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LA-8515-MS

Informal Report

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**Geology, Hydrothermal Petrology, Stable
Isotope Geochemistry, and Fluid Inclusion
Geothermometry of LASL Geothermal Test
Well C/T-1 (Mesa 31-1), East Mesa,
Imperial Valley, California, U.S.A.**

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This work was supported by the U.S. Department
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UNITED STATES
DEPARTMENT OF ENERGY
CONTRACT W-7405-ENG. 36

LA-8515-MS
Informal Report

UC-66b
Issued: September 1980

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Well C/T-1 (Mesa 31-1), East Mesa,
Imperial Valley, California, U.S.A.**

Prepared for
Los Alamos Scientific Laboratory

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Under Contract N294953H-1
August 1980

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GEOLOGY, HYDROTHERMAL PETROLOGY, STABLE ISOTOPE GEOCHEMISTRY, AND
FLUID INCLUSION GEOTHERMOMETRY OF LASL GEOTHERMAL TEST WELL C/T-1
(MESA 31-1), EAST MESA, IMPERIAL VALLEY, CALIFORNIA, U.S.A.

by

Keith R. Miller and Wilfred A. Elders

ABSTRACT

Borehole Mesa 31-1 (LASL C/T-1) is an 1899-m (6231-ft) deep well located in the northwestern part of the East Mesa Geothermal Field. Mesa 31-1 is the first Calibration/Test Well (C/T-1) in the Los Alamos Scientific Laboratory (LASL) U.S. DOE-Div. of Geothermal Energy, Geothermal Log Interpretation Program.

The purpose of this study is to provide a compilation of drillhole data, drill cuttings, well lithology, and formation petrology that will serve to support the use of well LASL C/T-1 as a calibration/test well for geothermal logging. In addition, reviews of fluid chemistry, stable isotope studies, isotopic and fluid inclusion geothermometry, and the temperature log data are presented. This study provides the basic data on the geology and hydrothermal alteration of the rocks in LASL C/T-1 as background for the interpretation of wireline logs.

I. INTRODUCTION

This study has its origins in technical problems encountered in the drilling and evaluation of the geothermal wells, which are often hampered by lack of experience and knowledge of geothermal systems. These wells are drilled and evaluated using equipment and practices developed by the petroleum industry, which may be inappropriate for geothermal systems. A particular example of this problem is found in wireline logging of boreholes. The physical properties measured may be temperature dependent and the high temperatures encountered in geothermal systems may damage logging equipment. Even more important, the logging tools may respond differently from the response expected in petroleum wells because the rocks themselves are different. In geothermal systems, dissolution and cementation cause significant porosity, permeability, and density changes. At the highest temperatures encountered, hydrothermal alteration may even change the sediments into low-grade metamorphic rocks of greenschist facies (Muffler and White, 1969).

To examine these problems further, the U.S. Department of Energy (DOE), in cooperation with LASL, has made the geothermal well East Mesa 31-1 available for calibration and testing of well logging tools, as part of their Geothermal Log Interpretation Program (GLIP). Mesa 31-1 is a good choice for such a project because its geologic conditions do not deviate greatly from those encountered in petroleum operations. The lithology is a sand and shale sequence and the temperature is less than 200°C. Therefore, this geothermal well is a convenient place to begin log calibration tests.

Bottom-hole temperature is about 165 to 170°C and fluid salinity is approximately 2600 mg/l total dissolved solids (TDS). The lithology is a sequence of alternating sands and shales. The sands appear to be largely poorly cemented and unconsolidated. The sediments above 840 m (2750 ft) form a clay-rich cap over the geothermal reservoir. Montmorillonite and mixed-layer clays are abundant and kaolinite is stable in the cap rock shales. Sand bodies in the cap rock unit have uniformly high porosities and permeabilities and low densities, and produce fluids that are isotopically different from the deeper geothermal waters. At the base of the cap rock an inflection point occurs in the temperature profile. Heating is by conduction in the cap rock and by convective fluid transfer in the underlying reservoir sediments. Below 840 m (2750 ft) the sandstones have decreasing porosities and permeabilities. As in the Mesa 6-2 well, two episodes of post-depositional alteration occur, but the extent of alteration has been much more limited in the Mesa 31-1 well. A very limited episode of silica/pyrite cementation preceded calcite, and occasionally barite, cementation. The extent of cementation increases downhole, but not in a consistent manner. The type and extent of cementation may vary rapidly between nearby sand bodies. These characteristics are also found in sandstones of the 1923-m (6310-ft) to 2728-m (8950-ft) interval of a nearby well, Mesa 38-30.

At temperatures of 150°C, kaolinite is destroyed and authigenic chlorite begins to form. Dolomite seems to disappear from Mesa 31-1 at lower temperatures than other East Mesa or Salton Sea wells.

Reliable temperature data may be difficult to obtain from geothermal wells. In Mesa 31-1, a range of about 20°C is observed in logging runs by five different tools. Fluid inclusion homogenization temperatures are almost identical to present temperatures, indicating that higher temperatures have not occurred in this part of the system in the past. This is in contrast to Mesa 6-2, where fluid inclusion evidence indicates temperatures at least 15°C hotter than the present existed in the past.

Fluids in the well appear to be chemically stratified. A salinity reversal is thought to occur in the middle part of the well. Limited fluid inclusion freezing studies indicate past fluid salinities were similar to present values. Isotopic exchange between fluids and country rock has been slight. Some sandstones are more depleted in ^{13}C than others, probably due to the presence of organic material that is very low in ^{13}C . Oxygen isotopic exchange has caused calcite in the shales at the bottom of the well to be about 5‰ lighter than in near-surface samples. Calcite in sandstone is lighter still, but this is mainly the result of cementation, which has added isotopically lighter calcite to the rock. Temperature calculations using present water and vein calcite isotopic compositions give temperature estimates that are about 50°C less than observed temperatures. The geothermal waters are too depleted in ^{18}O to be in equilibrium with any vein calcite in Mesa 31-1. The earlier fluids that deposited the calcite may have been enriched in ^{18}O by evaporation in ancient lakes (similar to Lake Cahuilla) and then flushed from the system by recharge from the Colorado River.

In contrast to the intensely silicified rocks found in Mesa 6-2, the rocks in Mesa 31-1 seem to be remarkably unaffected by hydrothermal processes. This is probably caused mainly by lower temperatures in this part of the system. Other factors that may contribute to the relative lack of alteration include locally different hydrologic and chemical conditions or a shorter duration of the heating event.

This study attempts to understand the geology and hydrothermal alteration of rocks penetrated by the Mesa 31-1 borehole as a background for interpreting the wireline logs. Optical microscopy and X-ray diffraction studies were made to characterize the downhole lithology and describe the nature and extent of hydrothermal alteration. Stable isotope studies were performed on selected samples of shales, sandstones, and calcite veins. Oxygen isotope studies are particularly useful in determining temperatures of equilibration and describing the water-to-rock ratio involved in a hydrothermal reaction. Fluid inclusion studies supplemented the isotope studies. They provide an excellent geothermometer for comparison with isotope temperatures. By using the freezing-point depression method, it was also possible to determine fluid salinity at the time of mineral formation. These methods give a time dimension to chemical and temperature studies, and they may make it possible to distinguish systems that are heating from those that are cooling. The geothermometers are especially valuable in providing temperature data in rocks disturbed by recent drilling.

See Appendix B for more information on the use of C/T-1 calibration well and how to find the well location near Holtville, California.

II. WELL LOCATION AND PREVIOUS WORK

The East Mesa Geothermal Field is located in the Imperial Valley of Southern California on the eastern flank of the Salton Trough. It is about 13-km east of Holtville (Fig. 1). There are none of the usual surface manifestations of a hydrothermal system, such as geysers, mud pots, or hot springs. The area was discovered solely by coincident geophysical anomalies. There is a 5.5 mgal residual gravity anomaly, which is probably caused by an increase in sediment density associated with cementation and hydrothermal mineralization (Biehler, 1971). The positive gravity anomaly could also be due to emplacement of higher density intrusive igneous rocks (Biehler and Combs, 1972). However, the lack of a corresponding magnetic anomaly suggests that this is not the case (Swanberg, 1976). Thermal gradients were measured in shallow boreholes (an average of 600 ft deep) have values up to 0.18°C/m (10°F per 100 ft) (Combs, 1971). An anomalous electrical resistivity low is also present because both elevated temperatures and the higher fluid salinity generally associated with hydrothermal systems increase electrical conductivity (Meidav and Furgerson, 1972).

Five deep wells were drilled by the U.S. Bureau of Reclamation between 1972 and 1974 (Fig. 2). Other deep wells have been drilled for commercial purposes by Magma Power, Inc. and Republic Geothermal, Inc. since 1975.

Previous studies of the East Mesa geothermal area are abundant. They include geochemistry and geophysical studies of gravity, heat flow, and electrical resistivity, by Rex et al. (1971, 1972), microearthquake and heat-flow studies by Combs (1974, a and b), and telluric mapping by Maas (1975). Stable isotope investigations were carried out by Coplen (1976). Fournier (1973, 1976) has studied cuttings from wells 6-1 and 6-2 by optical and X-ray diffraction methods. The petrology and hydrothermal alteration of well 6-2 was described by Hoagland (1976) and Hoagland and Elders (1977).

Major studies have been performed by TRW Systems Group (1976) and Howard et al. (1978). The geology and commercial development of the geothermal field are discussed by Smith (1979).

Figure 1.
Generalized geologic map of the
Imperial Valley and adjacent areas
(after Hoagland and Elders, 1977).

III. REGIONAL SETTING

The Imperial Valley occupies the central part of the Salton Basin, which is an elongate, triangular-shaped sedimentary basin. The Salton Basin has been isolated from the Gulf of California by the progradation of the Colorado River delta across the axis of the basin (Dibblee, 1954; Longwell, 1954; Van de Kamp, 1973). The Salton Basin is part of a structural province known as the Salton Trough.

The Salton Trough is the landward extension of the Gulf of California (Elders et al., 1972; Elders, 1979; Garfunkel, 1972). It is a region of high seismicity (Hileman et al., 1973), high regional heat flow (Rex, 1966)

Figure 2.
Map of the East Mesa Geothermal Field, showing well locations, lease boundaries and depth (in feet) to the 150°C (300°F) isothermal surface (from Smith, 1979).

and Quaternary volcanism (Robinson et al., 1976). The Salton Trough is a complex rift valley and is thought to be a region of transition from the crustal spreading of the East-Pacific Rise to the transform motion of the San Andreas Fault system. Rhombochasms, or pull-apart basins, occur along a series of en echelon faults. In general, faults in this region strike northwest and exhibit right-lateral strike-slip displacement. This is then a transform fault system which "leaks" because of crustal spreading, thinning, and extension (Elders et al., 1972; Elders and Biehler, 1975; Elders, 1979).

The region exhibits a positive Bouguer gravity anomaly even though there is a very thick fill of low-density sediments. It is suggested that this positive anomaly is due to marked crustal thinning under the trough (Thatcher, in Elders and Biehler, 1975). The trough has steep step-faulted margins and a relatively flat basement floor. The maximum thickness of the sediments is about 6.4 km (Biehler, 1964). However, in the East Mesa area the thickness is less, about 3.5 km (Swanberg, 1976).

Sedimentation in the Salton Basin began in the Miocene. A major marine incursion occurred in late Miocene and Pliocene (Woodard, 1974). The basin then became isolated from the Gulf of California and only minor marine incursions are recorded in the subsequent sedimentary rock record. Deposition and subsidence rates of the basin appear to have been approximately equal (Muffler and Doe, 1968). Sedimentation proceeded as the Colorado River repeatedly changed course, causing alternating cycles of deposition and dessication. The process was probably very similar to that in which the Salton Sea was created by the accidental diversion of the Colorado River in 1905 (Sykes, 1937).

Most of the sediments in the trough are fine-grained sands, silt, and clay deposited by the Colorado River. In general, the sediments show a decrease in grain size and volume percent of sand with increased distance from the apex of the delta near Yuma, Arizona (Randall, 1971; Van de Kamp, 1973). Near the apex of the delta, meandering channel deltaic deposits are dominant. These grade into lacustrine muds, silts, and beach sands toward the northwest distal end of the delta (Van de Kamp, 1973). There is little mineralogic variation in these deposits (Muffler and Doe, 1968; Hoagland, 1976).

Small amounts of coarser sediments are found in marginal alluvial fans and braided stream deposits. These sediments are derived from the nearby mountains that border the trough and are distinctly different from the basin sediments in mineralogical composition (Merriam and Bandy, 1965; Muffler and Doe, 1968; Van de Kamp, 1973).

IV. DRILLHOLE DATA

Borehole Mesa 31-1 was drilled by the U.S. Bureau of Reclamation (USBR) in June 1974, in an area of high heat flow (>7 HFU) about 3.2-km north of the earlier wells 6-1 and 6-2 to evaluate the geothermal potential of this previously undrilled portion of the reservoir. The well was drilled to 309 m (1013 ft) and 13-3/8-in. casing was cemented to that depth. The well was then deepened to 1899 m (6231 ft). Seven and 5/8-in. casing was cemented with slotted liner in the following intervals:

1652-1714 m (5420-5623 ft)
1726-1776 m (5662-5828 ft)
1788-1804 m (5868-5920 ft)
1826-1851 m (5990-6072 ft)
1864-1882 m (6115-6175 ft)

Drill chip cuttings were sampled at 3- to 10-m (10- to 30-ft) intervals. The well was cored from 1720 to 1736 m (5643 to 5969 ft). Unfortunately, the core barrel blocked off and only a few meters of shale and siltstone were recovered.

A temperature probe was run after the well had been shut in for one week. A bottom-hole temperature of 163°C was recorded. The well produced fluids that are predominantly bicarbonate-rich sodium chloride solutions. Total dissolved solids are 2311 mg/l (USBR, 1974).

Geophysical logs were run initially in the uncased hole in 1974. Since the initiation of the GLIP program by LASL in 1977 the hole has been repeatedly logged by tools applicable to cased holes.

V. EXAMINATION OF DRILL CHIP CUTTINGS

A. Problems Involving the Use of Drill Chip Cuttings

The rock materials available for study of this well are mainly drill chip cuttings. A few feet of core exist, but the samples are highly fragmented and consist mainly of gray shale and siltstone.

Many problems complicate the study of cuttings. The most obvious problem is the small sample size; most chips are only a few millimeters long. This means that only very small-scale lithologic features may be seen directly.

An inherent lack of precision is caused by the drilling and sampling methods. In Mesa 31-1, a single sample of cuttings represents a drilled interval of 3 to 10 m (10 to 30 ft). The cuttings are a composite sample that are taken to be representative of the drilled interval. Thus, the lithology can never be resolved in discrete units finer than the sampling interval. Density contrasts between different rock types and cutting size differences cause differential settling velocities within the mud. Therefore, the cuttings may not be representative of the drilled interval. This mixing problem is compounded by slumping of poorly consolidated material from shallower levels in the well. Caliper logs of the uncased hole show that many intervals of the well have an anomalously large diameter. Slumping has involved both sands and shales although unconsolidated or poorly cemented sands are most frequently involved.

Because of slumping and mixing problems, the most reliable indicator of lithology change is the first appearance of a new rock type in the cuttings. Slumping is a possibility if the same rock type persists at deeper intervals. Discrimination between drilled and slumped material is difficult in Mesa 31-1 because the lithology is a repetitive sand and shale sequence. However, because of the increasing induration of shales, and the systematic lithologic and mineralogic changes observed in the cuttings with increasing depth, it appears that, although it may be significant locally, slumping has not obscured the overall changes in lithology and mineralogy.

Another minor problem is that the depth reported was not corrected for lag time, the time taken for the cuttings to travel from the drill bit to the well-head. This means the labelled depth of the cuttings is greater than the true depth. The lag time is a function of depth, the rate of drilling, and the rate of circulation. Lynch (1962) suggests, as a rule-of-thumb, that lag time is approximately one minute for every 27 m (90 ft) of drilled interval. Depth correction would then depend on the drilling rate. Very rough calculations, using averaged drilling rates (Westec, 1978) indicate that depth lags would be less than 23 m (75 ft).

Cuttings are frequently contaminated with casing cement, metal fragments, paint chips, insects, lost circulation material (mica flakes, wood chips, or

other fibrous materials). These can usually be identified and removed. More troublesome are the clays (chiefly bentonites) and barite additives of the drilling mud. These were removed by washing. Unfortunately, washing may remove clay from poorly consolidated shales or mudstones and disaggregate loosely cemented sandstones and thus modify the lithology. Well-indurated rock types remain unaffected.

Another problem occurs when circulation is lost while drilling and the cuttings cannot be recovered.

In spite of these limitations, analyses of cuttings can yield valuable information. Consistent chemical, petrological, and mineralogical trends are observed (Clayton et al., 1968; Muffler and White, 1969). Cores avoid the cutting problems discussed. Unfortunately, coring is very expensive, time-consuming, and potentially hazardous to the borehole.

B. Lithology

The cuttings were examined under a binocular microscope to determine the various rock types present and their abundance. The cuttings are composed mostly of sandstone, unconsolidated sands, and shales. Sandstones and shales are interlayered in beds of varying thickness. Very small-scale interlayering is seen in the cuttings, with layers on the order of millimeters thick. Siltstones are found only in small quantities, never more than 10% of the cuttings. Siltstones are most common in the intervals of 1173 to 1280 m (3850 to 4200 ft) and 1432 to 1493 m (4700 to 4900 ft), where they consistently make up 5 to 10% of the cuttings. In color and appearance, the siltstones do not form a distinct group, but resemble either fine-grained sandstones or coarse-grained shales. Figure 3 is a cuttings log showing the percent total sands observed in the cuttings. Data are not given for the samples above 365 m (1200 ft) because of the poor quality of the samples.

As a check against operator bias in visual estimation of lithology, 23 randomly selected samples were examined twice by the senior author and 10 of these samples were also examined by a second observer. Seventy percent of all repeat measurements by the senior author were within $\pm 5\%$ of the original estimate. Seventy-two percent of the estimates by the second observer were within $\pm 5\%$ of the original estimate.

C. Sands and Sandstones

As mentioned previously, most of the sand in the cuttings is unconsolidated. This is true at all depths in the well. Although this may be partly due to slumping and disaggregation of the sandstones during drilling, it appears that many sandstones penetrated by the well are uncemented or very loosely cemented.

The first well-cemented sandstone occurs at about 838 m (2750 ft). Above this only trace amounts of consolidated sandstone are found. At greater depth both sandstone and more abundant unconsolidated sand exist in the cuttings.

Grain size in the sandstones varies from very fine to medium (Wentworth size scale), with the majority being very fine to fine-grained. Sorting is variable, but in general decreases with increasing grain size. Above 1160 m (3800 ft), most sandstones are red-brown. Below that depth red-brown and light gray to white sandstones are found.

The unconsolidated sands present in the cuttings are highly variable in grain size and sorting. They are very fine to coarse-grained. Larger granules and pebbles occur frequently. Sorting may vary significantly between different sand horizons. The unconsolidated sands are always more poorly sorted than sandstones from the same interval. Sand grains are mostly quartz with small

Figure 3.
Cuttings log showing percent total sands (sandstone plus unconsolidated sand) found in the cuttings, for sample intervals of 3 to 10 m.

Figure 4.
A comparison of volume percent sands per 76 m (250 ft) interval as indicated by the gamma-ray log (TRW Systems Group 1976) and the cuttings log (Fig. 3). The cuttings have generally lower sand abundance than indicated by the geophysical logs, especially in the lower part of the well.

amounts of feldspar. The grains are generally subangular to subrounded, with occasional intervals of more rounded or angular grains.

Pebbles occur sporadically above 1524 m (5000 ft). Many pebble occurrences are probably due to slumping from shallower levels. A conglomeratic sand occurs in the cuttings at about 945 m (3100 ft). Pebble-bearing horizons occur at 290 m (950 ft) and between 1280 and 1310 m (4200 and 4300 ft). Varicolored chert pebbles are the most common pebble-sized fragments. Various volcanic and granitic rocks, schists, and gneisses are also found in smaller amounts.

It was impossible to correlate the cuttings log (Fig. 3) with geophysical logs that discriminate sand and shale horizons, notably self-potential (S.P.), gamma-ray and Saraband logs*. There is good agreement among the geophysical logs, but lithologic boundaries cannot be correlated with the cuttings log, nor do the cuttings have the same abundance of rock types. Sands are under-represented in the cuttings relative to the wireline logs, and this bias against sand increases with depth. Similar results have been found for wells Mesa 6-1 and Mesa 6-2 (Hoagland, 1978; Fournier, 1973, 1976). This is illustrated in Figure 4. These graphs compare volume percent sands per 76-m (250-ft) intervals as indicated by the gamma-ray log (TRW Systems Group, 1976) and the cuttings log (Fig. 3). Below 1295 m (4250 ft) the disparity between the logs is extreme. We believe the geophysical logs to be more reliable in this case because most

*The Saraband log is a proprietary log of the Schlumberger Company which calculates porosity and other factors from various other logs.

methods of sampling cuttings lose fine-grained sands through the shaker table screens (Hills, 1949). A large percentage of the total sand in the cuttings is unconsolidated and much of it appears to have been lost in the sampling process. This bias against sands makes it impossible to use the cuttings as a factual indicator of the relative proportions of the lithologies drilled.

D. Faulting in Well Mesa 31-1

Using geophysical data, many faults have been inferred to exist in the vicinity of East Mesa (Smith, 1979; Howard et al., 1978). Both strike-slip and normal faults are postulated. Of particular interest in this study is a fault which, using interpreted dipmeter data, was inferred to be present in Mesa 31-1 at 1704 m (5590 ft) (TRW Systems Group, 1976, p. A-47). There is an abrupt and well-defined change of attitude of the dipmeter data at that point. Electric log correlations with a nearby well, Mesa 38-30, 400 m to the east, (Fig. 2), suggested to us that the section had also been shortened by at least 30 m (100 ft). This is interpreted as a strike-slip fault, and in contrast to the normal faults in the area, it is believed to act as a partial barrier to fluid flow on the basis of well tests (TRW Systems Group, 1976).

A possible alternative explanation for the dipmeter data is that the dips are not controlled by faulting, but by soft-sediment deformation and slumping. Evidence for this is seen in the deformed shales and siltstones in core samples from 1721 m (5645 ft).

Although the cuttings were examined closely with a binocular microscope to check for any evidence of this fault, none was seen. Unfortunately, the cuttings are small in size near this interval because the drill bit was badly worn (Driller's log, Westec Services, 1978). There is no obvious evidence in the cuttings of a fault near the 1704-m (5590-ft) interval: no signs of fault gouge, no zone of mineralization, and no significant change in the character of the rocks being drilled. No loss of circulation was reported while drilling through this interval (Driller's log, Westec Services, 1978). On the sonic log at the depth of the postulated fault there is a thin shaly zone (about 1 m thick), which has an apparently very low-sonic velocity. This may be an anomalous reading, however, because it also appears to be a slightly washed-out zone on the caliper log. If this thin zone truly is a fault zone, it would be very difficult to detect in the cuttings because of mixing and sampling problems.

VI. SANDSTONE PETROLOGY

Seventy-one polished thin sections were prepared from sandstone fragments hand-picked as being representative of the cuttings. Samples were collected at frequent intervals through the well and from all intervals where sandstones were abundant in the cuttings. Several fragments from each of the chosen intervals were mounted in an epoxy plug. The plug was ground and polished to standard thickness and examined with a petrographic microscope.

In addition, hand-picked sandstone samples were analyzed by X-ray diffraction to determine the mineralogical modal composition. Samples for X-ray diffraction were analyzed on an automated Picker X-ray diffractometer using $\text{CuK}\alpha$ radiation, a graphite crystal monochromator, and a scanning speed of $2^\circ 2\theta$ per minute. This instrument is interfaced with a small computer and graphics display for collecting, interpreting, and displaying data. A quantitative modal analysis is calculated by the computer and subsequently corrected by a skilled operator using an interactive computer terminal (Johnson et al., 1977). The

data may also be plotted on a spindle diagram, which shows the downhole variations in mineralogy.

The key to mineral abbreviations, the detection limit and "standard error" for each mineral is listed in Table I (Elders et al., 1978). The "standard error" is a measure of precision and accuracy, and represents one standard deviation from the mean values obtained in numerous calibration runs. For additional information see Elders et al. (1978).

The bulk mineralogy of X-ray analyses of sandstones are shown in Fig. 5 and Table A-I in App. A. It should be remembered that X-ray analysis does not by itself distinguish between detrital, diagenetic, and hydrothermal minerals.

Most of the sandstones may be classified as subarkoses or subarenites, depending on the amount of lithic clasts present. Wackes (>15% matrix) are present, but less abundant. Lithic clasts are variable in abundance and type. Chert is the most abundant type. Also common are small rounded shale particles, which were locally derived. Fragments of volcanic and metamorphic rocks are much less common in rocks studied petrographically.

Graded bedding is observed in many thin sections. In some samples an abrupt change from a fine-grained to a coarser-grained layer is observed.

A. Bulk Mineralogy

Detrital quartz is the most abundant mineral in all samples below 853 m (2800 ft). It occurs mostly as monocrystalline grains, which frequently exhibit undulatory extinction. Some quartz occurs as syntaxial authigenic overgrowths on detrital quartz grains. This will be discussed later.

Feldspars are the next most common framework grains. Most of the alkali feldspar is detrital microcline. Alkali feldspar is always more abundant than plagioclase.

Calcite occurs mostly as intergranular authigenic cement, although some detrital clasts and fossils are occasionally seen. The proportion of calcite present varies greatly according to the degree of cementation, but in general decreases with depth.

X-ray diffraction indicates dolomite to be present in the upper part of the well. Most dolomite occurs between 975 and 1280 m (3200 and 4200 ft). Isolated occurrences are found at about 869 m and 1402 m (2850 ft and 4600 ft). It is absent at greater depths. Ankerite is indicated in two samples from 1030 and 1082 m (3380 and 3550 ft). Dolomite and ankerite were not positively identified in thin-section studies. Rhomb-shaped carbonates were occasionally seen in sandstones that had only a little cement, but distinguishing tests were not performed.

Pyrite and barite occur locally as cementing minerals. Barite was tentatively identified optically in two samples and by X-ray diffraction in four samples. Pyrite was easily seen macroscopically in the cuttings and seems to be most abundant in the sandstones between 1067 and 1219 m (3500 and 4000 ft).

Kaolinite is present in most sandstones above 1631 m (5350 ft). It is absent below that depth. Chlorite is indicated by X-ray diffraction to be present only below 1631 m (5350 ft). However, small amounts of detrital chlorite are seen in many thin-section samples from more shallow levels. Illite is present in small amounts throughout most of the well.

Several accessory detrital minerals occur. They include muscovite, biotite, garnet, tourmaline, chlorite, zircon and undifferentiated opaque minerals. These minerals are sporadically distributed with the exception of a single sample from 1521 m (4990 ft), which exhibits a zone of accumulated zircons and opaque minerals (pyrite?).

TABLE I
X-RAY DIFFRACTION:
MINERAL AND DETECTION LIMITS AND STANDARD ERRORS
(MODIFIED FROM ELDERS et al., 1978)

Symbol	Mineral	True Value, %	Standard Error (1 S.D.), %	Detection Limit, %
QUAR	Quartz	60	±3.1	<1 ^a
PLAG	Plagioclase	10	±2.7	<2 ^a
ALKA	Alkali Feldspar	10	±2.6	<3 ^a
MICA	Biotite and/or muscovite	5	±1.0	>2
ILLI	Illite	5	±1.0	>2 ^b
1 MIX	Illite/montmorillonite mixed layer clay, 25% expandables	--	---	>5 ^a
2 MIX	Illite/montmorillonite mixed layer clay, 50% expandables	--	---	>5 ^a
MONT	2 H ₂ O montmorillonite	--	±5.5 ^a	>5 ^a
KAOL	Kaolinite	5	±1.0	} greater if both are present
CHLO	Chlorite	10	±1.5	
CALC	Calcite	10	±1.9	
DOLO	Dolomite	5	±1.5	>2
AKER	Ankerite	--	---	--
ZEOL	Wairakite and/or analcite	5	±2.7	<1
TALC	Talc and/or pyrophyllite	5	±1.8	<1
PRYI	Pyrite	5	±3.7	>4
HEMA	Hematite	--	---	--
PHRR	Pyrrhotite	--	---	--
AMPH	Hornblende	5	±1.0	<1
EPID	Epidote	5	±1.0	>4
PREH	Prehnite	5	±1.0	>4
ANHY	Anhydrite	5	±2.1	<1
GYPS	Gypsum	5	±2.0	<1

--Not Determined

^a Value reported in Elders et al., 1977

^b Estimated value

VII. SANDSTONE CEMENTATION AND POST-DEPOSITIONAL ALTERATION

The most significant post-depositional features of the sandstones are the cementation effects. Several cementing minerals are found in the rocks. Calcite is by far the most abundant. Quartz, pyrite, and barite cements also occur. Where more than one cementation episode has occurred, textural evidence indicates that quartz and/or pyrite cementation has preceeded calcite and/or barite cementation. These are the same relationships found by Hoagland (1976) in well Mesa 6-2. Kaolinite disappears and authigenic chlorite begins to form near the bottom of the well.

Figure 5.

XRD spindle diagram showing downhole variations in bulk mineralogy of sandstones. The width of the black bars is proportional to the abundance of the mineral concerned. SAMP shows the location of X-rayed samples. The mineral symbol codes are shown in Table I. BARI = Barite.

A. Quartz

Authigenic quartz occurs as optically continuous syntaxial overgrowths on detrital quartz grains. Below 1768 m (5800 ft), it also occasionally occurs as small elongate, euhedral crystals growing on detrital quartz grains. In this habit the authigenic quartz crystals are in optical continuity with the detrital grains and elongate parallel to the "c" crystallographic axis. No epitaxial quartz overgrowths were found on detrital feldspar grains, nor were any syntaxial feldspar rims found. Quartz veins were not found in the cuttings.

Authigenic quartz cementation begins at about 1176 m (3860 ft) where euhedral quartz overgrowths are found on some detrital grains.

The degree of quartz cementation increases with depth, but not in a consistent manner. Quartz overgrowths may be quite abundant at one depth and virtually absent at some deeper interval. This hints that the aquifers might be stratified chemically with respect to silica saturation and thus horizontal flow may be dominant in the system.

In most cases the formation of quartz overgrowths has been quite limited. Rocks subjected to quartz cementation but without calcite cement almost always have high porosities. Only a single small chip from 1841 m (6040 ft) exhibited quartz cementation, which had drastically reduced porosity by filling in the pores.

Grain-to-grain contacts at shallow depths are mostly point contacts. Long contacts and concavo-convex contacts become slightly more abundant with depth, and much more abundant with increasing quartz cementation.

B. Pyrite

Pyrite first appears in the cuttings at 390 m (1280 ft). It is also seen macroscopically in some of the coarse-grained shales. As a cementing mineral, pyrite is most abundant between 1067 and 1219 m (3500 and 4000 ft). Some of the associated framework quartz grains have authigenic quartz overgrowths. More commonly pyrite occurs as a vein mineral that is frequently found below 975 m (3200 ft). Pyrite veins are consistently present in the cuttings from 1082 to 1219 m (3550 to 4000 ft). Below 1219 m (4000 ft) they are less common, but are still found at the bottom of the well.

C. Chlorite

X-ray diffraction shows chlorite to be consistently present below 1798 m (5900 ft), and sporadically present between 1646 to 1798 m (5400 to 5900 ft). In thin section, authigenic chlorite is seen as epitaxial coats on detrital quartz grains in highly porous sandstones below 1829 m (6000 ft). The chlorite coat consists of tiny individual crystals, which are elongate perpendicular to the grain surface. The chlorite coats are not present on all exposed grain surfaces in a given sample, but are sporadically distributed and thus are not so extensive as to inhibit quartz cementation.

D. Mica

A pale green authigenic mica is occasionally seen in some of the samples below 1676 m (5500 ft). This mica occurs as coatings on detrital quartz or feldspar grains in slightly impure sandstones or replacing clay minerals in the matrix material.

E. Calcite

Calcite cement occurs throughout the well and is extremely variable in abundance. It occurs in a complete size range from micrite to large, optically continuous crystals that enclose many detrital grains. The fine-grained calcite has frequently recrystallized into larger sparry crystals. This process often causes rims of enlarged calcite crystals to form around detrital minerals or lithic clasts. The cement grain size appears to be largely controlled by permeability. Good sorting and high porosities result in higher initial permeabilities, and such rocks usually have larger calcite cement crystals than do rocks with lower initial permeabilities.

Calcite cement is occasionally observed as discrete small, elongate crystals that are growing into pore space. Below 1824 m (5985 ft) calcite is occasionally seen attacking detrital framework grains. This has been very limited and the replacement textures are very poorly developed.

F. Barite

Barite occurs at irregular intervals in the well. It was identified in two X-ray samples and two thin sections between 896 m (2850 ft) and 1024 m (3360 ft), and again by X-ray at 1242 m (4075 ft) and 1268 m (4159 ft).

In thin section, barite is seen as a cementing mineral in loosely packed medium-grained sandstones. It occurs as a single large optically continuous crystal that encloses all the framework grains of the sample chip.

G. Vein Calcite

Vein calcite is found sporadically between 549 m (1800 ft) and 1615 m (5300 ft). Below 1615 m (5300 ft) it is found consistently in almost every sample of cuttings. It is usually clear or white, and occasionally red or gray with veinlets that are usually thin (0.5 to 3 mm thick) and tabular. However, a large (8 mm long) euhedral dogtooth calcite crystal was found at 1720 m (5640 ft), which indicates that some fractures larger than 8 mm are present.

No veins were observed in the sandstones. They are restricted to shales and siltstones or the shale-sandstone lithologic boundary. This is probably because the sands are so poorly indurated that brittle fracture does not occur.

H. Ferric Oxides

Ferric oxides occur as reddish, dusty coatings on detrital grains. This is most easily seen in the relatively uncemented sandstones above 1524 m (5000 ft). These ferric oxide coatings occur in calcite cemented sandstones and are probably responsible for their red-brown color.

VIII. DEEP SANDSTONES OF MESA 38-30

It would be desirable to know what changes in sandstone petrology occur at depths greater than those reached by Mesa 31-1. A much deeper well, Republic Geothermal Mesa 38-30, was drilled only 381 m (1250 ft) to the northeast of Mesa 31-1 (Fig. 2). Mesa 38-30 is 2736 m (8976 ft) deep and is about 10°C hotter than Mesa 31-1. Excellent correlation has been established between the two wells using well logs (TRW Systems Group, 1976). A reconnaissance petrographic study was made of samples selected from twelve sandstone horizons between 1923 m (6310 ft) and 2728 m (8950 ft). The petrology of these rocks should be quite similar to rocks that would be encountered in the 1981 m (6500 ft) to 3048 m (10000 ft) interval below Mesa 31-1.

A. Silicate Minerals

The degree of quartz cementation increases with depth, although in an irregular fashion. Syntaxial quartz overgrowths become larger and more abundant. The discrete quartz crystals growing on detrital grains are larger than those seen in Mesa 31-1 and they become larger with increasing depth.

Between 2438 m (8000 ft) and 2536 m (8320 ft), the sandstones are highly cemented with silica. The grains become tightly interlocking with planar to sutured contacts.

From 2600 m (8530 ft) to 2706 m (8880 ft) at least two distinct types of sandstones are present. There is a fine-grained, poorly sorted variety with limited quartz cement and a coarser-grained variety that has been highly quartz cemented. The fine-grained varieties commonly contain pale green mica. The abundance of this mica seems to vary inversely to abundance of calcite cement. The intensely silica cemented sandstones are quite similar to those from Mesa 6-2 described by Hoagland (1976). No authigenic adularia or epitaxial quartz overgrowths on feldspar grains were observed. No epitaxial chlorite coats were observed. Some coarse-grained authigenic chlorite crystals were seen at 1923 m (6310 ft).

The trend of increasing silica cementation is suddenly terminated in the samples from 2719 m to 2728 m (8920 to 8950 ft). These are the deepest sandstones sampled and they consist entirely of calcite cemented, fine-grained sandstones. The formation of quartz overgrowths has been extremely limited. The calcite appears to be actively attacking the quartz grains.

B. Calcite

Calcite cement occurs as before in a variety of grain sizes. It is most abundant where silica cementation has been least intense. A few sands have been little cemented by either quartz or calcite.

The general trend of increasing silicification is disturbed by zones that are less silicified and contain more calcite cement. The most notable example of this is the change from intensely silica cemented sandstones at 2697 m (8850 ft) to the almost unsilicified, calcite-cemented sands at 2719 m (8920 ft).

C. Barite

Barite was also found in Mesa 38-30 at 2015 m (6610 ft) and 2444 m (8020 ft). The barite cement was deposited after an episode of silica cementation. Some quartz grains with euhedral overgrowths are entirely surrounded by the barite cement. Calcite cement is also present in part of the sample, but there is insufficient textural evidence to discern which came first.

IX. SHALES

The shales have been divided into three distinctive color groups: tan, red-brown, and gray. Gradational tan to red-brown or red-brown to gray varieties are much less abundant and yellow, green or orange shales occur only in trace amounts.

Figure 6 shows the occurrence of the three main shale colors according to depth. This is similar to the distribution found by Hoagland (1976) in Mesa 6-2. Tan shales are abundant above 914 m (3000 ft), but occur only in minor amounts below that depth. Red-brown shales occur sporadically above 610 m (2000 ft), but are abundant below that depth. Gray shales are common below 762 m (2500 ft).

In general, the shales are slightly silty and minor amounts of very coarse varieties also occur. These very coarse shales and siltstones are often found with dark micaceous (chlorite?) or carbonaceous laminations. Euhedral pyrite crystals are found in some gray shales.

Red-brown shales are frequently found mottled with gray spots, or altered to gray shale along fractures and sandy lamellae. The color change probably is the result of oxidation-reduction reactions. Virtually all tan and red-brown shales contain hematite, but gray shales do not. Gray shales and siltstones frequently contain pyrite.

Figure 6.
Color occurrence of shales versus depth.

A. X-ray Diffraction Analysis of Shale

Homogeneous samples of each shale type were hand-picked from the cuttings and X-rayed to determine their bulk mineralogy. The spindle diagrams are shown in Figs. 7, 8, and 9 for tan, red-brown and gray shales, respectively. The computer-calculated modal analyses are listed in App. A. The data concerning the various clay minerals (ILLI, IMIX, 2MIX, MONT, KAOL

Figure 7.

XRD spindle diagram showing bulk mineralogy of tan shales. The width of the black bars is proportional to the abundance of the mineral concerned. SAMP shows the location of samples X-rayed. The mineral symbol codes are shown in Table I. BARI = Barite.

Figure 8.

XRD spindle diagram showing bulk mineralogy of red-brown shales. The width of the black bars is proportional to the abundance of the mineral concerned. SAMP shows the location of samples X-rayed. The mineral symbol codes are shown in Table I. BARI = Barite.

and CHLO) are not precise in these bulk analyses. To obtain better diffractograms, clay-size fractions also extracted from every second sample and oriented clay mounts were prepared. The samples were prepared by crushing and sonic disaggregation. The >50 micron fraction was removed by sieving and the <2 micron fraction was separated by centrifuge. This latter fraction was air dried and then settled onto glass slides in a slurry. The samples were X-rayed from 2.6° to 40° 2θ at 2° 2θ per min. using CuKα radiation. Because of equipment failure with the automated Picker diffractometer, these samples were X-rayed on Norelco diffractometers. The samples were glycolated and run again on the same equipment.

Figure 9.

XRD spindle diagram showing bulk mineralogy of gray shales. The width of the black bars is proportional to the abundance of the mineral concerned. SAMP shows the location of samples X-rayed. The mineral symbol codes are shown in Table I. BARI = Barite.

B. Clay Mineral Identification Methods

The clay minerals were identified by examining the basal X-ray reflections. Kaolinite (Kaol) was identified by the (001) 7.15 Å peak and the (002) 3.58 Å peak. Chlorite (Chl) was identified by the 14 Å peak (001) and by a peak of about 3.56 Å (004). The chlorite (001 to 003) peaks are often masked by reflections from other minerals. Therefore, the samples were scanned at slow speed ($1/2^\circ$ 2θ per min.) from 24° to 26.5° 2θ to resolve the Kaol (002)/Chl (004) peaks.

Discrete montmorillonite (Mo) was identified by the (001) 15 Å peak (corresponding to 2- H_2O interlayers), which shifts to about 17 Å upon glycolation. The unglycolated peak may be at or shifted toward 12.5 Å. This indicates that 1- H_2O interlayer montmorillonite is present. Discrete illite (Ill) is identified by the (001) peak at about 10 Å, which is not shifted by glycolation.

Interlayered illite/montmorillonite clays were identified using certain composite peaks that represent average reflections from different basal spacing. Unglycolated samples exhibit a hump or shoulder on the low-angle (10 to 15 Å) of the Ill(001) peak. This hump is composed of reflections from a nonintegral

series of interlayered I11(001) reflections and Mo(001) reflections (Weaver, 1956). Upon glycolation a composite peak will appear between 8.7 and 10.4 Å, generally on the high angle side of the I11(001) peak. This is the composite I11(001)/Mog(002) peak, where Mog indicates the expanded (glycolated) 17 Å montmorillonite lattice. The position of this peak was used to estimate the percentage of expandable layers (Reynolds and Hower, 1970).

The percentage of expandable layers may also be estimated from the position of the I11(001)/Mog(001) peak using the mixing curve of Weaver (1956). However, data obtained from Weaver's (1956) curve may lead to serious errors because: (1) the I11(001)/Mog(001) peak is often obscured by the presence of chlorite, and (2) the model assumes entirely random interlayering. When the percentage of expandables drops below 40%, the interlayers may be ordered and this gives rise to ordered reflections (Reynolds and Hower, 1970). Nevertheless, where it was possible to use the two methods there was general agreement.

It was not possible to determine the iron and magnesium composition of chlorite in shales using the method of Petruk (1964) because of the consistent presence of kaolinite. However, it was possible to utilize this method for clay separations from three deep sandstones. The intensities of the (002) and (004) peaks are added and then divided by the (003) peak intensity. This ratio can then be related to the number of heavy atoms (Fe, Mn, Cr) present in the octahedral layers (Table II). These limited data suggest an increase in the heavy atom content of chlorite with depth.

No attempt was made to estimate quantitatively the abundance of the various clay minerals in these oriented clay mounts. The calibration and detection studies performed with the automated Picker diffractometer could not be easily used with the Norelco diffractograms and there are very large discrepancies among some of the other methods presented in use (Pierce and Siegel, 1969).

C. Bulk Mineralogy

All shale samples contain quartz, plagioclase, potassium feldspar and calcite. Dolomite is present in all tan shales, but disappears from the red-brown and gray shales below 1524 m (5000 ft). Dolomite persists in the gray shales to greater depths than the red-brown shales. Virtually all tan and red-brown shales contain hematite. Some deep gray shales contain pyrite and all lack detectable amounts of hematite.

TABLE II
HEAVY ATOM (Fe, Mn, Cr)
CONTENT OF CHLORITE

Sample No.	Depth		No. of Heavy Atoms in Octahedral Layers (6 possible) ^a
	(m)	(ft)	
096	1761-64	5777-87	2.1
098	1824-27	5985-95	2.9
100	1859-62	6098-10	4.4

^a The number shown is an average of the glycolated and unglycolated patterns.

D. Clay Mineralogy

Clay mineral assemblages are shown in Tables III and IV.

E. Tan Shales

Tan shales are characterized by abundant discrete montmorillonite. Both 1-H₂O and 2-H₂O layer montmorillonites are indicated by the presence of a (001) 12.5 Å (001) 15 Å composite peak which expands to 17 Å upon glycolation (Fig. 10a). Interlayered illite:montmorillonite is present and contains generally high percentages of expandable layers. Discrete illite is also present in all samples.

Kaolinite is present in all samples. Chlorite is present below 610 m (2000 ft), but is less abundant than kaolinite.

F. Red-brown Shales

Kaolinite and chlorite are present in all red-brown shales. Kaolinite is predominant over chlorite to about 1585 m (5200 ft). This situation is reversed below 1737 m (5700 ft). Chlorite clearly becomes the more dominant phase and kaolinite is sharply reduced.

Discrete montmorillonite is abundant near 731 m (2400 ft), but decreases sharply in abundance below that depth and disappears at about 1219 m (4000 ft). This is thought to be mostly 2-H₂O montmorillonite (Fig. 10b).

Discrete illite occurs in all samples and increases in abundance with depth. Interlayered illite/montmorillonite also occurs in all samples. It increases in abundance with the loss of discrete montmorillonite. The illite and mixed-layer diffractogram trends are illustrated in Fig. 11. The percentage of expandable layers shows a general slight decrease with depth.

Three zones may be distinguished in the red-brown shales: (1) a shallow zone above 1219 m (4000 ft) in which montmorillonite is present and illite occurs in small amounts; (2) a medium depth zone, 1219 to 1737 m (4000 to 5700 ft) which has more abundant illite and interlayered illite/montmorillonite; and (3) a deep zone below 1737 m (5700 ft) which has abundant illite and chlorite.

G. Gray Shales

The gray shales resemble zones 2 and 3 of the red-brown shales. Montmorillonite is absent. Illite is consistent in all samples. Kaolinite and chlorite are present in all samples. Chlorite gradually increases in abundance and exceeds kaolinite at 1707 m (5600 ft). Most of the interlayered illite/montmorillonite clays have 30 to 55% expandable layers. However, several samples between 1311 and 1743 m (4300 and 5720 ft) have unusual diffraction peaks at 10 Å, which are sharply or slightly reduced upon glycolation (Fig. 12). These are marked with an asterisk in Table I. This has been interpreted as a phase containing <10% expandable layers. This causes a high intensity peak in the unglycolated sample. Upon glycolation, the layers expand and the peak intensity is reduced to the normal Ill(001) intensity. These diffractogram trends may be seen in Fig. 13.

X. TEMPERATURE MEASUREMENTS IN MESA 31-1

At first glance, it seems that measuring borehole temperatures would be simple and straightforward. This is not true, however. Different logging tools vary in response and there is increased risk of tool failure in the high temperatures encountered in geothermal wells.

Mesa 31-1 has been logged for temperature on at least seven separate occasions: In 1974 (USBR, 1974), twice in 1977 (Urban et al., 1978), and four times

TABLE III

CLAY MINERALOGY OF TAN SHALES

Sample No.	Depth		Chl	Kao1	Ill	Mo	Ill/Mo	Expandables
	(m)	(ft)						%
106	392-403	1286-1321	-	t	t	++	+	>50
103	593-602	1944-1975	-	t	t	++	+	>50
104	631-641	2070-2102	t	≤ t	t	++	+	80
112	726-735	2382-2413	t	≤ t	t	++	+	>50
105	821-830	2693-2722	t	< t	+	++	+	60-80

RED-BROWN SHALES

002	726-735	2382-2413	t	< +	t	++	+	>40
004	868-878	2848-2880	t	< +	t	+	+	40-80
008	1104-1108	3621-3635	t	< +	t	t	+	40
010	1194-1198	3919-3929	t	< +	+	t	+	50
012	1242-1245	4075-4085	t	< +	+	-	+	50
014	1289-1292	4230-4240	t	< +	+	-	+	60
016	1335-1340	4380-4395	t	< +	+	-	+	50
018	1376-1379	4516-4526	t	< +	+	-	+	45
020	1492-1495	4895-4905	t	< +	+	-	+	40
022	1574-1578	5164-5177	t	< +	+	-	+	50-60
024	1742-1745	5714-5724	+	> t	+	-	+	30
026	1811-1815	5943-5954	+	> t	++	-	+	30
028	1843-1846	6047-6057	+	> t	++	-	+	40
030	1872-1875	6141-6151	+	> t	++	-	+	25-30

t = present in trace or small amounts
+ = present in moderate amounts
++ = present in abundant amounts
- = absent

TABLE IV

CLAY MINERALOGY OF GRAY SHALES

Sample No.	Depth		Chl	Kao1	Ill	Mo	Ill/Mo	Expandables
	(m)	(ft)						%
034	726-735	2382-2413	t < t	+	-	+		>40
036	1030-1040	3381-3412	t < t	+	-	+		30-50
038	1104-1108	3621-3635	t < t	+	-	+		30-50
040	1194-1198	3919-3929	t < t	+	-	+		40-60
044	1302-1305	4271-4281	t < t	+	-	+		40-60
046	1356-1358	4448-4457	t < t	+	-	+		*40-80
048	1402-1406	4600-4612	t < t	+	-	+		*40
050	1472-1475	4828-4839	t > t	+	-	+		*30-40
052	1520-1524	4988-5000	t ≈ t	+	-	+		*20-45
054	1600-1603	5249-5259	t ≈ t	+	-	+		*30-50
056	1649-1654	5410-5426	t ≈ t	+	-	+		*20-40
058	1711-1714	5613-5623	t ≥ t	+	-	+		*30-60
060	1742-1745	5714-5724	t ≥ t	+	-	+		*30-50
062	1811-1815	5943-5954	t ≥ t	+	-	+		30-50
064	1843-1846	6047-6057	t > t	+	-	+		30-50
066	1872-1875	6141-6151	+ >> t	+	-	+		50

* Denotes presence of a phase with <10% expandable layers (See Fig. 12).
t = present in trace or small amounts
+ = present in moderate amounts
++ = present in abundant amounts
- = absent

Figure 10.

- (a) Example of a 1-H₂O and 2-H₂O composite peak at about 14Å. Upon glycolation the montmorillonite expands to about 17Å (dashed line in glycolated sample). Tan shale from 631 to 641 m (2070 to 2102 ft).
- (b) Part of a diffractogram of red-brown shale from 868 to 878 m (2748 to 2880 ft) containing 2-H₂O montmorillonite, which expands to 17Å on glycolation (dashed line in glycolated sample). Interlayered illite/montmorillonite forms large hump on low-angle side of Ill(001) peak in unglycolated samples. After glycolation a broad Ill(001)/Mog(002) composite peak forms on left of Ill(001) peak.

This is from zone 1 of the red-brown shales.

Figure 11.

General diffractogram trends of red-brown shales. Samples contain abundant inter-layered illite/montmorillonite as indicated by hump on low-angle side of Ill(001) peak in unglycolated samples (dashed line in glycolated sample). Upon glycolation a composite Ill(001)/Mog(002) peak forms on the high-angle side of the Ill(001) peak. The abundance of illite increases slightly with increasing depth.

11(a) is from 1492 to 1495 m (4895 to 4905 ft), zone 2 of the red-brown shales.

11(b) is from 1843 to 1846 m (6047 to 6057 ft), zone 3 of the red-brown shales.

Figure 12.

Gray shale from 1640 to 1654 m (5410 to 5426 ft) containing abundant interlayered illite/montmorillonite. Sample is thought to contain an interlayered phase containing less than 10% expandable layers because of the decrease in intensity of the Ill(001) 10Å (ΔI) which occurs upon glycolation. Sample contains other illite/montmorillonite with 40% or more expandable layers.

Figure 13.

General diffractogram trends of gray shales with increasing depth. The interlayered illite/montmorillonite hump on low-angle side of Ill(001) contracts and becomes more intense. In glycolated samples (dashed line) the Ill(001)/Mog(002) composite peak shifts toward Ill(001) with depth, indicating a decrease in the percent expandable layers of the interlayered clay. Illite abundance remains nearly uniform with depth.

13(a) Gray shale from 1030 to 1040 m (3381 to 3412 ft).

13(b) Gray shale from 1872 to 1875 m (6141 to 6151 ft).

in 1978 and 1979 by different service companies as part of the Geothermal Log Interpretation Program (Mathews et al., 1979). No temperature data are available for one of the latter logging runs because of tool failure. The other temperature profiles are shown in Fig. 14. Fluid inclusion filling temperatures, which are discussed in the following section, are also shown on this graph. Temperatures recorded at 152-m (500-ft) intervals from all five logs are listed in Table V.

It is immediately apparent that the temperature profiles are not duplicated by different tools. At the same depth the measured temperatures vary by as much as 20°C. This does not appear to be caused by temperature changes in the well. Urban et al. (1978) used the same tool for two logging runs spaced 72 days apart and found the average temperature changed only about +0.2°C. This is negligible compared to the discrepancies between the other logs and the wide range of fluid inclusion temperatures.

There is reasonable agreement among the temperature logs when one ignores the GO Wireline measurements, that are consistently 10 to 20°C higher than the others. This emphasizes the importance of properly calibrating logging tools for the higher temperatures encountered in geothermal systems.

XI. FLUID INCLUSION GEOTHERMOMETRY

Fluid inclusions are small portions of fluid which were trapped in a solid crystal as it grew or recrystallized. The fluids are thus samples of the original brine from which the crystal grew or a later fluid which bathed the crystal during recrystallization (Roedder, 1972). If the crystal cools, the liquid will contract at a faster rate than the solid and a vapor bubble will form. Sorby (1858) was the first to recognize that this bubble-forming process could be reversed to determine the temperature of mineral formation. The inclusion is heated until the fluids homogenize in a single phase (i.e. the bubble disappears). This homogenization temperature (T_H) must then be corrected to the equivalent temperature at the original confining pressure because higher pressure requires a higher temperature to achieve homogenization (Potter, 1977).

The inclusions may also be frozen to determine the salinity of the fluid by depression of the freezing point (Roedder, 1962, 1963). As salinity increases, the freezing point decreases. The inclusions are frozen and slowly warmed until the last crystal of ice melts. This melting temperature (T_M) is taken as the freezing point (Roedder, 1962). NaCl equivalent salinity is calculated using the equation of Potter et al. (1978).

Because formation temperature and fluid salinity may change with time, it is important to know the origin of the fluid inclusions. Primary inclusions are those that form during growth of the host crystal. Secondary inclusions form at some later time, often in healed fractures or along cleavage planes. Pseudo-secondary inclusions form when a crystal fractures and heals during growth, and thus are similar to primary inclusions in temperature and salinity (Roedder, 1967a, 1976). For distinguishing criteria, see Roedder (1967a, 1972, 1976).

Fluid inclusions add a time dimension to temperature and salinity studies. The past thermal and chemical characteristics of a geothermal system are recorded by the fluid inclusions which form at that time, and may be compared to present conditions to see how the system has changed.

Data were collected using a combined heating/cooling stage (Freckman, 1978). Thermocouples were used to measure the temperatures, with an accuracy of $\pm 2^\circ\text{C}$ for the range of heating temperatures and $\pm 0.25^\circ\text{C}$ for the range of freezing temperatures in this study.

Figure 14.

Temperature logs and fluid inclusion homogenization temperatures for Mesa 31-1. Note the wide range of temperatures at a single depth between the different logging tools. Fluid inclusion temperatures bracket the general range of measured temperatures, indicating that temperatures in the past were about the same as the present.

Key to different logging runs:

- U.S.B.R., 1974
- Urban et al., 1978
- Dresser Atlas, 1978
- × GO Wireline, 1978
- △ NL McCullough, 1978
(Schlumberger tool failed)

Fluid inclusions: The bracket shows the range, average and number of T_H measurements. If no number is present, the value is for a single inclusion.

TABLE V
A COMPARISON OF TEMPERATURES RECORDED BY DIFFERENT
LOGGING TOOLS AT 152-M (500-FT) INTERVALS

Depth m (ft)	USBR ^a (°C)	Urban et al. ^b (°C)	N.L. McCullough ^c (°C)	Dresser Atlas ^c (°C)	GO Wireline ^c (°C)
305 (1000)	77	69	64	61	78
457 (1500)	92	86	83	80	100
610 (2000)	107	104	104	101	129
762 (2500)	120	118	120	118	134
914 (3000)	126	127	130	129	143
1067 (3500)	132	133	136	134	153
1219 (4000)	138	140	144	141	156
1372 (4500)	144	146	151	146	162
1524 (5000)	150	152	157	153	169
1676 (5500)	155	159	164	161	177
1829 (6000)	157	163	167	165	180

^a USBR, 1974

^b Urban et al., 1978

^c Mathews et al., 1979

The only authigenic mineral from well Mesa 31-1 found to contain fluid inclusions was vein calcite picked from the cuttings. Most of the calcite samples were not suitable for measurement because they lacked suitable inclusions or were turbid. The majority of the inclusions measured were secondary.

A. Fluid Inclusion Heating Results

All heating measurements were performed twice to verify reproducibility and check for leakage. Successive runs were within 2°C of the original. The two runs were averaged to obtain a single T_H value.

Figure 14 shows homogenization temperatures (corrected to hydrostatic pressure) of primary and secondary inclusions from six different depths in Mesa 6-1. The range of T_H variability at a single depth is as much as 40°C. The T_H range generally brackets the measured borehole temperatures.

The variability of T_H values at a single depth may be real, although some variation may be caused by secondary processes such as leakage or necking-down of

large inclusions to form several smaller ones. This occurs frequently. It causes some of the resulting inclusions to have increased T_H values and others to have decreased T_H values.

Overall, the thermal regime appears to be consistent through time. The primary inclusions homogenized in the same temperature range as the secondary inclusions, although they formed at different times. The same approximate temperatures have persisted to the present time.

In a previous study of East Mesa, Hoagland and Elders (1977) performed heating experiments on calcite and quartz from Mesa 6-2. Their uncorrected data indicate that past reservoir temperatures were at least 15°C greater than the measured borehole temperatures. If a pressure correction is made, the past temperatures become 25 to 30°C higher than the present temperatures. As we have seen, however, fluid inclusions suggest that this has not occurred in the rocks penetrated by Mesa 31-1.

In evaluating these results it should be remembered that most of the data came from vein material, and the thermal history of vein material may be quite complex. Fractures open and become sealed by deposition of authigenic minerals. The permeability of a given fracture system varies with time. Even if hydrologic conditions are consistent on a large scale, many small-scale irregularities may persist. In the case of hot fluids passing through rocks, the fluid may be hotter or colder than the surrounding rock and cause a temperature gradient to develop between the fluid and the rock. As calcite is precipitated in the vein, fluid of different temperatures would be trapped in the inclusions (Freckman, 1978).

B. Fluid Inclusion Freezing Results

Freezing data on fluid inclusions are more difficult to obtain than heating data. The host crystal must be very clear in order to see the ice. The inclusions must be larger because small inclusions will persist in a metastable supercooled state even at temperatures below -78°C (Roedder, 1962).

However, the major problem in this freezing study is the persistence of ice in a metastable superheated state. In most inclusions the ice would not melt until temperatures of +3.4 to 5°C had been attained. Then the remaining ice would suddenly melt and the vapor bubble reappear. The phenomenon of superheated ice in fluid inclusions has been described by Roedder (1967b). It occurs in inclusions that form at relatively low temperatures and therefore contain only a small vapor bubble. The liquid expands upon freezing and eliminates the vapor bubble. When the ice melts a vapor bubble does not immediately nucleate. The liquid is in a "stretched" condition under high negative pressure (Roedder, 1967b). Obviously, inclusions which contain superheated ice are useless for salinity measurements.

Only nine inclusions proved suitable for freezing experiments. All were below 1356 m (4448 ft). Their depths are marked on Fig. 15. The inclusions showed no depression of the freezing point. This means that the fluids trapped in these inclusions are virtually fresh.

The fluids presently produced from the well are also quite fresh, 0.23 to 0.29 wt% total dissolved solids (Howard et al., 1978). The NaCl equivalent salinity would depress the freezing point only by about -0.15°C. Thus, within the limits of analytical error, the fluids trapped in the fluid inclusions are identical to the fluids presented produced from the well.

C. Fluid Salinities from Log Interpretation

Wireline logs may be used to calculate or qualitatively describe fluid salinities. Mathews et al. (1979) used the conductivity curve of the Dual Induction-

Figure 15.
Salinity profile, Mesa 31-1. The line is the smoothed average of the log-derived SP salinity and resistivity salinity. The black dots are analyzed waters from drill stem tests (after Littleton and Burnett, 1977). The open circles indicate depths of samples used for salinity determinations in fluid inclusion freezing studies.

Laterolog (Schlumberger trademark log run in 1974) to infer five zones of different water salinity in Mesa 31-1. The boundaries of these zones are:

- 304 m (1000 ft) to 580 m (1900 ft)
- 580 m (1900 ft) to 915 m (3000 ft)
- 915 m (3000 ft) to 1190 m (3900 ft)
- 1190 m (3900 ft) to 1340 m (4400 ft)
- 1340 m (4400 ft) to 1890 m (6200 ft)

TRW Systems Group (1976) inferred a salinity reversal in Mesa 31-1 at 1158 m (3800 ft). Salinity was interpreted to increase with depth from the surface to 1158 m and decrease below that depth.

Littleton and Burnett (1977) computed salinity profiles from SP and Dual Induction logs for several East Mesa wells. Their computations also show a salinity reversal, although it occurs near 915 m (3000 ft). Fig. 15 is their smoothed profile of the average of the SP-derived salinity and the resistivity-derived salinity for Mesa 31-1. Two control points exist where analysis of fluids from drill stem tests are plotted. These are within a few hundred ppm of the calculated salinities, actually plotting between the SP and resistivity-derived salinity curves. Where data exist, there is good agreement between measured and log-derived salinities and fluid inclusion salinities.

Below about 1310 m (4300 ft) the calculated salinities are rather constant at about 2000 to 3000 ppm TDS. Although salinities may be slightly

higher at shallower depths, the fluids are still very fresh, less than 1 wt% TDS. It would probably be impossible to use fluid inclusions to distinguish the more saline fluids of the shallow aquifers from the deeper, less saline fluids, because the error limits of measurement are wider than the salinity difference.

All of the fluid inclusions that were frozen for salinity determination are from the deep part of the well. As stated previously, the fluids in the inclusions, which are fluids trapped at some time in the past, cannot be distinguished from the fluids now present in the well. Hoagland (1976) and Hoagland and Elders (1977), in their studies of Mesa 6-1 and 6-2, suggest that a more saline, hotter fluid caused extensive quartz mineralization found in those wells. Fluid inclusions and previously discussed mineralogical evidence give no indication that a hotter or more saline fluid has ever been present in Mesa 31-1.

XII. STABLE ISOTOPE STUDIES

Stable isotope studies have proven extremely valuable in the study of active geothermal systems. They provide information about the origin of the geothermal waters, the amount of water which has passed through the system, the temperatures at which minerals and/or fluids equilibrated within the system, or conversely whether or not equilibrium was attained between various phases (Taylor, 1967).

It is not necessary to know the absolute isotopic ratios for each substance, but only the difference in the ratios between a substance and a standard. This difference is expressed by the δ notation. δ is the difference in parts per thousand (per mil or ‰) of the isotope ratio of a sample from that of a standard. Thus,

$$\delta x = \left(\frac{R_x}{R_{\text{standard}}} - 1 \right) \times 1000 \quad ,$$

where R is the ratio of the heavy isotope to the light isotope ($^{18}\text{O}/^{16}\text{O}$, $^{13}\text{C}/^{12}\text{C}$, D/H). The standard for oxygen analyses is SMOW (Standard Mean Ocean Water) (Craig, 1961). For carbon the standard is PDB (Pee Dee Belemnite) (Epstein et al., 1951; Craig, 1957). Positive δ values indicate that the sample is enriched in the heavy isotope relative to the standard. This δ value is the quantity actually measured on isotopic ratio mass spectrometers (Friedman and O'Neil, 1977).

If two substances (A and B) exist in isotopic equilibrium, the isotope fractionation factor α is defined as:

$$\alpha_{A-B} = \frac{R_A}{R_B} = \frac{1 + (A/1000)}{1 + (B/1000)} \quad .$$

XIII. ISOTOPIC GEOTHERMOMETRY

Urey (1947) first described how the oxygen isotope fractionation of two substances in equilibrium could be used as a geothermometer. If an isotopic fractionation occurs between two phases which are in equilibrium with each other, the magnitude of the fractionation will be determined only by the temperature and the nature of the phases (Clayton and Epstein, 1961). The temperature dependence of the fractionation factor may be determined theoretically by using thermodynamic and spectroscopic data (e.g. McCrea, 1950), or experimentally by equilibrating two phases over a range of temperature and measuring the isotopic

fractionation (e.g. O'Neil et al., 1969). The numerical value of the fractionation factor, α , will decrease with an increase of temperature.

It is possible to use any two oxygen-bearing cogenetic phases or minerals as geothermometers as long as they have achieved equilibrium, and the temperature dependence of the fractionation factor is known. The calcite-water fractionation is commonly used because it is well established and widely applicable. The variation of $10^3 \ln \alpha$ is linear with $1/T^2$ over the temperature range of 0° to 500° (O'Neil et al., 1969). The most recent equation for the calcite-water fractionation is:

$$10^3 \ln \alpha_{\text{calcite-water}} = 2.78 (10^6 T^{-2}) - 2.89$$

(Friedman and O'Neil, 1977).

The samples used for isotopic analysis were hand-picked from the cuttings. Samples were picked to be homogeneous with respect to rock type, color, grain size and mineralogy. By choosing homogeneous samples, the isotopic trends of the separate rock types may be determined. The following rock types were sampled: red-brown shale, gray shale, tan shale, sandstone and vein calcite. Only calcite in each sample was analyzed. Samples were picked at 23- to 46-m (75- to 150-ft) intervals, or where good material was available.

The samples were first washed with water to remove all contaminants and oven-dried. The fragments were ground with a mortar and pestle to pass through a 100 mesh sieve. This powder was dissolved in 100% phosphoric acid (H_3PO_4) at 25°C to evolve carbon dioxide gas (McCrea, 1950). The CO_2 gas was analyzed on a double-focusing, double-collecting isotope ratio mass spectrometer (Coplen, 1973).

A. Previous Work on East Mesa

Coplen (1976) performed a major isotopic study of the East Mesa system. He measured oxygen and carbon isotope ratios of shallow and deep well waters, bulk cuttings and homogeneous rock types picked from cuttings and cores. He determined that the water presently in the system is largely evaporated Colorado River water, perhaps evaporated from ancient lakes. This water is not in equilibrium with any mineral phase at the present temperatures.

XIV. CARBON ISOTOPIC RESULTS

A. Water Composition

The East Mesa geothermal waters do not have a uniform ^{13}C composition. Wells Mesa 6-2 and Mesa 8-1, in the central part of the field, produce fluids with a $\delta^{13}\text{C}$ of about -3.7 ‰. Mesa 31-1, on the northwest periphery of the field, produces water of -2 ‰ (Coplen, 1976, p. 30). Carbon is present in the water principally as HCO_3^- .

Water produced from shallow levels is lighter because of the influence of CO_2 biologically generated in the soil zone. Shallow [less than 229 (750 ft)] groundwater wells in the vicinity of East Mesa have $\delta^{13}\text{C}$ compositions of -10 to -14 ‰. Two slightly deeper wells of 247 m (810 ft) and 429 m (1406 ft) have $\delta^{13}\text{C}$ of -8.9 and -5.6, respectively (Coplen, 1976).

B. Mesa 31-1

Figure 16 is a plot of $\delta^{13}\text{C}$ versus depth for calcite samples. Immediately obvious is the narrow range of $\delta^{13}\text{C}$ values. The extreme variation among all samples is only about 5 ‰.

Figure 16.
 $\delta^{13}\text{C}$ of calcite versus depth
 in Mesa 31-1. The heavy
 line on the right indicates
 zones where abundant organic
 material was found in the
 cuttings.

Key to symbols

- - vein calcite
- - calcite in sandstone
- - calcite in tan shale
- △ - calcite in red-brown shale
- ▲ - calcite in gray shale

C. Shales

All of the shales, with the exception of the tan shales above 610 m (2000 ft), lie within a narrow range of -1.2 to -2.3 ‰. This is only slightly heavier than the -2.2 ‰ reported for total carbonate of recent Colorado River sediment (Clayton et al., 1968). The consistent ^{13}C composition of these shales suggests that they may not have exchanged carbon isotopes at all. This is due to several reasons, including insufficient temperature, low permeability, similar water ^{13}C composition, and a very small ^{13}C supply in the water. Mesa 31-1 fluids contain only about 845 mg/l HCO_3^- (USBR, 1977), in contrast to about 8 wt% calcite in the shales.

Data from the tan shales above 610 m (2000 ft) are unreliable indicators of isotopic exchange because these were bulk samples of finely ground cuttings and may have been contaminated by casing cement or other substances. Evidence for contamination also comes from unusual $\delta^{18}\text{O}$ values for some of these samples (Fig. 19).

D. Sandstones

The $\delta^{13}\text{C}$ of sandstones is highly variable. A curve drawn through the data points would exhibit a trend toward heavier values with depth. However, the general trend is disturbed by local features causing some abrupt changes in $\delta^{13}\text{C}$ values. Kendall (1976) was successful in correlating $\delta^{13}\text{C}$ minima with horizons of high permeability in her study of three Salton Sea wells. There are some correlations of high-permeability zones (Fig. 21) and isotopic minima in Mesa 31-1.

There is another consideration, however. Many samples of cuttings contain abundant fragments of black organic material. Organic carbon is much lighter than the carbon in the geothermal waters. The organic material also greatly increases the carbon reservoir. This organic material is found in many places throughout the well, but is most abundant in or near the zones of $\delta^{13}\text{C}$ minima. These zones are marked on Fig. 16.

Thus the aquifers are probably stratified with different ^{13}C compositions. Zones with abundant organics cause calcite cement with lighter $\delta^{13}\text{C}$ values to be deposited. This difference shows up in Fig. 17 which is a plot of $\delta^{13}\text{C}$ versus CO_2 yield. One group of points of $\delta^{13}\text{C}$ heavier than -2.8‰ represents calcite with a ^{13}C composition near that of detrital calcite. The other groups represent distinctly lighter calcite cements which have received organic carbon depleted in ^{13}C .

Another possibility is that decarbonation reactions are occurring in some intervals; more negative $\delta^{13}\text{C}_{\text{calcite}}$ values would result from the loss of ^{13}C -rich CO_2 . The plot of $\delta^{13}\text{C}$ versus yield (Fig. 17) shows that sandstones with more negative $\delta^{13}\text{C}$ values also have high CO_2 yields. This is the opposite of what would be expected if decarbonation reactions were occurring. Davis and Sanyal (1979), using geophysical logs, suggest that gas (CO_2) may be found in some zones in the East Mesa field, though not in Mesa 31-1. One possible zone is in Mesa 18-28, located 3.2-km east of Mesa 31-1. This possible gas zone, however, is 2225 m (7300 ft) deep, about 335 m (1100 ft) deeper than Mesa 31-1. Decarbonation reactions do not appear to be of major significance in Mesa 31-1.

Figure 17.
 $\delta^{13}\text{C}$ vs CO_2 yield in calcite from sandstones.

E. Vein Calcite

Vein calcite data exhibits a simple trend toward heavier $\delta^{13}\text{C}$ values with increasing depth. A major disturbance of this trend is caused by a single sample from 1770 m (5808 ft) which is 3‰ lighter than neighboring vein calcites. This sample also has an abnormally heavy $\delta^{18}\text{O}$ value. There are two possible explanations for this: (1) this vein formed from an earlier, probably colder, fluid and did not equilibrate with later fluids, or (2) this material has slumped from a much higher interval.

Above 1600 m (5250 ft) the vein calcites have $\delta^{13}\text{C}$ values similar to those of neighboring sandstones. This

is not true of vein calcites below that depth. These deeper vein calcites are distinctively heavier in ^{13}C than sandstones. This suggests: (1) that the fluids which deposited calcite in the deep veins did not originate in the nearby sandstone aquifers, but came from elsewhere in the system; or (2) that the deep veins were deposited at a different time from different fluids than those that deposited the sandstone cement. Either explanation would be additional evidence for a chemically heterogeneous fluid reservoir.

XV. OXYGEN ISOTOPIC RESULTS

A. Water Isotope Chemistry

The East Mesa field produces waters of different ^{18}O compositions. The lightest waters (-14‰ , Mesa 5-1) are produced in the southeast. This is in the direction of recharge from the Colorado River. The waters become enriched in ^{18}O toward the northwest. Mesa 31-1 produced the heaviest water at about -12.4‰ (Coplen, 1976). These numbers represent an average $\delta^{18}\text{O}$ of water for the entire production interval of the well. Drill stem tests of fluids from shallower zones have sampled lighter fluids, but these samples may have been enriched in ^{18}O by boiling during the sampling process (Coplen, 1976).

B. CO_2 Yield and Calcite Content

Figure 18 is a plot of the CO_2 yield of each sample versus depth. CO_2 yield is in $\mu\text{moles CO}_2$ gas evolved per mg sample dissolved. Pure calcite yields 10 $\mu\text{moles CO}_2/\text{mg}$ calcite. Thus multiplying the yield by 10 gives wt% calcite of each sample. The results are similar to those obtained previously by X-ray diffraction.

It is seen that the calcite content of each rock type remains almost constant with increasing depth. Most shales contain 5 to 10% calcite, although a few of the more silty shales may contain an additional 5 to 10%. Sandstones contain 15 to 45% calcite, with the majority containing 20 to 30%. Samples containing greater than 35% calcite are rare. The calcite content of sandstones is a function of permeability, porosity, and degree of cementation.

Only two veins contain less than 78% calcite. These two veins may be contaminated by other minerals or shaly material.

Figure 19 is a plot of $\delta^{18}\text{O}$ vs depth for 118 selected samples. They may be considered in three lithologic groups: (1) shales (red-brown, tan and gray); (2) sandstones; (3) vein calcite. All analyses are of calcite only. Analytical results ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and CO_2 yield) are listed in Table A-I in Appendix A.

C. Shales

There is only a small $\delta^{18}\text{O}$ variation between the different colored shale samples from the same sampling interval. This variation is usually less than 1‰ , and is not consistent with respect to color. That is, red-brown shales are heavier than gray shales in some intervals and lighter in others. Tan shales have about the same $\delta^{18}\text{O}$ as the red-brown or gray shales of the same interval.

It was originally expected that gray shales would be consistently lighter than red-brown shales from the same interval. It has been suggested that gray shale is more highly altered by hydrothermal fluids than red-brown or tan shales (Fournier, 1973, 1976; Hoagland, 1976; and this study). However, this increased hydrothermal alteration (e.g. red-brown shales to gray shales) does not seem to be accompanied by any significant oxygen isotope exchange in calcite in Mesa 31-1.

Figure 18.

CO₂ yield of calcite versus depth in Mesa 31-1. Multiply the yield by 10 to obtain wt % calcite in the sample.

Figure 19.

δ¹⁸O (SMOW) - Calcite (‰) versus depth in Mesa 31-1.

Key to symbols

- - vein calcite
- - calcite in sandstone
- - calcite in tan shales
- △ - calcite in red-brown shales
- ▲ - calcite in gray shales

Shales have exchanged oxygen with the fluids, but only on a very limited basis. Clayton et al. (1968) report a calcite δ¹⁸O of 22 ‰ for unaltered Colorado River delta sediments. Extrapolation of shale δ¹⁸O values to zero depth yields similar values (Coplen, 1976). Oxygen in the shales is largely unexchanged to a depth of about 1524 m (5000 ft); approximately where the 150°C isotherm is encountered. Below this depth, oxygen exchange has been more significant, although shales are only 4 to 5 ‰ lighter at the bottom of the well than at the top.

Some of the tan shales above 610 m (2000 ft) have unusually light δ¹⁸O values. As mentioned previously, these samples may have been contaminated. A δ¹⁸O of about 22 to 24 ‰ is expected for these samples and there is no apparent reason for some shales to be 2 to 3 ‰ lighter, especially since the CO₂ yields are all similar.

D. Sandstones

Calcite cement in sandstones also exhibit a simple trend toward lighter δ¹⁸O values with depth. Sandstones from 1828 m (6000 ft) contain calcite which

is about 5 ‰ lighter than sandstones from 914 m (3000 ft). Over short intervals most $\delta^{18}\text{O}$ values lie within a 4 ‰ range, although some may vary as much as 8 ‰.

E. Vein Calcite

Vein calcite is, with one exception, always the phase most depleted in ^{18}O at any given depth. The one exception is the sample from 1779-1773 m (5808-5818 ft), which has been discussed previously.

Calcite veins are abundant below 1615 m (5300 ft), but much less common at shallower depths. This is probably because the rocks become more indurated with increasing depth, and are consequently more subject to brittle fracture.

Between 671 and 1524 m (2200 and 5000 ft), the vein calcite becomes lighter with increasing depth. However, below 1524 m (5000 ft) the vein calcite have a remarkably uniform ^{18}O composition. Most samples have a $\delta^{18}\text{O}$ of 2 to 3 ‰. Mesa 6-2 exhibits similar behavior. Vein calcites below 610 m (2000 ft) also have uniform $\delta^{18}\text{O}$ values of 2 to 4 ‰ (Coplen, 1976). These results are somewhat puzzling because one would expect the $\delta^{18}\text{O}$ values to continue their shift toward lighter values with increasing temperature.

The problem is compounded by the temperature data. Below 1524 m (5000 ft), Mesa 6-2 is 25 to 30°C hotter than Mesa 31-1, but the $\delta^{18}\text{O}$ values of vein calcite are nearly identical. This 30°C temperature difference would cause Mesa 6-2 to be 2 ‰ lighter than Mesa 31-1. If Mesa 6-2 was 15 to 25°C hotter at some time in the past, as Hoagland and Elders (1977) fluid inclusion data suggest, then the difference should be even greater, closer to 3 ‰. That these differences are not observed suggests that the reservoir fluids have been quite heterogeneous in ^{18}O composition. The uniform ^{18}O composition of many vein calcites may be the result of rapid calcite deposition from fluid that came from a different part of the reservoir and had not equilibrated chemically and/or thermally in its new environment. Thus the evidence indicates that the reservoir fluids were not homogeneous, either vertically or laterally.

F. Factors Affecting ^{18}O Composition

The variations in ^{18}O composition among different samples and different rock types are a function of how much authigenic calcite has been added to the rock and how much isotopic exchange has occurred. Shales appear to have received little authigenic calcite because they have consistently low CO_2 yields (Fig. 18). Low permeability has prevented significant isotope exchange with the formation waters. Sandstones have higher calcite contents than shales, and much of it is authigenic cement. This authigenic calcite was deposited in equilibrium with the fluid, and the result is a mixture of low ^{18}O authigenic calcite and high ^{18}O detrital calcite, which may be partially exchanged. The veins, of course, are low ^{18}O calcite that presumably crystallized in equilibrium with the fluid, making isotope exchange unnecessary. High permeability is important because it allows fluid to enter the rock and deposit the authigenic mineral phase.

This process is illustrated by the low ^{18}O shales at 1723 m (5652 ft) and 1219 m (4000 ft). These are coarse-grained shales which have higher CO_2 yields (indicating higher calcite content) than the fine-grained shales. The additional calcite has reduced the overall $\delta^{18}\text{O}$ value by 2 to 6 ‰ (Fig. 19).

G. Oxygen Isotope Geothermometry

Equilibration temperatures were calculated for vein calcite using the calcite-water fractionation equation of Friedman and O'Neil (1977) and the present water $\delta^{18}\text{O}$ of -12.4 ‰ (Coplen, 1976). The calculations, however, give temperatures which are about 50°C too cold when compared to the measured borehole

and fluid inclusion temperatures. The vein calcite is thus not in equilibrium with the present fluids at the present temperatures.

The vein calcite did not simply form at cooler temperatures as the geothermal system began heating up. The fluid inclusion data show that the veins were deposited at or near the present temperature. The next step is to calculate the $\delta^{18}\text{O}$ of the fluids which would be in equilibrium with the vein calcite at present temperatures. These calculations show the vein calcite to be in equilibrium with water which has a $\delta^{18}\text{O}$ of -7.2 to -11 ‰. This is 1.5 to 5 ‰ heavier than the fluids now produced. This is identical with the results obtained by Coplen (1976) for Mesa 6-1 and 6-2.

This suggests that the vein calcite precipitated from a fluid which has now been flushed from the system. Coplen (1976) suggests that earlier fluid was derived from ancient lakes (similar to Lake Cahuilla) which had undergone evaporation. The source for this water is the Colorado River (Loeltz et al., 1975) at about -14 ‰. Evaporation of this water from a lake would enrich the remaining water in ^{18}O and D, and this more saline water would infiltrate the sediments. When the lakes disappeared, Colorado River water would begin to flush this fluid from the system with fluid movement from southeast to northwest. Such a model is also supported by Hoagland and Elders (1977), who suggest that a hotter, more saline fluid has been flushed from the system because the fluid presently produced from Mesa 6-2 is not in equilibrium with the earlier potassium-silicate hydrothermal mineral assemblage.

Generally this model seems to work well. It fits the overall isotope chemistry pattern found in the wells in that waters enriched in ^{18}O are found to the northwest, farthest from the recharge point. The fluids become isotopically more like the Colorado River water towards the southeast (Coplen, 1976). One must assume that we simply have not sampled any vein calcite that may have crystallized from the fluid now present in the system.

The flushing may have decreased temperatures slightly in the central part of the field, as suggested by Hoagland and Elders (1977) fluid inclusion data, but it does not appear to have done this in the vicinity of Mesa 31-1. This flushing action may even have been responsible for the introduction of hot fluids to this part of the system.

A problem arises in that evaporation increases the salinity of the remaining water. This saline water should be found in the fluid inclusions of minerals which precipitated from it. It has not been found in this study of Mesa 31-1. Salinity data from fluid inclusion freezing studies and calculations from geophysical logs indicate very fresh fluids, less than 1 wt% NaCl. Salinity measurements on fluid inclusions from Mesa 6-1 and 6-2 would be desirable to better describe the past salinity variations in the entire field.

XVI. DISCUSSION AND CONCLUSIONS

A. Petrophysics and Petrology

It is now possible to combine information gathered from the cuttings with information derived from geophysical logs to more fully describe the rocks penetrated by Mesa 31-1. Figure 20 contains simplified summaries of the density and sonic logs for the sand intervals only. Sandstone densities and interval transit times (Δt , the reciprocal of the sonic velocity) were averaged over 15 m (50 ft) intervals. Figure 21 contains graphs of log-derived porosity (ϕ) and horizontal permeability (K) of sandstones (TRW Systems Group, 1976). Some caution should be used in interpreting these graphs because the ϕ and K values are calculated,

Figure 20.

Sandstone density and interval transit time (reciprocal of sonic velocity), averaged over 15-m (50-ft) intervals, from Schlumberger formation density and sonic logs.

Figure 21.

Porosity and permeability profiles of Mesa 31-1, calculated from geophysical logs (TRW Systems Group, 1976).

not measured, and tool response may be greatly affected by sloughing, which caused enlarged borehole diameters. Nevertheless, the qualitative relationships are thought to be correct.

The sandstone bulk density graph (Fig. 20) indicates only a small increase in density from 335 m (1100 ft) to the bottom of the well. Sandstones are locally more or less dense. Density is not always uniform within a single lithological unit. A discrete sand horizon may exhibit gradational or abrupt changes of density from one part of the bed to another. Similar behavior is seen on the sonic log. These variations in density depend mainly on the extent of cementation.

The sonic log (Fig. 20) indicates that below 840 m (2750 ft) the Δt values generally decrease with depth, but the Δt values above 840 m (2750 ft) are consistently large. The porosities and permeabilities of these shallow rocks are also consistently high (Fig. 21). A regular reduction of porosity and permeability only begins to occur below 914 m (3000 ft).

The rocks above about 840 m (2750 ft) are the cap rocks of the geothermal system. The sand intervals of the cap rocks are uncemented. They have high porosity and permeability, low density and large Δt values (Figs. 20 and 21). The waters in these cap rocks are enriched in ^{18}O and depleted in ^{13}C relative to the waters in the reservoir rocks (Coplen, 1976). The shales, however, are the dominant rock type above 840 m (2750 ft) and they form an impermeable cap over the reservoir. These shales are mostly tan and red-brown. Near the base of the cap gray shales begin to appear and tan shales begin to disappear from the cuttings (Fig. 6). Montmorillonite is the dominant clay mineral in the cap shales.

The lower boundary of the cap is also marked by the inflection point on the temperature profiles (Fig. 14). Heating occurs by conduction in the cap and by mass fluid transfer in the reservoir below.

The shales below 840 m (2750 ft) generally lack montmorillonite. Kaolinite is present only at temperatures below 150°C. At higher temperatures kaolinite is destroyed and authigenic chlorite is present. Illite abundance is uniform except for a slight increase in the deepest red-brown shales.

Sandstones in the reservoir exhibit regular changes in petrophysical properties. Density increases slightly, porosity and permeability decrease. In Mesa 31-1 some sands have experienced a very limited episode of silica/pyrite cementation, usually followed by calcite and/or barite cementation. The low densities, large Δt values and the paucity of well-cemented sandstones in the cuttings suggest that most of the sands are uncemented and unconsolidated. This suggestion is supported by the poor core recovery record for sandstone intervals (Swanberg, 1976), by the cement bond logs which show good bonding between the slotted intervals and the formation because the rocks have collapsed around the casing (Mathews et al., 1979), and by well tests, which show Mesa 31-1 to have the highest transmissivity of any of the USBR wells (transmissivity is the product of permeability of producing formation times the thickness of the producing interval (Howard et al., 1978; TRW Systems Group, 1976)).

Thus, in contrast to the intensely silicified rocks in the center of the geothermal field (Hoagland, 1976), the rocks penetrated by Mesa 31-1 are remarkably unaltered by hydrothermal processes. The primary reason for this is probably that lower temperatures in the outer part of the geothermal system have caused alteration processes to occur more slowly or not to occur at all. Other reasons for the lack of alteration in Mesa 31-1 may include:

- (1) Locally different hydrologic and chemical environments caused by such things as dilution, mixing of different fluids, stratification of fluids or different fluid flow regimes; and
- (2) Hot fluids may have entered this part of the reservoir at a later time than the middle part of the reservoir.

B. Temperature

It may be difficult to obtain reliable temperature data in boreholes as illustrated by the >20°C range of variation among different logging tools. Bottom-hole temperatures in the 165 to 170°C range appear to be most reasonable for this well. Fluid inclusion filling temperatures are in general agreement with the measured borehole temperatures. This suggests that the same approximate temperature has persisted since the rocks were initially heated, in contrast to Mesa 6-2 where fluid inclusions give evidence of previous hotter conditions.

The mineralogical changes observed in the cuttings are also in accord with the present temperature regime. The distribution of various mineral phases is shown in Figure 22. Hydrothermally induced changes in mineralogy generally occur first in the sandstones because of the increased fluid flow (Elders et al., 1978). In Mesa 31-1 authigenic chlorite forms in the 1650 to 1740 m (5400 to 5700 ft) range (Fig. 5). This corresponds to a borehole temperatures of about 160°C. These depths and temperatures also mark the disappearance of kaolinite. This is in good agreement with data from Cerro Prieto (Elders et al., 1978).

Montmorillonite was not found in any of the sandstone samples. It is present in the shales at temperatures below 130°. There is a slight increase in illite abundance in the red-brown shales at temperatures above 160°C.

Dolomite disappears from the sandstones in the 1280 to 1400 m (4200 to 4600 ft) interval. This corresponds to a temperature of about 145 to 150°C (Fig. 19). However, this may not be a simple temperature control. In Mesa 6-2 dolomite

Figure 22.
Generalized diagram of distribution of mineral phases. Dashed lines indicate minor or sporadic occurrence.

occurs down to 1494 m (4900 ft) and sporadically down to 2018 m (6620 ft). Mesa 6-2 contains dolomite to the bottom of the well (about 1830 m or 6000 ft) (Fournier, 1973, 1976). This is equivalent to temperatures of 185°C. Muffler and White found dolomite disappeared at temperatures in the 200 to 260°C range in the Salton Sea geothermal field. Even the shales from Mesa 31-1 appear to lose dolomite near the 1830 m (6000 ft) depth (Figs. 8 and 9). Dolomite thus disappears from Mesa 31-1 at lower temperatures than in other East Mesa or Salton Sea wells.

C. Hydrology

The chemistry of the East Mesa fluid has changed through time as evidenced by:

- (1) An early episode of silica/pyrite cementation;
- (2) A later episode of calcite±barite cementation, with some dissolution of pre-existing silicates; and
- (3) The lack of isotopic equilibrium between the present fluids and hydrothermal calcite. The fluids now produced are too depleted in ¹⁸O to have deposited the vein calcite at the required temperatures.

The various cementation and dissolution features are more poorly developed in Mesa 31-1 than in Mesa 6-2.

Previous fluids have now been flushed from the system by recharge from the Colorado River. However, the hydrologic history has undoubtedly been very complex. Diversion of the Colorado River to the northwest would cause large lakes to form (such as ancient Lake Cahuilla) and change the present regional hydraulic gradient. Such changes would greatly affect water chemistry by altering or inhibiting fluid flow, and adding or flushing components from the system (Hoagland and Elders, 1977).

The geothermal fluids are chemically stratified on a large scale. The fluids produced from below 2000 m (6560 ft) in Mesa 6-1 are more saline and chemically distinct from the more shallow waters (USBR, 1974). Only one isotopic analysis of these deep waters is available, and it indicates that they are enriched in ^{18}O by 3 to 5 ‰ relative to the more shallow waters. Small chemical differences may also exist between fluids from neighboring aquifers. Sandstones have frequently exhibited quite different alteration and cementation characteristics over short intervals in Mesa 31-1 and 38-30, such as the dramatic change from intensely silicified sandstones to calcite cemented sandstones at the bottom of Mesa 38-30.

The sporadic and sparse occurrence of barite (BaSO_4) is additional evidence of small-scale variations in fluid chemistry. The brine produced from Mesa 31-1 is considered to be undersaturated with respect to BaSO_4 (Howard et al., 1978), but local saturation and precipitation of barite has occurred.

Geophysical logs indicate chemically stratified waters with different salinities (Mathews et al., 1979; Littleton and Burnet, 1977), with a salinity reversal occurring near the middle part of the well (Fig. 15). Fluid inclusion evidence indicates that vein calcite in Mesa 31-1 precipitated from water with approximately the same salinity as the present fluids, but with different isotopic composition. The present fluids in the East Mesa field are not isotopically homogeneous but become more depleted in ^{18}O and ^{13}C toward the northwest (Coplen, 1976).

For these aquifers to remain horizontally stratified vertical convection must be minimized. There are many shale beds in the rocks which would act as impermeable barriers to vertical flow. Fracturing and faulting may provide vertical channels through impermeable layers. Fracturing in Mesa 31-1, as evidenced by calcite veins, has been rare above 1516 m (5300 ft). This suggests that vertical fluid flow and fluid mixing is largely restricted to below 1615 m (5300 ft) and that lateral flow is more important above that depth. This may also explain why the vein calcites below 1615 m (5300 ft) have such similar ^{18}O compositions.

Urban et al. (1978) used the shape of the temperature profile to describe the thermal-hydrologic regime of Mesa 31-1. Their models suggest the temperature profile shape may be better explained by a hot zone of horizontal water movement near the base of the cap (about 840 m) than by vertical convection.

It seems likely that convection occurs at some locations in the system, such as along permeable fault zones. These fluids then become isolated because they are confined to separate aquifers as they flow laterally through the system. Waters confined to separate aquifers become chemically heterogeneous because they mix with different fluids and experience different water/rock reactions, such as precipitation of cement or dissolution of existing minerals. The end result is a chemically stratified reservoir.

ACKNOWLEDGMENTS

This report is an adaptation of a thesis submitted to the University of California, Riverside, in partial satisfaction of the requirements for the degree of Master of Science, under the supervision of W. A. Elders (Miller, 1980).

It is the result of contributions from a great many people. We wish to thank Mark Mathews of LASL for initiating this research, and Peter M. Sadler and John L. "Bill" Smith for their review of the thesis. We are also indebted to the people who provided technical support for this project. Rich Coffman prepared the thin-sections, Carolyn Randall performed the isotopic analyses, John Freckman assisted with fluid inclusions, Linda Jankov drafted many of the figures, and

Lori Lovell typed the manuscript. We owe a special depth of gratitude to Pete Collier for X-ray analyses and other support during this project. We also benefited from scientific discussions with Jim Hoagland, Steve Sanford, Dan Seamount, Jeff Sternfeld, and Al Williams.

Samples for well Mesa 38-30 were provided by Republic Geothermal, Inc. This research was supported by LASL sub-contract number N294953H-1 (Wilfred A. Elders, principal investigator), and is a contribution to the U.S. DOE-Division of Geothermal Energy Geothermal Log Interpretation Program.

This is report 80/9 for the Institute of Geophysics and Planetary Physics, University of California, Riverside.

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TABLE A-I
XRD ANALYSIS OF BULK SANDSTONES

DEPT	QUAR	PLAG	ALKA	MICA	ILLI	MXED	1MIX	2MIX	MONT	KAOL	CHLO	CALC	DOLO	AKER	PRYI	HEMA	EPID	BARI
1538.0	43.7	4.2	6.6	-	2.5	-	-	-	-	-	-	42.9	-	-	-	-	-	-
2195.0	41.2	4.6	6.2	-	2.2	-	-	-	-	-	-	45.8	-	-	-	-	-	-
2848.0	50.0	3.6	6.0	-	2.6	-	-	-	-	2.4	-	31.6	2.0	-	-	-	-	1.9
2974.0	56.8	3.4	6.8	-	-	-	-	-	-	-	-	27.3	-	-	-	-	-	5.6
3099.0	46.7	4.6	6.6	-	1.2	-	-	-	-	0.9	-	40.0	-	-	-	-	-	-
3195.0	45.3	2.9	6.2	-	1.5	-	-	-	-	0.8	-	43.4	-	-	-	-	-	-
3284.0	52.8	6.1	7.9	-	2.6	-	-	-	-	2.4	-	24.5	3.7	-	-	-	-	-
3319.0	61.6	6.1	8.8	-	2.8	-	-	-	-	6.8	-	7.3	6.6	-	-	-	-	-
3381.0	55.2	7.1	9.4	-	2.2	-	-	-	-	4.8	-	11.1	5.5	4.7	-	-	-	-
3550.0	50.3	3.5	6.6	-	2.6	-	-	-	-	2.4	-	29.7	2.7	2.1	-	-	-	-
3621.0	56.3	3.3	5.9	-	-	-	-	-	-	0.9	-	31.6	2.0	-	-	-	-	-
3738.0	55.7	5.0	8.2	-	-	-	-	-	-	1.5	-	29.7	-	-	-	-	-	-
3857.0	53.2	5.3	9.4	-	1.6	-	-	-	-	2.3	-	22.8	3.8	-	1.6	-	-	-
3919.0	46.5	3.9	6.4	-	1.2	-	-	-	-	1.2	-	37.4	2.6	-	0.9	-	-	-
4001.0	49.9	6.4	7.7	-	1.9	-	-	-	-	1.9	-	29.2	3.0	-	-	-	-	-
4075.0	43.3	3.5	4.9	-	-	-	-	-	-	3.3	-	6.7	-	-	-	-	-	38.4
4159.0	50.0	3.0	4.1	-	-	-	-	-	-	3.2	-	21.9	1.3	-	-	-	-	16.6
4230.0	59.7	3.8	8.5	-	-	-	-	-	-	1.9	-	26.1	-	-	-	-	-	-
4292.0	49.0	4.5	7.3	-	1.6	-	-	-	-	1.2	-	36.4	-	-	-	-	-	-
4448.0	69.6	7.6	12.0	-	2.2	-	-	-	-	-	-	8.6	-	-	-	-	-	-
4525.0	56.7	4.6	7.2	-	-	-	-	-	-	0.9	-	30.5	-	-	-	-	-	-
4600.0	55.8	5.9	7.1	-	2.5	-	-	-	-	2.0	-	23.7	3.0	-	-	-	-	-
4828.0	58.5	6.4	8.1	-	2.9	-	-	-	-	1.2	-	22.8	-	-	-	-	-	-
4895.0	55.5	4.2	8.1	-	1.2	-	-	-	-	-	-	31.0	-	-	-	-	-	-
5249.0	59.3	4.3	6.0	-	-	-	-	-	-	-	-	30.4	-	-	-	-	-	-
5352.0	64.5	5.3	8.1	-	2.5	-	-	-	-	1.3	-	18.4	-	-	-	-	-	-
5410.0	65.2	7.4	9.4	-	1.0	-	-	-	-	-	1.7	15.3	-	-	-	-	-	-
5613.0	79.0	7.2	10.6	-	2.9	-	-	-	-	-	-	0.3	-	-	-	-	-	-
5777.0	55.8	5.1	7.8	-	2.8	-	-	-	-	-	2.0	26.7	-	-	-	-	-	-
5943.0	50.9	5.6	7.6	-	2.5	-	-	-	-	-	-	33.4	-	-	-	-	-	-
5985.0	59.0	6.2	8.1	-	3.3	-	-	-	-	-	2.8	20.6	-	-	-	-	-	-
6047.0	56.2	7.0	7.4	-	3.5	-	-	-	-	-	2.8	23.1	-	-	-	-	-	-
6098.0	59.5	6.8	7.6	-	2.4	-	-	-	-	-	0.9	22.9	-	-	-	-	-	-
6141.0	60.8	7.6	10.4	-	2.0	-	-	-	-	-	2.4	16.9	-	-	-	-	-	-
6209.0	56.5	7.0	9.8	-	1.2	-	-	-	-	-	1.5	24.0	-	-	-	-	-	-

APPENDIX A

See Table I in text for key to mineral abbreviations, detection limits, and "standard error." DEPT = depth in feet; BARI = barite.

TABLE A-II
XRD ANALYSES OF BULK TAN SHALES

DEPT	QUAR	PLAG	ALKA	MICA	ILLI	MXED	1MIX	2MIX	MONT	KAOL	CHLO	CALC	DOLO	AKER	PRYI	HEMA	EPID	BARI
1286.0	25.3	4.6	5.4	-	3.9	-	1.0	13.2	21.0	4.7	5.8	11.2	3.9	-	-	-	-	-
1414.0	22.0	4.6	5.6	-	6.0	-	2.4	7.7	15.8	6.1	5.6	18.8	3.8	-	-	1.5	-	-
1538.0	30.0	4.6	5.3	-	4.2	-	1.4	9.1	14.9	5.1	6.5	11.7	4.9	-	-	2.3	-	-
1724.0	22.5	4.6	4.8	-	4.1	-	6.1	11.3	18.9	6.4	6.4	9.4	3.5	-	-	1.9	-	-
1944.0	32.7	3.9	5.8	-	6.8	-	2.8	6.0	9.9	6.5	4.6	14.1	5.3	-	-	1.7	-	-
2070.0	28.5	4.1	4.9	-	4.8	-	3.7	9.4	15.5	5.3	5.0	12.6	4.6	-	-	1.6	-	-
2195.0	28.5	4.9	5.6	-	4.6	-	1.4	7.7	15.4	7.5	6.6	11.0	4.4	-	-	2.5	-	-
2291.0	27.4	4.2	5.1	-	3.7	-	4.6	11.0	14.3	5.3	5.5	11.8	4.4	-	-	2.7	-	-
2382.0	24.3	4.3	5.3	-	7.2	-	6.4	10.3	12.3	6.4	5.9	10.6	4.6	-	-	2.4	-	-
2693.0	30.7	6.1	5.7	-	6.4	-	3.7	9.1	9.2	7.3	7.0	8.1	5.2	-	-	1.5	-	-
3099.0	26.1	3.4	5.3	-	8.9	-	8.9	10.3	8.2	7.0	7.2	8.0	4.7	-	-	2.1	-	-

See Table I in text for key to mineral abbreviations, detection limits, and "standard error." DEPT = depth in feet; BARI = barite.

TABLE A-III
XRD ANALYSES OF BULK RED-BROWN SHALES

DEPT	QUAR	PLAG	ALKA	MICA	ILLI	MXED	1MIX	2MIX	MONT	KAOL	CHLO	CALC	DOLO	AKER	PRYI	HEMA	EPID	BARI
1944.0	24.8	3.4	4.0	-	8.0	-	8.1	12.9	14.0	6.2	5.2	8.3	3.6	-	-	1.6	-	-
2382.0	23.6	5.1	5.1	-	7.8	-	7.4	10.2	14.4	5.5	4.4	9.8	4.0	-	-	2.8	-	-
2538.0	27.5	5.4	5.2	-	7.9	-	6.6	11.2	8.1	6.9	5.3	9.1	4.3	-	-	2.5	-	-
2848.0	26.4	4.8	5.1	-	10.3	-	10.9	11.2	6.6	5.3	3.8	8.9	4.4	-	-	2.4	-	-
2974.0	22.3	3.8	4.3	-	9.5	-	12.9	10.2	8.3	5.3	4.8	12.5	4.6	-	-	1.6	-	-
3319.0	23.3	4.3	5.7	-	12.4	-	13.5	11.7	4.2	5.9	4.7	7.0	4.5	-	-	2.9	-	-
3550.0	24.4	3.2	5.2	-	10.0	-	12.9	11.4	2.1	8.0	5.9	7.7	5.0	-	-	4.1	-	-
3621.0	25.9	3.4	5.1	-	12.3	-	13.7	11.5	-	7.5	6.3	6.8	5.6	-	-	2.0	-	-
3770.0	20.4	3.1	4.1	-	13.4	-	17.7	17.4	-	6.0	5.8	6.1	4.1	-	-	1.8	-	-
3919.0	23.0	4.2	4.6	-	12.0	-	13.9	11.7	-	7.1	5.7	10.7	4.4	-	-	2.7	-	-
4001.0	22.1	3.3	4.0	-	11.9	-	14.8	17.3	-	4.7	3.6	11.8	4.0	-	-	2.3	-	-
4075.0	25.1	3.6	3.7	-	13.0	-	16.4	13.3	-	5.2	5.1	6.2	6.0	-	-	2.5	-	-
4159.0	22.2	4.3	4.8	-	10.6	-	13.6	14.5	-	6.6	4.1	11.5	4.7	-	-	3.0	-	-
4230.0	22.2	2.9	3.9	-	14.3	-	16.1	12.6	-	6.1	7.3	7.5	4.5	-	-	2.5	-	-
4292.0	23.2	3.6	4.5	-	10.1	-	16.0	18.8	-	6.1	4.2	5.7	4.1	-	-	3.7	-	-
4380.0	23.5	3.5	4.1	-	13.9	-	16.1	14.5	-	5.8	4.5	6.0	4.7	-	-	3.2	-	-
4448.0	23.9	3.7	4.2	-	13.9	-	18.6	16.0	-	3.9	3.6	6.0	3.8	-	-	2.4	-	-
4525.0	25.6	4.6	5.1	-	11.9	-	17.1	12.8	-	4.1	4.3	6.2	4.2	-	-	3.9	-	-
4600.0	19.9	3.1	3.9	-	13.1	-	19.3	18.8	-	4.7	3.9	6.0	4.0	-	-	3.5	-	-
4895.0	21.7	3.5	4.2	-	13.3	-	19.4	19.8	-	2.2	3.8	6.2	2.4	-	-	3.5	-	-
4988.0	21.5	3.0	4.4	-	13.7	-	20.5	18.6	-	3.7	4.1	4.9	3.1	-	-	2.4	-	-
5164.0	27.1	4.5	6.0	-	11.4	-	12.4	7.2	-	5.2	5.0	9.9	8.7	-	-	2.7	-	-
5489.0	24.6	3.8	4.8	-	13.3	-	19.2	16.5	-	2.1	4.3	8.3	1.1	-	-	1.9	-	-
5714.0	26.0	3.9	4.1	-	17.1	-	18.6	14.0	-	2.2	5.6	6.8	-	-	-	1.7	-	-
5777.0	23.7	4.1	4.8	-	17.9	-	19.7	12.2	-	1.6	5.1	8.9	-	-	-	2.1	-	-
5943.0	27.4	3.6	4.2	-	15.1	-	17.0	14.1	-	1.6	4.0	10.2	-	-	-	2.8	-	-
5985.0	31.8	4.3	6.8	-	15.0	-	18.4	5.8	-	-	6.5	8.9	-	-	-	2.5	-	-
6047.0	25.9	3.7	5.7	-	16.1	-	17.4	10.5	-	-	6.4	11.5	-	-	-	2.9	-	-
6098.0	30.7	4.3	5.9	-	12.9	-	16.5	8.3	-	-	5.7	13.3	-	-	-	2.4	-	-
6141.0	30.1	4.8	6.4	-	13.8	-	17.0	8.9	-	-	6.6	9.2	-	-	-	3.2	-	-
6209.0	26.8	4.0	5.1	-	18.8	-	19.8	10.7	-	-	4.9	7.3	-	-	-	2.5	-	-

See Table I in text for key to mineral abbreviations, detection limits, and "standard error." DEPT = depth in feet; BARI = barite.

TABLE A-IV

XRD ANALYSES OF BULK GRAY SHALES

DEPT	QUAR	PLAG	ALKA	MICA	ILLI	MXED	1MIX	2MIX	MONT	KAOL	CHLO	CALC	DOLO	AKER	PRYI	HEMA	EPID	BARI
2382.0	22.7	4.0	5.3	-	7.5	-	9.3	12.8	11.4	6.9	6.0	9.6	4.5	-	-	-	-	-
2848.0	26.9	4.3	5.2	-	11.7	-	10.8	12.0	-	7.6	7.5	9.2	4.9	-	-	-	-	-
3194.0	26.6	2.3	4.9	-	10.6	-	12.9	11.9	-	6.5	7.4	10.4	6.7	-	-	-	-	-
3284.0	29.5	2.3	5.9	-	9.4	-	11.2	8.8	-	8.4	8.6	9.4	6.6	-	-	-	-	-
3381.0	29.5	2.4	5.7	-	10.9	-	14.5	7.5	-	7.9	7.4	9.1	5.1	-	-	-	-	-
3550.0	28.2	3.2	5.3	-	10.2	-	12.8	9.3	-	8.1	9.9	8.0	4.9	-	-	-	-	-
3621.0	24.7	2.3	5.3	-	10.3	-	15.5	11.8	-	7.9	8.6	8.6	5.1	-	-	-	-	-
3770.0	26.4	3.6	5.1	-	11.7	-	13.3	13.0	-	5.7	8.0	8.3	4.9	-	-	-	-	-
3919.0	29.6	4.9	5.7	-	9.7	-	11.5	10.2	-	7.1	6.9	8.5	5.8	-	-	-	-	-
4075.0	28.4	4.4	5.0	-	9.4	-	11.3	12.0	-	6.5	9.0	8.7	5.2	-	-	-	-	-
4159.0	24.8	3.2	5.1	-	13.5	-	17.1	12.8	-	6.0	5.8	6.7	5.0	-	-	-	-	-
4230.0	27.6	2.4	5.5	-	12.2	-	13.2	11.0	-	6.4	8.3	8.1	5.2	-	-	-	-	-
4292.0	32.0	4.1	6.2	-	10.2	-	10.6	7.9	-	7.3	7.6	8.2	6.0	-	-	-	-	-
4380.0	29.7	3.5	5.3	-	9.1	-	14.3	9.9	-	6.9	6.5	9.2	5.5	-	-	-	-	-
4448.0	28.0	3.2	5.0	-	9.9	-	14.9	10.2	-	6.7	8.4	8.5	5.2	-	-	-	-	-
4525.0	27.1	3.1	5.1	-	12.2	-	13.9	11.5	-	5.7	8.4	7.9	5.2	-	-	-	-	-
4600.0	31.3	3.5	5.7	-	8.2	-	9.2	7.6	-	10.0	11.6	7.7	5.1	-	-	-	-	-
4674.0	27.3	3.4	4.5	-	10.7	-	13.3	10.1	-	5.9	9.4	9.9	5.4	-	-	-	-	-
4828.0	29.9	3.2	4.8	-	8.6	-	10.4	8.3	-	6.9	10.8	11.5	5.5	-	-	-	-	-
4895.0	25.5	3.3	4.3	-	10.5	-	15.9	12.4	-	5.9	8.8	8.8	4.4	-	-	-	-	-
4988.0	26.3	3.4	4.6	-	10.1	-	14.1	11.5	-	6.8	9.6	9.1	4.5	-	-	-	-	-
5164.0	27.2	3.1	4.6	-	12.8	-	12.9	10.8	-	6.0	7.3	10.4	4.8	-	-	-	-	-
5249.0	28.6	3.8	4.8	-	10.2	-	13.0	11.7	-	5.2	9.6	9.2	2.1	-	1.6	-	-	-
5352.0	27.9	3.9	4.6	-	10.9	-	12.4	10.3	-	5.8	9.8	10.9	3.4	-	-	-	-	-
5410.0	25.0	3.2	5.1	-	11.0	-	15.0	13.7	-	5.0	9.2	10.7	2.1	-	-	-	-	-
5489.0	28.6	3.8	6.3	-	10.7	-	13.6	9.6	-	4.9	9.9	9.9	2.7	-	-	-	-	-
5613.0	25.3	3.6	4.5	-	16.8	-	18.1	9.3	-	4.3	8.3	8.4	1.4	-	-	-	-	-
5652.0	39.1	5.7	6.0	-	8.3	-	6.8	4.9	-	3.0	7.8	12.8	5.6	-	-	-	-	-
5714.0	24.4	3.5	3.4	-	13.4	-	16.1	11.7	-	4.4	8.4	11.6	3.0	-	-	-	-	-
5777.0	26.3	3.7	5.0	-	11.7	-	17.2	15.1	-	3.6	7.9	7.8	1.6	-	-	-	-	-
5943.0	28.7	2.6	5.4	-	13.8	-	13.2	9.9	-	3.9	9.2	11.7	1.6	-	-	-	-	-
5985.0	30.6	3.9	6.0	-	13.3	-	12.5	7.3	-	4.0	9.0	11.9	1.3	-	-	-	-	-
6047.0	26.2	3.5	4.4	-	13.6	-	13.8	13.0	-	3.0	8.3	12.3	1.8	-	-	-	-	-
6098.0	27.3	3.7	5.4	-	10.5	-	16.2	12.1	-	2.4	7.9	12.0	-	-	2.5	-	-	-
6141.0	32.4	5.1	7.3	-	10.2	-	9.7	7.3	-	4.0	9.2	12.6	-	-	2.2	-	-	-
6209.0	31.3	4.3	5.5	-	8.7	-	17.4	10.2	-	2.4	8.5	11.9	-	-	-	-	-	-

See Table I in text for key to mineral abbreviations, detection limits, and "standard error." DEPT = depth in feet; BARI = barite.

TABLE A-V
EAST MESA RESULTS WELL 31-1

Interval	*Rock Type	Yield	δ^{13} (PDB)	δ^{18} (SMOW)
Feet		$\mu\text{mol/mg}$	‰	‰
1286 - 1381	bulk tan sh.	0.50	-3.95	22.80
1414 - 1445	bulk tan sh.	0.77	-4.14	22.04
1507 - 1538	bulk tan sh.	0.83	-4.34	20.03
1538 - 1569	bulk tan sh.	0.62	-4.17	20.12
1724 - 1755	bulk tan sh.	0.82	-4.02	18.95
1818 - 1913	bulk tan sh.	0.62	-3.54	21.09
2164 - 2195	ss	3.47	-4.18	8.96
2195 - 2228	v. cal.	8.87	-4.38	7.07
2382 - 2413	r/b sh.	0.77	-1.90	23.32
2848 - 2880	ss	3.51	-3.21	9.40
2848 - 2880	gr. sh.	0.97	-1.44	22.23
2848 - 2880	bulk tan sh.	0.85	-1.50	23.30
2974 - 3005	ss	1.97	-3.72	8.55
3099 - 3131	ss	1.73	-4.30	8.26
3099 - 3131	ss	2.68	-3.45	10.50
3280 - 3319	gr. sh.	0.91	-1.54	21.91
3288 - 3319	ss	2.71	-2.56	9.05
3381 - 3412	gr. sh.	0.87	-1.72	22.01
3381 - 3412	ss	1.98	-2.31	10.85
3550 - 3560	ss	2.50	-4.12	13.70
3550 - 3560	ss	1.86	-2.40	8.46
3560 - 3587	v. cal.	9.27	-2.99	6.19
3621 - 3635	r/b sh.	0.80	-1.91	21.62
3621 - 3635	gr. sh.	0.85	-1.50	22.27
3621 - 3635	ss	2.59	-3.69	8.83
3770 - 3780	r/b sh.	0.85	-1.52	22.66
3770 - 3780	gr. sh.	0.83	-1.67	22.40
3857 - 3867	ss	2.66	-5.86	8.33
3857 - 3867	ss	2.70	-4.02	9.36

(Page 2)

Interval	Rock Type	Yield	δ^{13} (PDB)	δ^{18} (SMOW)
Feet		$\mu\text{mol/mg}$	‰	‰
3919 - 3929	r/b sh.	0.80	-1.69	23.38
3919 - 3929	gr. sh.	0.86	-1.65	22.49
4001 - 4011	ss	2.95	-4.24	7.36
4001 - 4011	r/b sh.	1.93	-5.16	16.14
4075 - 4085	ss	1.33	-1.78	9.73
4075 - 4085	ss	2.28	-2.46	6.88
4159 - 4168	ss	2.28	-3.57	7.77
4159 - 4168	r/b sh.	1.10	-2.52	21.05
4230 - 4240	ss	3.08	-2.27	7.68
4230 - 4240	ss	4.30	-5.66	12.41
4281 - 4292	r/b sh.	0.79	-1.48	21.58
4281 - 4292	gr. sh.	0.80	-1.46	21.95
4292 - 4312	ss	2.01	-2.20	6.67
4448 - 4457	r/b sh.	0.70	-1.91	21.78
4448 - 4457	gr. sh.	0.79	-1.28	22.35
4525 - 4535	ss	2.11	-2.39	4.24
4600 - 4612	ss	2.54	-2.80	5.21
4600 - 4612	r/b sh.	0.99	-1.43	21.54
4600 - 4612	gr. sh.	0.84	-1.51	22.08
4600 - 4622	v. cal.	7.75	-2.49	3.73
4674 - 4684	gr. sh.	1.04	-1.73	19.92
4828 - 4839	gr. sh.	0.88	-1.56	21.65
4828 - 4839	ss	1.84	-2.21	5.29
4849 - 4895	v. cal.	5.39	-3.06	1.35
4895 - 4905	gr. sh.	0.83	-1.26	21.84
4895 - 4905	ss	2.76	-2.53	7.33
4988 - 5000	gr. sh.	0.84	-1.37	22.02
5164 - 5177	gr. sh.	0.76	-1.62	22.31
5164 - 5177	ss	1.56	-1.92	10.50
5164 - 5177	r/b sh.	0.62	-1.46	22.72
5249 - 5259	r/b sh.	0.79	-1.85	19.63

(Page 3)

Interval	Rock Type	Yield	δ^{13} (PDB)	δ^{18} (SMOW)
Feet		$\mu\text{mol/mg}$	‰	‰
5249 - 5259	gr. sh.	0.79	-1.44	20.29
5290 - 5310	v. cal.	6.18	-1.63	2.25
5352 - 5363	ss	1.88	-2.52	3.47
5352 - 5363	gr. sh.	0.78	-1.50	20.57
5395 - 5410	v. cal.	8.98	-1.74	2.39
5410 - 5426	gr. sh.	0.83	-1.56	19.92
5410 - 5426	ss	1.53	-3.79	7.73
5489 - 5499	gr. sh.	0.87	-1.65	20.31
5489 - 5499	r/b sh.	0.77	-1.76	20.73
5499 - 5509	v. cal.	8.73	-1.46	2.43
5592 - 5602	v. cal.	8.69	-1.31	2.26
5613 - 5623	ss	2.34	-2.17	3.90
5613 - 5623	gr. sh.	0.69	-2.02	19.38
5652 (a)	gr. sh.	1.35	-2.24	14.76
5652 (b)	gr. sh.	1.47	-2.20	13.58
5652 (c)	gr. sh.	1.34	-2.10	15.49
5652 (d)	gr. sh.	1.37	-2.25	15.32
5652 (e)	gr. sh.	1.36	-2.07	15.61
5652 (f)	gr. sh.	1.44	-2.26	14.33
5652	gr. sh.	0.84	-2.14	17.98
5653	gr. sh.	1.00	-2.07	19.26
5714 - 5724	gr. sh.	0.84	-1.57	19.96
5714 - 5724	r/b sh.	0.71	-1.72	19.30
5777 - 5787	v. cal.	8.24	-1.32	2.44
5777 - 5787	ss	2.46	-2.12	6.17
5777 - 5787	gr. sh. (a)	0.70	-1.77	19.36
5777 - 5787	gr. sh. (b)	0.87	-1.70	20.17
5777 - 5787	gr. sh. (c)	0.88	-1.58	19.44
5777 - 5787	gr. sh. (d)	0.87	-1.50	19.26
5777 - 5787	r/b sh. (a)	0.65	-1.68	20.43
5777 - 5787	r/b sh. (b)	0.60	-2.00	20.00

(Page 4)

Interval	Rock Type	Yield	δ^{13} (PDB)	δ^{18} (SMOW)
Feet		$\mu\text{mol/mg}$	‰	‰
5777 - 5787	r/b sh. (c)	0.70	-1.62	19.95
5777 - 5787	r/b sh. (d)	0.61	-1.56	20.26
5808 - 5818	ss (a)	2.33	-2.43	6.16
5808 - 5818	ss (b)	2.98	-2.39	5.59
5808 - 5818	v. cal.	8.72	-1.61	2.13
5808 - 5818	v. cal.	8.01	-4.50	8.72
5943 - 5954	r/b sh.	0.61	-1.67	20.68
5943 - 5954	v. cal.	9.15	-1.30	2.16
5943 - 5954	gr. sh.	0.83	-1.67	19.54
5954 - 5964	ss	2.35	-2.46	6.34
5954 - 5964	v. cal.	8.72	-1.23	2.15
5985 - 5995	ss	2.04	-2.43	5.46
6047 - 6057	v. cal.	7.80	-1.48	2.01
6047 - 6057	gr. sh.	0.89	-1.63	18.47
6047 - 6057	ss	2.48	-5.74	5.31
6047 - 6057	r/b sh.	0.71	-1.92	18.58
6098 - 6110	ss	2.71	-2.65	7.73
6098 - 6110	r/b sh.	0.72	-1.89	18.53
6098 - 6110	gr. sh.	0.98	-2.00	17.51
6098 - 6110	ss	2.64	-2.32	5.14
6098 - 6110	v. cal	9.07	-0.97	2.28
6141 - 6151	ss	1.79	-3.59	9.60
6141 - 6151	r/br sh.	0.76	-2.06	18.14
6141 - 6151	gr. sh.	0.75	-1.53	19.30
6174 - 6183	v. cal.	8.39	-1.71	2.70
6174 - 6183	r/b sh.	0.68	-1.82	19.58
6174 - 6183	gr. sh.	0.92	-1.65	18.27
6124 - 6183	v. cal.	8.82	-1.53	2.47
6183 - 6192	v. cal.	8.05	-1.63	2.62
6192 - 6201	v. cal.	8.89	-2.20	3.60
6209 - 6226	v. cal.	8.74	-1.59	2.02

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Interval	Rock Type	Yield	δ^{13} (PDB)	δ^{18} (SMOW)
Feet		$\mu\text{mol/mg}$	‰	‰
6209 - 6226	r/b sh.	0.62	-2.10	18.55
6209 - 6226	gr. sh.	0.64	-1.83	18.82
6209 - 6226	ss	1.80	-2.78	6.90

*Legend for Rock Type abbreviations:

bulk tan sh. = bulk tan shale
ss = sandstone
v. cal. = vein calcite
r/b sh. = red-brown shale
gr. sh. = gray shale

East Mesa

INTRODUCTION TO WELL C/T-1

The U.S. Department of Energy has established a Geothermal Test Facility (GTF) at East Mesa in the Imperial Valley of California. The purpose of the Facility is to promote the more rapid commercialization of geothermal energy potential. The site was originally established by the Department of the Interior, Bureau of Reclamation for desalination research. The site has since been turned over to the Department of Energy and the produced fluids are now made available to experimenters.

On the site there are three producing

wells, an injection well, and an observation well which is not used for production or injection. This observation well, Mesa 31-1 has been set aside in agreement with Los Alamos Scientific Laboratory for the calibration and testing of geothermal well logging tools.

Data base logging was performed by four companies, Dresser-Atlas, GO Wireline, NL McCullough, and Schlumberger. Tool testing has been performed by GO Wireline, GeoThermal Services, and United Wireline.

East Mesa, California

LOCATION BACKGROUND

The East Mesa KGRA is located about 20 miles east of El Centro, about 140 miles east of San Diego, and 10 miles north of the Mexican border in the Imperial Valley of California.

The Imperial Valley is supported mainly by agriculture with some heavy industry. The weather of the area is typified by hot summers and pleasant winters. Average annual rainfall is about 3.5 inches. About half the rain usually precipitates in high intensity summer showers and about half in gentle winter rains.

A four lane freeway, Interstate 8, traverses Imperial Valley from San Diego

to Yuma, Arizona. It runs adjacent to El Centro and about 2 miles from the GTF. To the north, State Highway 111 connects El Centro to Indio and Palm Springs. The Van Der Linden Road exit from Interstate 8 provides access to Even Hewes Road which parallels the freeway on the north side. East of Van Der Linden Road on Even Hewes Road is the Test Facility sign located by the access road to the Facility; the Facility is located 1.7 miles north of this sign. All roads to the Facility are paved. The GTF is also accessible by regularly scheduled airline flights to the airport in El Centro.

Temperature and pressure log comparison
LASL Well C/T-1 (East Mesa 31-1)

WELL LOG CALIBRATION/TEST APPLICATION

Application for: ☐ C/T-1, East Mesa, California
☐ C/T-2, Roosevelt Hot Springs, Utah

Organization:

Application No.: _____ (Office Use)

Address:

Contact:

Phone:

Test/Calibration Name:

Test/Calibration Objective (Please elaborate):

- ☐ Test, or
☐ Calibration?

Parameter(s) Measured, Technology Under Study:

Equipment Type (type trucks, conductor, sonde(s), etc.):

Support or Resources Required:

Schedule (Best Estimate):

1. Arrival Date: _____
2. Test/Calibration Start Date: _____
3. Test/Calibration Completion Date: _____
4. Equipment Removal Date: _____

Equipment Size/Weight: (Note: Must include drawing or exterior dimensions of tool)

Special Hazards:

Signed: _____ Date: _____

Principal Investigator