The Hydrology and Mineralogy of Deep Springs Lake Inyo County, California

GEOLOGICAL SURVEY PROFESSIONAL PAPER 502-A

The Hydrology and Mineralogy of Deep Springs Lake Inyo County, California

By BLAIR F. JONES

C L O S E D - B A S I N   I N V E S T I G A T I O N S

GEOLOGICAL SURVEY PROFESSIONAL PAPER 502-A

A study of the relation between hydrologic factors and saline mineralogy for a small playa in the western Great Basin

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CLOSED-BASIN INVESTIGATIONS

THE HYDROLOGY AND MINERALOGY OF DEEP SPRINGS LAKE, INYO COUNTY, CALIFORNIA

By Blair F. Jones

ABSTRACT

Deep Springs Lake is a small intermittent saline lake at an altitude of about 5,000 feet within a completely closed basin in northern Inyo County, Calif. The basin is a relatively recently formed graben, in which the most active faulting occurred on the east, within or near the White-Inyo Mountains horst block.

The Deep Springs Valley was formerly occupied by a large lake which may have covered more than one-half of the valley floor as recently as 1,000 years B.P. Recent normal faulting has cut off outflow, tilted old shorelines, controlled the location of the playa in the southeast corner of the valley, and provided a locus for a major belt of springs which flow into the lake.

In the bedrock geology of the Deep Springs Valley the feature of greatest importance to the valley hydrology is the contact between granodioritic igneous rocks to the north and east and ancient sedimentary strata lying to the south and west. An intermediate belt of contact metamorphic rocks and associated mineralized zones is also very influential. The alluvium of the northern part of the valley is composed largely of fan materials which are derived from the Wyman-Crooked Creeks system, the major perennial drainage to the valley. The northern valley alluvium contains abundant zones of high permeability related to old channel deposits. Lacustrine materials can be recognized in the central part of the valley 5 miles from the present lake area and may contain impermeable confining layers composed of very compact clays or carbonate muds. The present playa area has three types of lacustrine deposits listed from the center of the playa outward or with increasing depth as (1) the saline crusts, (2) carbonate muds, and (3) fine-grained well-sorted silts and sands.

Annual precipitation in the Deep Springs Valley averages from about 5 inches on the valley floor to 15 inches on the basin divide. Distribution is intermediate between the summer minimum pattern of the California coast and the more uniform annual distribution of the Great Basin. The evaporation rate of the saline waters of Deep Springs Lake may be as low as 30 inches per year, but the mean is probably almost 52 inches per year.

The hydrography of the Deep Springs Valley is dominated by the perennial streams of the Wyman-Crooked Creeks and Birch-Antelope Creeks systems, and the three major spring groups around Deep Springs Lake—the Corral, Buckhorn, and bog-mound springs. Flow patterns suggest (a) that the Birch-Antelope Creeks system is connected with the western bog-mound springs, (b) and that the Wyman-Crooked Creeks system is connected with the central and eastern bog-mound springs, and, to a lesser extent, (c) with the Corral and Buckhorn Springs along the prominent fault zone east of the lake. The Corral Springs discharge ground water from igneous rocks along the fault zone north of the lake; the Buckhorn Springs discharge largely ground water from sedimentary and contact metamorphic rocks intercepted by the fault zone south of the lake.

Discharge estimates indicate that inflow to Deep Springs Lake from the Wyman-Crooked Creeks, Corral Springs, or Buckhorn Springs systems is roughly equal. Seasonal fluctuation in the discharge of the springs with change in storm patterns suggests substantial contribution from sources other than the main valley ground-water reservoir.

Deep Springs Lake is generally pan shaped, and there is a roughly linear logarithmic relation between its area and volume. The lake demonstrates greater areal variation and shorter response time than anything considered in the closed lake compilation of Langbein (1961) and is thus most similar to the numerous small playas so common in the Great Basin. In spite of uncertainties in volume computations and the homogeneity of surface waters at Deep Springs Lake, a relation between volume and tonnage of dissolved salts was established. This relation follows, perhaps fortuitously in detail, the "ideal cycle" for a closed lake proposed by Langbein (1961). Continuity in the relations of water volume and total concentration is interrupted by the precipitation of sodium sulfate. Loss of dissolved salts by entrapment in marginal sediments during lake-level recession was quantitatively verified at Deep Springs Lake; lacustrine muds contain as much as 30 percent readily water-soluble material, and some contain over 80 percent of material of precipitate origin (that is, including alkaline earth carbonates).

Analytical studies at Deep Springs Lake have indicated that most presently available data on carbonate in concentrated natural waters can be considered only semiquantitative at best. Of all the constituents for which saline waters are normally analyzed, only sulfate and chloride are consistently reliable.

Chemical variation in more than 125 water samples from the Deep Springs Valley is most effectively shown by trilinear plotting. The chemical compositions of most waters considered inflow to the lake fall within fairly well-defined limits characteristic of the distinct hydrologic units. The general trend is toward increase of the percentage of alkali, sulfate, and chloride simultaneously with increase in total dissolved solids. The major factors affecting the differences in chemical composition of inflow units are gross environmental lithology and processes taking place within and between each unit. The chief processes involved are the precipitation of alkaline earth carbonate, sulfate reduction, and additional solution of alkali material. Cation exchange is considered a minor factor in affecting compositional change. The chemical data generally support the flow patterns deduced for the valley. Variations in water composition may also be related to seasonal fluctuations in discharge, temperature and biologic activity, but original lithologic control appears more important than any climatic factor. A shift of source areas due to the seasonal change of storm patterns may affect the composition of waters significantly, as indicated by the northern Buckhorn Springs. Substantial compositional changes take place between inflow springs and the lake itself; the major cause of these changes is evaporation associated with alkaline earth carbonate precipitation, and re-solution of capillary salts.

Ephemeral inflow to the lake very definitely reflects the immediate area from which it is derived, and tends to retain its identity within the playa area for some time.

Waters from Deep Springs Lake fall into relatively narrow ranges in percentage composition depending on the immediate environment. Surface waters generally have less than 10 percent total alkalinity, or about 5 percent below the average values for intercrustal brine. Sulfate and chloride vary about 20 equivalent percent, and potassium is stable at a relatively high level for natural waters. In surface waters, sulfate to chloride ratios fluctuate significantly with changes in stage and show the dominance of sulfate in the readily soluble salts of the saltpan.

The equation derived by Langbein (1961) for the calculation of salinity from hydrographic data for permanent closed lakes cannot be applied to a playa such as Deep Springs Lake because of the importance of solution kinetics, relative salt solubilities, thickness of exposed crust, and residual brines.

More than 15 different species of saline and carbonate minerals occur at Deep Springs Lake. There are three distinct patterns in their distribution: (1) areal zoning of precipitate minerals, (2) layer variation in saline crusts, and (3) local variation in capillary efflorescences. The areal sequence of mineral zones from playa margin to center is calcite and (or) aragonite, dolomite, gaylussite, thenardite, and burkeite. Dolomite, in part primary, is the most abundant mineral of the carbonate muds. Thenardite is the dominant mineral of the saline crusts. Burkeite is the most characteristic mineral of the central salt pan. The layer sequence of crustal minerals in ascending order is nahcolite, thenardite, burkeite, trona, and halite. Capillary efflorescent crusts are typical of the western part of the playa. These deposits are dominated by halite and thenardite, but variation is great and is indicative of the anion composition of ephemeral inflow. The silicate mineralogy is dominated by quartz and 10-A clay species, particularly $2M_1$ mica, but chlorite, talc, and poorly defined expandable lattice types were also identified. The major effects of the saline environment on the clays appear to be the "degradation" (probably oxidation and magnesium loss) in chlorite and possibly the reorganization of montmorillonite.

The data on the hydrochemistry and saline mineralogy indicate that Deep Springs Lake is basically a sodium sulfocarbonate system. The most characteristic mineral of the central lake crusts is burkeite. In general, the formation of burkeite is favored by relatively high temperature, salinity, and low pCO_2. Comparison of mineral assemblages at Deep Springs Lake with experimental data on burkeite stability (Jones, 1962) with results of work by Teeple (1929) on the quinary Na_2CO_3–$NaHCO_3$–Na_2SO_4–$NaCl$–H_2O suggests that the salines reflect their original sequence of precipitation but trend toward local equilibrium with associated brines.

INTRODUCTION

The intermontane basins of the Western States provide numerous examples of complete hydrologic units where the end product is a brackish or saline lake, a saltpan, or a playa. In such closed basins, both the derivation and disposition of solutes should be accountable in terms of processes taking place within a single confined area. Langbein (1961) has attempted to explain the variable salinity of closed lakes in terms of their geometric properties and hydrologic environment. A further step would be to relate the specific chemical nature of such waters and associated evaporite deposits to the hydrologic processes operative in each basin.

The waters of closed basins are often classified on the basis of the dominant anion, such as carbonate, sulfate, or chloride (Hutchinson, 1957). In the Great Basin, two main types are represented; the Bonneville basin, which is high in chloride, and the Lahontan basin, where alkali carbonates predominate. Further south in the Mojave Desert region, sulfate becomes a major component of the waters. These differences have been attributed to the source of solutes and conditions of evaporation, but outside of Great Salt Lake and Death Valley (Hunt, 1960) no detailed study has been made.

After a reconnaissance of the western Great Basin early in 1959, the Deep Springs Valley was chosen as a workable and reasonably representative area for the study of a closed basin of the sulfocarbonate type. Fieldwork was done during late spring and early fall from October 1959 to September 1961. The chemical variations in the waters could not be related either to hydrologic environment or source of solutes alone but to an intimate interrelation of both as evinced by the deposits which the waters leave behind.

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work was performed by Miss Shirley L. Rettig, Quality of Water Branch laboratory, Washington, D.C.

GEOGRAPHY

The Deep Springs Valley is a small intermontane basin (alt. about 5,000 ft) in northern Inyo County, Calif., which has a total drainage area of 200 square miles. To the west and northwest of this northeast-trending elongate basin are the White Mountains (altitudes up to 11,278 ft. in the valley drainage area). Most of the drainage to the valley is derived from precipitation in these mountains. The valley rim to the east and south rises gradually to a maximum altitude of 8,924 feet at the southernmost end of the valley, then merges with the White Mountains to form the Inyo Mountains. Beyond the mountains west of the Deep Springs Valley lies the Owens Valley and the high Sierra Nevada, which intercepts much precipitation that would otherwise fall within the Deep Springs Valley drainage area. To the east of the Deep Springs Valley lies the much larger and much lower Eureka Valley; to the north of both is the Fish Lake Valley. An index map of the Deep Springs Valley and vicinity is shown in figure 1.

The Deep Springs Valley floor is about 15 miles long, has a maximum width of 5 miles, and encloses an area of approximately 47 square miles. Deep Springs Lake lies in the southeast corner of the valley. The saltpan and surrounding playa cover an area of about 5 square miles. The alluvial floor of the valley has an average gradient of about 50 feet per mile from the north end to the playa, where the average altitude is 4,920 feet; 4,917 feet is the lowest measured points in the lowest measured point. Altitudes of several points in the Deep Springs Lake area are given in figure 2.

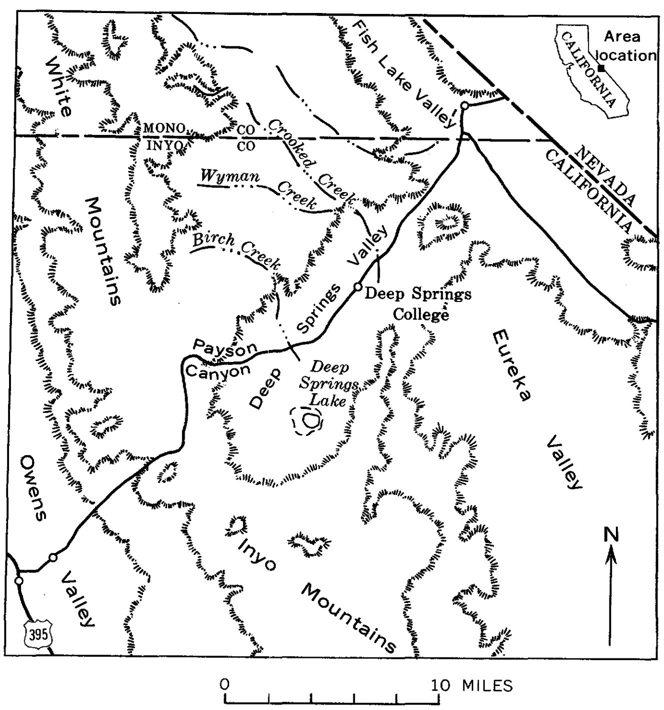

FIGURE 1.—Map of the Deep Springs Valley and vicinity, California, showing the major topographic and drainage features. The Owens Valley occupies the southwest corner of the map.

The playa which constitutes Deep Springs Lake may be readily subdivided on the basis of distinctive surficial features (fig. 3). The east-central one-third of the playa is lowest and is marked by a thick porous varicolored salt crust broken into irregular polygonal units up to 100 feet across. Surrounding the large polygonal units is a narrow zone of thinner but coherent salt crust; this zone widens to the south and gradually passes westward into an area of mud flats, which widen markedly to the north. The western third of the playa area is 2 to 5 feet higher than the saltpan, is conspicuously marked by very shallow alluvial channels, and is normally covered by a thin efflorescent salt crust.

The most distinctive feature of the playa is a low levee up to 2 feet high enclosing the northeastern third of the area. This levee was built of material dredged from an adjacent trough and was part of an attempt in the early 1920's to create a solar evaporation pond for the commercial production of potash salts (Palmer, 1922). The levee and trough seem to have little direct effect on the hydrography of the playa.

Southeast of Deep Springs Lake are several distinctive topographic features associated with recent faulting. This area has been described in detail by Miller (1928, p. 520–523). Here the mountain front stands behind the lake like a great wall, rising more than 2,000 feet in less than a mile. The base of the front is marked by distinct fault scarps for about 2 miles (fig. 4). The alluvial fans along the steepest part of the front are transected by a pronounced fault trough for a distance of about half a mile. This trough is terminated at both ends by sheer fault scarps up to 50 feet in height at the base of the mountain. The trough floor ranges from 100 to 300 feet in width and averages about 25 feet in depth. Lakeward from the main trough are some broader sags and ridges which trend roughly parallel, but somewhat arcuate, to the major fault zone, and decrease westward in mean altitude. The crest of the outermost ridge is slightly higher than the floor of the main trough. These sags and ridges are apparently the result of slumping in alluvial material associated with movement along the major fault.

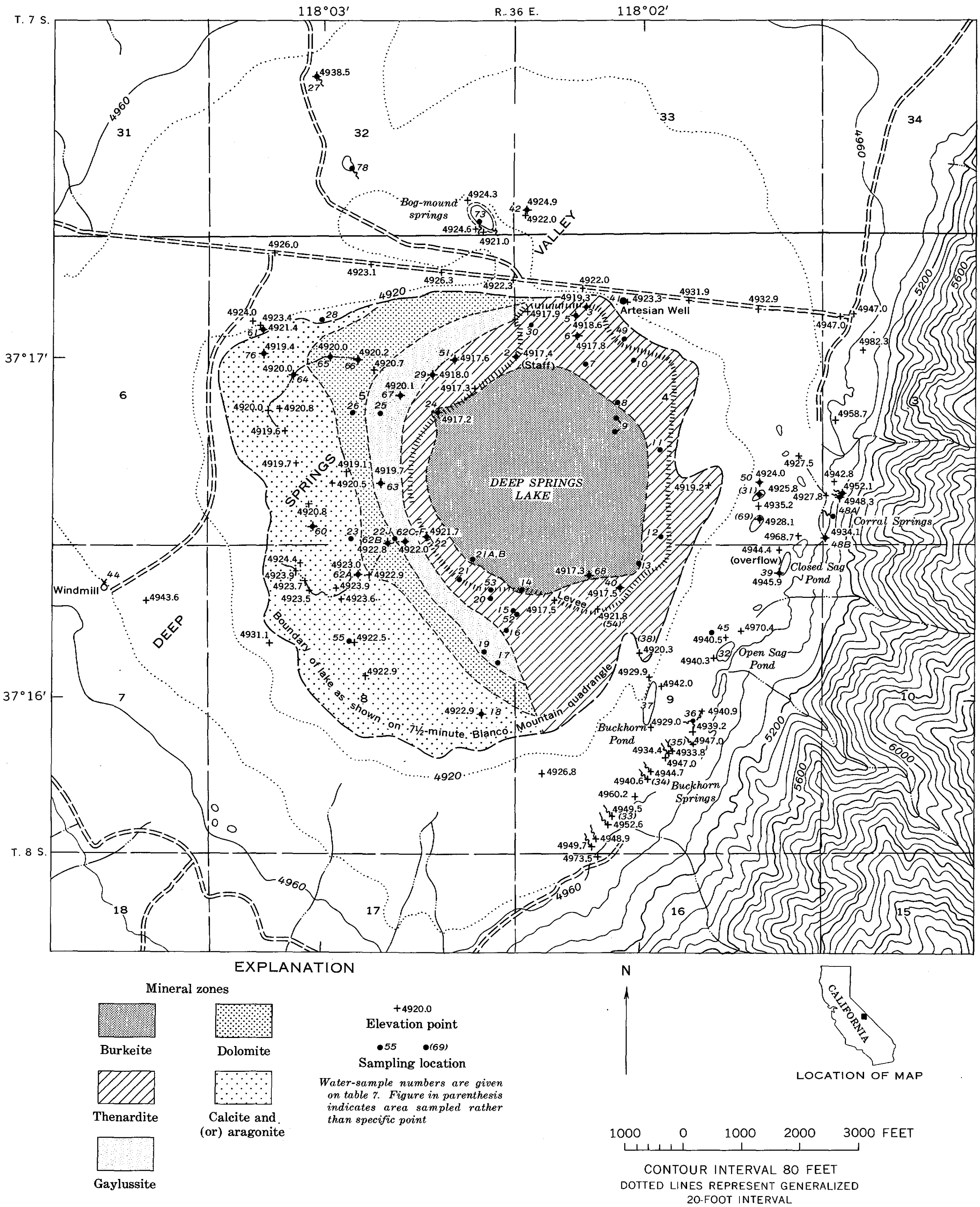

Figure 2.—Map of the Deep Springs Lake area including altitudes of key points and sample localities. Dashed lines delineate the general saline mineral zoning in the surficial deposits of the playa.

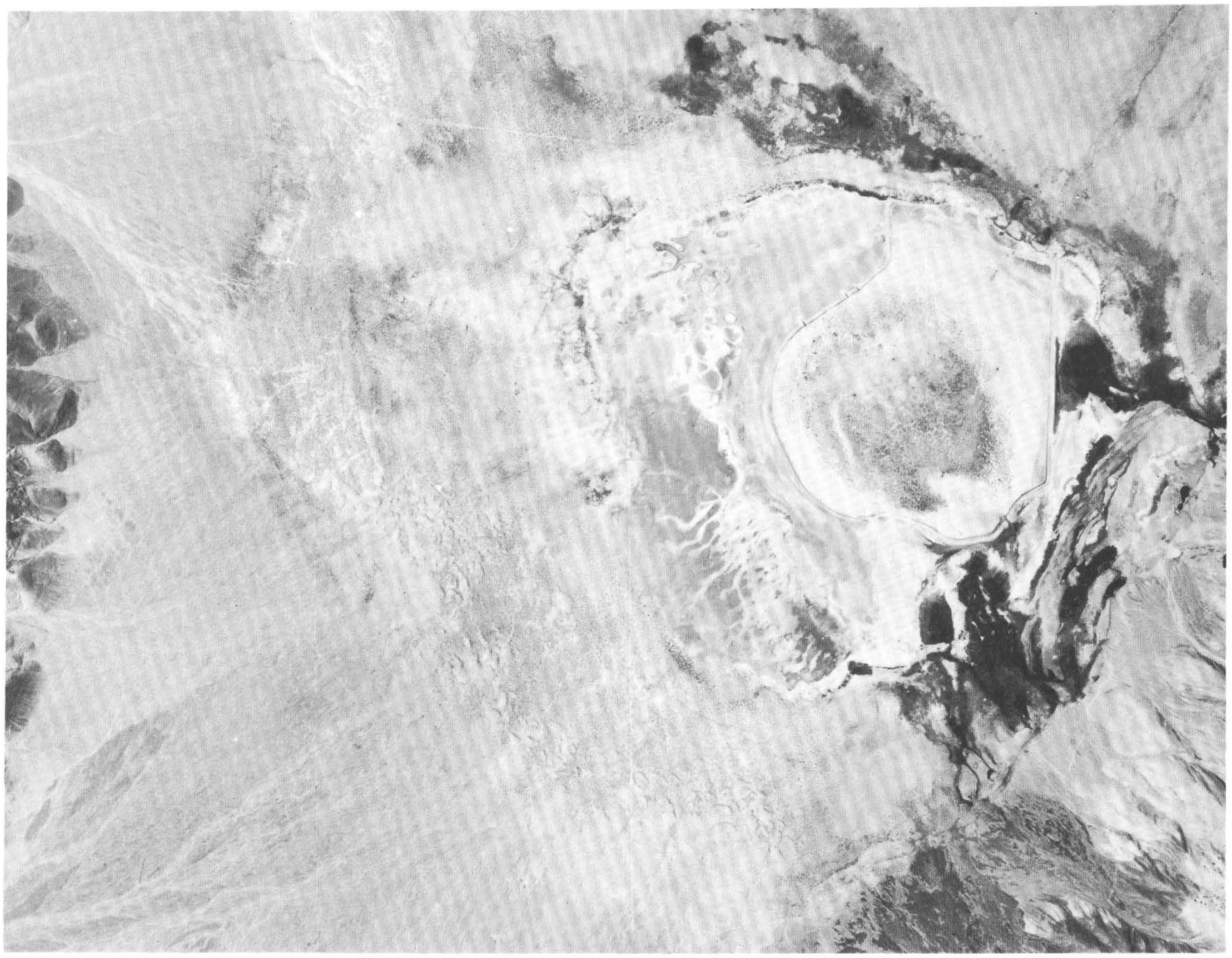

FIGURE 3.—Aerial photograph of Deep Springs Lake and surrounding area at dry stage, July 1947. The faint border around the central saltpan is the levee mound. The straight part on the east side is about three-fourths of a mile long. The bog-mound springs are the dark spots north of the playa area. The Corral Springs and their inflow area are immediately east of the central saltpan, and the Buckhorn Springs are to the south. Two sag ponds may be seen in the fault trough between the two spring areas along the fault zone. Note the shoreline traces around the saltpan and the ancient lake shorelines at the south and west (left) extremes of the photograph. The sand dunes southwest of the playa are also conspicuous.

North of the playa is an elongate broad shallow depression that trends west-northwest and is separated in part from the playa by a low ridge. The floor of this depression slopes gently east and is marked by several circular or oval springs. On the ridge to the north and to the west and south of the playa, dune topography is common. Near the playa the dunal ridges are generally circular or oval, about 3 feet high, and are separated by broad smooth flats. The size and relief of the dunes increase with distance away from the playa to a maximum of about 500 feet long and 15 feet high about half a mile southwest. Most of the dunes are now stabilized by vegetation, but many show signs of recent activity.

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Evidence indicates that at least the south half of the Deep Springs Valley was formerly occupied by a large lake. Although, as pointed out by Hubbs and Miller (1948, p. 89), shore features are not readily apparent in the valley, such features can easily be identified on aerial photographs of the region (fig. 3). Shoreline benches can be definitely delineated and examined at only a few places in the field, however. Such features can be distinguished with certainty up to 5,200 feet in altitude west of the playa or nearly 300 feet above the present lake level. However, faulting and a marked lowering of the southern part of the valley floor since the lake stood at the 5,100-foot level are suggested by the coincidence in altitude of

FIGURE 4.—View looking southeast from the north side of Deep Springs Lake, showing the prominent fault scarps in alluvial fans east of the playa.

fairly prominent benches on the southwest side of the valley and the top of prominent fault scarps on alluvial fans east of the lake. This hypothesis is further supported by the gradual transgression of strand lines by contours on the broad pediment formed by coalesced alluvial fans southwest of the playa.

As pointed out by Miller (1928, p. 525), the most likely outlet for the lake was through Soldiers Pass into the much lower Eureka Valley. Soldiers Pass at an altitude of about 5,500 feet is the lowest point on the east rim of the valley and consists of a deep canyon and a well-defined channel. Lustig (1962, p. 25) described remnants of a carbonate-cemented pebble conglomerate in the pass and noted that the beds dip to the west, which suggests local tilting. He also noted that C. A. Nelson (oral commun.) had mapped the large fault east of the lake as continuing through the pass and suggested a displacement of at least 200 feet on the basis of the difference in altitude from one side of the mountain mass to the other. Lustig (1962, p. 26) also described stream gravels which indicate uplifts of at least 200 feet in the northeast corner of the valley. Such tectonic movement makes it impossible to ascertain the maximum depth of the former lake. Miller's (1928, p. 525) estimate of 400 to 500 feet is conjectural, and it is questionable whether or not the lake level was ever sufficiently high to reach the far north end of the valley.

Vegetation in the Deep Springs Valley is sparse. The playa itself supports no vegetation. Saltgrass and some pickleweed grow around the edge of the playa. Many species of grasses grow in the depression area north of the lake and around the spring channels east of the lake. Marsh grasses, reeds, and cattails grow around the ponds and spring orifices. The vegetation on the remaining valley floor is composed of shrubs dominated by rabbitbush *(Chrysothamnus sp.)*, saltbush *(Atriplex canescens?)*, greasewood *(Sarcobatus vermiculatus)*, and hopsage. Sagebrush becomes common only far outside and higher than the playa area. At altitudes above 7,000 feet the pinyon pine and, to a lesser extent, juniper are common, but on the valley floor the only trees outside the Deep Springs College and the State highway maintenance station are a small clump of cottonwoods at the north end of the spring zone east of the saltpan. A study of the distribution of xerophytes, phreatophytes, and plants indicative of saline conditions could provide additional information on the hydrology of Deep Springs Valley, but such a study was not undertaken here.

BEDROCK GEOLOGY OF THE DEEP SPRINGS VALLEY

The bedrock exposed within the drainage area of the Deep Springs Valley is divided about equally between intrusive rocks related to the Sierra Nevada batholith and a thick sequence of Cambrian and (or) Precambrian sedimentary strata (Knopf and Kirk, 1918; Miller, 1928; Nelson, 1962). A metamorphic aureole is conspicuous along the contact between the igneous and sedimentary rocks. The contact trends north-northwest and roughly bisects the valley (pl. 1), as shown on the geologic map of the area by Nelson (1963) (table 1).

The oldest rocks in the area are thought to be Precambrian. Kirk (in Knopf and Kirk, 1918, p. 9) attributed to the Precambrian three lithologic units with an aggregate thickness of 3,600 feet as follows: "a series of sandstones and thin-bedded impure dolomites at the bottom, the Reed Dolomite above these, and locally, the Deep Springs Formation at the top." Nelson (1962) assigned Precambrian(?) age to the lowermost unit, the Wyman Formation, and indicated that

TABLE 1.—*Generalized column of sedimentary bedrock in the Deep Springs Valley drainage area*

[After Nelson (1962, 1963)]

Age	Stratigraphic unit and thickness (feet)		Lithology
Middle Cambrian			Gray shale, mudstone.
Lower Cambrian	Mule Spring Limestone 1000		Massive blue-gray algal limestone.
	Harkness Formation 2000		Gray-green shale, platy fine-grained sandstone, and siltstone.
	Saline Valley Formation 850		Brown siltstone and quartzitic sandstone; locally metamorphosed to siliceous hornfels.
	Poleta Formation 1200	Upper	Gray-green shale, blue-gray limestone beds and quartzite.
		Lower	Massive gray-blue limestone, some dolomite.
	Campito Formation 3500	Montenegro Member	Gray shale interbedded with siltstone and sandstone.
Precambrian(?)		Andrews Mountain Member	Massive black quartzitic sandstone interbedded with gray siltstone and shale.
	Deep Springs Formation 1500		Black quartzitic sandstone, dolomite, and limestone.
	Reed Dolomite 2000		Massive gray to buff dolomite.
	Wyman Formation 9000+		Thin bedded brown to dark-gray argillite, sandstone, siltstone, and interbedded lenticular gray-blue oolitic limestone or dolomite.

the overlying Reed Dolomite and Deep Springs Formation rest on regional unconformity and may possibly be Cambrian in age. These rocks have undergone some metamorphism and are commonly schistose. They underlie much of the upper parts of the major drainage systems which enter the Deep Springs Valley from the northwest.

Overlying the Deep Springs Formation is the Campito Formation of Precambrian (?) and Early Cambrian age, about 3,500 feet thick. Lower Cambrian rocks conformably overlying the Campito Formation include the Poleta, Harkless, Saline Valley, and Mule Spring Formations, in ascending order (Nelson, 1962, p. 141–143). These formations have somewhat similar lithologies consisting of limestones (commonly dolomitic), quartzitic sandstones, and gray shales in varying proportions. Except where extensively metamorphosed by granitic intrusives, these rocks lie outside the drainage of any perennial streams and affect only groundwater and ephemeral surface flow in the southwestern part of the area.

Some Middle Cambrian rocks which immediately overlie the Mule Spring Limestone (Nelson, 1963) are included in the hornfels exposed near intrusive contact in the conspicuous canyon southeast of Deep Springs Lake.

The igneous rocks of the Deep Springs Valley have been described in general terms by Miller (1928, p. 512–514). He divided the intrusive rocks into two plutons. One is an older "quartz diorite varying to gabbro-diorite" and occurring "nowhere in large bodies," but "in the form of very irregular inclusions ranging in size from less than an inch to half a mile long." The other and much larger pluton, according to Miller, ranges in composition from granite to monzonite.

In the White Mountains north of the Deep Springs Valley, Anderson (1937) has delineated two principal units of plutonic rocks which he has called the Boundary Peak and Pellisier Granites. According to Anderson (1937, p. 8–9) the Boundary Peak Granite is a fairly uniform intrusive mass, whereas the Pellisier is decidedly variable in texture and composition. Anderson suggested that the Pellisier Granite was formed by the alteration and recrystallization of sediments under the influence of subjacent magma and is actually part of the Boundary Peak Granite contact aureole. In general, as Anderson described the two units they are quite similar except that the Pellisier Granite is foliated, somewhat more mafic, usually contains hornblende, and has been more strongly affected by secondary albitization. Although Anderson's map shows only the Boundary Peak Granite extending southward toward the Deep Springs Valley, rocks which have petrographic affinities to both of Anderson's units were recognized in the area. Anderson (1937, pl. 11) also mapped a body of older quartz diorite in the southeastern White Mountains; inclusions of such composition were also noted in Deep Springs Valley.

Emerson (1959) maintained that the Pellisier Granite as mapped by Anderson actually includes two distinctive granodiorite plutons plus metamorphic rocks and that all the granitic bodies are in sharp intrusive contact.

The mountains immediately east and a little north of the lake area are composed of a monzonite mentioned by Knopf (1918, p. 60). Contrary to Miller's (1928, p. 513) contention, the monzonite mass appears to be a separate pluton, rather than a more basic facies of the granitic rocks already described.

The plutonic rocks of the Deep Springs Valley are generally similar to rock units of the east-central Sierra Nevada batholith described by Bateman (1961). They underlie not only the lower segments of the perennial Wyman-Crooked Creeks system to the northwest, but also nearly all the ephemeral drainage area east of the valley. In addition to the plutonic rocks, basaltic lava occurs in patches at the north end of the Deep Springs Valley.

Contact metamorphic rocks are prevalent in three of the major drainageways leading into Deep Springs Valley. The South Fork and Lower Birch Creek Canyons are cut largely in such rocks. Contact metamorphic rocks are also exposed for more than a mile in Wyman Canyon, and all along the largest canyon immediately southeast of Deep Springs Lake (pl. 1). The rocks cover a wide range of calc-silicate and quartzofeldspathic hornfels and represent the contact metamorphism of nearly all the sedimentary rocks types exposed in the valley area.

Several ore zones, most of which are of minor proportions, are associated with areas of pronounced contact metamorphism, particularly in the drainage of Birch Creek, although ore zones are also present near the sources of Wyman and Crooked Creeks. Some study was made of compositional variations in the plutonic and metamorphic rocks and ore zones adjacent to the major stream courses. The author intended to correlate such variations with data on minor elements in solution in Deep Springs waters. Analytical difficulties have not been fully resolved, however, and further work is necessary.

QUATERNARY GEOLOGY

As with all the intermontane valleys of the Great Basin, the Deep Springs Valley floor is a great mass of alluvium[1] derived from the erosion of the surrounding ranges. During drilling operations for the irrigation well at Deep Springs School, highly varied alluvium, including layers of clay and coarse gravel, was penetrated to a depth of more than 600 feet. Alluvial fans are conspicuous on all sides of the valley, especially on the south and west. The largest single fan covers nearly all the north end of the valley. This deposit has been built by the perennial streams from the White Mountains. Delineation of individual fans at the base is difficult as they coalesce and overlap. The material in the fans ranges in size from fine silt to large boulders. As noted by Kesseli and Beatty (1959, p. 11), differences in size distribution of particles on the fans can be related to the bedrock of the catchment area as well as distance from source. Weathering along prominent joint planes in granitic rocks produces an abundance of large blocks which are generally absent form fans built with metamporphic and sedimentary debris. Though there is a very general decrease in particle size toward the center of the valley, the trend is highly irregular. Mudflows appear to be the dominant agents in transportation and deposition at present. Detailed information on the alluvial fan deposits in the Deep Springs Valley is given by Lustig (1962).

The alluvium on the valley floor decreases in particle size southward to the lake in a broad belt up to 2 miles wide between the alluvial fans at the valley margin. Aerial photographs of the region indicate that most of these central valley deposits are related to drainage from the far northwest end of the valley and that these sediments represent deposition by streamflow which reached part way or possibly all the way to the lake. Some of these deposits are well stratified and very fine grained. They may have been reworked by waters of the former lake.

Lacustrine deposits are not confined to the present area of Deep Springs Lake. Their extent, however, is difficult to ascertain, though they have been recognized to a depth of 30 feet and a distance of 5 miles north of the present playa. The lacustrine deposits of the playa may be divided into three facies: (1) the saline crusts, (2) carbonate muds, and (3) marginal silts and sands.

The saline crusts which occupy the central part of the playa are largely lacustrine rather than efflorescent in origin. They are fairly uniform in thickness, except where much efflorescent material is superposed upon them. From a maximum thickness of nearly 2 feet at the center of the salt pan (16 inches was the maximum measured thickness; material which appeared thicker was inaccessible to direct measurement) the crusts thin shoreward to a featheredge. The thicker crusts are composed of four or five distinct layers. Thinner crusts normally contain at least two layers. The average thickness of lacustrine crusts for the entire present area of Deep Springs Lake is about 2 inches.

[1] As used here, the term "alluvium" denotes all detrital deposits resulting from fluvial agents and includes lacustrine detritus.

A substantial part of the playa surface, especially in the south and west, is covered by a saline crust formed by capillary efflorescence. The saline crusts of lacustrine origin may be roughly distinguished from efflorescent crusts by their textures. Efflorescent crusts lack uniform layering and are highly variable in thickness. They form very bumpy dusty irregular surfaces marked by contorted ridges and mounds. The saline material usually includes substantial amounts of the underlying sediment, which is normally more granular than the sediment found underneath the lacustrine crusts. Where saline crusts are both lacustrine and efflorescent in origin, the distinction between the two types, of course, is difficult. The saline crusts are discussed in greater detail in the section entitled, "Mineralogy."

Throughout most of the playa area, carbonate muds underlie the saline crusts. These muds are most typically very fine grained, smooth and even textured, plastic, and generally gray green. The top one-half inch or so is commonly light brown, apparently due to oxidation. In the central part of the saltpan, immediately beneath the thickest saline crusts, the mud is fluid, is dark green to black, and contains much organic matter. This black layer persists outward from the center of the lake at increasingly greater depths below the normal gray-green material, and finally lenses out. This distribution is illustrated by sections taken in shallow auger holes on both the north and south sides of the playa (fig. 5). Near its outer limits, the layer is faintly laminated in places.

The stratigraphy of the upper 26 feet of lacustrine carbonate mud in the playa area is given in the log of a shallow hole drilled with a jeep-mounted auger under the direction of Ward C. Smith of the U.S. Geological Survey (near loc. DL2, fig. 2; stratigraphic notes and material courtesy W. C. Smith and R. C. Erd). Alternating light and dark laminae mentioned below may be true varves (annular layers). These features are discussed in greater detail in the section entitled, "Mineralogy" (p. A43).

Feet	
1.0– 2.0	Wet light green plastic mud.
2.0– 3.5	Laminated black and white mud, green near the top. Laminae about 1 to 3 mm thick.
3.5– 6.0	Black plastic clay, smell of H_2S, grading to unit below.
6.0– 7.0	Gray-green mud which contains lenses of medium to coarse sand.
7.0–10.0	Laminated black and green mud in which green dominates. Laminae are about 2 mm thick.
10.0–26.0	Crudely banded gray-green to yellow-green mud.

The lacustrine carbonate muds intertongue near the margins of the playa with the lenses of coarser grained alluvial material, which ranges in particle size from fine silt to sand. A typical section is represented by a shallow auger hole in the southwest part of the playa (loc. 55, fig. 2):

Feet	
0.2– 0.7	Moist granular textured silt and clay.
.7– 1.1	Dark brown silt to medium sand. Faintly bedded.
1.1– 1.6	Light brown silty mud.
1.6– 2.3	Dark brown fine silt to fine sand.
2.3– 3.0	Gray-green dense plastic mud.

All contacts are somewhat irregular and gradational.

Lacustrine deposits outside the playa have been extensively reworked by runoff and wind. These deposits are not readily distinguished from finer grained fluvial deposits presently being transported from the mountains, except where associated with ancient strand lines or where characterized by lacustrine faunal remains, chiefly gastropods of the family Planorbidae, probably *Gyraulus* (K. V. Slack, written commun., 1961).

Lustig (1962, p. 98) obtained a carbon-14 date of 1380±250 years B.P. for a fire hearth in deposits bordering the lower channel from Antelope Canyon (alt 5,080 ft). This site is close to some distinctive linear ridges presumably formed at a much higher stage of Deep Springs Lake. From consideration of the rates of erosion and deposition, Lustig (p. 99) dated this high stage at 700 to 1,000 years B.P.

HYDROLOGY

PRECIPITATION

Precipitation and temperature records have been obtained by the U.S. Weather Bureau at Deep Springs School (alt 5,225 ft) in the northern part of the valley since 1948 and at two research laboratories operated by the University of California on the crest of the White Mountains since 1955. The lower station (White Mountain No. 1, alt 10,150 ft) lies within the Deep Springs Valley drainage at the head of Crooked Creek; inasmuch as most of the inflow to the Deep Springs Valley is ultimately derived from the mountains to the northwest, this station is very advantageously located.

Climatic data for Deep Springs School and for White Mountain No. 1 for the period 1948–60 (table 2) indicate that maximum precipitation generally occurs in the winter or spring. Although no data are available for the specific area of Deep Springs Lake, the annual distribution of precipitation is presumably very similar to that at the school. Valley residents believe that the lake area receives more precipitation than does the school, and inasmuch as the White Mountain No. 1 station is near the highest altitude

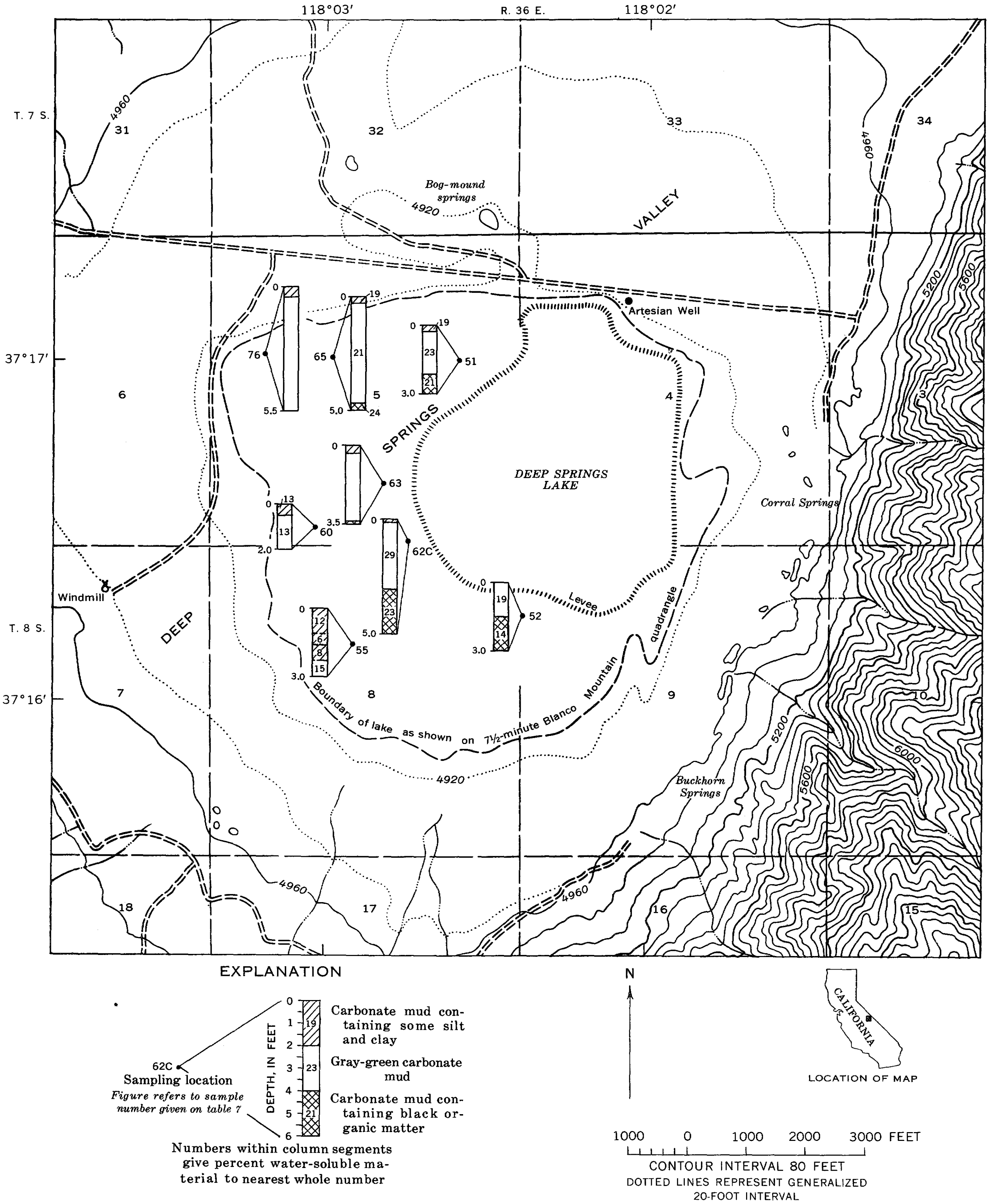


FIGURE 5.—Map of the Deep Springs Lake area showing the location of sites outside the saltpan where cores were taken or pits dug to establish the stratigraphy of the surficial deposits. The general stratigraphy is shown in the small columns adjacent to each site. Numbers within the columns refer to the percent readily water soluble material within each stratigraphic unit.

TABLE 2.—*Climatic data for Deep Springs Valley and vicinity, California*

[M, missing; E, estimated]

Date	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Annual
AIR TEMPERATURE (°C)													
Deep Springs School													
1949	12.4	22.2	37.0	52.5	56.5M	----	75.0M	70.2	68.0	51.0	47.5	30.9M	----
1950	26.2	38.9	43.1	51.6	59.5	65.3	71.2M	72.2M	62.3M	57.1M	44.8M	----	----
1951	29.3	----	42.0	54.6	59.8	67.5	76.1M	71.7M	67.8M	52.7M	40.7M	28.7M	----
1952	23.7	33.3	----	51.4	62.1M	67.4M	74.2M	----	65.9M	58.6M	----	----	
1953	38.3	37.9	44.2	50.4	52.2	66.4M	77.2	70.8	66.8	52.0M	42.7M	32.3	52.6E
1954	33.9	42.3	41.0	56.3	63.5	70.5	78.5	71.2	65.4M	53.3	44.5	32.2	54.4
1955	21.6	33.3	42.2M	47.4	57.5	----	74.2	----	----	55.7M	41.2	----	----
1956	36.8M	33.8	44.2	50.4	59.8	69.6	72.4	69.6	66.6	52.5	40.8	----	----
1957	25.6	41.6	44.4M	50.5M	52.0	68.1	70.2M	70.2	----	49.6	38.3	----	----
1958	----	40.4M	38.7M	48.8M	61.5M	69.5M	74.4M	74.9M	66.9M	59.3M	42.5M	40.0M	----
1959	36.8	31.1	45.5M	54.3M	57.4	71.2	77.7M	73.4M	63.2	55.2	44.3	36.6	54.0
1960	28.4	36.3	48.6	53.6	60.9	74.7	75.1	73.7M	68.3M	52.9	40.6	35.6	54.0
Average	28.5	35.6	42.8	51.8	58.6	69.0	74.7	71.8	66.1	54.2	42.5	33.8	53.8
White Mountain No. 1 [Alt 10,150 ft]													
1955	----	----	----	----	----	----	----	----	----	38.5	30.0	23.0	----
1956	20.8M	16.5	23.9	25.3	34.2	46.1	49.3	47.6	46.3	33.7	28.6	24.5	33.1
1957	13.0	24.7	23.4	26.6	31.8	47.6	49.9	49.4	44.0	31.2M	23.8	24.2	32.5
1958	20.2	20.7	12.7	24.3	38.6	44.0	50.3	52.6	46.4	39.5	27.7	30.3	33.9
1959	23.9	14.8	26.1	35.5	34.7	49.7	55.4	51.1	44.0	39.4	31.8	24.9	35.9
1960	15.6	19.9	29.9	32.2	37.9	49.9	53.0	52.4	47.2	35.4	25.3	19.2	34.8
Average	18.7	19.3	23.2	28.8	35.4	47.5	51.6	50.6	45.6	36.3	27.9	24.4	34.0
White Mountain No. 2 [Alt 12,470 ft]													
1955	----	----	----	----	----	----	----	----	----	33.8	24.2	----	----
1956	14.5	8.8	18.4	18.3	26.5	38.7	43.7M	42.7	41.9	26.5	26.1	22.0	27.4
1957	9.6	18.3	15.7	20.2	23.7	40.5	44.7	44.5	39.6	25.8	18.7	20.0	26.8
1958	17.2	21.5	14.2	22.3	31.5	33.1	41.3	44.9	37.8	32.0	20.2	22.6	28.2
1959	15.1	7.0	16.2	24.4M	21.3	39.7	46.9	41.8	33.7	29.8	26.3	----	----
1960	5.3	7.9	18.4	20.9	27.1	41.6	43.0	44.8	39.5	29.2	19.5	16.3	26.1
Average	12.3	12.7	16.6	21.2	26.0	38.7	43.9	43.7	38.5	29.5	22.5	20.2	27.1
PRECIPITATION (INCHES)													
Deep Springs School													
1949	0.61	0.40	0.15	0.03	1.20	----	----	0.13	0.00	Tr.	0.43	0.04	----
1950	.04	.25	.65	.00	Tr.	0.00	0.26	.11	1.46	0.29	.55	----	----
1951	.04	----	.39	1.23	.30	.25	.31	.25	.00	.04	.00	1.98	----
1952	2.07	.00	1.55	2.58	.00	.30	.91	Tr.	.08	.00	.25	.83E	8.57E
1953	.08	.15	.29	Tr.	.30	.00	.50	Tr.	.00	.06	.42	.02	1.82
1954	.64	.97	2.05	1.00	.20	.47	.52	.00	.03	.00	.42	.23	6.53
1955	.94E	.11	Tr.	.22	1.36	.00E	Tr.	.28E	.20E	.00	.01	1.75E	4.87E
1956	1.26	.00	.00	3.80	.34	.00	.66	.00	.00	.42	.00	.00	6.48
1957	.72E	.31	.02	.66	1.46	.02	.06	.13	.20	1.42	.28	1.61E	6.89E
1958	.98	1.66	1.52	1.33	.52	Tr.	.00	.19	.32	.03	.17	.00	6.72
1959	.04	1.61E	.00	.04	.17	Tr.	.15	.49	.17	Tr.	.00	.36E	3.03E
1960	.23E	.83E	.07E	.13	.00	.00	.05	Tr.	.36	.03	3.46	.00	5.16E
Average	0.64	0.57	0.56	0.92	0.49	0.09	0.31	0.14	0.24	0.19	0.50	0.62	5.56
White Mountain No. 1													
1955	----	----	----	----	----	----	----	----	----	0.00	0.36	6.03	----
1956	2.79	0.16	0.02	4.15	1.79	Tr.	3.30	0.03	0.02	.96	Tr.	.07	13.29
1957	2.03	1.25	.86	1.90	2.74	Tr.	.06	.00	1.50	1.54	1.22	1.22	14.32
1958	1.61	2.05	3.05	2.28	.94	Tr.	Tr.	1.09	.30	.38	.84	.07	12.61
1959	.29	3.10	.04	.04	1.07	Tr.	2.05	.75	2.25	.69	.00	.36	10.64
1960	.44	1.69	.14	.43	.09	Tr.	.86	.57	.30	.72	4.29	.34	9.87
Average	1.43	1.65	0.82	1.76	1.33	Tr.	1.25	0.50	0.87	0.72	1.12	1.35	12.15
White Mountain No. 2													
1955	----	----	----	----	----	----	----	----	----	0.08	0.93	----	----
1956	2.30	0.45	0.02	6.65	1.67	0.18	2.93	0.26	0.00	1.95	.08	0.31	16.80
1957	1.85	.92	2.41	1.59	4.66	.26	.10	Tr.	.49	3.33	1.56	.83	18.00
1958	1.41	2.30	2.39	1.29	.79	.48	.32	4.30	.74	.48	1.23	.15	15.88
1959	1.07	3.83	.50	.62	3.11	Tr.	.65	.40	1.58	2.26	.00	----	----
1960	1.37	1.21	.11	1.11	.38	.25	1.08	.75	.09	1.12	3.17	.19	10.83
Average	1.60	1.74	1.09	2.25	2.12	0.23	1.02	1.14	0.90	1.54	1.16	0.37	15.38

within the Deep Springs Valley drainage, the two station records probably give a reasonable range for weather conditions in the entire basin.

According to Kesseli and Beatty (1959, p. 18), the White Mountains lie in a transition zone between two areas which have markedly different precipitation regimes. The Owens Valley, which has a typical California coastal regime, has little or no precipitation in summer, whereas the areas east of the main White Mountains, such as the Fish Lake Valley, have a fairly uniform distribution of precipitation throughout the year. Seasonal precipitation distribution within the Deep Springs basin appears to be intermediate between the two regimes (fig. 6).

Storm patterns in the Deep Springs area show marked seasonal variation. Winter precipitation is derived from cyclonic storms moving eastward across the Sierra Nevada where much of their moisture is lost. Thus, though precipitation may be of relatively long duration, it normally is only moderate to light in intensity. Strong northwest winds are common during the winter storms. During the summer, precipitation accompanies thunderstorms which frequently build up and move northward along the crests of the ranges on either side of the Deep Springs Valley. The summer thunderstorm period is often marked by strong southeasterly winds. Kesseli and Beatty (1959, p. 18) associated increased thunderstorm activity east

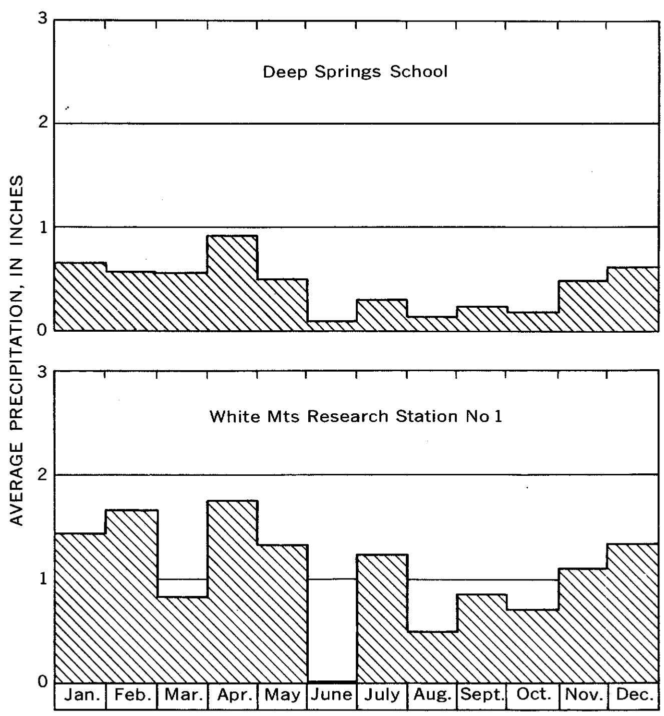


FIGURE 6.—Seasonal distribution of precipitation for the Deep Springs Valley.

of the White Mountains with a flow of warm moist air from the Gulf of Mexico, and they suggested that summer thunderstorms in this area are likely to produce at least as much precipitation as weakened winter cyclonic storms. As discussed in some detail by Kesseli and Beatty (1959, p. 22–23), thunderstorm precipitation is sporadic, both in time and place, and reliable accounts of intensities and total amounts are commonly not available in the established weather station records. However, the data for Deep Springs School and White Mountain No. 1 indicate that maximum precipitation in the Deep Springs drainage area usually occurs in April (fig. 6).

EVAPORATION

The Deep Springs Valley—because of size, higher altitude associated with lower mean annual temperature, and complete enclosure by steep mountain masses—does not have as high an evaporation rate as many nearby intermontane basins. Extrapolation from the evaporation maps prepared by the U.S. Weather Bureau (Kohler, Nordenson, and Baker, 1959) indicates that the average class-A-pan evaporation is about 65 inches per year, and lake evaporation is about 45 inches per year. Similarly, from the relation derived by Langbein (1961, p. 3), lake evaporation would be about 48 inches per year at a mean annual temperature of 54°F. (See table 2a.) At Deep Springs Lake, shallow relatively dilute surface waters probably evaporate at nearly the maximum pan rate, but the rate of intercrustal and subsurface brine evaporation is probably close to or even lower than the rate for lakes where surface area and storage of radiation energy at depth are important factors. Because of the additional effect of the lowering of vapor pressure by high solute content, the evaporation rate for the perennial intercrustal brines at Deep Springs Lake may be as low as 30 inches per year. An estimate of the net annual evaporation from lake waters indicates a rainfall equivalent to, or slightly higher than, that recorded at Deep Springs School, about 5 to 7 inches per year. (See table 2b.) On this basis the net evaporation rate for surface waters would range from 40 to 60 inches per year depending on salinity and depth. A reasonable estimate of the mean is about 4.25 feet per year.

HYDROGRAPHY OF THE DEEP SPRINGS VALLEY

Perennial surface waters are a major factor in the hydrology of Deep Springs Valley. In the mountains to the northwest of the valley, one stream is perennial and segments of three streams flow during most of the year. Most of the time, the total streamflow in the mountainous part of Deep Springs Valley drainage area is equaled or exceeded by discharge from springs near the mountain fronts and the lake. Several fairly large ephemeral drainage systems, which are normally dry for extensive periods, may carry considerable quantities of flow during times of storm runoff.

STREAMS AND DRAINAGE OF THE MOUNTAIN AREAS

The largest stream entering the Deep Springs Valley, Wyman Creek, has a drainage area of 29.7 square miles exclusive of its major tributary, Crooked Creek. Wyman Creek rises in a series of wet meadows called Roberts Ranch, about 1½ miles above the contact zone between granitic rocks and Precambrian metasediments, and maintains perennial flow over a total length of about 7 miles. Throughout most of the distance of perennial flow, the Wyman Creek channel is narrow but well defined, and except in a few reaches closely bounded by bedrock, the channel is bordered and overgrown with brush. Two or three miles upstream from the head of its fan, the Wyman Creek channel is carved directly in granitic bedrock. Wyman Canyon is very narrow and steep sided through most of the distance underlain by igneous rocks. In contrast, above Roberts Ranch where Wyman Creek is largely ephemeral and underlain by sedimentary rocks, the valley is broader and is floored with much alluvium. Although distinct channels are present, no flow was observed in the Wyman Creek drainage above Roberts Ranch at any time except from Cedar Spring, a tributary channel about 3 miles above Roberts Ranch. Flow from this spring is light; less than 0.01 cfs (cubic feet per second) disappears into alluvium a very short distance from the source.

In the short reach of Wyman Canyon between Roberts Ranch and the first outcrops of intrusive rock, the canyon gradually narrows and the valley of Wyman Creek contains less fill as the metamorphic grade of underlying sedimentary sequences increases. The effect of decreased permeability in surrounding and underlying bedrock is suggested by comparing flow estimates made on Wyman Creek just below the mouth of Mill Canyon and estimates made just above the confluence with Crooked Creek (table 3). Nearly two-thirds of the total flow of Wyman Creek was measured less than 1 mile from its source.

The main drainageway of Wyman Creek extends from the White Mountain ridge for a total linear distance of about 12 miles and enters the northwest corner of the Deep Springs Valley at the head of an extensive, low, and complex alluvial fan. The main channel curves very sharply at this point and for about 2 miles follows the outer limits of a small series of secondary fans before reaching the central floor of the valley.

TABLE 3.—*Estimated discharge, in cubic feet per second, of surface water units in Deep Springs Valley, 1959–61*

[Numbers in parentheses refer to sample localities shown in fig. 2; n.m., not measured]

Source	Oct. 1959	June 1960	Sept. 1960	May 1961	Aug. 1961
Wyman Creek:					
At Mill Canyon	....	2.87	....	....	....
Above confluence with Crooked Creek (47)	3.71	....	....	4.12	1.92
Below confluence with Crooked Creek (56)	....	4.32	2.05	....	....
Crooked Creek	.65	.62	No flow.	.52	No flow.
Total, Wyman Creek plus Crooked Creek	4.36	3.70	2.05	4.64	1.92
Corral Springs:					
Group 1 (48A)	1.32	n.m.	.69	1.44	1.00
Group 2 (48B)	.57	n.m.	.31	.62	.30
Total	1.89	....	1.00	2.06	1.30
Outflow from the open sag pond (32)	.33	n.m.	.60	1.67	.54
Buckhorn Springs:					
Nos. 1–5 (33)	.38	n.m.	.30	.43	.16
Nos. 6–7 (34)	.40	n.m.	.21	.69	.49
Nos. 8–10 (35)	.21	n.m.	.13	.32	.15
Group 11 (36)	.74	.44	1.06	1.59	1.04
Total	1.73	....	1.70	3.03	1.84

NOTE.—Bog-mound springs discharge is very difficult to estimate but apparently near constant annually; northwest bog mound, about 0.01–0.02 cfs; western bog mound, at least 0.05 cfs; central bog mound, at least 0.20 cfs; and old artesian well (eastern bog mound) spring, about 0.03 cfs.

The valley channel is normally dry, as the waters of the creek are diverted into a canal which leads to the Deep Springs Junior College. This canal is headed by a sediment trap, which is periodically flushed; at such times water does occupy the former channel. However, through most of the year all the water of Wyman Creek is utilized by the school. Discharge from the school is released into a series of shallow channels directly south of the school lands, and what is not consumed or lost by evaporation eventually reenters Wyman Creek drainage. Inasmuch as creek water is stored in an open reservoir behind the school, evaporation losses may be high.

By far the largest tributary to Wyman Creek, Crooked Creek, enters the main stream about a mile above the sediment trap. Crooked Creek, about 14 miles long, drains 21.8 square miles. Although not perennial, Crooked Creek through most of the year maintains some flow from Dead Horse Meadow to its confluence with Wyman Creek, a distance of about 4 miles; the valley is especially narrow and the channel is carved in bedrock through most of this reach. This segment represents a former small tributary of Wyman Creek which by headward erosion and capture has diverted upper Crooked Creek drainage from the Fish Lake Valley. The valley of Crooked Creek above Dead Horse Meadow contains considerable fill, but the channel is commonly cut through almost to bedrock. Much of the Crooked Creek drainage area is underlain by granitic rocks and basalt. Discharge estimates (table 3) indicate that Crooked Creek contributes up to 20 percent of the total flow of the Wyman-Crooked Creeks system.

Other major drainage systems tributary to the Deep Springs Valley on the west include Birch Creek, Antelope Creek, Beer Creek, and Payson Canyon. Birch Creek enters the valley about 6 miles southwest of Wyman Creek and about 4 miles due north of the lake. Though Birch Creek drains an area of 16.0 square miles, flow is perennial only in a short reach of bedrock channel which extends about 1½ miles from a meadow at the confluence of the north and south forks to the head of the relatively wide alluviated valley immediately north of its alluvial fan. The Birch Creek Canyon below the confluence of the north and south forks is cut through granitic and contact metamorphic rocks—chiefly calc-silicate hornfels, marbles, and schists. The actual source of Birch Creek surface water is a group of 16 springs in the wet meadow area at the confluence of the north and south forks. Discharge is relatively stable at the source, but streamflow decreases by more than two-thirds and fluctuates diurnally downstream. Maximum measured discharge is 0.5 cfs, but normal flow is about one-tenth as much. The active channel of the stream is tightly cemented by precipitated calcite throughout its length (Barnes, 1962).

A similar situation to the one at Birch Creek is found about 3 miles farther south. A little more than a mile above Antelope Spring, a very small surface stream originates from the base of alluvium at the head of a narrow gorge and flows with gradually decreasing discharge about ⅓ to ½ mile before disappearing. The exposed bedrock in the gorge consists of calcareous and argillaceous sediments which are not intensively metamorphosed. In contrast to Birch Creek, bed cementation with calcite is not as apparent in Antelope Canyon, which drains a total of 7.0 square miles and, in addition to its creek, includes a small spring near its mouth, where rapid downcutting along the main channel has cut under the alluvium filling a tributary valley.

The only surface waters outside the major drainage lines on the west side of the valley are south of Antelope Spring, where a small group of springs discharge from cemented fanglomerate about one-half mile south of the mouth of Antelope Canyon. These springs give rise to small vegetated channels which flow a maximum distance of about 100 yards before disappearing into valley alluvium. Though ostensibly derived from a small drainage upslope, much of the discharge seems to be derived from other systems by downvalley movement through local jointing along the mountain front; this hypothesis is also supported by the water chemistry (p. A33).

Payson Canyon contains the largest fully ephemeral channel tributary to the Deep Springs Valley. Field observation and aerial photographs indicate that large quantities of surface flow may be carried from this drainage during periods of heavy runoff.

On the east side of the valley, the most conspicuous drainageway is a deep narrow canyon which lies very close to the contact of plutonic and sedimentary rocks southeast of the lake (pl. 1). The channel at the floor of the canyon is carved in bedrock for the most part, but a few small clumps of vegetation give some evidence of seepage.

There are many other ephemeral stream systems tributary to the Deep Springs Valley. All such drainage from the mountains discharges onto alluvial fans at the edge of the valley. Most of the fan surfaces are cut by several well-defined channels. The most marked channels, those of the Payson and Antelope Canyons systems, are on the largest fans of the southern valley area. Most channels reach a maximum depth where they cross remnant shorelines of the ancient lake which filled the valley. These systems rarely carry surface flow and are a result of intensive storms in their catchment areas. During the period of study, 1959–61, flash flooding was observed only in the Antelope Canyon system and in one of the drainage systems south of the playa; in each case flooding followed heavy thunderstorms. Only the flood from Antelope Canyon reached the lake. However, William A. Jenkins of the California Division of Highways has given the author accounts of flash flooding of major proportions in Payson Canyon in 1953 and 1958 and in Birch Creek Canyon in 1952. The Birch Creek flood carried a massive amount of material, and its effects are still evident. A road which formerly ran up the main stem and North Fork Birch Creek to Roberts Ridge was completely destroyed at this time. L. K. Lustig (oral commun.) reported minor flooding in Wyman Canyon in late August 1960; sediment-laden water moved down the old Wyman Creek channels to the center of the valley. Kesseli and Beatty (1959, p. 68–69) reported a flood of Antelope Canyon in August 1957.

In their study of the entire White Mountain range, Kesseli and Beatty (1959, p. 89) suggested that in the whole area there may be as many as 15 floods per decade of which 5 may be of major proportions. Flood frequency within individual drainages is more difficult to estimate, but major flooding has probably not occurred in many of the canyons in over 50 years, though the larger drainage systems probably undergo minor flooding in part every few years.

Most of the surface inflow to Deep Springs Lake is directly derived from springs near the lake area. These springs may be readily subdivided into groups based on location, setting, genesis, and chemical composition.

SPRINGS NORTH OF DEEP SPRINGS LAKE

The springs which occupy the broad shallow depression north of the playa have been given the field name bog-mound springs because they characteristically give rise to marshy muskeglike mounds which are roughly circular and which rise above the immediate surrounding area. Smaller mounds are occupied by marsh grass and reeds, commonly growing in clumps, and by standing water in the intervening spaces; no flow is apparent. The larger mounds have flowing springs. The large springs discharge from about 5 to 30 gpm (gallons per minute) into the surrounding reed-filled marsh which may contain discontinuous areas of open water. Flow can readily be detected as much as 150 feet from the spring orifice. Surface water does not usually extend farther than about 100 yards beyond the mound before disappearing. Much of this water, however, seems to reach the lake or nearby springs through a series of seepage lines trending southeast and marked by green vegetation. Material from holes put down in these areas has a much higher moisture content and shows more soil development than surrounding sediments.

The situation is somewhat analogous to the conditions described by Bunting (1961). Some seepage apparently escapes evapotranspiration and proceeds under the surface to enter Deep Springs Lake through a breech in the levee mound on the northeast side of the playa. In most of the bog mounds, the materials in which the springs arise are largely fine silts and clays admixed with a great deal of organic matter which is held together only through the binding action of roots. However, in the northwesternmost part of the bog-mount area, one of the springs arises in material which is largely of sand size, and the result is a treacherous quicksand pocket.

The vegetated mat around many bog mounds is spongy; trampling of this mat produces gas bubbles in abundance. Much of the gas is trapped air or carbon dioxide. However, around the orifice and immediately downstream from actual flowing springs, there is abundant white fungus similar to that found around hot springs. Traces of hydrogen sulfide may also be detected here.

The largest of the bog-mound springs is about a quarter of a mile due north of Deep Springs Lake. The mound and marsh area is about 200 feet long and 75 to 90 feet wide. The mound and the springs are about 4 feet higher than the surrounding area.

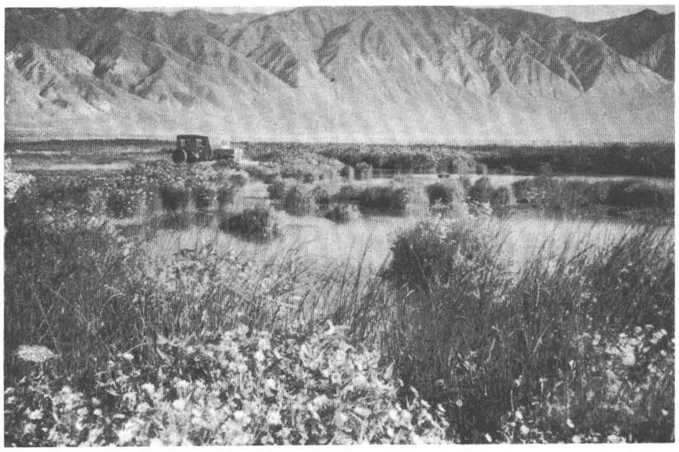


FIGURE 7.—View looking southeast across the western bog-mound marsh.

Another large spring is about one-half mile west (fig. 7). Though the marsh area is large, the source of flow is difficult to detect, the mound is poorly formed, and there is much open water in a pond as much as 2 feet deep.

SPRINGS EAST OF DEEP SPRINGS LAKE

The major discharge of surface flow to Deep Springs Lake is derived from springs along the prominent fault zone to the south and east of the playa. These springs are separated into two groups by the fault trough and alluvial ridges at the steepest part of the mountain front. The northern group, referred to as the Corral Springs, lies approximately 1½ miles north of the southern group, known as the Buckhorn Springs. The actual number of springs in the Corral Springs group is variable. The springs consist of rapid flows of clear water which emerge from beneath outcrops of cemented fanglomerate directly on the fault scarp and a series of marshy seeps which coalesce not far from their point of origin at the base of the scarp. These springs all appear fresh and clear, except for the southernmost of the group which is brackish, has an odor of hydrogen sulfide, and supports the growth of white fungus. About 100 feet from the point of origin, all outflow from the Corral Springs coalesces into a channel, as much as about 3 feet wide, which is partly filled with marsh vegetation. Although this channel leads directly westward toward the playa, much of the flow is diverted into a canal trending northward. Leakage from this canal toward the playa occurs at several places, and the canal is totally dry after carrying some flow a maximum distance of 500 feet. Most of the seepage from the canal rejoins flow

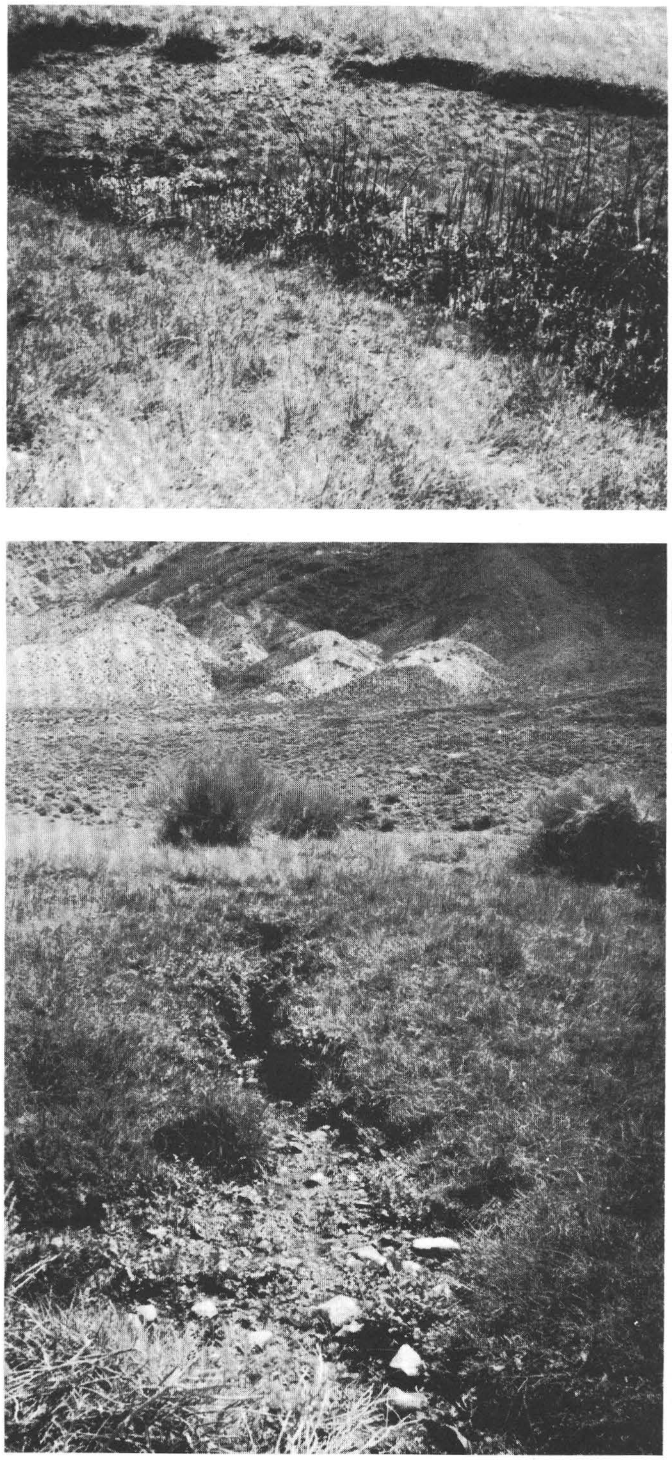


FIGURE 8.—Buckhorn Springs 5 and 7, selected as representing extremes of channel types at the source. The broad vegetation-choked spring area is No. 5 (upper panel), whereas the trenched rocky channel is No. 7 (lower panel).

from the natural channel in a gently sloping grassy flat with little detectable flow. Eventually, the water is impounded east of the saltpan by the levee mound. A large pond forms here in the wet season, but normally the area is simply a damp mudflat. Apparently, the water eventually reaches the saltpan by seepage beneath the levee mound.

The Buckhorn Springs consists of 11 points of emergence where flow arises beneath the outermost coarse alluvial material; the line of springs parallels the approximate extension of the fault zone, which becomes less well defined southward. All the springs originate from depressions at the foot of an escarpment which results from faulting and may coincide with an old beach level. From south to north, the first five springs arise in rocky depressions and flow through channels choked with marsh vegetation for as much as 150 feet from the points of origin (fig. 8). The outflow channels are shallow, and many have little detectable flow which carries beyond the source. These channels coalesce in a large marshy area near the southeastern part of the playa and supply most of the surface water which accumulates on the lake muds south of the levee mound. Flow from springs 6 and 7 (from south to north) arises from sharp depressions as much as 4 feet deep at the foot of the escarpment and continues several feet through narrow channels (fig. 8). Springs 8 and 9 are small seeps which originate in wet grassy areas beneath coarse alluvial material. Springs 10 and 11 arise in the depression at the foot of a vertical fault scarp as much as 50 feet high and marked by outcrop of cemented gravel fanglomerate (fig. 9). Spring 11 results from the coalescence of flow from 17 seeps in the broad depression at the base of the scarp and provides the largest discharge in the Buckhorn group. The channel leading from spring 11 is narrow in its upper part and is as much as 5 feet deep. Fineness of bed material, low altitude, and the proximity of the steep scarp suggest that these springs originate directly from the fault zone.

Outflow from springs 6–11 discharges down a series of vegetated channels and coalesces in a large marsh

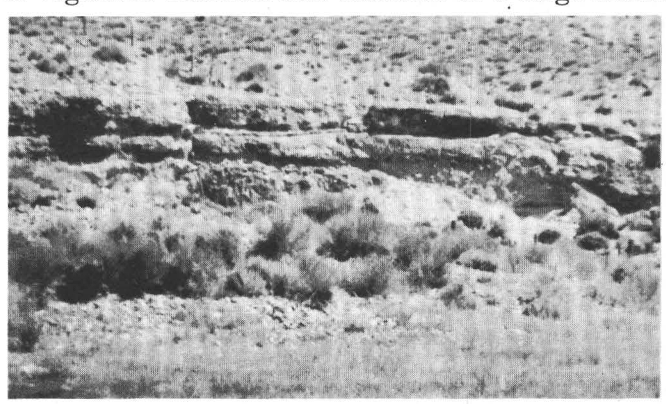


FIGURE 9.—Cemented fanglomerate on fault scarp above Buckhorn Springs 10 and 11. Buckhorn Spring 11 arises immediately below the center foreground.

which extends lakeward to the south end of the levee. Some of this flow is diverted from the main channel into the depressions between ridges in the alluvial mass which separates the Buckhorn from the Corral Springs. During the period of this study the southwesternmost depression contained a large pond which was not present when the aerial photographs were taken in 1947. Flow diverted into this pond returns to the marshy area south of the spring by seepage through the impounding ridge.

Attempts have been made to channel surface waters from several of the Buckhorn Springs into small canals lying transverse to the outflow channels. The purpose was to spread water over the grasslands lakeward from the springs. Most of these attempts have not greatly interfered with the general flow systems or patterns. All surface water from the springs which does not evaporate eventually appears to reach the central playa area through marshes and breeches in the southern part of the levee.

The temperature (about 20° C) of the waters from the Corral and Buckhorn Springs is consistently 5 to 10° C higher than that measured in the bog-mound springs or any of the mountain streams. This higher temperature suggests that the fault-zone springs issue water which has come from considerable depth.

SAG PONDS

In the fault trough between the Corral and Buckhorn Springs are two perennial sag ponds. The northern pond is about 500 feet long and 50 feet wide at its widest point. It is nearly surrounded by alluvial ridges as high as 50 feet, except at the northwest corner, where the confining ridge is only about 5 feet above the average pond level. At highest stage, outflow spills over this ridge, but normally the pond is closed. The pond is fed by seepage around the margin, which comes to the surface in small irregular channels on the north and east sides. A few reeds apparently supported by the seepage grow at the south end and along the east border (fig. 10). At all stages the waters of the pond are amber in color, and contain abundant algal material. In the spring high stage waters teem with brine shrimp. At lower stages, in late summer, the pond contains a salt crust as much as 2 inches thick which is partly submerged in the brine and irregularly eroded at the edges by seepage inflow from the margins. Efflorescent crust is abundant in the flat areas at the south and northeast ends of the pond at almost all stages.

A second sag pond occurs about 2,000 feet southwest of the northern pond. This one is as much as 750 feet long and about 75 feet wide at its widest point. It is also surrounded by high ridges except on the northwest side. This pond, though also fed by marginal seepage, has a perennial outflow at the northwest and maintains fresh or slightly brackish water throughout the year. Marginal vegetation and aquatic life are abundant and green algal mats are often conspicuous. Overflow is through a relatively deep channel to the northwest. Some of this flow is diverted into a canal in an attempt to spread water over the grassland. A large breech in this old canal, however, allows much of the water to escape into a large interridge depression and locally gives rise to standing water. Ephemeral bodies of water also accumulate in other interridge areas lakeward from the sag ponds, especially during wet seasons. Marsh vegetation is common in those depressions which receive water most often. Water loss apparently takes place from all these depressions by seepage through the ridges toward the lake as well as by evaporation.

FIGURE 10.—The closed sag pond at low stage; shown are the marginal seepage inflow to the pond and the growth of reeds in the seepage areas. The photograph is taken from the north end looking at the eastern shore. The white patches are efflorescent crusts.

GROUND WATER

Information on ground water in the Deep Springs Valley is limited to what could be obtained from five wells in the valley area and from observations at the springs. The only fully utilized wells are in the north end of the valley at Deep Springs College and at the highway maintenance station. Although the college obtains most of its water supply from Wyman Creek, a large irrigation well has been installed on the north side of the school. This well was drilled more than 600 feet; unconsolidated alluvium was penetrated throughout the section. Depth to the water table averages about 185 feet. According to personnel at the college, extensive pumping has resulted in little draw-down in a short period. The well at the highway maintenance station has been drilled to approximately 280 feet, the depth to the water table being about 250 feet. This well supplies all water for operations of the maintenance station, but is pumped only for long enough periods to fill the reservoir tank. The water composition is very close to that of the school irrigation well and total solids average about 300 ppm (parts per million). There are two stock wells in the valley, one very close to the center of the valley and the other about one-half mile due west of the lake area. The well at the center of the valley is pumped infrequently and only in sufficient quantity to fill the stock tanks; depth to the water table is 45 feet. The well to the west of the lake is windmill operated and has a depth to water of 16 feet. There is very little increase in total dissolved solids in well water downvalley. The fifth valley well is a remnant of the old Inyo Chemical Co. workings, and was presumably used as a water supply for the pilot plant. This well was put down in the area of one of the bog-mound springs, and, even though the casing apparently is heavily corroded, it maintains at least 3 feet of artesian head.

Inasmuch as there is little subsurface data, it is very difficult to ascertain the extent of the ground-water reservoir in the Deep Springs Valley. The altitude and occurrence of springs, especially on the north side of Deep Springs Lake, suggest that the valley alluvium and thus the occurrence of ground water is subject to considerable variation both vertically and laterally. However, the available well data indicate that the piezometric surface slopes roughly parallel to the surface of the valley floor at a somewhat lower gradient from the mouth of Wyman Canyon to somewhere north of the playa (fig. 11). In the area north of the playa the ground water in the valley alluvium appears under artesian head, aided by gas pressure, as the bog-mound springs. The rough correspondence in maximum altitude of the mounds with the head measured in the old artesian well suggests that the mounds were built with material carried by upwelling ground waters.

Information on the areal pattern of water movement in the Deep Springs Valley can be obtained from field observations and inspection of aerial photographs (fig. 12). A complex pattern of distributary channels lead from Wyman Creek toward the lake and extend over a wide area in the central part of the valley. The most recent of these channels supports a substantial amount of vegetation as compared to the surrounding area. Prior to diversion, these channels carried the flow of Wyman Creek by the school. Their number and size suggest generally higher discharges than those now observed in the Wyman Creek system. The most recent channels are on the east side of the valley floor and two-thirds of the way downvalley they come very close to the mountain front. Although these channels are normally dry, they may carry some of the heaviest runoff from the Wyman Creek system. The multichannel patterns probably indicate the position of underlying gravel deposits related to former channels and mark the route of ground water contained therein. Channels related to the Wyman Creek system appear to cut off or deflect the lower parts of other drainages from the western mountain front farther south, such as the drainages from Birch Creek and Antelope Canyons. The Wyman Creek channels become less distinct near the lake and are more poorly defined below each old lake level. Most of the later channels may be traced

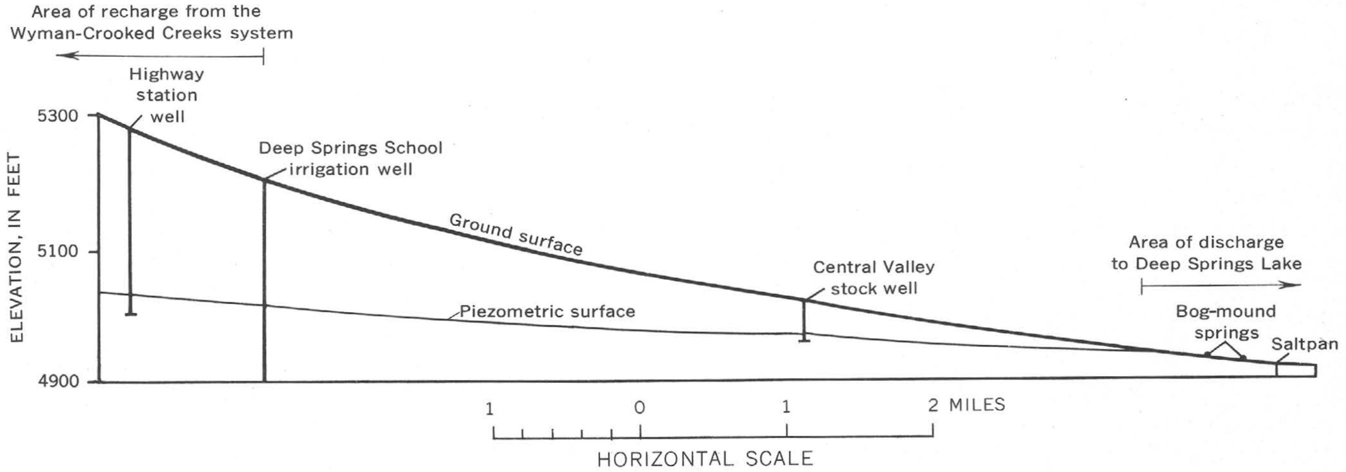


FIGURE 11.—North-south section of the alluvium in the Deep Springs Valley showing the estimated position of the piezometric surface, 1959–61.

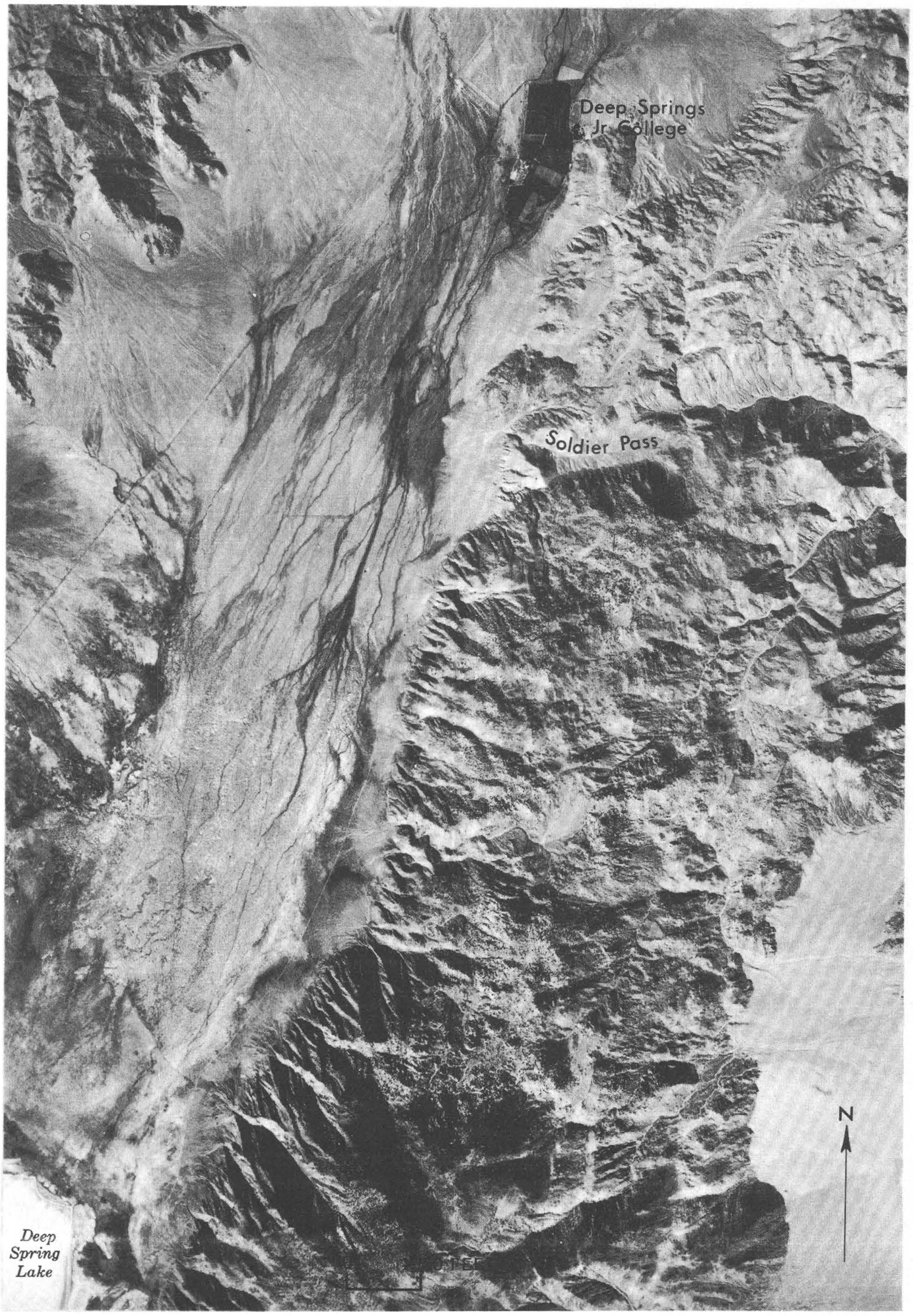


FIGURE 12.—High-altitude aerial photograph of the Deep Springs Valley north of Deep Springs Lake illustrating the Wyman Creek channel patterns and the mountain front fault zone on the east side of the valley. The dark patchwork at the north end of the photograph is the fields of the Deep Springs Junior College.

toward the fault zone north of the Corral Springs area. Older channels to the west may be traced directly into the bog-mound springs, but the foregoing observations suggest that recharge from the Wyman-Crooked Creeks system not only feeds some of the bog-mound springs but, to a lesser extent, the fault-zone springs as well. The slope of the piezometric surface in the valley (fig. 11) and the orientation of the major fault zone on the east side of the valley (pl. 1) strongly suggests that the fault-zone springs tap flow at depth under head in excess of the 30-foot differential between the highest spring altitudes and the saltpan. The discharge of the Corral Springs and especially the Buckhorn Springs may also consist wholly or in part of waters from drainage areas to the south and east from which waters are channeled through the major fault zone to the springs. Ground-water discharge in the area of Deep Springs Lake itself must be very slow because of the impermeability of the dense plastic carbonate muds which constitute a major part of the central lacustrine deposits.

The western bog-mound springs are probably recharged principally from the Birch and Antelope Creek systems. Small channels around the northwesternmost bog-mounds spring may actually be traced directly back into the Antelope Creek fan. Flash floods from Antelope Canyon enter the playa at the northwest corner. Similarly, the channels of the Payson Canyon system extend into this area. Inflow to the southwest corner of the playa is made exclusively by flash floods from the mountains to the waters and may be readily discerned from the very shallow alluvial channels in the southwest part of the playa.

The pattern of water movement to Deep Springs Lake is summarized in plate 1.

COMPARISON OF DISCHARGE FROM STREAMS AND SPRINGS

An effort was made to obtain some comparative data on the quantities of surface flow in the various water units of the Deep Springs Valley. Initial measurements of streams and spring outflow were made with a pygmy current meter. Such attempts were complicated by an abundance of vegetation, coarse alluvial material, and very shallow depths in many channels. Often, artificial improvements had to be made in the measuring section. The difficulties encountered rendered the resulting data rather poor. Subsequent to the initial measurements, estimates of discharge were made utilizing a folding pocket rule to obtain section parameters and velocities by measurement of differences in head on the face and edge. Such methods give semi-quantitative results at best, but they provide some means of comparing flow in different water units. Data were obtained during each period of fieldwork in the course of the study and are presented in table 3. In addition to measurements on the major streams and springs, an estimate was obtained just northwest of the bog-mound-spring area on a minor flash flood from the lower reaches of Antelope Canyon. The measured discharge was half a cubic foot per second with a maximum velocity of 0.1 foot per second and was composed of 67 percent by weight of solid material.

The data presented in table 3 cannot be taken as truly representative of the range of discharges under so-called normal conditions, but the autumn measurements collected over the 3-year period 1959–61 might be considered as being close to minimum mean values for the total period. Though the valley flow patterns seem to substantiate the observation that water levels of the lake and associated springs respond principally to runoff from the White Mountains through Wyman Creek and its tributaries, the flow estimates lend support to the hypothesis that significant inflow to the lake area from the fault zone springs is derived from sources other than surface waters of the Wyman-Crooked Creeks system. The data of table 3 indicate that during each period of measurement, outflow from the Corral and Buckhorn Springs nearly equaled or exceeded the maximum possible surface water input from the Wyman-Crooked Creeks system. Furthermore, the substantial late spring increase (May, 1961) following a wet year suggests considerable contribution from sources in addition to the valley ground-water reservoir.

HYDROGRAPHY OF THE LAKE

Attempts were also made to obtain data on the hydrologic variations of Deep Springs Lake itself. Special attention was paid to variations in the salinity of the waters and resultant effects on the saline deposits.

Langebein (1961) showed that the salinity of closed lakes and, indirectly, of associated lacustrine salt deposits is strongly dependent on their hydrographic and hydrologic character. Among the most important parameters are the annual net evaporation, mean depth, the variability of lake area, and the geometric shape of the lake basin. Hydrographic data for Deep Springs Lake were difficult to obtain because of inaccessibility and frequent large fluctuations in level.

A staff gage was installed at the northwest corner of the leveed area close to the margin of the perennial brine body (fig. 2, loc. DL2). Staff levels were correlated with lake areas by plane-table leveling on nearby remnant shorelines which resulted from relatively stable intervals during an overall receding phase. Such shorelines are not everywhere continuous or readily apparent but may usually be distinguished by variation of moisture content, color, and texture retained in the surficial lacustrine salt crust. These shorelines are also conspicuous, though discontinuous, features on aerial photographs of the playa. Shorelines roughly corresponding to individual staff levels were sketched from aerial photograph enlargements, and their position compared with snapshots, tape measurements, field notes, and a plane-table map (scale, 1 inch=1,000 ft) of the lake area made during the dry stage of the early September 1960. Even so, it is not altogether certain that the selected shorelines were fully continuous or representative of a specific stage. The resultant approximation of the detailed topography of the lake as indicated by the shorelines is given in plate 2. The most questionable details are in the area around the southern part of the levee, where the faint shoreline traces are interrupted by alluvial wash, low mounds, and the Buckhorn Springs inflow patterns.

Plate 2 indicates that Deep Springs Lake is generally pan shaped, having a flat bottom and very gently sloping sides. Level shots in the central leveed area indicated that the altitude of the saltpan surfaces was about 0.15 feet below the base of the staff (4,917.4 ± 0.2 ft, as controlled by given nearby point altitudes, U.S. Geol. Survey Blanco Mountain quad.). This level is thought to be virtually uniform on both sides of the central leveed area, with the exception of the thrusted edges on salt polygons and the delta around the Buckhorn Springs inflow. However, when seasonal desiccation begins, surface waters retreat to the southern part of the leveed area where inflow from the Buckhorn Springs stagnates. This is indicated by the successive lettered shorelines of plate 2, which represent nearly equal levels but successively smaller area.

For hydrographic calculations, water surface areas at Deep Springs Lake were obtained by planimeter from the plot of shoreline positions. Volumes were computed from water surface areas and corresponding staff-level measurements on the assumption of a trapezoidal cross section for the lake as a whole. Staff-level measurements could not be obtained for low stages because water levels fell below the base of the staff, whose position was determined by its stability and accessibility to an observer. Most of the staff-level data and samples for the relatively high lake stages were collected by William A. Jenkins of the California Division of Highways, inasmuch as water surface levels usually fell below the base of the staff during the author's periods of observation. Volumes for lake stages below the staff base were obtained from crude depth measurements on the salt crust, and area estimates were made from aerial photographs.

Values for hydrographic parameters derived by use of methods just outlined must be viewed with caution. Lake area measurements may be in error by at least 5 percent, depending on the continuity and interpretation of shoreline positions. Such errors are compounded in the calculation of lake volumes, especially at low stage, by deviations from the assumed cross-sectional configuration. Nevertheless, estimates of area and volume were made to show, at least semiquantitatively, the relations among hydrographic properties at Deep Springs Lake.

Derived values of area and volume for stages observed at Deep Springs Lake are given in table 4. The graphical relation between area and volume is shown in figure 13. This plot suggests a linear-logarithmic relation between area and volume for readily measurable stages of Deep Springs Lake. Such a rela-

TABLE 4.—*Data on water surface area, volume, and tonnage of dissolved solids at Deep Springs Lake*

[Data on surface area are considered accurate to within 5 percent, volume and total tonnage to within 10 percent]

Sample [1]	Date collected	Staff reading (feet)	Water surface area (acres)	Volume (acre-ft)	Concentration (ppm)	Total tonnage dissolved solids [2]
DL13	10/3/59	------------	260	13.0	310,000	7,840
2D	1/30/60	0.97	518	450	67,400	43,040
2E	2/4/60	.75	459	350	81,400	40,850
2F	2/20/60	.56	434	262	93,300	35,480
2G	2/29/60	.42	402	176	108,000	27,950
2H	3/19/60	.24	366	110	124,000	20,330
2I	3/31/60	.04	314	57	122,000	10,460
2C	5/5/60	Trace.	284	27	224,000	9,840
2B	6/5/60	------------	265	13.25	295,000	6,750
2K	1/14/61	1.20	563	558	21,900	16,660

[1] With the exception of DL13, all samples listed were collected at or near the staff gage (loc. DL2). The first sample was collected toward the center of the saltpan near locality DL13 (fig. 2) at very low stage.
[2] Calculated on the basis of the equation of Langbein (1961, p. 8): 0.00135×volume in acre-feet×concentration in parts per million by weight×density.

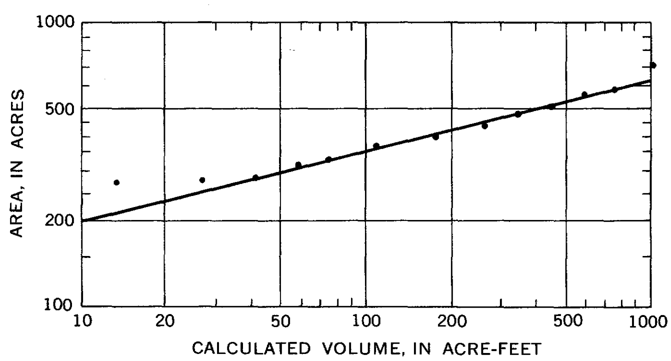


FIGURE 13.—Derived relation of water surface area to volume for Deep Springs Lake, October 1959 to Sepetember 1961.

tion was employed by Langbein (1961, p. 5) in calculating the variation in stage of much larger closed lakes. For comparison with closed-lake data computed by Langbein (1961, table 1), response time and area variation were also calculated for Deep Springs Lake.

Response time (coefficient K) is the measure of time, in years, that a closed lake takes to react to inflow fluctuation. Response time is defined by the equation (Langbein, 1961, p. 6) $\frac{V'' - V'}{E(A'' - A')}$, where V'' (in acre-ft) is the value for a high lake stage, V' is a low lake stage, A'' and A' correspond to water surface areas (in acres), and E is net evaporation in feet per year (4.25 for Deep Springs Lake), which is assumed constant annually. Response time was computed utilizing volume and area data (table 5) for the highest level readily reached by the lake (based on textural and mineralogic criteria) when water covers about two-thirds of the unvegetated playa area and for the lowest level at which a coherent body of water still exists on the saltpan. Obviously, these limits are arbitrary.

Lake area variability may be characterized by the coefficient of variation, which is calculated from the equation (Langbein, 1961, p. 6)

$$U=\frac{0.26n}{D}\sqrt{\frac{E(A_T/A_L)K}{(2+1/K)}}$$

where n is the exponent in the relation of lake area (A_L) to volume ($A_L \sim V_T^{26}$ for Deep Springs Lake), D is mean depth, and A_T is tributary area. Inasmuch as the mean depth for the period of study apparently remained just below the level of the staff base, the coefficient of area variation was computed for a mean depth of 0.1 foot. The results of these computations and the basic data from which they were made are compared with data of some of the more saline of Langbein's examples given in table 5.

The calculated response time, K, is somewhat less than that observed for the well-documented recession of lake waters during early 1960 (to be discussed on p. A24) but is considerably more than that observed in October 1959 or September 1961. However, fluctuations resulting from protracted annual spring runoff might be expected to be of longer duration than fluctuations based on isolated, aperiodic runoff. The values of K and U given in table 5 should be close to a reasonable average for the two types of inflow events. As these values suggest, Deep Springs Lake more nearly resembles the typical small playa of the Great Basin than do any examples given by Langbein, who noted the lack of data on such lakes.

Actually, the period of study in the Deep Springs Valley was one of somewhat below-normal precipitation throughout the region. Local residents were able to recall times during the past when lake waters extended beyond the present playa margins. The highest

TABLE 5.—*Hydrologic data from Deep Springs Lake as compared with some major closed lakes cited by Langbein* [1]

[Explanation of symbols: A_T, tributary drainage area in square miles; E, net evaporation in feet per year; U, coefficient of lake area variation; K, response time in years; D, mean depth in feet; A_L, lake area in square miles]

Lake	A_T	E	U	K	Salinity		D	A_L
					Date	Parts per million		
Deep Springs	200. 85	4. 25	7. 63	0. 32	Jan. 1961	21, 900	1. 05	0. 88
					Oct. 1959	310, 000	. 1	. 41
Great Salt	21, 000	2. 70	. 125	9. 0	1877	138, 000	18. 0	2, 200
					1932	276, 000	13. 0	1, 300
Owens	2, 900	5. 00	. 10	10	1876	60, 000	24. 0	105
					1905	213, 700	11. 0	76
Abert, Ore	900	2. 50	. 5	6	1902	76, 000	5. 0	50
					1956–59	20, 000	10. 0	60
Lake Eyre, Australia	550, 000	7. 0	2. 5	1. 5	1950	40, 000	8. 5	3, 100
					1951	240, 000	2. 8	740
Tuz Golu, Turkey	4, 400	2. 4	. 5	1	1959	250, 000	2	650

[1] Langbein (1961, p. 18, table 1).

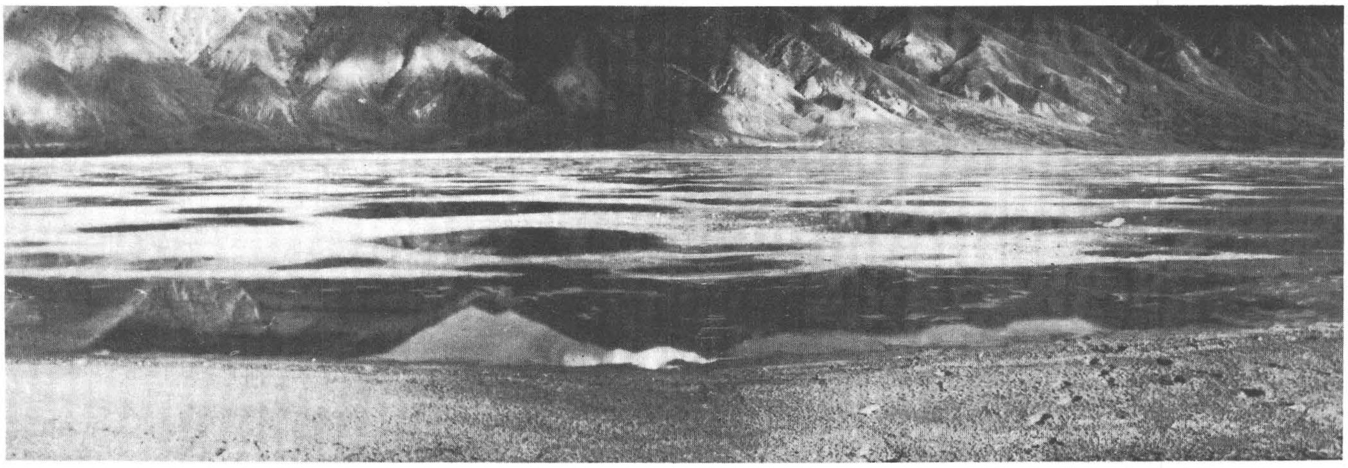


FIGURE 14.—Deep Springs Lake at low stage prior to total desiccation, August 1961, as seen looking southeast from the staff gage, locality DL2. Footprints give scale.

stages are normally recorded after periods of maximum winter precipitation, February through April, and are the result of snowmelt and associated runoff in the White Mountains. Marked short-term level rises result from intense thunderstorms in the surrounding mountains or directly over the lake in the late summer. The recession of lake waters usually takes place in a sporadic fashion as evaporation fluctuates with weather conditions; rapid late-summer level rises are usually followed by very rapid recession. The migration of receding shorelines is highly irregular locally, and on the flats just outside the leveed area, remnant puddles are left behind by receding lake waters in the most inconspicuous topographic depressions. Fractionation of lake waters into a group of isolated puddles becomes more conspicuous in the salt crust areas, where polygonal fracturing and thrusting of fractured crustal edges gives rise to sharp microrelief and promotes separation of lake waters into myraid shallow ponds prior to complete desiccation (fig. 14). At low stages, lake waters are commonly divided into three units separated by the physical features of the playa and density gradients due to variable salt concentrations. These units include: (1) inflow waters which may become somewhat stagnant and accumulate near the inflow channels of the bog-mound springs, Corral Springs, and particularly the Buckhorn Springs, (2) standing surface waters in the central or south-central part of the leveed area, and (3) a perennial brine body which occurs interstitially in the salt crusts and in a few stagnant ponds around large crustal fractures or in the trough just inside the levee. Mixing of these water units takes place more slowly as the stage rises, although circulation can be aided considerably by wind action. At lowest stage, surface

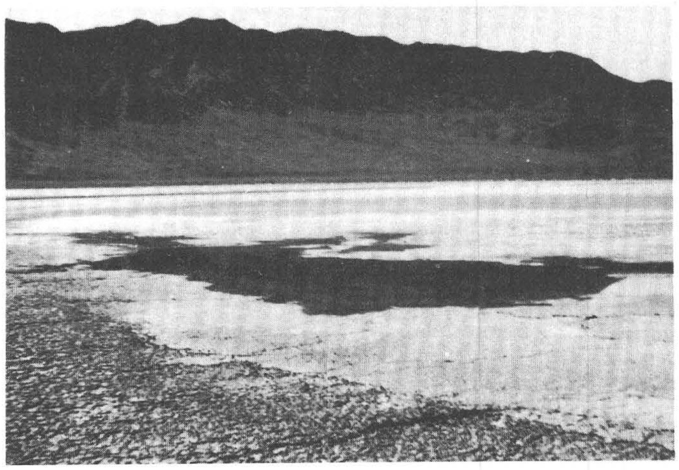


FIGURE 15.—Stagnant pool of highly colored brine, north side of Deep Springs Lake.

water at Deep Springs Lake is confined to a few small stagnant pools of dense amber-colored brine (fig. 15).

In spite of the problems of inhomogeneity in surface waters, as just outlined, an attempt was made to find what relations might exist between volume and the amounts of salts in solution in the lake waters. Samples were collected in the northwestern part of the leveed area, near the staff gage, where presumably at least the inhomogeneities resulting from inflow source would be held to a minimum.

Data on the relation between lake volume and tonnage of salts in solution at Deep Springs Lake, have been given in table 4, and are presented graphically in figure 16. It should be emphasized that whereas the points corresponding to the receding phase of the curve are derived from consecutive samplings, the other sampling points are not. For example, the sample of January 1961 was collected near the peak of a rising

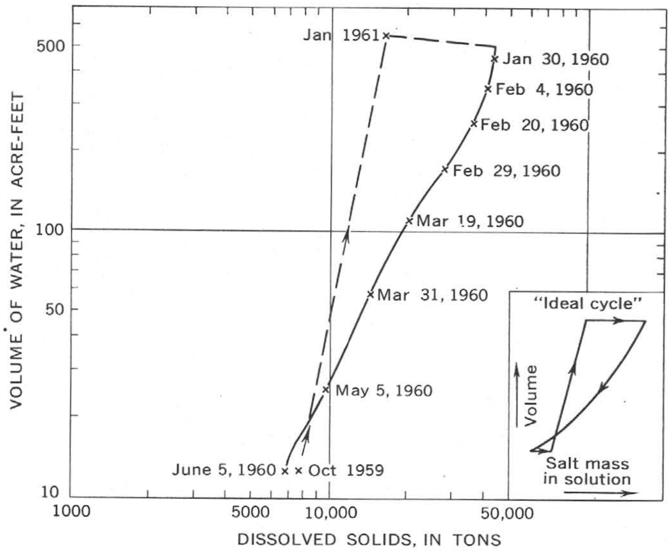


FIGURE 16.—Relation of lake volume to total dissolved solids at Deep Springs Lake for samples collected from October 1959 to January 1961. The insert shows the "ideal cycle" of Langbein (1961, p. 9).

stage, but the area became inaccessible shortly thereafter. In spite of the lack of samples during rising and initially receding stages, the dashed lines are thought to approximate the trends at Deep Springs Lake. The trend of the recession curve is rather regular, although it must be remembered that the salt tonnage in solution is derived in part from the volume data, and small errors in the volume values will have a large effect on the continuity of the volume-tonnage relation.

It should be noted in connection with figure 16 that salt precipitation apparently did not take place until just prior to March 19. Thus, it is apparent that nearly 50 percent of the total dissolved solids in the lake waters was lost from solution prior to solid precipitation.

The loss of dissolved salts during fluctuations in the volume of saline lakes has been discussed by Langbein (1961, p. 10). To explain such losses, he has proposed entrapment of salts in bordering sediments and removal of salines from dried parts of the lakebed by wind action. To test the first part of Langbein's hypothesis, samples of the sediment from Deep Springs Lake were leached by repeated distilled-water washing. Data from the playa sediments outside the leveed area are incorporated with the stratigraphic information given in figure 5. The sediments were found to contain more than 30 percent material which was readily water soluble. Higher values were obtained for muds within the leveed area. Some samples contained more than 17 percent readily soluble material in the absence of a detectable saline phase, as determined by microscope or X-ray techniques. In fact, some of the muds of Deep Springs Lake contain more than 80 percent material of precipitate origin (including alkaline earth carbonate as well as soluble salts).

The determination of the solubles content of muds of Deep Springs Lake indicates that dissolved salts may indeed be incorporated with bordering sediments as the surface waters recede. The stirring up of carbonate muds in the lake waters by wind action has been directly observed. The deposit of such sediment, one-tenth of a foot deep and containing 25 to 30 percent soluble salts on the ½-square mile area left exposed by recession of lake waters from January 30 to March 19, 1960, could readily account for the loss of 30,000 tons of dissolved salts (fig. 16) from the receding waters in that time.

The effect of wind action on the removal of salts from central Deep Springs Lake is considered to be relatively minor. Although the wind frequently stirs up clouds of salt-laden dust from efflorescent crusts on the west side of the playa (fig. 17), the well-indurated crusts of the central lake area do not contribute much to aeolian transport. Observations throughout Deep Springs Valley suggest that most wind-transported material either remains within the drainage area or that the material lost is largely replaced by similar material carried into the basin.

Another mechanism for the removal of salts available for solution in Deep Springs Lake waters is the burial of saline muds and crusts by an influx of clastic sediments from one of the large ephemeral drainage systems. There is little evidence, however, that in recent years such an influx has carried as far as the saltpan.

Langbein (1961, p. 8) stated

> that the variation of salinity with lake volume is not directly a simple matter of concentration or dilution of a fixed mass of salts in a changing volume of lake water. The mass of salts in solution appears to decrease with contraction in lake volume,

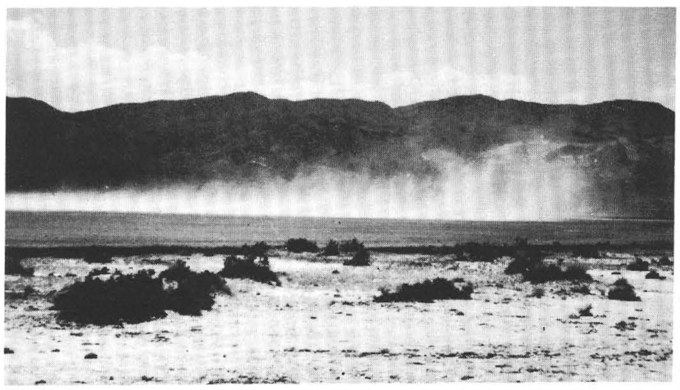


FIGURE 17.—Salt-laden dust being blown off the west side of the Deep Springs Lake playa.

but although the mass in solution may increase with a recovery in lake volume, the total mass may be less than before the recession began. Completion of the cycle is accomplished during slow accumulation of salts carried in solution by the influent streamflow.

Langbein (1961, p. 9) also presented a schematic volume-salinity cycle and suggested that "actual volume-tonnage curves * * * may never describe the idealized cycle * * * but may consist of a series of zigzag curves of which each part is related to one of the limbs of the hypothetical cycle."

The volume-tonnage relations for Deep Springs Lake illustrated in figure 16 appear to follow Langbein's schematic cycle closely, but if one considers the potential errors in the volume data, the agreement in detail could be simply fortuitous. The only apparent disparity between the relations shown for Deep Springs Lake and Langbein's ideal cycle is in the lower part of the recession curve. This disparity could be explained as an effect of direct salt precipitation from the lake waters.

In most closed lakes the proportions of constituents in solution remain virtually uniform throughout any decrease in volume and increase in concentration. This uniformity exists because of peripheral effects and lack of any significant precipitation, within the receding waters, of salts intermediate in solubility between alkaline earth carbonates and sodium chloride. As the concentration increases, each further reduction in volume is associated with a proportionately greater decrease of the total salts in solution. At Deep Springs Lake, however, the precipitation of sodium sulfate—the major salt in solution at high stage—apparently stabilizes the ratio of further volume-tonnage decrease. This would explain the linearity of the lower part of the curve representing the receding phase (fig. 16).

The effect of sulfate precipitation is more apparent in the relation of volume to concentration (fig. 18). During the receding phase, concentration increased regularly but slowly until precipitation began. The concentration remained constant until some sort of equilibrium with the precipitate was reached, and then increased rapidly, as the lake decreased further in volume, until complete desiccation was reached. This rapid rise was not noted by Langbein (1961, p. 10) in his volume-concentration relations hypothesized for a playa lake.

For the filling cycle of a playa lake, Langbein (1961, p. 10) hypothesized that fresh-water inflow will cause a rapid increase in dissolved-solids concentration owing to solution of crustal salts until the point of saturation is reached and then a gradual decrease in concentration as further increase in lake volume takes place after the crust is dissolved. These general trends

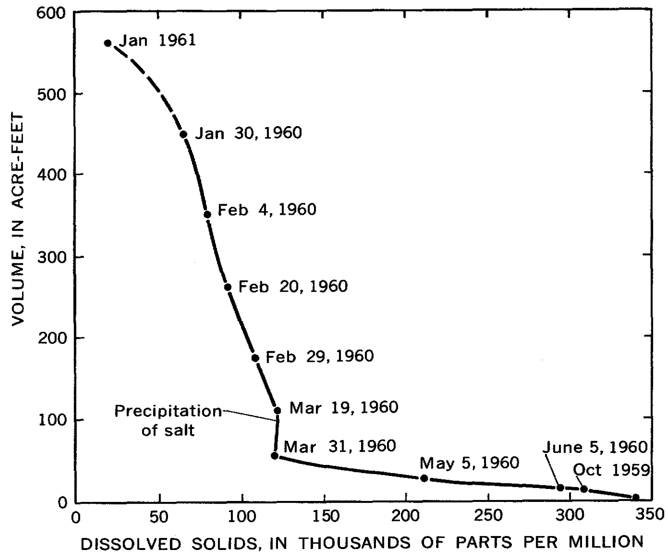


FIGURE 18.—Relation of surface-water volume to dissolved-solids concentration for a receding phase of Deep Springs Lake. The salt precipitated is mirabilite ($Na_2SO_4 \cdot 10H_2O$).

were noted at Deep Springs Lake, but the magnitude of an initial rise in concentration in the inflow waters was strongly dependent on the abundance of the different salts and their solubilities at the various points of inflow. The role of such factors is more evident in the section discussing the saline mineralogy of the lake area.

HYDROCHEMISTRY

ANALYTICAL METHODS

The procedures employed in the collection and analysis of waters from the Deep Springs Valley were those in general use by the U.S. Geological Survey. Though some preliminary measurements were made in the field, most of the analytical work was done in the laboratory by Shirley L. Rettig, advised by H. R. Feltz. The reproducibility and limits of accuracy in the methods are generally the same as those set forth by Rainwater and Thatcher (1960) and Hem (1959), although special problems exist in the analysis of brines, and potential sources of error were greatly magnified when compared to determinations in more dilute waters. Some of the problems encountered and special techniques employed are discussed in greater detail below.

SAMPLING

A variety of containers were utilized in the sampling of waters. Initially, 1-liter Pyrex bottles with paraffin inner linings and Teflon-coated stoppers were used. Although contamination from glass was thus held to a minimum, these containers were too bulky and fragile for convenient field use. Further, the paraffin linings tended to crack and flake. These problems outweighed the possibility of gaseous leakage in polyethylene bottles, which were employed in subsequent work.

Where there was a distinct problem in retaining gaseous components present in solution at time of sampling, "citrate-of-magnesia" bottles were used initially. However, the pressure seals of these bottles were frequently defective, and greater success (plus less contamination during storage) was achieved by use of gas analysis tubes or fixatives.

Waters from the Deep Springs Valley were sampled during five periods of field observation from October 1959 to September 1961. As previously mentioned (p. A21), some additional samples taken from Deep Springs Lake at high stages during early spring were collected by W. A. Jenkins of the California Division of Highways. Bottles were simply dipped into waters of sufficient depth; but for very shallow waters, samples were taken by bulb suction, and care was taken not to pick up solids. During the period of study, up to 6 samples were collected from each major source of inflow to the lake, usually at the time that discharge estimates were made. For both inflow and the lake, samples were collected to establish point-to-point compositional variation as well as fluctuations with time.

Temperature measurements were made at the source of all samples, but some early measurements were in error because of a faulty thermometer.

pH AND ALKALINITY

It is well known that significant changes in the pH-alkalinity-carbon dioxide balance commonly take place between the time of water sample collection and analysis (Hem, 1959). These changes are brought about not only by inorganic loss or gain of carbon dioxide but also by subsequent variations in temperature, oxidation, precipitation, or the activity of microbiota. These factors are most pronounced where there are marked changes in conditions between the field and laboratory. Therefore, the pH and alkalinity for a majority of samples from the Deep Springs Valley were measured in the field.

The pH was measured electrometrically in the field by using several standard instruments in the course of the study. Measurements were normally made directly at the point of collection. The electrodes were kept as moist as possible and were standardized against sodium carbonate (pH$\sim$10), borax (pH$\sim$9), and phosphate (pH$\sim$7), buffers. Suitable corrections were made for temperature, which was brought to within 5° C of the sample wherever possible. Both electrode fatigue and instrumental drift were noted, and empirical corrections were applied by frequent buffer standardization. Nearly all measurements were reproducible and are considered accurate to at least $\pm$0.15 pH unit, depending in many cases on temperature changes between collection and measurement on site. Sodium-ion and junction-potential effects expected in highly concentrated waters were found to be within the limits of error stated above by checking against saturated standard sodium carbonate solutions (Bedekar, 1955). These problems are discussed more fully by Bates (1954).

Alkalinity measurements in the field were made potentiometrically and were usually made immediately after the determination of pH. Quantitative distribution of CO_3^{-2} and HCO_3^{-1} were computed from the amount of standard sulfuric acid titrant necessary to reach the commonly assumed end points at pH of 8.2 and 4.5. These end points probably are sufficiently close to the true end points for waters having total alkalinity ranging from 100 to 450 ppm; therefore, no significant error is introduced. (For a detailed discussion of this problem, see Barnes, 1964.) However, for waters having higher alkalinity, errors resulting from variation of the end points from these standard values are to be expected. Because of the large amounts of titrant required and the expected interference of other weak acid radicals—such as borate, phosphate, and silica—alkalinity of the more concentrated waters from Deep Springs Lake was not measured in the field. Instead, an attempt was made to preserve the distribution of ionic species through use of gas analysis tubes. Use of this technique was only partially successful because of temperature variation and microbiotic activity. Comparison of gas sample tube analyses with field measurements in dilute waters suggests potential errors of more than 5 percent. Closer agreement was obtained where very small amounts of chloroform were added to the gas sample tubes to hinder biotic action, as suggested by K. V. Slack (oral commun.). A comparison of field data for pH and alkalinity with that determined in the laboratory on samples from bottles and gas tubes is presented in table 6. For waters having a total alkalinity of less than 500 ppm, the gas tube analyses reproduced field data to within 5 percent, but differences between field data and laboratory data on bottle samples exceeded 20 percent. For waters whose alkalinity is greater than 500 ppm, gas tube and standard laboratory data differed by even greater amounts. At high concentrations and pH's, there is a significant contribution to the total alkalinity by borate, phosphate, and silica, but inasmuch as there is no satisfactory data on the state of these elements in such solutions, no correction has been attempted.

In order to obtain additional data characterizing the carbonate-bicarbonate-carbon dioxide system in the Deep Springs waters, total carbon dioxide content of several samples was determined in the laboratory utilizing a manometric gas evolution technique modified from that described by Pro, Etienne, and Feeny (1959). These values, which were reproducible within 5 percent for highly concentrated waters, are also presented in table 6. Potential error in this method increases significantly at concentrations of less than about 800 ppm because of very small changes in manometer readings and difficulties in the calibration of the volume of the system. Bicarbonate concentrations calculated from the data on total carbon dioxide sometimes more nearly approximate field measurements than standard laboratory determinations, but the results are not consistent, and total range of variation in the three values may still exceed 20 percent.

Table 6 presents total carbon dioxide content data for many samples both as determined manometrically and by conversion of the titrated CO_3^{-2} and HCO_3^{-1} values, including free carbon dioxide calculated as suggested by Rainwater and Thatcher (1961, p. 139) for waters containing less than 800 ppm in total alkalinity. The calculations of free carbon dioxide are based on the ionization constant for carbonic acid at 25° C and are uncorrected for ionic strength. Such errors are probably significant only for waters which have a temperature differential of 10° C and a pH of less than 7.5. Nevertheless, significant differences between the calculated and measured values for total carbon dioxide concentration are readily apparent. Calculated values of total carbon dioxide content are substantially lower than those measured in waters of low concentration and higher than those measured in waters of high concentration.

In view of the difficulties just outlined, data on the carbonate-bicarbonate-carbon dioxide system at Deep Springs Lake cannot be considered to be more than semiquantitative. From information currently at hand, it is not possible to establish the cause of all anomalies.

MAJOR-ELEMENT ANALYSIS

Calcium and total hardness were determined by EDTA titration, and magnesium was determined by difference. The values reported for calcium and magnesium in the highly concentrated lake brines are probably no more accurate than an order of magnitude, because of very small concentrations and the possibility of large interferences. The determination of calcium and magnesium in dilute waters from the valley may also be in error because of the precipitation of alkaline earth salts between time of collection and analysis. (For particular reference to Birch Creek, see Barnes, 1962.)

Alkalies in Deep Springs waters were determined by direct-reading flame photometry. Standards were run coincidentally with sample, and multiple readings were frequently made. The large dilutions necessary to analyze concentrated waters may have introduced small errors, but results were reproducible within a few percent. Gravimetric analysis for sodium in a few samples gave values within 10 percent of those determined by flame photometry.

Sulfate and chloride were determined gravimetrically as barium sulfate and silver chloride, respectively, in the concentrated lake waters. Sulfate was analyzed by the thorin method and chloride was determined volumetrically by Mohr titration in the more dilute inflow. These determinations are probably the most reliable of all data on Deep Springs Valley waters.

SULFIDE

Sulfide in some samples was preserved by collection in gas analysis tubes or by adding approximately 2 grams of zinc acetate to the sample at the time of collection. The iodometric method was utilized for both types of analysis.

PRESENTATION AND INTERPRETATION OF DATA

The analytical results for 125 samples of water taken from the Deep Springs Valley during the 2½-year period of study are presented in table 7. Sample numbers correspond to the points of collection shown on the location maps (figs. 2 and pl. 1). The analyses are grouped roughly with decreasing distance from the perennial brine body at Deep Springs Lake. Thus, they are listed as: (1) streamflow including springs immediately within major drainage units, (2) wells, except for the old artesian well adjacent to the lake area, (3) springs of the lake area including the artesian well spring, (4) ponds, (5) surface waters of the lake area, and (6) interstitial brine. Except for the closed sag pond, all water outside the immediate playa area has been considered as inflow to Deep Springs Lake. Because of the analytical problems discussed previously, a strictly quantitative evaluation of the data is impossible, but significant qualitative relations are apparent.

TABLE 6.—*Total* CO_2, pH, *and alkalinity for samples of water from the Deep Springs Valley*

[Sample numbers refer to localities shown in fig. 2 and pl. 1. Only samples for which more than the normal laboratory measurement of alkalinity was made are included. Concentrations are in parts per million. Symbols: Asterisk (*) indicates data from gas analyses tube; n.a., indicates data not analyzed]

Sample	Total CO_2			pH		Alkalinity			
	Measured	Calculated from—		Field	Laboratory	HCO_3		CO_3	
		Laboratory data	Field or gas tube* data			Field	Laboratory	Field	Laboratory
DSH1		184	193	7.50	7.8	250	245		
DSS1		202		7.45	7.7		268		
DL1	21,500	24,400			9.15		12,600		20,900
2	19,500	23,250			9.4		12,400		19,500
2B	8,730	8,710			9.3		5,330		6,640
2C	6,890	6,800			9.4		3,770		5,570
2D	2,750	2,740			9.6		1,680		2,080
2E	2,760	2,980			9.5		2,000		2,100
2F	2,800	2,870			9.5		2,150		1,800
2G	3,260	3,410			9.3		3,160		1,540
2I	3,450	3,630			9.5		2,390		2,600
2J		18,400	* 18,600	9.58 * 9.60	9.4	 * 7,860	1,690	 * 17,600	23,400
2L					9.5		n.a.		22,100
2P		17,500			9.35		1,210		22,700
3A	19,000	24,700		9.30	9.35		15,300		18,400
3B	18,700	21,900					11,100		18,900
13	19,900	21,300		10.20	9.3		8,720		20,500
14	5,220	2,970		10.00	9.35		371		3,680
21	12,600	13,600		9.65	9.2		5,320		14,700
22	22,700	24,700		9.70	9.1		16,000		18,200
24	17,500	17,100		9.70	9.15		8,730		17,400
27C		135		7.23	7.3		169		.0
29A	10,300	10,700		9.75	8.8		11,200		3,630
30A	8,190	7,190		9.40	8.85		7,000		2,920
31	618	578		8.85	8.1		788		
32	218	251	211	8.35	7.7	290	334	9.8	.0
32C		206		8.79	7.8		304		
33		190	142	8.45	7.2	195	232		.0
33C		148		7.78	7.5		192		.0
34	133	152	157	7.75	8.4	210	209	.0	.0
34C		158		8.01	7.7		210		.0
35	256	274	172	7.55	7.4	225	350	.0	.0
35C		165		8.13	7.8		220		.0
36	280	349	185	7.65	7.7	245	464		
36D		176	* 180	* 8.6	7.6	* 216	232	* 20	.0
36E		178	177 * 175	7.97 * 7.75	7.65	240 * 233	236	.0	.0
37	258	317	241	7.70	7.80	320	426		
37C		215		8.11	7.8		291		
38A	1,080	1,310	771	8.10	7.80	1,050	1,750	.0	.0
38B		1,070	676	8.35	7.20	930	1,300	.0	.0
39A	50,600	73,100		10.20	9.7		46,800		53,600
39B	560	658	397	7.90	8.1	535	896	.0	.0
39C	29,500	30,900			9.6		18,700		23,800
39D–E		55,400	* 48,800	10.18 * 10.10	10.1	 * 10,400	12,500	 * 56,300	63,300
39G		22,200			9.7		11,000		19,500
40A	730	748	792	9.05	8.9	970	919	123	113
40B		2,340		10.15	9.2		2,220		1,020
41A	190	217	215	7.70	7.8	285	291	.0	.0
41F		212	228 *212	8.08 *7.9	7.6	310 *289	278	.0 *.0	.0
42	344	260	233	7.10	7.1	275	308	.0	.0
43	111	161		7.65	7.9		218		.0
44	110	143		8.80	7.8		192		.0
44B		191		7.93	7.5		248		.0
45		263	260	8.20	8.4	350	354	4.9	8
CC47	87	132		8.10	7.9		179		.0
WC47	186	212		8.10	7.7		282		.0
47E		195	201	8.60	7.9	265	263	12.8	.0
DL48A3		143	*135	*8.9	7.7	*154	190	*33	.0
48B3		353	*372	*8.7	7.3	*476	441	*38	
48A4		147		8.36	7.7		196		.0
48B4		279		8.61	7.75		371		.0
48C		186	184	8.29	8.0	253	252	.0	.0
51A		22,900	*20,500	9.58	9.7	*2,430	9,360	*25,500	22,100
BC70	73	104	*110	8.52 *7.6	8.0	*144	142	*.0	.0
AS71A		160	*158	7.22 *7.7	8.1	*210	218	*.0	.0
71C		151	153	7.89	7.8	207	203	.0	.0
DL73A	132	152	*160	8.01 *8.1	7.6	*218	200	*.0	.0
73C	152	157	156 *158	8.26 *8.0	7.65	213 *215	208	.0 *.0	.0
73D		325	388	7.95	7.8	525	436	.0	.0
78		145	137	8.84	7.8	188	194	17.2	.0
AS79		143	*145	8.69 *8.0	8.0	*200	197	*.0	.0

TABLE 7.—*Analyses of waters from the Deep Springs Valley, 1959–61*

[Sample numbers refer to localities shown in fig. 2 and pl. 1. Locations and samples are listed in approximate geographical order from valley margins to the playa area. Concentrations are in parts per million]

Location	Sample	Date of collection	Silica (SiO_2)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate (HCO_3)	Carbonate (CO_3)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F)	Boron (B)	Total dissolved solids	pH
Streamflow															
Crooked Creek:															
Just above confluence with Wyman Creek.	CC47B	Oct. 20, 1959	20	46	6.3	20	7.0	179	0	21	4.5	0.2	0.06	213	7.9
	47	June 7, 1960	28	42	5.8	9.8	2.3	150	0	18	2.5	.2	.01	183	7.8
	47D	May 20, 1961	24	46	5.1	11	1.8	170	0	20	4.0	.2	.07	172	7.6
Wyman Creek:															
Just above confluence with Crooked Creek.	WC47	Oct. 21, 1959	16	69	17	12	6.0	282	0	35	2.0	.2	.03	296	7.7
	47B	June 7, 1960	20	58	14	9.4	3.0	238	0	41	2.0	.2	.01	265	7.9
	47C	Sept. 1960	27	85	12	12	8.5	286	0	43	1.0	.3	.20	330	7.6
	47E	May 20, 1961	22	69	16	9.5	2.7	265	0	40	2.8	.1	.10	291	7.9
At gage just above sediment trap.	56A	June 7, 1960	21	59	13	9.5	2.9	226	0	37	2.5	.3	.02	256	8.0
	56B	Sept. 13, 1960	23	65	18	10	3.5	264	0	40	3.0	.2	.03	293	8.2
Spring beneath inclusion riddled outcrop, ½ mile above confluence with Crooked Creek.	57	June 7, 1960	16	55	11	7.3	2.3	198	0	36	2.0	.1	.00	228	7.6
¼ mile below Mill Canyon.	58	do	19	56	11	7.3	2.3	198	0	37	1.0	.2	.02	232	7.8
Source near Roberts Ranch.	59A	Sept. 27, 1959	20	53	16	12	6.0	234	0	17	4.5	.1	.08	244	7.5
	WS59B	June 7, 1960	37	112	19	18	3.0	366	0	60	5.0	.3	.04	435	7.5
Birch Creek:															
Near point of disappearance (other points on fig. 23 from Barnes, 1962).	BC70	Sept. 11, 1960	22	42	23	13	3.3	142	0	104	4.0	.7	.02	282	8.0
Antelope Creek:															
Near termination	AS79	Aug. 16, 1961	19	44	27	10	3.2	197	0	70	5.5	.1	.00	302	7.9
Antelope Springs:															
Southern spring 1	71	Sept. 11, 1960	20	56	26	13	4.0	218	0	93	6.0	.4	.01	325	8.1
	71C	Aug. 16, 1961	20	56	24	12	3.6	203	0	90	8.5	.3	.00	314	7.6
Northern spring 2	72	Sept. 11, 1960	21	55	25	12	4.3	202	0	100	5.0	.7	.03	323	7.9
Wells															
Highway Station	DSH1	Oct. 4, 1959	16	69	16	14	7.0	245	0	45	5.0	0.1	0.03	293	7.8
	2	June 9, 1960	21	64	15	22	10	230	0	88	6.0	.2	.07	339	8.0
	3	May 20, 1961	24	67	16	10	2.5	252	0	39	6.5	.1	.10	317	7.3
Deep Springs School irrigation well.	DSS1	Oct. 8, 1959	18	76	17	15	7.0	268	0	57	4.1	.1	.05	326	7.7
Stock well, Central Valley.	DL43	Oct. 20, 1959	35	42	10	88	15	218	0	121	10	2.5	.22	431	7.9
	43B	Oct. 10, 1961	39	35	12	55	3.3	186	0	100	5.1	2.5	-------	344	8.1
Stock well, west of Deep Springs Lake.	44	Oct. 20, 1959	26	32	31	45	12	192	0	133	12	1.4	.11	388	7.8
	44B	Aug. 16, 1961	73	49	32	26	10	248	0	109	9.5	1.1	.10	431	7.4
Springs of the Lake Area															
Bog-mound springs:															
Northwest bog-mound spring.	DL27A	Oct. 7, 1959	39	40	22	37	10	230	0	105	3.0	1.8	0.14	371	7.5
	27AS	Feb. 1960	86	46	26	39	9.0	270	0	98	9.0	1.8	-------	447	8.1
	27C	Aug. 14, 1961	48	38	22	28	7.0	169	0	100	9.0	1.8	.05	337	7.0
Western bog-mound pond.	78	do	42	28	16	39	12	194	0	59	8.5	1.4	.03	303	7.7
Central bog-mound spring.	73A	Sept. 12, 1960	48	37	13	18	7.6	200	0	30	4.0	.4	.06	257	7.6
	73B	May 17, 1961	48	39	14	20	7.7	200	0	30	7.0	.3	.14	267	7.4
	73C	Aug. 15, 1961	46	36	17	15	7.2	208	0	30	4.0	.2	.41	258	7.55
Central bog-mound marsh.	73D	do	56	78	31	26	11	436	0	27	4.0	.8	.34	484	7.6
Eastern bog mounds:															
Artesian well spring	41A	Oct. 19, 1959	41	14	14	40	45	291	0	1.0	4.0	1.2	.33	304	7.8
	41E	May 28, 1960	46	16	13	35	48	288	0	6.6	.0	------	.43	307	8.2
	41F	Aug. 12, 1961	46	14	14	36	46	278	0	8.0	3.8	1.1	.34	447	7.4
Northeast bog mound.	42	Oct. 19, 1959	42	19	18	40	46	308	0	17	5.0	1.2	.36	341	7.1
Corral Springs:															
Spring 1	48A	Mar. 28, 1959	46	17	14	44	16	186	4.8	36	3.6	.4	.10	273	8.3
	48A1	Oct. 28, 1959	41	18	10	48	20	202	0	40	6.0	.4	.05	284	7.8
	48A2	Sept. 9, 1960	42	17	9.1	46	16	188	0	37	6.5	.4	.09	267	7.9
	48A3	May 22, 1961	39	18	10	45	15	190	0	38	7.0	.3	.13	264	7.7
	48A4	Aug. 15, 1961	43	16	11	44	15	196	0	41	6.8	.3	.03	273	7.8
Spring 2	48B	Mar. 28, 1959	61	10	7.5	234	48	402	78	107	7.2	.9	1.0	754	9.1
	48B1	Oct. 1959	-------	6.4	2.9	230	55	420	14	193	57	.9	.87	774	8.4
	48B2	Sept. 9, 1960	29	8.0	2.4	280	52	496	0	143	87	1.0	1.0	848	7.9
	48B3	May 22, 1961	26	6.0	2.4	205	50	441	0	82	56	.8	.68	698	7.3
	48B4	Aug. 15, 1961	32	8.0	5.6	222	34	371	0	161	77	.8	.06	770	7.65
Spring intermediate between Nos. 1 and 2.	48Int	May 22, 1961	39	14	8.6	56	15	192	0	44	7.0	.4	.11	271	7.9
Outflow from Nos. 1 and 2 combined.	48C	Aug. 15, 1961	39	13	9.2	112	23	252	0	82	30	.5	.15	278	8.0
Corral Spring inflow to the lake.	50	Mar. 28, 1959	43	16	18	769	118	661	113	593	385	1.0	2.6	2,390	9.0
Buckhorn Springs:															
Springs 1–5	33	Oct. 13, 1959	41	40	20	60	18	232	0	121	1.5	.5	.09	416	7.2
	33B	May 22, 1961	45	39	21	48	10	192	0	119	10	.5	.14	387	7.4
	33C	Aug. 12, 1961	44	38	22	44	9.0	192	0	120	8.0	.5	.07	380	7.35
Springs 6–7	34	Oct. 13, 1959	37	34	21	75	22	209	4	138	11	.6	.14	446	8.4
	34B	May 22, 1961	42	39	19	68	13	216	0	134	10	.5	.15	431	7.4
	34C	Aug. 13, 1961	41	34	22	65	14	210	0	135	9.5	.5	.07	424	7.6
Springs 8–10	35	Oct. 13, 1959	38	35	20	88	18	350	0	127	2.0	.6	.12	501	7.4
	35C	Aug. 13, 1961	41	32	23	68	12	220	0	140	12	.6	.07	436	7.65

TABLE 7.—*Analyses of waters from the Deep Springs Valley, 1959–61*—Continued

[Sample numbers refer to localities shown in fig. 2 and pl. 1. Locations and samples are listed in approximate geographical order from valley margins to the play area. Concentrations are in parts per million]

Location	Sample	Date of collection	Silica (SiO_2)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate (HCO_3)	Carbonate (CO_3)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F)	Boron (B)	Total dissolved solids	pH
Springs of the Lake Area—Continued															
Buckhorn Springs—Con. Spring group 11	DL36	Oct. 13, 1959	34	35	19	100	22	464	0	62	2.0	0.6	0.15	504	7.7
	36C	Sept. 1960	39	36	27	89	16	233	0	178	16	.8	.15	517	7.8
	36D	May 22, 1961	41	38	24	82	15	232	0	188	14	.8	.20	517	7.6
	36E	Aug. 13, 1961	39	35	25	88	13	236	0	181	16	.6	.08	513	7.5
Buckhorn Springs inflow to the lake.	38A	Oct. 14, 1960	77	32	177	450	100	1,750	0	325	125	2.7	.78	2,150	7.8
	38B	do	68	18	130	605	110	1,300	0	691	250	2.8	1.6	2,520	7.2
	38D	Aug. 13, 1961	52	62	42	154	28	518	0	227	28	1.8	.36	850	7.75
	38E	do	------	26	42	434	94	598	15	556	155	2.3	1.3	1,650	8.3
	40A	Oct. 14, 1959	------	25	80	1,200	230	919	113	1,370	695	2.8	4.2	4,190	8.9
	40B	do	------	11	52	11,700	3,910	2,220	1,020	10,500	10,400	3.6	24	38,700	9.2
	40C	Aug. 13, 1961	------	------	------	19,500	2,430	2,180	2,140	17,300	14,100	29	44	56,800	9.65
	54A	June 2, 1960	------	26	60	800	105	735	0	1,080	390	1.7	2.4	2,940	8.1
	54B	do	------	22	37	1,620	188	1,570	45	1,960	648	2.4	4.6	5,370	8.5
Ponds															
Ephemeral: North end of slump ridges east of lake.	DL31	Oct. 9, 1959	------	51	134	597	84	78	0	1,440	246	------	2.27	2,974	8.1
Marsh fed by outflow from open sag pond.	45	Oct. 20, 1959	------	42	31	250	40	354	8	333	98	1.0	.43	1,010	8.4
Accumulated flash runoff northwest corner of playa.	61	Sept. 1960	------	12	3.4	1,200	552	828	90	638	1,360	4.8	22	4,320	8.4
Accumulated runoff, carbonate flats area. north of playa.	28	June 6, 1960	38	12	4.4	240	92	540	74	59	30	2.7	10	828	9.0
Open sag pond	32	Oct. 9, 1959	36	31	26	190	20	334	0	281	62	.8	.34	812	7.7
	32B	May 22, 1961	32	23	24	209	23	293	0	279	61	.8	.36	824	7.7
	32C	Aug. 15, 1961	34	24	27	186	23	304	0	276	63	.9	.07	845	7.7
Buckhorn pond	37	Oct. 14, 1959	28	46	24	125	25	426	0	199	1.0	.8	.01	692	7.8
	37B	May 22, 1961	18	51	32	118	18	331	0	235	21	.9	.20	674	7.6
	37C	Aug. 15, 1961	37	44	32	101	21	291	0	226	20	.8	.35	654	7.7
Closed sag pond: Low stage	39A	Oct. 14, 1959		------	------	92,200	9,150	46,800	53,600	50,100	49,000	------	303	279,200	9.70
Inflow seepage	39B	Oct. 19, 1959		13	7.7	400	50	896	0	200	50	1.1	.68	1,210	8.1
Low stage	39C	June 6, 1960		18	8.8	48,700	2,710	18,700	23,800	35,700	11,900	41	.00	132,000	9.6
High stage	39D&E	Sept. 9, 1960		------	------	101,000	15,700	12,500	63,300	40,500	58,100	171	532	286,000	10.1
Low stage	39G	May 22, 1961		12	3.0	36,600	2,400	11,000	19,500	27,200	8,460	11	36	97,300	9.7
Deep Springs Lake Playa Area															
Surface waters: Around gas vents, north side.	DL29A	Oct. 8, 1959		------	------	54,000	14,000	11,200	3,630	8,180	79,000	------	94	165,000	8.80
	30A	Oct. 9, 1959		------	------	37,900	10,100	7,000	2,920	11,400	53,800	------	75	120,000	8.85
	30E	May 20, 1961		50	158	3,650	1,510	2,040	0	365	5,930	2.4	3.1	13,000	7.4
Central lake area near staff gage	2B	June 5, 1960		18	15	95,500	3,760	5,330	6,640	152,000	31,700	20	.00	293,000	9.3
	2C	May 5, 1960		13	6.1	73,000	4,150	3,770	5,570	109,000	29,900	22	.00	224,000	9.4
	2D	Jan. 30, 1960		10	23	21,400	3,040	1,680	2,080	18,500	20,300	2.1	.00	66,300	9.6
	2E	Feb. 4, 1960		11	24	26,500	3,670	2,000	2,100	21,100	26,200	2.3	43	80,800	9.5
	2F	Feb. 20, 1960		12	20	29,800	2,790	2,150	1,800	40,500	16,800	1.6	.00	90,500	9.5
	2G	Feb. 29, 1960		13	11	37,900	2,670	3,160	1,540	51,100	16,400	2.0	.00	111,000	9.3
	2H	Mar. 19, 1960		11	52	39,900	3,820	2,270	2,790	51,700	23,200	1.8	22	123,000	9.5
	2I	Mar. 31, 1960		15	46	38,800	3,740	2,390	2,600	50,600	18,500	1.8	33	116,000	9.5
	2K	Jan. 14, 1961		25	2.3	7,370	926	1,400	89	7,640	5,280	1.4	11	21,900	8.6
Intercrustal Brines															
Trough, north side of central lake area.	DL1	Sept. 29, 1959		------	------	99,000	19,000	12,600	20,900	38,800	108,000	------	407	291,000	9.15
	3A	Sept. 30, 1959		------	------	103,000	22,600	15,300	18,400	27,700	127,000	------	364	314,000	9.30
	3B	Oct. 2, 1959		------	------	102,000	19,400	11,100	18,900	37,100	116,000	------	432	306,000	9.35
Trough at staff gage	2	May 31, 1960		6.1	3.6	110,000	24,500	12,400	19,500	57,800	120,000	41	230	340,000	9.4
	2J	Sept. 1960		1.5	3.0	106,000	21,200	1,690	23,400	62,600	113,000	40	335	329,000	9.4
	2L	May 17, 1961		3.1	1.3	116,000	24,800	------	22,100	56,900	118,000	------	290	334,000	9.4
	2M	do		------	------	110,000	18,600	------	24,000	52,200	119,000	------	358	330,000	9.45
	2N	May 20, 1961		1.5	1.1	108,000	24,000	------	22,500	55,200	120,000	------	255	333,000	9.5
	2P	Aug. 11, 1961		------	------	106,000	19,900	1,210	22,700	63,100	98,500	149	305	316,000	9.35
Southeast central lake area.	13	Oct. 3, 1959		------	------	102,000	21,000	8,720	20,500	43,000	111,000	13.0	544	310,000	9.30
Southern thin crust, near Buckhorn inflow.	14	Oct. 5, 1959		------	------	37,000	4,610	371	3,680	37,000	28,400	17.4	104	110,000	9.35
Southwest central area	21	Oct. 6, 1959		------	------	97,700	12,600	5,320	14,700	54,100	101,000	17.6	310	288,000	9.20
Western area, near gas vent.	22	Oct. 7, 1959		------	------	88,800	16,000	16,000	18,200	62,700	75,000	13.0	344	277,000	9.10
Northwest trough	24	do		------	------	102,000	20,400	8,730	17,400	41,000	119,000	12.0	392	311,000	9.15
South central lake area near toe of Buckhorn.	68	Sept. 9, 1960		6.2	3.6	105,000	18,800	12,800	20,400	74,100	94,600	99	935	321,000	9.6
Inflow delta	53	June 1, 1960		12	15	108,000	30,900	12,700	19,200	53,700	120,000	29	53	341,000	9.5
	53B	do		6.3	29	110,000	25,200	0	26,600	54,100	117,000	111	752	336,000	9.3
Interstitial Brines															
In muds	DL29C	Oct. 8, 1959		3.9	1.1	88,300	15,600	8,640	15,900	37,600	103,000	62	166	265,000	9.4
	51	Sept. 12, 1960		3.1	1.2	111,000	19,500	9,360	22,000	57,100	119,000	92	739	335,000	9.7

Along with the analyses, the equivalent percentages of the major cations and anions have been computed, compiled in table 7, and plotted on trilinear diagrams (figs. 19, 20 and 21) such as proposed by Piper (1944). This system brings out the chemical relations among waters more clearly than a simple numerical tabulation. However, these plots are chiefly illustrative, and processes suggested by the relations shown must be verified by further evidence, as pointed out by Hem (1959, p. 183). For example, a change in water composition of Deep Springs waters resulting from the precipitation of calcite cannot be resolved from changes brought about by combination of base exchange and sulfate reduction. In spite of limitations, trilinear plotting is very useful in showing systematic differences among waters, as also indicated by Hem (1959, p. 184). Probably the chief shortcoming of the method is that the plot does not indicate total concentration. In the diagram for Deep Springs Valley waters, total concentration generally increases to the lower right of the diagram; that is, toward the alkali and chloride corners. This is not true of streamflow, however, and this factor renders the Piper diagram less satisfactory for illustrating trends in streamflow units than for illustrating the valley waters as a whole. Also, the criteria for mixtures described by Piper (1944, p. 920) cannot be fully utilized because of solid precipitation and reactions taking place in many waters.

A summary of the compositional trends in Deep

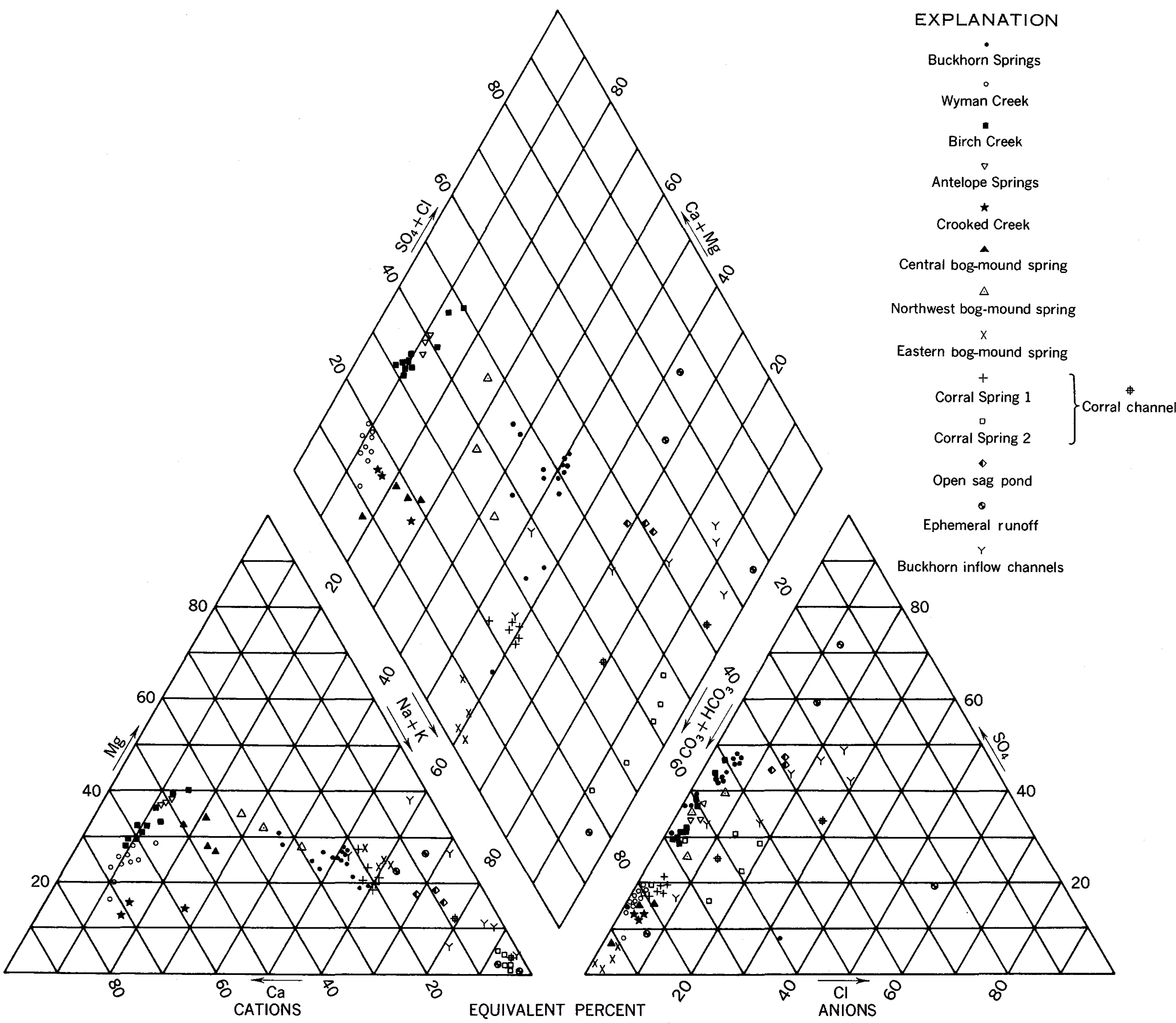


FIGURE 19.—Chemical composition of waters from the Deep Springs Valley which can be considered inflow to Deep Springs Lake.

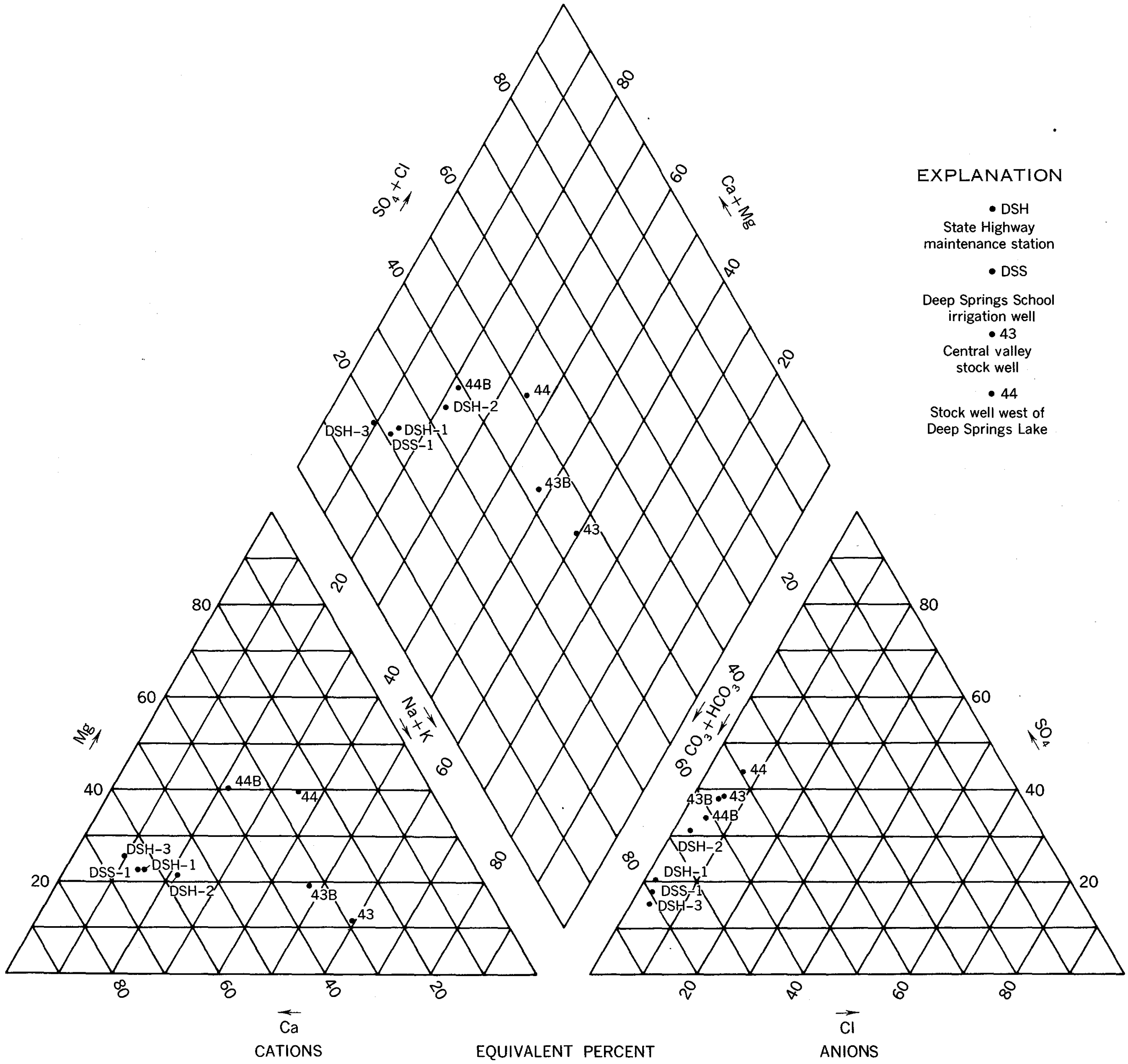


FIGURE 20.—Composition of well waters from the Deep Springs Valley. The numbers refer to localities shown on plate 1 and the corresponding analyses of table 7.

Springs Valley waters suggested by the trilinear plot is given in figure 22.

INFLOW

As illustrated in the central part of the Piper diagram (the diamond field, fig. 19), the chemical compositions of most waters which constitute inflow to Deep Springs Lake fall within distinctive limits; these limits are based on the ratio of alkaline earth to total alkalies and on the ratio of alkalinity to sulfate plus chloride. Towards the lake area, the concentration of alkalies, sulfate, and chloride increases simultaneously with an increase in total dissolved-solid concentration. To a lesser extent these trends may be seen on the individual cation and anion plots as well. The diamond field is the most useful for demonstrating trends in the inflow waters, fig. 19; a separate anion diagram (fig. 21) better illustrates differences in lake and ponded waters. It is interesting to compare the anion diagrams (figs. 19 and 21) with the diagram given by Hutchinson (1957, p. 566) for the general anionic composition of inflow and closed lakes in the western

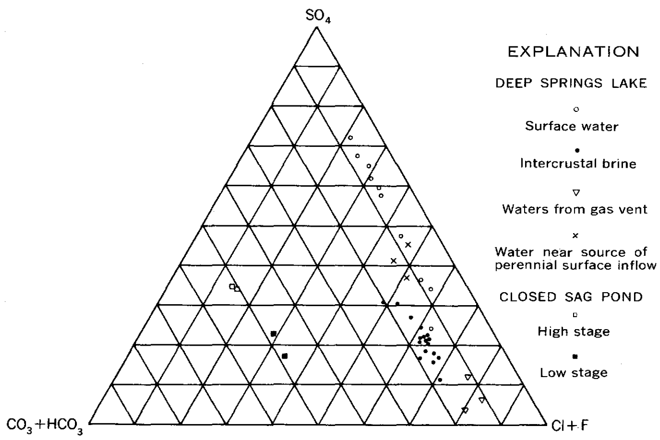


FIGURE 21.—Proportions of major anions, in equivalent percent, in waters from Deep Springs Lake. The points correspond to analyses given in table 7.

Great Basin. The waters of the Deep Springs Valley follow a trend similar to that in Hutchinson's diagram, but in the course of their evolution they show a pronounced offset toward sulfate.

Gross differences in chemical composition of inflow waters are readily apparent in the diagram (fig. 19). The waters of Birch Creek and Antelope Springs are very similar in ionic proportions, and they differ from the waters of the Wyman Creek and Crooked Creek drainages by containing higher concentrations of sulfate and magnesium. In the same way, Buckhorn Springs water can usually be separated from Corral Springs outflow. The easternmost bog-mound springs, on the other hand, are characterized by very low proportions of sulfate. Low concentrations of alkaline earth which result in higher alkali percentages serve to distinguish Crooked Creek from Wyman Creek waters. The open sag pond differs from the Buckhorn Springs by containing higher alkali and chloride percentages.

The major factors influencing gross differences in the chemical composition of inflow units are the original mineral source of the dissolved constituents and processes taking place within each hydrologic unit. Thus, the higher proportions of sulfate in Birch Creek as compared with Wyman Creek reflect the greater number of mineralized zones found within the Birch Creek drainage. Large segments of the channels of Birch Creek and its tributaries parallel contact aureoles formed in the Precambrian and Cambrian rocks by the intrusion of the granitic rock suite. (See pl. 1.) Hydrothermal veins and sulfide mineralization are closely associated with this metamorphism, especially in the drainage of the North Fork Birch Creek. Wyman Creek, on the other hand, crosses this zone at right angles, and a much smaller part of its drainage comes in contact with such rocks. This chemical composition of Antelope Springs may reflect either an extension of mineralization beyond the immediate contact zone or, more likely, a connection with Birch Creek drainage through fractures or joints related to faulting along the west margin of the valley.

There is adequate source for the calcium, magnesium, and carbonate in the calcareous sedimentary and metamorphic rocks in both the Wyman Creek and Birch Creek basins. The somewhat higher calcium content and the still higher magnesium content of Birch

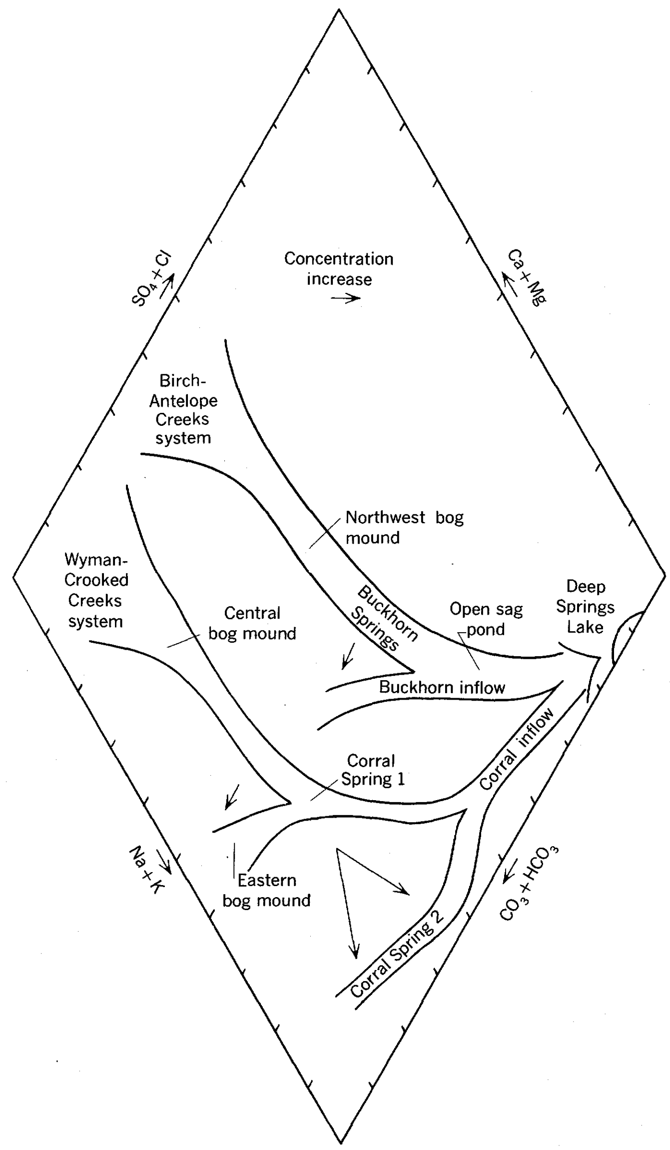


FIGURE 22.—Generalized trends in the composition of dissolved solids in inflow to Deep Springs Lake, based on the plot of water compositions shown in figure 19. Large arrows indicate the two major compositional trends in inflow to Deep Springs Lake. Small arrows indicate the compositional effect of organic respiration, decay, and sulfate reduction on each major trend.

Creek reflect the fact that a larger area is underlain by the Reed dolomite and its metamorphic equivalents in the Birch Creek drainage. The alkalies in both basins are primarily derived through leaching of the feldspathic and micaceous components of granitic rocks. This is shown by the low alkaline earth concentration and resulting high alkali concentration in the waters of Crooked Creek, where the drainage is principally underlain by granitic rocks.

Precipitation of calcite on the streambed of Birch Creek has resulted in a decrease in the concentration of calcium and alkalinity in the waters downstream and a concurrent decrease in total dissolved-solids concentration. Wyman Creek waters decrease slightly downstream in total concentration, but despite some calcite cementation of the bed, there is little change in percentage composition of the streamflow.

The effects of lithologic control and mineralogic source of solutes may also be noted in the compositional variation of the Buckhorn Springs water. From north to south along the spring belt, individual springs receive more drainage from areas underlain by Precambrian rocks which are at increasingly greater distance from the contact with intrusives. (See index map, fig. 1; table 7; and fig. 19; locality numbering from south to north is from DL33 to DL36.) The contact zone, a continuation of that found along Birch Creek, is exposed along the prominent canyon immediately southeast of the lake, and the effects of thermal metamorphism in the sedimentary rocks decrease steadily to the southwest. Correlatively, the proportion of alkalies in the Buckhorn Springs decreases from north to south. Total dissolved-solids concentration and sulfate concentration also decrease in the springs to the south.

The distinctive compositional differences of the bog-mound springs and Corral Springs are enhanced by secondary effects superposed on the lithologic control. The chief processes involved are the precipitation of alkaline earth carbonate, sulfate reduction, and additional solution of material as water moves through the valley deposits.

The patterns of water movement deduced from the hydrography of Deep Springs Valley suggest that waters from the Wyman-Crooked Creeks system move down the hydraulic gradient in the valley alluvium and reappear in the central and eastern bog-mound springs and Corral Springs. (See p. 81 and pl. 1.) No information is available on the flow rates of individual hydrologic units, but comparisons may be made on the assumption that the composition of Wyman Creek waters are not greatly altered between the time they enter the valley ground-water reservoir and the time they issue from the springs. The chief difference in chemical composition between Wyman Creek streamflow and the Corral Springs water is the higher percentage of alkalies and increased total dissolved-solids concentration in the springs; anionic proportions are strikingly similar. (See fig. 19 and table 7.) Alkali concentrations range from 20 to 25 percent of the cations in the central bog-mound spring (loc. DL73), from 55 to 60 percent in the eastern bog-mound group (DL41–42) and Corral Springs 1 (DL48A), and from 90 to 95 percent in Corral Spring 2 (DL48B). Such change in composition would at first appear to result primarily from cation exchange. Montmorillonite clays which might be capable of releasing alkali and fixing calcium have been identified in the surficial valley alluvium but they probably form a minor part of the total valley fill and even in the clay fraction are subordinate to calcium-saturated types. Although ion exchange may be a significant mechanism for alkali enrichment in waters permeating fine-grained lacustrine deposits such as in the bog-mound-spring area, it is more likely that calcite precipitation balanced by additional alkali solution from granitic alluvium accounts for the compositional difference between the streams and spring outlets. Calcite cement has been recognized in several places in the valley alluvium and is most conspicuous in the fanglomerate around the north end of the Corral Springs area. The fact that total dissolved-solids concentration does not decrease in the waters associated with the calcite deposits suggests that precipitation loss is balanced by additional alkali and bicarbonate picked up as the waters percolate through the granitic alluvium and old lacustrine deposits. The obvious and abundant calcite deposits around the orifice of the Corral Springs may be associated with the loss of gaseous carbon dioxide as ground waters come to the surface.

Biotically induced sulfate reduction is another process causing differences in chemical composition between streams and springs. Good evidence of sulfate reduction is the hydrogen sulfide detected in all the bog-mound springs and also at Corral Spring 2. White fungus growth is common too. In the easternmost bog-mound springs 0.1 ppm S^{-2} as hydrogen sulfide was measured, and at Corral Spring 2 up to 11 ppm was determined. The easternmost bog-mound springs contain the lowest sulfate concentrations and, in addition, are notably high in K^{+1}. In fact, were it not for the high potassium content the water of these springs would be cationically intermediate between the large central bog-mound spring and Corral Spring 1. The K^{+1} is probably released by hydrogen-ion reaction with micaceous clays; such a process is conceivable in a strong reducing environment resulting from organic decomposition. The higher dissolved-solids concentration and alkali carbonate enrichment in water of Corral Spring 2 (fig. 2, loc. DL48B) and of the inflow to the closed sag pond (loc. DL39B) as compared with water of Corral Spring 1 (loc. DL48A) are also most likely the result of biotic activity. Where the discharge of springs is slow, marsh vegetation has taken hold and flourished. As a result, evapotranspiration increases total concentration, and respiration and decay keep carbon dioxide concentration high. To a lesser extent, the same processes coupled with evaporation from open water can account for an increased dissovled-solids concentration and alkali proportion in the open sag pond (fig. 2, loc. DL32) and related waters (loc. DL45) as compared with the outflow from the northern Buckhorn Springs (loc. DL36). The effects of biota and evaporative processes on water composition in the immediate spring areas is shown by comparison of samples from the central bog-mound spring taken consecutively at the orifice (loc. DL73C) and at the edge of the surrounding marsh (loc. DL73D). From the spring orifice to the marsh edge total dissolved-solids concentration of the waters nearly doubles as alkalies increase 5 percent and alkalinity increases 10 percent. (See fig. 19 and table 7.)

The lower concentration of alkali in the central bog-mound spring waters as compared with that in the eastern bog-mound group and Corral Springs (fig. 19) may be indicative of variations in the lithologic source of the waters and valley alluvium. The Corral Springs most likely receive all contributions from the small drainages on the east side of the valley which are underlain exclusively by granitic rocks. Furthermore, what little return flow there is from irrigation at Deep Springs School enters old channels on the west side of the valley floor. The central bog-mound spring, on the other hand, is probably fed through alluvial deposits which contain a small amount of material derived from sedimentary and metasedimentary strata.

Waters which by their composition can be traced to Birch Creek-Antelope Springs drainage occur only in the two western bog-mound springs. Even here, some mixture with waters derived from the Wyman-Crooked Creeks system is suggested by the composition of waters from the larger of the two springs (loc. DL78). The chemical composition of its water plots in the diamond field of the Piper diagram (fig. 19) intermediately between the most northwestern bog-mound spring (loc. DL27) and Corral Spring 1 (loc. DL48A). The difference in samples from the most northwestern bog-mound spring (table 7) to some extent reflects different points of collection; sample DL27C was obtained closer to the spring orifice. The trend from the Birch-Antelope Creeks drainage system to the western bog-mound springs is similar to the trend from Wyman Creek to the Corral Springs (fig. 22); no doubt these similar trends from similar processes.

The single sample (AS79) obtained from surface waters in Antelope Canyon shows significant cationic differences from the waters of Antelope Springs; magnesium and sodium are both about 5 percent higher (table 7). This sample most closely resembles samples from the well due west of the lake, and is probably representative of waters in contact with relatively unmetamorphosed Cambrian strata or alluvium derived therefrom. Relatively high Mg^{+2} content might be expected to occur in waters in contact with dolomitic argillites rather than waters in contact with the metamorphic equivalents, in which Mg^{+2} is locked in insoluble silicates.

The composition of most well waters varies from place to place in accordance with the flow patterns suggested for the valley (pl. 1). In the northern part of the valley, well samples (fig. 23) either reflect the composition of the Wyman-Crooked Creeks system very closely (table 7, samples DSH-1 and DSS-1) or show the effect of calcite precipitation plus alkali and sulfate enrichment (sample DSH-2). Well-water samples from the central valley (samples DL43A and DL43B) unfortunately could only be obtained from the stock tank. Thus, the percentage composition reflects a rise in temperature, evaporation, loss of carbon dioxide, and the precipitation of alkaline earth carbonate as scum. The effect of these factors is also shown in a comparison of two samples from the well west of the lake area (table 7, samples DL44 and DL44B); the sample taken directly from the outflow pipe had higher calcium and bicarbonate concentrations than the sample collected from the stock tank when the well was not operating.

Variations in the chemical composition of inflow are also caused by seasonal fluctuations in discharge, source areas, temperature, and biologic activity.

Figure 23 shows the relations of constituent concentration and percentage composition to discharge for Wyman Creek. Except where there is some contribution from overland flow, total dissolved-solids concentration decreases with an increase in discharge which decreases the concentration of alkaline earths and bicarbonate; however, percentage composition remains relatively uniform. The addition of overland flow raises the amounts of calcium and bicarbonate significantly because calcareous rocks are plentiful in the upper

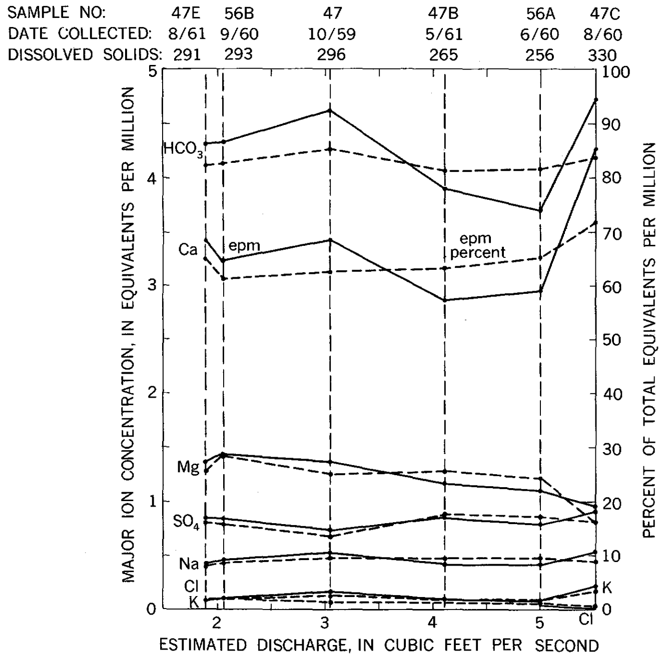


Figure 23.—Relation of major ion concentration and percentage composition to estimated discharge for Wyman Creek. Dissolved solids are given in parts per million.

parts of the drainage area where most such flow originates. Magnesium is the only constituent which shows a consistent trend; its concentration decreases as discharge increases.

No clear-cut relation of either constituent concentration or percentage composition to total dissolved solids concentration or discharge is apparent from the scanty data available for the Corral and Buckhorn Springs (figs. 24, 25 and 26), though the composition of the Buckhorn Springs water faintly suggests that calcium increases as the discharge increases. Whatever relation might exist is apparently masked by prior seasonal fluctuations, particularly in the waters feeding the springs. This masking is especially evident in the northern Buckhorn Springs. Samples collected after an early winter-type northwest storm in October 1959, in which most of the precipitation fell in the north half of the valley, showed the composition of the northern Buckhorn Springs water to be very close to that of the Corral Springs—higher bicarbonate, lower sulfate, and slightly higher alkali than usual. (See fig. 19 and table 7, sample DL36.) This effect was not as pronounced in the Buckhorn springs to the south. These conditions were probably the result of increased flow southward along the major fault zone. The southern Buckhorn Springs were not as strongly affected because of distance and a reversed hydraulic gradient to the south (the Buckhorn Springs farthest north, loc. DL36, are the lowest points in the system).

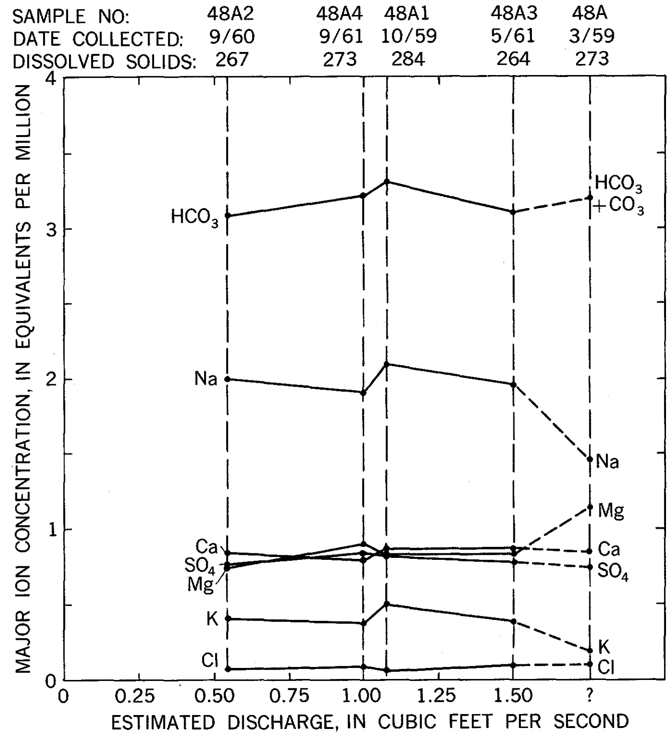


Figure 24.—Relation of major ion concentration to estimated discharge for Corral Spring 1. Dissolved solids are in parts per million. Dashed lines suggest trend associated with higher, but unknown, discharges, such as that in March 1959.

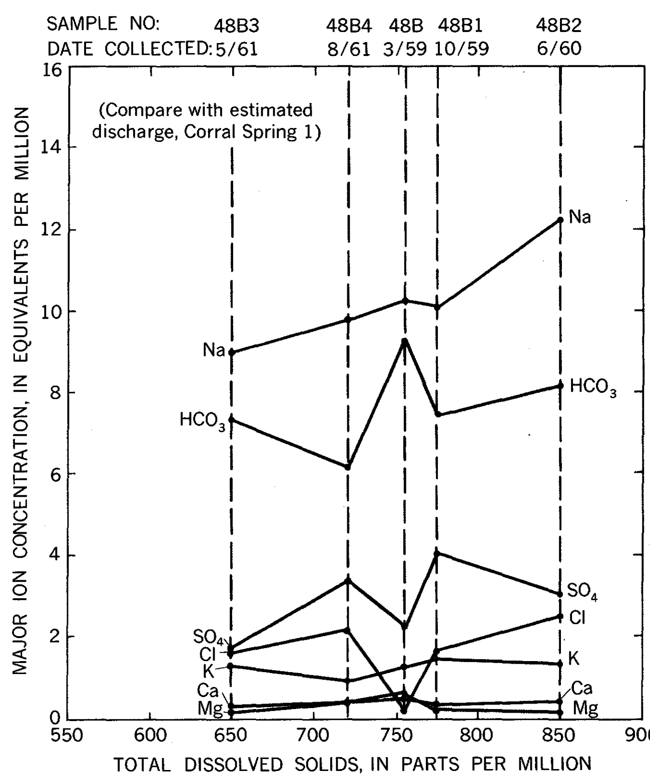


Figure 25.—Relation of major ion concentration to total dissolved solids for Corral Spring 2.

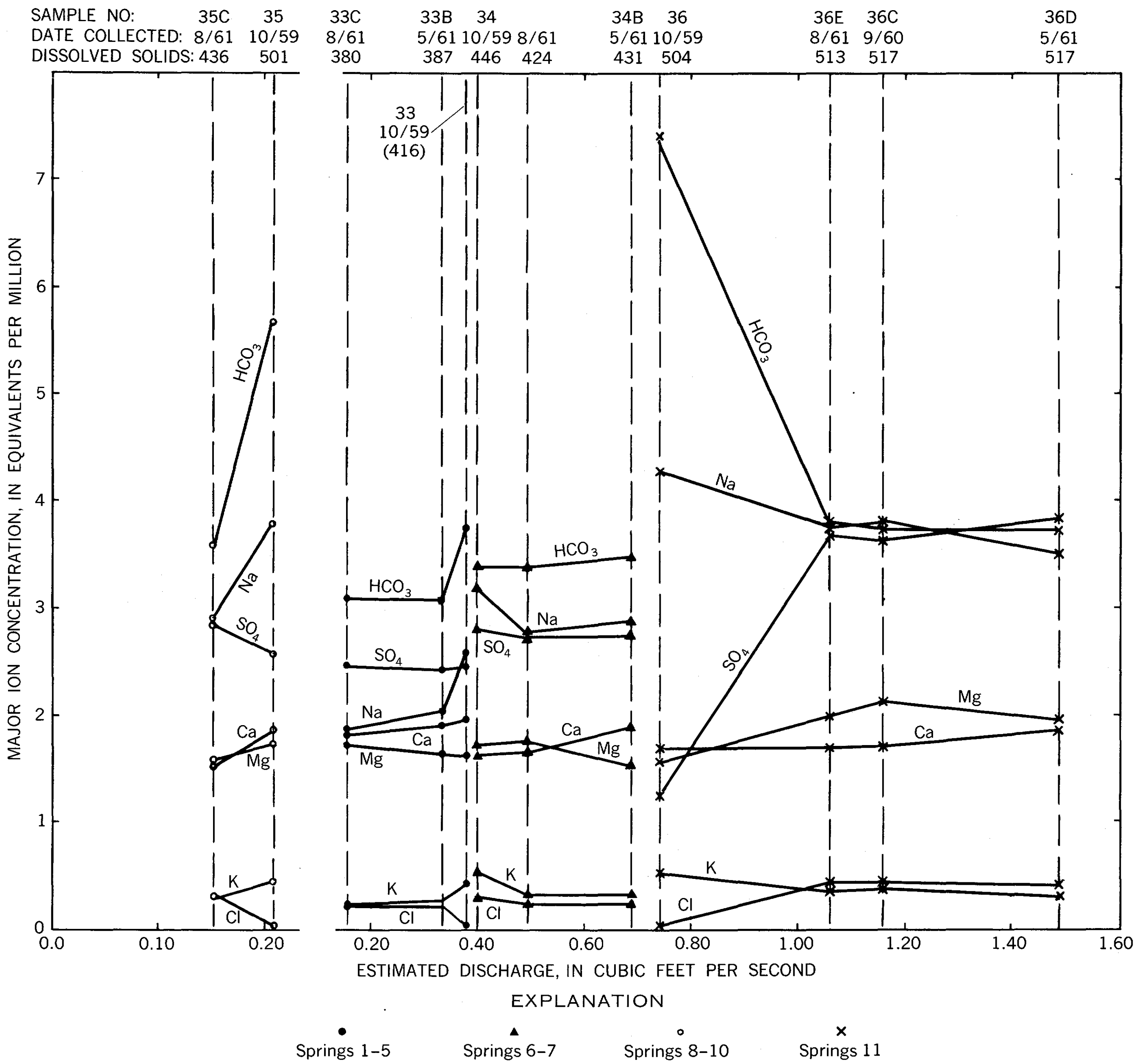


FIGURE 26.—Relation of major ion concentration to estimated discharged for the Buckhorn Springs. Dissolved solids are in parts per million.

Seasonal variation in the composition of water may also be noted in the Corral Springs (figs. 24 and 25), where the composition of water affected by winter storms bears little resemblance to compositional trends of waters during the summer. Such variation is also related to seasonal differences in temperature and biotic activity.

Substantial changes in total concentration and percentage composition of waters contributed to Deep Springs Lake take place between the orifices of inflow springs and the central lake area. The course of these changes can be traced definitely only when the waters which reach the lake remain on the land surface, as in the main inflow channels of the Corral and Buckhorn Springs. Lack of prior knowledge prevented the selection of the best sites to represent these trends in adequate detail, but general relations are given in figures 27, 28 and 29, in which the change of percentage composition with increasing total concentration is shown. In these plots, lines are drawn in a down-channel sequence and connect values for each major constituent at each site, but because of the small number of sites not much significance can be attached to point-to-point variation. Futrhermore, the sampling at the sites was not time sequential.

The major cause of the compositional changes downchannel is evaporation coupled with alkaline earth carbonate precipitation and re-solution of capillary salts. Biologic, particularly floral, activity probably plays a significant catalytic role in these processes which are all accelerated as the inflow channels become less distinct and, near the lake, open onto marshy flats having very low gradients.

Compositional change in the Corral Springs inflow (fig. 27) appears to follow a regular pattern, if one assumes that the small seeps intermediate in pattern between springs 1 and 2 and Corral Spring 2 itself are in downgradient sequence as the result of a progressive increase in evaporation at each site. Precipitation of calcite, probably impure, is indicated by near

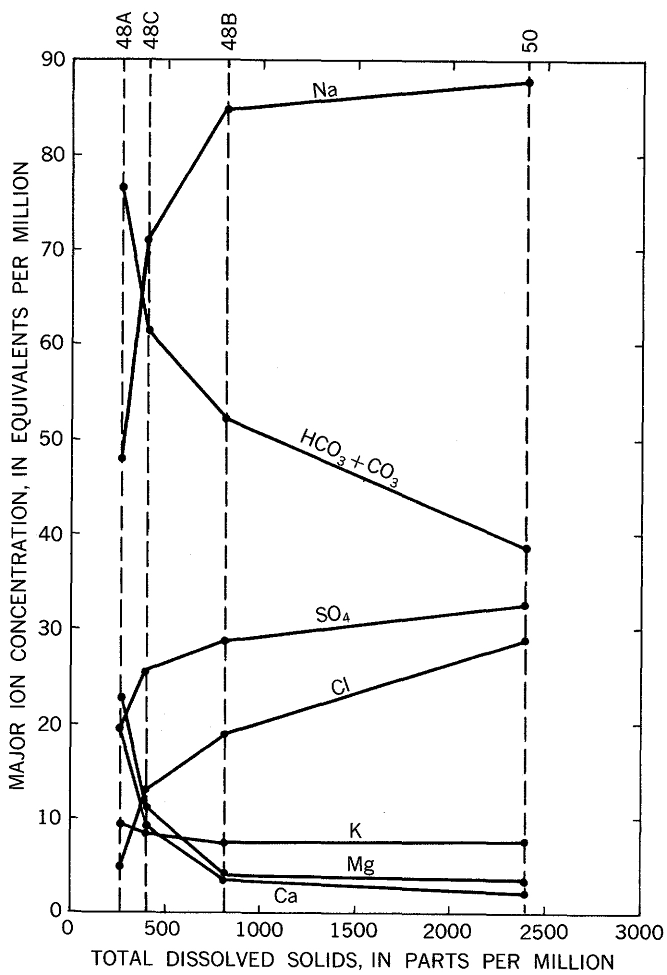


Figure 27.—Relation of major ion concentration to total dissolved solids for Corral Springs waters along the course of inflow to Deep Springs Lake. Sampling locations at top of graph: 48A, Corral Spring 1; 48C, Corral Spring Canal; 48B, Corral Spring 2; and 50, lower channel.

constancy of calcium and to a lesser extent magnesium. An increase of alkalinity due to evaporation is offset by carbonate precipitation and loss of carbon dioxide to the atmosphere; so, alkalinity declines as sulfate and chloride contents increase.

The fact that compositional trends shown by the main Buckhorn Springs inflow are generally the same as those in the Corral Springs sequence (compare fig. 27 with 28 and 29) suggests that the same processes are operative. In detail, however, there are much greater fluctuations, particularly in the anionic composition. The trend of each constituent during different time and seasonal intervals is roughly parallel but displaced relative to total dissolved-solids concentration. The same is true of the downchannel rise in pH (fig. 30). This displacement occurs because the higher temperatures and increased floral activity in summer bring about a loss of carbon dioxide, rise in pH, and precipitation of alkaline earth carbonate at lower levels of total dissolved-solids concentration. The cationic trends suggest that seasonal factors are especially important in the precipitation of a magnesium-bearing carbonate phase (dolomite). Such seasonal effects are somewhat similar to those described by Alderman and Skinner (1957) in their explanation of dolomite precipitation in the lagoons of southeastern Australia.

In summary, the chemical composition of most waters from the Deep Springs Valley falls into distinct units according to the ratios of alkaline earths to alkalies and of alkalinity to sulfate plus chloride—the basis of the Piper diagram (fig. 19). These units include the Wyman-Crooked Creeks system, Birch Creek plus Antelope Springs, the eastern, central, and western bog-mound springs, the Corral Springs 1 and 2, the Buckhorn Springs, and the open sag pond. In the course of movement from the mountains to the lake, the water composition is initially controlled by bedrock lithology and is subsequently influenced by other factors. If cations or anions are considered individually, compositional change follows a single trend from the drainage divides to the lake area; the percentages of alkalies and sulfate plus chloride generally increase as the total dissolved solids concentration increases. However, the combination of cations and anions suggests two major compositional trends in the inflow to Deep Springs Lake (fig. 22). One group of waters is predominantly associated with sedimentary lithologies and is characterized by higher concentrations of sulfate. This group includes Birch Creek, Antelope Springs, the northwest bog-mound springs, and Buckhorn Springs, and the open sag pond. The other group is predominantly associated with igneous rocks and includes the Wyman-Crooked Creeks system, the east

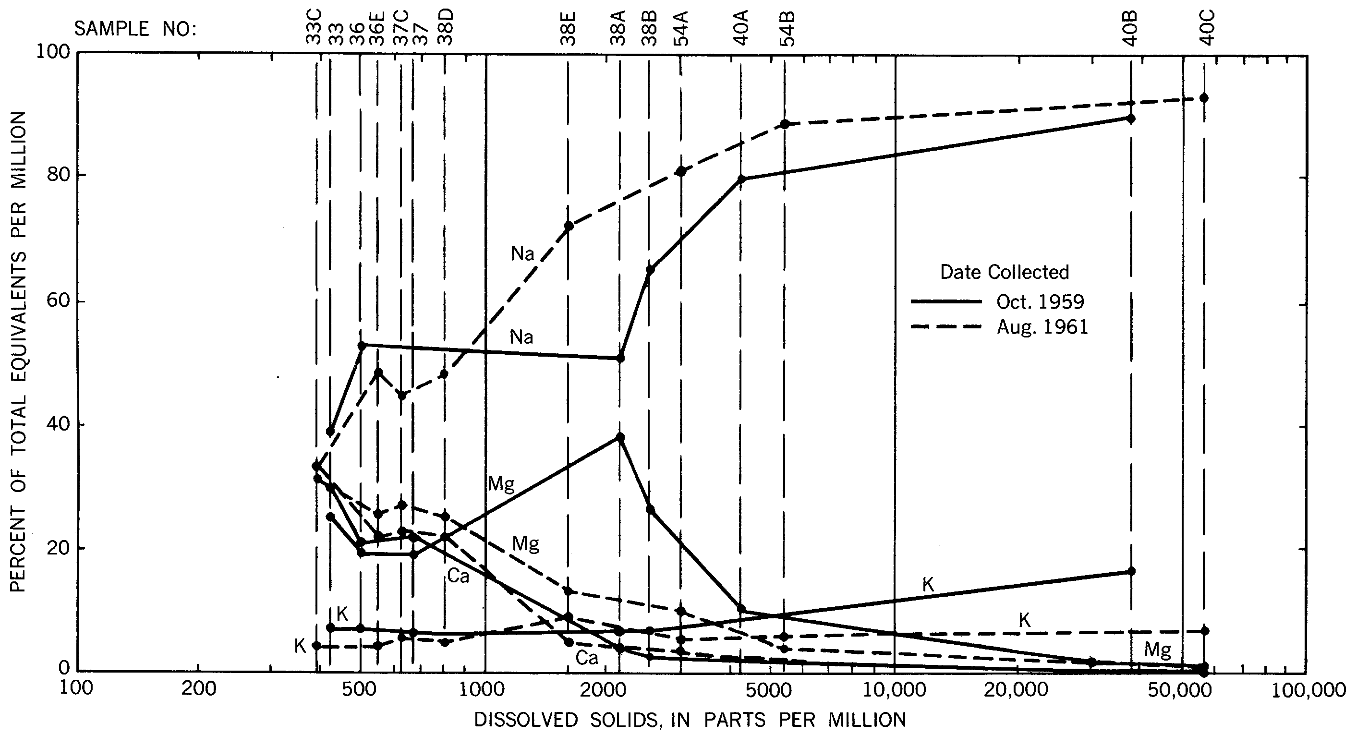


Figure 28.—Relation of major cation concentration to total dissolved solids for Buckhorn Springs waters along the course of inflow to Deep Springs Lake. Though collected in June 1960, samples DL54A and DL54B are included with samples collected in August 1961 because hydrologic conditions were very similar.

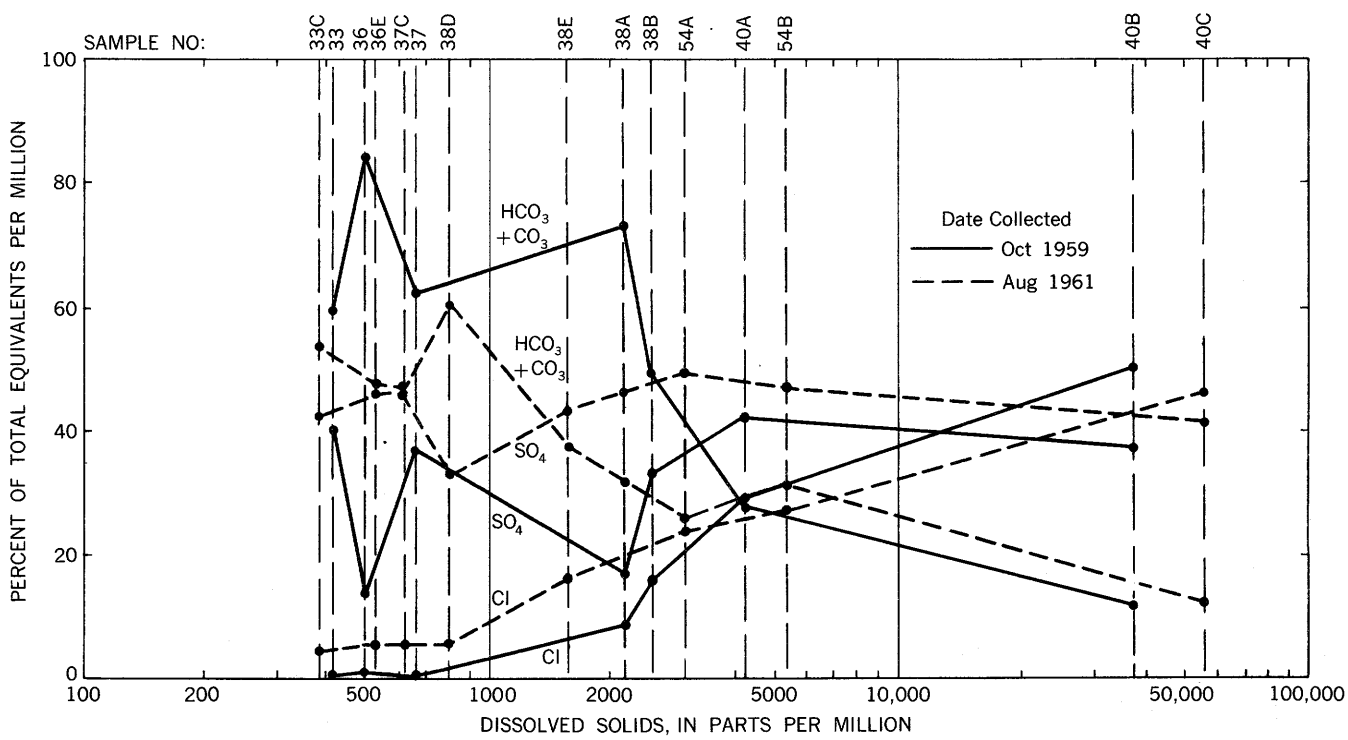


Figure 29.—Relation of major anion concentration to total dissolved solids for Buckhorn Springs waters along the course of inflow to Deep Springs Lake. Though collected in June 1960, samples DL54A and DL54B are included with samples collected in August 1961 because hydrologic conditions were very similar.

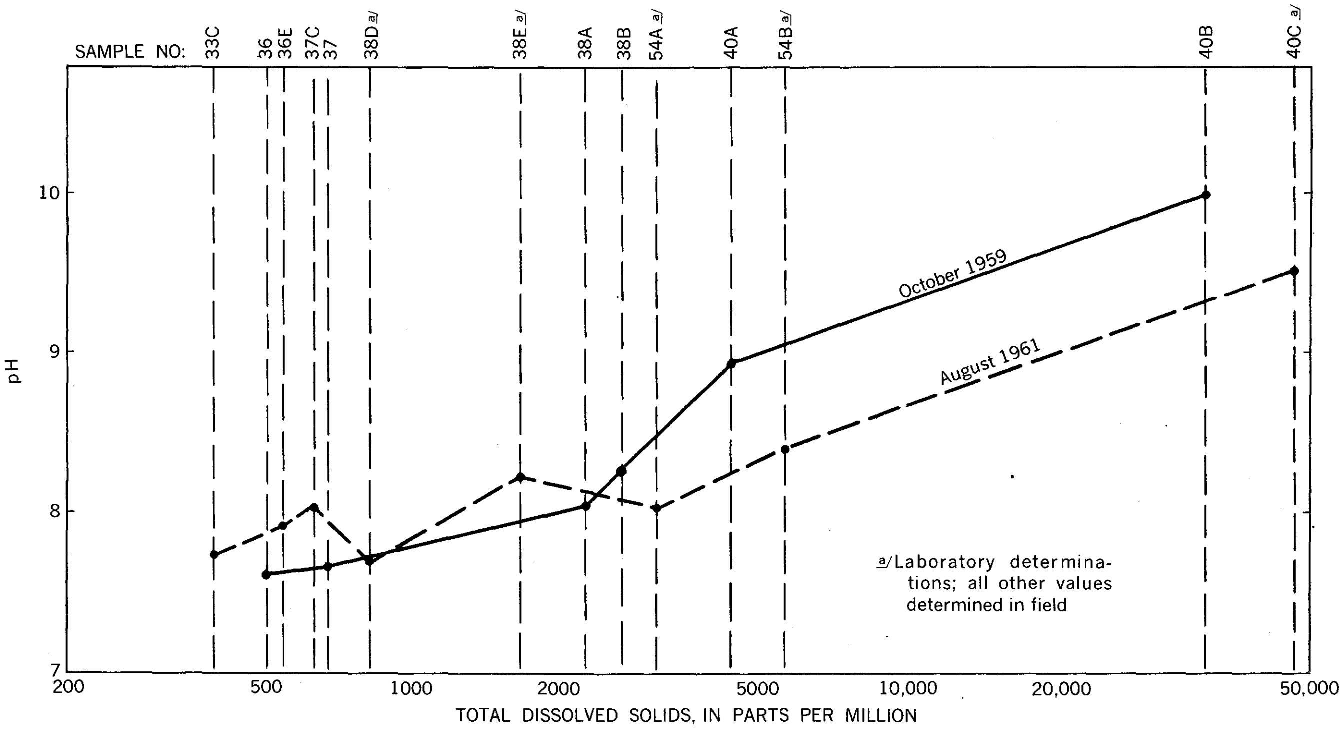


FIGURE 30.—Relation of pH to total dissolved solids in Buckhorn Springs waters along the course of inflow to Deep Springs Lake. Though collected in June 1960, samples DL54A and DL54B are included with the August 1961 sequence because hydrologic conditions were nearly identical.

and central bog-mound springs, and Corral Springs. The two groups have compositional trends which are distinct but roughly parallel up to the area adjacent to the playa, where they begin to converge (fig. 22). The compositional trends further corroborate the flow patterns worked out for the valley (p. A20). Compositional change in inflow units is brought about by precipitation of alkaline earth carbonate, sulfate reduction, additional alkali solution, and evaporation.

Some variation in stream composition is associated with changes in the amount of discharge, but spring composition is more affected by prior seasonal variation in waters feeding the springs. Compositional changes in all waters may be correlated with seasonal changes in temperature and biologic activity.

LAKE AREA

The chemical composition of waters from Deep Springs Lake varies through wide extremes dependent on stage and specific influences at the point of collection. Lake waters become more uniform in composition as the stage rises but at low stages the waters become physically separated, as previously described (p. A23), and their composition is more dependent on their immediate environment. Surface waters entering on the playa, especially at low stage, often become stagnant before mixing and retain chemical characteristics related to a particular inflow source. Thus, stagnant waters around the immediate inflow area of the Corral Springs may be distinguished from ponded waters near the mouth of the Buckhorn channels by higher ratios of alkalinity to sulfate. (See fig. 19; compare samples DL50 and DL54, table 7.) The most extreme examples of such differences are in puddles and ponds resulting from heavy local precipitation and runoff. (See fig. 31.) These waters reflect the most abundant and soluble surficial materials in the immediate area. Runoff from the carbonate-rich interdunal flats on the north side of the lake area was found to carry more than 80 equivalent percent bicarbonate (fig. 19; table 7, DL28), whereas ponded water from the northwest corner of the playa where there is much halite in the efflorescent crusts contained nearly 70 equivalent percent chloride (table 7, sample DL61). Interstitial solutions and surface pools derived from ground-water inflow are also closely related to their source (table 7, sample DL3A). Brines around the north margin of the central lake area are influenced by the low sulfate content of the eastern bog-mound springs.

The most distinctive waters associated with specific playa features are found around gas vents (fig. 32). These vents are roughly conical in shape and extend into the underlying mud. They are most common in the northern part of the central lake area and are associated with small pools even at lowest lake stage. Gaseous activity is seldom apparent in most of the vents, though one or two vents near the north limits of the leveed area may show vigorous bubbling. The gases are mostly air and carbon dioxide, but hydrogen sulfide and methane were also detected. The waters associated with these vents (fig. 21; table 7, samples DL29A, 30A, and 30E) are characteristically high in bicarbonate. They may also be high in potassium and chloride, though concentrations are ⅓ to ⅒ that of the perennial brine. These waters were probably once similar to those of the eastern bog-mound springs, but they have since picked up chloride by seepage in the northern lake area, which is characteristically high in chloride and in which are located the halite "beaches" described on p. A46. Samples of intercrustal brine from the northern part of the trough are also high in Cl^{-1} (table 7, sample DL3B).

Despite considerable variation in absolute quantities, equivalent percentages of the major anions in waters from the central lake area fall within narrow ranges

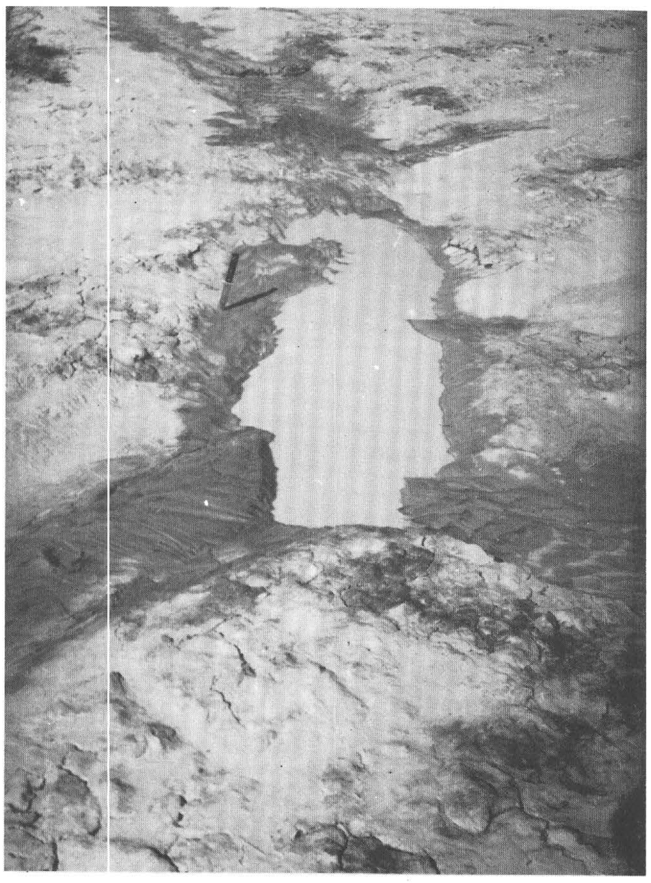


FIGURE 31.—Puddle of ephemeral runoff left behind on interdunal flats just north of Deep Springs Lake. Suspended material is almost entirely alkaline earth carbonate. Knife gives scale.

FIGURE 32.—Typical small gas vent on the north side of Deep Springs Lake saltpan. Gas bubbles were not present at the time the photograph was taken but appeared later. The surrounding crust was less than 2 inches thick. Hunting knife gives scale.

(fig. 21). All samples of surface water taken from the main lake area, regardless of sampling point, contained less than 14 equivalent percent carbonate species (less than 10 equivalent percent if associated with a lake stage in excess of 0.1 foot). In contrast, all samples of intercrustal brine had higher bicarbonate contents and ranged from 14 to 20 equivalent percent in total alkalinity (such values would be changed only 2 or 3 equivalent percent with errors up to 20 percent in the alkalinity determination). Ratios of SO_4:Cl generally are slightly higher in brines from the southern part of the lake area, but variations are less than 25 equivalent percent of either constituent in intercrustal brines regardless of the sampling point. Two samples of interstitial brines from shallow muds immediately outside the leveed area (table 7, samples DL29C, and DL51) did not differ significantly from the central area brines.

In the absence of almost any detectable alkaline earth, cation variation of the brines can be expressed primarily on the basis of the relative amounts of sodium and potassium. There appears to be little areal variation of Na:K ratios in the intercrustal brines of the central lake area, though the ratios (3.5–7.7) are consistently very low for natural solutions (even compared to such a brine as that from Owens Lake, 13+). When considered in the simplified reciprocal system Na-K-SO_4-Cl (fig. 33), the intercrustal brines show three times as much variation in the SO_4:Cl as in the Na:K ratios. This indicates that potassium does not precipitate solely as the chloride but also as a sulfate

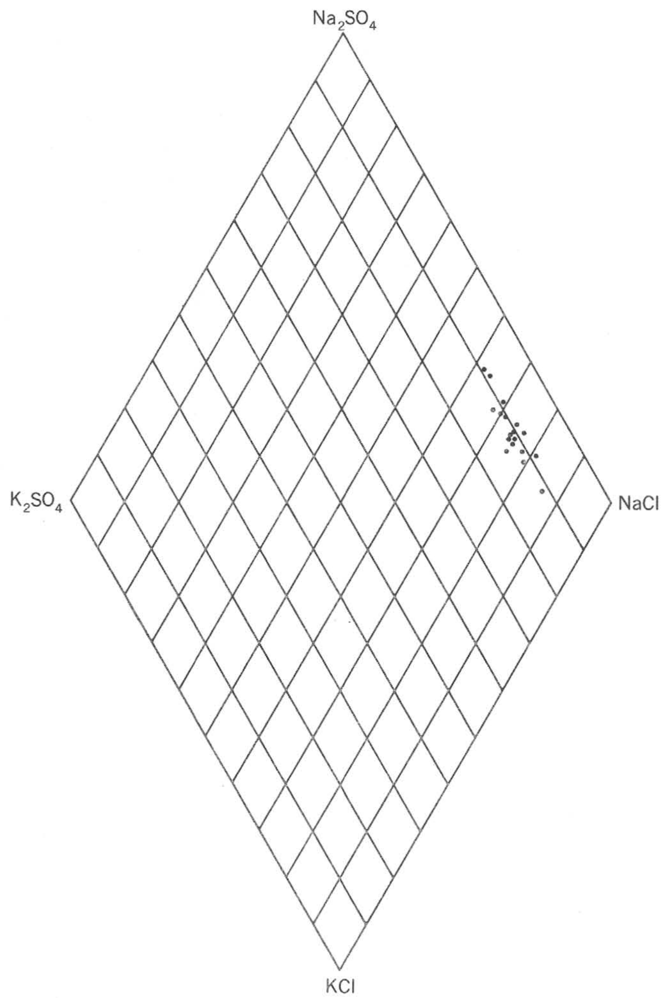


Figure 33.—Composition of intercrustal brines from Deep Springs Lake in terms of the reciprocal system Na_2SO_4–K_2SO_4–NaCl–KCl. Points represent analyses given in table 7.

salt (identified as aphthitalite, $K_3Na(SO_4)_2$). Slightly higher K^{+1} contents may be associated with brines from the trough, especially on the north side.

Significant variations in composition of the surface waters at Deep Springs Lake are associated with changes in stage. (See table 7, and fig. 34.) Rapid rises in stage resulting from rainfall directly on the lake area are characterized by SO_4:Cl ratios in the waters which are nearly 10 times as great as those in intercrustal brines. This change in the ratios reflects the quantitative dominance of sulfate over chloride in the saline crusts. The dominance of sulfate is also shown in changes in the SO_4:Cl ratios during recession of seasonal high stages. After a rise in stage, the SO_4:Cl ratios are initially somewhat similar to intercrustal brine, but continued re-solution of sulfate after most of the chloride has been redissolved brings about an increase in the SO_4:Cl ratios until saturation with sodium sulfate is reached. Precipitation of the sulfate temporarily reduces the ratio again, but continued desiccation eventually leads to even higher values (fig. 34). The relatively low SO_4:Cl ratios of intercrustal brine reflect higher chloride solubility, the formation of additional sulfate salts prior to chloride precipitation, and the bacterial reduction of sulfate in the brines. Substantial sulfide has been detected in all intercrustal brines analyzed.

In summary, waters from the Deep Springs Lake playa itself may be categorized as marginal inflow (ephemeral or perennial), surface waters of the main lake, or intercrustal brines. Marginal inflow is diverse in composition, depending on its source, but surface waters and brines show consistent compositional traits; intercrustal brines are higher in alkalinity and have a lower SO_4:Cl ratio than any surface waters of measurable stage.

CLOSED SAG POND

Variation in the percentage composition of waters (table 7, samples DL39A and DL39C-G) from the closed sag pond is limited almost exclusively to the SO_4:Cl ratio. Alkalinity varies less than 3 equivalent percent in all samples analyzed and is about 30 percent higher than the waters of the main lake. Available data indicate that the SO_4:Cl ratio remains at the level found in the inflow seepage (table 7, sample DL39B) until saturation with sodium sulfate is reached, and then the ratios shifts to lower values.

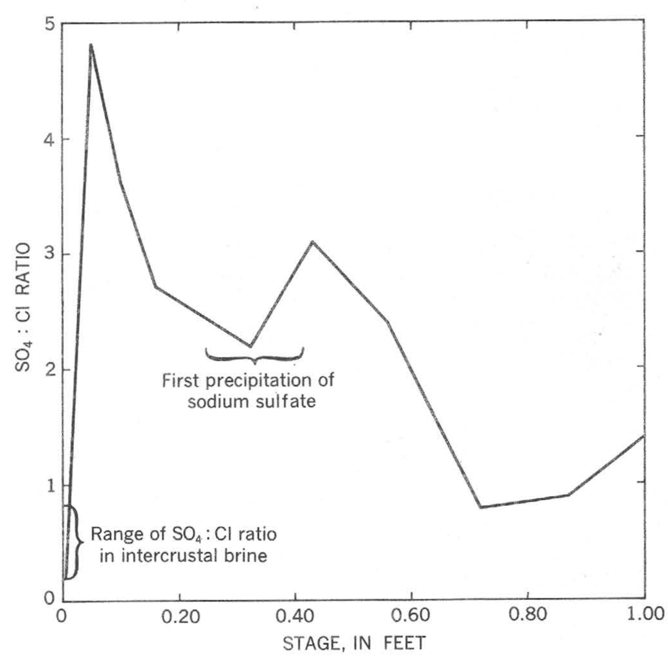


Figure 34.—Relation of lake stage to SO_4: Cl ratio for surface waters at Deep Springs Lake.

CALCULATION OF SALINITY FROM HYDROGRAPHIC DATA

From data compiled from 24 closed lakes in all parts of the world, Langbein (1961, p. 13—14) has derived an empirical equation for salinity based solely on hydrographic factors. Computations of solution concentrations for various stages of Deep Springs Lake utilizing this relation departed from actual measurements irregularly and by large amounts. However, only two or three of Langbein's examples could be considered playa lakes having response time of about 1 year or less. The rapid fluctuations, shallow depths, and high concentrations of water bodies such as Deep Springs Lake increase the importance of nonhydrographic factors, such as solution kinetics, relative salt solubilities, thickness of exposed salt crust, and residual brines. At Deep Springs Lake the thickness of salt crust exceeds the mean depth, and the major constituent of that crust—thenardite (sodium sulfate)—may not dissolve in appreciable quantities until the stage has exceeded a quarter of the highest annual level rise. The initial surge of fresh inflow is expended in the dilution of residual brine and the solution of surficial salts. A minimum estimate of the total quantity of salts available for solution can be derived by considering the salt crust as a spherical segment having a maximum thickness of 1 foot and a minimum diameter of 1 mile. If a minimum specific gravity of 2.0 is assumed, the crustal salts at Deep Springs Lake would exceed 10^6 tons. This quantity is more than could have been dissolved at the highest lake stage observed during the entire course of this study.

MINERALOGY

The deposits of Deep Springs Lake contain a rather wide variety of mineral constituents. Inasmuch as most of the material is very fine grained, it was seldom possible to precisely identify even the major mineral species in the field. Thus, samples were collected for detailed laboratory examination from more than 100 points in the playa area at or close to the localities shown in figure 2. Sample points were generally selected with reference to the major surficial features of the playa; location was in large measure controlled by obvious changes in texture or color. One great limitation on sampling was the inaccessibility of substantial parts of the playa which, especially within the leveed area, could not be reached with safety. At most points, particular attention was given to the deposits on or near the surface which might be related to present hydrologic conditions. Several replicate samples were collected in areas subject to seasonal hydrologic fluctuation.

Surficial materials were collected by cutting out small sections with a hunting knife or trowel. Where direct observation of material at depth was desirable and possible, pits were manually dug. Otherwise, samples from depths as much as 5 feet were obtained by a plunger-type hand auger having a 1-inch-diameter core barrel. Some samples were also examined from the 28-foot auger hole drilled through the levee on the north side of the lake near the level staff. Samples collected in the field were packed in two to three layers of heavy polyethylene to prevent drying and then placed in 1-pint cartons.

CARBONATE AND SALINE MINERALOGY

The saline minerals in the samples from Deep Springs Lake were identified principally by X-ray powder diffraction, utilizing a Norelco recording geiger-counter diffractometer. The instrumental techniques employed were virtually those of Zen (1957, p. 890). Small portions of all samples were first ground fine and analyzed in bulk. Both aluminum holders and glass slides were used for mounting. When possible, a part of the material was X-rayed in its original wet state and compared with another portion of the same sample ground in acetone. Where distinction was doubtful between reflections belonging to precipitate or detrital minerals, the samples were treated with distilled water and (or) dilute acid and reexamined.

Several attempts were made to separate the individual saline mineral species occurring in any single sample. Spot tests using several compounds that characteristically react with various salts were unsuccessful. Centrifuging in liquids of known specific gravity was also largely unsuccessful. Some alkali carbonate, however, was separated from saline crusts by centrifuging at high speed. In many samples, specific layers, spots, or individual mineral grains were picked out under the binocular microscope and X-rayed separately. In addition, most samples were examined under the petrographic microscope to provide textural information and some check on the X-ray results; however, examination was severely limited by the extremely fine-grained nature of most of the material. In the coarsely crystalline samples, some minerals were separated by disaggregation and sieving into selected size fractions. A few thin sections cut from material impregnated with wax by the method of Tourtelot (1961) were examined.

The carbonate and saline minerals and the general type of deposit in which they occur are given in approximate order of abundance in table 8. Identification is based on the comparison of bulk-sample X-ray patterns with standards and on an optical check of most of the samples.

The detailed mineralogy of the Deep Springs Lake deposits is strongly dependent on the hydrography of the area. The distribution and character of the carbonate and saline minerals are most readily discussed in terms of three distinct patterns found in the saline deposits: (1) relatively regular areal zoning of precipitate minerals in the surficial (top 2+ ft) lacustrine deposits of the entire playa, (2) layer variation in mineralogy of lake-deposited saline crusts, and (3) highly irregular and local mineralogical variation in capillary efflorescences.

TABLE 8.—*Precipitate mineral species at Deep Springs Lake*

[Minerals are listed in approximate order of abundance]

Mineral	*Occurrence*
Dolomite, $CaMg(CO_2)_2$	Mud.
Calcite, aragonite, $CaCO_3$	Do.
Thenardite, Na_2SO_4	Mud, saline crust, and efflorescence.
Halite, NaCl	Do.
Gaylussite, $Na_2Ca(CO_3)_2 \cdot 5H_2O$	Mud.
Burkeite, $Na_6(SO_4)_2CO_3$	Mud, saline crust, and efflorescence.
Trona, $Na_3H(CO_3)_2 \cdot 2H_2O$	Saline crust and efflorescence.
Aphthitalite, $K_3Na(SO_4)_2$	Do.
Pirssonite, $Na_2Ca(CO_3)_2 \cdot 2H_2O$	Efflorescence.
Nahcolite, $NaHCO_3$	Mud.
Thermonatrite, $Na_2CO_3 \cdot H_2O$	Efflorescence.
Glauberite, $Na_2Ca(SO_4)_2$	Do.
Sylvite, KCl	Saline crust and efflorescence.
(?)Analcite, $NaAlSi_2O_6 \cdot H_2O$	Saline crust.
(?)Bloedite, $Na_2Mg(SO_4)_2$	Do.

AREAL DISTRIBUTION IN THE LACUSTRINE DEPOSITS

Hunt (1960) has delineated mappable zones in the saltpan at Death Valley based on dominant anion composition and related to the sequence of precipitates formed on evaporation of an average brine. Similarly, the zoning of precipitate minerals in the deposits at Deep Springs Lake can be utilized in areal mapping of the playa (fig. 2).

Except for calcite and dolomite, zonal boundaries are drawn along the outer limit of key precipitate minerals in the surficial playa deposits. Calcite-aragonite and dolomite zones are based on the relative amounts of these minerals. The sequence of key minerals from playa margin to center is calcite and (or) aragonite, dolomite, gaylussite, thenardite, and burkeite. Except for aragonite and gaylussite, all the key minerals persist from the outer limits of their occurrence to the center of the saltpan. No aragonite could positively be identified in deposits containing thenardite. Gaylussite crystals persist only into the outer reaches of the thenardite zone.

No zone based on the relative amount of halite can be established, inasmuch as halite crystallizes from evaporating waters over a wide area under a variety of conditions. Other saline minerals occur irregularly or in response to conditions other than fluctuating lake levels.

Calcite and (or) aragonite strongly dominate over dolomite only on the playa's far west side. Typically, the surficial sediments of these areas contain a high percentage of detrital material. Most of the carbonate is present as an extremely fine-grained aggregate. A few relatively large anhedral grains suggest that at least some of the calcite is detrital in origin, but such occurrences are confined to the marginal channel-scarred parts of the playa which most frequently receive surface runoff from the surrounding area. In the borderlands immediately outside the playa, the interdunal flats are characteristically covered by white to gray sediment which has a shiny smooth surface. This surficial sediment remains in milky suspension for a considerable time in the standing puddles which accumulate. This sediment is composed chiefly of calcite but includes some aragonite, dolomite, and minor amounts of silicate clay. Such material has a physical appearance very similar to the sediments, described by Alderman and Skinner (1957), from the shallow saline interdunal basins of southeast Australia (fig. 34).

Dolomite is the dominant precipitate mineral constituent in the gray-green and black muds of the lake area. Though dolomite may make up more than 70 percent of some near-surface muds, the amount of dolomite relative to calcite and clay generally decreased with depth. In most samples, the dolomite is so fine grained that individual grains can not be resolved with very high magnification, but a few tiny euhedral rhombs were observed. The dolomite-clay aggregate is commonly coated with hydrous iron oxide.

It is difficult to establish a precise boundary that separates areas containing mostly dolomite from those containing mostly calcium carbonate. This boundary is closely dependent on the microrelief of the playa surface and apparently lies near the highest shoreline reached by the ephemeral lake during an annual climatic cycle. (See highest contour in fig. 12.) Precipitation of dolomite directly from solution is suggested by the occurrence of euhedral grains and the distribution of the dolomite; muds within areas of relatively frequent flooding by lake waters contain the highest percentage of dolomite. Bulk-sample X-ray data on the dolomite muds were obscured by the presence of substantial clay impurity, but variation in spacings, broad lines, and weak order reflections in many samples suggested the presence of primary calcic protodolomite (Goldsmith and Graf, 1958), though detrital dolomite was identified in sediment carried from Antelope Canyon to the lake by flash flood. More detailed X-ray and isotopic studies by Peterson, Bien, and Berner (1963) have since shown that much of the dolomite is definitely of recent primary origin.

Near the west-central part of the playa in the area of broad mud flats, gaylussite occurs as clear euhedral flattened wedge-shaped crystals disseminated in a carbonate-clay matrix that is mostly dolomite but that also contains calcite and white lamina of aragonite. The gaylussite crystals are fairly uniform in size and average 1 to 2 mm in maximum dimension; they are present only in permanently wet mud. They are oriented with long axes approximately parallel to traces of stratification and appear to interrupt such traces. Furthermore, their absence within aragonite laminae and their association with calcite suggest that the gaylussite formed by reaction of calcite with sodium-rich solutions. This reaction may have occurred at the bottom of the desiccating lake or diagenetically after burial. The distribution of the crystals within the mud varies considerably; in places the crystals make up nearly 35 percent of the total sediment, whereas a short distance away they may be entirely absent.

Examination of the two or three deeper core sections (locs. DL2, 62, and 65) in the central playa suggests that "varved" strata are characteristic of the older lacustrine carbonate muds of Deep Springs Lake. The "varves" are formed by the alternation of white aragonite lamina with darker colored layers composed primarily of dolomite and calcite in varying proportions, plus a little clastic quartz and 10-A clay. Megascopic gaylussite crystals are found only in the darker layers. The "varved" strata are very similar in appearance to some of the mud layers in the deposits of Searles Lake as described by Smith and Haines (1964) and may be the result of a true annual cycle of lake-level rise and all. Aragonite may be precipitated from relatively fresh spring inflow to the lake (probably stratified at high stages) whereas calcite is formed in more saline waters which result from late summer desiccation; this idea is supported by the absence of aragonite and the presence of calcite in association with thenardite at Deep Springs Lake. In the Searles Lake occurrences, the darker layers often contain considerable microscopic gaylussite, dolomite, and very little or no calcite. The dominance of calcite over gaylussite in the more recent dolomitic deposits of Deep Springs Lake suggests that the formation of gaylussite is continuous but very slow.

The playa immediately west and south of the levee consists of thin nearly monomineralic thenardite crust overlying dolomitic mud. Enclosed in the mud are individual crystals of thenardite. The proportion of crystals decreases abruptly with depth. Textures suggest that the very fine grained dolomite-clay aggregate has been forced aside by the growth of the thenardite crystals.

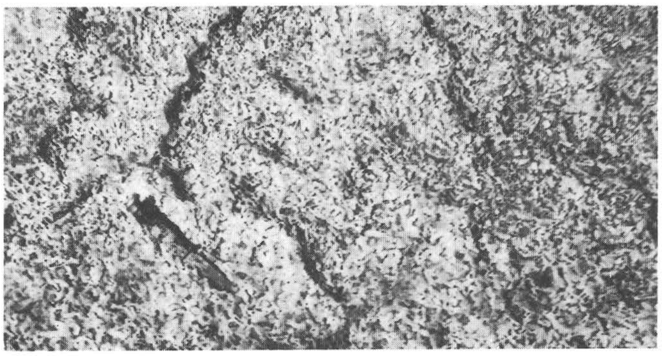


FIGURE 35.—View directly down on salt crust at Deep Springs Lake. Intricate pattern is formed by network of acicular mirabilite crystals now altered to powdery thenardite. Knife gives scale.

Thenardite is also the major constituent of the saline crusts which cover nearly the whole eastside of the playa. The upper surface of many crusts is covered with a network of skeletal powdery mirabilite ($Na_2SO_4 \cdot 10H_2O$) crystals altered to thenardite (fig. 35). The mirabilite forms with the solution and reprecipitation of sulfate as a result of fluctuation in lake levels. Subsequent alteration to thenardite takes place as the temperature rises or as evaporation and concentration of lake waters progress.

Inside the area enclosed by the levee, the layered crust of massive thenardite overlying brown and green dolomitic muds thickens abruptly and becomes brilliant white. As the crust becomes thicker, the mud underneath becomes darker and more fluid. In the central saltpan the average total thickness of saline layers exceeds 1 foot, and the crust is tan, pink, or red. Where the saline crusts exceed 2 or 3 inches in thickness, burkeite becomes a persistent component of the upper crustal layers. Thus, burkeite may be considered the characteristic mineral of the saltpan.

SALINE CRUSTS

Distinctive crustal layering may be detected in the saline crusts at Deep Springs Lake wherever the thickness exceeds about one-half inch. Although individual mineral species can seldom be identified megascopically, differences in mineral assemblage can be roughly discerned by variations in the texture of crustal layers. Layering becomes more distinctive as thickness increases.

The layering of the average saline crust is based on variations in texture and mineralogy and usually falls into a three- to five-fold sequence. Local mineralogic differences are common, but in a typical specimen of thick crust, the topmost layer is very thin, white or clear, highly discontinuous, and confined to the center of a crustal polygon surface. This layer is composed chiefly of halite, commonly in skeletal hopper-shaped crystals. A similar delicately thin layer at places coats vestigal pools of brine on crustal surfaces. In contrast, on the north and east sides of the leveed area, cuspate beachlike deposits of halite crystal mush accumulate on top of layered crust. These deposits are usually brilliant white when dry, and commonly tan when wet. They are formed through redistribution of the uppermost crustal layers by shallow, wind-driven, surface waters saturated with sodium chloride. As a result, a distinct surface layer of halite is commonly missing from the thick crusts in the southern part of the saltpan, where normally there is some surface water derived from the Buckhorn Springs.

Other minerals of the uppermost layer representing the final stages of precipitation from the lake waters include sylvite, bloedite(?), and analcite(?). The sylvite is present mostly as minute cubic inclusions in halite. Analcite has been identified interstitially in tiny icositetrahedrons, but its precipitate origin has not been definitely established.

Rarely, distinct, but very thin patchy lenses of trona occur at the base of the uppermost layer. More commonly, myriad trona needles occur as mats included in halite—as if the trona precipitated just prior to crystallization of halite.

Beneath the surface layer of the thicker crusts are white to pink flaky highly porous gently undulatory laminae usually dominated by burkeite. These laminae are at some places very thin or absent, especially in the crusts outside the central saltpan. Burkeite is commonly present only in traces in other layers. The burkeite is usually massive, anhedral, or tabular, but it may be faintly prismatic or spherulitic (probably due to twinning). (See Foshag, 1935, p. 52–53.) The material is clouded by myriad inclusions, both solid and liquid. Rather than distinct layers, burkeite may also form little clusters surrounding thenardite crystals. Optically, burkeite is most readily distinguished from thenardite by higher birefringence; both minerals are commonly tabular parallel to (110) or (111). The refractive indices of burkeite from Deep Springs are virtually the same as those given by Foshag (1935, p. 54) for material from Searles Lake, but the 2V is slightly lower (approximately 30° as compared with 34°).

Underlying the flaky laminae are massive relatively coarse-grained layers of thenardite which compose more than 75 percent of the total crustal thickness. For the most part, these layers are strikingly uniform in thickness, but in the area of the central saltpan they may be somewhat irregular and possess cavities a few inches across. Single layers have a maximum thickness of about 5 inches. Dry layers are usually dull white. Wet layers may be tan or, more commonly, pink, red, or deep purple. These colors result from algal or bacterial colonies which thrive in the interstitial brine. Surfaces around cavities are commonly darkest colored and colloform or "warty" in texture. The thickest layers are nearly 100 percent thenardite, but aphthitalite is a common minor accessory, and burkeite may be found in traces toward the top. Thenardite habit may differ between layers; most of the thenardite is massive anhedral, but some is dypyramidal or, rarely, even prismatic in form.

The bottom layer of saline crust commonly exhibits great irregularity, especially in the central saltpan area, where the layer appears as a group of crystal clusters projecting downward into the underlying mud. Thickness of the layer may thus range from 2 to 10 inches in a single sample. The crystals composing the clusters are generally dipyramidal; thenardite is usually the sole mineral species, although traces of aphthitalite are also present in the thicker masses. Saline crusts less than 2 inches thick have irregular bottoms, though the irregularity is not great.

The individual disseminated thenardite crystals previously noted in gray-green dolomitic muds near the levee are not found in the fetid black ooze immediately underlying the thick crusts in the central area of the lake. The only salt mineral identified in this material was nahcolite at one or two localities.

Thus, the layer sequence of saline minerals upward from the carbonate muds in the central area of the lake is nahcolite, thenardite, burkeite, trona, halite, and sylvite(?). Small quantities of aphthitalite are persistent throughout the thicker crustal layers.

A definite mineralogical sequence can also be found in the crust which forms on the small closed sag pond of the fault zone at low stage. Although the layer sequence of minerals is the same in the sag pond as it is in the main lake crust, relative quantities of the minerals are markedly different. In the sag pond, the thin surface layer dominated by halite is underlain by delicate porous somewhat granular orange laminae consisting largely of trona; burkeite occurs in the lower layers. These laminae may account for more than 50 percent of the crustal thickness. Distinct thenardite layers are present, but large crystal clusters are not found on the bottom of these layers. Nahcolite, associated with some gaylussite, is common in the underlying mud. Aphthitalite is confined to the upper parts of the thenardite layer.

The structure of the saline crusts in the central saltpan appears dependent largely on crystallization and growth in the thenardite layers. As is apparent in aerial photographs (fig. 3), the crusts are broken intq rough polygonal units, usually crudely hexagonal, which range in size from 5 feet across where first discernible outside the levee the more than 100 feet across in the center of the lake. These polygons all have distinct raised edges which are commonly thrust over one another. Crustal fragments show little or no growth on their outer margins that is not clearly secondary and the result of capillary draw along the already raised edge. Indeed, the outer margins of thrusted fragments on adjacent polygonal units may commonly be fitted almost exactly. This suggests than the main force which causes the raising and thrusting of the polygon edges comes from crystallization and resulting expansion near the center of the polygonal unit and beneath the thrusted layer. This relation may be seen in a section of thick crust (fig. 36). The thenardite layers generally dip toward the center of the crustal ploygon, and all layers thicken toward the center of the polygonal unit. One of the upper thenardite layers usually extends the farthest laterally to form the ploygonal edge. Though crustal porosity is extremely variable, the upper thenardite layer is usually rather massive as a result of frequent re-solution and recrystallization. Thus, additional crystallization in intralayer pore space would probably force crustal growth and expansion to take place.

The width of the raised edge seems to be roughly proportional to the overall size of crustal polygons. Polygons about 5 feet across have an edge 1 to 2 inches wide, whereas 15-foot polygons may have edges 4 to 5 inches wide. Larger polygons were inaccessible to detailed measurement. The relation is probably not regular, however, because of variation in thickness and in number of layers plus the effects of frequent resolution and redistribution by surface inflow.

A secondary crystal growth is common beneath the raised edge of a crustal polygon, as well as on the outer margin of the edge fragments. This material is dominated by halite but includes minor burkeite, sylvite, and aphthitalite. These minerals form through capillary draw and complete evaporation of underlying brine in the crustal cracks.

Two specific kinds of features of the central saltpan area are worth special attention.

1. The man-made trough adjacent to the levee is covered by a layered saline crust nearly identical in composition and structure with the thick crusts near the center of the playa. This crust averages about 3 to 5 inches thick and is underlain by fluid black to dark-green dolomitic mud. However, about 8 inches below the trough surface is another saline layer about 2 inches thick composed of coarsely crystalline thenardite and minor amounts of halite. A third layer of the same composition about 3 inches thick occurs about 12 inches below the trough surface and is separated from the layer above by the same highly fluid black mud. These subsurface saline layers have probably formed through alternate partial filling and evaporation of waters in the trough until the present playa level was reached. Interlayer muds were probably washed in or precipitated during high stages.

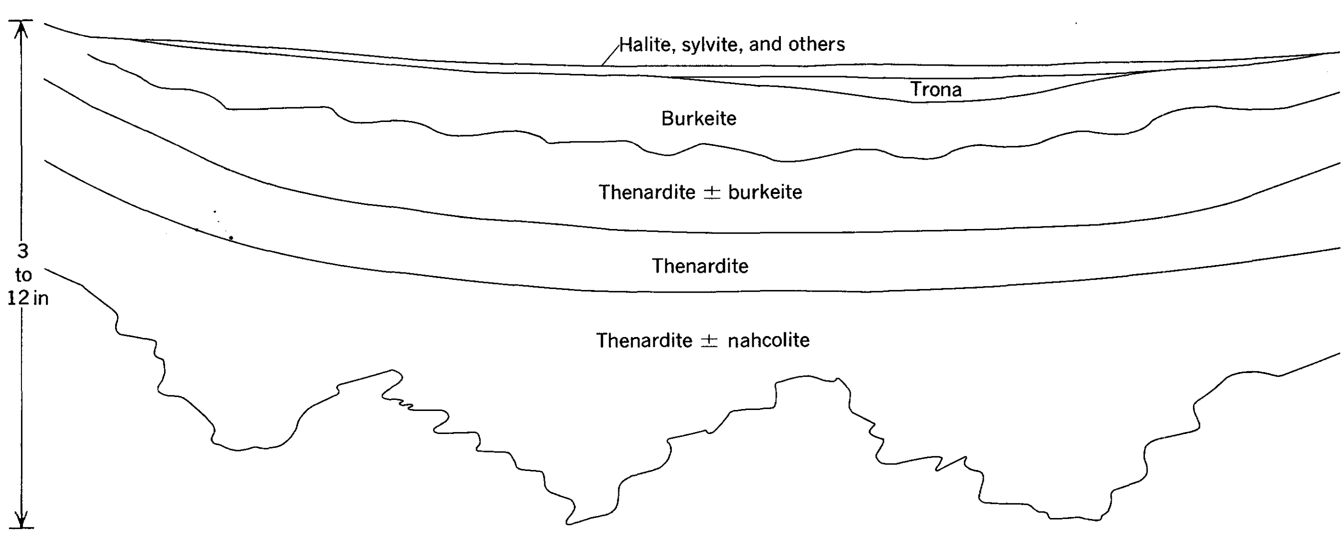


FIGURE 36.—Section of representative salt crust polygon from the saltpan at Deep Springs Lake. The most complete mineral sequence within the crustal layers is shown.

2. Other features of mineralogical interest are associated with the circular gas vents (fig. 32) in the areas marginal to the central saltpan. Only the smaller vents were accessible for detailed observation; these are as much as several inches in diameter and as previously mentioned (p. A41) are roughly conical. These vents are lined with black to dark-green dolomitic mud and are surrounded by salt crust of variable thickness. Gas bubbles vigorously from the smallest vents, which occur north of the central saltpan, but in the thick crusts, the vents appear to show little, if any, gaseous activity. Even when the lake is near complete desiccation, the vents are usually covered by pools of highly colored brine and seem to be surface outlets for subcrustal brine. Saline crusts adjacent to the vents (fig. 32) are usually orange and yellow. Mineralogically, the sequence is the same as elsewhere, except that trona is more abundant in the upper layers, and traces of nahcolite may be present.

In spite of the almost monomineralic composition of the thin thenardite crusts outside the central saltpan, these deposits are complex with regard to structure, texture, and color. Much of this complexity is due to the rapid fluctuations of water levels in this area. Frequent level changes leave their mark on the surface of the saline crust. When submerged beneath a thin sheet of highly mineralized water, the crustal surface is smooth. As surface waters retreat, recrystallization forms an irregular surface of new material. Mean temperature during a particular cycle determines whether the crust will consist of a layer of thenardite or a network of long prismatic mirabilite crystals (fig. 35). Further desiccation promotes the fracturing of the crust into large polygonal patterns. The raising and thrusting of the polygon edges isolates any remaining surface water into pockets for final evaporation. As desiccation and crystallization proceed, smaller polygons form within the larger features. As additional cracks form in the crustal surface, brine trapped underneath is drawn to the surface by capillary action, and ridges of halite form along the fractures. As desiccation proceeds to the underlying mud, further polygonal subdivision occurs and the halite ridges grow large, irregular, and very hard. The entire sequence, as just outlined, may be readily observed in the flats immediately south of the leveed area. Notable differences in color commonly reflect the amount of interstitial brine in wet crusts or underlying mud.

EFFLORESCENT CRUSTS

Capillary efflorescence is primarily responsible for salt crusts found on the west third of the playa and around inflow channels outside the playa area proper. The mineral assemblages in such deposits changed as seasonal variations in temperature and hydrology took place. Any zoning is subject to local variation.

Outside the playa, to the west and south, the saline efflorescences consist largely of halite mixed with small quantities of alkaline earth carbonates in thin white coatings on clays and silts. In the spring areas to the north and east of the lake, surfaces efflorescences are commonly dark in color (so-called black alkali) and commonly contain small amounts of pirssonite. Pirssonite also occurs in the efflorescences on the playa not far from the west edge. Nearer the saltpan, thenardite becomes a common constitutent of the efflorescent crust. In the playa area marked by alluvial channels, the surface efflorescences found on the channel bottoms contrast with those on the divides. The surficial materials on channel bottoms are soft, powdery, and mixed with underlying mud; the divides are covered with very hard, silty, and irregularly ridged efflorescence. Thenardite dominates in the channel bottoms; halite, on the divides. In the admixed mud, dolomite is prevalent in the channels, and calcite, on the divides. Where the channels coalesce and on the mud flats outside the leveed area, the efflorescences are chiefly thenardite and commonly contain small amounts of glauberite and trona.

Efflorescences dominated by halite or thenardite can be distinguished roughly throughout the lake area on the basis of texture. Thenardite crusts are usually soft and powdery. Halite forms hard ridges or large "popcorn balls" in a random pattern. Where the two minerals are present in nearly equivalent amounts, ridges high in halite are separated by smooth platy material containing mostly thenardite.

Efflorescences near springs or the outflow channels from springs reflect the nearby water composition. Seepage channels from the bog-mound springs on the north side of the lake commonly support soft crusts composed of trona, burkeite, halite, aphthitalite, and thermonatrite. The salt forming near channels and around roots near the Buckhorn Springs is mostly thenardite but also includes some halite and traces of aphthitalite. Capillary salts around seeps feeding the closed sag pond include trona, thenardite, aphthitalite, and thermonatrite.

SUMMARY OF CARBONATE AND SALINE MINERALOGY

There are three distinct patterns in the carbonate and saline mineralogy of Deep Springs Lake which include: (1) The areal zoning of precipitate minerals, (2) the layer variation in saline crusts, and (3) the local variation in capillary efflorescences. The minerals, their composition, and the type of deposit in which they occur are given in table 8. The areal sequence of mineral zones from playa margin to center is calcite and (or) aragonite, dolomite, gaylussite, thenardite and burkeite.

Calcite, aragonite, and dolomite are the major components of Deep Springs Lake muds which contain up to 80 percent material of precipitate origin. The dolomite is, at least in part, primary. Some of the carbonate muds which lie 3 feet or so below the top appear "varved" because of light-colored aragonite laminae alternating with darker layers containing calcite and dolomite.

The saline crusts of the Deep Springs Lake saltpan are composed primarily of thenardite (sodium sulfate), but other saline minerals are contained in specific crustal layers. The layer sequence of crustal minerals is, in ascending order, nahcolite, thenardite, burkeite, trona, and halite. Burkeite is a characteristic and abundant mineral in the thick crusts of the central saltpan. Aphthitalite also persists in small amounts in the thick layers. The crusts formed on the closed sag pond show a layer sequence similar to the crusts in the Deep Springs Lake saltpan, but trona (sodium sesquicarbonate) dominates rather than thenardite. Crustal expansion appears to take place predominantly by growth in intralayer pore space. A complete sequence in surface forms is associated with the desiccation saline crust. In the central saltpan, somewhat more sodium carbonate is common in the crusts near gas vents, and additional thenardite layers are present at depth in the trough. Capillary efflorescences are characteristic of the surficial deposits of the western part of the playa. These deposits are dominated by halite and thenardite; the most abundant of these salts is locally indicated by texture. Pirssonite, glauberite, and thermonatrite are restricted to such deposits. Efflorescences around specific inflow areas reflect the anion distribution in the associated inflow waters.

SILICATE MINERALOGY

Silicates compose from nearly 100 percent of the lacustrine deposits well outside the margins of the present playa area to less than 20 percent of the muds underlying the crusts of the central saltpan. The minerals are primarily quartz, feldspar, and clays and traces of detrital amphibole. All but the amphibole can be found even in submicron-size fractions.

Although feldspar was identified in most lacustrine materials, little attempt was made to subdivide the species inasmuch as very fine grain size and interference of other phases commonly obscured characteristic X-ray reflections. However, much of the feldspar is apparently albite or sodic plagioclase and is probably detrital.

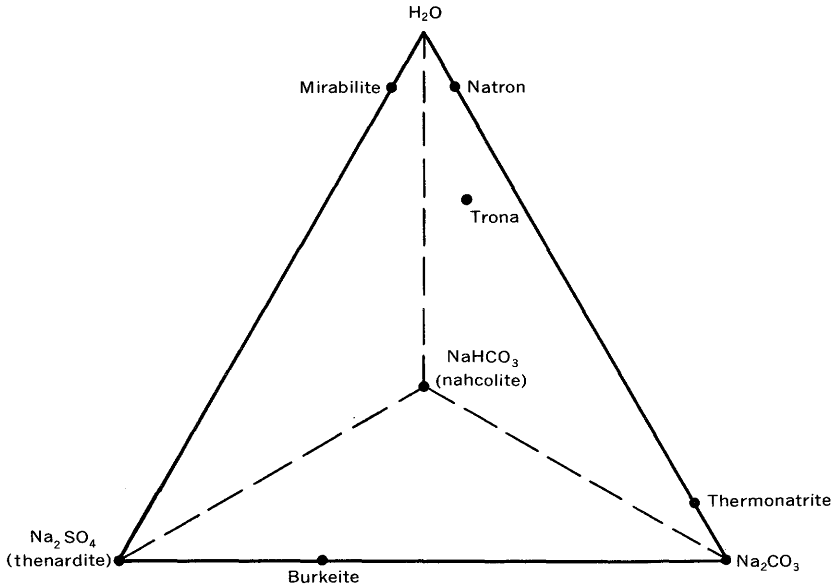
Generally, quartz and 10-A clay minerals are the dominant silicates in the lacustrine deposits. The assemblage quartz, dioctahedral and trioctahedral mica plus minor chlorite, talc, and expandable lattice clays is common throughout the Deep Springs Lake area. The muds of the central saltpan contain a very well ordered mica-type clay close to $2M_1$ muscovite (Yoder and Eugster, 1955) and minor amounts of a poorly defined high-iron "sedimentary chloride" (Warshaw and Roy, 1960, p. 1492), similar to that described by Nelson (1960). Evidence of montmorillonite is sparse, consisting only of very weak reflections about 12 A. Outside the central saltpan area, normal magnesium chlorite, talc, and expandable-lattice clays are more abundant, especially in the parts of the playa which receive detritus from sedimentary and metamorphic rock drainage. To the southwest and northeast of the actual saltpan, chlorite is more abundant, better crystallized, and apparently contains a high proportion of magnesium.

Subordinant amounts of montmorillonite in muds underlying the central saltpan, despite its abundance in marginal lacustrine deposits and its availability from the weathering of either sedimentary or igneous rocks in the basin, suggests diagenetic alteration in the saline environment. Such alteration is probably the result of the high potassium content of the brine at Deep Springs Lake. The chlorite alteration suggested by poor crystallinity and low magnesium content could be the result of leaching and iron oxidation.

GEOCHEMISTRY

The data on the hydrochemistry and saline mineralogy indicate that the Deep Springs Lake saltpan is basically an alkali sulfocarbonate system. Thus, the general geochemistry may be illustrated by phase equilibria in the system Na_2O–SO_3–CO_2–H_2O. The compositional relations of the major phases in the aqueous salt subsystem Na_2CO_3–$NaHCO_3$–Na_2SO_4–H_2O are shown in figure 37.

The interpretation of environment and brine composition in a natural salt system depends on the stability relations of the coexistent minerals. Milton and Eugster (1959) have considered the stability relations of the sodium carbonate minerals as functions of temperature and pCO_2 with reference to the Green River Formation. The assemblages at Deep Springs Lake illustrate primarily the addition of sulfate to the sodium carbonate system. The only mineral compound in the system involving both sodium sulfate and car-



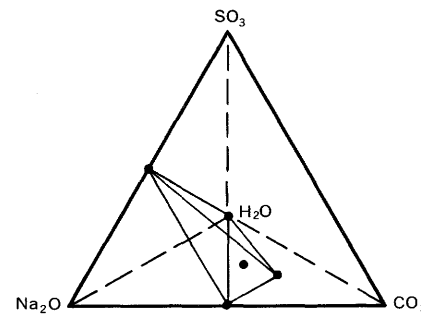


FIGURE 37.—Major phases in the aqueous salt system $NaHCO_3$–Na_2CO_3–Na_2SO_4–H_2O. The larger quaternary Na_2O–SO_3–CO_2–H_2O in which the salt subsystem is outlined, is shown in the smaller diagram to right.

bonate is burkeite, $2Na_2SO_4 \cdot Na_2CO_3$. The stability of burkeite has been studied experimentally (Jones, 1962) in solutions of nahcolite-trona and trona-natron ($Na_2CO_3 \cdot 10H_2O$) buffer assemblages which approximate the total pCO_2 range in nature. The lower stability limits of burkeite were found to be 47° and 24° C, respectively. Addition of saturated sodium chloride solution decreased the lower limit to 33° and 14° C.

The experimental results indicate that the formation of burkeite is favored by relatively high temperature and salinity. Also, above the lower stability limit, burkeite-bearing assemblages are favored over other sulfocarbonate mineral combinations by relatively low partial pressures of carbon dioxide. These results explain why at Deep Springs Lake burkeite is a persistent phase only in the thick central crusts of the main lake, where the concentration of the intercrustal brine and the average annual temperature are relatively high. In the marginal crustal areas, trona plus thenardite or mirabilite—in saturated sodium carbonate solution the mirabilite-thenardite transition is 28°C according to Makarov and Blidden (1928), and in saturated sodium chloride solution without carbonate the mirabilite-thenardite transition is 18°C according to D'ans (1933)—is a more common assemblage because of more dilute inflow and lower temperature. Similarily, trona plus thenardite (or mirabilite in cold weather) predominates in surficial crustal deposits because surface lake waters are relatively dilute.

The saline mineral assemblages at Deep Springs Lake also correlate with experimental conditions relative to pCO_2. Thus, the sodium sulfocarbonate mineral assemblage in the fetid muds beneath the crusts is nahcolite plus thenardite (high pCO_2 assemblage), except where thin crust may allow the formation of burkeite because of higher temperature. Where there is sufficient carbonate, trona plus thenardite occur in the upper parts of crusts where the temperatures or salinities are not high enough for burkeite formation. Thermonatrite, indicative of high temperature and (or) salinity or low pCO_2 occurs only in efflorescent crusts, usually accompanied by burkeite. Most thermonatrite was probably derived from the dehydration of natron. Natron plus mirabilite assemblages have never been observed, but might precipitate from relatively dilute surface waters in very cold weather and are not preserved. Much of the thenardite in surficial crustal deposits appears to have been derived from original mirabilite, but in most of the crustal layers, salinity was sufficiently high that thenardite crystallization was probably primary. A detailed itemization of the sodium sulfocarbonate assemblages identified at Deep Springs Lake, including location and nature of the deposit, is given in table 9. A specific example of the interpretative application of experimental data is at locality 22. Here, the coexistence of trona, nahcolite, thenardite, and burkeite fixes the temperature at 33°C to 47°C depending on the total salinity of the intercrustal brine. The total dissolved-solids content for brine sample DL22 (277,000 ppm) suggests that temperature in excess of 35°C may not have been necessary for burkeite to form.

TABLE 9.—*Saline mineral assemblages in the system* $NaHCO_3-Na_2CO_3-Na_2SO_4-H_2O$ *at Deep Springs Lake*

[In some assemblages, thenardite or thermonatrite have derived from their equivalent decahydrates, mirabilite or natron, prior to analysis]

Assemblage	Locality No. (fig. 2)	Type of deposit	Relative condition
Burkeite+thenardite	9, 11, 24, 53	Upper part of layered crust	Low pCO_2.
	21, 22, 68	Layered crust near pools	High temperature and (or) salinity.
	39	Lower part of layered sag pond crust	Do.
	51	Efflorescent crust	Do.
Burkeite+trona	39	Efflorescence and layered sag pond crust	Average pCO_2.
	22	Upper part of layered crust near gas vent	High temperature and (or) salinity.
Burkeite+trona+thermonatrite	49	Efflorescent crust around "seep channel" to lake from bog mound springs.	High temperature and average pCO_2. High salinity.
Burkeite+nahcolite	8, 11, 15	Mud beneath relatively thin saline crust	High temperature, very high pCO_2.
Burkeite+nahcolite+trona	22	Layered crust next to gas vent	High temperature and (or) salinity and high pCO_2.
Burkeite+thenardite+trona	22	Layered crust near gas vent	High temperature and (or) salinity, average pCO_2.
Thenardite+trona	1, 2, 3, 8, 10, 13, 39	Layered crust	Low temperature and (or) salinity, average pCO_2.
Thenardite+nahcolite	11, 12, 24, 29	Mud under layered crust	Low temperature and (or) salinity, high pCO_2.
Thenardite+trona+thermonatrite	37	Efflorescence around Buckhorn pond seepage	Low temperature and salinity, low pCO_2.

To a considerable extent, the sodium sulfocarbonate mineral sequences indicate original sequences of precipitation as well as gradients in temperature or present solution composition. Thus the lower and outermost lacustrine crusts consist of sulfate, alkali carbonate occurring areally inward or vertically above. Where burkeite is present, it is between and (or) intermixed with the sulfate and carbonate phases. In the seasonally formed crusts of the sag pond where carbonate predominates, nahcolite and thenardite are succeeded upward by burkeite and thence by trona alone. This mineral layering reflects both upward decrease in pCO_2 and the original precipitation sequence. Generally, it appears that the saline minerals and associated brines at Deep Springs Lake have reached local equilibrium in the sense of Thompson (1959, p. 430).

Equilibrium between brines and sulfocarbonate mineral assemblages may be tested in part for conditions at 20°C, the only temperature for which data on the system Na_2CO_3–$NaHCO_3$–Na_2SO_4–$NaCl$–H_2O are available (Teeple, 1929). The data are shown graphically in a triangular projection (fig. 38). This plot represents the composition of saturated solution surfaces with $NaHCO_3$ is projected to the Na_2CO_3 corner. The letters refer to the experimentally determined points; the connecting curves are reduced to straight lines because of lack of data. The divariant fields are labeled according to the solids which coexist with saturated solution; fields which have a single solid phase are undersaturated with carbonate. Teeple's (1929, p. 163) actual data are also listed in fig. 38. If additional data were available each divariant field might be contoured according to pCO_2 and total concentration, but the burkeite field very likely contains the points of maximum total concentration and minimum pCO_2 of solution in equilibrium with a carbonate phase.

Figure 38 also illustrates the relative composition of a few brines from Deep Springs Lake which had temperatures near 20°C at time of collection. The mineral assemblages found coexistent with these brines were as follows:

1	Trona+thenardite.
2	Thenardite.
3A	Trona+halite.
3B	Trona+thenardite.
13	Do.
24	Thenardite.
29C	Thenardite+nahcolite.

If one realizes that much thenardite may be derived from the evaporative dehydration of mirabilite during sampling (as observed at No. 13) or analysis, then the overall agreement is good. The few disparities can readily be explained by the effect of additional components in the system. The additional components in Deep Springs Lake brines would tend to lower the activity of H_2O and carbon dioxide in the system and expand the fields representative of brines which are in equilibrium with solids of lower hydration and carbonation state. An increase in temperature would have the same effect. The burkeite and thenardite fields would be especially favored by such trends and would expand primarily at the expense of mirabilite. Thus, it is not surprising that brines coexistent with these phases at higher temperatures would plot within the mirabilite field on the 20°C diagram. (Compare fig. 38 and fig. 21.) The relations emphasize the need for additional data on the quinary system; data for 30° to 35°C should be particularly useful to the interpretation of saline occurrences at Deep Springs Lake.

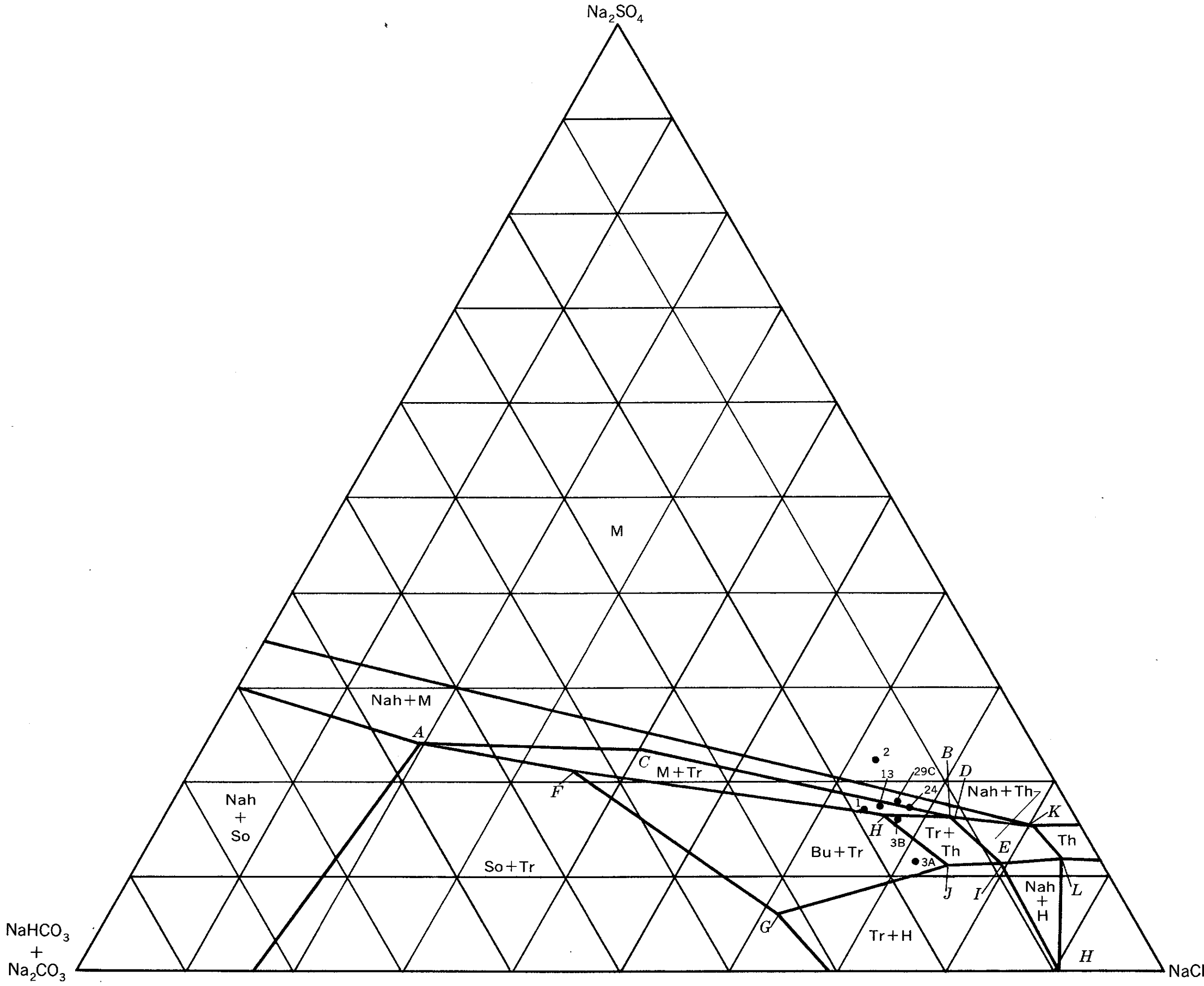


FIGURE 38.—Mineral assemblages in equilibrium with saturated solution in the system $NaHCO_3$-Na_2CO_3-Na_2SO_4-NaCl-H_2O at 20° C and 1 atmosphere total pressure based on the data of Teeple (1929, p. 163). Isobaric invariant assemblages consist of points where there are four solid phases plus saturated solution. Field boundaries in the diagram represent univariant curves with three solids plus saturated solution and are shown straight for lack of further data. Numbered points refer to brine samples which were near 20° C at time of collection. Nah, nahcolite; So, natron; Tr, trona; Th, thenardite; M, mirabilite; Bu, burkeite; H, halite.

Invariant point on fig. 38	Teeple's number	All phases plus solution	Percent of equivalents per million		
			HCO_3+CO_3	SO_4	Cl
A	253	Nah + So + M + Tr	56.0	24.3	19.8
B	256	Nah + M + Th + Tr	11.7	16.2	72.1
C	----	Nah + M + Tr	36.0	23.2	40.8
D	----	Nah + Th + Tr	9.8	12.7	77.5
E	258	Nah + Th + H + Tr	9.1	11.4	79.5
F	254	So + M + Bu + Tr	43.8	21.3	34.8
G	255	So + H + Bu + Tr	32.3	5.9	61.8
H	257	M + Th + Bu + Tr	17.4	16.7	65.9
I	----	H + Th + Tr	9.0	11.2	79.8
J	259	H + Th + Bu + Tr	14.2	11.1	74.6
K	250	M + Th + Nah	4.2	15.6	80.2
L	251	Th + Nah + H	3.2	12.0	84.8
	242	Nah + M	34.8	65.2	----
	1	Th + H	----	11.7	88.3
	243	Th + M	----	15.6	84.4
	247	Nah + Tr + So	83.8	----	16.2
	248	Tr + So + H	30.9	----	69.1
	252	Nah + Tr + H	9.5	----	90.5
	245	So + Nah + M	70.0	30.0	----

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