Clay Mineralogy at the Brine-sediment Interface in the South Arm of Great Salt Lake, Utah

by Leonard L. Hedberg and William T. Parry

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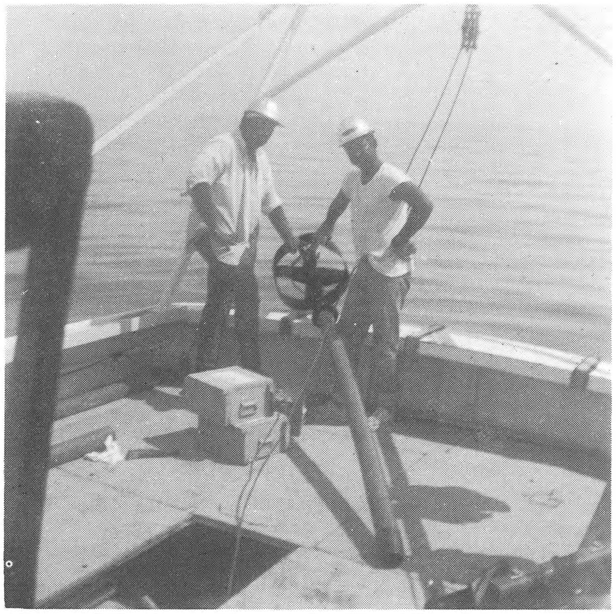

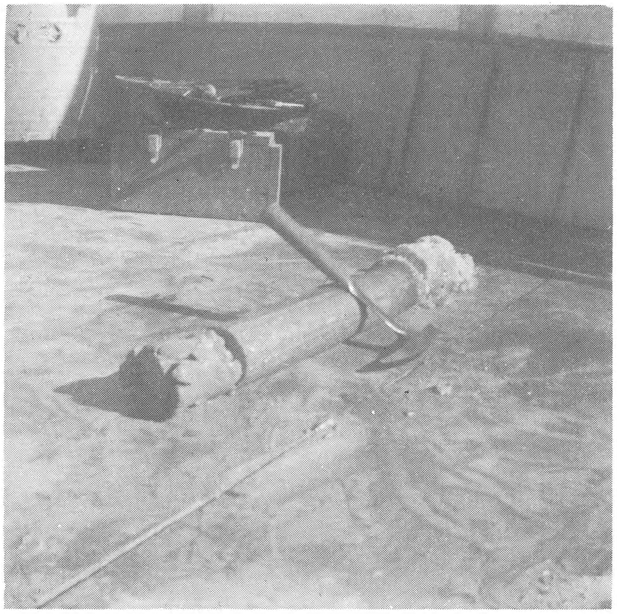

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CLAY MINERALOGY AT THE BRINE-SEDIMENT INTERFACE
IN THE SOUTH ARM OF GREAT SALT LAKE, UTAH

by

Leonard L. Hedberg[1] and William T. Parry[2]

ABSTRACT

The clay mineralogy of 23 cores collected at specified locations in the south arm of Great Salt Lake was determined using X-ray diffraction.

The mean composition of clays from the lake is 51 percent K-mica, 39 percent montmorillonite and interstratified illite-montmorillonite, and 10 percent kaolinite. Great Salt Lake clays are not in chemical equilibrium with lake brine. K-mica is a stable mineral in the lake brine and should form at the expense of kaolinite and montmorillonite.

Pore brines from the clays contain less sodium, potassium and magnesium than lake brine. These ions may be reacting with clay minerals through cation exchange in the process of diagenesis.

INTRODUCTION

Clay minerals in native soils are transported in suspension by streams from upland regions and are deposited in Great Salt Lake. The saline chemical environment of Great Salt Lake is vastly different from the environments of origin or transportation of the clays, thus, diagenetic reactions are expected between the clay minerals and lake brine.

This study (1) examines the clay mineralogical composition of the sediments in the brine-sediment interface of Great Salt Lake, (2) attempts to relate clay minerals of the lake to those of detrital source areas and suspended clays carried by inflowing streams, and (3) studies diagenesis with emphasis on the establishment of equilibrium between Great Salt Lake brine and clay minerals.

Geologic Setting

Great Salt Lake, in northern Utah, occupies a part of the northeast section of the Great Basin. In 1970 the lake had an area of approximately 1,000 square miles, was 27 miles at its widest extreme, was 73 miles long, and had a maximum water depth of 30 feet. Great Salt Lake is a remnant of Lake Bonneville, a Pleistocene lake which had an area of about 20,000 square miles and a depth approximating 1,000 feet at its highest stage. Overflow through Red Rock Pass, Idaho and subsequent dessication has led to the present saline lake (Gilbert, 1890; Morrison, 1966).

The axis of Great Salt Lake trends northwest. The mountains which form islands and associated tombolos trend northward and form bays, mainly at the mouths of tributary valleys. The lake is saucer-shaped in cross section and has wide areas of shallow water bordered by broad mudflats.

The three major tributary rivers are the Jordan, Weber and Bear. The Jordan River, draining the south portion of the Wasatch Mountains and west flank of the Uinta Mountains, enters Great Salt Lake from the south. The lithology of its drainage area includes the following major rock types: quartzite, granite, limestone, sandstone, shale and extrusive volcanics.

The Weber River originates in the Uinta Mountains and drains much of the local Wasatch region east of Great Salt Lake. Abundant quartzite, limestone and a large metamorphic zone including schist and gneiss occur within the Weber drainage.

The Bear River also has its origin in the Uinta Mountains, but meanders northward into southwest Wyoming and south Idaho before returning to Utah. Quartzite, limestone, sandstone, shale, claystone, dolomite, basalt, tuff and small amounts of rhyolite occur in this drainage.

Four basic kinds of bottom sediments are deposited in Great Salt Lake: salt, oolitic sand, algal reef and clays (figure 1). Eardley was the first to study and map the distribution of clays, oolites and bioherms in his classical paper on Great Salt Lake (1938). Cores cut through the salt crust in the deepest portion of the basin show thick sequences of laminated clay.

Climatology and Precipitation

The semiarid continental climate of the Salt Lake basin is marked by large variation in mean temperature resulting from topographic differences. The mean annual temperatures range from 53.2° F at Antelope Island in Great Salt Lake to 44.9° F at Snowville, Utah. The highest temperature recorded in the vicinity of the lake was 111° F at Antelope Island in July 1959; the lowest temperature on record is –32° F in December 1924 at Corrine, Utah. Thus the annual range of extreme temperatures in the lake area has been 143° F. From 1951 to

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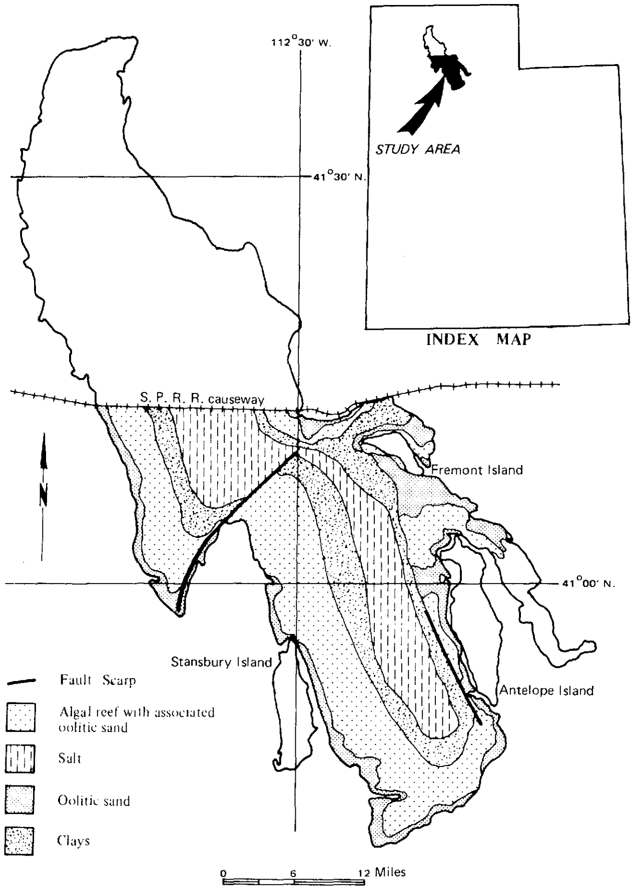

Figure 1. Map showing the distribution of bottom sediments in the south arm of Great Salt Lake.

1960, temperatures of 90° F or higher were recorded for an annual average of 72 days at Antelope Island (Peck and Richardson, 1966).

The average annual precipitation ranges from 5 to 15 inches over the lake to approximately 60 inches in the Wasatch Mountains. The period of heaviest precipitation is during the spring; the driest interval is from June to September. The direct precipitation into the lake varies from 4.5 to 6.0 inches (11.4–15.2 cm) on the west shores and 14 to 18 inches (35.6–45.7 cm) on the east shores. The annual runoff and river discharge vary from .8–1.3 million acre-feet during low water periods to 1.9–3.5 million acre-feet during the high stages (Cohenour, 1968).

Humidity near Great Salt Lake represents extremes in climatology. In December the mean humidity at sunrise and sunset respectively are 81 and 73 percent. During July the mean humidity is 49 and 18 percent. Extremes in relative humidity vary from 100 percent during winter fogs to 4 percent on hot summer afternoons (Peck and Richardson, 1966).

Topography

Great Salt Lake has an average elevation of 4,200 feet (1,280 meters) above mean sea level. It occupies a depression between adjacent fault block mountains which form islands or nearshore uplands in the vicinity of the lake. The watershed of Great Salt Lake is approximately 23,000 square miles (59,570 km^2) and has a relief of 8,285 feet (2,526 m). The highest point in the Great Salt Lake drainage is Hayden Peak which stands at 12,485 feet (3,805 m) in the Uinta Mountains. Islands in Great Salt Lake vary in relief from 35 to 2,445 feet (10.7–745 m) above mean lake level.

One of the most prominent topographic features near Great Salt Lake is the north-south trending Wasatch Mountains, locally greater than 11,000 feet in elevation. This range of mountains meets the west-flanking basin valleys very abruptly along the Wasatch Fault.

Brine Chemistry

Interest in the chemistry of Great Salt Lake was shown as early as 1850 (Richardson, 1906, p. 34). Before the 1960's when Utah Geological and Mineralogical Survey and United States Geological Survey commenced water sampling programs, data were collected infrequently and usually without accurate identification of the location or depth from which the samples were taken.

It is believed that previous to 1957, Great Salt Lake brine was laterally a homogeneous solution and only varied in composition vertically with laminar stratification. Since 1957 when the Southern Pacific Railroad completed a permeable causeway dividing the lake into two bodies of water, the brine chemistry has undergone drastic changes. The north one-third of Great Salt Lake, separated from its parent body by the causeway, acts as an evaporation pond where mineral solids are concentrated. This body of brine remains nearly saturated with respect to sodium chloride. During the warm summer months when evaporation is most active, sodium chloride is precipitated from the water.

Three distinctive brines from the Great Salt Lake have been identified: (1) the sodium chloride-saturated brine north of the Southern Pacific causeway, (2) the typical Great Salt Lake brine south of the causeway from the surface to a depth of 24 feet, and (3) the low-lying dense brine south of the Southern Pacific causeway below a depth of 24 feet (table 1). The brine south of the causeway from the surface to a depth of 24 feet is much less concentrated than the north brine. The three main tributary rivers to Great Salt Lake enter south of the causeway and dilute the brine in the south arm. The only observable link between the two bodies of water is two culverts 15 feet wide and approximately 20 feet high which permit brine to flow through the causeway.

Table 1. Mean composition of three distinctive Great Salt Lake brines compared with ocean brine.

Ion	Oceanic[1]		Great Salt Lake, Southern portion less than 24 ft.[2] (January, 1970)		Great Salt Lake, Northern portion[2] (January, 1970)		Great Salt Lake, Southern portion deeper than 24 ft.[2] (October, 1969)	
	gpl	percent	gpl	percent	gpl	percent	gpl	percent
Cl	19.70	55.4	111.0	53.3	185.50	58.4	146.60	53.1
Mg	1.32	3.7	7.05	3.4	13.95	4.4	8.89	3.2
Ca.	0.41	1.2	0.21	1.0	0.23	.7	0.29	1.0
Na	10.93	30.7	66.60	32.0	85.20	26.8	89.70	32.5
K	0.39	1.1	6.28	3.0	8.91	2.8	6.67	2.4
SO_4	2.74	7.7	14.95	7.18	23.43	7.4	23.66	8.6
					ppm			
Br	65	.07	80.0	.09	162.8	.199	118.7	.142
Li	0.1	.0001	32.2	.037	59.2	.072	36.6	.044
B	--	--	44.4	.051	53.8	.066	40.9	.049
Total Solids (gpl)	35.57		208.22		317.5		276.01	
Density (g/ml)	1.035		1.138		1.220		1.195	

[1] Whelan, J. A., 1969, p. 9.

[2] Calculated from results of sampling by Utah Geological and Mineralogical Survey.

The third chemically distinctive brine in the lake occupies the deepest portion of the south basin below 24 feet. This brine is much more concentrated than the upper brine in the south arm of the lake and is only slightly less concentrated than the brine north of the causeway. Its distinctive characteristics are a strong odor of hydrogen sulphide and a murky appearance caused by suspended organic matter. This low-lying dense brine is divided from the upper brine by an abrupt interface. The brine is believed to come from the north arm through the culverts where a counterflow at depth from north to south is observed.

The deep dense brine south of the causeway immediately overlaid most of the cores studied in this investigation. Clays taken from samples in water shallower than 24 feet, along the margins of the lake, were overlain by the less concentrated typical Great Salt Lake brine.

DIAGENESIS OF CLAY MINERALS AND REVIEW OF LITERATURE

Brine Concentration and Clay Mineral Diagenesis

In playa lakes such as Great Salt Lake, the chemical composition of water is contingent upon weathered source rocks of the surrounding area. Such factors as rates of weathering and evaporation and length of time involved in the dessication process are critical in determining brine chemistry. Brine concentrations and constituent ions are important in understanding the diagenesis of clay minerals.

Cation Exchange in Clay Minerals

Three basic causes of cation exchange are (Grim, 1953, p. 132-134): (1) broken bonds around the edges of the silica-alumina units give rise to unsatisfied charges which are balanced by adsorbed cations; (2) substitutions within the lattice structure of trivalent aluminum for quadrivalent silicon in the tetrahedral sheet and of ions of lower valence, particularly magnesium, for trivalent aluminum in the octahedral sheet, resulting in unbalanced charges in the structural units of some clay minerals; (3) replacement of exposed hydrogen atoms in the hydroxyl ion by exchangeable cations. (Volk and Jackson, 1964; Kerns, 1967).

The replaceability of interlayer potassium is affected by the magnitude of the layer charge, particle size, presence of hydrogen, nature of the replacing cation and the fixed or native character of the potassium ion (Barshad, 1954). Differences in hydroxyl orientation may also be critical for interlayer potassium replacement (Bassett, 1960).

Recent experiments have shown that potassium in muscovite, biotite, phlogopite, illite and vermiculite-hydrobiotite can be replaced by sodium in times ranging from less than 10 hours to more than 45 weeks, depending on sodium and potassium activities in the reacting solutions (Scott and Smith, 1966). Chloritic and illitic clay types may ultimately develop from montmorillonite in the marine environment (Whitehouse and others, 1956). Gibbsite $Al(OH)_3$ and kaolin-type crystals were produced in the laboratory from dilute solutions of aluminum and silica at a temperature near 25° C (Polzer and others, 1967).

Diagenic Reactions and Interstitial Waters

To understand the possible diagenetic reactions occurring in a specified depositional environment, it is necessary to examine the interstitial waters. Marine waters differ significantly from interstitial waters in marine sediments (Swanson and others, 1967). MacKenzie and Garrels (1965) measured the rate of dissolution of clay minerals in sea water and showed that rates are high enough to indicate that the ocean responds rapidly, chemically, to added detrital silicates.

Surface brine composition in Great Salt Lake differs from ground water brine in the Great Salt Lake Desert near Wendover, Utah. Ground water brine near Wendover contains excess KCI where Great Salt Lake brine contains excess sulfate (Nolan, 1927). It is believed that some of the observed differences in brine chemistry can be explained by the diagenetic reaction of siliceous clay minerals with the associated brines.

Clay Mineral Stability and Change

Glauconite, phillipsite, illite and chlorite are assumed to be stable minerals in a marine environment, and kaolinite and montmorillonite are stable in fresh water (Weaver, 1967). Conway (1945) proposes that the amount of potassium in sea water increased gradually from early Precambrian reaching a peak concentration in late Precambrian when the large scale development of organic material aided in fixing quantities of potassium ion in clays. Potassium ion continued to be extracted at a faster rate than it was being supplied to the sea, and from the Mesozoic era onwards glauconite and presumably illite played the dominant role in removing potassium ion. Clay minerals are presently sorbing from the ocean more sodium and magnesium ions than potassium ion (Weaver, 1967).

Potassium in interstitial waters from Pacific Ocean muds was reported 30 times more concentrated than that in the overlying ocean waters (Shishkina, 1959). Siever and others (1965) studied the interstitial waters in a wide variety of marine muds and determined that potassium is usually more abundant, commonly by a factor of two, than in sea water. Carroll and Starkley (1960) found that the replaceability of hydrogen ions from the clay minerals they examined follow the order $Ca^{++} > Mg^{++} > Na^{+} > K^{+}$.

Phase Equilibria

Hess (1966) studied phase equilibria of some minerals in the $K_2O{-}Na_2O{-}Al_2O_3{-}SiO_2{-}H_2O$ system at standard temperature and pressure. He proposed a series of activity diagrams to explain the stability of selected minerals with varying concentrations of SiO_2, K, H and Na. The work of Hess (1966) and Helgeson and others (1969) give a firm thermodynamic basis for correlating brine chemistry and diagenetic reactions with detrital sediments.

Clays of the Great Salt Lake Region

Modern lake sediments of Great Salt Lake and those of post-Provo age immediately preceding the present lake have approximately the same clay-mineral composition: montmorillonite shows poor organization resulting from numerous sodium ions within the clay structures; illite is well organized showing only a small amount of degrading; kaolinite composed approximately 10 percent of the clay fraction; attapulgite-sepiolite clay minerals were not found (Grim and others, 1960).

The Alpine and Provo sediments of Great Salt Lake were somewhat different from modern lake clays. These Bonneville clays were distinguished by a highly degraded chlorite, a very well-ordered but low content montmorillonite, and abundant illite (Grim and others, 1960).

Guven and Kerr (1966) studied the effects of weathering on the structures of mica-type clay minerals at Willard Bay near Great Salt Lake. They concluded that the mica clay minerals had been subjected to intensive weathering which attacked the interlayer portion of the structure of layer silicates most intensively where the bonds are weakest between the mica-type layers.

Clays from six playa lakes in the Great Basin contained illites, vermiculites and montmorillonites (Guven and Kerr, 1966).

Eardley and Gvosdetsky (1960) examined the clay minerals taken from a continuous core from the surface to a depth of 650 feet near Saltair on the south shore of Great Salt Lake. They found the samples to contain approximately 20 percent kaolinite, 45 percent illite and 35 percent montmorillonite.

The clay minerals reported in a study of the Oquirrh Mountains include (Gilluly, 1932): phillipsite, chlorite, sepiolite, kaolinite, halloysite and montmorillonite.

Alpha sepiolite was observed and photographed from clay samples taken in Little Cottonwood Canyon (Kerr and others, 1950).

PROCEDURES

Collecting Samples

Sample cores were collected from Great Salt Lake using a hydroplastic corer on loan to the Utah Geological and Mineralogical Survey from the Environmental Science Services Administration (figure 2). The corer successfully penetrated 29 cm of halite, permitting collection of clay samples underlying the thickest salt crusts in the south arm of Great Salt Lake (figures 3, 4).

The 8.89 cm (3.5 inch)-diameter core barrels were constructed of high impact polyvinyl chloride pipe. Forty-four cores from 2 to 9 feet long with a mean of 5 feet were obtained during the summer of 1969. After removal from the lake, the cores were capped and stored in the core barrels.

Twenty-three selected cores were extruded and clay samples for X-ray identification were taken (figure 5). In some of the longer cores where lithologic change was obvious, two or three samples were taken from the same core.

Treatment of Samples

Core samples were disaggregated using a high speed mixer and washed in distilled water until all soluble salts were removed. The washed clays were dispersed in 10 ml of 20 percent Calgon solution and placed in a graduated cylinder. The upper 10 cm of the sedimentation column was removed after an 8-hour interval.

Each sample was divided into three portions. The first portion was repeatedly washed in .3 molar CaCl solution and glycerated using a 20 percent glycerol solution. The second clay sample was repeatedly rinsed in .3 molar KCl solution. The final sample was left without chemical treatment.

X-ray mounts were prepared by smearing clay suspended in a water droplet on a petrographic microscrope slide. One slide was prepared from the calcium saturated-glycerated sample and one from the untreated sample. Two slides were prepared from the potassium-saturated sample. The first potassium slide was heated to 250° C for one hour and the second to 550° C for the same period of time.

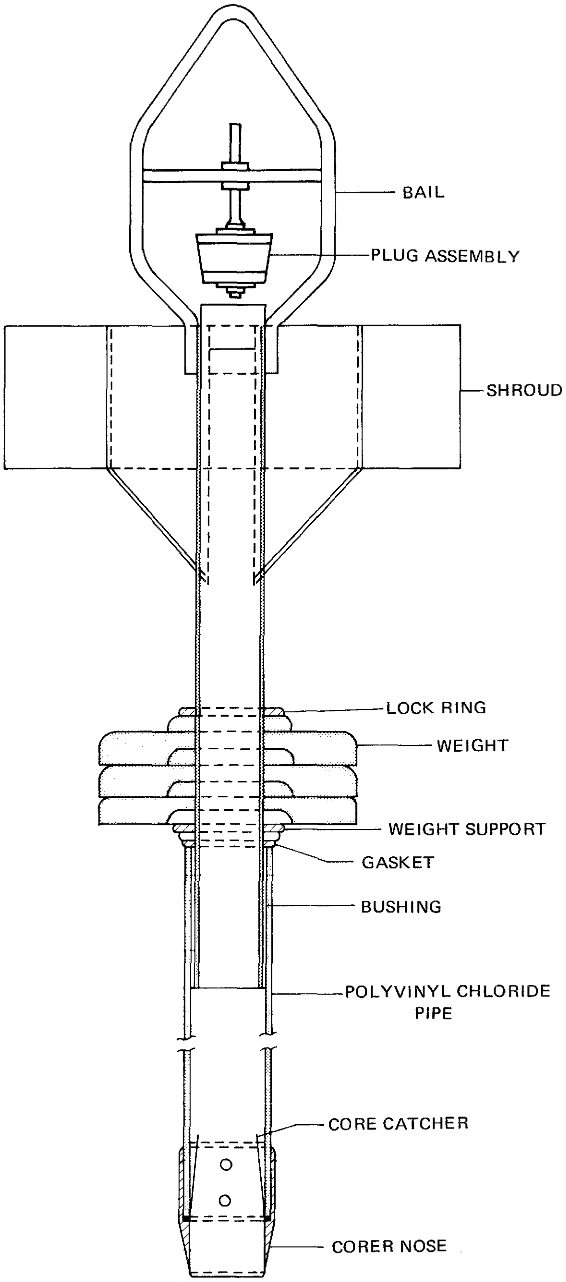

Figure 2. Schematic drawing of hydroplastic corer.

X-ray Analysis

Each sample was examined on a General Electric XRD-5 X-ray diffraction unit. The X-ray diffractograms were obtained using Ni filtered CuK α radiation with 35 kvp and 29 ma. The angular range of 2° to 35° 2 θ was scanned at rates of .5° and 1° per minute. X-ray peaks were identified and d-spacings in angstroms computed from tables. Study in this fashion provided information on basal spacings in the clay lattices.

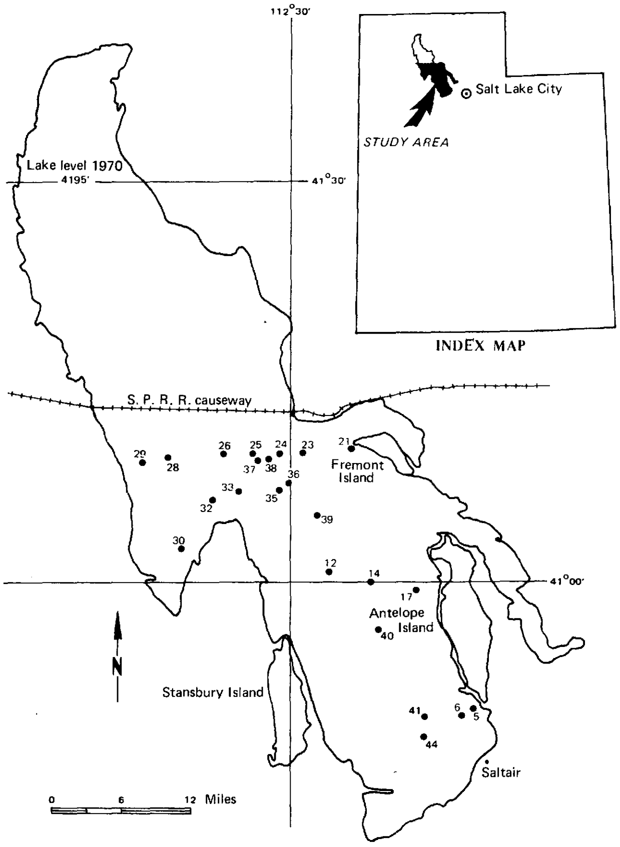

Figure 3. Locations of selected cores in the south arm of Great Salt Lake.

The criteria used for identification of clay minerals were:

1. Illite or K-mica. 10, 5, and 3.3 Å spacings which were unaffected by glyceration or heat treatments (figure 4).
2. Montmorillonite. 18 Å spacing when Ca saturated and glycerated. Collapsed to 10 + Å when heated to 250° C. Also 10 + Å when potassium saturated and air dried (figure 5).
3. Kaolinite. 7, 3.5 Å spacings destroyed at 550° C (figure 4).
4. Chlorite. 14, 7, 3.5 Å spacings not destroyed at 550° C. 14 Å may be missing or very weak.
5. Interstratified clays. Behavior intermediate to the particular clays involved.

DATA PRESENTATION

Clay Mineralogy

Kaolinite, illite, montmorillonite and interstratified illite-montmorillonite were found in all samples taken from Great Salt Lake (table 2). The samples were studied using the quantitative analysis technique of Pierce and Siegel (1969). Interstratified illite-montmorillonite was grouped with montmorillonite in this quantitative presentation. The relative value for K-mica was obtained by multiplying the area of the 10 Å peak by a factor of four. The 7 Å peak area was multiplied by two for the approximate amount of kaolinite. The integrated intensity of the 17 Å glycolated peak is equal to the relative amount of montmorillonite. Integrated intensity showing total area under the X-ray peak was used rather than peak intensity (Brown, 1961, p. 512).

The mean composition for all samples X-rayed is 51 percent K-mica, 39 percent montmorillonite and 10 percent kaolinite. The K-mica and kaolinite showed moderate to good crystallization. Montmorillonite was generally poorly crystallized and not significantly more pronounced in peak intensity than the interstratified clays.

Samples of water from each of the three major tributary rivers were collected near their point of discharge into Great Salt Lake and the suspended matter allowed to settle from the water. The water was decanted and the sediment removed for X-ray analysis (table 3).

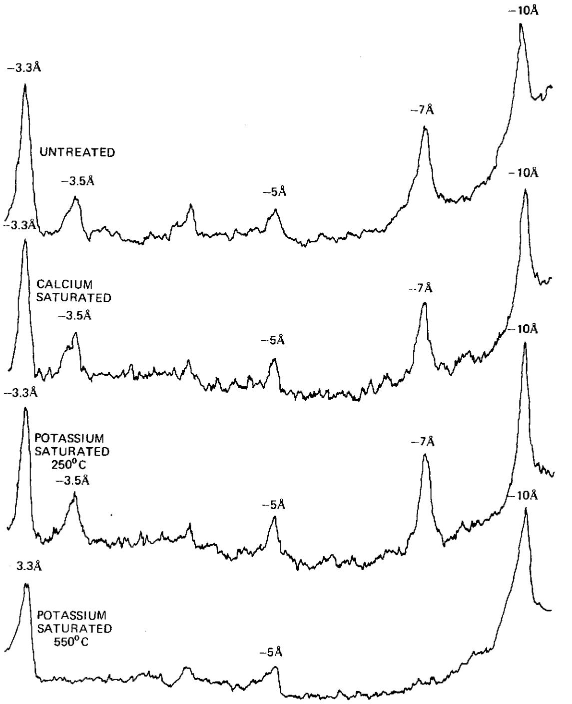

Figure 4. X-ray diffractograms of core 23 at 275 cm depth. K-mica is identified by the 10, 5, and 3.3 Å spacings which are unaffected by glyceration or heat treatments. Kaolinite is identified by 7 and 3.5 Å spacings which are destroyed when potassium saturated and heated to 550° C.

Samples of lake brine from the sediment-brine interface were collected at each coring location and analyzed (table 4). Clay samples were washed in distilled water and the solution analyzed to determine the composition of interstitial brine (table 5). Mean compositions for the ions of both brines were tabulated (table 6).

Silicon analyses were run on the Hellige turbidimeter. These analyses were considerably higher than expected from previous work completed by the United States Geological Survey (Hahl and Handy, 1969). A molybdate blue method of analysis yielded more satisfactory results (table 7).

Porosity tests on samples taken from five cores showed little variation in percent of pore spaces laterally or at depth to 275 cm. Specific gravity of lake muds was measured using a pycnometer (table 8).

Miscellaneous Mineralogy

In addition to the clay minerals, the presence of aragonite, calcite, dolomite, halite and quartz was noted on the X-ray diffractograms of Great Salt Lake sediment.

Table 2 Clay mineralogy of Great Salt Lake cores.

Core No.	Sample Depth (cm)	percent		
		Montmorillonite	K-mica	Kaolinite
5	9	40	54	6
5	24	36	54	10
5	45	49	45	6
5	113	42	49	9
6	9	37	52	11
6	30	64	27	9
6	74	32	60	8
12	46	49	44	7
14	10	27	66	7
14	50	19	70	11
14	100	35	54	11
17	10	27	60	13
17	50	28	62	10
17	93	36	57	7
21	178	55	37	8
23	275	26	62	12
24	114	38	52	10
25	220	41	54	5
26	65	38	55	7
28	107	56	35	9
29	8	38	54	8
30	81	44	46	10
32	46	52	38	10
33	89	38	54	8
35	4	59	28	13
35	70	34	58	8
36	55	35	52	13
36	163	45	45	10
36	183	33	57	10
37	84	46	48	6
38	190	41	47	12
39	122	30	56	14
40	110	40	53	7
41	107	28	61	11
44	60	30	48	22

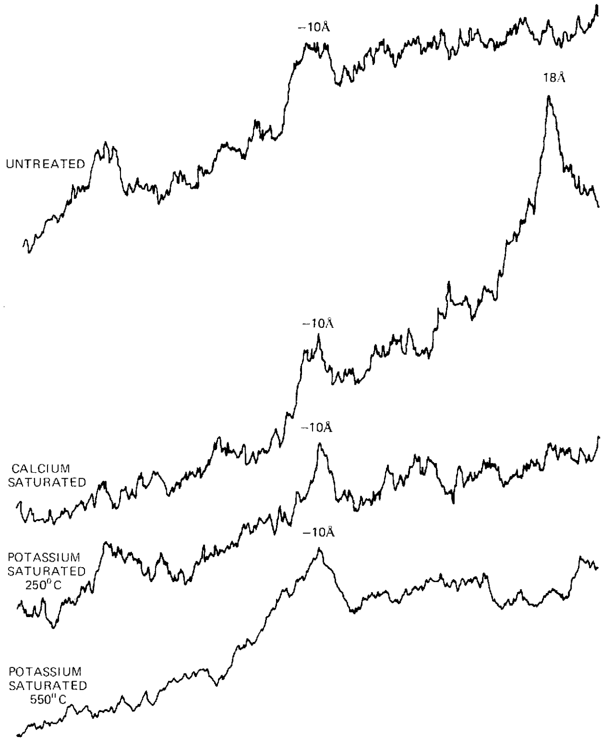

Figure 5. X-ray diffractograms of core 44 at 60 cm depth. Montmorillonite is identified by the 18 Å peak which is prominent on the calcium saturated and glycerated diffractogram. The peak collapses when the sample is potassium saturated and heated.

Table 3. Clay mineralogy of suspended river sediment.

Tributary river	percent		
	Montmorillonite	K-mica	Kaolinite
Weber River	32	63	5
Bear River	37	56	7
Jordan River	28	60	12

Calcite and quartz were found in the suspended load of the Bear, Weber and Jordan rivers. Quartz was three times more abundant in the Bear River than either the Jordan or Weber rivers.

Dolomite was not found in any of the three major tributary rivers to Great Salt Lake but occurred abundantly in lake sediments.

INTERPRETATION OF DATA

The phase relations in the K_2O–Na_2O–Al_2O_3–SiO_2–H_2O system as calculated by Hess (1966) are shown in figure 6. Figure 7 is an isoplethic section with computed equilibrium data for Great Salt Lake brine at the brine-sediment interface for each coring location. Figures 8 and 9 are isoplethic sections with data from

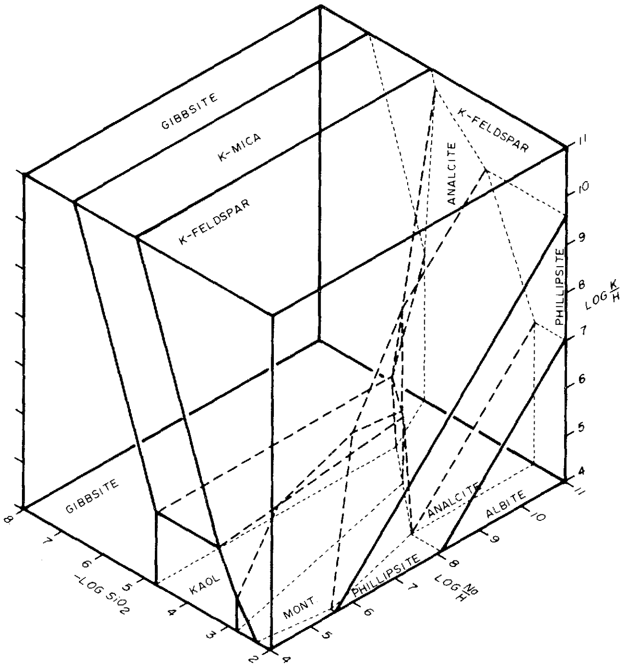

Figure 6. Complete phase diagram for the system K_2O–Na_2O–Al_2O_3–SiO_2–H_2O at 25° C and 1 atmosphere pressure (after Hess, 1966).

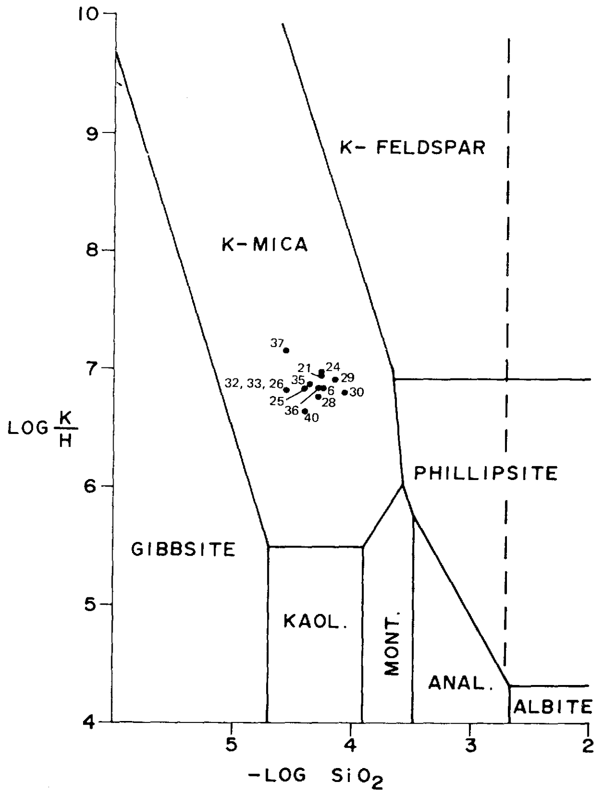

Figure 7. Isoplethic section for lake brine at log Na/H = 8.5. Dotted line equals solubility of amorphous silica at 25° C (after Hess, 1966).

interstitial brine analyses plotted for each core sample. The activity ratios shown in these figures were obtained by assuming activity coefficients of one and calculated from the brine analyses of tables 4 and 5.

Hess constructed the phase diagram (figure 6) using a relatively small number of equilibria:

K-mica – Kaolinite
Gibbsite – Kaolinite
Phillipsite – K-feldspar
K-mica – K-feldspar
Kaolinite – Na-montmorillonite
Albite – Analcite

These equilibria were plotted as planes in three dimensional space. Where these planes intersect, a plane representing the equilibria between the two planes exists. Hess's phase diagram was made from plotting the intersection of these planes.

The brine compositions shown in tables 4 and 5 fall in the K-mica stability field while the actual assemblage is K-mica, kaolinite and montmorillonite. These minerals probably do not represent an equilibrium assemblage but rather a mixture of unstable minerals. Montmorillonite and kaolinite are unstable in lake brine (figure 7). K-mica is forming at the expense of both kaolinite and montmorillonite. The montmorillonite–K-mica reaction is as follows:

$$9Na_{0.33}Al_{2.33}Si_{3.67}O_{10}(OH)_2 + 7K^+ + 3H^+ = 7Kal_3Si_3O_{10}(OH)_2 + 3Na^+ + 7H^+ + 12SiO_2$$

The Kaolinite–K-mica reaction is written as:

$$3H_4Al_2Si_2O_9 + 2K^+ = 2Kal_3Si_3O_{10}(OH)_2 + 2H^+ + 3H_2O$$

Samples of suspended clay taken from the Jordan, Weber and Bear rivers averaged 8.5 percent more K-mica than clays X-rayed from the sediments of Great Salt Lake. This could be the result of (1) inaccuracies in quantitative analysis, (2) change in weathering intensity in source area producing less kaolinite and montmorillonite and more K-mica now than when sediments were deposited, (3) the K-mica has been used up in lake sediments to form other minerals such as K-feldspar or phillipsite, or (4) the present lake clays are largely the result of wind blown detritus from valley soils. Eardley (1938) concluded the origin of Great Salt Lake clays to be 20 percent transported by lake currents derived largely from tributary rivers, 44 percent transported by wind and 36 percent precipitated from solution.

Grim (1968, p. 514-515) reviewed a study of clay mineral composition of soils by Alexander and others (1939):

> . . . kaolinite is the dominant clay-mineral component of red and gray podsolic soils, with illite frequently present but in quantities usually less than about 10 percent. In Prairie soils both illite and kaolinite are usually present, with the former slightly the more abundant. Small amounts of smectite are often present in the hardpan horizon of Prairie soils. In the series of Chernozem soils studied by these investigators, illite was the dominant component, and kaolinite and smectite were also present in about equal amounts; and in a series of Desert soils from California, illite and smectite were the dominant clay minerals.

Work by Knox (Grim, 1968) shows that smectite (including montmorillonite) and illite predominate the soils of arid regions. Assuming the Great Basin to be arid to semiarid in climate and illite the dominant mineral of such climatic zones, the anomaly of Great Salt Lake containing less illite and more kaolinite and montmorillonite than found in tributary rivers is not explained by the wind transport hypothesis.

Table 4. Chemical composition of Great Salt Lake brine at the brine-sediment interface from coring locations, in g/l.

Core No.	Na	K	Mg	Ca	Cl	SO_4
21	59.0	5.02	6.80	.10	109.74	15.06
24	73.5	6.7	9.10	.09	138.06	19.79
28	87.0	10.73	13.50	.09	161.07	24.32
29	66.0	4.62	7.00	.09	122.13	15.27
32	88.0	8.58	12.05	.09	177.00	26.71
33	88.0	8.58	12.05	.09	177.00	26.79
35	57.5	4.79	6.95	.11	111.51	15.10
36	65.0	5.84	7.90	.11	125.67	21.73
37	86.0	8.54	11.90	.11	166.38	22.80
40	94.0	6.40	8.85	.13	177.00	23.46
25	90.0	8.65	11.75	.08	171.69	25.89
6	76.0	5.84	8.35	.11	143.37	20.37
26	94.0	8.65	12.50	.09	178.77	27.62
30	61.5	5.02	7.00	.09	115.05	15.27

The mean interstitial brine analyses are compared with the mean lake brine analyses at the specified coring locations (table 6). All ions except calcium are more concentrated in the lake brine than in the interstitial brine. The observed differences in concentrations of magnesium, potassium and sodium are likely to result from diagenetic reactions of clay minerals.

Clay minerals have the property of sorbing certain anions and cations and retaining these in an exchangeable state around the outside of the silica-alumina clay-mineral structure unit. This exchange generally occurs in a water solution treated with specific ions (Grim, 1953, p. 126). An ideal environment exists for ion exchange in Great Salt Lake. Sodium, potassium and magnesium cations are present in interstitial waters surrounding clay mineral tetrahedron-octahedron structures, but are not as abundant as the same ions in the mobile lake brine overlying the sediments. Great Salt Lake brine overlying the clays is more concentrated in the following ions than the interstitial brine (table 6): (1) sodium, 9 g/l, (2) potassium, 1.5 g/l and (3) magnesium, 7.5 g/l. These ions may be less abundant in the interstitial brine because of attachment to the clay minerals through cation exchange.

Table 5. Chemical composition of interstitial brine from clay samples of Great Salt Lake in g/l.

Core No.	Depth (cm)	Na	K	Mg	Ca	Cl	SO_4
25	220	97.4	7.83	1.20	2.89	118.5	24.1
5	113	41.3	3.68	1.49	2.18	71.3	11.85
29	8	53.6	4.63	1.81	1.92	98.9	39.6
36	163	94.7	8.38	3.67	3.9	121.8	24.9
33	89	75.9	5.54	2.17	3.37	139.2	24.5
6	56	57.8	4.57	1.76	2.57	103.7	15.9
36	55	54.4	4.7	1.38	2.89	100.5	16.2
35	70	56.2	5.53	1.47	2.82	95.8	17.6
24	114	76.9	6.29	2.68	2.80	155.7	20.6
40	110	83.2	6.74	2.17	3.37	139.0	23.2
26	65	79.8	6.64	2.17	3.14	157.4	25.8
38	190	87.7	6.62	3.08	3.65	113.0	22.1
21	178	47.6	3.72	1.13	2.6	79.7	10.25
28	107	47.6	3.46	1.19	2.63	91.4	13.85
35	4	82.4	7.00	2.37	2.93	177.4	23.6
30	81	56.0	4.64	1.47	2.49	112.0	13.9
32	46	51.5	4.47	1.69	2.53	91.7	16.25
37	84	96.6	7.89	2.99	3.59	149.0	28.00
12	46	65.8	5.3	1.93	2.77	122.3	18.1

Aragonite ($CaCO_3$) is a precipitate of Great Salt Lake and is concentrated in the lake sediment. The near-surface waters with free calcium in solution are exposed to abundant CO_2 where the carbonate ion is produced which combines with the calcium cation to precipitate aragonite. Thus calcium is 29 times more concentrated in the interstitial waters than in the mobile lake brine. Eardley (1938) identified aragonite as the original precipitate of the lake which converts to calcite to produce the radiating structure in oolites.

Calcite occurs in the suspended load of all major tributary rivers to Great Salt Lake as well as in lake sediment. Dolomite occurs in lake sediment but not in the suspended load of tributary rivers. The calcium ion in calcite may be replaced by a magnesium ion when coming in contact with the brine and its high magnesium

Table 6. Mean concentrations of Great Salt Lake brine at coring locations and interstitial brine from clay samples in g/l.

	Na	K	Mg	Ca	Cl	SO_4
Interstitial brine	68.76	5.66	1.99	2.90	117.81	20.54
Lake brine at coring locations	77.54	7.00	9.69	.10	148.17	21.44

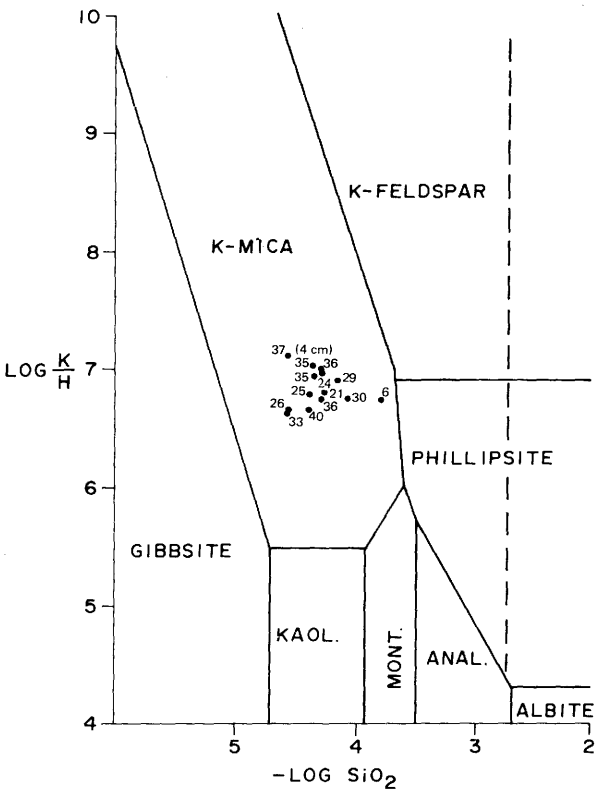

Figure 8. Isoplethic section for interstitial brine at log Na/H = 8.5. Dotted line equals solubility of amorphous silica at 25° C (after Hess, 1966).

Table 7. Specific gravity, pH, and SiO_2 concentrations of brine from the brine-sediment interface at specified coring locations.

Core No.	Specific Gravity	pH	(Molybdate blue method) SiO_2 ppm	(Hellige Turbidimeter method) [1]SiO_2 ppm
21	1.128	7.85	3.0	15
24	1.164	7.75	2.8	11
28	1.196	7.35	2.6	17
29	1.129	7.85	3.9	11
32	1.207	7.5	1.4	23
33	1.207	7.5	1.4	18
35	1.129	7.8	2.4	14
36	1.148	7.7	2.7	8
37	1.197	7.83	1.4	16
40	1.205	7.45	2.0	13
6	1.171	7.7	3.0	
25	1.204	7.5	2.0	
26	1.208	7.55	1.4	
30	1.132	7.7	4.7	

[1] Results were unsatisfactory and not used in equilibrium determinations.

content (7 g/l), or perhaps the dolomite observed in the lake is primary in origin. Eardley (1938, p. 1353) concluded the dolomite crystals in Great Salt Lake clays to be primary in origin.

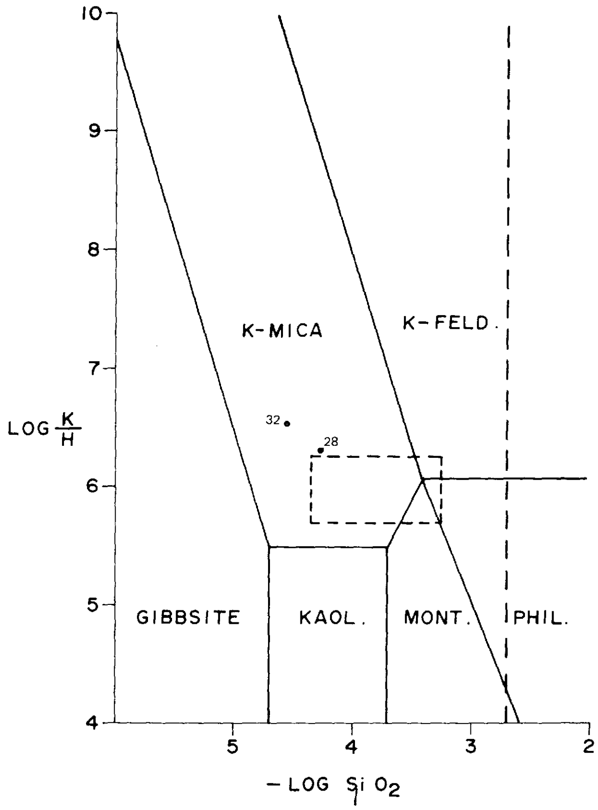

Figure 9. Isoplethic section for interstitial brine at log Na/H = 7.5. Dotted line equals solubility of amorphous silica at 25° C. Dotted box corresponds to oceanic environment (after Hess, 1966).

CONCLUSIONS

The mean composition of clays examined from Great Salt Lake is 51 percent K-mica, 39 percent montmorillonite and interstratified illite-montmorillonite, and 10 percent kaolinite. No chlorite or attapulgite-sepiolite minerals could be identified. There is no systematic change in the kind or percent of clay minerals with depth or laterally in the south arm of Great Salt Lake.

X-ray analyses of sediment samples collected from the suspended load of three major tributary rivers to Great Salt Lake showed K-mica to be slightly more abundant and kaolinite and montmorillonite less abundant in tributary rivers than in the lake sediment.

Great Salt Lake clays are not in equilibrium with lake brine. K-mica should be forming at the expense of kaolinite and montmorillonite.

The interstitial brine in lake clays is less concentrated in sodium, potassium, magnesium, chlorine and sulfate ions and more concentrated in calcium ion that the lake brines overlying the sediment. Sodium, potassium and magnesium may be less concentrated in the interstitial brines because they are diagenetically reacting with clay minerals through cation exchange.

Table 8. Porosities and specific gravities of core samples from Great Salt Lake.

Core No.	Sample depth (cm)	Porosity = $\frac{\text{Weight loss}}{\text{Weight loss}}$ x 100	Porosity = $\frac{\text{Weight loss}}{\text{Oven-dried weight}}$ x 100	Specific gravities of core muds
23	275	56.6	79.16	2.54
41	107	55.2	75.83	2.50
39	122	51.83	68.02	2.51
36	183	59.5	86.81	1.92
44	60	51.5	67.38	2.41

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