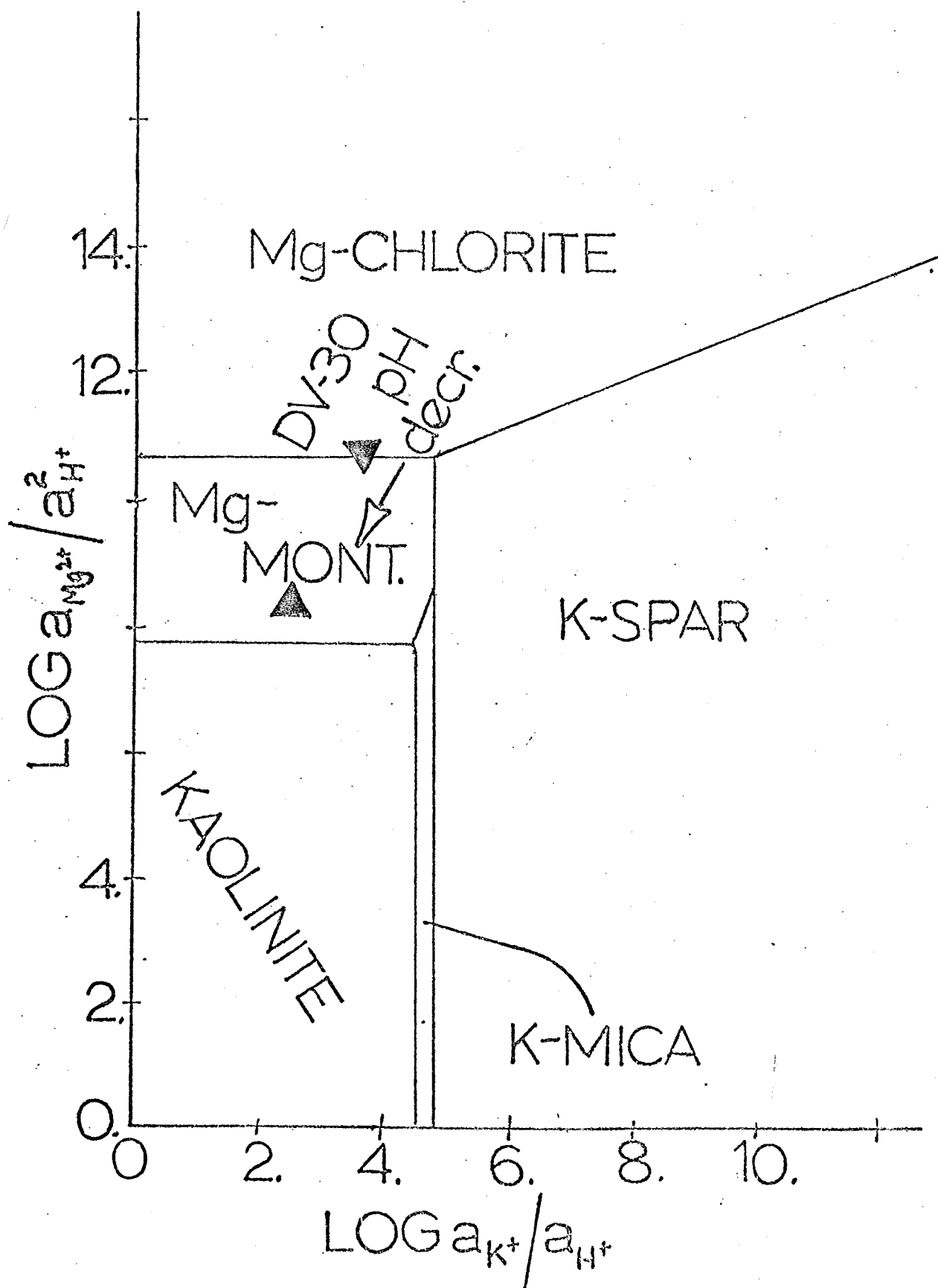


Figure 41. Plot of water sample DV-30 from SR2A at 100°C.

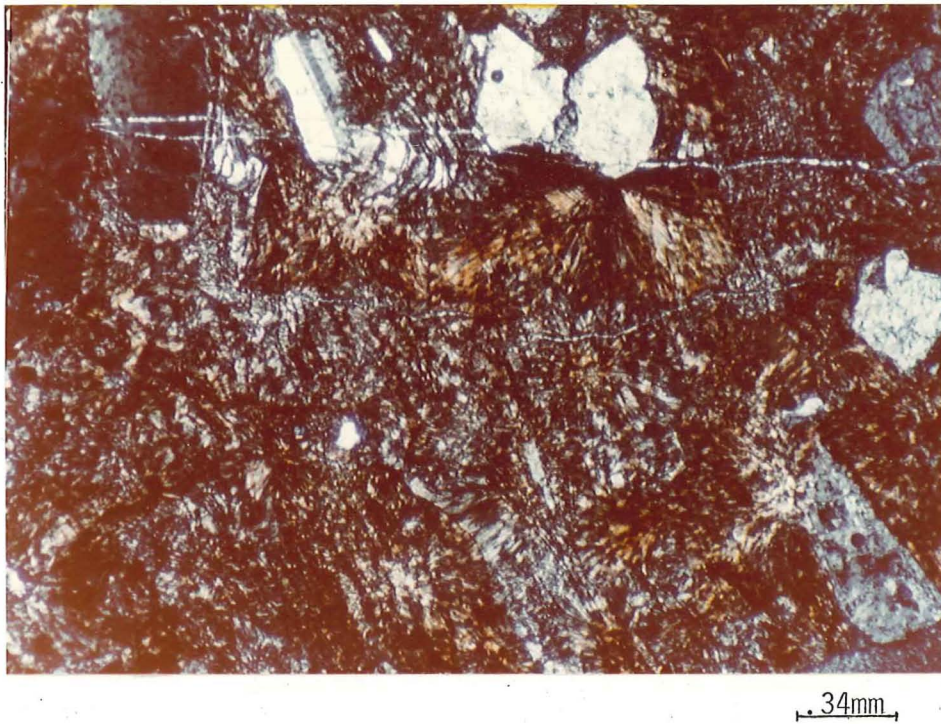


Figure 13. Photomicrograph of tuff unit under crossed polars. (thermal gradient hole S-8)

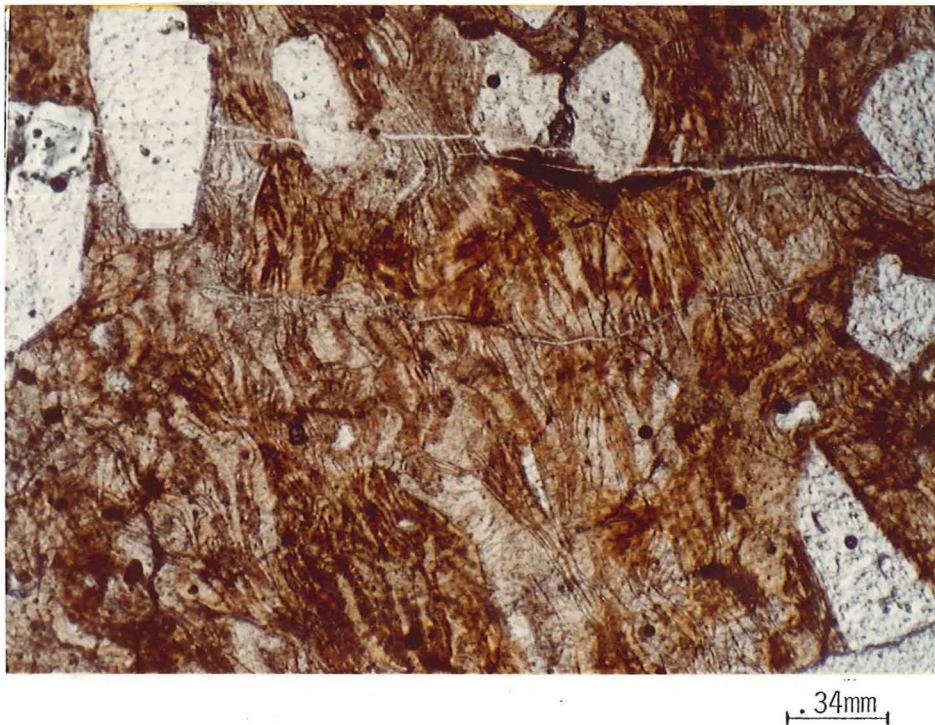


Figure 14. Same as Figure 13 without crossed polars. Note flow structures and development of clay.



Figure 15. Sericitization of plagioclase under crossed polars. Note relatively unaltered rim. (DD-9, 1200 - 1300 feet)

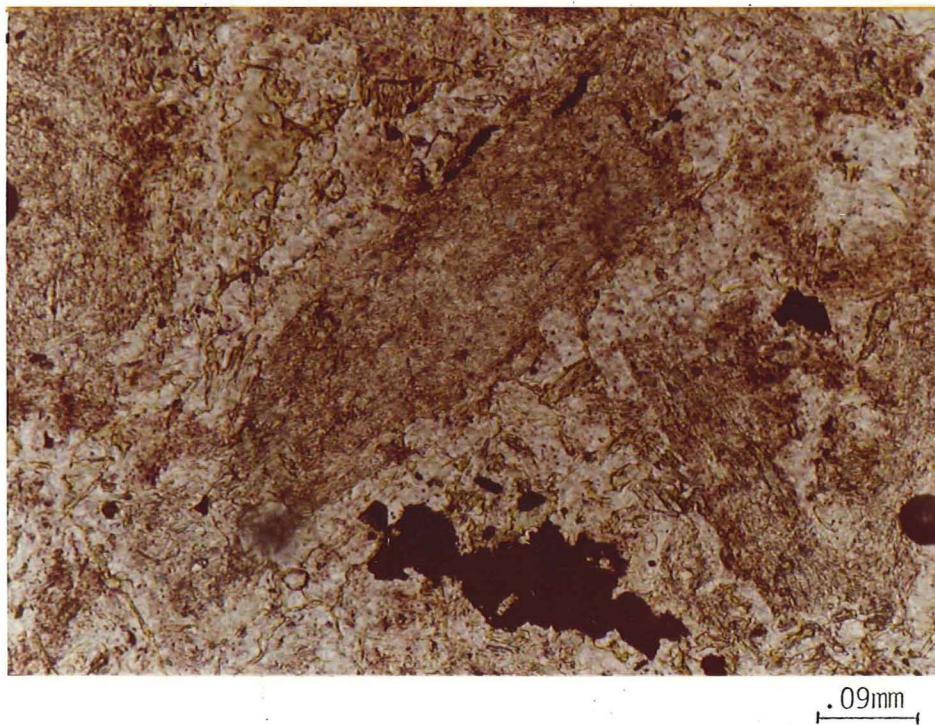


Figure 16. Same as Figure 15 without crossed polars. Note 'dirty' appearance of crystal.

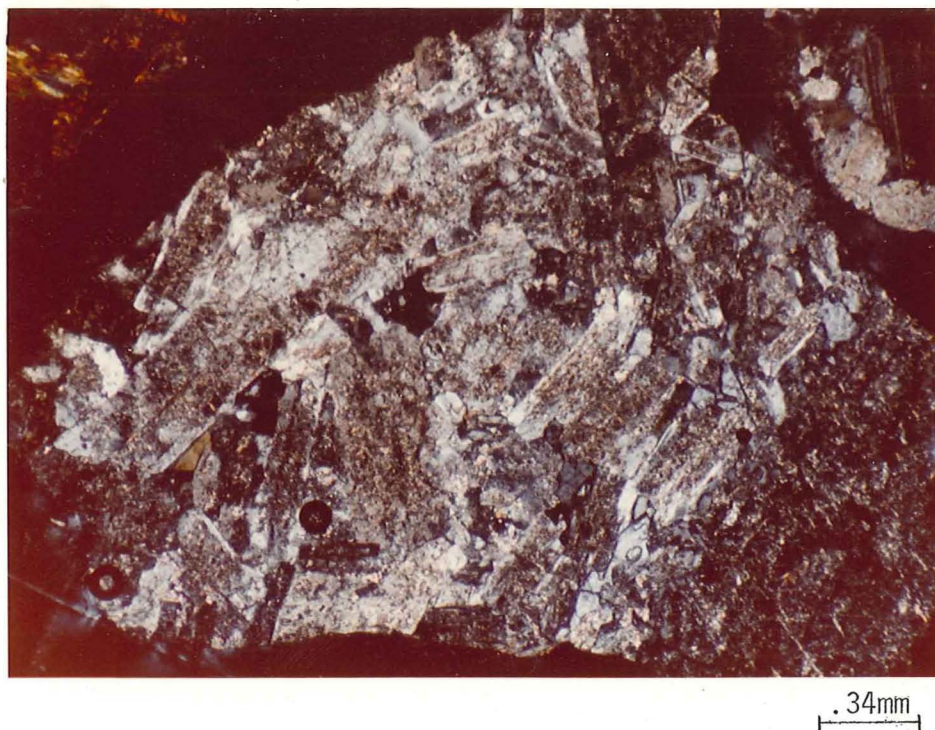


Figure 17. Sericitized plagioclase lathes in volcanic rock under crossed polars. Note the sericitization proceeds from the cores of the crystals outward.

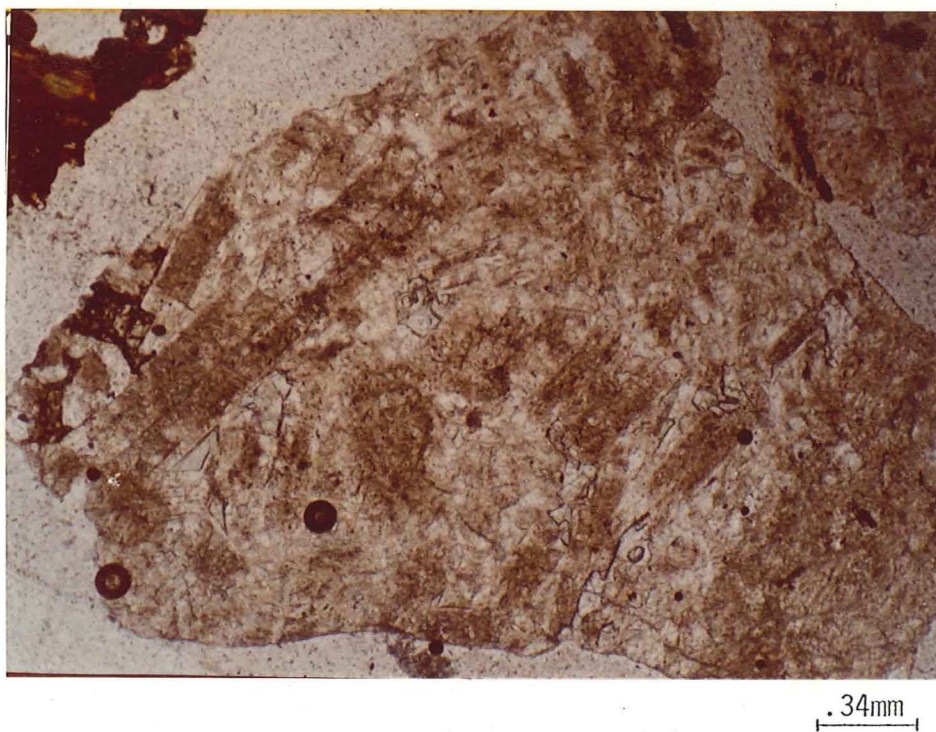
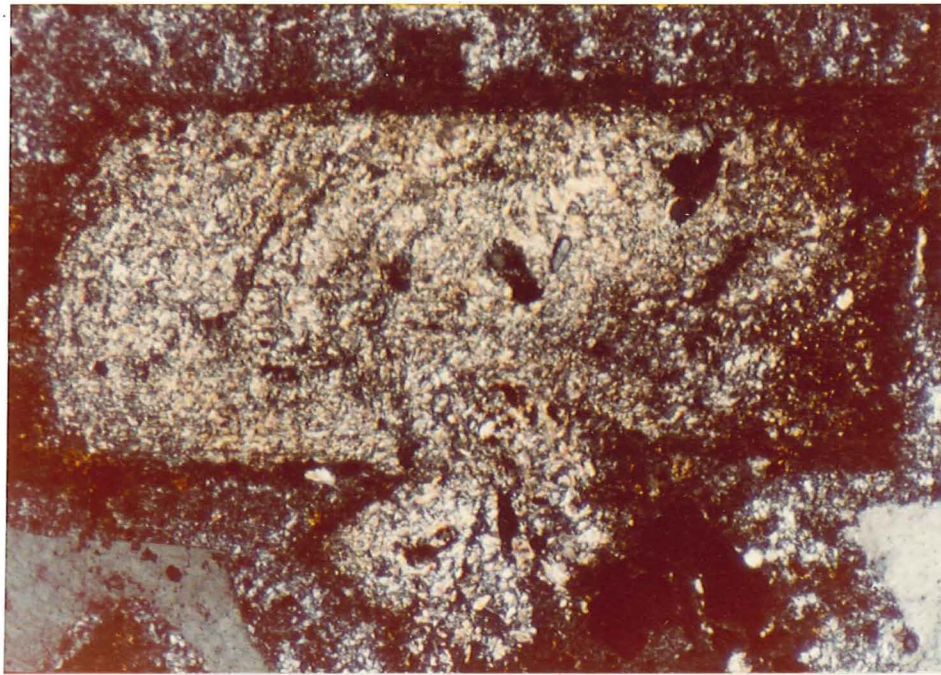
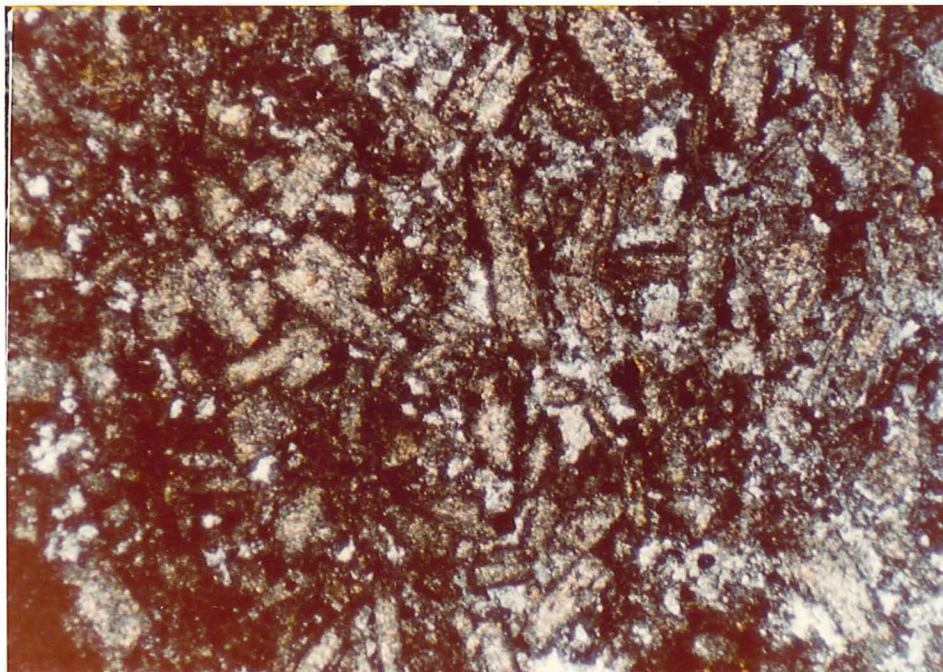


Figure 18. Same as Figure 17 without crossed polars. Note the lack of clarity of the crystals.



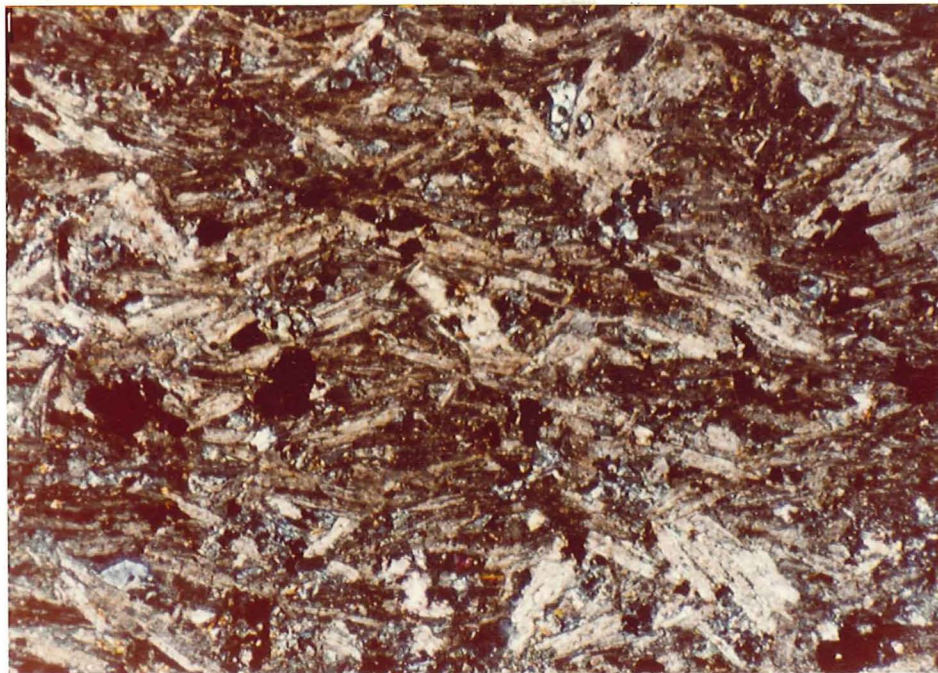
.27mm

Figure 19. Completely sericitized plagioclase crystal under crossed polars. Note the relict form of the plagioclase. (H-1, 0 - 100 feet)



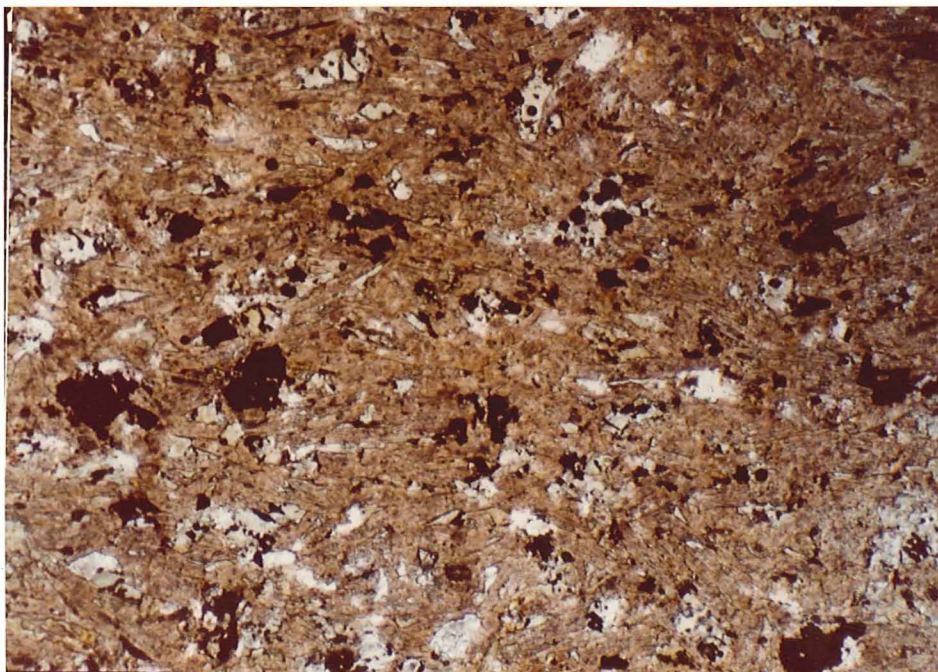
.27mm

Figure 20. Ghosts of plagioclase crystals in volcanic rock under crossed polars. (DD-9, 1100 - 1200 feet)



.34mm

Figure 21. Alteration of oriented plagioclase lathes in an andesite under crossed polars. (H-2, 500 - 600 feet)



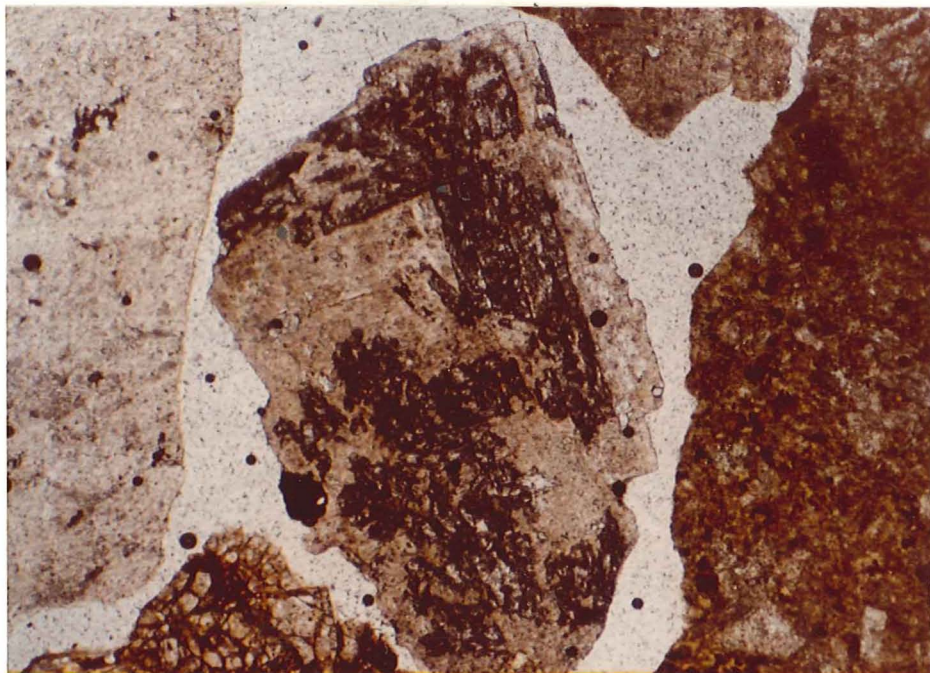
.34mm

Figure 22. Same as Figure 21 without crossed polars.



.34mm

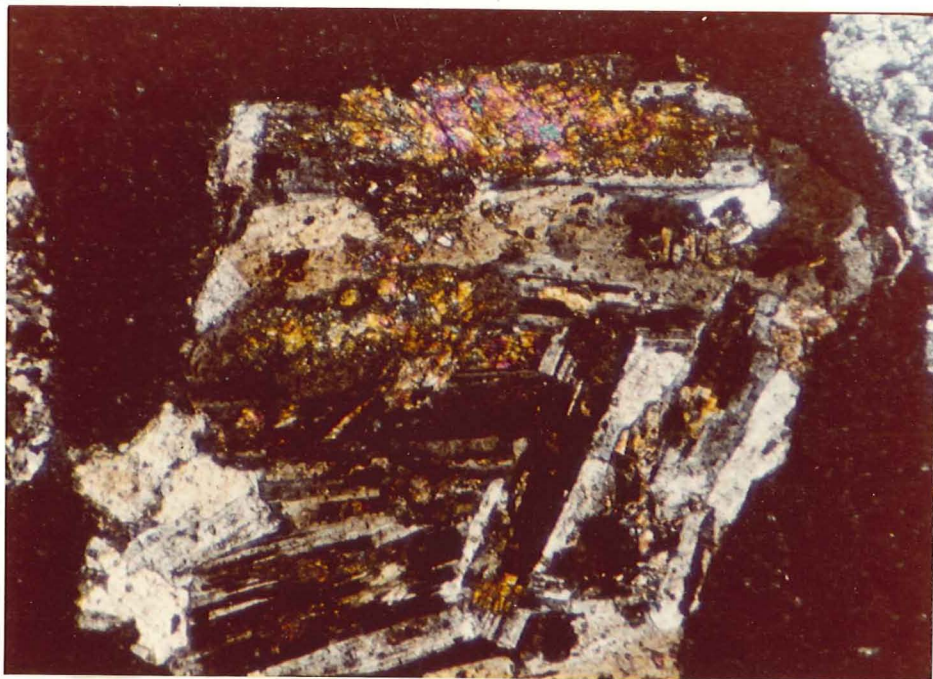
Figure 23. Epidote after plagioclase under crossed polars. (SR2A, 600 - 700 feet)



.34mm

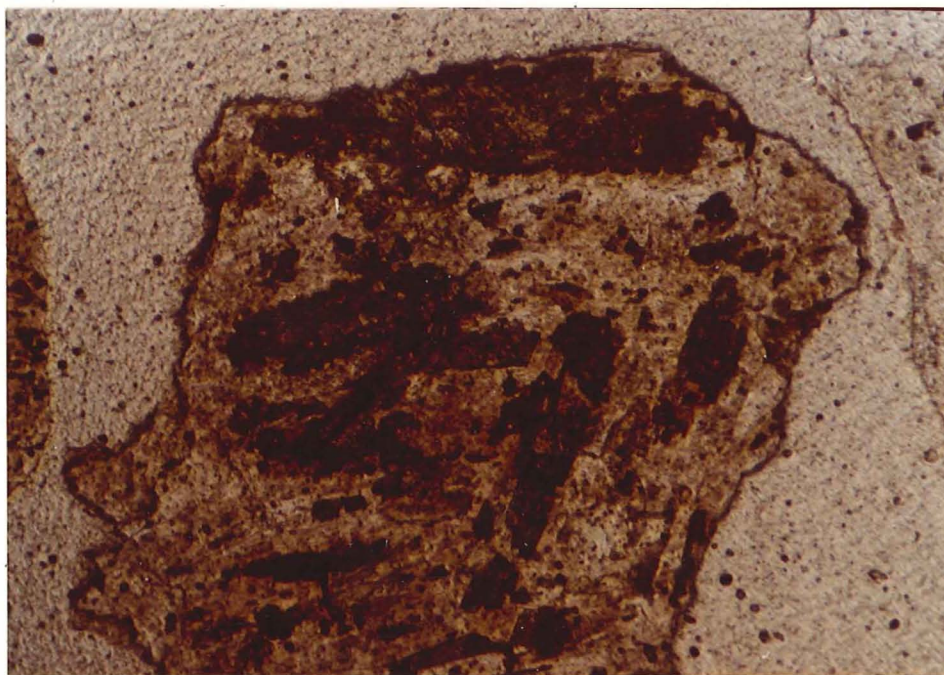
Figure 24. Same as Figure 23 without crossed polars.





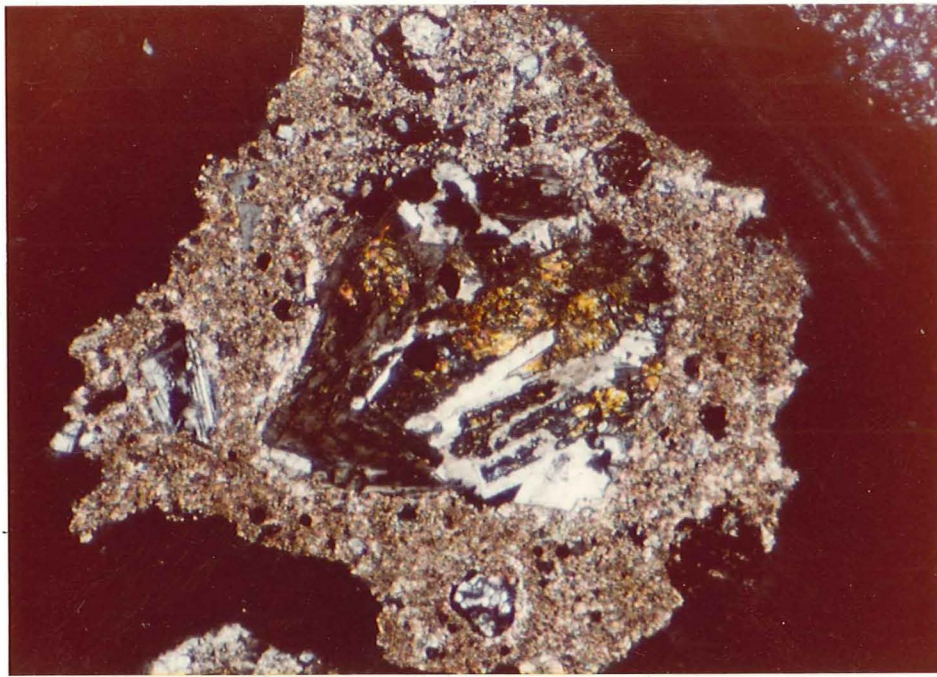
.27mm

Figure 25. Epidote after plagioclase under crossed polars. (H-1, 900 - 1000 feet)



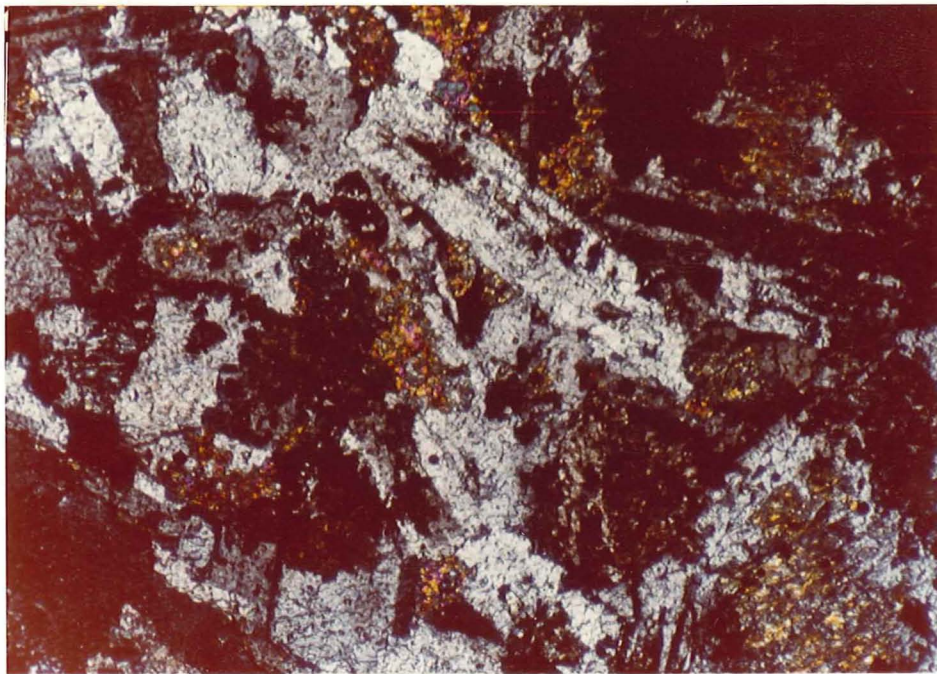
.27mm

Figure 26. Same as Figure 25 without crossed polars.



.27mm

Figure 27. Epidote replacing plagioclase under crossed polars. Note the alteration surrounding the fragment. (H-1, 100 - 200 feet)



.34mm

Figure 28. Minor epidote after plagioclase under crossed polars. (DD-9, 600 - 700 feet)

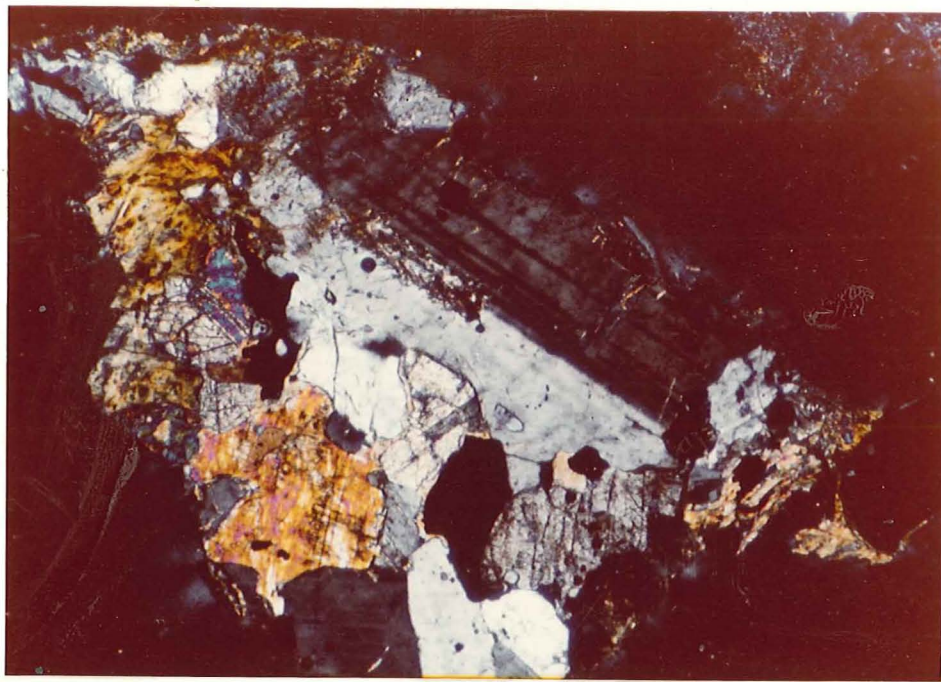


Figure 29. Relatively fresh diorite/gabbro under crossed polars. (H-1, 1200 - 1300 feet)

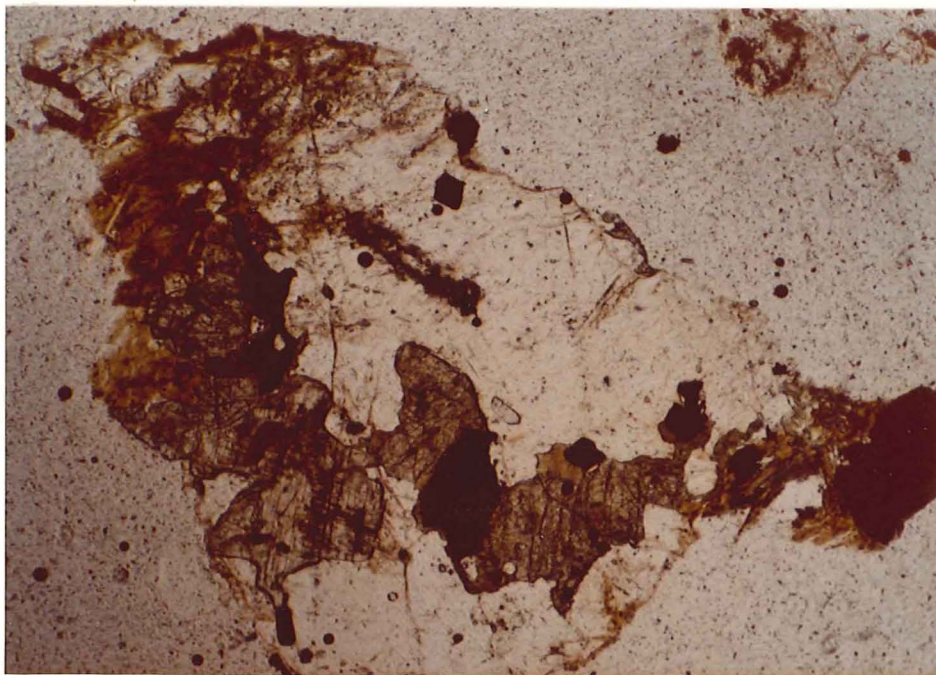
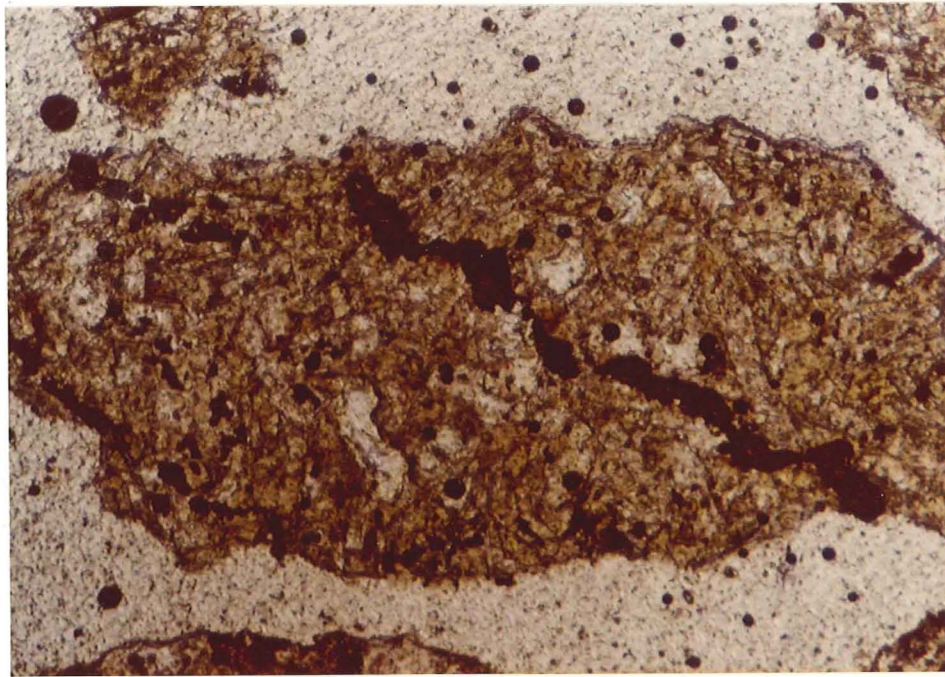
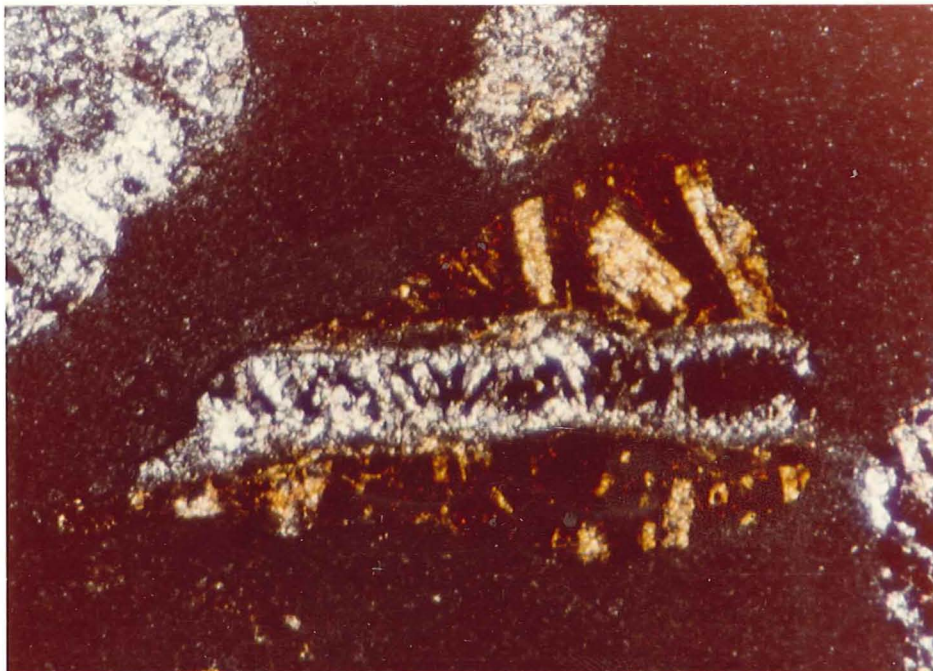


Figure 30. Same as Figure 29 without crossed polars. Note clarity of plagioclase crystal.



.22mm

Figure 39. Epidote vein in altered volcanic rock without crossed polars. (H-2, 400 - 500 feet)



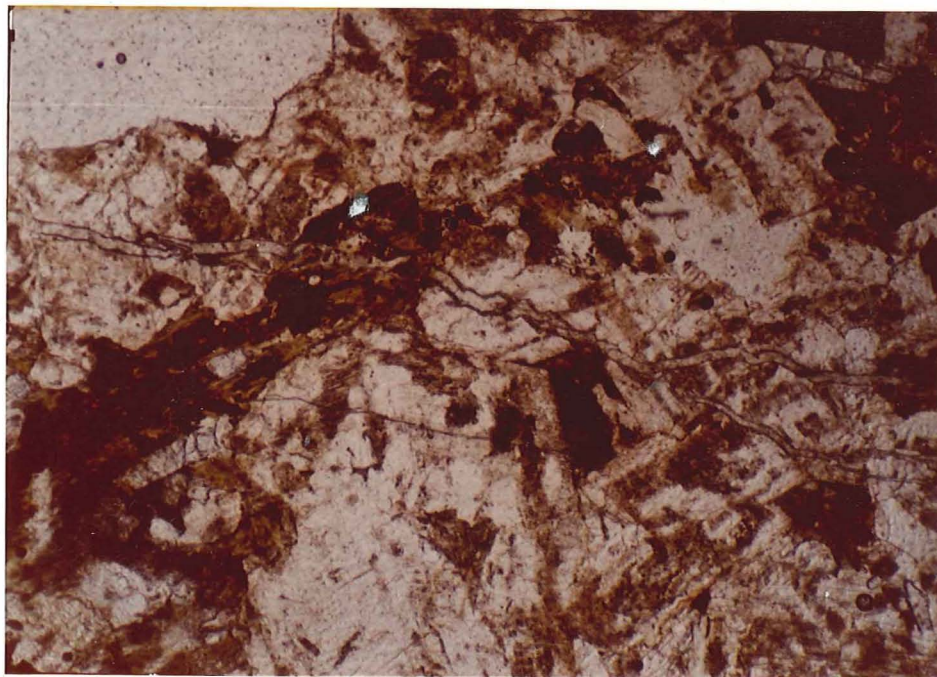
.22mm

Figure 40. Quartz vein in iron-stained volcanic rock under crossed polars. Note the comb structure of the quartz. (DD-9, 600 - 700 feet)



.34mm  
└──────────┘

Figure 37. Calcite vein in diorite/gabbro under crossed polars. (H-1, 100 - 200 feet)



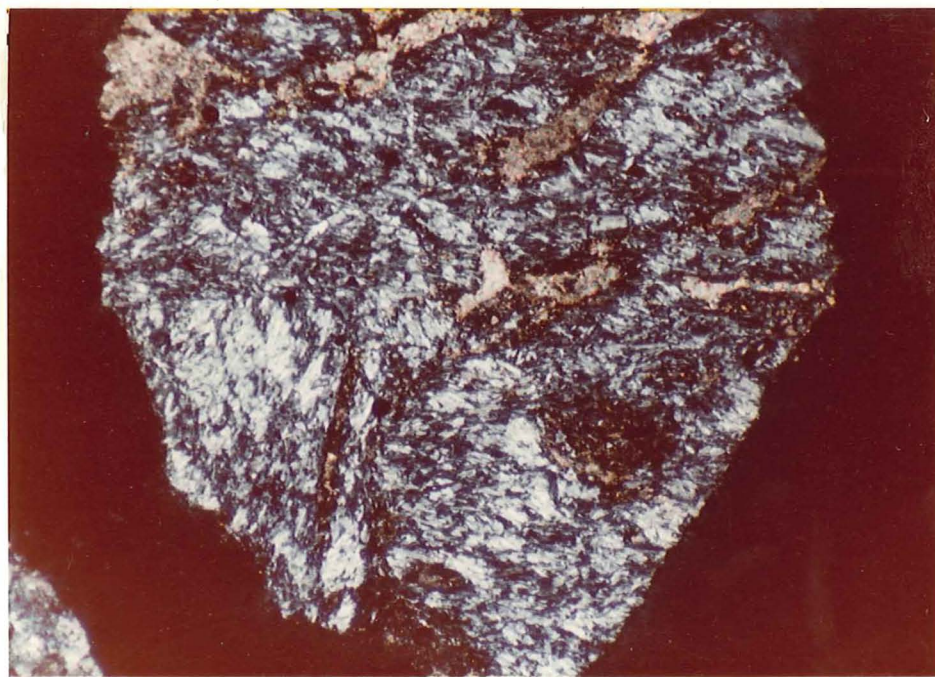
.34mm  
└──────────┘

Figure 38. Same as Figure 37 without crossed polars.



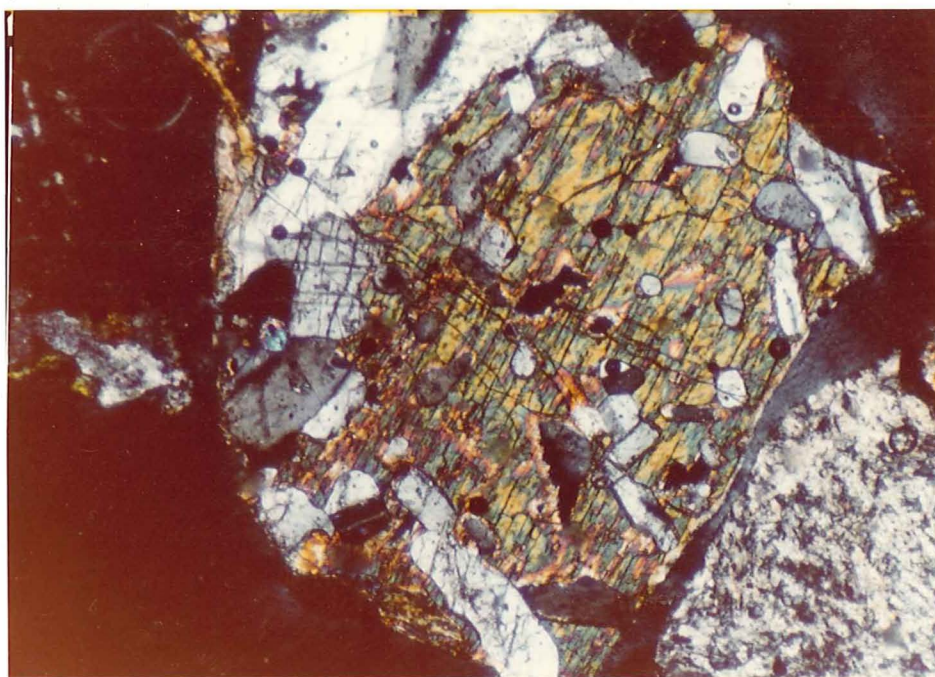
.34mm  
|-----|

Figure 35. Calcite vein in altered volcanic rock under crossed polars. (SR2A, 500 - 600 feet)



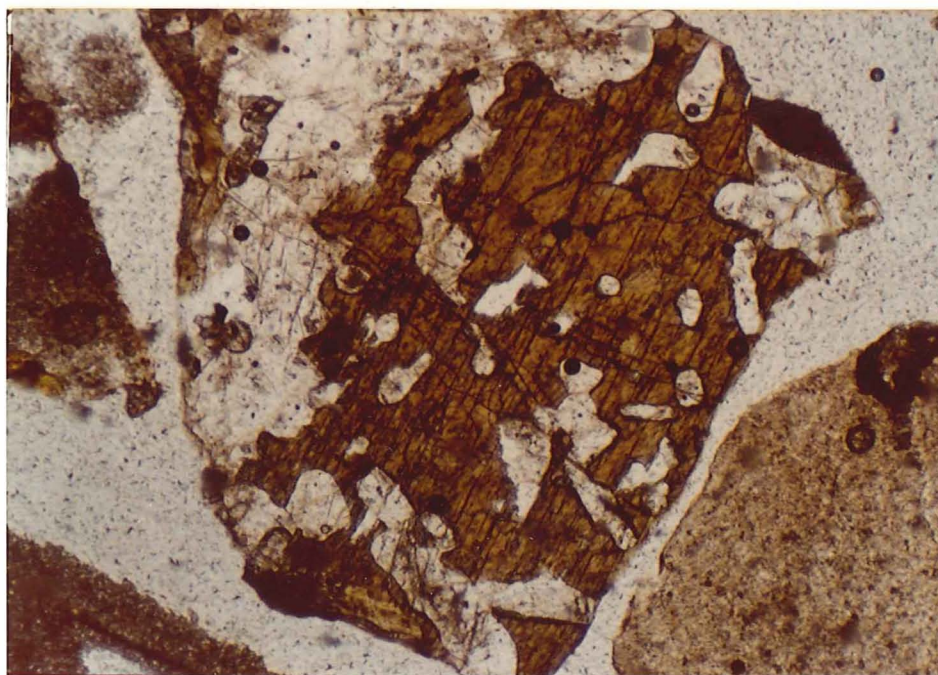
.34mm  
|-----|

Figure 36. Calcite vein in altered andesite under crossed polars. (H-2, 1100 - 1200 feet)



.27mm  
|-----|

Figure 33. Poikilitic hornblende in diorite/gabbro under crossed polars. (H-1, 1300 - 1400 feet)



.27mm  
|-----|

Figure 34. Same as Figure 33 without crossed polars. Note the relatively unaltered appearance of the rock.

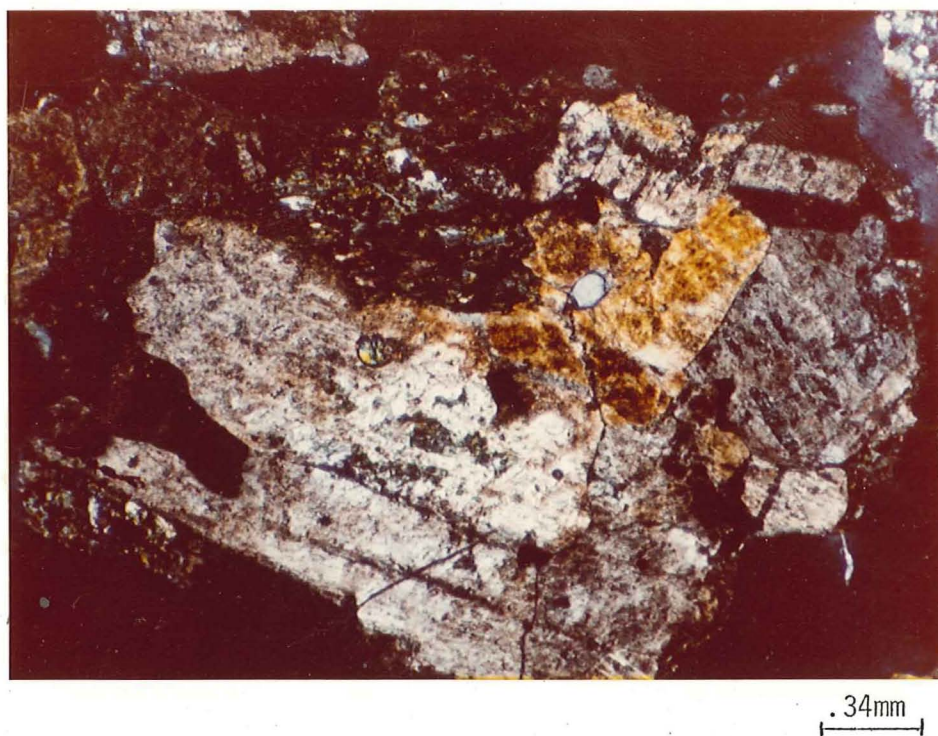


Figure 31. Altered diorite/gabbro under crossed polars.  
(H-1, 1100 - 1200 feet)

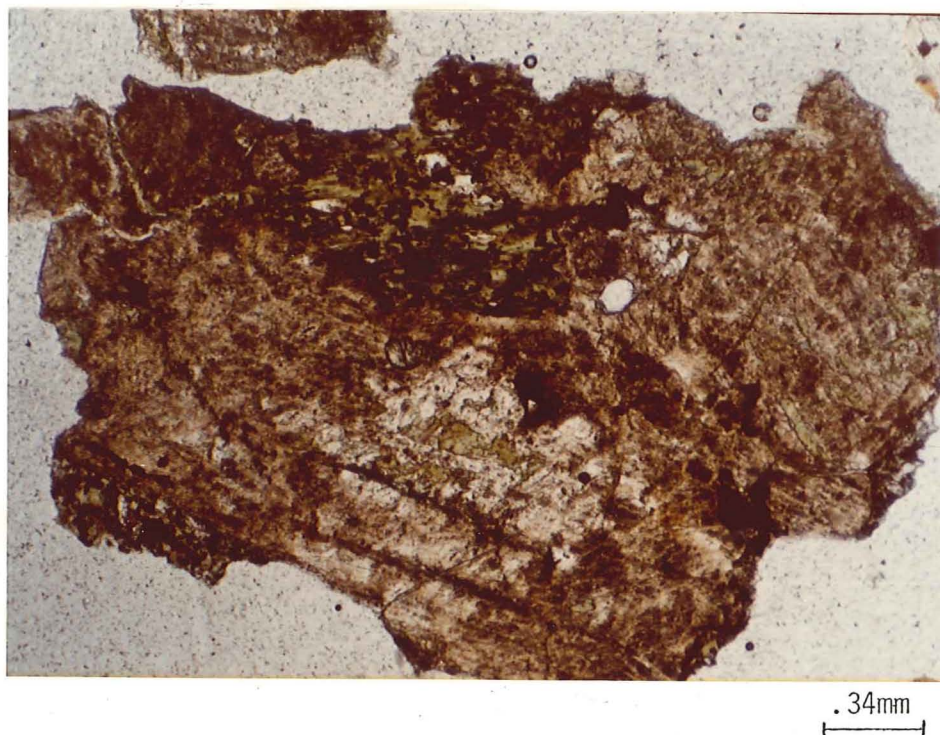
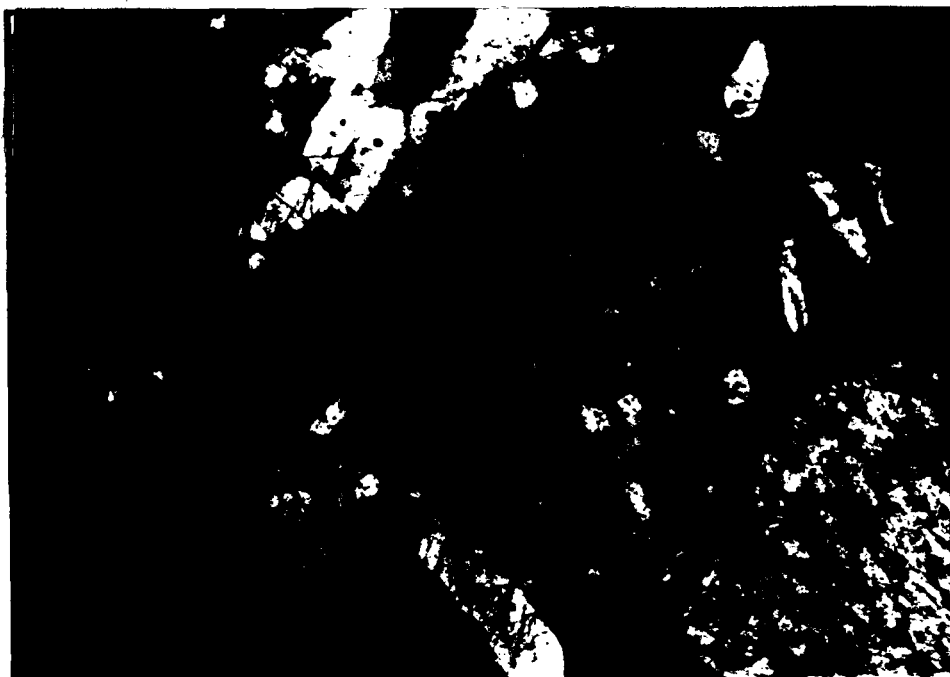


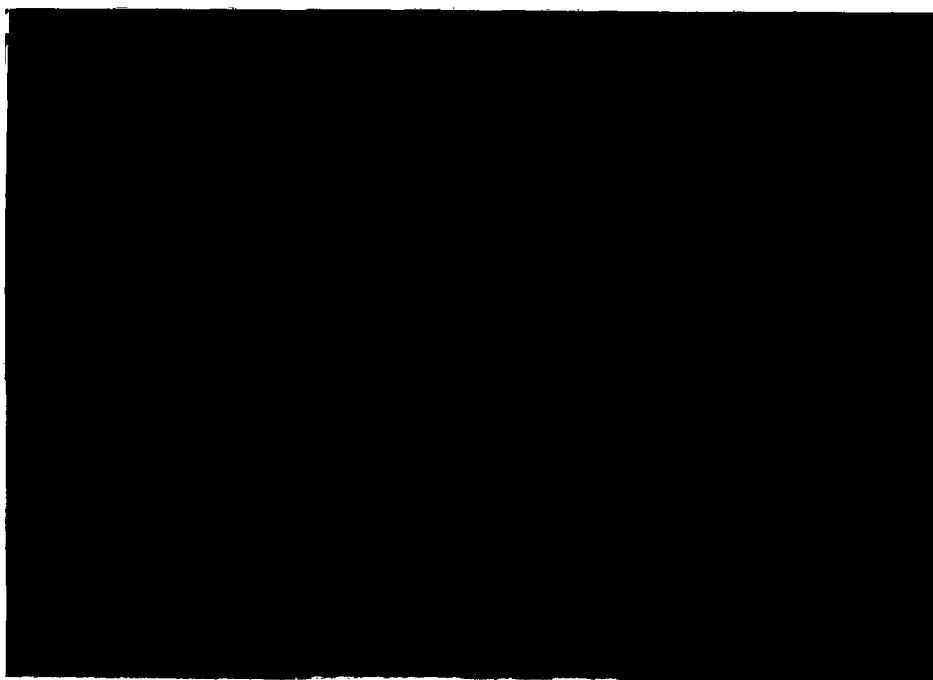
Figure 32. Same as Figure 31 without crossed polars.  
Note altered appearance of plagioclase.





.27mm  
└──────────┘

Figure 33. Poikilitic hornblende in diorite/gabbro under crossed polars. (H-1, 1300 - 1400 feet)



.27mm  
└──────────┘

Figure 34. Same as Figure 33 without crossed polars. Note the relatively unaltered appearance of the rock.

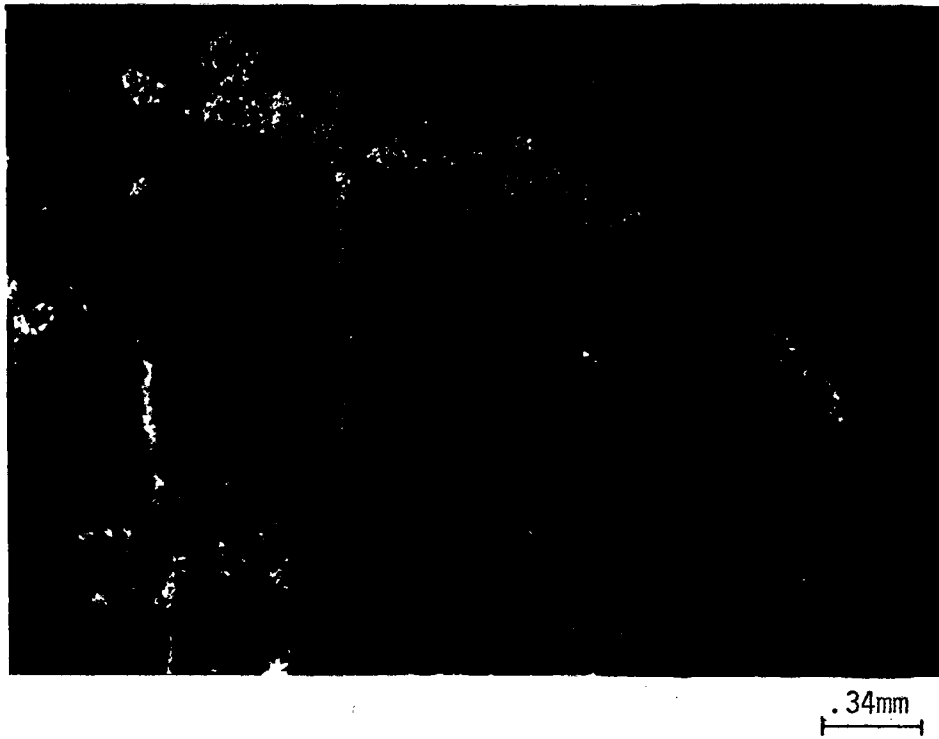


Figure 35. Calcite vein in altered volcanic rock under crossed polars. (SR2A, 500 - 600 feet)



Figure 36. Calcite vein in altered andesite under crossed polars. (H-2, 1100 - 1200 feet)



Figure 37. Calcite vein in diorite/gabbro under crossed polars. (H-1, 100 - 200 feet)

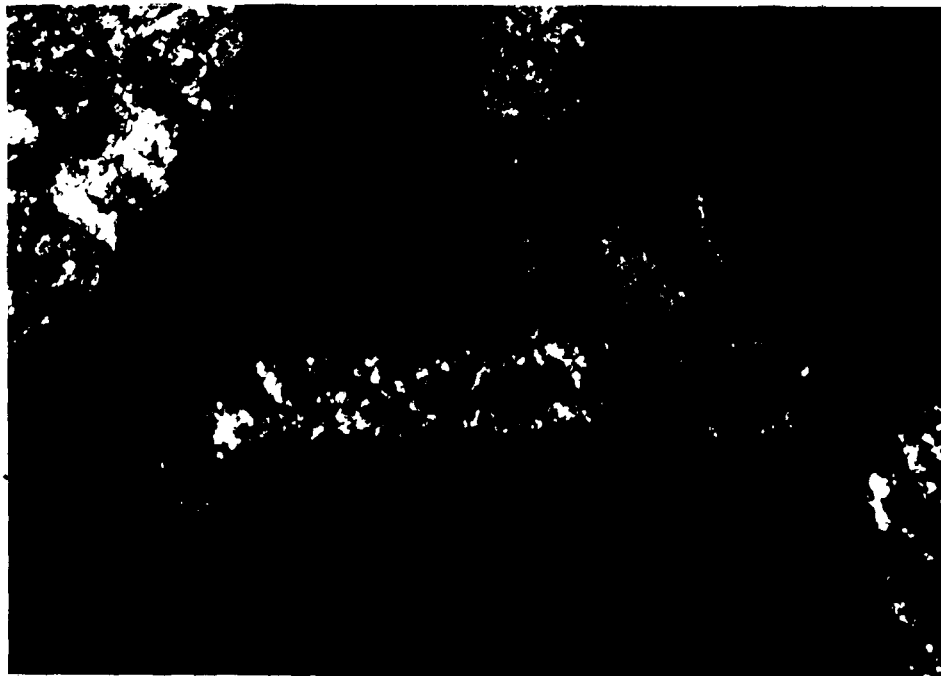


Figure 38. Same as Figure 37 without crossed polars.



.22mm  
|-----|

Figure 39. Epidote vein in altered volcanic rock without crossed polars. (H-2, 400 - 500 feet)



.22mm  
|-----|

Figure 40. Quartz vein in iron-stained volcanic rock under crossed polars. Note the comb structure of the quartz. (DD-9, 600 - 700 feet)