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THERMAL WATERS OF UTAH

Topical Report

by

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Bonneville Salt Flats

INTRODUCTION

The Bonneville Salt Flats are in western Tooele County adjacent to the Nevada border. Wendover is the only town, and the principal industry is the extraction of salt from brines that underlie the salt flats. The wells drilled to explore for and recover the brines provide the principal information about subsurface temperature.

SUMMARY of OCCURRENCES of HOT and WARM WATER

Many shallow and deep wells drilled to extract brines from the Bonneville Salt Flats have reported warm and hot water. Two deep wells that seem to straddle a northeast-trending fault reached water of 43° at 1200 feet and 88°C at 1636 feet, respectively.

GEOLOGIC and HYDROLOGIC ENVIRONMENT

The Bonneville Salt Flats are in an asymmetric two-step graben whose axis trends about N50°E (Turk, 1973, fig. 3). The northwestern boundary of the graben is the Silver Island Mountains formed largely of Paleozoic sedimentary rocks, but "intruded by five stocks of unknown age, ranging in composition from quartz monzonite to granodiorite (Whelan and Petersen, 1974, p. 75). In addition, there are dikes of various compositions and "seven volcanic flows of rhyolite or andesitic composition in the southern Silver Island Range" (Whelan and Petersen, 1974, p. 77). The southeastern flank of the graben probably is composed of volcanic rock but it has no surface expression and is hidden by the valley fill.

"The basin overlying the volcanic rocks was filled with fluvial and later lacustrine sediments of Plio-Pleistocene age" (Turk, 1973, p. 1). The youngest deposits form the salt crust which underlies about 150 square miles "and ranges in thickness from a feather edge to nearly 5 feet in the center...The sediments underlying the salt crust are saturated with sodium chloride brine" (Turk, 1973, p. 1).

Turk (1973, p. 1) has identified three aquifers: 1) an alluvial fan aquifer off the southeast slope of the Silver Island Mountains that supplies brackish water to 27 wells, 2) "a deep stratified aquifer holding low-grade brine recoverable by deep wells, and 3) a shallow aquifer of lacustrine sediments containing high-grade brine which is harvested for its potassium chloride content."

OCCURRENCE of HOT and WARM WATERS

Turk (1973, p. 5) reports one spring, Blue Lake Spring, (C-4-19)6d, about 15 miles south of Wendover, that yields moderately saline water at a temperature of 29°C (84°F). All other temperature data come from wells, of which only a few records are given here in table 2. Most of the wells, even one, (C-1-19)23cbc, that is 1496 feet deep, have water temperatures that range from 21°C to 35°C, but deep wells 1 and 3 reached water of 43°C at 1200 feet and 88°C at 1636 feet, respectively. These last two wells seem to straddle a southwestward extension of the inner fault on the southeast side of the graben.

Whelan and Petersen have computed reservoir temperatures using the Na-K-Ca method of Fournier and Treusdell, and have determined that the temperature for well #5 is 270°C, for DBW13, 199°C, and for DBW8, 285°C. Because the waters of DBW8 and 13 are so briny, Whelan and Petersen feel that the temperature at well #5, 270°C, may be the best indicator of the reservoir temperature (1974, p. 77).

Whelan and Petersen (1974, p. 78) conclude that "Bonneville Salt Flats just south of Wendover, Utah, possibly contain a geothermal reservoir. All of the theoretical requirements for the system could be present. There is a buried intrusive which could be the heat source; water in the form of brines appears to be present; the faulting of the Wendover Graben could provide the required permeability; and buried volcanics could provide the cap rock."

Although Turk (1973, p. 74) doesn't specifically deny the existence of a thermal reservoir, in his short discussion on the Great Salt Lake Desert he suggests that the warmth of the water may be due to a high geothermal gradient which results from a hindering of heat flow by the thick porous clay beds that underlie the desert. "The implication is that areas of thick clay accumulation, like the Great Salt Lake Desert, may have a low-temperature geothermal potential without a near-surface source of heat..."

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Cove Fort - Sulphurdale

Cove Fort-Sulphurdale, in an area of Quaternary volcanic rocks, is listed as an Administrative KGRA in the section on Known Geothermal Resource Areas, above. No records of water wells in the area were found in the literature, and Mundorff (1970, p. 50) reported an unsuccessful search for "Sulphurdale Hot Springs" that may have been related to the mining of sulfur deposits in the area.

Rush (1977) reported temperature measurements that he made in 13 wells or drill holes in the Cove Fort-Sulphurdale area, and I have computed geothermal gradients from his reported measurements:

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BONNEVILLE SALT FLATS—A POSSIBLE GEOTHERMAL AREA?

by J. A. Whelan[1] *and Carol A. Petersen*[2]

INTRODUCTION

In his paper on the hydrogeology of the Bonneville Salt Flats, Utah, L. J. Turk (1973) noted indications of an abnormally high geothermal gradient in several brackish water wells, several deep brine wells, and two warm springs in the Bonneville Salt Flats area (see table 1 and figures 1 and 2). From Turk's data, Whelan noted both sets of wells are on essentially north-south lines (figure 3) and that water and brine temperatures are consistently higher to the south (figure 4).

In the Utah Geological and Mineral Survey's program of developing and publishing data on potential geothermal areas, it was considered appropriate to expand on the data furnished by Turk.

DESCRIPTIONS OF WARM WELLS AND SPRINGS

Deep Brine Wells

Thirteen deep brine wells (drilled between 1939 and 1951) near the western edge of the Bonneville Salt Flats playa range in depth from 1,070 to 2,069 feet (Turk, 1973). These wells are in a south-southeasterly line, about six miles long, as shown on figure 3. Temperature data on water from seven wells indicate that temperatures increase from north to south along the line, as shown in figure 4. Drillers' logs of those deep brine wells for which temperature data are available are given in the Appendix. For drillers' logs of the other deep brine wells, the reader is referred to Turk (1973).

Hydrologic data on the deep brine aquifer are sketchy. Turk (1973, p. 5) reports specific capacities (Q/s, where Q = discharge, and s = drawdown in feet) between 11.5 gpm/ft and 52.4 gpm/ft. The brines contain 120,000 to 130,000 ppm total dissolved solids, as shown in table 2.

Brackish Water Wells

The data given in this section are abstracted from Turk (1973, p. 2-5).

Table 1. Temperatures of water from brackish water wells, deep brine wells, and fault-line springs (Turk, 1973, tables 5 and 8a).

Source of sample	Date	Temperature		Remarks
		°C	°F	
FW3	8-4-67	35	95	
FW5	8-4-67	31	88	
	9-8-67	31	88	
FW9-A	8-4-67	25.5	78	
FW11	8-4-67	25.5	78	
DBW1	1-22-48	41	106	
	1-23-48	42	108	
	1-24-48	42	108	
	1-25-48	42	108	
	1-26-48	42	108	
	1-27-48	42	108	
	1-28-48	43	109	
	1-29-48	43	109	
	1-30-48	43	109	
DBW3	–	56	133[1]	Depth 1,502 ft
	–	88	190[1]	Depth 1,636 ft
DBW6	7-24-67	27	80	
	8-14-67	27	80	
	9-13-67	27	80	
DBW7	6-16-67	24.5	76	
	7-24-67	25	77	
	8-14-67	24.5	76	
	9-13-67	24.5	76	
DBW8	6-16-67	28	82	
	7-24-67	28	82	
	8-14-67	28	82	
	9-13-67	28	82	
DBW10	6-16-67	25	77	Temperature fluctuation probably the result of short pumping time before sample was collected
	7-24-67	27	80	
	8-14-67	24.5	76	
	9-13-67	23	73	
DBW13	7-25-67	22	71	Same as above
	8-14-67	24	75	
	9-13-67	24.5	76	
Spring No. 1, Blue Lake	9-14-67	29	84	
Spring No. 2, Pilot Valley	7-23-67	24.5	76	

[1] Temperatures of mud bailed when drilling.

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Utah Geology, Vol. 1, No. 1, p. 71 to 82.

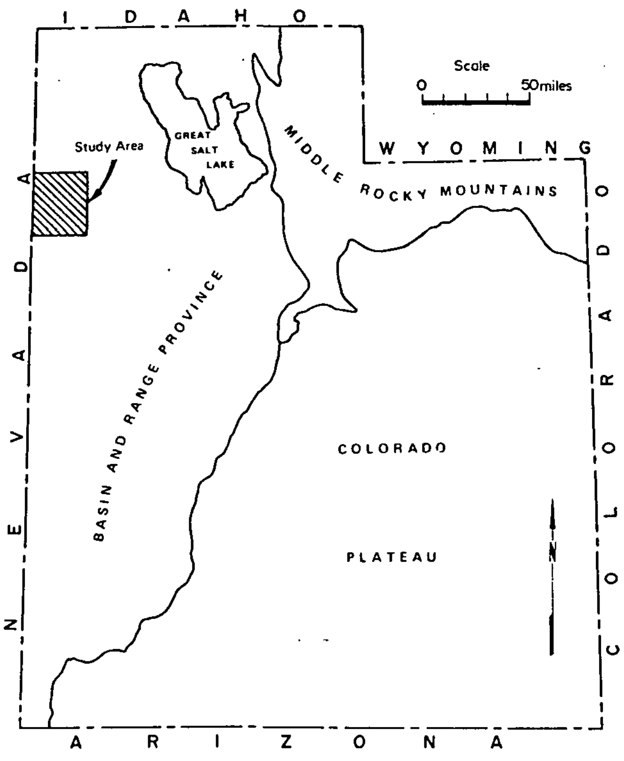


Figure 1. Map of Utah showing physiographic divisions and study area. (Turk, 1973)

Several alluvial fans along the southeast flank of the Silver Island Range are important aquifers yielding large volumes of brackish water. The fanglomerates interfinger with lacustrine sediments near the margins of the salt flats and consist of poorly sorted angular to rounded cobbles, pebbles, sand, and silt.

Twenty-seven wells, most of them less than 100 feet deep, were drilled for Bonneville, Ltd. in the 1940's and 1950's. The water is used for the daily operation of the Bonneville, Ltd. potash plant. Turk noted abnormally high temperatures on four of these wells (1973). A driller's log of one of these wells is given in the Appendix.

Turk (1973, p. 3) states that tests on two wells yielded transmissivities of 159,000 gpd/ft and 412,000 gpd/ft for the alluvial fan aquifers; storage coefficients ranged from 0.00023 to 0.00046. The brackish water wells were all flowing artesian wells when first drilled, but most of them are now being pumped. The potentiometric surface was above ground in 1960, but dropped to more than 19 feet below the surface in 1965 (Turk, 1973). The water levels in these wells have been rising since 1966 (Turk, 1973).

The brackish water contains total dissolved solids of 6,800 to 8,200 mg/l (Turk, 1973). An analysis of water from one brackish water well and water analyses from two warm fault-line springs are given in table 3. Turk (1973) indicates the sources of the water are: (1) rainfall on the fans and runoff from adjacent slopes; (2) brine from the playa; and (3) upward leakage of warm water along the border fault of the range, which is covered by the fans. Turk suggests that surface recharge is significant but not abundant, and that contribution from the playa is indicated by an increase

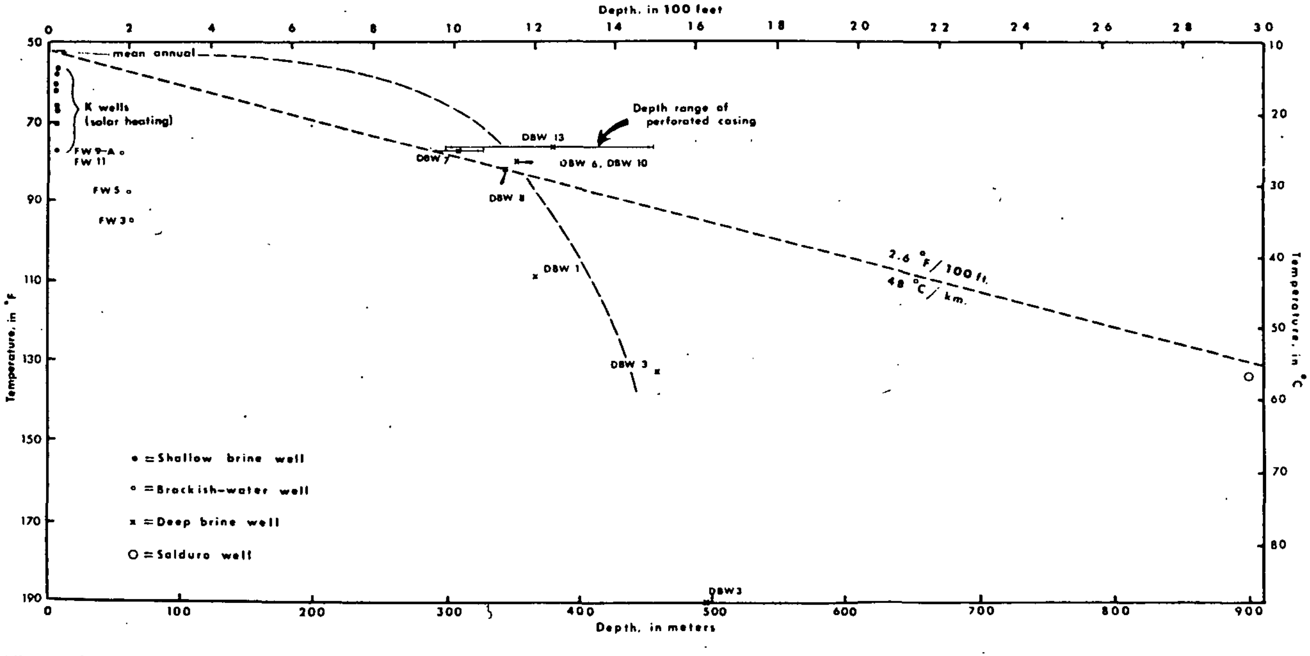
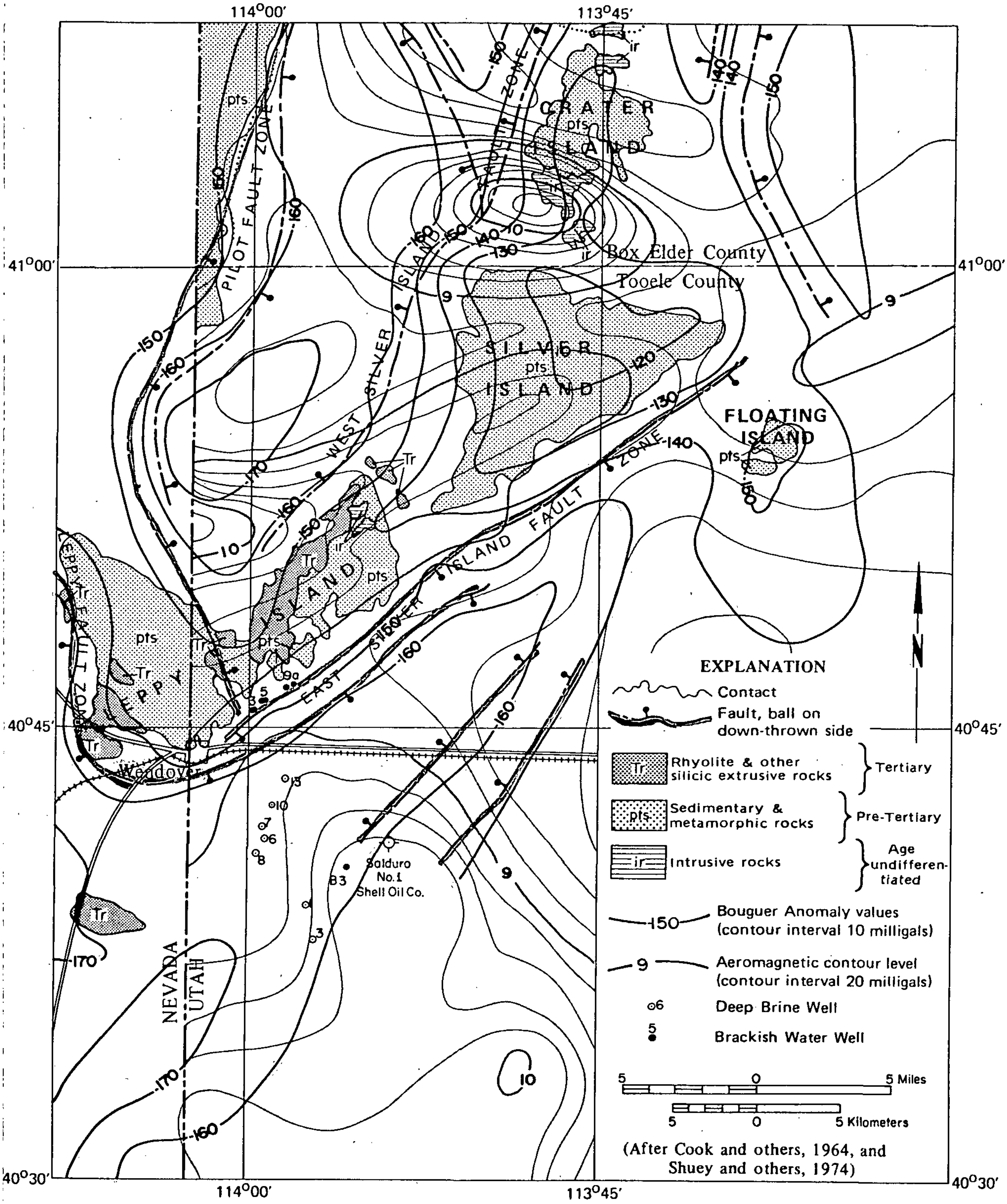


Figure 2. Depth-temperature profile, Bonneville Salt Flats. After Turk (1973, p. 29). Profile of deep brine well gradient added by Whelan and Petersen.



(After Cook and others, 1964, and Shuey and others, 1974)

Figure 3. Bouguer gravity, aeromagnetic, and generalized geologic map of the Bonneville Salt Flats.

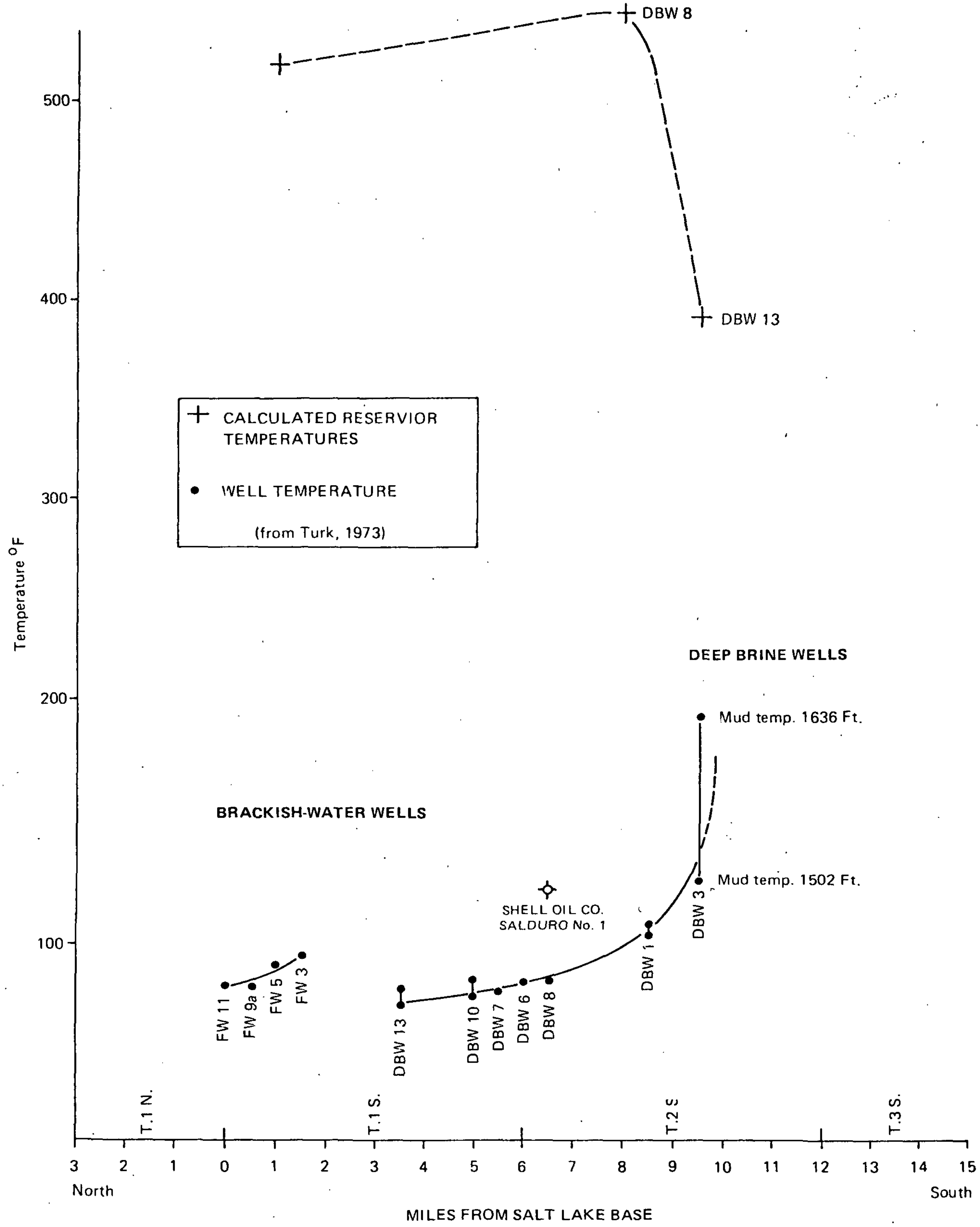


Figure 4. Calculated reservoir temperatures and observed fluid temperatures of brackish water wells and deep brine wells, compared with distance north or south of the Salt Lake Base (latitude).

Table 2. Composition of brine from deep wells (analyses by Kaiser Chemicals, San Leandro, California). (From Turk, 1973, p. 9).

Sample no.	Source	Constituents in parts per million						
		Ca	Mg	Na	Li	K	SO_4	Cl
1.14	DBW8	1,600	1,400	41,400	16	1,800	6,000	70,000
1.15	DBW13	1,500	1,400	46,000	17	2,000	6,200	72,800

in salinity of the water from the wells with time. He attributes the warm water to the same aquifer that supplies hot water to the deep brine wells.

Fault-Line Springs

Turk (1973) notes that the water issuing from fault-line springs on the southeast side of the Silver Island Range and in Pilot Valley is warm. One such spring, Pilot Valley Spring, is 76° F.; and another spring, Blue Lake Spring, is 84° F. Analytical data on the spring waters are given in table 3.

GEOLOGY

Generalized geology of the Bonneville Salt Flats and adjoining mountain ranges, together with gravity data, is shown on figure 3. The geology and gravity data were adapted from Cook *et al.* (1964). Figure 3 also shows aeromagnetic data from Shuey (1974) and well locations from Turk (1973, plate 1).

Bonneville Salt Flats

The Bonneville Salt Flats are located within a large playa in the western part of the Great Salt Lake Desert, near the Utah-Nevada border. The average elevation of the land surface is about 4,215 feet above sea level, with a relief of only 1.53 feet (Turk *et al.*, 1973). The salt crust occupies about 150 square miles and is up to five feet thick in the center (Turk *et al.*, 1973, p. 68).

Wendover Graben

Gravity data by Cook *et al.* (1964) indicate the salt flats are underlain by a structure that they designated as the Wendover Graben, which trends parallel to the front of the Silver Island Range. The graben is probably more than 35 miles long and at least 10 miles in maximum width (Cook *et al.*, 1964, p. 731). Beneath the salt crust, this graben is filled with lacustrine sediments underlain by fluvial sediments. At depths of about 1,200 feet, the deep brine wells encountered "hard rock" or "conglomerate" (Turk, 1973, p. 51-55). Turk *et al.* (1973, p. 66) indicate these hard rocks may be volcanic breccia, corresponding with the post-Early Pliocene and pre-Late Pleistocene volcanic rocks of the Silver Island Range as described by Schaeffer and Anderson (1960, p. 143).

The log of the Shell Salduro #1 well, located in the NW¼ Sec. 4, T. 2 S., R. 18 W. (Salt Lake Base and Meridian) is given as table 4. The volcanics of this well are probably equivalent to the volcanic breccia described by Turk *et al.* (1973). The fact that the hole bottomed in basic igneous rocks, interpreted as an intrusion, is of considerable interest, because similar intrusions might act as heat sources for geothermal systems.

Silver Island Range

The Silver Island Range forms the northwest border of the Bonneville Salt Flats. This range trends northeast and is about 32 miles long. The southwest end of the range is six miles west of the Utah-Nevada border, due west of Wendover. The range has a maximum relief of about 3,000 feet.

There are no outcrops of Precambrian or Mesozoic rocks in the Silver Island Range. However, every system of the Paleozoic Era is represented in the 24,000 feet of exposed sediments. The Tertiary is represented by about 4,000 feet of lacustrine and volcanic strata. Fluvial and lacustrine Quaternary deposits outcrop along the flanks of the range.

The range is intruded by five stocks of unknown age, ranging in composition from quartz monzonite to granodiorite. In the southern part of the range, a light green to black diorite-porphyry stock, about one-half

Table 3. Composition of brackish water and water from fault-line springs (analyses by Kaiser Chemicals, San Leandro, California). (From Turk, 1973, p. 5).

Sample no.	Source	Constituents in parts per million						
		Ca	Mg	Na	Li	K	SO_4	Cl
1.16	FW5	100	80	2,100	1.2	100	300	3,700
1.17	Spring 1	200	50	1,400	1.4	100	200	2,600
1.18	Spring 2	270	50	2,000	1.7	130	100	3,400

Table 4. Log of Shell Salduro #1 well (Sec. 4, T. 2 S., R. 18 W.). (From Utah Oil and Gas Conservation Division.)

Feet		Percent	
From	To		Shows italicized
40	340	100	*Clay*, light grayish green, calcareous, very soft and gummy.
340	570	70	*Clay*, as above.
		30	*Gypsum*, transparent, fractured (or cleaved).
570	610	90	*Gypsum*, as above.
		10	*Clay*, as above.
610	630	50	*Gypsum*, as above.
		50	*Limestone*, light gray, oolitic, very fossiliferous.
630	710	80	*Gypsum*, as above.
		10	*Clay*, as above.
		10	*Limestone*, as above.
710	730	60	*Gypsum*, as above.
		30	*Limestone*, light brownish gray, very fossiliferous.
		10	*Clay*, as above.
730	790	90	*Limestone*, as above.
		10	*Gypsum*, as above.
790	890	100	*Siltstone*, gray, calcareous, argillaceous.
890	970	100	*Limestone*, *siltstone*, *gypsum*, interbedded, as above (Ls., I VFA).
970	1,320	100	*Shale*, *gypsum*, interbedded, shale: medium gray, silty, gummy.
1,320	1,350	100	*Limestone*, medium gray, I VFA, very argillaceous.
1,350	1,489	100	*Volcanic fragments*, composed of basalt, andesite, tuff, etc.
1,489	1,504		Core #1, recovered 11.0 feet.
			Conglomerate, variegated dark reddish brown and dark green, irregular splotches of dark green, poorly consolidated, massive with no apparent bedding. 50% clasts: average diameter ¼ inch, ranging up to 1 inch, composed predominantly of volcanics (andesite or basalt) with rare vitric tuff and buff I VFA, Ls. clasts angular to subrounded. 50% matrix: clay to fine sand, calcareous, soft, predominantly red, rare green, common subrounded coarse grained quartz fragments. Probable age: Tertiary
1,504	1,610	100	*Conglomerate*, composed chiefly of volcanic fragments, altered reddish brown and green, clasts composed of andesite, rare tuff.
1,610	1,740	100	*Volcanics* (andesite?), variegated, green, red, black, fine grained to aphanitic, calcite veinlets common.
1,740	1,800	100	*Limestone*, light gray, sandy, tuffaceous, contains biotite, chert, and rare pyrite.
1,800	1,840	100	*Tuff* (?), bright yellow, altered.
1,840	1,860	100	*Sandstone*, light gray, calcareous, tuffaceous, contains some chert.
1,860	1,900	100	*Volcanic breccia*, composed predominantly of light brown to black andesite, few tuffaceous clasts, fractures filled with calcite.
1,900	1,915		Core #2, recovered 11.0 feet.
			Volcanic breccia, greenish, indurated, no bedding apparent. 60% clasts: very heterogeneous in rock type, composed predominantly of different varieties of andesite, light brown to black, aphanitic to porphyritic, some vesicular, few clasts tuffaceous and in part altered to bentonite. Diameter ranges from ¼ inch to 5 inches, 1 inch average. 40% matrix: probably tuffaceous, noncalcareous, rare opaline cement. Probable age: Tertiary
1,915	2,260	100	*Volcanic breccia*, as above.
2,260	2,280	100	*Bentonite* or *tuff*, medium reddish brown, calcareous.
2,280	2,300	100	*Volcanic breccia*, as above.
2,300	2,340	100	*Bentonite* or *tuff*, light gray, calcareous.
2,340	2,560	100	*Volcanic breccia*, as above.
2,560	2,640	100	*Limestone*, white to light brown, sandy, tuffaceous, some interbedded calcareous sandstone and tuff.
2,640	2,720	100	*Andesite*, generally black and dark green, fine-grained, calcite veins.
2,720	2,740	100	*Limestone*, mottled brown.
2,740	2,820	100	*Basalt*, mottled black, brown, green, contains feldspar laths and probably some olivine.
2,820	2,830		Core #3, recovered 10.0 feet.
			Basalt or possibly *microgabbro* or *diabase* dark gray to black with some green mottling, fine to medium grained, hard, massive, generally fresh but partially altered to chlorite and serpentine. Composed of 70% mafic minerals, chiefly pyroxene with amphibole (hornblende?) and possible biotite, 30% plagioclase feldspar. Minor olivine (?). Mafics partially altered to chlorite and serpentine. Fractures predominantly 45° and 90° filled completely with chlorite and serpentine.
2,830	2,941	100	*Microgabbro*, mottled black, dark gray, brown, green, fine to medium grained, composed of pyroxene, hornblende, plagioclase feldspar, chlorite, serpentine, olivine. Partially altered.

Table 4. (continued)

Feet		Percent	Shows italicized
From	To		
2,941	2,950		Core #4, recovered 9.0 feet.
			Olivine augite diabase, black to dark gray with greenish cast, massive, in part altered to greenstone, commonly fractured–irregular (general) with one well-defined fracture surface dipping 45%, fractures commonly healed with serpentine and chlorite and pyrite, appear tight. Texturally and mineralogically–the rock appears essentially same as in core #3, but with melaphyric dike–3 to 4 inches thick at 2,947 feet.

mile in width, intrudes one of the granodiorite-monzonite stocks (Schaeffer and Anderson, 1960, p. 121). Perhaps the dioritic stock is equivalent to the basic rock encountered at the bottom of the Salduro #1 well.

Dikes of andesite, aplite, lamprophyre, rhyodacite, rhyolite porphyry, dacite, and quartz latite porphyry are found in the Silver Island Range. The dikes range from a few inches to 20-feet wide and are as long as 600 feet.

Schaeffer (Schaeffer and Anderson, 1960, p. 123-124) notes seven volcanic flows of rhyolitic or andesitic composition in the southern Silver Island Range. His "early" volcanic group consists of rhyolites and one andesite, and he gives the group a post-Permian and pre-Pliocene age. He assigns a post-Early Pliocene and pre-Late Pleistocene age to rhyolites and andesites which overlie Tertiary sediments, and he designates them as the "late volcanics". Armstrong (1970, p. 210-211) states that the tridymite rhyolites of Schaeffer's "early" volcanics are 11.6 ± 0.4 million years old.

The Silver Island Range consists of several alternating anticlines and synclines, cut by both normal and reverse faults (Schaeffer and Anderson, 1960, p. 133-139). Schaeffer and Anderson described a border fault parallel to the southeastern margin of the range. The last movement on this fault occurred after Early Pliocene and before Late Pleistocene; the displacement on this fault is 1,100 to 5,000 feet. This fault is one of the western border faults of the Wendover Graben, and it is thought that the warm water of the brackish water aquifer has risen through it.

According to Stokes (1963), the rock outcrops in the vicinity of the warm spring at Blue Lake consist of Late Tertiary rhyolite-dacite-quartz latite flows. Turk (1973, p. 4) considers the warm spring to be a fault-line spring. Alluvial fans on the eastern flank of the Silver Island Range are the site of the brackish water wells with the abnormal temperature gradients described above.

Great Salt Lake Desert

Shuey (1974, personal communication) indicates the gravity high shown on the southern part of figure 3 may represent a buried intrusion. The gravity data of Cook *et al.* (1964) are incomplete in this portion of the Great Salt Lake Desert.

TEMPERATURE OF THE RESERVOIR

Estimates of reservoir temperatures were made for those wells for which Turk (1973) supplied sufficient analytical data. The sodium-potassium-calcium method of Fournier and Truesdell (1973) was used; results are shown in table 5.

Possible reservoir temperatures of 177° to 285° C. (351° F. to 545° F.) indicate adequate heat for a hot water geothermal system. However, the extremely high salinities of the deep brine wells may affect the empirical model used, and the calculated temperatures may not be very accurate. The reservoir temperatures calculated from the brackish water well (FW5) and the warm springs are perhaps better indicators of actual reservoir temperatures.

LAND OWNERSHIP

All of the deep brine wells are on private land owned by Bonneville, Ltd. Some of the brackish water wells are on State School Section 2, T. 1 N., R. 19 W. (Salt Lake Base and Meridian).

Bonneville, Ltd. has rights to the brackish water and brines of the wells. In Utah, under the Geothermal Energy and Associated Resources Act (Section 73-1-20, Utah Code Annotated 1953), geothermal rights are considered to be water rights. Because Bonneville, Ltd. has surface rights and also brine and water appropriations, it appears the company controls the area containing the possible geothermal system.

Land in the surrounding area is federally owned, but withdrawn, as military lands of the Wendover Bombing and Gunnery Range.

Table 5. Analytical data, fluid temperatures, and calculated reservoir temperatures of wells and springs, Bonneville Salt Flats area.

Well or spring	Constituents in parts per million			Specific gravity	Temperature of sampled water °C	Calculated temperature of reservoir	
	Ca^{++}	Na^{+}	K^{+}			°C	°F
DBW8	1,600	41,400	1,800	1.0945-1.0960	28	285	545
DBW13	1,500	46,000	2,000	1.1098-1.0995	22-24.5	199	390
FW5	100	2,100	100	1.039	31	270	518
Spring No. 1 (Blue Lake)	200	1,400	100	1.003[1]	29	181	358
Spring No. 2 (Pilot Valley)	270	2,000	130	1.004[1]	24.5	177	351

[1] Calculated from table A-1 of Levorsen (1958, p. 663).

CONCLUSIONS AND RECOMMENDATIONS

The Bonneville Salt Flats just south of Wendover, Utah, possibly contain a geothermal reservoir. All of the theoretical requirements for the system could be present. There is a buried intrusive which could be a heat source; water, in the form of brines, appears to be present; the faulting of the Wendover Graben could provide the required permeability; and buried volcanics could provide the cap rock.

Land in the area is privately owned, state-owned school sections, or federally owned but withdrawn for military purposes. Because the private land is owned by Bonneville, Ltd., and the state school sections are under lease with brine rights to the same company, the decision to test the geothermal potential or not will rest with that company.

Initial exploration should probably consist of temperature gradient holes, particularly in the center and southern parts of T. 2 S., where the highest-temperature deep brine wells and the highest calculated reservoir temperatures occur. Gradient wells along the eastern edge of the Wendover Graben and over the magnetic high 18 miles southeast of Wendover, which probably represents a buried intrusion, would also be useful.

Analyses of the deep brines, brackish water, and warm springs water should be made for their silica contents, in order to estimate the reservoir temperatures by the method of Fournier and Rowe (1966). If initial results of the gradient holes and silica analyses are favorable, more elaborate geophysical studies should be made, followed by test drilling.

Areas of geothermal potential may be present in other valleys of Utah. Indicators may be:

(1) Abnormal temperatures in water or oil wells,

(2) Aeromagnetic data indicating buried intrusions,

(3) Gravity data indicating faults bordering grabens, which may supply fracture permeability, and

(4) Young intrusive or extrusive rocks exposed in adjacent mountain ranges.

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APPENDIX

Drillers' logs of deep brine wells and brackish water wells (From Turk, 1973).

WELL NO. DBW1

Location: SE¼NE¼ sec. 14 (on east section line), T. 2 S., R. 19 W.
Year drilled: Started 1939, completed 1943
Total depth: 1,200 ft
Casing: 1,175 ft, 8-inch

Depth (ft)	Description	Remarks
0-5	Salt, white, hard	Water in hole
5-50	Clay, light gray, soft	Water in hole to 40 ft; dry hole at 45 ft
50-80	Clay, dark gray, soft	Little water from 55 ft to 70 ft
80-100	Clay, light gray, soft	Dry hole
100-120	Clay, dark gray, soft	Little water 85 ft to 105 ft
120-180	Clay, light gray, soft	Dry hole except little water at 140 ft
180-205	Clay, dark gray, soft	
205-270	Clay, light gray, soft	
270-340	Gypsum and little clay, medium hard	Lots of water at 290 ft; water level 100 ft
340-345	Gypsum and light gray, hard	
345-395	Gypsum and little clay, medium hard	
395-410	Straight gypsum, medium hard	Lots of water at 405 ft
410-415	Clay and showing of gypsum	Water level 30 ft after standing 15 hrs
415-460	Gypsum and some clay, medium hard	
460-525	Gypsum and some clay, medium hard	Water level 268 ft after standing 12 hrs
525-585	Gypsum and little clay, medium hard	Hole caving at 565
585-595	Gypsum and clay, medium hard	
595-620	Straight gypsum, medium hard	
620-660	Gypsum and clay, medium hard	
660-676	Blue clay	
676-677	Hard pan	
677-690	Blue clay	
690-695	Salt and sand	More water; water level 76 ft after 12 hrs
695-705	Blue clay	
705-710	Blue clay and gypsum	
710-760	Blue clay	
760-765	Blue clay and gypsum	
765-791	Blue clay	Trace sand at 780 ft
791-792	Blue clay and gypsum	Drills harder than clay
792-844	Blue clay	
844-852	Blue clay and gypsum	Hole began filling with water at 844 ft, apparently from gypsum
852-866	Blue clay	
866-867	Gypsum	
867-870	Blue clay	
870-875	Blue clay and gypsum	
875-885	Blue clay	
885-898	Blue clay and gypsum	
898-918	Blue clay	
918-920	Fine sand	Water
920-923	Blue clay	
923-930	Fine sand and clay	Water
930-940	Blue clay	
940-950	Fine sand with streaks of clay	
950-955	Blue clay	Water level 420 ft
955-975	White clay	
975-1,000	White clay	
1,000-1,015	White clay	
1,015-1,025	Blue clay	Water broke in
1,025-1,027	Sand, hard	Water level 215 ft
1,027-1,030	White clay	After standing three days the water level is 125 ft
1,030-1,040	Blue clay	Water level 85 ft
1,040-1,058	Blue clay	Water level 90 ft
1,058-1,062	Blue clay	Water level 105 ft
1,062-1,071	Blue clay	
1,071-1,074	Blue clay	
1,074-1,092	White clay	Water level 85 ft
1,092-1,104	White clay and gypsum	
1,104-1,106	Blue clay	
1,106-1,107	Blue clay	
1,107-1,109	Sand	
1,109-1,113	Blue clay	Water level 150 ft
1,113-1,115	Blue clay	
1,115-1,118	Sand	Water level 105 ft
1,118-1,135	White clay, sandy	Water level 85 ft
1,135-1,139	Sandy blue clay	Water level 65 ft
1,139-1,142	Hard sand, gypsum and clay	
1,142-1,145	Blue clay	
1,145-1,151	Blue clay	
1,151-1,152	Yellow clay	
1,152-1,157	Dark blue clay	Water level 80 ft
1,157-1,163	Blue sticky clay	
1,163-1,166	Hard sand	
1,166-1,168	Hard sand	
1,168-1,171	Rock	
1,171-1,177	Rock	
1,177-1,180	Rock (black sand running in from above)	
1,180-1,185	Black volcanic rock	
1,185-1,196	Black volcanic rock	
1,196-1,198	Black volcanic rock	
1,198-1,200	Black volcanic rock	Water level 25 ft (water cannot be bailed below 25 ft)

Appendix (continued)

WELL NO. DBW3

Location: NE¼SW¼ sec. 24, T. 2 S., R. 19 W.
Year drilled: 1949 (?)
Total depth: 2,068 ft
Casing: 36 ft, 20-inch; 418 ft, 16-inch

Note: This well was abandoned and later covered by a pond; no trace of the well at the surface.

Depth (ft)	Description	Remarks
0-249	Blue clay	
249-251	Gypsum	
251-280	Blue clay	
280-282	Gypsum	
282-293	Blue clay	
293-296	Gypsum	
296-400	Clay and gypsum	
400-420	Gypsum (trace gravel)	
420-450	Clay	
450-555	Clay, gypsum, gravel (hard)	
555-575	Gypsum, clay	
575-620	Clay, sticky	
620-630	Gypsum (trace gravel)	
630-646	Clay	
646-652	Gypsum	
652-675	Clay	
675-687	Blue clay	
687-690	Gypsum, clay, gravel	
690-697	Clay, sticky	
697-700	Gypsum (hard)	
700-705	Gray sandy clay	
705-719	Conglomerate	
719-785	Clay, sticky, gray	
785-788	Gypsum (hard)	
788-820	Clay, sticky, gray	
820-822	Conglomerate, hard	
822-835	Clay, gray	
835-836	Gypsum, hard	
836-879	Clay, blue and gray	
879-881	Gypsum	
881-920	Clay, dark, sticky	
920-922	Gypsum	
922-1,017	Clay, gray, sticky to hard	
1,017-1,018	Gypsum	
1,018-1,061	Clay, sticky to hard	
1,061-1,070	Conglomerate, hard	
1,070-1,075	Clay, sticky	
1,075-1,079	Sand and gypsum	
1,079-1,147	Clay, sticky	
1,147-1,148	Conglomerate, hard	
1,148-1,166	Clay, sticky, sandy	
1,166-1,170	Clay, sticky, and gravel (first iron)	
1,170-1,193	Clay, sticky to sandy	
1,193-1,195	Gravel	
1,195-1,210	Conglomerate	
1,210-1,214	Conglomerate, hard	
1,214-1,215	Clay, sticky	
1,215-1,218	Gravel, tight	
1,218-1,220	Clay and gravel	
1,220-1,277	Conglomerate	
1,277-1,281	No log	
1,281-1,289	Conglomerate	
1,289-1,291	Clay, sticky, and gravel	

WELL NO. DBW3 (continued)

Depth (ft)	Description	Remarks
1,291-1,386	Conglomerate	
1,386-1,420	Gravel and sand	
1,420-1,423	Conglomerate	
1,423-1,427	No log	
1,427-1,429	Sand and gravel	
1,429-1,432	Sand and clay, very sticky	
1,432-1,435	Conglomerate, sticky	
1,435-1,449	Conglomerate, hard	
1,449-1,452	No log	
1,452-1,479	Conglomerate, sticky	
1,479-1,514	Sand and gravel, hard; some blue clay (132.8° F at 1,502 ft)	
1,514-1,557	Conglomerate, hard	
1,557-1,558	Clay, sticky	
1,558-1,712	Hard (190.4° F at 1,634 ft)	
1,712-1,768	Conglomerate and gravel	
1,768-1,788	Conglomerate, sticky	
1,788-1,834	Conglomerate	
1,834-1,858	Conglomerate, sticky	
1,858-1,890	Conglomerate and gravel	
1,890-1,907	Conglomerate, sticky	
1,907-1,914	Conglomerate, hard	
1,914-1,965	Conglomerate, light brown	
1,965-2,006	Conglomerate, gray, sticky	
2,006-2,012	Core (no description)	
2,012-2,042	Conglomerate, gray, sticky	
2,042-2,068	Hard	

WELL NO. DBW6

Location: SE¼SW¼ sec. 34, T. 1 S., R. 19 W.
Year drilled: No record
Total depth: 1,153 ft
Casing: No record

Depth (ft)	Description	Remarks
0-265	Clay	
265-268	Gypsum	
268-315	Clay	
315-318	Gypsum	
318-455	Clay	
455-459	Gypsum	
459-487	Clay	
487-489	Gypsum	
489-492	Clay	
492-500	Gypsum	
500-634	Clay with gypsum showing	
634-645	Gypsum	
645-714	Clay	
714-716	Gypsum	
716-835	Clay	
835-875	Hard clay	
875-885	Hard clay with a small amount of gypsum and gravel showing	

Appendix–WELL NO. DBW6 (continued)

Depth (ft)	Description	Remarks
885-902	Sticky clay	
902-918	Hard clay	
918-932	Sand, gravel and clay	
932-944	Hard clay	
944-1,153	Conglomerate	

WELL NO. DBW8

Location: SE¼NW¼ sec. 34, T. 1 S., R. 19 W.
Year drilled: 1950
Total depth: 1,126 ft
Casing: No record

Depth (ft)	Description	Remarks
0-268	Clay	
268-272	Gypsum	
272-320	Clay	
320-323	Gypsum	
323-380	Clay	
380-381	Gypsum	
381-390	Clay	
390-391	Gypsum	
391-400	Clay	
400-416	Clay and gypsum	
416-460	Clay	
460-463	Gypsum	
463-475	Sticky clay	
475-485	Clay	
485-487	Gypsum	
487-496	Sticky clay	
496-507	Gypsum	Caving, hole filled up 11 ft
507-513	Clay	
513-515	Gypsum	
515-565	Clay	
565-571	Gypsum	
571-608	Clay	
608-610	Gypsum	
610-628	Clay	
628-634	Hard gypsum	
634-638	Clay	
638-645	Gypsum	
645-700	Clay	
700-704	Gypsum	
704-745	No log	
745-754	Clay	
754-757	Gypsum	
757-872	Clay	
872-894	Gypsum	
894-918	Clay	
918-930	Clay, gravel showing in sample	
930-944	Gravel	
944-1,039	Conglomerate	
1,039-1,046	Gravel	
1,046-1,060	Hard conglomerate, brown in color	
1,060-1,126	Conglomerate	

WELL NO. DBW8 (continued)

Depth (ft)	Description	Remarks
		Pumping test: 1,300 gpm produced at 1,800 rpm. Hole filled up to 1,045 ft Pumping test: 1,270 gpm with 85 ft drawdown 1,000 gpm with 70 ft drawdown (T≈25,000 gpd/ft)

WELL NO. DBW10

Location: NW¼NE¼ sec. 34, T. 1 S., R. 19 W.
Year drilled: 1951 (?)
Total depth: 1,152 ft (?)
Casing: No record

Depth (ft)	Description	Remarks
0-60	No log	
60-190	Clay	
190-192	Gypsum	
192-262	Clay	
262-265	Gypsum	
265-268	Clay	
268-272	Gypsum	
272-278	Clay	
278-285	Gypsum	
285-321	Clay	
321-323	Gypsum	
323-350	Clay	
350-352	Gypsum	
352-370	Clay	
370-381	Gypsum	
381-498	Clay	
498-505	Gypsum	
505-518	Clay	
518-521	Gypsum	
521-531	Sticky clay	
531-533	Hard	
533-600	Sticky clay	
600-605	Gypsum	
605-632	Sticky clay	
632-643	Gypsum	
643-677	Clay	
677-687	Sticky clay	
687-692	Gypsum	
692-831	Sticky clay	
831-836	Hard	
836-840	Sticky clay	
840-853	No log	
853-860	Hard	
860-862	Gypsum	
862-866	No log	
866-867	Hard	
867-880	Clay	
880-882	Hard	
882-904	Clay	
904-914	Sticky clay	
914-926	Hard	

Appendix–WELL NO. DBW10 (continued)

Depth (ft)	Description	Remarks
926-933	Clay	
933-965	Sticky clay	
965-1,016	Hard	
1,016-1,018	Conglomerate	
1,018-1,115	Hard	
1,115-1,121	No log	
1,121-1,129	Hard	
1,129-1,130	Sticky clay	
1,130-1,131	Hard	
1,131-1,137	Conglomerate	
1,137-1,152	Hard	

WELL NO. DBW7

Location: SW¼NW¼ sec. 3, T. 2 S., R. 19 W.
Year drilled: 1950
Total depth: 1,070 ft
Casing: 138 ft perforated casing on bottom

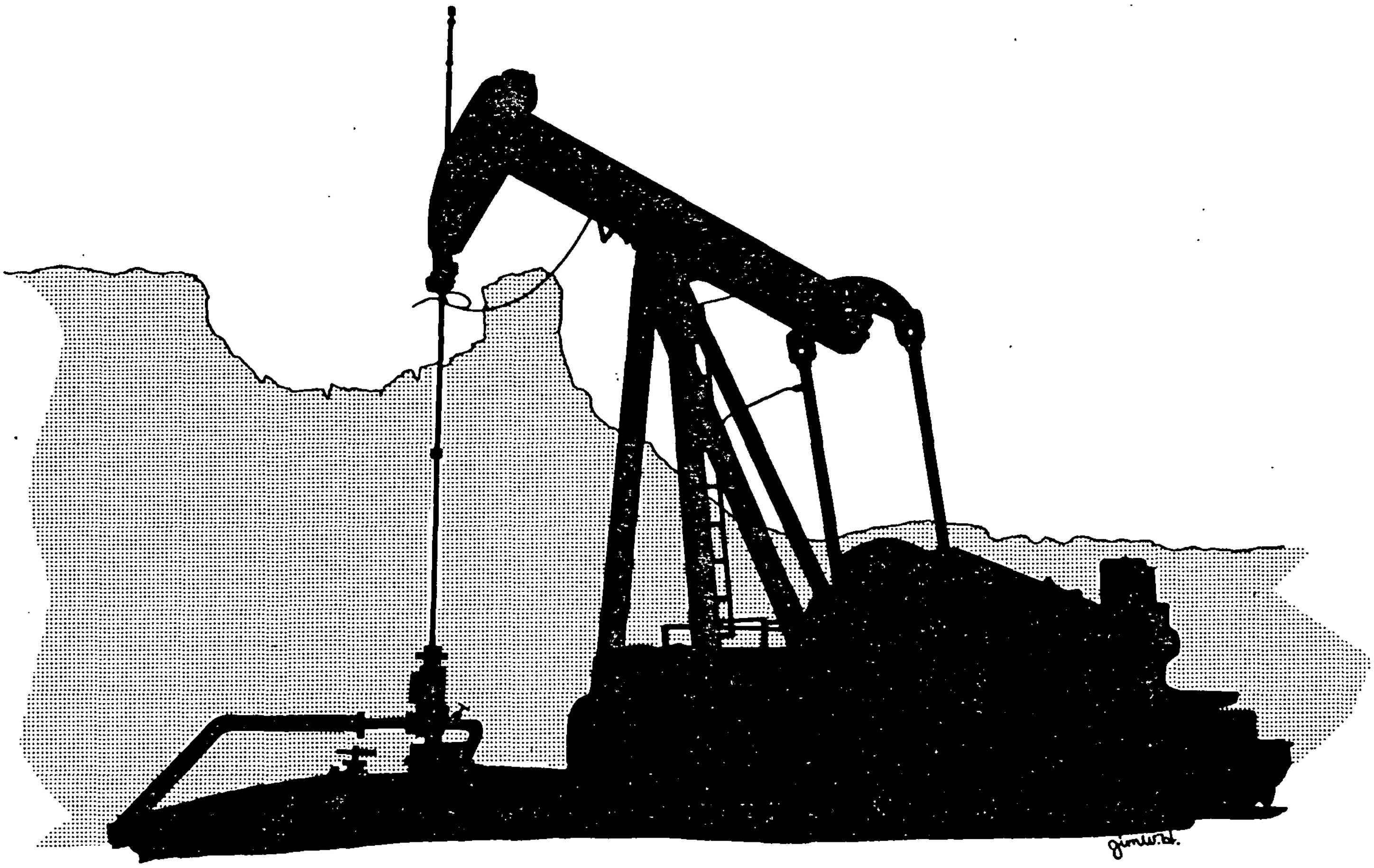
WELL NO. DBW13

Location: NW¼SW¼ sec. 23, T. 1 S., R. 19 W.
Year drilled: 1951
Total depth: 1,496 ft
Casing: 511.5 ft of 10-inch perforated casing on bottom

WELL NO. FW9-A

Location: 713.4 ft E. and 1,555.9 ft N. from SW cor. sec. 2, T. 1 S., R. 19 W. (541.7 ft NE of FW9)
Year drilled: 1947
Total depth: 193 ft

Depth (ft)	Description	Remarks
0-96	Clay	
96-101	Gravel	
101-110	Gravel and clay	
110-121	Gravel	
121-139	Clay	
139-150	Gravel	
150-152	Hard pan	
152-169	Gravel and clay	
169-190	Gravel and conglomerate	
190-193	Loose gravel	



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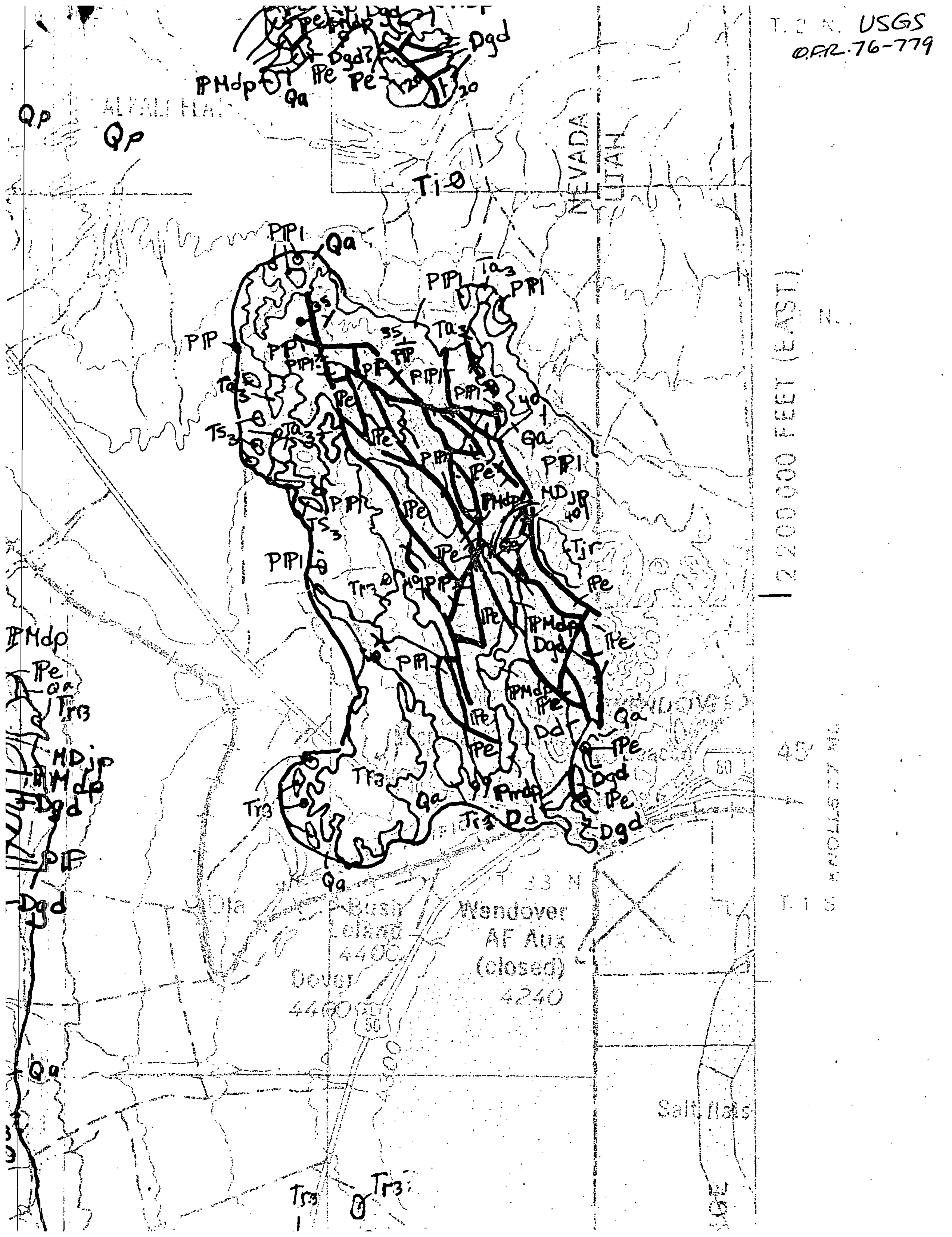


DESCRIPTIONS OF MAP UNITS*

Qf	ARTIFICIAL FILL
Qp	PLUVIAL LAKE DEPOSITS Includes beach and bar gravels and playa silts
Qls	LANDSLIDE DEPOSITS AND COLLUVIUM
Qg	GLACIAL MORAINES AND ROCK GLACIERS Locally, includes outwash gravels
Qa	ALLUVIUM Silt, sand, and gravel along present streams. Includes alluvial fans
QTa	BOULDER GRAVEL AND SAND Deposits on high-level stream terraces
QTs	TUFFACEOUS LIMESTONE, SILT, AND SAND, LOCALLY INDURATED Includes Hay Ranch Formation of Regnier (1960)
QThs	HOT SPRING TUFA AND SINTER, MOSTLY CALCAREOUS
Tt_3	PYROXENE DACITE IGNIMBRITE Phenocryst-poor, with oligoclase, sanidine, pigeonite, and augite
Trt	UNCONSOLIDATED PYROXENE RHYODACITE TUFF
Tgr	ALASKITIC GRANITE (upper Miocene) Extremely variable in texture
Ta_3	PYROXENE ANDESITE AND HORNBLENDE ANDESITE Dark gray, weathering dark brown [3, 4]
Tbg	BANBURY FORMATION (10±0.5 m.y.) Tholeiitic olivine basalt, overlying and underlying gravel, Minor amounts of rhyolitic tuff directly beneath basalt
Tb	BASALT FLOWS Part of Banbury Formation
Tbc	BASALTIC CINDER, TUFF, AND LAVA CONES Part of Banbury Formation
Tb_3	BASALT FLOWS In Sheep Creek Range [19], also Dairy Valley quadrangle [8]
Tcp	COUGAR POINT WELDED TUFFS Gray to brown pyroxene rhyodacite ignimbrites
Tts	IGNIMBRITE, TUFF, AND SEDIMENTARY ROCKS Includes Idavada Formation, and locally, Cougar Point Welded Tuff
Ts_3	TUFF, VITRIC ASH, TUFFACEOUS SILTSTONE AND SANDSTONE, CONGLOMERATE, AND LIMESTONE Includes Humboldt Formation
Tg_3	GRAVEL Includes Slide Creek Gravel [6], Young America Gravel of Bushnell (1967) [5]
Tls	LANDSLIDE DEPOSITS
Tr_3	RHYOLITIC TO DACITIC FLOWS AND DOMES Sempatic quartz and sanidine rich, locally vitrophyric. Hornblende and fayalite sparse [19] [2]
Tjr	JARBIDGE RHYOLITE Light brown to gray phenocryst-rich ferroaugite rhyolite, mostly as flows, but also with some domes and minor tuff
$Tr_{3}p$	PORPHYRITIC RHYOLITE AND RHYODACITE Commonly as domes. Phenocrysts abundant, commonly sanidine, quartz, and oligoclase [19]
Tpi	PIGEONITE ANDESITES Medium gray, holocrystalline hybrid rocks, characterized by xenocrystic quartz, orthoclase, and sodic plagioclase, and by pseudo-uniaxial pigeonite
Tt_2	RHYOLITE TO DACITE IGNIMBRITE Locally vitrophyric. Commonly with phenocrysts of biotite, often also hornblende, quartz, plagioclase, and sanidine
Tr_2	RHYOLITIC FLOWS AND DOMES Relatively fine grained, generally holocrystalline. Characterized by small sparse crystals of biotite. Locally topaziferous
Ta_2	ANDESITIC AND LATITIC FLOWS AND PYROCLASTIC ROCKS Commonly with phenocrysts of hornblende and pyroxene. Includes andesites of Jones Creek [3, 4] and those

RMIAN,
NNSYLVANIAN
, MISSISSIPPIAN

Tr_2	RHYOLITIC FLOWS AND DOMES Relatively fine grained, generally holocrystalline. Characterized by small sparse crystals of biotite. Locally topaziferous
Ta_2	ANDESITIC AND LATITIC FLOWS AND PYROCLASTIC ROCKS Commonly with phenocrysts of hornblende and pyroxene. Includes andesites of Jones Creek [3, 4] and those near Cornucopia [1]
Tb_2	ALKALI-OLIVINE BASALT AND BASALTIC TUFFS AND TUFF-BRECCIAS Includes some minor intrusions. Commonly porphyritic, with common to sparse, large to very large phenocrysts of labradorite. Groundmass includes purplish augite, olivine, and locally biotite and alkali-feldspar. Includes Seventy Six Basalt [4, 6, 14]
Ts_2	TUFFACEOUS AND CLASTIC SEDIMENTARY ROCKS Includes minor amounts of tuff and welded tuff. Near Mountain City, contains Arikareean fossils (C. A. Repenning, written commun.)
Tg_2	GRAVEL Locally tuffaceous
Tw	TUFFACEOUS SEDIMENTARY ROCKS, VITRIC ASH AND TUFF, WELDED TUFF, AND LAVA (Oligocene) Includes limestone, tuffaceous limestone, sandstone. Exposed near Indian Well [43]. Includes Meadow Fork Formation [6] and unnamed gravel [3]

EVONIAN(?)
ILURIAN(?)
RDOVICIAN(?)
ND CAMBRIAN(?)

Tc_1	RHYOLITIC TO DACITIC IGNIMBRITE Commonly micaceous; also may contain hornblende, augite, and hypersthene. In many places with high concentrations of phenocrysts, and of xenoliths of Valmy chert.
Tr_1	RHYOLITIC TO DACITIC FLOWS AND DOMES May include some ignimbrites. Includes Ottawanah Rhyolite [3, 4]
Ta_1	ANDESITIC TO LATITIC FLOWS AND PYROCLASTIC ROCKS Phenocrysts generally include plagioclase, hornblende, pyroxene, locally biotite, or any combination of these mafics. Includes andesite of Summit Creek [3, 4]
Tgd	GRANODIORITE, QUARTZ MONZONITE, AND GRANODIORITE AND QUARTZ MONZONITE PORPHYRY Commonly biotitic, locally with augite or hornblende
$\mathrm{T}\ell_1$	LATITIC ROCK OF BASALTIC HABIT Dark gray to black, holocrystalline to hypocrystalline, very fine grained rocks, locally vesicular, containing olivine, hypersthene, labradorite and potash feldspar. Flows and dike feeders
Ts_1	LIMESTONE, LOCALLY CHERTY, CONGLOMERATE, SANDSTONE, CLAYSTONE, SILTSTONE, SHALE (INCLUDING CARBONACEOUS SHALE AND OIL SHALE), AND TUFF
Tc_1	CONGLOMERATE Fragments generally chert and quartzite, commonly well-rounded locally angular. Locally up to two feet in fragment size, but commonly a few inches. May be stained by iron oxide. May have siliceous cement; also a tuffaceous matrix in some places
Kgr	GRANITE Commonly biotitic
Kgd	GRANODIORITE May be characterized by biotite, hornblende, or both
Kqm	QUARTZ MONZONITE May be characterized by biotite, hornblende, or both
Knc	NEWARK CANYON FORMATION Non-marine conglomerate, sandstone, siltstone, shale and limestone [43]. Clastic rocks commonly gray, tan, brown, and red. Clasts include volcanic rocks, sandstone, quartzite, chert, limestone, and silicified limestone; sizes may range up to one foot. Limestone, dense, silty, gray, tan, to creamy tan in color
KJgd	GRANODIORITE
Jd	DIORITE, includes locally Jgr, granite and Jgd, granodiorite

tuffaceous matrix in some places

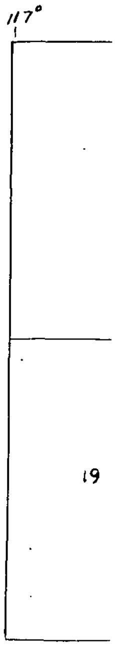
Kgr	GRANITE Commonly biotitic
Kgd	GRANODIORITE May be characterized by biotite, hornblende, or both
Kqm	QUARTZ MONZONITE May be characterized by biotite, hornblende, or both
Knc	NEWARK CANYON FORMATION Non-marine conglomerate, sandstone, siltstone, shale and limestone [43]. Clastic rocks commonly gray, tan, brown, and red. Clasts include volcanic rocks, sandstone, quartzite, chert, limestone, and silicified limestone; sizes may range up to one foot. Limestone, dense, silty, gray, tan, to creamy tan in color
KJgd	GRANODIORITE
Jd	DIORITE, includes locally Jgr, granite and Jgd, granodiorite, not everywhere mapped separately
Jf	FRENCHIE CREEK RHYOLITE Rhyolite flows and other volcanic rocks; some sedimentary rocks
JŦs	SANDSTONE AND SHALE Correlative with Nugget and Aztec Sandstones and Chinle Formation of southern Nevada
Ŧs	MARINE SEDIMENTARY ROCKS Includes Moenkopi and Thaynes Formations, and unnamed Lower Triassic rocks [13, 14]
ŦPs	SILTY LIMESTONE, SHALE, AND MINOR GREENSTONE At north end Adobe Range [29]
ŦPc	MARINE CONGLOMERATE [14]
Pem	EDNA MOUNTAIN FORMATION Coarse sandstone, locally conglomeratic to siltstone, buff-weathering. Typical sandstone is chert-quartz arenite
Pph	PHOSPHORIA FORMATION Chert, phosphatic mudstone, siltstone and limestone
Pmp	MEADE PEAK PHOSPHATIC SHALE, REX CHERT, AND DOLOMITE MEMBERS OF PHOSPHORIA FORMATION
Pgp	GERSTER AND PHOSPHORIA FORMATIONS, UNDIVIDED Carbonate rock, chert, and phosphorite
Ppc	PARK CITY GROUP Carbonate rock and sandstone. Mapped in southern part of Elko County
Ppcg	GRANDEUR MEMBER OF PARK CITY GROUP Carbonate Rock and chert
Pbl	UNNAMED BIOCLASTIC LIMESTONE
Phm	SANDSTONE AND SILTSTONE OF HORSE MOUNTAIN Medium-grained, brown weathering sandstone, largely chert fragments, and dark gray to black siltstone

*Numbers in brackets refer to areas on index map.

PRELIMINARY GEOLOGIC MAP

BY

Pp	PEQUOP FORMATION OF STEELE (1969) Fusulinid limestone. In southern part of the county, includes Arcturus Formation
Pℙhr	HAVALLAH AND RESERVATION HILL FORMATIONS Metagraywacke, fine-grained dolomitic sandstone and siltstone, gray siliceous dolomitic limestone. Some meta-andesite
Pℙcs	CARLIN SEQUENCE OF ROBERTS AND THOMASSON (1964) Includes Buckskin Mountain Beacon Flat and Carlin Canyon Formations of Fails (1966), also Strathearn Formation
Pℙs	STRATHEARN FORMATION Limestone, conglomeratic toward base. Includes Sunflower Formation of Bushnell (1967) [5], [43], [4]
Pℙ	LIMESTONE AND DOLOMITE Includes Winecup Formation of Riva (1970) in HD Range [17] (Upper Pennsylvanian), Rib Hill Sandstone (Permian), Riepe Spring Limestone (Lower Permian) of Steele (1960) in Spruce Mountain [46] and Ferguson Mountain Formation (Lower Permian) of Berge (1960) in southeastern part of county
Pℙl	UNDIVIDED LIMY ROCKS (Lower Pennsylvanian to Lower Permian) In Leppy and Pilot Ranges [36, 42]
ℙmc	MITCHELL CREEK FORMATION Limestone and andesitic tuffs [3]
ℙq	QUILICI FORMATION OF RIVA (1970) [17] Limestone, siltstone and sandstone, and chert conglomerate (DesMoinesian)
ℙe	ELY LIMESTONE (Morrowan to middle Desmoinesian) Limestone, largely of bioclastic origin. Includes Hogan Formation of Robinson (1961) [40]
ℙcd	MOLEEN AND TOMERA FORMATIONS (lower and Middle Pennsylvanian) Cherty limestone, sandy and silty limestone, and conglomerate
ℙMdpc	DIAMOND PEAK AND CHAINMAN FORMATION (Upper Mississippian and Lower Pennsylvanian) Conglomerate, sandstone, and shale, with some limestone
ℙMdp	DIAMOND PEAK FORMATION (Lower Pennsylvanian and Upper Mississippian) Conglomerate, sandstone, shale, some limestone
Ms	SCHOONOVER FORMATION OF FAGAN (1962) (Chesterian) [11] Sandstone, chert, with minor limestone and andesitic lava flows
Mc	CHAINMAN SHALE (Upper Mississippian) Shale and sandstone. Includes Mountain City Formation [3, 4] of shale and calcareous shale, metamorphosed to schist and slate
Ma	ARGILLITE UNIT OF LEE CANYON (Lower Mississippian) [43] Black siliceous argillite
Mw	WEBB FORMATION (Lower Mississippian) [43] Mudstone and claystone, some sandstone and limestone.
Mbn	BANNER AND NELSON FORMATIONS (Osagian to Meramecian) Limestone with quartzite cobble conglomerate at base, grading upward through péperite to meta-andesite [3, 4]
Ml	TRIPON PASS LIMESTONE OF OVERSBY (1973) (Late Kinderhookian) [26] Clastic limestone, argillite, quartz siltite and quartz arenite, and quartz-chert arenite
PMl	LIMESTONE, SHALE, CHERT, ORTHOQUARTZITE, AND QUARTZ SILTITE (Lower Permian to Upper Mississippian) Includes parts of Poorman Peak and Hammond Canyon Formations of Coash (1967) [13] [8, 14]
PMvd	VAN DUZER LIMESTONE OF DECKER (1962) [9, 3, 4, 5] Limestone and shaly limestone



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Mdp	DIAMOND PEAK FORMATION (Lower Pennsylvanian and Upper Mississippian) Conglomerate, sandstone, shale, some limestone	
Ms	SCHOONOVER FORMATION OF FAGAN (1962) (Chesterian) [11] Sandstone, chert, with minor limestone and andesitic lava flows	
Mc	CHAINMAN SHALE (Upper Mississippian) Shale and sandstone. Includes Mountain City Formation [3, 4] of shale and calcareous shale, metamorphosed to schist and slate	
Ma	ARGILLITE UNIT OF LEE CANYON (Lower Mississippian) [43] Black siliceous argillite.	
Mw	WEBB FORMATION (Lower Mississippian) [43] Mudstone and claystone, some sandstone and limestone.	
Mbn	BANNER AND NELSON FORMATIONS (Osagian to Meramecian) Limestone with quartzite cobble conglomerate at base, grading upward through péperite to meta-andesite [3, 4]	
Ml	TRIPON PASS LIMESTONE OF OVERSBY (1973) (Late Kinderhookian) [26] Clastic limestone, argillite, quartz siltite and quartz arenite, and quartz-chert arenite	
PMl	LIMESTONE, SHALE, CHERT, ORTHOQUARTZITE, AND QUARTZ SILTITE (Lower Permian to Upper Mississippian) Includes parts of Poorman Peak and Hammond Canyon Formations of Coash (1967) [13] [8, 14]	
PMvd	VAN DUZER LIMESTONE OF DECKER (1962) [9, 3, 4, 5] Limestone and shaly limestone	6
MDg	GROSSMAN FORMATION (Mississippian(?) or Devonian(?)) [3, 4] Coarse conglomerate, siltstone, sandstone, and phyllite	
MDjp	JOANA LIMESTONE AND PILOT SHALE (lower Mississippian and Upper Devonian) Limestone and argillaceous limestone; carbonaceous shale	7
Dw	WOODRUFF FORMATION (Devonian) Dark siliceous mudstone, shale, and chert, siltstone dolomitic siltstone, and dolomite, with some limestone, sandy limestone and calcareous sandstone [43]	8
Dt	PLATY SILTSTONE, LIMESTONE, AND SHALE (Devonian) [15]	
Dgd	GUILMETTE AND DEVILS GATE FORMATIONS (Upper Devonian) Dolomite and limestone	8A
Dsa	SHALE, CHERT, AND LIMESTONE [29, 30]	
Dl	LIMESTONE (Devonian) [1, 15, 21, 30]	9
Dd	SIMONSON AND SEVY DOLOMITES AND NEVADA FORMATION (Lower and Middle Devonian) Light to dark gray dolomite; limestone	10
DSrm	ROBERTS MOUNTAINS FORMATION (Silurian and Lower Devonian) Platy silty limestone and dolomite [1, 21, 30]	11
DSd	DOLOMITIC LIMESTONE AND DOLOMITE (Silurian and Lower Devonian(?)) Includes Lone Mountain and Laketown Dolomites	12
DOd	PREDOMINANTLY DOLOMITIC ROCKS (Devonian, Silurian, and Ordovician)	
Sl	LONE MOUNTAIN DOLOMITE (includes some Devonian rocks locally in Sulphur Spring Range)	13
Ss	SHALE AND CHERT (Silurian) Includes Noh Formation (Middle Silurian) of Riva [17] and unnamed rocks near Lone Mountain and in Adobe Range [29, 30]	14
SOh	HANSON CREEK FORMATION (Lower Silurian and Upper and Middle Ordovician) Limestone and dolomite	
SOd	DOLOMITE (Lower Silurian and Upper Ordovician) Includes Laketown and Ely Springs Dolomites	15

	Limestone and dolomite
SOd	DOLOMITE (Lower Silurian and Upper Ordovician) Includes Laketown and Ely Springs Dolomites
Ov	VALMY AND VININI FORMATIONS (Ordovician) Chert, shale, siltstone, gray quartzite, greenstone
Ovℓ	LIMESTONE IN VALMY OR VININI FORMATIONS (Ordovician) Clastic and bioclastic locally biohermal
Oa	AURA FORMATION OF DECKER (1962) [9] Brown to black phyllite, partly calcareous, with chert and quartzite
Oe	EUREKA QUARTZITE (Upper Ordovician) White, brown-weathering orthoquartzite
Op	POGONIP GROUP (Middle and Lower Ordovician) Limestone
WF	WESTERN FACIES (Devonian, Silurian, and Ordovician) Mudstone, shale, chert, siltstone, gray quartzite, greenstone, minor limestone
WFℓ	LIMESTONE IN WESTERN FACIES (Devonian, Silurian, and Ordovician) Bioclastic and biohermal limestone, sandy limestone, dolomitic siltstone and dolomite
OЄs	SHALE, PHYLLITE, AND LIMESTONE (Ordovician and Cambrian) Includes Tennessee Mountain Formation of Bushnell (1967 [5] and unnamed shale in Snake Mountains [15, 25]
Єt	CARBONATE ROCK, MINOR QUARTZITE AND PHYLLITE (Cambrian) Includes Edgemont Formation and Porter Peak Limestone of Decker (1962) [9], and schist and limestone of Bushnell (1967) [5]
Є	CARBONATE ROCKS AND MINOR QUARTZITE (Cambrian) Dolomite, limestone, minor shal
ЄpЄpm	PROSPECT MOUNTAIN QUARTZITE (Lower Cambrian and Precambrian Z) Quartzite, with phyllite interbeds
pЄ	MCCOY CREEK GROUP OF MISCH AND HAZZARD (1962) [36] Includes unnamed Precambrian rocks [6, 13], quartzite and phyllite
Dgd'	CALCITE MARBLE Probably part of Guilmette(?) and Devils Gate(?) Formations
Oe'	EUREKA QUARTZITE
OЄ'	CALCITE MARBLE (pogonip(?) and indifferentiated Cambrian(?))
DOd'	DOLOMITE MARBLE (Devonian(?), Silurian(?), and Ordovician(?))
ЄpЄpm'	PROSPECT MOUNTAIN(?) QUARTZITE
pЄ'	SCHIST (Precambrian Z(?)) Sillimanite biotite schist, quartzitic schist
m	METAMORPHIC ROCKS (Lower Paleozoic and Precambrian Z) Granitic to dioritic gneiss, biotite and muscovite schists, locally with sillimanite, quartzitic schist, quartzite, calc-silicate rocks, marble. Includes some granitic intrusives of later age
bx	BRECCIA, MOSTLY FAULT BRECCIA (Age unknown)
i	INTRUSIVE ROCKS (Age unknown) Mostly plutons of silicic rock
gn	GNEISS (Age unknown, locally Mesozoic)

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BONNEVILLE SALT FLATS ---
A POSSIBLE GEOTHERMAL AREA?

by

J. A. Whelan[1]

and

C. A. Petersen[2]

Introduction

In a paper on the hydrology of the Bonneville Salt Flats, L. J. Turk (1973) noted indications of an abnormally high geothermal gradient in several brackish-water wells, several deep brine wells, and two warm springs in the Bonneville Salt Flats area (UGMS Water-Resources Bulletin, Tables 5 & 8a). From Turk's data, Whelan noted that both sets of wells are on essentially north-south lines and that water and brine temperatures are consistently higher to the south.

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Note: Maps and figures to accompany this report will be published along with the text in Utah Geology, vol. 1, no. 1.

In the Utah Geological and Mineral Survey's program of developing and publishing data on potential geothermal areas, it was considered appropriate to expand on the data furnished by Turk.

Descriptions of Warm Wells and Springs

Deep Brine Wells

Thirteen deep brine wells near the western edge of the Bonneville Salt Flats playa range in depth from 1070 to 2069 feet. The wells were drilled between 1939 and 1951 (Turk, 1973). These wells are in a south-southeasterly line, about six miles long. Temperature data on seven wells indicate that temperatures rise from north to south along the line. For drillers' logs, the reader is referred to Turk (1973).

Hydrologic data on the deep brine aquifer are sketchy. Turk (1973, p. 5) reports specific capacities (Q/s, where Q=discharge, and s=drawdown in feet) between 11.5 gpm/ft and 52.4 gpm/ft. The brines contain 120,000 to 130,000 ppm total dissolved solids.

Brackish Water Wells

The data given in this section are abstracted from Turk (1973, p. 2-5).

Several alluvial fans along the southeast flank of the Silver Island Range are important aquifers which yield large volumes of brackish water. The fanglomerates consist of poorly-sorted angular to rounded cobbles, pebbles, sand and silt that interfinger with lacustrine sediments near the margins of the salt flats.

Twenty-seven wells, mostly less than 100 feet deep, were drilled for Bonneville Ltd. in the 1940's and 1950's. The water is used for the daily operation of the Bonneville Ltd. potash plant. Turk noted abnormally high temperatures on four of these wells (1973). A driller's log of one of these wells is given in Appendix A.

Turk (1973, p. 3) states that tests on two wells yielded transmissivities of 159,000 gpd/ft and 412,000 gpd/ft for the alluvial fan aquifers; with storage coefficients ranging from .00023 to .00046. The brackish-water wells were all flowing artesian wells when first drilled, but most of them are now being pumped. The potentiometric surface was above ground in 1960, but dropped to more than 19 feet below the surface in 1965 (Turk, 1973). The water levels in these wells have been rising since 1966 (Turk, 1966).

The brackish water contains total dissolved solids of 6,800 to 8,200 mg/l (Turk, 1973). An analysis of water from one brackish-water well and also analyses of waters from two warm fault-line springs are given in Turk (1973). Turk (1973) indicates that the sources of the water are: (1) rainfall on the fans and runoff from adjacent slopes; (2) brine from the playa; and (3) upward leakage of warm water along the border fault of the range, which is covered by the fans. Turk suggests that surface recharge is significant but not abundant, and that contribution from the playa is indicated by an increase of salinity of the water from the wells with time.

He attributes the warm water to the same aquifer that supplies hot water to the deep brine wells.

Fault-Line Springs

Turk (1973) notes that the water issuing from fault-line springs on the southeast side of the Silver Island Range and in Pilot Valley is warm. One such spring, Pilot Valley spring, is 76°F; and another spring, Blue Lake Spring, is 84°F. Analytical data on the spring waters are given in Turk (1973, p. 5).

Bonneville Salt Flats

The Bonneville Salt Flats are located within a large playa in the western part of the Great Salt Lake Desert, near the Utah-Nevada border. The average elevation of the land surface is about 4215 feet above sea level, with a relief of only 1.53 feet (Turk and others, 1973). The salt crust occupies about 150 square miles and is up to five feet thick in the center (Turk and others, 1973, p. 68).

Wendover Graben

Gravity data by Cook and others (1964) indicates that the salt flats are underlain by a structure that they designate as the Wendover Graben, which trends parallel to the front of the Silver Island Range. The graben is probably more than 35 miles long and at least 10 miles in maximum width (Cook and others, 1964, p. 731). Beneath the salt crust, this graben is filled with lacustrine sediments that are underlain by fluvial sediments. At depths of about 1200 feet, the deep brine wells encountered "hard rock" or "conglomerate" (Turk, 1973, p. 51-55). Turk and others (1973, p.66) indicate that these hard rocks may be volcanic breccia, corresponding with the post-Early Pliocene and pre-late Pleistocene volcanic rocks of the Silver Island Range as described by Schaeffer and Anderson (1960, p. 143).

The log of the Shell Salduro Number One well, is located in the northwest quarter of section 4, T. 2 S., R. 18 W. (Salt Lake Base and Meridian) (well log is on record at Utah Oil & Gas Conservation Division). The volcanics of this well are probably equivalent to the volcanic breccia described by Turk and others (1973). The fact that the hole bottomed in basic igneous rocks, interpreted as an intrusion, is of considerable interest, because similar intrusions might act as a heat source for a geothermal system.

Silver Island Range

The Silver Island Range forms the northwest border of the Bonneville Salt Flats. This range trends northeast and is about 32 miles long. The southwest end of the range is six miles west of the Utah-Nevada border, due west of Wendover. The range has a maximum relief of about 3000 feet.

There are no outcrops of Precambrian or Mesozoic rocks in the Silver Island Range. However, every system of the Paleozoic era is represented in the 24,000 feet of exposed sediments. The Tertiary is represented by about 4,000 feet of lacustrine and volcanic strata. Fluvial and lacustrine Quaternary deposits outcrop along the flanks of the range.

The range is intruded by five stocks of unknown age that range in composition from quartz monzonite to granodiorite. In the southern part of the range, a light-green to black diorite-porphyry stock, which is about one-half mile in width, intrudes one of the granodiorite-monzonite stocks (Schaeffer and Anderson, 1960, p. 121). Perhaps the dioritic stock is equivalent to the basic rock encountered at the bottom of the Salduro Number One well.

Dikes of andesite, aplite, lamprophyre, rhyodacite, rhyolite porphyry, dacite and quartz latite porphyry are found in the Silver Island Range. The dikes range in width from a few inches to 20 feet, and are as long as 600 feet.

Schaeffer (Schaeffer and Anderson, 1960, p. 123-124) notes seven volcanic flows of rhyolitic or andesitic composition in the southern Silver Island Range. His "early" volcanic group consists of rhyolites and one andesite, and he gives the group a post-Permian and pre-Pliocene age. He assigns a post-early Pliocene and pre-late Pleistocene age to rhyolites and andesites which overlie Tertiary sediments, and he designates them as the "late volcanics". Armstrong (1970, p. 210-211) states that the tridymite rhyolites of Schaeffer's "early" volcanics are 11.6 ± 0.4 million years old.

Alluvial fans on the eastern flank of the Silver Island Range are the side of the brackish-water wells with abnormal temperature gradients that are described above.

The Silver Island Range consists of several alternating synclines and anticlines, cut by reverse and normal faults (Schaeffer and Anderson, 1960, p. 133-139). Schaeffer and Anderson described a border fault parallel to the southeastern margin of the range. The last movement on this fault occurred after early Pliocene and before late Pleistocene; the displacement on this fault is 1,100 to 5,000 feet. This fault is one of the western border faults of the Wendover Graben, and it is thought that the warm water of the brackish-water aquifer has risen through it.

According to Stokes (1963), the rock outcrops in the vicinity of the warm spring at Blue Lake consist of late Tertiary rhyolite-dacite-quartz latite flows. Turk (1973, p. 4) considers the warm spring to be a fault-line spring.

Great Salt Lake Desert

Shuey (1974, personal communication) indicates that a gravity high in T. 2 S., R. 18 W.; T. 3S., R. 18 W.; and T. 3 S., R. 17 W. may represent a buried intrusion. The gravity data of Cook and others (1964) are incomplete in this portion of the Great Salt Lake Desert.

Temperature of the Reservoir

Estimates of reservoir temperatures were made for those wells for which Turk (1973) supplied sufficient analytical data. The sodium-potassium-calcium method of Fournier and Truesdell (1973).

Possible reservoir temperatures of 177° to 285°C (351°F to 545°F) indicate adequate heat for a hot-water geothermal system. However, the extremely high salinities of the deep brine wells may affect the empirical model used, and the calculated temperatures may not be very accurate. The reservoir temperatures calculated from the brackish-water well (FW5) and the warm springs are perhaps better indicators of actual reservoir temperatures.

Land Ownership

All of the deep brine wells are on private land owned by Bonneville, Ltd. Some of the brackish-water wells are on State School section 2, T. 1 N., R. 19 W. (Salt Lake Base and Meridian).

Bonneville Ltd. has rights to the brackish water and brines of the wells. In Utah, under the Geothermal Energy and Associated Resources Act (Section 73-1-20, Utah Code Annotated 1953), geothermal rights are considered to be water rights. Because Bonneville Ltd. has surface rights and also brine and water appropriations, it would appear that the company controls the area containing the possible geothermal system.

Surrounding areas are federally owned but are withdrawn military lands of the Wendover Bombing and Gunnery Range.

Conclusions and Recommendations

The Bonneville Salt Flats just south of Wendover, Utah, possibly contain a geothermal reservoir. All of the theoretical requirements for the system could be present. There is a buried intrusive which could be a heat source; water, in the form of brines, appears to be present; the faulting of the Wendover Graben could provide the required permeability; and buried volcanics, the cap rock.

Land in the area is privately owned, or state-owned school sections, or federally owned but withdrawn for military purposes. Because the private land is owned by Bonneville, Ltd., Inc., and the state school sections are under lease with brine rights to the same company, the decision to test the geothermal potential or not will rest with that company.

Initial exploration should probably consist of temperature gradient holes, particularly in the center and southern parts of T. 2 S., R. 18 & 19 W., where the highest-temperature deep brine wells and the highest calculated reservoir temperatures occur. Gradient wells along the eastern edge of the Wendover Graben and over the magnetic high 18 miles southeast of Wendover, which probably represents a buried intrusion, would also be useful.

Analyses of the deep brines, brackish waters, and warm spring waters should be made for their silica contents, in order to estimate the reservoir temperatures by the method of Fournier and Rowe (1966). If initial results of the gradient holes and silica analyses are favorable, more elaborate geophysical studies should be made, followed by test drilling.

Areas of geothermal potential may be present in other valleys of Utah: Indicators may be:

(1) Abnormal temperatures in water or oil wells.

(2) Aeromagnetic data indicating buried intrusions.

(3) Gravity data indicating faults bordering grabens, which may supply fracture permeability.

(4) Young intrusive or extrusive rocks exposed in adjacent mountain ranges.

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Hydrogeology of Lacustrine Sediments, Bonneville Salt Flats, Utah

L. J. Turk, S. N. Davis, and C. P. Bingham

Abstract

Bonneville Salt Flats is a large salt pan deposited by an intermittent playa lake in western Utah. Magnesium and potassium chlorides are recovered from brines collected in a system of ditches and concentrated in large solar evaporation ponds. Recharge to the brine system is by rain and saline well water. Some recharge to the area of brine production may come from overland flow during wet winters.

The upper ten feet of the Bonneville sediments constitutes an aquifer of high transmissivity which is attributed to, first, the high hydraulic conductivity of the salt bed at the surface, second, fractures in some of the underlying layers of silty clay, and, third, permeable lenses of sand-size brine-shrimp pellets. Transmissivities in parts of the aquifer exceed 50,000 gallons per day per foot. The occurrence of open fractures in silty clay is explained either by osmotic desiccation of buried clays induced by an increase in salinity of near-surface waters or by syneresis.

Introduction

Regional setting

Physiography: The Bonneville Salt Flats is near the Nevada-Utah border in the western part of the Great Salt Lake Desert (Fig. 1). The salt crust, which covers approximately 150 square miles, occupies the western edge of a large, flat playa. Average altitude of the land surface is about 4,215 feet above sea level. A survey of the area in 1966 yielded a maximum difference of only 1.53 feet in the altitude of the natural surface. Before the area was disturbed by man it had less physiographic expression than a table top.

Silver Island Mountains border the salt flats on the northwest, rising abruptly to altitudes greater than 6,500 feet. Shoreline features of several levels of ancient Lake Bonneville can be seen clearly on the flanks of this range as well as other high mountains to the west and southwest. To the northeast, east, and south, the nearly flat desert surface slopes upward to effectively close the topographic depression that holds the salt crust.

The salt flats have no exterior surface drainage. Runoff from Silver Island Mountains and a small amount from the surrounding flatland creates a shallow lake in the depression during wet seasons. In the spring of 1967 the playa lake exceeded a depth of 1.5 feet in some places, but levees, roads, and a railroad bed prevented equal flooding of the entire area.

Climate: The average annual temperature at the U. S. Weather Bureau station in Wendover is 52° F. The maximum recorded temperature was 112° F; the minimum was —19° F.

Yearly precipitation at Wendover during the period 1912–1967 ranged from 1.77 inches to 10.13 inches, and averaged 4.74 inches. Although the long-term average precipitation at a given location is distributed rather evenly throughout the year, scattered thunderstorms commonly cause rainfall to be highly sporadic during any particular year. Orographic effects probably produce measurable differences in mean annual rainfall from one place to another, particularly along the western margin of the salt flats.

Geology: Silver Island Mountains consist mainly of Paleozoic limestone and dolomite but also contain widespread Tertiary volcanic rocks in the southwestern end of the range. Several large silicic stocks are exposed northeastward on Silver Island (Schaeffer and Anderson, 1960). The apparent lack of Mesozoic sedimentary rocks is because the area was part of the Mesocordilleran geanticline shown by Eardley (1951, pls. 15 and 16).

Present relief of Silver Island Mountains is a result of Basin-and-Range faulting which began in early Tertiary and probably continued to late Pliocene. A border fault along the southeastern edge of the moutains near Wendover has 1,100 to 5,000 feet of displacement (Schaeffer and Anderson, 1960, p. 148–149). A gravity survey verified the existence of such a fault, and suggested that several additional faults are present beneath the Plio-Pleistocene sediments of the salt flats (Cook and others, 1964).

Several deep brine wells on the western edge of the playa topped "hard rock" or "conglomerate" at

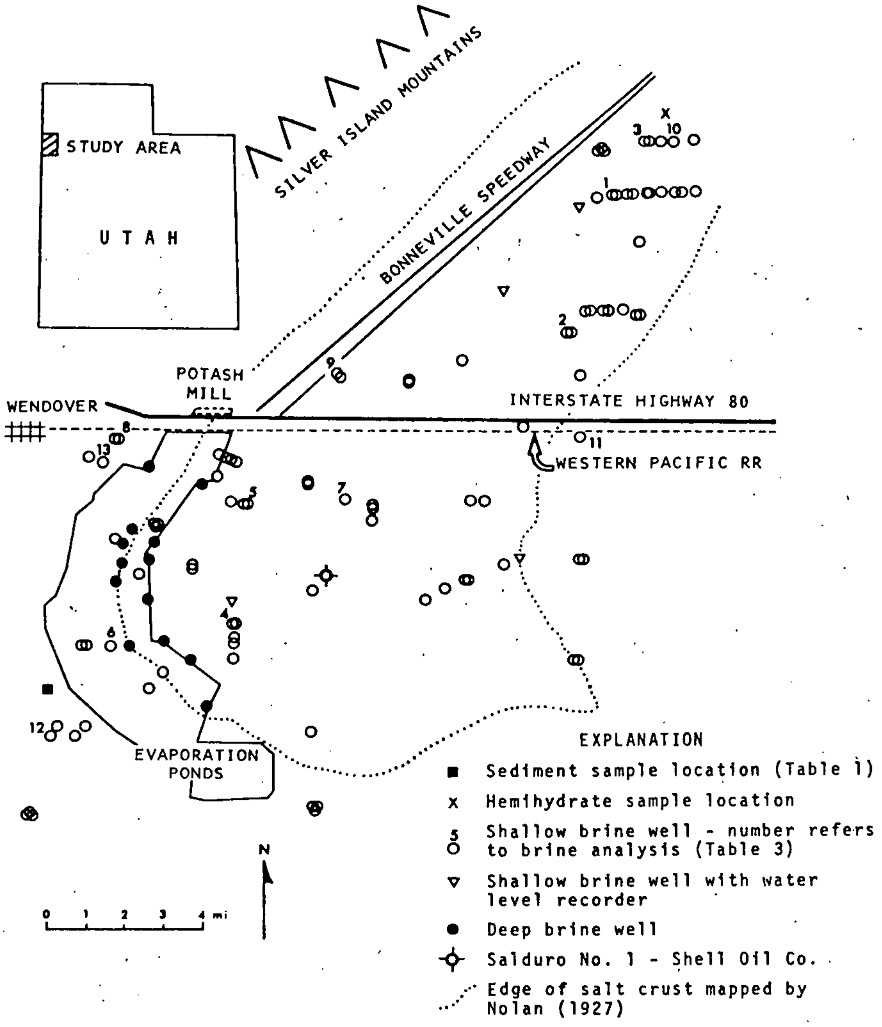

FIG. 1. Index map of Bonneville Salt Flats.

depths of 1,168 feet to 1,200 feet. The hard rock is probably volcanic breccia corresponding with the post-early Pliocene and pre-late Pleistocene volcanic rocks described by Schaeffer and Anderson (1960, p. 143). Shell Oil Company's Salduro No. 1 well (Fig. 1) reportedly encountered volcanic rocks at a depth of 1,375 feet and penetrated nearly 1,400 feet of volcanic rocks before the hole passed into basic intrusive rock of possible Cretaceous age at 2,742 feet (Heylmun, 1965, p. 28).

The basin above the volcanic rocks was filled with fluvial, and later, lacustrine sediments, which are the focus of this paper.

Placement of the salt crust by isostatic adjustment: Gilbert (1890) first proposed the theory that upwarping of the central part of the Lake Bonneville basin, as shown by the tilting of old shorelines, was caused by isostatic compensation as weight was removed by evaporation of the lake water. Geophysical data presented by Crittenden (1963) confirmed Gilbert's conclusions. Eardley (1962) presented the following explanation of the adjustment, which accounts for the present location of the Bonneville salt crust.

Original closure of the desert basin was centered approximately 23 miles east of the present center of the salt crust. The western side of the basin (Wendover area) was at a slightly higher altitude during the early stages of desiccation of the lake. As the lake level dropped and exposed the western part of the basin, erosion, mostly by wind, removed some of the younger sediments. With continued gradual unloading of the basin, upward isostatic rebound eventually elevated the central part of the basin, now the eastern side of Great Salt Lake Desert, higher than the western side. And hence the salt crust, originally deposited in the center of the basin, shifted to the west by periodic dissolution by rainwater and re-

TABLE 1. Composition of Salt Crust (From Utah State University Engineering Experiment Station, 1960)

Depth Below Surface	Total Soluble Solids[2]	Cations				Anions		
		Ca	Mg	Na	K	Cl	SO_4	HCO_3
Inches	Percent	Percentage on chemical equivalent basis						
0 - 1.0	94	3.3	0.14	96.3	0.25	97.0	2.7	0.35
1.0- 2.5	100	2.6	0.00	97.1	0.24	97.8	1.9	0.35
2.5- 4.0	98	2.3	0.13	97.3	0.18	97.9	1.7	0.33
4.0- 5.5	100	2.2	0.23	97.3	0.19	98.0	1.7	0.26
5.5- 7.0	100	5.0	0.12	94.7	0.21	95.2	4.6	0.27
7.0- 8.5	100	6.9	0.00	93.0	0.15	93.3	6.3	0.33
8.5-10.0	100	2.7	0.02	97.1	0.18	97.4	2.4	0.25
10.0-11.5	100	1.9	0.07	97.8	0.15	98.2	1.5	0.29
Average	99	3.4	0.09	96.3	0.19	96.8	2.8	0.30

[1] Sample location was listed as 3.5 miles east from the Utah-Nevada state line along U. S. Highway 80 route, then 1 mile south.

[2] Percent of dry weight of the sample.

precipitation. The closed basin shifted farther westward each time the cycle was repeated until the salt crust finally came to rest in its present location at the foot of Silver Island Mountains. The migration of the salt crust was the result of differential isostatic adjustment—areas which held deeper water rebounded more than areas of shallower water. The resulting reversal of slope created the closed topographic depression which now contains the salt pan.

Economic importance: The Bonneville Salt Flats is one of the few places in the United States where potash has been recovered from natural brines by solar evaporation. Production of potash began in 1914, ended in 1922 as a result of the post-World War I decline in demand for potash, then resumed in 1935. The brine deposit is owned by Kaiser Aluminum & Chemical Corporation.

The system for collecting brine consists of more than 90 miles of ditches, dug with draglines to a depth of 20 to 22 feet and an initial width of four to six feet. The ditches are cleaned about every two years and eventually are widened to as much as 20 feet. Brine enters the ditches by gravity flow, then is elevated by booster pumps along its route to a series of solar evaporation ponds. Most of the NaCl precipitates in large primary ponds; the remaining brine, high in KCl content, flows into smaller "harvest" ponds where sylvinite (a mixture of KCl and NaCl) precipitates. Effluent from the harvest pond proceeds through a final evaporation pond where carnallite precipitates, then the residual liquor, rich in $MgCl_2$, is stored in open ponds until it is sold in its liquid state as a by-product. Large mechanical earth-moving equipment harvests the sylvinite and transports the ore to a mill where it is concentrated by a flotation process to the grade required for marketing (about 96 percent KCl).

Previous work

Since G. K. Gilbert included the Bonneville area in the first U. S. Geological Survey Monograph (1890), the Bonneville Salt Flats has been a subject of continued geological interest. The area has been described in many studies of Basin-and-Range structure (e.g., Gilbert, 1928; Nolan, 1943). The Utah Geological and Mineralogical Survey has sponsored several studies in the area. Eardley (1962) worked out the history of Great Salt Lake Basin on the basis of C^{14} dates obtained on samples from several locations across the flats. Kaliser (1967) presented results of a short hydrogeological investigation based on pumping tests in shallow wells near the Bonneville Speedway.

Gale (1914) was one of the first to discuss the economic potential of extracting potash from brines underlying the salt, but Nolan (1927) published the first intensive scientific study of the Bonneville brines, including numerous brine analyses and a map of Great Salt Lake Desert showing the distribution of potash in the near-surface brines. Hutchinson investigated the potash economics for a company called Bonneville Limited, but the results of his work were never published. Additional unpublished private reports prepared over the years (by Nackowski and Mehrhoff in 1960 and 1961; Nackowski in 1962; and Davis in 1966 and 1967) led to this investigation, which in part is a continuation of Davis' work.

Perhaps the largest quantity of data available for use in this study was assembled by the Utah State Department of Highways from foundation and route investigations for the new U. S. Interstate Highway 80 (Utah State University Engineering Experiment Station, 1960, 1962, 1963).

TABLE 2. Composition of Playa Sediments[1]

		Sample Number[2]				
Particle Size		1	2	3	4	5
		Weight Percent of Total[3]				
Sand		—	—	—	—	—
Silt		21.3	71.3	13.0	13.3	76.5
Clay		78.7	28.7	87.0	86.7	23.5
		Calcium Carbonate,[4] Weight Percent of Size Fraction				
Sand		—	—	—	—	—
Silt		50.6	57.2	68.8	76.8	91.8
Clay		74.7	82.7	77.7	78.3	87.0
Total Sample		69.6	64.5	76.5	78.1	90.7
Mineral[5]		Weight Percent of Size Fraction				
Sand	Aragonite	—	—	—	—	—
	Quartz	—	—	—	—	—
	Montmorillonite[6]	—	—	—	—	—
Silt	Aragonite	50.6	57.2	67.3		
	Calcite	—	—	1.5	76.8	91.8
	Quartz	49.4	26.2	—	21.4	3.8
	Gypsum	—	16.6	—	—	2.2
	Montmorillonite	—	—	31.2	—	2.2
	Illite	—	—	—	1.8	—
Clay	Aragonite	74.7*	75.2	77.7*	78.3*	87.0*
	Calcite		7.5			
	Quartz	21.4	2.6	17.6	16.3	13.0
	Gypsum	—	6.1	—	—	—
	Montmorillonite	—	6.0	4.7	5.4	—
	Illite	—	2.6	—	—	—
	Stilbite	1.2	—	—	—	—
Total Sample	Aragonite	69.6*	62.4	76.5*	78.1*	90.7*
	Calcite		2.1			
	Quartz	29.5	19.5	15.4	17.2	5.9
	Gypsum	—	13.6	—	—	1.7
	Montmorillonite	—	1.7	8.1	4.7	1.7
	Illite	—	0.7	—	0.2	—
	Stilbite	0.9	—	—	—	—

[1] Analyses by J. A. Whelan, Utah Geological and Mineralogical Survey.
[2] Source of samples shown on Figure 2A.
[3] Determined by centrifuge.
[4] Weight loss in cold dilute hydrochloric acid.
[5] Determined by X-ray diffraction of whole and acid leached size fractions; quantities *estimated* from limited standards.
[6] Probably entrapped.
* Aragonite and Calcite.

Hydrogeology

Mineralogy and lithology

Salt crust: The Bonneville salt crust is lens-shaped in cross section, ranging in thickness from a feather edge to almost five feet near its center. Eight samples analyzed by the Utah Soil and Water Laboratory are listed in Table 1. Very little potash (KCl) is present in the salt crust. Because of its high solubility, most of the KCl remains in solution in the ground water. The salt samples contained no detectable amounts of lithium or strontium.

A black, odorous substance occurs beneath the crust in some places where the salt crust is thin and adjacent to areas where the crust is underlain by a thin coating of light-green algae. The substance exhibits most of the physical properties of black organic mud. Professor J. A. Whelan of the University of Utah analyzed a sample of the substance using X-ray and differential thermal analysis techniques (see Fig. 1 for sample location). The substance contains a hemihydrate of calcium sulfate, which is intermediate between anhydrite and gypsum, or approximately $CaSO_4 \cdot \frac{1}{2} H_2O$. The substance appears to be similar to the hemihydrate from Clayton Playa, Nevada, described by Moiola and Glover (1965). However, this, as well as the Clayton Playa occurrence, may be the result of dehydration of the sample before analysis (B. F. Jones, 1972, written communication).

Clastic sediments: Shell Oil Company drilled 1,375 feet of claystone, gypsum, oolitic limestone, and siltstone in the Salduro No. 1 well (Fig. 1), and several other wells in the desert penetrated more than 1,000 feet of Late Pliocene and Pleistocene sediments deposited unconformably on volcanic rocks (Heylmun, 1965, p. 29).

The upper 20 feet of these sediments comprise the aquifer which yields the commercial brine of primary interest in this study. Bedding in the upper playa sediments can be traced for thousands of feet, and in some places even miles, along exposures in production ditches. Dips seldom exceed two or three feet per mile. Silt- and clay-sized particles dominate the near-surface sediments. Sand-sized particles comprised mainly of brine-shrimp fecal pellets deposited on the bottom of Lake Bonneville occur in thin, discontinuous lenses, except in the northeastern part of the area where relatively dry, hard beds of fecal-pellet sand are encountered at depths of 14 to 18 feet. Average particle size increases from west to east because of an increase in the amount of fecal-pellet sand. Compact, nodular layers of sand-sized aragonite are commonly interbedded with the fecal-pellet sand.

Determinations of the clay mineralogy of the shallow sediments have been somewhat contradictory. Ten samples taken near the Bonneville racetrack and near Salduro by the Utah State Highway Department reportedly contained 14 to 42 percent montmorillonite and 0 to 11 percent kaolinite; the bulk of the samples was aragonite with some quartz and feldspar. Kaiser's Permanente Laboratory in San Leandro, California, X-rayed 25 sediment samples from eight shallow wells and found a general absence of clay minerals. Aragonite again was the predominant mineral in all but a few samples, halite was next most common, and traces of calcite, quartz, and dolomite were present in all the samples. Gypsum and feldspars occurred sporadically.

Inasmuch as knowing the abundance of various clay minerals is essential in determining the cause of fracturing in the sediments, samples were taken from five different layers exposed in a ditch face in the southwestern part of the salt flats (Fig. 2A). X-ray analyses of the various size fractions (essentially all silt and clay sizes) are listed in Table 2. The analytical data clearly show a large variation in the clay mineral content both from layer to layer and from size fraction to size fraction within a single layer. Montmorillonite varies from 0 to 8.1 percent of the total sample, from 0 to 31.2 percent in the silt sizes and from 0 to 6.0 percent in the clay sizes. Thus the earlier conflicting sets of data merely prove the extreme range of clay mineral content. Neither set of analyses is necessarily less valid than the other.

To sum up, aragonite is the dominant mineral in all size fractions (more than 50 percent in every sample), quartz is the most common secondary mineral, and montmorillonite is the most abundant clay mineral. Minor amounts of gypsum, illite, and stilbite occur throughout the section.

Fractures in clays

Some layers of the shallow sediments contain many vertical fissures, up to one inch wide, which are the principal conduits that transmit brine to the production ditches. Fissures may be exposed in the ditch face when the brine level is drawn down by pumping (Figs. 2A,B,C). Spacing between the cracks is variable, but averages about one foot in most places. The orientation of the fissures in plan view is only poorly known. A few shallow excavations west of the evaporating ponds uncovered open cracks with a hexagonal pattern. Similarly, benchlike exposures along old ditches suggest that the cracks have a hexagonal pattern in plan view. Cracks in the overlying salt crust also display well-developed hexagonal patterns which may be upward extensions of the patterns of the clay fissures. Not all the beds are highly fissured; some have few or no cracks, especially the sandy or more indurated beds.

Aquifer characteristics

Porosity and specific yield: Porosity of the aquifer was determined by numerous wet and dry bulk density measurements by the Utah State Highway Department (1960), Nackowski and Mehroff, and Bingham. The average porosity is 45 percent. Nackowski and Mehrhoff calculated the average specific yield as 6.3 percent, based on sediment samples from above the water table, but the samples doubtless were within the capillary fringe and contained brine held in capillary pores. Bingham's laboratory tests, which included gravity drainage of saturated samples from below the water table, indicated a specific yield of about 10 percent.

Aquifer tests: More than 90 shallow wells, six inches in diameter, were augered by hand during 1965–67. Most of the wells were lined with four-inch slotted plastic casing and gravel packed. Depths of the wells ranged from 13 to 30 feet; the most common depth was 23 feet. Transmissivities and storage coefficients of the playa sediments were determined by 70 aquifer tests, most of which were of less than 3 hours' duration. Pumping rates generally were one to ten gallons per minute. Tests in 1965–1966 were performed with a gasoline-powered suction pump, which gave acceptable results although its discharge fluctuated somewhat. Aquifer tests in 1967 were with a small electric-powered pump driven by a gasoline generator. Discharge was remarkably constant and results generally excellent. Results of twelve previous tests by engineers of Utah State University (1960) were also incorporated in the study.

Data from aquifer tests were analyzed by the nonequilibrium method of Theis (1935) and the Theis-

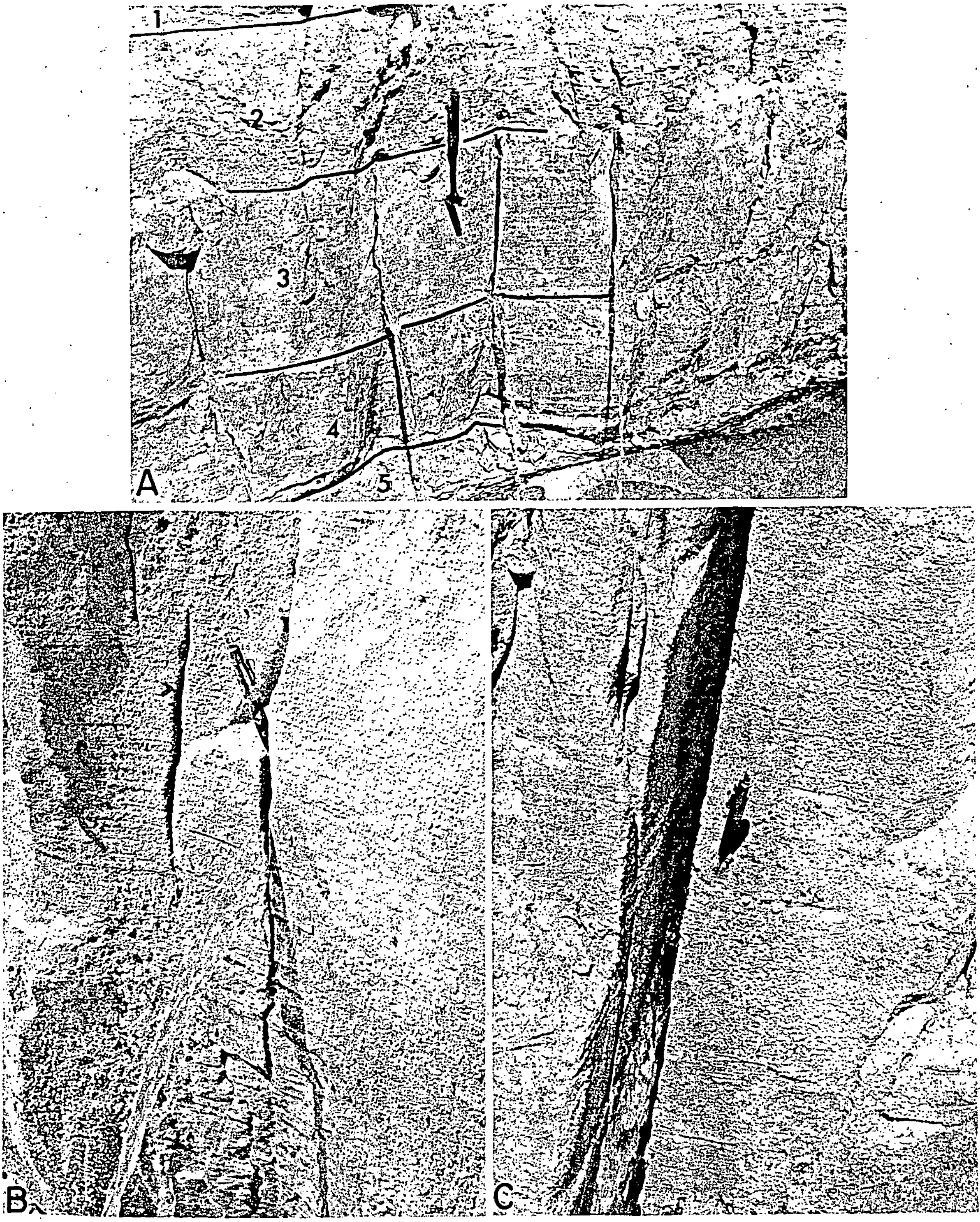

FIG. 2A. Ditch face showing fissures in near-surface sediments. Note the brine issuing from fissures. Sediment samples were taken from the numbered layers. (Sample location is shown on Figure 1, analyses are listed in Table 1.) FIG. 2B–2C. Closeup views showing shape of the fissures. Ballpoint pen is for scale.

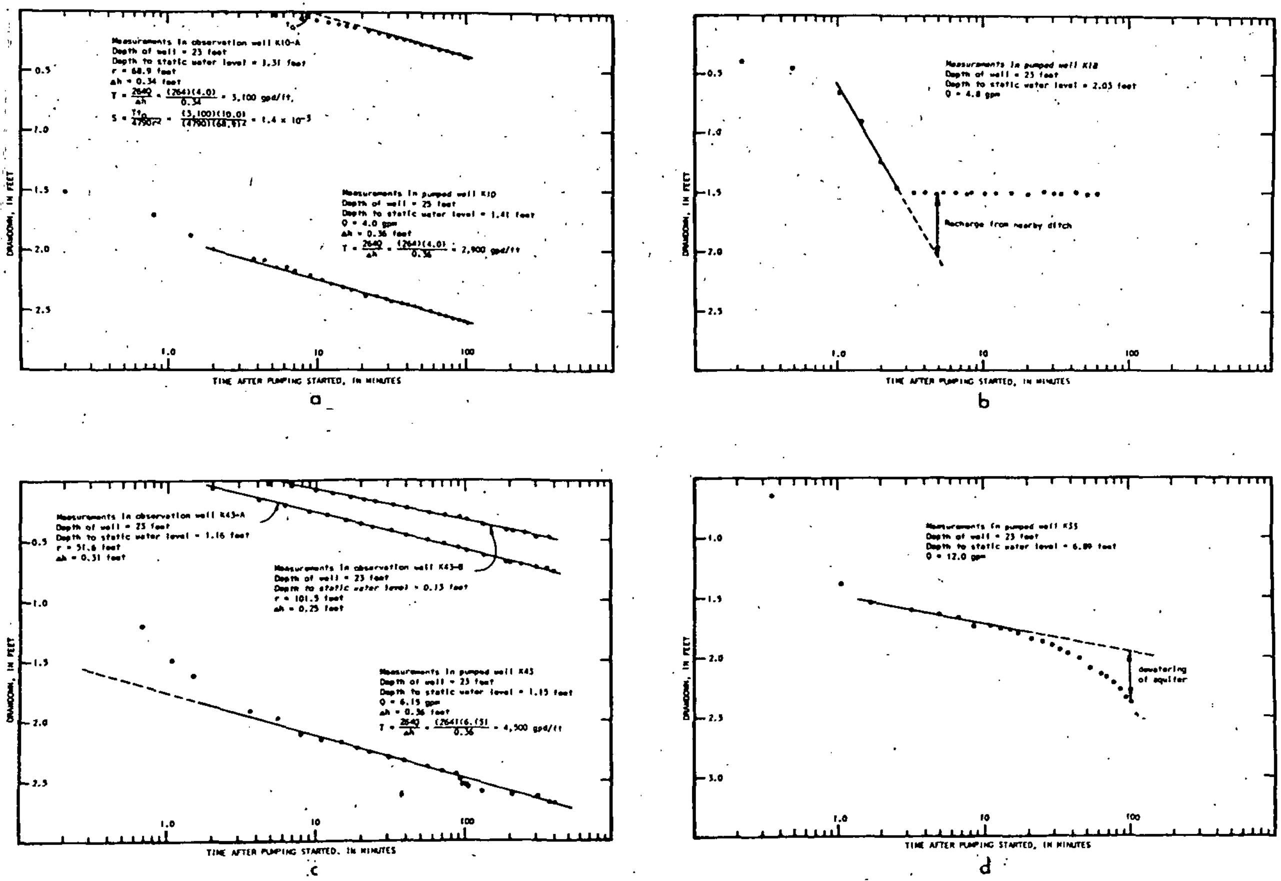

FIG. 3. Hydrographs of wells during pumping tests. Q = pumping rate in gallons per minute (gpm), r = radial distance to observation well in feet, Δh = drawdown in feet per log cycle of time, T = transmissivity of aquifer in gallons per day per foot (gpd/ft) under a hydraulic gradient of one foot per foot, S = storage coefficient. See Jacob (1950) and Walton (1962), for a description of the method of interpretation. Discussion of the hydrographs is in text.

Jacob semilogarithmic method (Jacob, 1950). Where applicable, various vertical and horizontal boundary conditions were taken into consideration (Ferris, 1949; Hantush, 1956). Many test wells penetrated several permeable layers which were most commonly salt at the surface with one or more layers of shrimp-pellet sand and fractured clay in the subsurface. Wells which reacted in a manner suggesting water-table conditions were analyzed by the method of Boulton (1963) and Prickett (1965). Due to the uncertainty of the number and exact character of brine-yielding layers, the analyses, although yielding numerical values, must be considered to be only semiquantitative. As Neuman and Witherspoon (1969a) have pointed out, even the leakage analyses of Hantush (1956) can be quite misleading provided aquitards yield significant amounts of fluid from storage. Construction of wells so that isolated permeable zones could be tested separately for long periods of time was not feasible for this study. Hence, the more exact methods of analyses outlined by Neuman and Witherspoon (1969b) could not be applied.

Storage coefficient:[1] Twenty-two tests were in paired wells, allowing calculation of the coefficients of storage, which ranged from 1.2×10^{-1} to 5.0×10^{-5}. This indicates a wide range of aquifer conditions, from unconfined to well-confined. The aquifer is most commonly unconfined to semi-confined.

Transmissivity:[2] Drawdown curves from four pumping tests are shown in Figures 3a through 3d. If a curve approaches a straight line shortly after pumping starts, then an aquifer of uniform characteristics is suggested. Both curves of Figure 3a are of this type. The delay time before the

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[1] *Storage coefficient* is the volume of water in storage released from a column of aquifer with unit cross section under a unit decline in head. The quantity is dimensionless.

[2] *Transmissivity* is the volume of water which can pass through a unit width of aquifer, in unit time, under a unit hydraulic gradient, at standard temperature. Transmissivity has the dimensions $L^3/T/L = L^2/T$.

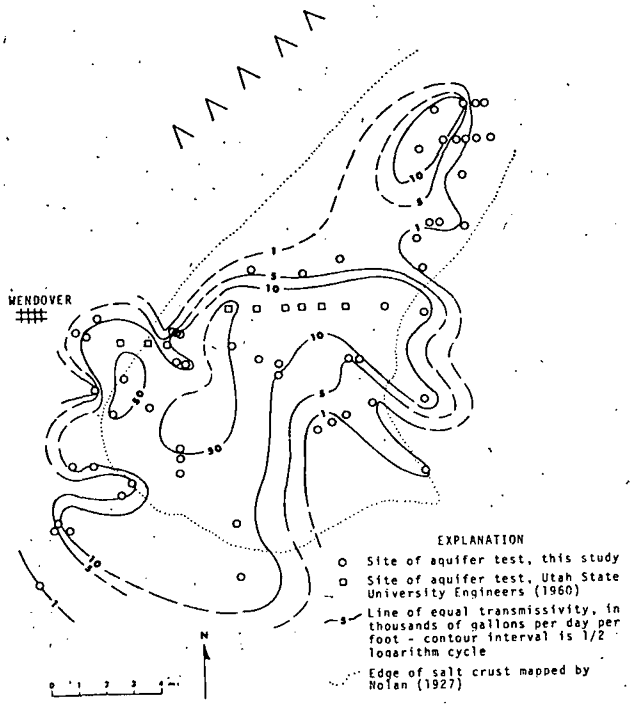

FIG. 4. Distribution of transmissivity within near-surface playa sediments.

curve becomes a straight line is partly a function of the distance from the pumping well. Thus the upper curve, which reflects conditions 68.9 feet from the pumping well, takes about 15 minutes to approach a straight line, whereas the lower curve, indicative of conditions at the pumping well, takes less than two minutes. A total of 19 of the 70 tests had a straight-line drawdown when plotted on semilogarithmic paper.

The curve in Figure 3b is typical of tests made near sources of large amounts of recharge brine. In this case, the recharge was from a production ditch. Nearby recharge was indicated in the results of 22 tests. The opposite reaction, that of a sharp downward inflection of the curve, would be produced by vertical barriers such as impervious clay-filled trenches which have been constructed along the sides of evaporation ponds. Only two tests suggested this type of reaction and only one of the two was near an impervious trench. Two of the wells near clay-filled trenches indicated recharge across the barrier instead of an impermeable boundary. This change from an impermeable boundary to a recharge boundary is probably due to fissuring of the clay subsequent to construction of the impervious trench.

Vertical leakage from overlying or underlying beds was shown in 18 of the tests. The drawdown curves are similar to those influenced by rapid recharge but lack the sharp inflections shown in the curve of Figure 3b. Leakage is easily detected when hydrographs of wells at varying distances from the pumped well are plotted as in Figure 3c. Note that the slope of the drawdown curve decreases as the distance from the pumped well increases (see discussion by Hantush, 1956).

If water table conditions exist during pumping, the transmissivity will decrease with time. This in turn will increase the rate of lowering of the brine surface. During the first 10 or 20 minutes, however, the tendency to lower the brine levels of pumping will be offset by a rapid vertical drainage of brine. The result is a peculiar double-S type of curve shown by the curve of Figure 3d. Only eight tests clearly indicated this type of reaction. If the tests were made for a period of several days, it would be reasonable to expect fully half the wells to eventually show a water-table response.

Figure 4 shows the general distribution of transmissivity over the salt flat. Although there are local variations, the presence of an elongate area having much higher transmissivities than the surrounding area is well documented. Note that the area of highest transmissivity coincides almost perfectly with the axis of the salt flat, where the salt is thickest. The transmissivity varies from less than 500 gpd/ft (gallons per day per foot) in wells near the edge of the salt crust to 50,000 gpd/ft or more in some wells near the center of the crust. The higher transmissivities are phenomenal for such a fine-grained aquifer—equal to the values expected for a very well-sorted medium to coarse sand! If all the brine flows through a section of aquifer 10 feet thick, the hydraulic conductivity (K) would be:

$$K = \frac{T}{b} = \frac{50{,}000\ \text{gpd/ft}}{10\ \text{ft}}$$

$$K = 5{,}000\ \text{gpd/ft}^2$$

where b = the saturated thickness of the aquifer. Some of the brine produced during a few of the tests actually moved through the highly permeable salt crust. Therefore, a few of the exceptionally high transmissivities measured are of doubtful value in evaluating the hydraulic properties of the playa sediments under the salt crust. On the other hand, moderately high transmissivities which cannot be attributed to flow through the salt crust are a result of the fissuring in the sediments, which was described earlier in this paper, and to the thin layers of fecal-pellet sand which are more abundant near the center of the salt-crust area. The areas of high transmissivity correspond closely with the areas of highest KCl concentrations shown in Figure 5.

Several sharp re-entrants of the lines of lower transmissivity occur around the edges of the salt crust (Fig. 4). These are interpreted as old drainage channels from which the lacustrine muds have been eroded and replaced by fluvial sediments which are less permeable than the fractured lacustrine sediments.

Selective aquifer tests: The highest hydraulic conductivities are localized in the uppermost part of the playa sediments. To estimate more closely the vertical distribution of hydraulic conductivity, bailer tests were run at successive depths during the sinking of nonperforated casing. Qualitative indications of the variations were obtained to a depth of about 20 feet. Permeable zones up to 2 feet thick were tested at depths of 5 to 18 feet; hydraulic conductivity ranged from 52 gpd/ft^2 to 409 gpd/ft^2 with the highest conductivities in the uppermost 10 feet of sediments. Even if the average hydraulic conductivity were represented by the largest value, the transmissivity for all the sediments to a depth of 20 feet would be only (20 ft × 409 gpd/ft^2) = 8,180 gpd/ft which is considerably less than 10,000 to 20,000 gpd/ft measured in the same area in 1965. A large part, perhaps as much as 70 percent, of the transmissivity in areas of thick salt crust may be due to brine flowing through the highly permeable salt.

Engineers from Utah State University (1962) reported the results of tests on a drill hole at mile post 15 (15 miles east of the Nevada-Utah boundary along U. S. Highway 40). Their first and second tests were made when the hole was 8.7 feet deep, the third after the hole was deepened to 12.2 feet. Transmissivity of the aquifer was 82,700 gpd/ft and 73,000 gpd/ft, respectively, at 8.7 feet, and 49,000 gpd/ft at 12.2 feet. The authors suggested that during the deepening of the well the upper part of the aquifer was "muddied up." A contributing factor may have been that the flow of brine from vertical gravity drainage was significant in the first test, somewhat less in the second test, but nearly exhausted in the third. Also, transmissivity may have declined with time owing to the dewatering of the uppermost part of the aquifer, which would effectively reduce the saturated thickness of the aquifer.

Additional selective aquifer tests were made during the construction of two pairs of wells in 1967. In the first pair, the transmissivity declined from 2,200 gpd/ft to 1,700 gpd/ft as the well was deepened from 10.2 feet to 21.0 feet. In the second pair the transmissivity was 40,900 gpd/ft at 10.3 ft, 41,700 gpd/ft at 15.0 feet, and 37,000 gpd/ft at 21.0 feet. These tests clearly indicate that little brine is obtained from depths greater than 10 feet, at least in the areas tested.

TABLE 3. Rates of Infiltration into Various Types of Playa Surface

Type of Surface	Infiltration Rate (ft/day)
Salt crust in abandoned ponds	3.5
Natural salt crust	2.5 –4.0
Gypsum sand	0.87–1.4
Clay-silt	0.4 –1.4

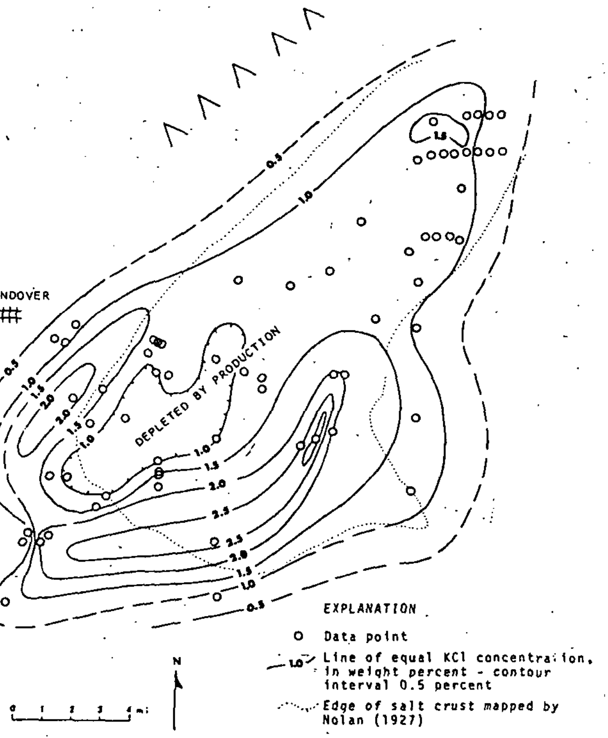

FIG. 5. Distribution of maximum KCl concentration within near-surface playa sediments during 1965–1967.

Hydrologic balance

Recharge: Recharge to the shallow aquifer is largely from local rainfall. However, during wet years, some surface runoff from higher areas, particularly to the west, enters the salt crust by overland flow and through production ditches. Some brackish water is imported from distant wells into the plant area for the flotation process. The excess water is discharged into the salt-crust area. Saline water is also pumped from deep wells and discharged at the surface.

The infiltration capacity of the playa surface was tested by ponding fresh water on the surface and measuring the rate at which the pond surface lowered. A total of 33 tests were made at eleven different locations with the results shown in Table 3. Despite the approximate nature of the tests, it is clear that all rains of reasonable intensity should be able to infiltrate almost immediately into the subsurface in all areas except possibly those having very moist clay at the surface.

TABLE 4. Composition of Brines in Near-Surface Sediments

Sample No.[2]	Constituents in Milligrams per Liter[1]											
	Ca	Mg	Na	Li	K	SO_4	Cl	Br	B	SiO_2	Sr	HCO_3
1	1,200	1,900	81,500	36	5,800	3,900	153,500	40	3	—	—	—
2	1,100	3,600	70,300	15	7,900	5,600	143,200	40	5	—	—	—
3	1,300	1,000	83,900	19	4,300	5,400	155,600	20	3	—	—	—
4	1,300	1,000	82,200	29	4,900	3,500	156,500	40	5	—	—	—
5	1,000	2,300	80,800	41	7,200	4,100	158,800	40	6	—	—	—
6	1,100	700	87,400	17	3,300	3,600	159,500	50	4	—	—	—
7	1,300	1,700	75,000	34	6,000	4,500	149,000	40	4	—	—	—
8	1,100	1,100	44,500	25	4,400	4,000	78,600	40	4	—	—	—
9	1,500	3,100	88,300	29	4,200	3,800	156,900	—	—	—	—	—
10[3]	1,200	2,100	81,000	22	3,200	4,000	148,900	—	—	—	—	—
11[3]	1,400	1,900	78,200	24	3,600	4,500	142,400	—	—	—	—	—
12	900	500	45,400	22	2,800	3,200	74,700	—	—	—	—	—
13	1,500	2,200	58,500	29	3,300	6,100	101,900	—	—	—	—	—
14[4]	1,130	1,430	96,800	41	2,660	3,680	159,000	—	—	10	57	42
15[5]	1,770	1,360	96,200	18	2,930	3,770	156,000	35	4.2	7.4	—	40

[1] Analyses by Kaiser Chemicals, Division of Kaiser Aluminum & Chemicals, San Leandro, California, except Nos. 14 and 15.

[2] Sample localities are shown on Figure 1.

[3] Insoluble clay in sample bottle.

[4] Composite sample of surface brine; analysis by Polzer and Roberson (1967, p. 116); pH at 25° C = 7.1; density at 20° C = 1.203 gm/cm³.

[5] Brine beside U. S. Highway 40, collected 4-6-58; analysis reported by Whitehead and Feth (1961, table 1) additional determinations: CO_3 = 0 mg/l; NO_3 = 4.8 mg/l, undetectable amounts of Fe, Mn, As, PO_4; Al = 1.8 mg/l; F = 1.3 mg/l; pH = 7.1; density at 20° C = 1.201 gm/cm³.

Hydrographs of observation wells indicate that rainfalls in excess of 0.1 inch during the summer and 0.05 inch during the winter contribute to recharge in the area of thick salt crust. Whenever brine levels are drawn down below the surface, most of the rainfall in excess of these amounts will become recharge. Examination of daily rainfall records shows that the excess rainfall available for recharge averages about 2.3 inches per year, or roughly half the total precipitation. An additional unknown amount of water is added by surface runoff into the playa during wet years; in dry years replenishment from this source may be zero.

Lateral inflow through the playa sediments adds some recharge, but the amount is insignificant compared with the brine extraction rates. Virgin conditions were such that fluid gradients of less than 0.1 foot per mile existed. Therefore, even if the area had a transmissivity of 100,000 gpd/ft, only 1,000 gpd would have moved through the sediments for each mile of the periphery of the salt flats. In fact, the actual hydraulic gradients may have been even smaller, because brine near the center of the playa has a greater density than brine near the edge of the salt crust. A column of brine 15 feet high with a density of 1.1 gm/cm could be balanced by a column of brine 13.75 feet high with a density of 1.2 gm/cm. Thus, if there were a 0.1 gm/cm decrease in brine density over a distance of five miles, an apparent hydraulic gradient of 0.25 ft/mile could be balanced by the density contrast, and no flow would occur. Of course, this example assumes an impermeable layer at a constant depth of 15 feet, which obviously is not true. Nonetheless, it is safe to assume that original hydraulic gradients were very small, and lateral brine flow must have been exceedingly slow. It is possible, however, that over a period of several thousand years, sufficient brine movement has occurred to contribute a significant amount of the dissolved solids now present in the central part of the playa.

A small component or recharge may be contributed through vertical movement of brine from below. Vertical hydraulic conductivity (K′) is extremely low, however, and hence, the small vertical gradients are not large enough to cause significant flow rates. Pumping test data analyzed by the leaky aquifer method indicate vertical hydraulic conductivities of 0.37 gpd/ft ² to 0.57 gpd/ft ² in the clays overlying the alluvial fan aquifer northwest of the salt crust, and about 0.58 gpd/ft ² in the center of the salt crust. Although these values for K′ are at best order-of-magnitude estimates, they show that the vertical hydraulic conductivities are on the order of 1/100 to 1/10,000 as large as the horizontal conductivities, and thus only minute quantities of brine can flow vertically under the observed hydraulic gradients. Osmotic effects, however, may reinforce natural hydraulic gradients in some areas so that upward flow may be more significant. Other possible effects of chemical osmosis are discussed in a subsequent section.

Two analyses of the tritium content of brine from shallow wells, collected in 1965, yielded 50 ± 3 and

405 ± 20 tritium units. The high tritium values suggest that recharge to the near-surface aquifers is primarily by recent rain rather than by slow lateral or vertical movement of brine.

Evaporation: The primary natural discharge from the playa is by evaporation, although a few salt-tolerant plants grow along the margins of the salt flat, and probably transpire a small amount of water. Evaporation of brine from the sediments is much slower than evaporation from the standing water which floods the playa during wet seasons. The evaporation rate decreases as the salinity of the brine increases (Turk, 1970). Late summer evaporation from ground water at depths of less than 1.0 foot below the salt crust doubtless is less than 10 percent of the evaporation from a free body of fresh water in the same location. In fact, under special conditions, water will condense rather than evaporate from the brine.

Hydrochemistry

Chemical composition of brine: Table 4 lists analyses of samples from 13 wells. Samples no. 1 through no. 8 were collected in 1965, the other five were collected in 1967. Each well was pumped a minimum of 10 minutes before a sample was taken. Analyses of two composite samples of the Bonneville brine by Polzer and Roberson (1967, p. 116) and by Whitehead and Feth (1961) are also included in the table.

Dissolved constituents of the shallow brine are mostly sodium and chloride. General distribution of KCl in brine of the playa sediments is shown in Figure 5. Data for this map were partial analyses of hundreds of samples collected during the period 1965–1967. Only the maximum concentration measured at each point during the period was used in constructing the map. This selection of the data tended to minimize the number of anomalously low values caused by dilution by antecedent rainfall. To be sure, the brine quality does fluctuate widely with time, and the sampling method could not entirely eliminate such variations; thus the map must be considered an approximation to the actual conditions. A second isoconcentration map, which has been prepared showing the distribution of $MgCl_2$, is not included here because of its similarity to the KCl map.

Three important features can be observed on Figure 5. First, the lines of isoconcentration are approximately parallel with the original border of the salt crust as mapped by Nolan (1927), which suggests that the brine quality, at least initially, was influenced by the same factors that determined the position of the salt crust. Second, KCl (and $MgCl_2$) concentrations tend to increase toward the center of the salt crust, except in the area where brine has been mined for many years. Third, the highest concentrations of KCl (and $MgCl_2$) are in the southeastern part of the salt crust where brine production has not depleted the reserves.

Inasmuch as the NaCl content is not significantly reduced in the areas where brine has been produced, we believe that the more valuable constituents have been selectively removed. Dissolution of the salt crust has maintained the NaCl content while the KCl-rich brine has been extracted. Brine analyses from various depths at 23 locations suggest that draining of KCl (and $MgCl_2$)-rich brines from near-surface sediments is most pronounced in areas of long-term production. There appears to be a near-surface enrichment of these salts in the virgin areas.

Brine quality varies widely with the seasons. During the abnormally wet spring of 1967 the dissolved solids concentration of the brine was low. As production of the brine proceeded and evaporation progressed, brine quality improved steadily—except when rainfall diluted it. Data obtained during the summer of 1965 indicate that heavy rainfall in May and June could reduce the KCl concentration of brine produced from normal values of about 2.0 percent by weight to less than 1.0 percent.

Origin of brine: The dissolved solids in the shallow brine were derived from two sources—those brought in by inward-draining waters and those leached from the clay-size material (Nolan, 1927, p. 40–42). Nolan believed that the second of these sources was more important. Our data appear to substantiate his conclusion, although the solids are more likely "drained" from pores, rather than "leached" from the solids.

Most of the solids now dissolved in the brine were deposited within the lacustrine sediments, either as chemical (or biochemical) precipitates or in "connate" water. Precipitation of aragonite and gypsum removed most of the $CaCO_3$ and $CaSO_4$ from the brines. Slow lateral subsurface inflow of water toward the lowest point of the playa has brought in more KCl and $MgCl_2$ derived from adjacent sediments.

The ultimate source of the dissolved solids is the same as the source of salts in Great Salt Lake. Feth (1959) postulated that bedded evaporites, deposited in pre-Lake Bonneville time, have contributed most of the salt load to the Bonneville Basin. High percentages of chloride indicate dissolution processes over a drainage area of sedimentary rocks of marine origin (B. F. Jones, 1966, p. 198). Marine sedimentary rocks west of the Bonneville area may be the ultimate source of most of the salts.

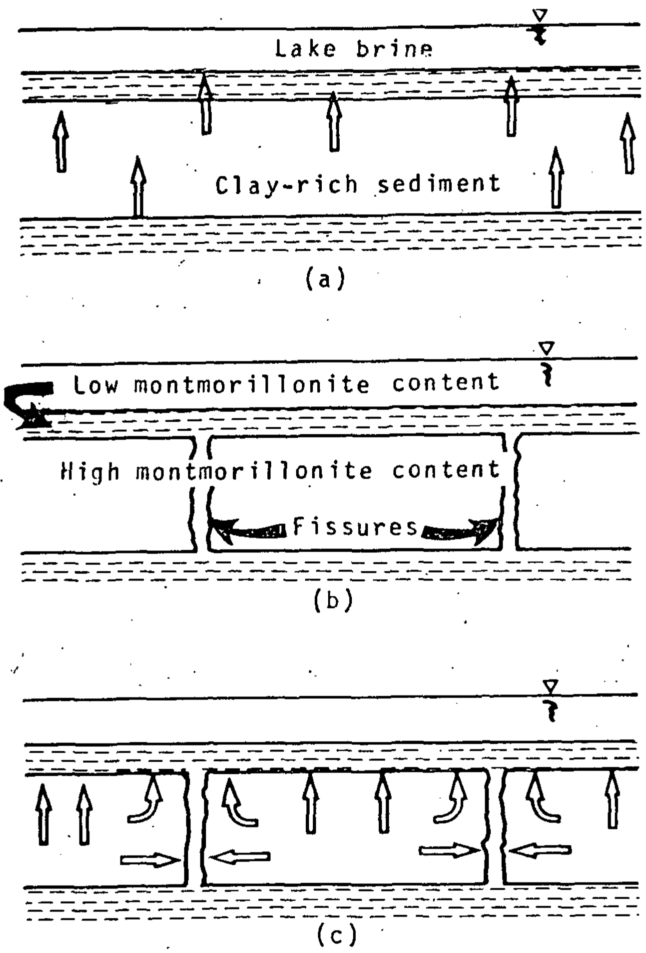

FIG. 6. Proposed model for the development of fissures by osmotic desiccation.

Interpretation of Data

Possible explanations of phenomena observed

Utah State University Engineers suggested that the fissures within the Bonneville sediments are desiccation cracks formed by subaerial drying of the lake bed sediments when they occupied the surface of the lake bed. There are two physical reasons why this explanation may be inadequate:

1. The cracks have vertical sides, not the typical V-shape expected in subaerial desiccation cracks.
2. The cracks are not filled with sediment from subsequent deposition.

Similar reasoning, plus the low coefficient of thermal expansion for salt, seems to preclude thermal expansion and contraction of the playa sediments as a cause of the cracking. Tectonic fracturing also can be ruled out because the more competent layers above and below layers of fissured clay are not disturbed.

The most reasonable explanation of the mechanism of fissuring is that the fine-grained sediments have undergone subaqueous shrinkage, by a process of osmosis or syneresis, or both. The following hypothesis is proposed as a possible model for the fissuring.

Osmotic desiccation

During the desiccation of Lake Bonneville, sediments were deposited in an increasingly saline environment; younger sediments contained saltier water than older sediments. A differential osmotic potential developed between the fine-grained sediments and the more saline brine in the lake, thus water could move by osmotic flow from the sediments to the brine (Fig. 6a). The loss of water caused the sediments to shrink and crack (Fig. 6b). If some of the fissures reached the top of the sediments, concentrated brine would enter the openings, changing the osmotic flow pattern but accelerating shrinkage (Fig. 6c). Jones has proposed the term *osmotic desiccation* to describe a similar dewatering process which occurs in Gulf of Mexico clays sandwiched between sand layers (P. H. Jones, oral communication, 1969). The term seems to be appropriate for the phenomenon observed at Bonneville.

Several investigators have demonstrated the effectiveness of clay as an osmotic membrane (e.g., McKelvey and Milne, 1962; von Engelhardt and Gaida, 1963; Young and Low, 1965). Other workers have concluded that osmotic flow, on a large scale, is responsible for the anomalously high hydrodynamic fluid potentials found in deep sedimentary basins (e.g., Berry, 1959; White, 1965; Jones, 1968, 1969).

Eardley (1966, p. 108) described fissures in the bottom sediments of Great Salt Lake which are probably of the same origin as those of the salt flats. Eardley stated that

> . . . Dehydration and compaction are occurring in the beds at the surface and down possibly 100 feet or more. A polygonal fissure system is an interesting consequence in certain places. From Rozell Point northwestward for several miles is a polygonal system in a blue-grey clay with each polygon 20 to 40 feet across, and the lines or lanes between polygons about 3 to 5 inches wide of slightly lighter, and almost soupy clay. . . . The lanes are believed to be the surface expression of fissures along which water from the dehydrating clays is emerging.

Osmotic desiccation cracks more likely would form in layers with a high clay mineral content than in layers with a low clay mineral content, because clay-rich sediments would provide the required semipermeable membrane and would subsequently undergo greater shrinkage than coarser sediments. Table 2 and Figure 2A show that the most fissured beds indeed contain more clay minerals than the less fissured beds; thus there is good correlation between clay mineral content and degree of fissuring.

Evidence supporting the hypothesis that the fissures formed by osmotic desiccation may be summed up as follows:

1. the cracks have straight sides and approximately constant width;
2. there is positive correlation between clay mineral content and degree of fracturing;
3. fissured beds are sandwiched between nonfissured, more competent beds, which precludes tectonic fracturing; and
4. the sediments were deposited in an increasingly saline environment, which would be conductive to osmotic flow.

The problem is not completely solved, however, because much of the evidence cited above can also be cited in support of another process, *syneresis,* which is the expulsion of part of the liquid from the solid in a liquid-solid suspension when internal forces of attraction (van der Waals' forces) are greater than internal forces of repulsion between particles of the solid.

Jüngst (1934) and White (1961) produced syneresis cracks in synthetic geologic environments and Mielenz and King (1955, p. 238) stated that syneresis is a widespread phenomenon whose development usually passes unrecognized because of its superficial resemblance to drying shrinkage. Nevertheless, meager field evidence indicates that syneresis cracks *do* exist in the geologic record. Rich (1951, p. 14) attributed cracks in shales and carbonate rocks of deep water origin to subaqueous shrinkage, although he did not specifically state that the shrinkage was a result of syneresis. Rooney et al. (1969) suggested that contraction as a result of syneresis may have produced the cracks in Alkali Lake, Oregon, which now contain vein-fillings of magadiite. White (1961, p. 566–567) reported that fossil syneresis cracks in the Tradewater Group in Illinois were evidence of subaqueous fissures of this type. He stated that syneresis cracks are most commonly found in thin beds of clay between harder strata such as fissile shale, limestone, or sandstone. The Bonneville cracks occur in just such a stratigraphic sequence.

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Regional Gravity Survey of the Northern Great Salt Lake Desert and Adjacent Areas in Utah, Nevada, and Idaho

bstract: From 1957 to 1961 a regional gravity urvey was made over the northern part of the reat Salt Lake Desert and adjacent areas in Utah, stern Nevada, and southeastern Idaho. A total f 1040 stations were taken over an area of about 00 square miles. The results were compiled as a ouguer gravity anomaly map with a contour terval of 2 mgal. The Bouguer values ranged from high of about —120 mgal over the outcrop areas a low of about —196 mgal over the alluvium-vered graben areas.

The gravity high over the Raft River Mountains pparently corresponds with the Raft River ountains anticline. A belt of gravity contours, ith a total relief of 15–20 mgal, extending for 40 iles between the Wildcat Hills and the southern rt of the Grouse Creek Mountains and beyond, is terpreted provisionally as caused partly by rupt thickening and/or downwarping of the cks of late Paleozoic age in a general northwest-rd direction, perhaps to form a foredeep (part of e Butte-Deep Creek trough of Steele, 1960) uth of the Raft River Mountains; however, this omaly could also be partly caused by overthrust-g.

Many Basin and Range faults, grabens, and rsts are indicated by the gravity data. In the rthern part of the surveyed area, Junction Valley, pper Raft River valley, and Curlew Valley are icated to be grabens. The Newfoundland Range, the northeastern part of the Great Salt Lake Desert, is a horst flanked by a graben on each side. The northwestern margin of the Great Salt Lake Desert comprises a complex pattern of Basin and Range fault blocks, large and small, that lie along a generally northward-trending belt or zone. The Silver Island-Pigeon fault block, which comprises the Silver Island Mountains, the Little Pigeon Mountains, and Pigeon Mountain, forms an elongate, arcuate horst that is flanked by a belt of grabens on the west (Pilot Valley, Lucin, and Grouse Creek grabens) and east (Wendover, Crater Island, Little Pigeon, and Pigeon grabens). A major fault zone is indicated along the east margin of the Pilot Range. The Pilot-Grouse Creek rift belt, at least 90 miles long, extends northward between Pilot Valley and the Upper Raft River valley and constitutes a major lineament in the earth's crust along which the graben blocks were displaced downward relative to the adjacent mountain blocks. In the southern part of the rift belt, the grabens are separated by blocks (the Lemay and Lucin horsts) that were probably downfaulted relative to the large mountain blocks, but became lodged at intermediate height; in the northern part, however, the blocks (for example the Junction Valley graben), apparently merely broke from the main crustal unit along this belt of weakness.

The indicated thickness of the valley fill of Cenozoic age in some of the graben areas ranges up to about 6000 feet.

CONTENTS

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INTRODUCTION

A regional gravity survey was conducted intermittently during the period June 1957 to March 1961 in the northern part of the Great Salt Lake Desert and adjacent areas in Utah, Nevada, and Idaho (Fig. 1) as part of a geophysical study of the eastern portion of the Basin and Range province (Cook and others, 1964). The area surveyed lies between lat. 40° 30′ N. and 42° 15′ N., and long. 112° 45′ W. and 114° 15′ W. The southern and central parts of the area are dominated by the flat expanse of the Great Salt Lake Desert. Rising abruptly from the desert flats are several northward- to northeastward-trending mountain ranges, some of which attain local relief of more than 5000 feet.

The Toana Range lies in the southwestern portion of the area (Pl. 1). The Pilot Range lies north and east of the portion of the Toana Mountains included in Plate 1. East of the Pilot Range and within the Great Salt Lake Desert are the Silver Island Mountains, Little Pigeon Mountains, Pigeon Mountain, Newfoundland Mountains, Terrace Mountains, and Hogup Mountains. In northwestern Utah the northward-trending Grouse Creek Mountains merge to the northeast with the eastward-trending Raft River Mountains, to the north (at Junction Creek) with the Albion Range, and to the northwest with the northwestward-trending Junction Mountain and Vipont Mountains and the Goose Creek Mountains. A belt of low hills extends northeastward from Pigeon Mountain to Curlew Valley (Pl. 1). Within the Idaho portion of the area surveyed, there are five mountain ranges, which from west to east are: the Middle Creek Southern Mountains, the Albion Range, the Cotterell Range, the Black Pine Mountains, and the Sublett Range. These ranges consist in part of dissected block mountains that trend northward to plunge beneath the Snake River lava plains (Anderson, 1931, p. 7).

The purpose of the gravity survey was to determine, insofar as possible, the regional lithology and structure of the underlying rocks. The authors hoped that the gravity survey would assist in delineating the main structural trends in the areas covered with valley fill, assist in the deciphering of the geologic history of the region, and provide basic structural information that would assist in ground-water and mineral-deposit studies currently being made in the region.

ACKNOWLEDGMENTS

The following students of the University of Utah assisted in the field and/or compilation work incident to the survey: W. W. Johnson, W. H. Dolan, J. D. Morgan, L. H. Rausher, R. P. Quitzau, J. I. Tanis, Elwood Hardman, and M. G. Best. A. J. Eardley and J. K. Costain gave helpful advice on several interpretive problems. Miss Emily J. Roberts assisted in the reduction of the data. The gravity contours in the Curley Valley area were taken from the detailed gravity map of the area made available by the Utah Southern Oil Company. Many

tives of the region assisted with the probms of logistics that arose in connection th the hazardous travel on the salt flats of the reat Salt Lake Desert and in other isolated rts of the surveyed area.

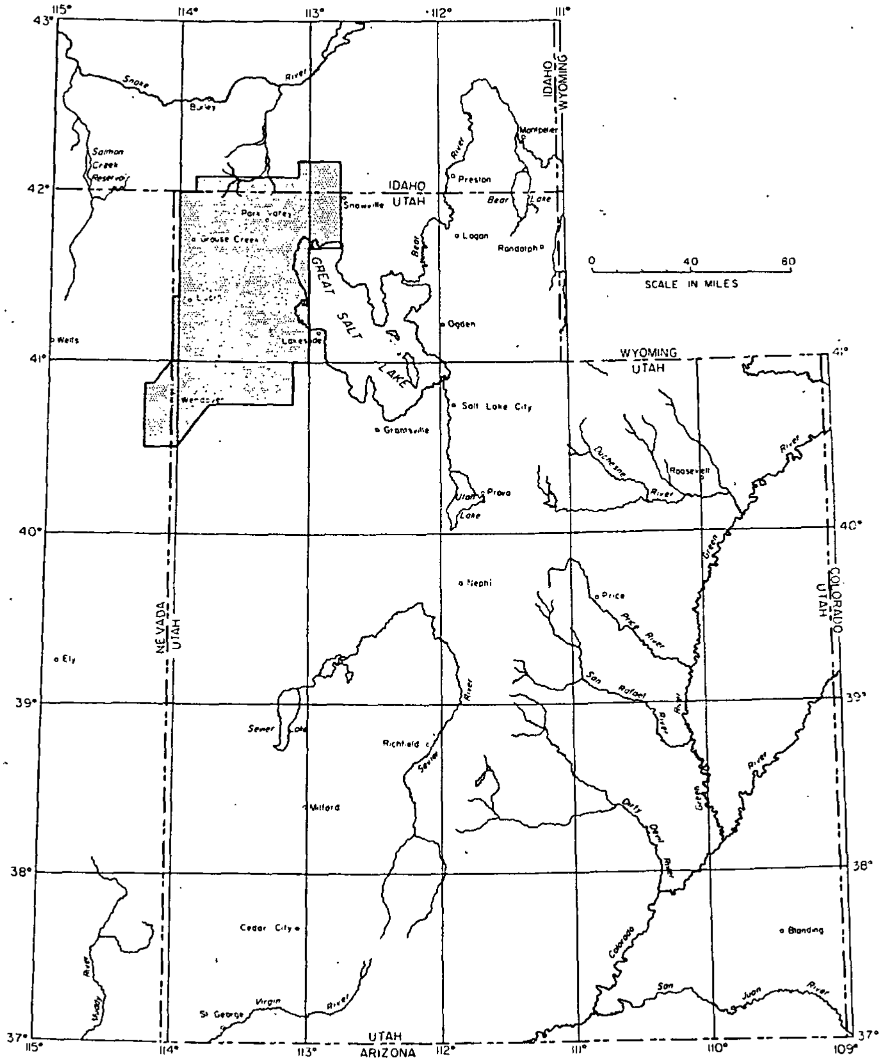

Figure 1. Index map of parts of Utah, Nevada, and Idaho showing area covered by regional gravity survey

Financial support for the survey and the reparation of the map and diagrams was given y the National Science Foundation under esearch Grants NSF-G2766, NSF-G5646, nd NSF-G13649, and the Utah Engineering xperiment Station of the University of Utah. he Worden gravimeter used on the survey was obtained through the Kennecott Copper Company Research Fund of the University of Utah; and the Carter gravimeter used was given to the Department of Geophysics of the University of Utah by the Carter Oil Company (now Jersey Standard Oil Company). The paper is based partly on material contained in M.S. theses submitted by M.O. Halverson and J. C. Stepp to the University of Utah during December 1960 and May 1961, respectively; however, several major revisions in the preliminary interpretations have been made in this paper. During the academic year 1959–1960, J. C. Stepp was the recipient of the Texaco Incorporated Graduate Fellowship in Geophysics, which helped make this study possible. The authors take the full responsibility for the interpretations given.

TECHNIQUES AND BACKGROUND DATA

Two gravimeters were used in taking the gravity field data. The major portion of the data were taken with a Worden gravimeter having a sensitivity of 0.1047 mgal per dial division; the remainder were taken with a Carter gravimeter having a sensitivity of 0.0744 mgal per dial division.

A total of 1040 gravity stations were established within the surveyed area of about 7000 square miles. In addition, the reduced and contoured Bouguer gravity data, which were supplied by the Utah Southern Oil Company for stations taken with a 1-mile grid spacing in the Curlew Valley area, were incorporated into this report. The area of this detailed survey is outlined on Plate 1, but the individual stations are not shown. Six permanent base stations were established and used throughout the field work and were tied together by alternating two or more pairs of gravimeter readings. One of these base stations was similarly tied to the U.S. Coast and Geodetic Survey pendulum station No. 49. The descriptions and gravity values of the base stations relative to this pendulum station are given in the appendix.

Elevation control was maintained by: (1) occupation of bench marks, points with spot elevations on both the U.S. Geological Survey and the U. S. Army topographic maps, and points surveyed by the Southern Pacific Railroad; (2) estimation from the surface elevation of both the Great Salt Lake and the salt flats; and (3) altimeter traverses. For the altimeter stations, where errors in elevation were greatest, the estimated maximum error was 30 feet (which corresponds to an error in gravity of 1.8 mgal) for several stations in areas of great topographic relief and 20 feet (1.2 mgal) for stations in areas of moderate topographic relief. For most of the stations in the surveyed area, however, which were either on the salt flats or in other areas of small topographic relief and excellent elevation control, the estimated maximum error in elevation was 10 feet (0.6 mgal).

Standard data reductions were made for the effects of instrument drift, latitude (U.S. Dept. of Commerce, Coast and Geodetic Survey, 1942), total elevation (combined free-air and Bouguer effects), and terrain. Isostatic corrections were not made. A combined elevation correction of 0.06 mgal/ft, which includes a Bouguer correction for an infinite slab of material of density 2.67 g/cc, was used. The datum for the final reductions was mean sea level. The reference for observed absolute gravity was the U.S. Coast and Geodetic Survey pendulum station No. 49, in Salt Lake City. The simple Bouguer anomaly value which omits the terrain correction was taken for this station as −183 mgal (Duerksen, 1949, p. 8); the value of all other gravity stations of the survey are in reference to this value.

Terrain corrections were made for 253 of the 1040 gravity stations with Hayford-Bowie zone charts of the U.S. Coast and Geodetic Survey corresponding to the Bullard modification (Swick, 1942). These 253 stations were carefully selected so that their corrections could be contoured on a map and the terrain corrections of the remaining stations could then be obtained by interpolation. Of the 253 terrain corrections, 78 were carried out to include zone K (18.8 km), 9 out to include zone L (28.8 km), and 176 out to include zone 0 (166.7 km) of the Hayford-Bowie zone charts. The terrain corrections for about 204 stations were less than 0.1 mgal and considered negligible. The largest correction was 6.8 mgal for a station on the north slope of the Raft River Mountains.

Plate 1 shows the terrain-corrected Bouguer anomalies, with a contour interval of 2 mgal, that were obtained by adding the terrain corrections to the simple Bouguer anomaly values. For stations in areas of moderate topographic relief, the estimated maximum error in the terrain-corrected Bouguer gravity value is 2.4 mgal considering the effect of errors from instrument reading and drift, latitude correction, elevation correction, and terrain correction. For most of the stations however, which were in areas of small topographic relief, the estimated maximum error in the terrain-corrected Bouguer gravity value is 1.0 mgal. This estimated error does not include possible additional errors that would occur if the average density of the rocks differs from the assumed value of 2.67 g/cc.

The rock densities may be divided into three groups of rock types: (1) sedimentary rocks of pre-Tertiary age that are dominantly mas-

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sive limestones and dolomites, (2) intrusive silicic igneous rocks, and (3) volcanic and clastic rocks of Tertiary and/or Quaternary age. Insofar as feasible, this grouping by approximate rock density has been used in the generalized geologic map in Plate 1. Because of the reconnaissance nature of the survey and the difficulty of access to most of the bedrock areas, no field samples were collected for making density determinations.

The average density of the rocks of groups (1) and (2) is estimated by the authors to be about 2.67 g/cc (Birch and others, 1942, p. 8–26), and that of the extrusive igneous rocks is estimated to be about 2.4–2.5 g/cc. Densities of the Tertiary and Quaternary clastic rocks are estimated to range from 1.9 g/cc near the surface to 2.3 g/cc at depth. When the rate of increase of density with depth and thickness of these clastic rocks is considered, an average density of about 2.1–2.2 g/cc is a reasonable estimate for gravity computations.

On the basis of the densities given heretofore, the average density contrast between the Tertiary and Quaternary clastic deposits and volcanic rocks that constitute the valley fill and the pre-Tertiary bedrock of the mountain ranges is estimated as 0.5 g/cc. This value compares favorably with the density contrast previously used in gravity studies in the Basin and Range province (Stewart, 1958, p. 1153). Some lateral change in density probably occurs in the rocks of Cenozoic age, especially those bordering the mountains; but too little is yet known of such subsurface features to warrant an attempt to include them in the calculations at present.

GRAVITY PROFILES AND INTERPRETIVE GEOLOGIC CROSS SECTIONS

The authors plotted seven gravity profiles (Figs. 2–8) over some of the major anomalies. Each gravity profile shows the observed gravity, the assumed regional gravity, and the residual gravity. Quantitative interpretation of the residual gravity profiles resulted in the interpretive geologic cross sections. The theoretical gravity, shown as points of "computed gravity" on the profiles, was computed for each cross section using a two-dimensional Jung graticule (Heiland, 1940, p. 154) and an assumed density contrast of 0.5 g/cc between the valley fill of Cenozoic (Tertiary and Quaternary) age and the bedrock of pre-Tertiary age. No end corrections were applied to the assumed two-dimensional structural models.

The geologic structures shown are considered a reasonable interpretation based on the available gravity and geologic control; however, they should not be considered a unique interpretation. If Basin and Range faulting causes the residual gravity, an equally good fit of the computed and residual gravity could be obtained by assuming a larger number of step faults than those actually shown. However, the relative movement of the Basin and Range fault blocks often occurs by large displacements along a relatively few parallel or subparallel master faults bordering the blocks rather than by small displacements along a large number of faults (*See* Osmond, 1960, p. 261). The angle of dip shown on the faults in the interpretive geologic cross sections is subject to much uncertainty; the authors believe that the values assigned are reasonable.

GENERAL GEOLOGY

The surveyed area lies within the Basin and Range province, about 60–70 miles west of the Wasatch Mountains. The Great Salt Lake Desert comprises salt flats and lake beds of former Lake Bonneville, alluvial material derived from the adjoining mountains, and wind-blown sand. Little is known of the subsurface lithology and structure of the Great Salt Lake Desert. Mapping of the outcrops in mountainous areas within and bordering the Great Salt Lake Desert indicates a complex structural and geologic history.

The surveyed area lies east of the central part of the Paleozoic Cordilleran miogeosyncline. The Antler orogenic belt, about 100 miles west of the surveyed area, was an area of positive movement and the locus of intense folding and faulting that began in Late Devonian or Early Mississippian time and continued probably into Cretaceous time (Roberts and others, 1958, Fig. 6, p. 2825, 2813).

Some places in the area record submergence and sedimentation of the miogeosynclinal type throughout the Paleozoic. In other places several Paleozoic periods are not represented. In the Silver Island Mountains, for example, approximately 23,600 feet of miogeosynclinal strata representing the entire Paleozoic are exposed (Schaeffer and Anderson, 1959; Schaeffer, 1960a). However, the northeast Nevada highland (Steele, 1959, p. 1105) was decidedly positive during the Paleozoic: rocks of Precambrian age are exposed, and the Paleozoic section includes only Cambrian and Pennsylvanian systems (Stokes, 1952; Bissell, 1962, Fig. 2, p. 1085). The Antler orogenic belt may be a westward extension of the highland (Steele, 1960, p. 91; Schaeffer, 1960c, p. 131). A second highland that came into existence in early Pennsylvanian time and persisted intermittently through late Cretaceous-early Tertiary time has been designated variously as the Sevier arch (Harris, 1959), the west-central Utah highlands (Steele, 1960) and the west-central Utah highland (Bissell, 1962, Fig. 2, p. 1085). In Permian and Pennsylvanian time, this highland separated the Ely basin of eastern Nevada and the Oquirrh basin of Utah and Idaho. Between this highland on the south, the Antler orogenic belt on the west, and the northeast Nevada highland on the northwest, was the Butte-Deep Creek trough that joined the Oquirrh basin on the east (Steele, 1960, p. 91–92). This was possibly the depocenter of thousands of feet of bioclastic and lithoclastic sediments and may have extended into Idaho to the Sublett Range area (Bissell, 1962, p. 1107). However, the limits of the Permian basins are still unknown (Bissell, 1962, p. 1107).

In the Newfoundland Mountains area, R. E. Paddock (1956, M. S. thesis, Univ. Utah)[4] found evidence of intermittent positive tendencies during late Paleozoic time. Rocks of early Devonian, Mississippian, and Pennsylvanian ages are absent; rocks of Permian age lie upon rocks of late Devonian age with profound unconformity. In the Silver Island Mountains, Schaeffer (1960c, p. 143) found evidence of positive tendencies that demonstrate the unstable nature of the Cordilleran miogeosyncline in pre-Middle Devonian time. Additional evidence of positive tendencies toward the end of the Devonian time has been found in parts of the Oquirrh and Stansbury ranges, which lie east and southeast, respectively, of the mapped area (Osmond, 1960, p. 256; Rigby, 1958; 1959; Stokes and Arnold, 1958).

According to Bissell (1962, p. 1108), a large area of the Great Basin stood above ocean waters during latest Permian time.

No Mesozoic rocks have been found in the surveyed area. During the Nevadan or Laramide orogenies, the area was subjected to forces that resulted in northward-trending folds (Osmond, 1960, p. 258).

In the Raft River Mountains, all the sedimentary rocks of early (?) Paleozoic age are interpreted as thrust over the rocks of Precambrian age (Felix, 1956). Because the eastward extent of this thrusting is unknown in the region south of the northeast Nevada highland (Misch, 1960, Pl. 1, p. 18, 36), the possibility that the frontal segment of the thrust may break through the overlying sequence should be recognized. Steele (1960, p. 94) believes, however, that any large-scale displacements during the Antler orogeny in the areas of the northeast Nevada highland and the Butte-Deep Creek trough occurred during late Devonian-Mississippian time and, therefore, that the Pennsylvanian-Permian sedimentary rocks were basically unaffected.

Block faulting that resulted in the uplift and tilting of the ranges was initiated probably in early Oligocene time (Nolan, 1943, p. 141) and has continued to the present. The prefaulted surface of the region probably had some topographic relief. Sedimentary rocks of Tertiary age in the region are generally exposed along the margins of the valleys and in most cases dip toward the valleys. Subsequent erosion on the uplifted blocks and accompanying accumulation of sediments in the basins has resulted in the present physiographic features of the area. Igneous intrusive and extrusive rocks are widespread in the area.

The regional gravity patterns reflect existing density contrasts in the earth's crust that are the final product of the past tectonic activity in the region. The largest gravity anomalies in the surveyed region are believed to be caused by Basin and Range structures. In particular, the density contrast between the pre-Tertiary rocks of the relatively uplifted mountain blocks and the sediments and volcanic rocks of Cenozoic age overlying the relatively downfaulted valley blocks give rise to the striking gravity gradients observed in the area. These steep gradients, with a total relief of up to 35 mgal in a horizontal distance of 2 miles, occur along one or both sides of the fault blocks. Also, some gravity anomalies are believed to be related to earlier orogenies, such as the Laramide orogeny, and to the possible thickening and/or downwarping of rocks of Paleozoic age within the eastern part of the Paleozoic Cordilleran miogeosyncline. Density contrasts within the Precambrian basement complex also may cause some of the observed anomalies.

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[4] Paddock's geologic map of the Newfoundland Mountains was used in compiling the geology shown on Plate 1.

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REGIONAL GRAVITY PATTERNS OF LARAMIDE AND OLDER STRUCTURES

General Statement

Some of the highest gravity values in the surveyed area occur in the outcrop areas of Precambrian and Paleozoic age that apparently correspond with the positive Paleozoic elements in the region. In general, the gravity values in these bedrock areas decrease in a westward or southwestward direction. There are exceptions, however, and the manner in which the maximum gravity values vary from range to range indicates that the positive Paleozoic elements may comprise complex lithologic or structural units.

The broad regional decrease of gravity in a westward and southwestward direction is probably caused largely by the greater thickness of rocks of late Paleozoic age in the western part of the surveyed area. Because the average elevation of this western part is somewhat greater than that of the eastern part, the broad regional gravity decrease is probably partly an isostatic effect (Mabey, 1960).

Raft River Mountains Area

The geology of parts of the Raft River Mountains and adjoining areas had been described by Piper (1923), Anderson (1931), Peterson (1942), and Felix (1956). The eastern part of the Raft River Mountains has been mapped in detail by Felix (1956). Except as otherwise noted, all geology presented here is taken from Felix's (1956) report and accompanying geologic map.

The stratigraphy consists of: metaquartzites and schists of the Harrison(?) Series of Precambrian age; sedimentary and metamorphic rocks of early (?) Paleozoic age and Pennsylvanian age, especially near the north, east, and south flanks of the mountains; and some sedimentary rocks of Tertiary age. The Raft River Mountains form a great asymmetrical doubly plunging anticline. The axis of the anticline trends nearly east and is arcuate and slightly concave southward. As mapped by Felix (1956) all the rocks of Paleozoic age are thrust over rocks of Precambrian age.

A gravity high with a closure of about 6 mgal overlies the Raft River Mountains. In the west-central part of the mountains, the westward-trending axis of the gravity nose on the west side of the gravity high is inferred (on the basis of sparse gravity data) to coincide approximately with the axis of the Raft River anticline. Thus the gravity high with 6 mgal closure is probably related principally to the density contrasts incident to the anticlinal structure of the range.

The gravity contours, with a steep gradient and a total relief varying between about 15 and 20 mgal, extend southwestward from the east side of the Raft River valley and nearly encircle the Raft River Mountains except for the east and northeast sides. On the basis of the geology as it is known at present, the most probable cause of the gravity gradient is the contrast between the presumed higher density of Felix's unit A of the Harrison(?) Series (a dark mica schist at least 1300 feet thick) and the rocks forming the core of the eastern part of the Raft River Mountains, on the one hand, and the lower density of the overlying (and younger rocks of Felix's unit B of the Harrison (?) Series (a series of metaquartzites and quartz schists alternating with mica schists, chlorite-mica schists and minor phyllites, with a total thickness of at least several thousand feet) on the other. The exposures of unit A, which has been designated the Harrison Formation by Stokes (1963), are confined principally to the eastern crestal part of the Raft River Range, whereas those of unit B, which has been designated the Dove Creek Formation by Stokes (1963), are widespread throughout the western crestal part of the range, as well as along the northern and southern flanks of the range (Bronson Stringham, 1963, oral communication; Felix, 1956, Fig. 14).

In addition to these density contrasts, the gravity gradient along the north and south flanks of the Raft River Range is probably caused partly by the density contrasts between the rocks of Paleozoic age exposed along these flanks and the Precambrian rocks forming the central part of the Raft River Mountains uplift. An abrupt thickening and/or downwarping of the rocks of Paleozoic age is indicated along the north and south flanks of the range. South of Stanrod, the observed gravity gradient along the northwestern flank of the Raft River Mountains corresponds partly with the location of the overthrust fault, as mapped by Felix. Here the beds of early(?) Paleozoic and Pennsylvanian age are thrust over beds of Precambrian age. The Paleozoic rocks are probably less dense than the Precambrian rocks, and the gravity gradient is therefore in the right direction to be caused by the thrust fault.

Silver Island Mountains Area

The northeastward-trending Silver Island Mountains consist of three main segments: the southern segment, known as the Leppy Range, the central segment, known as Silver Island, and the northern segment, known as Crater Island. The designation "island" is used because these rock masses rise in isolation from the Great Salt Lake Desert. The range has a maximum relief of about 3000 feet above the surrounding salt flats.

The geology of the Silver Island Mountains has been mapped in detail by Schaeffer (1960a; 1960b; 1960c) and Anderson (1960a; 1960b). The lithology is dominantly limestones and dolomites of Paleozoic age. Tertiary extrusive igneous rocks of silicic composition are extensive in the Leppy Range. Four major intrusive stocks of silicic composition occur on Crater Island (Anderson, 1960a) and two stocks of silicic composition occur in the northern part of the Leppy Range (Schaeffer, 1960a). Structural relationships throughout the Silver Island Mountains are complex, and only the salient structural features are shown on Plate 1. Two major reverse faults, of possible Laramide age, exist. According to Schaeffer (1960c, p. 134), the first, the Lost Canyon fault, which extends across the northwestern part of Silver Island, dips north and has a vertical displacement of 6000 feet. The second, the Jenkins Peak fault, in the central part of Silver Island, dips 55° NW. and has an estimated vertical displacement of 2000 feet. Several major normal faults of Basin and Range type, with vertical displacements of as much as 8000 feet, exist. Normal border faults, or fault zones, of the Basin and Range type were postulated by Schaeffer (1960c, p. 137–139) along both the southeastern margin of the Silver Island Mountains and the northwestern margin of the Leppy Range. Anderson (1960b, p. 126) found no evidence to support the existence of a border fault along the west side of Crater Island but suggested the possibility of a border fault along the east side of Crater Island.

SILVER ISLAND MOUNTAINS GRAVITY HIGH: The complexity of the gravity pattern over and adjacent to the Silver Island Mountains reflects the corresponding complexity of the geologic structure in this area. A gravity maximum, herein designated the Silver Island Mountains gravity high, lies over the entire Silver Island Mountains area. Because of sparse gravity data over or near the outcrop area of Silver Island, the gravity contours over the interior of Silver Island must be largely inferred. Although the total closure over the Silver Island Mountains is only about 18 mgal, the gravity relief between the Silver Island Mountains and the bottoms of the surrounding valleys and salt flats of the Great Salt Lake Desert is as much as 56 mgal.

The steep gravity gradient along the north side of Silver Island indicates that this side of the island is probably bounded by an eastward-trending Basin and Range fault or fault zone that crosses the Silver Island Mountains in the Donner Reed Pass area. In the area east of Donner Reed Pass, the northward-dipping Lost Canyon fault, a major reverse fault (Schaeffer, 1960c, p. 134), probably extends eastward beneath the salt flats; a large amount of subsequent normal movement of the Basin and Range type may have occurred along this old line of weakness and produced the largest portion of the observed gradient here. In the area west of Donner Reed Pass, an eastward-trending and northward-dipping normal fault of Basin and Range type is interpreted from the gravity data to lie about 2 miles north of, and parallel to, the Lost Canyon fault, and to extend westward into Pilot Valley as a major structural lineament. This faulting also may have occurred along old lines of weakness parallel to the Lost Canyon fault.

Part of the large gravity gradient along the west margin of Silver Island is caused by the Silver Island fault, a westward-dipping normal fault with an estimated stratigraphic displacement of 8000 feet which places rocks of Permian age against rocks of Ordovician age (Schaeffer, 1960c, p. 135).

EAST SILVER ISLAND FAULT ZONE: The pronounced gravity gradient along the northeastward-trending contours along the southeastern margin of the Leppy Range and Silver Island indicates a major fault zone, herein designated the East Silver Island fault zone. The gradient —and hence the fault zone—extends at least 22 miles from the Wendover area to a point several miles west of Floating Island. The sinuous trend of the contours in the areas of West Peninsula Peak and East Peninsula Peak conforms with the curving trend of the mountain front and thus indicates that the edge of the land mass in this region is probably fault-controlled. Although the gravity contours bend around and trend abruptly southeastward in the area west of Floating Island, the East Silver Island fault zone probably continues northeastward between Silver Island and Floating Island, then bends north, and continues northward to join the east Pigeon fault zone. In the area between Silver Island and Floating Island, Schaeffer (1960c, p. 139) postulates a great normal fault downthrown on the east with a maximum displacement of approximately 5000 feet. A branch of that part of the East Silver Island fault zone lying about 3 miles northeast of the east tip of Silver Island probably extends south-southeastward to form the east margin of the Floating Island fault-block spur. A border fault along the northeastern margin of Crater Island suggested by Anderson (1960b, p. 126), is supported by the gravity data, although the data here are too sparse to determine the location of the fault. If this fault were considered to be a northwestward branch of the East Silver Island fault zone, an alternate interpretation to that shown on Plate 1 would be to extend the contours associated with the East Silver Island fault zone along the east margin of Crater Island. On the basis of the available gravity data, however, this alternative is considered less probable than that shown.

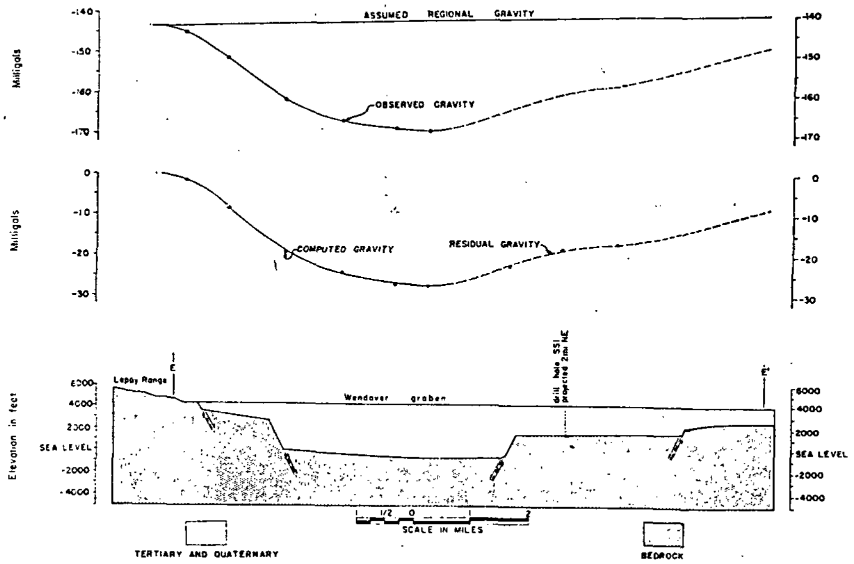

Figure 6. Gravity and interpretive geologic cross section along profile E-E′ across Wendover graben. Assumed density contrast is 0.5 g/cc.

An interpretive geologic section across the East Silver Island fault zone along southeastward-trending profile E–E′ (Fig. 6) shows a total vertical displacement of 4200 feet that occurs along the fault zone over two normal faults only.

LEPPY FAULT ZONE: In the area west and southwest of Wendover, the gravity contours associated with the East Silver Island fault zone bend abruptly westward and then northward and indicate another major fault, herein designated the Leppy fault, which forms the southwestern margin of the Leppy Range. The extent of the Leppy fault to the northwest and its relationship to the northward-trending Toana Range cannot be determined from the available gravity data. This fault may possibly coincide with a transverse fault in the Wendover area (Nolan, 1943, p. 177).

WENDOVER GRABEN: The northeastward-trending gravity low lying southeast of the Leppy Range and Silver Island indicates a graben, herein designated the Wendover graben. The graben is probably more than 35 miles long and at least 10 miles in maximum width. The southeastern margin of the graben lies beneath the salt flats of the Great Salt Lake Desert with no apparent topographic expression. Because of the dangerously wet conditions of the salt flats in this area, no data are available, and the continuity of the steep gravity gradient along the east-central margin of the gravity low can only be inferred.

Along southeastward-trending profile E–E′

(Fig. 6) across the Wendover graben, the indicated thickness of the valley fill in the deepest part of the graben is about 5000 feet. On the east end of the profile, the gravity control was projected from the trend of the contours about 5 miles northeast of the profile; the indicated structure and thickness of the valley fill here is therefore tentative. The Shell Oil Company Salduro No. 1 well (SS1 of Pl. 1) is shown on profile E–E′ as projected about 2 miles northeastward along the trend of the gravity contours. The well penetrated rocks of Cenozoic age to a depth of 2740 feet, where intrusive igneous rocks of granitic type were penetrated. No formations of Paleozoic age were penetrated. The depth from the surface to the granite was used for geologic control in compiling the cross section in Figure 6. The sparse gravity data in this area give no definite indication of the intrusion or its possible horizontal extent.

The gravity low with a closure of at least 8 mgal lying southwest of the south tip of the Leppy Range indicates a downfaulted block that probably forms a westward arm of the Wendover graben. Thus the Wendover graben is probably a great complex downfaulted block that comprises several subsidiary blocks that, in turn, have been dislocated by different amounts relative to each other.

FLOATING ISLAND FAULT-BLOCK SPUR: The gravity data indicate that Floating Island is a large fault-block spur of the Silver Island Mountains. The block was downfaulted relative to the Silver Island Mountains, but not as much as the adjacent graben blocks. As indicated by the striking gravity high over it, the Floating Island fault-block spur is about 7½ miles long in a north direction and about 3 miles wide. The block is probably bounded by normal faults on the west, northwest, east, and south margins. The crest of the southern part of the Floating Island fault-block spur is buried under probably less than 1000 feet of valley fill; only the highest part of the spur protrudes above the surface of the mud and salt flats to form Floating Island.

CRATER ISLAND GRABEN: The elongate northward-trending gravity low, with a closure of at least 8 mgal, which extends from the area east of Floating Island to the area east of Pigeon Mountain, appears to be caused by either two separate grabens or a single great downfaulted block with two subsidiary grabens. Although the contours shown on Plate 1 were drawn to infer the latter interpretation, the data are insufficient to prove it. The southern and northern grabens are herein designated the Crater Island and Pigeon grabens, respectively.

The Crater Island graben is about 26 miles long and several miles wide. On profile D–D‴, which crosses the north end of the Crater Island graben, the indicated thickness of the valley fill is about 3100 feet. The graben is apparently bounded on the west by the East Silver Island fault zone.

Pilot Range Area

The Pilot Range is composed of sedimentary rocks of Paleozoic age and intrusive and extrusive igneous rocks; the geologic structure is complex due to faulting and regional folding (Butler, 1920, Blue, 1962). The northern part of the range displays several stages of faulting and folding, with possible gravity sliding phenomena in the area of Patterson Pass (D. M. Blue, 1960, M.S. thesis, Univ. Utah, p. 71–72). This part of the Pilot Range is bounded on the east and west by normal faults that have truncated the older structures and caused tilting of the range to the east. A large body of silicic composition has intruded the beds of Paleozoic age from Patterson Pass northward for a distance of about 6 miles along the northeastern side of the range.

A major eastward-striking fault transverse to the trend of the Pilot Range, originally noted by Butler (1920), has been confirmed by D. M. Blue (1960, M.S. thesis, Univ. Utah[7]) in the area of Patterson Pass. On the basis of stratigraphic relationships, Blue estimates a minimum of 9500 feet of right-lateral horizontal displacement along the Patterson Pass fault and an equally large vertical displacement with relative downthrow on the north (D. M. Blue, 1960, M.S. thesis, Univ. Utah p. 66). This transverse fault is apparently terminated on the east by a northwestward-trending normal fault downthrown on the southwest, which was mapped by Blue near the eastern end of Patterson Pass. This northwestward-trending fault is probably part of the Lemay fault zone.

PILOT VALLEY GRABEN: Pilot Valley extends northward along the east side of the Pilot Range for about two-thirds of the length of the range; it is bounded on the north by an ancient Lake Bonneville tombolo that now occurs as a ridge that joins Lemay Mountain to the Pilot Range near Patterson Pass. The

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[7] Blue's geologic map of the northern part of the Pilot Range was used in compiling the geology shown on Plate 1.

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Broad Regional Features

Several broad regional structural features, larger than the individual Basin and Range fault blocks, form part of the over-all tectonic framework of the region.

EASTERN RIFT BELT: The series of great elongate grabens that includes the Wendover, Crater Island, and Pigeon grabens forms an arcuate rift belt or zone about 80 miles long that is herein designated the eastern rift belt. This belt is apparently continuous except for the interruption by the Floating Island fault-block spur. The Little Pigeon graben lies within or adjacent to this belt.

PILOT-GROUSE CREEK RIFT BELT: The belt of grabens trending northward at least 90 miles between the southern part of Pilot Valley and the Upper Raft River valley is herein designated the Pilot-Grouse Creek rift belt. The belt includes the Pilot Valley, Lucin, Grouse Creek, Junction Valley, and Upper Raft River valley grabens. The belt is probably a major lineament in the earth's crust along which great fault blocks were displaced downward relative to the adjacent mountain blocks. In the southern part of the belt, the grabens are separated by blocks (the Lemay and Lucin horsts) that were probably downfaulted relative to the adjoining large mountain blocks but became lodged at intermediate height. In the northern part of the belt, the grabens, for example the Junction Valley graben, apparently merely broke from the main crustal unit along the belt of weakness; this graben has *en echelon* pattern with respect to the extreme northern part of the Grouse Creek graben.

SILVER ISLAND-PIGEON FAULT BLOCK: The gravity data indicate that the Silver Island Mountains, the Little Pigeon Mountains, and Pigeon Mountain probably form a great arcuate fault block nearly 50 miles long that is herein designated the Silver Island-Pigeon fault block. The Silver Island-Pigeon fault block, although principally a horst, is not a rigid structural unit. It is cut transversely in places by Laramide or older faults, as well as by Basin and Range faults. The block is flanked on the west by the southern part of the Pilot-Grouse Creek rift belt and on the east by the eastern rift belt.

COMPARISON WITH FEATURES OF THE WASATCH RANGE AREA: The broad regional features related to Basin and Range faulting just described are similar, although apparently on a smaller scale, to those found in the Wasatch Range area. The Silver Island-Pigeon fault block is a horst: so is the Wasatch Range (Gilbert, 1928, p. 62; Cook and Berg, 1956; 1958). Also the eastern and Pilot-Grouse Creek rift belts are somewhat similar to (1) the Wasatch structural trough, which is a major rift extending about 160 miles in length and 16 miles in maximum width along the western side of the Wasatch Range (Cook and Berg, 1961; Cook, 1962, p. 325), and (2) the back valleys (series of grabens) along the eastern side of the northern and central parts of the Wasatch Range (Gilbert, 1928, p. 62). This striking structural correspondence between the Silver Island-Pigeon fault-block area on the west and the Wasatch horst area on the east is considered tectonically significant because the former feature is at the west margin of the great Bonneville basin and the latter is at the east margin.

SUMMARY AND CONCLUSIONS

The gravity survey has resulted in the discovery of new geologic structures and the delineation and correlation of known geologic structures. Most of the structures indicated by the gravity data are of the Basin and Range type, but some are of Laramide age or older.

The gravity high over the Raft River Mountains apparently corresponds with the Raft River Mountains anticline. The belt of steep-gradient gravity contours that nearly encircles the Raft River Mountains is probably partly caused by (1) the contrast between the presumed higher density of the Precambrian Harrison Formation and the rocks forming the core of the eastern part of the Raft River Mountains on the one hand and the lower density of the overlying Precambrian Dove Creek Formation on the other; and (2) the density contrast between the rocks of Paleozoic age exposed along the flanks of the range and the Precambrian rocks forming the central part of the Raft River Mountains uplift. The southwestward extension of this anomaly belt in the area between the Wildcat Hills and the southern part of the Grouse Creek Mountains and beyond is interpreted provisionally as partly caused by a rapid thickening and/or downwarping of the rocks of upper Paleozoic age in a general northwestward direction, perhaps to form a foredeep (part of the Butte-Deep Creek trough of Steele, 1960) between this belt and the area now occupied by the Raft River Mountains. Additional geologic investigation is warranted to appraise the possibility that overthrusting may partly cause this anomaly belt.

Many Basin and Range faults, grabens, and horsts were delineated by the gravity survey. Many of the valleys are shown to be grabens in which the thickness of the valley fill probably exceeds several thousand feet. The northwestern margin of the Great Salt Lake Desert comprises a complex pattern of Basin and Range fault blocks, large and small, that lie along a generally northward-trending zone.

The thickness of the great expanse of valley fill comprising sediments and possibly volcanic rocks of Cenozoic age in the northern part of the Great Salt Lake Desert is largely predicated upon the Basin and Range structures. The valley fill is thick—6000 feet or more—in the graben areas and thin—several hundred feet—in buried horst areas.

The significant southwestward-trending regional gravity gradients of up to 3 mgal per mile over the bedrock areas west of the Silver Island Mountains and Little Pigeon Mountains are probably caused by an increase in the thickness of the sedimentary rocks of Paleozoic age toward the southwest. However, part of this regional trend may be due to regional isostatic adjustments with increasing average elevations of the surface to the southwest.

The authors believe that the results of the gravity survey indicate key areas where future geologic work can be concentrated to help decipher the many unsolved structural and stratigraphic problems that still remain in this region.

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Contribution No. 39, Dept. Geophysics, University of Utah, Salt Lake City, Utah

APPENDIX: DESCRIPTION AND VALUES OF GRAVITY BASE STATIONS

Descriptions of the important gravity base stations used in the survey are given here. The accompanying value is the gravity reading at the designated station, in milligals, relative to the reading at the U.S. Coast and Geodetic Survey pendulum station 49, located in the Temple Grounds in Salt Lake City and established in 1894 (Duerkson, 1949, p. 8). The value shown is the algebraic quantity "base station value minus pendulum station value."

	Milligals
Knolls base station	+40.10

Station is at Knolls, Utah, approximately 195 feet west of west end of westernmost building of Knolls Motel. Station is 75 feet south of center line of U.S. highway 40, and approximately 4 feet above U.S. Coast and Geodetic Survey bench mark Y–77.

Wendover base station +29.53

Station is along Western Pacific Railroad 1.4 miles east of Utah-Nevada state line at Wendover. Station is 66 feet west of mile post 807 on Western Pacific Railroad and 6 feet north of northernmost rail of railroad, in center of a single-lane secondary road leading from U.S. highway 40 to railroad.

Lucin base station +30.65

Station is at U.S. Coast and Geodetic Survey bench mark A–59, which is 50 feet south of intersection of roads from towns of Park Valley and Grouse Creek, Utah, about 1 mile north of Lucin. Station is 50 feet north of center line of abandoned Southern Pacific Railroad bed.

Snowville base station +62.40

Station is in Snowville, Utah, 10 feet west of U.S. Coast and Geodetic Survey bench mark C–89, which is 9 feet north of southwest corner of Nelson Cafe. Station is 4 feet below elevation of bench mark. Station is just off Plate 1 east of Curlew Valley.

Park Valley base station + 8.66

Station is 5 feet due south of grounds of Park Valley Church of Jesus Christ of Latter Day Saints, Park Valley, Utah. Station is on center of sidewalk and is due south of church steeple. (Park Valley, Utah, quadrangle)

Groome base station (SL–128) +99.58

Station is at base of north rail on railroad tie due south of mile post 711 on Southern Pacific Railroad about 0.1–0.2 mile west of Groome, Utah, siding.

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[9] Because this excellent map became available after the generalized geology shown on Plate 1 was compiled on final tracings in preparation for publication, data from it were used for making corrections and additions on Plate 1 principally in areas where previously published data were not available.

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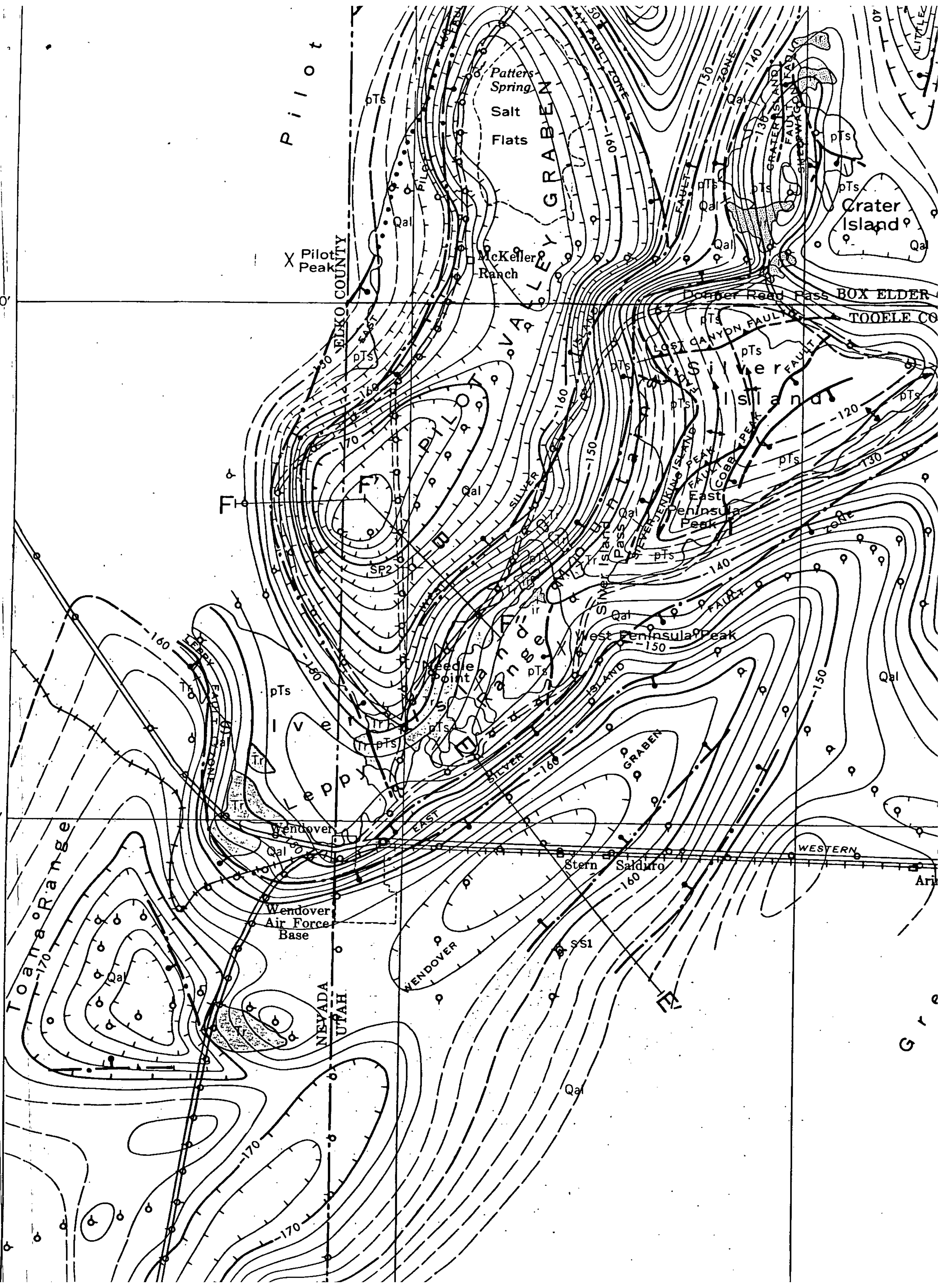
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Leppy Hills
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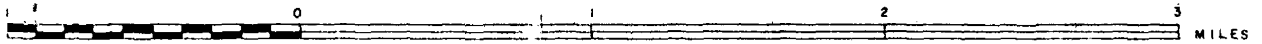
PLATE IC

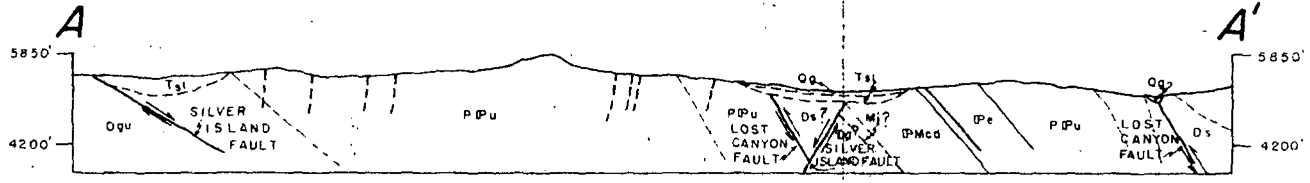
GEOLOGIC CROSS SECTIONS OF THE CENTRAL AND SOUTHERN SILVER ISLAND MOUNTAINS, UTAH AND NEVADA

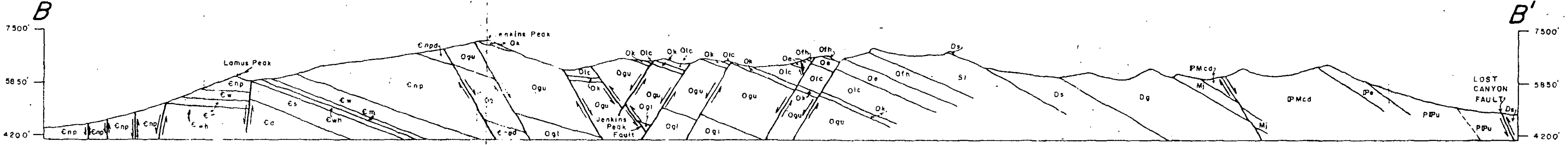
Geology by Frederick E. Schaeffer

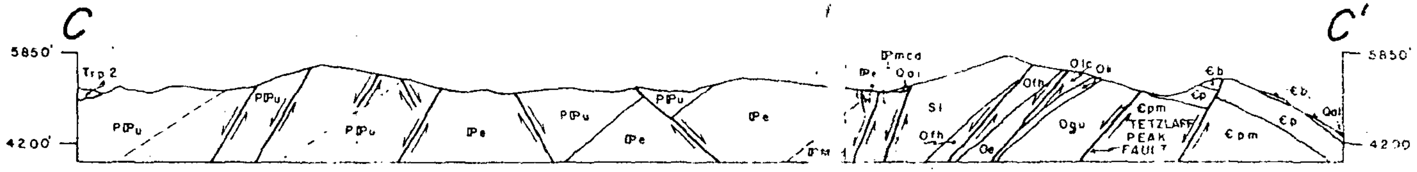
Drafted by Peggy A. Schaeffer

1960









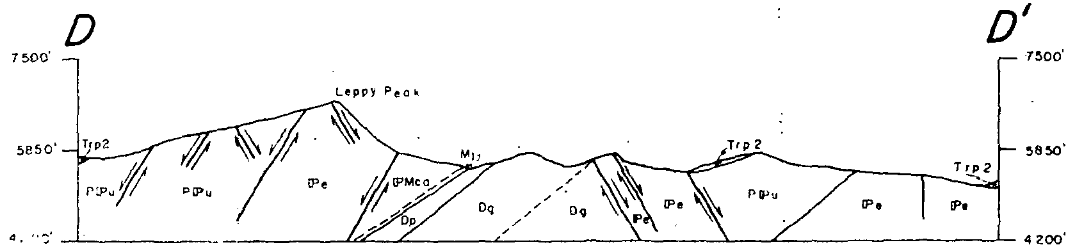


PLATE 1C

GEOLOGIC CROSS SECTIONS OF THE CENTRAL AND SOUTHERN SILVER ISLAND MOUNTAINS, UTAH AND NEVADA

Geology by Frederick E. Schaeffer

Drafted by Peggy A. Schaeffer

1960