University of Nevada
Reno

Geochemical and Isotopic Evidence for Hydrologic Processes at Owens Lake, California.

A thesis submitted in partial fulfillment of the requirements for the Degree of Master of Science in Hydrology/Hydrogeology

by

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Abstract

Dust mitigation at Owens Dry Lake requires an understanding of the hydrologic system because one proposed remedial strategy is flooding dust-producing areas with groundwater pumped from deep aquifers. This study uses geochemical evidence to define hydrologic processes. Solutes in shallow groundwater (<10 m) are controlled by evaporation, mineral dissolution, convective mixing, changes in salt morphology, diffusion, and partial lake desiccation. Deep groundwater in east and west portions of the basin is relatively old (>20,000 $^{14}$C yr) compared to groundwater recharged from the north (~7,760 $^{14}$C yr). Modern (<100 $^{14}$C yr) recharge discharges in margin springs. Groundwater from different depths mixes in a fault zone in the northeast portion of the lake bed. Geochemical evidence of downward groundwater flow contradicts current conditions of upward hydraulic gradients. Groundwater flow out of the basin required at least 12 m of surface water.
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CHAPTER I. INTRODUCTION

Owens Dry Lake is located in the southwestern portion of the Great Basin, at the south end of the Owens Valley, approximately 80 kilometers west of the California–Nevada border (Figure 1). Owens Lake desiccated in the early twentieth century as a result of the diversion of the Owens River into the Los Angeles Aqueduct. Surface characteristics vary across the dry lake bed (280 km²). A brine pool occupies the central western portion and provides the saline water necessary for the formation of evaporite deposits several meters in thickness. In the eastern and southern portions of the lake bed, a thinner surface salt crust forms, replenished by evaporation of shallow groundwater, and varies in stability with temperature and humidity. Efflorescent salts, eroded by wind and sand, mainly originate from areas where the salt crust is thinner and less stable. Blowing dust contains PM–10 particles (dust particles are less than 10 microns in diameter), which have been determined hazardous to human respiratory systems (Saint-Amand et al., 1986). The goal of dust mitigation has prompted much hydrologic research (Lopes, 1988, Kranz, 1994), including this study, because one of the suggested remedial measures is flooding the dust–producing areas of the lake with groundwater pumped from deep aquifers.

Purpose and goals of this study

The purpose of this study is to define the physical processes and lithologic characteristics that control the shallow (defined as less than 10 m below lake bed surface) and deep (defined as greater than 30 m below lake bed surface) groundwater chemistry at Owens Dry Lake. Specifically, the study evaluates the effects of evaporation, groundwater mixing, salt chemistry, geologic structure, stratigraphy, and paleohydrology on groundwater chemistry.

A corollary purpose of this research is to apply the conclusions based on geochemistry to a description of some aspects of groundwater movement at Owens Dry Lake. Such a
Figure 1: Study area and surroundings
description will aid the development of a conceptual groundwater model for the lake basin, as well as increase understanding of some aspects of the Owens Dry Lake water budget. A conceptual groundwater model and quantification of components of the water budget are necessary precursors to numerical analysis of the groundwater system.

Future research that could build on this study includes the prediction of geochemical outcomes of dissolving salt crust components in groundwater pumped from deep aquifers, as well as the geochemical effects of long-term pumping on deep aquifers. These topics are relevant because one suggested mitigation strategy is the flooding of dust-producing areas of the lake with groundwater pumped from deep aquifers.

CHAPTER II. STUDY AREA DESCRIPTION

Surrounding geology

The Owens Valley is a graben, continuously dropping relative to the mountain ranges to the east and west. The Owens Valley most recently became tectonically active in late Tertiary time (<25 my) (Carver, 1969; Pakiser et al., 1964) resulting in the current 3,000 meter relief between the valley floor and the surrounding mountain ranges. Carver (1969) depicts Quaternary faulting along the Sierra, Inyo, and Coso Ranges. Pakiser et al. (1964) delineate the major fault zones that define the Owens Dry Lake basin (Plate 1, sheet 3 of Pakiser et al., 1964). An eastern fault trends N30W, which intersects a south–trending fault 4 km south of Keeler. A series of faults on the western side of the basin parallel the Sierra Nevada Mountains and the Alabama Hills. These features are located in Figure 2.

Owens Dry Lake is at an elevation of approximately 1,097 m above sea level. The Sierra Nevada Mountains rise to over 4,270 m above sea level to the west of the Owens Valley. To the east the Inyo Mountains reach elevations ≤ 3,360 m above sea level. At the southern end of the Owens Valley is the Coso Range, rising to more modest elevations of approximately 2,440 m above sea level.
The Sierra Nevada Mountains in the area of Owens Dry Lake are composed predominantly of intrusive rocks such as quartz monzonites and granites with minor amounts of diorites and gabbros (Knopf, 1918). The Inyo Mountains consist of 11,000 m of sedimentary rocks including pre-Cambrian dolomites and sandstones, Paleozoic and Triassic sandstones, shales and limestones (Knopf, 1918), as well as metavolcanics (Bateman and Merriam, 1954). In the southern part of the study area, lithologies consist of the Coso volcanics (dacite, basalt, rhyolite), intrusive rocks such as quartz monzonites, and arkosic deposits (Bacon et al., 1982).

Alluvial fans originating in the Inyo Mountains, on the east side of the lake bed, are slightly steeper and less broad than alluvial fans emanating from the Sierra Nevada Mountains.
Multiple periods of alluvial deposition are evidenced by older alluvial fans which are still exposed in some canyons of the Inyo Mountains; older alluvial fans on the west side have been partially buried by younger alluvial and glacial outwash (Carver, 1969).

The valley fill sediments in the Owens Dry Lake Basin are at least 1,830 m thick (Pakiser et al., 1964) in the southern portion of the lake basin. These sediments consist of massive to laminar clays to depths of 225 m and interbedded sands and clays to depths of 280 m (core DH–1, Smith and Pratt, 1957. Located in Figure 3). Diatoms and ostracods vary from sparse to abundant throughout the length of the DH–1 core. A 2.5 cm layer of pumice is present at 128 m, and 5 cm of limestone are present at a depth of 125 m.

Similar layers are found in the OL–92 core, completed at 322 m below lake bed surface in the south–central portion of the lake (Smith and Bischoff, 1993), but the depths do not consistently correlate with the DH–1 core. In the OL–92 core, an 11 cm layer of impure limestone is found at approximately 74 m (compared to 125 m in DH–1). In OL–92, a 2 m thick tephra layer exists at 50 m and a 1 cm tephra layer is found at 130 m (compared to a 2.5 cm layer of pumice at 128 m in DH–1). Other thin (less than 10 cm) tephra layers are present between 220 and 320 m. Five meters of tephra are encountered at 303 m and are the fallout and reworked ash of the Bishop Tuff (0.73 Ma, Wilson and Hildreth, 1991). The Bishop Tuff originated from the Long Valley caldera, which is directly north of the Owens Valley. Sand becomes more abundant in the predominantly silty–clay OL–92 core below a depth of 225 m, which correlates well with the similar transition in the DH–1 core. The upper 5 m of OL–92 consists of bedded salts, oolites, and oolitic sand.

Subsurface geology of the southeastern margin of the lake basin has been documented to a depth of 1,070 m (Schaer, 1981). Lithologies of two borings (Figure 3) at the southern end of the lake near highway 190 (OV–D, OV–E) are similar: less than 10 m of fanglomerate (gravels) underlain by 60 to 300 m of clay lake sediments, underlain by fine– to
Figure 3: Sample locations. Lopes (1988) samples indicated by "L." OV and DH locations mentioned in surrounding geology section.
coarse-grained arkosic sand and gravel to the total depth of the borings. A boring drilled to 660 m, approximately 1 km southwest of the intersection of highways 190 and 136, penetrated 24 m of gravels, underlain by 140 m of sand with gravels, underlain by 380 m of clay. Below this depth (~540 m), four basalt flows are interlayered with clays and 91 m of siltstone.

**Climate**

The Owens Valley, and Owens Dry Lake in particular, are characterized by an arid climate. Annual precipitation at Lone Pine, located 10 km directly north of Owens Dry Lake, is generally less than 15 cm. Owens Dry Lake lies in the rain shadow of the Sierra Nevada Mountains, which intercept most moisture-laden clouds, and near its crest receives an average annual precipitation greater than 60 cm (Lopes, 1988). Most of this precipitation falls as rain or snow between October and April (Hollet et al., 1991). The three types of precipitation that affect Owens Valley are winter/spring storms that originate in the Pacific Ocean, scattered summer thunderstorms that originate in the Gulf of Mexico, and summer thunderstorms that originate in the Sierra Nevada Mountains (Houghton, 1969). Summer temperatures often exceed 37°C, with an average summer temperature of 27°C. Average winter temperatures are slightly above freezing (~3°C), although temperatures as low as -18°C have been recorded (Hollet et al., 1991). Vegetation is characteristic of the arid climate, consisting of brush-type phreatophytes such as sage and greasewood, and occasional cacti. Sierra Nevada washes and stream channels on the alluvial fans and the western lake margin are populated by willows and cottonwoods (Carver, 1969).

**Physiography**

The physiographic characteristics of the dry lake environment have been described by Lopes (1988, p. 45–50). Areas covered by a salt crust are separated by sandy areas and occasional mudflats. Minor flow from the Owens River provides increased moisture relative to other
areas of the lake at the river's former delta. A brine pool located in the west central portion of the lake varies in size seasonally. Springs outcrop along the margins of the lake where permeable alluvial fan materials contact less permeable lake sediment. Spring mounds are concentrated in three areas of the lake, located at least 0.4 kilometers from the lake margin: the southernmost portion, the central-eastern portion, and the northeastern portion of the lake.

Current hydrology and water balance

Owens Dry Lake is a discharging, rather than a recharging, playa. Groundwater recharged at higher elevations is discharged from lake margin springs, through evapotranspiration, evaporation, and from pumped and flowing wells (Lopes, 1988).

The main components of hydrologic input into the Owens Dry Lake basin are precipitation and groundwater inflow. The Owens River and mountain stream runoff are captured by the Los Angeles Aqueduct and diverted from the basin. Of the surrounding mountain ranges, the Sierra Nevada Mountains receive the majority of precipitation and, hence, produce the most runoff and subsurface recharge. The amount of subsurface recharge also influences the degree of mixing that occurs between lake brines and recharge groundwater from the alluvial fans (Lopes, 1988).

Groundwater flow from the southwest side of the basin may be impeded by faulting (Perrine, 1983), but Lopes (1988) suggests that faults northeast of Keeler would have little effect on groundwater flow because there are no thick impermeable sediments to juxtapose against alluvial fan sediments in this location. Lake clays are capable of storing and leaking water (Lopes, 1988).

Lopes (1988) infers that a thick sandy aquifer deposited by the Owens River exists in the center of the basin from the upper Owens Valley to Haiwee Reservoir, 8 km south of Owens
Dry Lake. Other continuous aquifers are alluvial fan deposits which follow the lake margins and may be up to 900 m thick.

The Sulfate Well in the east-central portion of the lake (Figure 3) appears to be completed in an alluvial aquifer, and to be recharged from the Inyo Mountains (Lopes, 1988). Based on sedimentation rates determined for Searles lake (Smith et al., 1983), Lopes disproves continuity between the Sulfate Well aquifer and the sandy layers found in the DH-1 core in the central portion of the lake (Smith and Pratt, 1957).

Lopes (1988) developed a water budget for the area bounded by the Sierra Nevada crest, the Inyo and Coso Mountain crests, the northern border of Owens Dry Lake, and the northern border of Haiwee Reservoir. He used hydrologic data collected between 1940 and 1980 in his calculations, and a summary of Lopes’ results is presented in Table 1.

| Table 1: Water budget from Lopes (1988). Results have been converted to cubic meters. |
|---------------------------------|-----------------|-----------------|-----------------|
| **Inflow**                      |                 |                 |                 |
| component                       | maximum         | minimum         | mean            |
| precipitation                  | $9.64 \times 10^7$ | $8.95 \times 10^7$ | $9.30 \times 10^7$ |
| runoff                          | $8.67 \times 10^7$ | $7.18 \times 10^7$ | $7.93 \times 10^7$ |
| subsurface flow                | $3.14 \times 10^7$ | $1.53 \times 10^7$ | $2.32 \times 10^7$ |
| Owens River                    |                 |                 | $1.32 \times 10^7$ |
| Total (rounded)                | $2.28 \times 10^8$ | $1.90 \times 10^8$ | $2.08 \times 10^8$ |
| **Outflow**                    |                 |                 |                 |
| component                       | maximum         | minimum         | mean            |
| evapotranspiration             | $6.34 \times 10^7$ | $5.23 \times 10^7$ | $5.80 \times 10^7$ |
| aqueduct diversions            | $3.16 \times 10^7$ | $2.91 \times 10^7$ | $3.03 \times 10^7$ |
| lake bed evaporation           | $4.77 \times 10^7$ | $4.29 \times 10^7$ | $4.54 \times 10^7$ |
| lake evaporation               | $1.09 \times 10^8$ | $8.18 \times 10^7$ | $9.55 \times 10^7$ |
| Total (rounded)                | $2.50 \times 10^8$ | $2.06 \times 10^8$ | $2.29 \times 10^8$ |
Paleohydrology of Owens Dry Lake

Closed basin lakes such as Owens Dry Lake have been looked to as records of paleoclimate variations. Street-Perrott and Harrison (1985) noted that the lakes that are the most sensitive to climatic variation and provide the most useful records of that variation are amplifier lakes. These are lakes in which nearly 100 percent of the input to the lake comes from surface runoff, but are not characterized by a high rate of flushing. Closed basin amplifier lakes are the most useful because of the continuous and complete sedimentary record. Owens Dry Lake has not always been a terminal lake, but it has been for at least the past 8,900 years (Friedman et al., 1993).

Gale (1915) was likely the first to describe the prehistoric connection between Owens Lake and other closed-basin lakes (China, Searles, Panamint) to the southeast toward Death Valley (Figure 4). Since late Pliocene time, changes in temperature and moisture have prompted these lakes to respond by overflowing and carrying dissolved constituents into successively lower elevation lakes, to maintain perennial water bodies below their respective outflows, or to desiccate completely and accumulate deposits of evaporite minerals and coarse-grained sediments. Owens Lake was the first in the series in the Pleistocene lake system (except rare occurrences when Mono Lake overflowed into a smaller lake which connected to the Owens river). Gale estimated the Owens lake outflow at 1,146 m above sea level, and Lee (1912) noted that the uppermost water line was at 1,155 m. Owens lake was 65 m deep (Smith and Street-Perrott, 1983) when it most recently overflowed (~8,900 years ago). Significant flow from Owens Lake into Rose Valley is indicated by eroded channels near Little Lake; one channel circumvents a volcanic flow that dammed the former channel approximately 0.44 my ago (Duffield and Smith, 1978).

A 930 meter core was drilled at Searles Lake, from surface to bedrock, which provides an enormous amount of climatic information about the past 3.2 million years (Smith et al.,
Figure 4: Pleistocene lake chains of the southeast Great Basin. From Jannik et al., 1991; Flint and Gale, 1958.
The upper 693 m of the core consists of lacustrine sediments and evaporites, underlain to total depth by alluvial sand and gravel deposits. Smith et al. (1983) and Smith (1984, Table 1) have developed a chronology divided into ten regimes (i.e., climatic periods) based on changes in the deduced temperatures and humidities that prevailed. These periods of wetting and drying can be consulted when interpreting the sedimentary and geochemical record of Owens Lake. The first regime, Upper Salt, comprises the last 10,000 years and has been dry relative to the other time periods represented in the Searles core. In the last 3.2 million years, the only other dry periods extended from 0.57 to 0.31 my and 2.56 to 2.04 my. All other periods have been intermediate or wet. Wet periods have occurred between 1.28 and 1.00 my and 3.18 and 2.56 my.

The natural shrinking of Owens Lake that began ~10,000 years ago was exacerbated in 1872 when irrigation began to significantly deplete the Owens river flow into the lake. The lake desiccated completely in 1913 when the city of Los Angeles completed an aqueduct that diverted the entire flow of the Owens River. The lake has remained desiccated from that time except for rare periods, such as in 1969, when the river flow has exceeded the capacity of the aqueduct.

Findings from previous studies

Hydrologic studies

The hydrology of the Owens Valley and Owens Lake were first investigated in the early part of this century. Lee (1906) and Lee (1912) performed the most extensive explorations of the area, documenting the water resources of the Owens Valley prior to diversion of the Owens River into the Los Angeles aqueduct.

Danskin (1988) used a finite difference groundwater model (McDonald and Harbaugh, 1984) to evaluate the hydrogeologic system of the Owens Valley. The modeled area included the Owens Valley as far south as Lone Pine, 10 km north of Owens Dry Lake. No hydraulic
head measurements from Owens Dry Lake were available, and therefore, no calibration was done in the Owens Dry Lake area. The modeling effort attempted to account for complexities in the system such as a high degree of faulting, the presence of highly transmissive volcanic deposits, groundwater inflow from valleys to the north, and groundwater–surface water interactions. In spite of these accommodations, certain complexities (such as transition zones of reworked sediments at the toes of alluvial fans), and data limitations (such as a lack of any hydraulic head data on alluvial fans) were recognized. Danskin’s (1988) model was a steady state simulation in which groundwater flowed in two distinct layers in the middle of the valley, and in one layer on the alluvial fans. Six categories of hydrogeologic materials were defined, and Owens Lake was treated as lacustrine sediments surrounded by alluvial deposits, with deltaic sediments at the northern extreme of the current lake bed. The calibrated transmissivity values for a range of deposits are listed in Table 2. These values were judged inaccurate near the toes of alluvial fans and in the vicinity of volcanic deposits. In these areas, the difference between measured and modeled hydraulic head values was greater than 4.57 m.

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<th>Transmissivity lower layer</th>
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<td>alluvial fan deposits (mtn. front)</td>
<td>50.2</td>
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<tr>
<td>alluvial fan deposits with faults</td>
<td>25.1</td>
<td>62.2</td>
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<tr>
<td>stream channel, deltaic</td>
<td>195</td>
<td>1,950</td>
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<tr>
<td>lake bed deposits*</td>
<td>121</td>
<td>1,210</td>
</tr>
<tr>
<td>permeable volcanic deposits</td>
<td>306</td>
<td>3,066</td>
</tr>
</tbody>
</table>

* lake bed deposits are defined as covering the largest, most recent extent of Owens Lake.

Hollett et al. (1991) summarized much of the previous work and combined it with more recent findings such as Danskin (1988) in an updated survey of the water resources of Owens Valley. Hollett et al. (1991) is a comprehensive study which includes data on water balance and an extensive treatment of the aquifers that comprise the hydrogeologic system, including
aquifer test results from 20 wells installed as part of their research. Several results from Hollett et al. (1991) are used later in this study.

Great Basin Unified Air Pollution Control District (Great Basin UAPCD), Bishop, CA, has completed three new wells around Owens Dry Lake, which the Desert Research Institute (DRI) designed and documented: the River Site Wells (Mihevc and Lyles, 1989), Mill Site Well (Mihevc and Lyles, 1990), and Keeler-Swansea Well (Lyles and Mihevc, 1989), each located in Figure 3. A brief description of each of those wells and available aquifer test data is presented below.

**River Site Wells**: The River site wells, upper and lower, are completed in two sandy aquifers separated by 62 m of clay. The upper well is screened between 47 and 69 m below ground surface. The deeper well is completed between 148 and 169 m below ground surface. There is also a multiple completion monitoring well located approximately 90 m west of the production wells. Analysis of data from a three day aquifer test at the River site (Jacobson et al., 1990) indicated a transmissivity of 720 m²/day in the lower River site aquifer. The average transmissivity (T) of the upper aquifer is 531 m²/day (Jacobson, et al., 1992) based on a three month aquifer test. The potentiometric head in the upper aquifer is 15 m above ground surface (measured Jun. 23, 1993). The potentiometric head in the lower aquifer is 16.3 m above ground surface (measured Jun. 23, 1993).

**Mill Site Well**: Great Basin UAPCD also completed the Mill Site Well on the eastern margin of the lake (Mihevc and Lyles, 1990). The well is completed in a mixed sand and gravel aquifer, and the unconfined water table is approximately 2.7 m below ground surface. The well is screened in three intervals: between 33.5 and 41 m, 52 and 58 m, and 64 and 78 m below ground surface. A nested monitoring well is completed in the shallowest and deepest screened zones approximately 75 m north of the production well. The transmissivity of the
Mill site aquifer is approximately 930 m$^2$/day, based on a three month aquifer test (Jacobson et al., 1992).

**Keeler-Swansea Piezometer:** Three piezometers are nested in one boring at the Keeler-Swansea site (Lyles and Mihevc, 1989). They are referred to later in the text as upper, intermediate, and lower. The two shallowest piezometers are screened between 30 and 36 meters and 49 and 55 meters below ground surface. The deepest piezometer is screened in three separate zones between 67 – 98 meters below ground surface. The two shallowest piezometers are completed in mixed sand, gravel, and clay. Aquifer material below approximately 61 meters of depth consists of boulders, cobbles, gravels, and sands. The three piezometers all have a potentiometric surface above ground surface. The hydraulic heads are approximately 3.16 meters, 3.29 meters, and 3.54 meters above land surface in the upper, intermediate, and lower piezometers, respectively. The chemistry and isotopic composition of the groundwater in each of the three piezometers is virtually identical, and usually the chemical analysis from the intermediate piezometer is used in this study.

**OL-92:** The lithology of the core that was collected during the drilling of this piezometer is described in the Surrounding Geology section. The core was collected and analyzed by the U.S. Geological Survey (Smith and Bischoff, 1993). The piezometer, which was designed by DRI and installed under DRI supervision, is screened between 227 and 236 m below ground surface, and the hydraulic head is approximately 15 m above ground surface.

**Cottonwood Wells:** Several wells have been completed in the west-central portion of the lake near the Cottonwood Springs area (CH2M Hill, 1991; Luhдорff and Scalmanini, 1993a, 1993b). A production well (CW-PW1) and a nested monitoring well (CW-MW2S, CW-MW2D) are described below. The production well is screened in an alluvial aquifer between 61 and 131 m below ground surface (ground surface at this location is
approximately 1,113 m above sea level). Depth to water in the production well is 23 m below ground surface. Monitoring well CW-MW2 is located approximately 24 m lower in topographic elevation and 250 m to the east of CW-PW1. The shallow monitoring well (CW-MW2S) is screened between 18 and 21 m below ground surface, and the deeper monitoring well (CW-MW2D) between 39 and 91 m below ground surface. Depth to water in the shallow and deep monitoring wells is 1.85 m and 0.78 m, respectively.

**Sulfate Well:** The Sulfate Well is not a recently installed well, and the estimated specifications of the well are described in Lopes (1988, p. 41). Recently, however, a six hour shut-in test was performed at the Sulfate Well (Lyles et al., 1994) and the approximately transmissivity of the penetrated aquifer was determined to be 60 m²/day.

**Isotopic studies**

Research on the isotopes of the deep groundwater in the area of Owens Dry Lake has yielded a great deal of paleoclimatic information relevant to Owens Dry Lake. A study of the isotopic composition of groundwater in southeast California and how it relates to modern precipitation was completed by Smith *et al.* (1992) and Friedman *et al.* (1992). Deuterium values in modern precipitation (both rain and snow) in several hydrologic basins including and south of Owens Dry Lake averaged -56‰ during the summer and -78‰ during the winter. The weighted average δD values in precipitation samples collected at Lone Pine were -62‰ (summer) and -96‰ (winter). Friedman and Smith (1970) obtained snow cores from several locations in the Sierra Nevada Mountains and reported their δD values. They noted a consistent relationship between δD content and elevation on the west side of the Sierra (a decrease in 40‰ for each 1,000 m elevation gain), but only a vague correlation between these parameters on the eastern side of the Sierra. Snow samples from the southeastern Sierra (from Bishop Pass to Whitney Portal Road) ranged in δD content from
-159‰ to -128‰. The average δD content of the Owens River, -120‰, (Friedman et al., 1976) indicates contributions from both snowmelt and rain.

Smith et al. (1992) analyzed 82 groundwater samples in conjunction with the precipitation study described above. They noted that approximately half of the groundwater samples were depleted in δD by more than 10‰ relative to average modern winter precipitation. These samples were interpreted to be of Pleistocene origin, having condensed during a cooler, wetter climate, 4°C to 8°C cooler than today (Smith et al., 1979).

**Geochemical studies**

One of the first investigations of Owens Lake to include geochemical data was compiled by Gale (1915) in which he described the formation and distribution of saline deposits on the lake bed, inferred the Pleistocene lake chain including Owens, China, Searles, Panamint, and Manly Lakes, and estimated the elevation of Owens Lake outflow.

An opportunity for study of the surface and shallow subsurface geochemical processes at Owens Dry Lake was presented in 1969 when the Owens River flow surpassed the capacity of the Los Angeles Aqueduct and the lake bed was flooded to a depth of 2.4 m. During the summer of that year, the lake began to desiccate and was dry again by the summer of 1971. Friedman et al. (1976) found that concentrations of all the major ions except Ca and Mg increased in lake water as the lake continued to desiccate. Deviations from increasing trends were attributed to precipitation of minerals containing Ca, Na, CO₃, SO₄ and Cl. Lithium behaved as a conservative substance and was used to estimate the fraction of water remaining in the lake as evaporation progressed. Comparison of the increasing D/H ratio in the lake water to the decreasing volume of lake water indicated a Rayleigh-type evaporation, in which the water vapor leaving the lake was assumed to be in isotopic equilibrium with the lake water. The isotopic fractionation factor determined from the field data began to decrease at the highest salinity concentrations attained in the evaporating lake water.
Smith et al. (1987) found that temperature fluctuations (diurnal and seasonal) were as important as the evolving brine composition in determining the salts that crystallized. During the solution stage (from February 1970 to August 1970), $0.9 \times 10^6$ tons of carbon dioxide were removed from the lake water as the following reaction proceeded:

$$2\text{HCO}_3^- + \text{CO}_3 + \text{CO}_2 + \text{H}_2\text{O}.$$  

The reaction was driven by the partial pressure of CO$_2$ in the lake water, which exceeded that of the atmosphere. Carbon dioxide was expelled to the atmosphere or consumed by algae. From September 1970 to August 1971, the groups of primary salts to crystallize ranged from natron–trona to natron–mirabilite–trona and finally trona–burkeite–halite. However, these primary salt assemblages varied with diurnal temperature changes as well as with depth. Smith et al. (1987) concluded that paleoclimatic data cannot be extracted from salt mineral assemblages because the observed mineral assemblage is not necessarily representative of the primary salts to crystallize.

Smith and Friedman (1986) as well as Matsuo et al. (1972) observed the fractionation of $\delta D$ between hydrated mineral phases and the brines from which they precipitated. Nahcolite and trona become depleted in $\delta D$ and mirabilite becomes enriched in $\delta D$ relative to their parent brines. Observations made following the lake’s desiccation (1971–1977) indicated a cyclicity in salt formation, with more hydrated phases (natron, mirabilite) forming in the winter months, and dehydrated phases (trona, burkeite) forming in the summer months. The $\delta D$ values in the precipitated salts and in the brines stabilized (i.e. seasonal variation decreased) after numerous seasonal cycles.

Saint-Amand et al. (1986) produced a comprehensive study of the dust production from the lake bed and the atmospheric and mineralogic conditions of the crust that increases the severity of the dust storms. Saint-Amand et al. (1986) also report variation in salt structure and chemistry with seasonal changes. The salt crust is stable as long as the surface remains
dry for several weeks, but becomes friable if wetted. When the salt crust becomes wet and heavier (during the winter months), there is better contact between the underlying clays and the crust. This increased contact enhances the capillary action that supplies concentrated groundwater (brine) to the crust, and the dissolution of some of the salt into the uppermost centimeters of the clays leads to osmotic transport of less saline waters from depth to the surface. Crust formation in the wetter months, therefore, is more rapid than in the drier months. Saint-Amand et al. (1986) reports the average ionic percentages by weight from his samples of crust in the southern and central portions of the lake bed and from samples of lake water (Table 3), although he doesn't specify if the samples he has termed lake water are from the brine pool or from shallow groundwater.

Table 3: Average ionic percentage by weight, from Saint-Amand et al. (1986).

<table>
<thead>
<tr>
<th>sample</th>
<th>Na</th>
<th>K</th>
<th>Cl</th>
<th>SO₄</th>
<th>CO₃</th>
<th>HCO₃</th>
<th>SiO₂</th>
</tr>
</thead>
<tbody>
<tr>
<td>crust</td>
<td>37.64</td>
<td>1.90</td>
<td>21.69</td>
<td>23.21</td>
<td>14.58</td>
<td>0.98</td>
<td></td>
</tr>
<tr>
<td>lake water</td>
<td>38.78</td>
<td>2.02</td>
<td>26.45</td>
<td>9.45</td>
<td>18.44</td>
<td>4.91</td>
<td>0.24</td>
</tr>
</tbody>
</table>

* CO₃ and HCO₃ in crust samples reported together as CO₃

Saint-Amand et al. (1986) discuss the major crustal constituents but do not address the discrepancy between ionic concentrations in the lake water versus the crust. The analyses of salt crust indicate that the most important crust-forming minerals are halite, natron, thermonatrite, thenardite, mirabilite, and trona. They report that nahcolite will not normally form due to insufficient carbon dioxide or temperature. Burkeite was not identified in their crustal samples.

Lopes (1988) used geochemical and isotopic signatures of groundwater to determine source areas, occurrence of groundwater mixing, and flow paths. Lopes' sampling locations are presented in Figure 3.
Based on the geochemical and isotopic data obtained from 31 wells and springs, Lopes differentiated five geochemical signatures in Owens Dry Lake groundwater:

1. west: $\text{Na+K-Ca-HCO}_3$
2. northeast: $\text{Na+K-Mg-HCO}_3$
3. southeast: $\text{Na+K-Mg-HCO}_3-\text{Cl}$
4. playas brines: $\text{Na+K-HCO}_3-\text{Cl}$
5. mixing of 1–3 with 4

Lopes (1988) concluded that the overall characteristics of the groundwater chemistry are controlled predominantly by the mineralogy of the source rocks and carbonate mineral precipitation, with lesser influence from groundwater mixing and sulfur redox reactions.

The findings of many of these previous studies are referenced in the following chapters. Lopes' (1988) conclusions raise several questions that will be addressed in this study. Why do some locations on the west margin of the lake, adjacent to the Sierra Nevada Mountains, show more influence from mixing with saline lake water than locations on the east side of the lake, where recharge from the Inyo range is significantly less than that from the Sierra? Could there be a connection between groundwater the northern part of the lake and that in the northeastern part of the lake bed? With additional data now available from wells and spring mounds in the northeastern portion of the lake (not sampled by Lopes) is there evidence of fault-controlled groundwater movement? A major factor that Lopes did not discuss in determining the most important controls on groundwater chemistry, is the importance of varying paleohydrologic conditions.
CHAPTER III. METHODS OF ANALYSIS

General hypothesis

The geochemistry of the shallow groundwater system (less than 10 m below lake bed surface) is mainly controlled by evaporation, but mineral dissolution and precipitation, mixing between waters of different salinities, and salt morphology changes with temperature contribute to local variations in groundwater composition. The deep groundwater geochemistry (>30 m) depends on groundwater residence time in aquifers, paleoclimate, groundwater mixing at lake margins, structurally–controlled flow, and vertical groundwater movement.

Methodology

This study begins with a description of the tools used to evaluate the hypothesis, i.e., geochemical analysis and computer models. Justification for the use of those tools and some examples of how they have been applied are presented.

The Owens Dry Lake groundwater system is addressed in two chapters: Chapter IV examines the shallow system (within 10 m of the lake bed surface), and Chapter V is devoted to understanding the deep system (generally >30 m beneath ground surface). Data from the shallow groundwater system were collected from a network of 162 piezometers installed by Great Basin UAPCD. The major ion chemistry from the shallow groundwater is used to discern overall relationships between ions and the relationship between the shallow groundwater and the composition of the surface crust. The distribution of solutes in shallow groundwater is examined and potential explanations for the distribution are discussed. A presentation of isotopic and geochemical variation and possible controlling factors is the final topic of the shallow groundwater chapter.

The fifth chapter of this thesis attempts to identify the controls on the deep groundwater geochemistry. By identifying these controls, conclusions about the hydrology and
paleohydrology of Owens Dry Lake are made. The first section looks at hydraulic head measurements, and the stable and radiogenic isotopic data from deep groundwater sites and assesses various hydrologic scenarios that could generate the observed values. The next section of the deep groundwater chapter builds upon a conclusion from Lopes (1988) concerning fault-controlled groundwater flow. This section provides evidence for this process and shows that mixing of different groundwaters could contribute to observed geochemical and isotopic variations in the deep groundwater. The issue of the next section in Chapter V is vertical groundwater flow. Geochemical and isotopic data are used to infer vertical movement of groundwater between aquifers. The final section of the deep groundwater chapter is a brief discussion of potential groundwater flow out of the Owens Dry Lake basin, as evaluated by other studies (Smith and Bischoff, 1993) and the results of this study.

**Description of methods of analysis (tools)**

The tools employed in order to evaluate the hypotheses are analyses for geochemical constituents that are indicative of physical processes or are useful for distinguishing recharge times or locations. Any one particular analysis is not used alone to formulate conclusions, but rather is used to suggest a controlling process that can be further evaluated with other data. Owens Dry Lake studies to date (Lopes, 1988; Friedman et al., 1976; Smith and Friedman, 1986) have mainly used major ion chemistry, salt chemistry, and stable isotopes to relate groundwater at different locations, and to identify the physical controls (evaporation, temperature) on the chemistry of dissolved and precipitated salts. This study employs those methods but also uses groundwater chemistry, stable isotopes, and radiogenic isotopes to identify important hydrologic processes within the Owens basin.

Geochemical models are also an important interpretive tool used in this study. Models are used to test the feasibility of a range of scenarios, to eliminate certain processes, or to predict
outcomes for a range of initial conditions. Limited geochemical modeling has been done at Owens Dry Lake. Lopes (1988) used the program SALTNORM (Bodine and Jones, 1986) to predict precipitated mineral phases from evaporated brines.

The specific tools used in this study and justification for their use are presented below. Appendix A summarizes the protocol of groundwater sample collection. The appendix also describes the analytical procedures performed on the samples, laboratory preparations, and instrumentation used in each analysis.

Oxygen and hydrogen stable isotopes

The stable isotopes of oxygen and hydrogen, $\delta^{18}O$ and $\delta D$, are extremely useful in evaluating the effects of evaporation because of the differential fractionation of deuterium (D) and oxygen-18 ($^{18}O$), described below. Friedman et al. (1976), Smith et al. (1987) follow the progression of evaporation of the brine pool at Owens Dry Lake by comparing the evolution of D to the dissolved major ion chemistry and chemistry of precipitating salts. Smith et al. (1992) and Friedman et al. (1992) use stable isotopes to distinguish groundwater recharged during the Pleistocene from groundwater recharged by modern precipitation. Deuterium and $^{18}O$ concentrations are expressed as delta units, relative to standard mean ocean water (SMOW), as first suggested by Craig (1961a):

$$
\delta D = \frac{D_{samp} - D_{smow}}{D_{smow}} \times 1000 \quad \text{and} \quad \delta^{18}O = \frac{^{18}O_{samp} - ^{18}O_{smow}}{^{18}O_{smow}} \times 1000.
$$  \hspace{1cm} (1)

Differences between the vapor pressures of hydrogen and oxygen result in different isotopic fractionation factors. The global meteoric water line (MWL) (Craig 1961b), which has a slope approximately equal to eight, reflects the different fractionation factors of $^{18}O$ and D (Figure 5).
The slope of the meteoric water line expressed mathematically is:

$$\frac{e^*_D}{e^*_{^{18}O}} \equiv 8 \quad \text{(Craig, et al., 1963).}$$

where $e^*$ is the fractionation factor ($R_{\text{vapor}}/R_{\text{liquid}} - 1$) and $R$ is the ratio of the heavy isotope to the light isotope (i.e. $D/H$, $^{18}O/^{16}O$).

The deuterium fractionation factor that Craig et al. (1963) use ($e^*_D = -69 \, ^\circ\text{C}$) is from Miles and Menzies (1936). The experimental oxygen fractionation factor determined by Craig et al. (1963) is $e^*_{^{18}O} = -9 \, ^\circ\text{C}$.

The process of precipitation (or condensation of liquid water from atmospheric water) is a Rayleigh process in which the condensate is in isotopic and thermodynamic equilibrium with the water vapor at 100% vapor saturation (Craig 1961b). That process abides by the equation of an equilibrium batch distillation:
\[ \lambda - \lambda^o = e \ln f. \]  

(3)

where \( \lambda = \ln (I + \delta) \) at any time, \( \lambda^o = \ln (I + \delta) \) initially, and \( f \) = the fraction remaining of the original reservoir (Craig et al., 1963)

Craig (1961b, 1963) deduced that unlike precipitation, which is a Rayleigh process, evaporation is controlled by kinetic processes where isotopic equilibrium does not exist between the vapor phase and the liquid phase. Fritz and Fontes (1980, p. 37) note the differential kinetic effects that result in evaporative slopes less than eight. They introduce a term, \( \Delta \epsilon \), the kinetic excess separation factor, such that:

\[ I - I/\alpha_e + \Delta \epsilon = e. \]  

(4)

where \( \alpha_e = R_{\text{liquid}}/R_{\text{vapor}} \) and \( e \) is the effective separation factor.

The \( \Delta \epsilon \) term, which is dependent upon humidity, is approximated for D and \( ^{18} \text{O} \) by Merlivat, 1978, as \( 15 \cdot (1-h)\% \) and \( 13 \cdot (1-h)\% \), respectively, where \( h \) = relative humidity (as a fraction). The effect of this term at decreasing humidities is much greater when added to the \( ^{18} \text{O} \) separation factor (0.009) versus the D separation factor (0.069). Consequently, the slope (\( \Delta \delta_D/\Delta \delta^{18} \text{O} \)) decreases due to the differential increases in enrichment. Craig (1961b) found that evaporated waters that had originated on the MWL tended to consistently plot on slopes around five. Experiments conducted by Craig et al. (1963) also confirmed evaporative slopes of 5.5 ±1.)
Isotopic fractionation factors also vary with the salinity of the water, as increased salinity reduces the vapor pressure of hydrogen and oxygen. A very general representation of this concept as it applies to the amount of $\delta D$ in the liquid phase is from Gat (1980, p. 15):

\[(1 - 1/\gamma) \cdot 10^3 = 6.1M \text{CaCl}_2 + 5.1M \text{MgCl}_2 + 2.4M \text{KCl} + 0.4M \text{NaCl}. \quad (5)\]

where $\gamma$ is the activity of HD$^{16}O$ and $M$ is the molality of the solution.

Not only are the effects of evaporation reflected in the $\delta^{18}O$ and $\delta D$ isotopic signatures, but comparison of the isotopic values between samples may also reveal groundwater recharged under different climatic conditions, or the mixture of groundwaters of different origins. Groundwater mixing is most discernable with $\delta^{18}O$ and $\delta D$ when the proportions of each of the input waters approaches 50 percent. Because the heavier isotopes are relatively rare (one D atom in ~5,000 H atoms), small amounts of groundwater mixed into a larger supply will not alter the isotopic value of the original groundwater to an extent detectable by the analysis, unless the end members are significantly different. This study also employs models of $\delta^{18}O$ and $\delta D$ evolution to identify processes that have determined the isotopic signatures observed in deep and shallow groundwater.

**Strontium isotopes**

Another useful analysis used in this study is the ratio of Sr isotopes ($^{87}\text{Sr}/^{86}\text{Sr}$) in the groundwater. This analysis is accurate to at least four significant figures and variation among groundwater samples is immediately evident. Although no Sr isotope ratio data have been derived from Owens Dry Lake groundwater prior to this study, Neumann (1993) successfully used Sr isotope ratios to differentiate the various source areas that supply Mono Lake, California. The Sr ratio in rocks depends on the original abundance of $^{87}\text{Rb}$, which radioactively decays to $^{87}\text{Sr}$. Rubidium is chemically similar to K and will therefore be enriched in minerals containing significant K (micas, K feldspars, and K-rich clays). The Sr ratios in groundwater, however, will not necessarily be reflective of the whole rock Sr ratios.
in their recharge areas. As Neumann (1993) and Blum et al. (1994) have found for some eastern Sierra Nevada streams, differential weathering of minerals results in Sr ratios in surface water elevated above the whole rock ratios in the bedrock. Blum et al. (1994) suggest that the more rapid weathering of biotite from granodiorite is responsible for the elevated Sr ratios in the surface water, biotite having a relatively higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratio than other minerals in the rock. Because of the high precision of the Sr analysis, close agreement of the isotopic ratios from different groundwater samples in this study is often evidence that the groundwaters have followed similar geochemical evolutionary paths from the same recharge area. Mixture of groundwaters from different recharge areas may also be detected by comparison of Sr ratios.

Carbon isotopes

Prior to this study, no one has analyzed Owens Dry Lake groundwater for radiogenic carbon. However, Lyons et al. (in press) have collected two groundwater samples from approximately 20 km north of Owens Dry Lake and determined their $^{14}\text{C}$ content. Lyons et al. (in press) compare the $^{14}\text{C}$ data with calculations of travel times based on hydraulic and stratigraphic data. Carbon-14 analyses are most useful when used in conjunction with other geochemical data because the radiogenic carbon content can be altered by rock–water interactions and microbiologic activity. Radiogenic carbon analyses in groundwater are therefore more difficult to interpret than the results from stable isotopic analyses. Carbon-14 is produced in the upper atmosphere, and carbon dioxide containing $^{13}\text{C}$ is incorporated into surface and atmospheric water. If this water becomes isolated from the atmosphere (such as groundwater in a confined aquifer), the decay of $^{14}\text{C}$ proceeds with a half life of 5,730 years (Mazor, 1991, p. 164). The process of $^{14}\text{C}$ decay is represented by:

$$t = \frac{5,730}{\log 2} \frac{\log A_{\text{nd}}}{A}$$  \hspace{1cm} (Wigley et al., 1978)  \hspace{1cm} (6)
where, $t$ is the adjusted age of the water, $A$ is the measured $^{14}$C activity, and $A_{rd}$ is the $^{14}$C activity due to non-decay processes. The non-decay processes that may alter the original $^{14}$C concentration are dissolution and precipitation of carbonate minerals, production of methane, oxidation of organic matter (Wigley et al., 1978), and diffusion of modern carbon in areas of low recharge (Walker and Cook, 1991). Dissolution of carbonate minerals results in a dilution of the original $^{14}$C activity with inactive carbon; oxidation of organic matter supplies CO$_2$, and methane production depletes the CO$_2$ supply. Each of these processes may also be accompanied by an isotopic fractionation that alters the $^{14}$C activity.

The $\delta^{13}$C values of the groundwater and phases that have interacted with the groundwater are often used to modify the measured $^{14}$C age, to correct for the influence of non-radiogenic processes. Carbon-14 and $\delta^{13}$C analyses are reported in units relative to two established standards, oxalic acid and a belemnite carbonate formation in South Carolina (Pee Dee Belemnite), respectively. Ideally, the $\delta^{13}$C value is indicative of the origin of the carbon dioxide, whether from the atmosphere ($\delta^{13}$C $\equiv -7\%$) (Drever, 1988, p. 377), from biologic production in the soil zone ($\delta^{13}$C $\equiv -25\%$ to $-12\%$ depending on the plant communities) (Wigley et al., 1978), or from carbonate minerals ($\delta^{13}$C $\equiv 0\%$) (Mazor, 1991, p. 173). If the $\delta^{13}$C values of the groundwater, aquifer material and soil gas can be measured or estimated, several models are available to calculate the $^{14}$C age based on the processes the groundwater has undergone. Fontes and Garnier (1979) summarize correction methods that deal only with congruent dissolution of carbonate minerals (e.g. Ingerson and Pearson, 1964; Tamers, 1975), and those that address the effects of isotopic exchange as well (Mook, 1972; Fontes and Garnier, 1979).

The percent of modern carbon (oxalic standard is considered 100 percent) in groundwater may also be artificially altered by incorporation of current atmospheric carbon dioxide that has been subjected to increased $^{14}$C content from the detonation of nuclear weapons in the
1950s and 1960s. Consequently, a very young groundwater sample may yield greater than 100 percent modern carbon. Tritium (3H) analysis is examined in conjunction with 14C analysis, as this isotope of hydrogen would only be detectable in unmixed groundwaters less than approximately 50 years of age.

Each of the analyses mentioned, 14C, δ13C, and 3H are used in a limited number of samples in this study. The information available for interpretation includes the measured 14C of the groundwater, the measured δ13C of the groundwater, measured δ13C from lake sediments (Benson and Bischoff, 1993), carbonate saturation indices, pH and laboratory alkalinity, knowledge of surrounding geology, and estimates of soil zone CO2 based on the vegetation. All this information is consulted in assembling a reasonable model of the processes that have affected the carbon isotope chemistry of the groundwater. Information derived from these analyses is useful in this study in assessing relative recharge rates from different areas of the Owens basin, and in interpreting diverse flow paths of shallow and deep groundwater.

Geochemical computer models

Three geochemical computer models have been used for purposes of simplifying calculations, i.e. determining saturation indices, as well as testing the viability of numerous possible hydrologic scenarios. Comparison of saturation indices is useful to distinguish groundwaters at different stages of geochemical evolution, and also to determine what geochemical reactions are thermodynamically possible. Two examples of studies that have used geochemical models for similar purposes as this study are Donovan and Rose (1994) and Thomas et al. (1989). Donovan and Rose (1994) use PHRPITZ (Plummer et al., 1988) to predict the composition of groundwaters in which pCO2 is lowered to atmospheric concentrations and in which evaporation is simulated by addition of pure solutes in the proportions of the assumed source water. The results are then compared to compositions of evaporative, groundwater-fed lakes. Thomas et al. (1989) use BALANCE (Parkhurst et al.,
1982), a precursor to NETPATH (Plummer et al., 1991), described below, to model potential mass balance between recharge groundwater in alluvial fans and evolved brines in a closed basin in central Nevada.

The models used in this study are PHREEQE (Parkhurst et al., 1990), PHRQPITZ (Plummer et al., 1988), and NETPATH (Plummer et al., 1991). PHREEQE and PHRQPITZ are forward models, defined as models that predict an equilibrium outcome from given input of geochemical information. Neither model incorporates kinetic factors into its calculations. PHREEQE relies on extended Debye–Huckel or Davies equations to calculate ion activities, and an ion-pair model to calculate mineral saturation indices and species distribution. The components of the Davies equation are:

\[
\log \gamma_i = \frac{-A z_i^2 \sqrt{I}}{1 + B a_0 I} + b I \quad (\text{Drever, 1988, p. 25})
\]  

(7)

where \(\gamma_i\) is the activity coefficient of ion \(i\), \(A\) and \(B\) are constants dependent on temperature and pressure, \(z_i\) is the charge of ion \(i\), \(I\) is the ionic strength of the solution, \(a_0\) is the hydrated radius of ion \(i\), and \(b\) is an empirical constant for \(I > 0.1\). The use of an ion pair model to perform speciation is appropriate until the ionic strength of the solution exceeds that of seawater (Drever, 1988, p.26). Beyond these concentrations, specific interaction models must be used (eg. Pitzer, 1973). The Pitzer treatment introduces two additional coefficients that add effects of binary and ternary ionic interactions to a Debye–Huckel term. These coefficients are determined semi-empirically, and are available for a limited number of the constituents of natural water. Harvie and Weare (1980) and Harvie et al. (1984), developed Pitzer models for the seawater system and the carbonate system, respectively. PHRQPITZ incorporates Pitzer parameters into a speciation code with an input program identical to
PHREEQE. The PHRQPITZ data base is significantly reduced from that of PHREEQE, and relies heavily on the findings of Harvie and Weare (1980) and Harvie et al. (1984).

In this study, PHRQPITZ is used more frequently than PHREEQE because the shallow groundwaters considered in the reactions generally exceed an ionic strength of 1.0, and the minerals of interest are evaporites. When reactions involving silicates are of interest e.g. clay minerals, PHREEQE must be used because Pitzer interaction coefficients have not been calculated for the silicate system. PHREEQE and PHRQPITZ are used to test potential hydrologic situations such as mixing two groundwaters of different compositions, equilibrating a groundwater with a particular mineral, or reacting one or more minerals in proportions determined by the program NETPATH.

NETPATH is an inverse model, defined as a model that calculates an intermediate mass transfer reaction, given initial and final concentrations. NETPATH also requires a set of reacting phases (e.g. minerals, ion-exchange, isotopic exchange) determined by the modeler. NETPATH calculates the amount of the specified phases that would need to dissolve or precipitate to attain the final concentration, based on a series of linear equations, but does not consider whether the reaction is thermodynamically possible (i.e. the model assumes an infinite supply of reactants and does not consider mineral saturation indices). NETPATH is used in this study to model hypothetical dissolution and ion exchange reactions, as well as to determine groundwater mixing proportions. The reactions suggested by NETPATH are modeled in PHREEQE or PHRQPITZ to determine their thermodynamic feasibility.

Each of the tools described above, in addition to analyses for the major ionic constituents of groundwater, are used to discern the various hydrologic processes that have influenced the chemistry of the groundwater at Owens Dry Lake.
CHAPTER IV. SHALLOW GROUNDWATER SYSTEM

An understanding of the hydrologic and chemical processes which govern the shallow groundwater system (<10 m deep) is important for several reasons. Evaporation of shallow groundwater at the lake bed surface promotes the formation of minerals which comprise the salt crust. Certain combinations of physical and chemical processes dictate the mineralogy and morphology of the surface precipitates, and the ease with which the precipitates are eroded. The geochemistry of the shallow groundwater also documents the effects of total lake desiccation, a record that may not be evident from deeper groundwater. However, processes that have influenced the geochemical signature of the shallow groundwater, prior to lake desiccation, may have acted on the deeper groundwater at one time.

The hypothesis to be tested is that shallow groundwater geochemistry is controlled mainly by evaporation, with mineral (salt) dissolution, groundwater mixing, and morphologic changes in salts exerting some influence. This hypothesis is evaluated by examining the relationship between chemical composition of shallow groundwater and chemical composition of salt precipitates, and describing solute distribution in the shallow groundwater. Several physical processes, including evaporation, are evaluated as potential controls of shallow groundwater geochemical and isotopic signatures.

Locations and lithologic characteristics

Data from the analysis of numerous shallow groundwater samples were provided by Great Basin UAPCD. Great Basin UAPCD installed 162 piezometers extending from north to south along the eastern margin of the lake bed (Figure 6). Piezometers are completed at approximately 1 m (95 piezometers), 3 m (61 piezometers) and 10 m (six piezometers). The individual piezometer sites are specified by the transect (e.g. N7, M8), the number of the piezometer in the transect (numbers increase with distance from the lake bed margin), and the depth. For example, the N7(6) piezometer is the sixth piezometer from the lake bed margin.
Figure 6: Piezometer network installed by Great Basin UAPCD in summer of 1992.
margin in the N7 transect, and the piezometer is completed at a depth of 1 m. Great Basin UAPCD sampled the piezometers during April and May, 1993, collected field measurements, and had groundwater samples analyzed for major ions, selected trace metals, and δD and 18O. Analytical results are presented in Appendix B. A potentiometric map of the 1 m piezometer network compiled from data collected in 4/94 is illustrated in Figure 7.

At the time of piezometer installation, Great Basin UAPCD documented the lithologic characteristics of the sediments in which each piezometer was completed. Generally, subsurface lithology consists of salt crust underlain by a range of silty sand to sand with gravelly layers, which overlie silty clays and clays. The 10 m piezometer locations are lithologically distinct from the shallower boreholes located closer to the lake margin. The 10 m locations are characterized by a very thin layer (5 to 30 cm) of sand under a salt crust, underlain by clays to their total depth. Salt crystals and pellets, when present, are generally concentrated in the upper coarser-grained materials and are more scarce in the clays. In most cases (70%) when layers of salt crystals or pellets are present in the sediments, they are in contact with groundwater. However, the thickest sequences of sediments with salt crystals tend to be above the water table. This is the case in the central portion of the lake bed (L9(5), M8(7), O6(8)). Subsurface salt distribution may reflect deposition that occurs with water table fluctuation. Thicker salt deposits (0.5 to 0.7 m) in two of the piezometers furthest from the lake bed margin (N7(9) and L9(5)) may reflect the extent of the brine pool prior to lake desiccation. The extent of the brine pool at various times is illustrated in Figure 8.

Elevation of the water table in the piezometers is generally within 0.75 m of the lake bed surface. Exceptions are the water levels in four of the 10 m piezometers (M8, N7, O6, P5) which range from 2 to 6 m below lake bed surface, and nine of the the 3 m piezometers (J10(8), K10(5), M8(5), N7(7), O6(6), P5(6), Q4(8), S3(4), and S3(5)), which are between 2 and 3 m below lake bed surface. Average seasonal variation in water levels varies from
Figure 7: Approximate piezometric contours of 1 m piezometers measured 4/94 by Great Basin UAPCD in meters above sea level.
Figure 8: Lake levels at Owens Lake including pre-irrigation (1,097.3 m) shoreline, the shoreline when the aqueduct was completed (1,092.7 m), and the maximum and minimum elevations from the 1969 flooding of the lake. Modified from Gale (1915) and Smith and Friedman (1986).
approximately 0.3 m in the 1 m and 3 m piezometers, to approximately 3 m in three of the 10 m piezometers (M8, N7, O6).

Oolites, CaCO₃ concretions, are important markers in the lake sediments. Oolites are not present in a continuous layer across the lake, but rather are encountered in only 13 of the piezometer cores and range in depth between 30 cm and 3.3 m below the current lake bed surface. A layer of oolitic sand is present between 0.6 and 1.0 m of depth in the N7(9) piezometer location. Depths at which oolites are found in the piezometer locations do not correlate with the depth of a 3.8 meter oolitic layer (and oolites mixed with sand and silt) encountered 1.3 m below ground surface at the site of OL-92. This lack of stratigraphic correlation is probably because the oolite deposits at OL-92 have filled in a channel. Oolites are informative markers because they are indicative of the point at which lake water reached calcite supersaturation.

**Geochemical characteristics**

**Overall**

For a better understanding of processes controlling shallow groundwater geochemistry, the concentrations of ions in the groundwater are examined for stoichiometric relationships, and groundwater constituents are compared to constituents in the precipitated salts.

Plots of the Na and Cl concentrations in shallow groundwater are shown in Figures 9a, 9b, and 9c. Concentrations in the 1 m, 3 m and 10 m piezometers indicate high correlations \( r^2 = 0.98 \). Sodium concentrations increase at roughly twice the rate of Cl concentrations. The slope of the 1 m piezometer data is 1.82, and slopes for the 3 m and 10 m data are 1.98 and 1.76, respectively.

Figures 10a, 10b, and 10c illustrate the Na and SO₄ concentrations in shallow groundwater. Increases in Na correlate reasonably well with increases in SO₄ in the 1 m and 3 m
Figure 9: Correlation between Na (Na) and Cl (Cl) in 1 m (a), 3 m (b), and 10 m (c) piezometer groundwater.
Figure 10: Correlation between Na and SO₄ concentrations in 1 m (a), 3 m (b), and 10 m (c) piezometer groundwater.
piezometers. Sodium increases approximately eight (7.8 - 7.9) equivalents for every SO$_4$ equivalent. Sodium and SO$_4$ correlation is less evident from the plot of the 10 m data. First, there are fewer data points to interpret, and second, SO$_4$ in the 10 m groundwater may be undergoing reduction, resulting in a lower correlation with non-redox ions such as Na.

The correlation between Na and HCO$_3$ ion concentrations (Figures 11a, 11b, 11c) is high in the 1 m and 3 m piezometers, and slightly less in the 10 m piezometers. Sodium increases, on average, 2.43 times as fast as HCO$_3$.

Correlation between the divalent cations (Ca + Mg) and the major anions is not evident from Figures 12a, 12b, and 12c. Concentrations of Ca and Mg are generally low compared to Na due to the precipitation of Ca and Mg carbonate minerals. Virtually all the piezometer locations are supersaturated with calcite. The five sites in Figures 12a, 12b, and 12c that appear to follow a trend of increasing Ca and Mg with anion concentration are not related by location, lithology, or field characteristics of the groundwater (pH, temperature).

Ionic relationships are evident between Na, Cl, SO$_4$, and HCO$_3$ based on high correlations of the plotted data. However, the slopes of the data do not reflect simple stoichiometric increases (e.g. Na and Cl would increase 1:1 as a result of halite dissolution). Also, scatter about the regression line increases with increasing solute concentration. This may indicate that increase in solute concentration in lower salinity groundwaters occurs by only one process, such as evapo-concentration, and solute increase in more saline waters occurs by several mechanisms, one of which may be dissolution of subsurface salts.
Figure 11: Correlation between Na and HCO₃ concentrations in 1 m (a), 3 m (b), and 10 m (c) piezometer groundwater.
Figure 12: Calcium and Mg concentrations plotted with (a) Cl, (b) HCO₃, and (c) SO₄ from the 3 m piezometers.
Relationship of shallow groundwater to salt crust

Since mineralogic analyses are not available on the salt encountered in the subsurface during piezometer installation, comparison is made with the constituents of the salt crust on the lake bed surface. The salt crust cannot be practically described as a grouping of particular minerals, as the crust composition varies greatly depending on temperature and humidity (Saint–Amand et al., 1986; Smith and Friedman, 1986). The morphology and composition of the surface salts can also be altered rapidly with changes in environment, which makes accurate analysis of the mineral composition difficult. Salt compositions are presented in Table 4.

Table 4: Chemical compositions of salts referred to in text

<table>
<thead>
<tr>
<th>Salt</th>
<th>Chemical Formula</th>
</tr>
</thead>
<tbody>
<tr>
<td>Burkeite</td>
<td>Na₆(SO₄)₂CO₃</td>
</tr>
<tr>
<td>Gaylussite</td>
<td>Na₃Ca(CO₃)₂·5H₂O</td>
</tr>
<tr>
<td>Mirabilite</td>
<td>Na₂SO₄·10H₂O</td>
</tr>
<tr>
<td>Halite</td>
<td>NaCl</td>
</tr>
<tr>
<td>Natron</td>
<td>Na₂CO₃·10H₂O</td>
</tr>
<tr>
<td>Natronite</td>
<td>NaHCO₃</td>
</tr>
<tr>
<td>Thenardite</td>
<td>Na₂SO₄</td>
</tr>
<tr>
<td>Thermonatrite</td>
<td>Na₂CO₃·H₂O</td>
</tr>
<tr>
<td>Trona</td>
<td>Na₃H(CO₃)₂·2H₂O</td>
</tr>
</tbody>
</table>

At each of the piezometer locations, an integrated sample of the surface crust down to the underlying clay or sand was collected by Great Basin UAPCD and analyzed for weight percentages of various salts. Salts are reported as NaSO₄, NaHCO₃, Na₂CO₃, and NaCl; variations in hydration state are not reported, but are not needed when comparing only amounts of gross ionic constituents (Na versus SO₄, for example). The reported percent–by–weight values are converted to relative mole amounts of the ionic constituents of the salts, and ratios between ionic constituents are averaged over all the piezometer sites. These ratios are presented in Table 5, as well as the standard deviations about the means of the
data, and are compared to the averages of the ionic ratios in groundwater (from slopes of Figures 9–11). The ratio of moles of Na in the crust compared to HCO$_3$ is 2.43, which is nearly identical to the ratio in groundwater (2.47). The average Na concentration in the shallow groundwater is approximately 16 times the SO$_4$ concentration in molar terms (from Figure 10, expressed as equivalents, Na is approximately eight times the concentration of SO$_4$). The average of moles of Na compared to moles of SO$_4$ in the crust is approximately 12.6. The Na to Cl ratio in the crust (3.77) is more than that ratio in the groundwater (1.85).

Table 5: Comparison of ionic ratios in groundwater to ionic ratios in the salt crust. Note that ratios of Na/Cl and Na/SO$_4$ have large standard deviations compared to the average ionic constituents in the crust.

<table>
<thead>
<tr>
<th>Ionic Ratio</th>
<th>Average Molar Ratio in Groundwater</th>
<th>Average Mole Ratio in Crust</th>
<th>Standard Deviation of Crust Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na/HCO$_3$</td>
<td>2.47</td>
<td>2.43</td>
<td>0.89</td>
</tr>
<tr>
<td>Na/Cl</td>
<td>1.85</td>
<td>3.77</td>
<td>2.39</td>
</tr>
<tr>
<td>Na/SO$_4$</td>
<td>15.8</td>
<td>12.6</td>
<td>10.0</td>
</tr>
</tbody>
</table>

Good agreement between the two averages (groundwater to salt crust) would be expected if the salts are ubiquitous and regularly distributed in the sediments. There is good agreement between the Na–HCO$_3$ ratios in the groundwater and salt crust. The remaining ratios are not in good agreement, but the standard deviation of the averages for the SO$_4$ and Cl minerals in the crust are so large that it cannot be said definitively whether the two averages are significantly different. The solubility of the minerals may explain some of the variation. The solubility of halite is extremely high relative to the other minerals ($K_{sp}$halite = 37.19) (calculated from Harvie et al., 1984). The solubility products of the two most common Na–SO$_4$ minerals, thenardite and mirabilite ($K_{sp}$thenardite = 0.724 and $K_{sp}$mirabilite = 0.0611) (Harvie et al., 1984), are significantly lower than that of halite, but still more than nine orders of magnitude more soluble than the Na–CO$_3$ minerals ($K_{sp}$trona = 4.13 x 10$^{-12}$, $K_{sp}$nahcolite = 1.81 x 10$^{-11}$) (Harvie et al., 1984). Because of low solubility, Na–CO$_3$ salts are the most likely to be present and regularly distributed in sediments. Sodium–sulfate salts and halite will be less common and perhaps more irregularly distributed. This distribution contributes
to the large standard deviation as well as lack of agreement between crust ratios and groundwater ratios.

The chemistry of the pre-1872 lake water is also informative because prior to 1872, all available salts were in solution. The ionic ratios of pre-1872 lake water should be mirrored in current groundwater unless other processes have altered the geochemistry. An analysis collected by Loew (1876) and cited in Saint-Amand et al. (1986) is presented in Table 6.

Table 6: Analyses of 1866 lake water from Loew (1876) as cited in Saint-Amand et al. (1986). Chemical composition was reported in weight percent of total dissolved solids. Conversion to mmol/kg is performed in order to compare ratios with ionic ratios in crust and current groundwater, as presented in Table 5.

<table>
<thead>
<tr>
<th>%</th>
<th>Na</th>
<th>K</th>
<th>Cl</th>
<th>SO4</th>
<th>CO3</th>
<th>HCO3</th>
<th>TDS</th>
</tr>
</thead>
<tbody>
<tr>
<td>mg/kg</td>
<td>35.58</td>
<td>4.55</td>
<td>22.28</td>
<td>15.46</td>
<td>21.86</td>
<td>6.56</td>
<td></td>
</tr>
<tr>
<td>mmol/kg</td>
<td>23.3</td>
<td>40</td>
<td>2,98</td>
<td>14.6</td>
<td>16</td>
<td>10.142</td>
<td></td>
</tr>
</tbody>
</table>

Ionic ratios:
- Na/HCO3 = 2.12
- Na/Cl = 2.46
- Na/SO4 = 9.67

The Na/HCO3 ratio is slightly lower in the 1866 lake compared to current groundwater and salt crust. The Na/Cl and Na/SO4 ratios are different among the three sources. The Na/Cl ratio of the 1866 lake water is intermediate between that of the current groundwater and salt crust. High solubility of halite may explain this distribution; as pre-1872 lake water evaporates, halite is more likely to stay in solution than to precipitate. Hence, Cl builds up in solution (current groundwater) as Na is precipitated in other less soluble salts (salt crust). The Na/SO4 ratio is significantly lower in the 1866 lake analysis than either the ratio in the current groundwater or salt crust. The implication, then, is that there was more SO4 relative to Na in the 1866 lake water than there is currently in either the groundwater or salt crust. This is probably explained by Na–SO4 and Na–CO3 salts being removed from the lake bed by wind. Both Na–CO3 and Na–SO4 salts are more plentiful and more easily wind eroded.
than halite and may have been disproportionately removed from the system. Mining of these two salts from the basin during the last century may also account for observed losses.

**Distribution of solutes in the shallow groundwater**

**Description**

As shallow groundwater moves from the lake margins toward the center of the lake as illustrated in Figure 7, patterns are evident that may reflect geochemical evolution, the influence of changing lithologies, or effects of physical processes. Major ion variations are examined in two directions: north to south along the eastern margin of the lake and through the central portion of the lake; and east to west from several points on the eastern lake margin toward the center of the lake.

The following figures illustrate Na, Cl, HCO₃ and SO₄ concentrations. Calcium and Mg are not included in the figures because the magnitude of the changes in divalent cations is generally less than 20 meq/L (compared to the several hundred meq/L changes in the other ions) and do not indicate discernable patterns along the transects. Chloride is omitted from some of the figures for the sake of clarity.

An examination of shallow groundwater chemistry variation from the north portion to the south portion of the lake bed is presented in Figures 13 and 14. Similar patterns between Na, SO₄, and HCO₃ concentration variations are observed in a north–south transect of piezometers along the eastern margin of the lake (Figure 13) and through the center of the lake bed (Figure 14). High and low concentrations correlate, as do the magnitude of the changes.

There is a general increase in salinity (Na, Cl, SO₄, and HCO₃) toward the center of the lake bed with lower concentrations at the north and south margins. Lower concentrations in the extreme north and south portions of the lake are probably due to recharge with fresher water.
Figure 13: Concentrations of Na, SO$_4$, HCO$_3$ from 3 m piezometers located along the eastern margin of the lake bed. Location 0 is the river delta.

Figure 14: Concentrations of Na, SO$_4$, HCO$_3$ from 3 m piezometers located in the center of the lake bed. Location 0 is the river delta.
in these areas. The influence of freshwater recharge in the northern, southern, and portions of the eastern margins is shown in Figures 15a and 15b. These figures are contour plots of saturation indices for mirabilite in the 1 m sites, and trona in the 3 m sites. Steep concentration gradients (e.g. from log SI = -12 to log SI = -4) are highlighted in Figure 15 as areas of freshwater recharge.

Figures 16 through 18 represent data from several piezometer transects from the eastern lake margin toward the center of the lake bed. In the 1 m piezometer transects, a general trend of increasing Na, Cl, HCO₃, and SO₄ is discernable but the slope of increase is not constant between adjacent locations (Figures 16a, 16b, 16c, 16d). Ionic concentrations tend to rise rapidly near the lake margin, then either stabilize or decrease rapidly and increase again with distance from the lake margin. Not all 1 m piezometer transects follow a pattern of rapid increase followed by decrease. For example, solute concentrations from transects Delta W and 6, pictured in Figures 17a and 17b, do not peak near the lake margin. In the 3 m piezometers, a trend of increasing concentration with distance from the lake margin is more consistent, although the magnitude of the increase still varies between adjacent piezometer locations. This discrepancy between the 1 m and 3 m piezometers is demonstrated by the 110 transect in Figure 18, as well as those transects in Figure 16. Transect P5, illustrated in Figure 16b, is an exception in that solute concentration in the 3 m piezometer groundwater also increases rapidly initially and then decreases.
Figure 15: Contour plots of (a) trona log saturation indices in 3 m sites, and (b) mirabilite log saturation indices in 1 m sites. Steeper gradients near north, south, and east lake margins indicate areas of highest freshwater recharge.
Figure 16(a): Piezometer transect M8. Correlation of Na, Cl, HCO₃, SO₄ with distance from lake margin (approximately 1097 meter contour). Sulfate scale is enlarged to the right.
Figure 16(b): Piezometer transect PS. Correlation of Na, Cl, HCO₃, SO₄ with distance from lake margin. Sulfate scale is enlarged to right. Note that scales are not identical between the two plots.
Figure 16(c): Piezometer transect N7. Correlation of Na, Cl, SO₄, HCO₃ with distance from lake margin. Sulfate scale is enlarged to the right.
Figure 16(d): Piezometer transect G9. 1 m piezometers only. Correlation between Na, Cl, SO$_4$, HCO$_3$ with distance from lake margin. Sulfate scale is enlarged.
Figure 17(a) and 17(b): Piezometer transects Delta W (a) and 6 (b), both from 1 m. These transects do not show a pattern of rapid salinity increase followed by decrease with distance, as in Figure 16. Flow direction and orientation identical for both plots.
Figure 18: Piezometer transect II0. Salinity trend (represented by Na and Cl concentrations) with distance from lake margin. One m piezometers follow trend of rapid increase followed by decrease; 3 m piezometers indicate a steady increase with distance. Note that scales are not identical between the two plots. Flow direction and orientation identical for both plots.
Explanation of solute distribution

Variation with sedimentologic and mineralogic characteristics

As expected from the preceding discussion of mineral solubility, variations in $\text{SO}_4$, $\text{HCO}_3$, and $\text{Cl}$ concentrations mimic variations in Na concentration. Variations from the correlated pattern probably reflect the types of salts available for dissolution (i.e., at a particular location, a Na–CO$_3$ salt may be much more plentiful than a Na–SO$_4$ salt). A correlation between presence of salts in the sediments and the concentration of dissolved ions in the groundwater is pictured in Figure 19. Precipitated salts become more abundant with distance from the lake bed margin. Solute concentration in groundwater from the N7 transect increases as groundwater moves onto the lake bed (from N7(1) to N7(5)). The highest solute concentrations are observed in groundwater from N7(9), which is completed in sediments of high salt content (Figure 19). This correlation suggests that one important controlling factor on the concentration of solutes in groundwater is the dissolution of salts precipitated in the sediments. This relationship, however, does not explain the rapid decrease in salinity between N7(5) and N7(6).

Deviation from the trend of increasing solute concentration with distance from the lake bed margin is observed in other 1 m transects in addition to the N7 transect. In these transects (M8, G9, I10) there is a rapid increase in solutes followed by a rapid decrease generally within 1.5 km from the lake bed margin. This characteristic is not observed in the 3 m piezometers at the same sites, and is therefore assumed to be due to one or more exclusively near-surface processes. These distinct changes in groundwater salinity occur in coarse-grained sediments. Ionic concentrations tend to stabilize as the center of the lake is approached and the sand content diminishes. These near-margin locations may be conducive to convective groundwater mixing between different density groundwaters. This possibility is discussed in the next section.
Surface salt crust. Ranges in thickness from <1.0 mm to 10 cm.

Silty sand to sandy silt. Sand ranges from fine to coarse-grained. Occasional gravels, clay pellets.

Clay to silty clay. Dark green to black, odoriferous.

Hard aggregated salt pan. Fine to coarse-grained sands cemented with salt; compacted sands and oolites.

Salt crystals and pellets to 2 mm diameter. Symbol is qualitative and representative only of depths where salt crystals observed. Symbol is not meant to indicate quantity of salt.

Figure 19: Chemical and lithologic variations in piezometer transect N7. Bar chart indicates Na, Cl, SO₄ concentrations in 1 m, 3 m, and 10 m piezometer groundwater.
Density driven groundwater mixing

The erratic behavior of dissolved ions in groundwater near lake margins may be due to the occurrence of groundwater mixing induced by groundwater density differences.

Vertical convection of dense, evapo-concentrated brines, initiated by physical instability, has been suggested as a transport mechanism in field studies (Duffy and Al-Hassan, 1988, Ferguson et al., 1992), and supported by laboratory studies and mathematical treatments confirming the occurrence of interfingering of different density fluids (Wooding, 1960; Tyler and Wooding, 1991). Instability is dependent on the difference in density between the two fluids, the evaporation rate, and the hydraulic conductivity of the sediments. The physical instability is gauged by the Rayleigh number:

\[ R_c = \frac{K \Delta \rho}{E} \]  

(8)

where \( R_c \) is the Rayleigh number, \( K \) is the hydraulic conductivity of the sediments, \( \Delta \rho \) is the difference in density between the two fluids, \( \rho \) is the density of the deeper fluid, and \( E \) is the evaporation rate.

The preceding expression for the Rayleigh number was modified by Tyler and Wooding (1991) from expressions that do not directly incorporate the effects of evaporation. The critical Rayleigh number represents the point at which the conditions are favorable for convection, and is approximately equal to 10 (Tyler and Wooding, 1991).

Calculations based on estimated hydraulic conductivity values and a range of evaporation magnitudes (Kranz, 1994) for Owens Dry Lake demonstrate the potential for Rayleigh instability (Table 7).
Table 7: Calculated Rayleigh numbers for a range of physical parameters.

<table>
<thead>
<tr>
<th>$K$ (mm/day)</th>
<th>$E$ (mm/day)</th>
<th>$\Delta Q$ (g/cm³)</th>
<th>$R_c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.864</td>
<td>0.50</td>
<td>0.086</td>
<td>0.149</td>
</tr>
<tr>
<td>0.864</td>
<td>0.10</td>
<td>0.086</td>
<td>0.74</td>
</tr>
<tr>
<td>8.64</td>
<td>0.50</td>
<td>0.086</td>
<td>1.49</td>
</tr>
<tr>
<td>8.64</td>
<td>0.10</td>
<td>0.086</td>
<td>7.43</td>
</tr>
<tr>
<td>8.64</td>
<td>0.05</td>
<td>0.086</td>
<td>14.9</td>
</tr>
<tr>
<td>8.64</td>
<td>0.50</td>
<td>0.060</td>
<td>1.01</td>
</tr>
<tr>
<td>25.9</td>
<td>0.50</td>
<td>0.060</td>
<td>3.03</td>
</tr>
<tr>
<td>25.9</td>
<td>0.10</td>
<td>0.086</td>
<td>22.3</td>
</tr>
<tr>
<td>25.9</td>
<td>0.10</td>
<td>0.060</td>
<td>15.1</td>
</tr>
<tr>
<td>25.9</td>
<td>0.05</td>
<td>0.086</td>
<td>44.6</td>
</tr>
<tr>
<td>86.4</td>
<td>0.50</td>
<td>0.060</td>
<td>10.1</td>
</tr>
<tr>
<td>8.64</td>
<td>0.10</td>
<td>0.060</td>
<td>5.05</td>
</tr>
<tr>
<td>86.4</td>
<td>0.05</td>
<td>0.060</td>
<td>50.5</td>
</tr>
<tr>
<td>43.2</td>
<td>0.10</td>
<td>0.060</td>
<td>25.3</td>
</tr>
<tr>
<td>8.64</td>
<td>0.05</td>
<td>0.039</td>
<td>6.44</td>
</tr>
<tr>
<td>25.9</td>
<td>0.10</td>
<td>0.039</td>
<td>9.65</td>
</tr>
</tbody>
</table>

The Rayleigh number is most sensitive to the evaporation rate and the hydraulic conductivity. If evaporation is very low (0.05 mm/day) then almost any combination of density difference and hydraulic conductivity could be unstable, as demonstrated in Table 7. If the evaporation rate is higher, sufficient density difference between the two fluids or greater hydraulic conductivity is required for an unstable situation. The results presented in Table 7 suggest that a hydraulic conductivity value of 25.9 mm/day is sufficient under most density differences and evaporation rates to allow convection. In situations where the upper layer of sediment is coarse-grained, the near surface hydraulic conductivity could be at least an order of magnitude greater than 25.9 mm/day, and instability is likely. If the evaporation rate is as high as 0.5 mm/day, then a hydraulic conductivity of at least 86.4 mm/day is required for density instability.

Sufficient density instability to initiate convection is more likely in the near-surface, higher permeability sediments (sands) found near the margins of the lake bed, rather than in the
clays that dominate the central lake bed region. A persistent source of brine is also necessary for the existence of continuous convective mixing. Although a brine source is found in the central–western portion of the lake, brine in the eastern lake bed could only originate from evapo–concentration of shallow groundwater.

Increased evaporation of the shallow groundwater with distance from the lake bed margin i.e. in the direction of groundwater flow, is reflected in the isotopic signatures of piezometer groundwater. Figures 20 through 24 show that as groundwater moves from the lake bed margin toward the center of the lake there is an overall trend of increasing δD content. However, rapid decrease in δD content correlates with decrease in solute concentrations (compare with Figures 16 through 18) in all the illustrated transects except M8 (Figures 16a and 20). Convective mixing may not be initiated until evaporation has progressed enough to increase the density of the shallowest groundwater to the point that the shallowest groundwater is unstable overlying less dense groundwater. This may explain why the rapid change in solute concentration tends to occur between 0.35 and 1.5 km from the lake bed margin, rather than closer to the lake bed margin.
Figure 20: Deuterium content piezometer transect M8. Groundwater becomes isotopically heavier as center of lake bed is approached.
Figure 21: Deuterium content piezometer transect PS. Deuterium profile is similar to pattern of solute concentrations illustrated in Figure 16b.
Figure 22: Deuterium content piezometer transect N7. Deuterium profile is similar to pattern of solute concentration illustrated in Figure 16c.
Figure 23: Deuterium content (a) piezometer transect G9 and (b) piezometer transect Delta W, both from 1 m depths. Deuterium profile similar to pattern of solute concentrations illustrated in Figures 16d and 17a.
Figure 24: Deuterium content piezometer transect II0. Deuterium profile similar to pattern of solute concentrations illustrated in Figure 18. Note that scales are not identical between the two plots.
A hypothetical density–induced mixing scenario that is supported by observed data, is illustrated in Figures 25 and 26. The N7 transect is chosen as an example, but similar behavior (rapid increase in solutes near the margin, followed by rapid decrease) is also observed in the M8, G9, and I10 transects. The mixing model requires an input groundwater, such as that from N7(1)1 with a relatively low Cl concentration of 627 mmol/L, and a low δD value of −105‰. As this groundwater moves further from the lake margin, salts precipitated in the sediment are dissolved and evaporation occurs (Figure 26). Both of these processes are necessary in order to make a hypothetical brine that has a high enough Cl concentration (2,000 mmol/L) to be density unstable as well as an enriched δD concentration (−50 ‰) relative to the starting δD value. The influence of both of these processes also agrees with the initially rapid increase in Cl and δD as groundwaters move west from the lake margin. This hypothetical brine becomes unstable overlying a less dense groundwater, perhaps similar in concentration to groundwater at 3 m of depth (Cl = 845 mmol/L, δD = −72 ‰). A mixture of these two waters in the proportions indicated in Figure 25 (47:53) results in a moderately saline (Cl = 1,383 mmol/L) groundwater with relatively enriched δD (−62 ‰), as measured in N7(6)1.

![Figure 25: Model of mixing scenario to explain erratic behavior of dissolved constituents in groundwater near lake margins. Upper values are δD‰; lower values are Cl in mmol/L. A cross sectional view of this scenario is presented in Figure 26.](image-url)
Mixing of shallower (1 m) and deeper (3 m) groundwater is suggested also by the change in gradients observed in three piezometer transects that demonstrate this variable concentration behavior. In the N7, M8, and I10 transects there is a decrease in the magnitude of the upward gradient (Δgradient = 0.1 m – 0.6 m) from the 3 m piezometers to the 1 m piezometers prior to or at the same location that the solute concentration decreases rapidly. This decrease in upward vertical gradient indicates an increase in vertical hydraulic conductivity (specifically permeability) at these locations. The decrease in groundwater salinity within 1.5 km of the lake bed margin, therefore, is most likely controlled by sedimentologic
characteristics such as sand thickness and permeability, rather than a process such as faulting that would promote the upward discharge of deeper, less saline groundwater. The generally increasing trend of salinity in the 3 m groundwater with distance from the lake bed margin also negates the possibility of upward discharge of less saline groundwater.

A pattern of rapidly changing groundwater salinity is displayed in transects in the central portion of the eastern margin of the lake bed. This solute distribution may be more ubiquitous than is obvious from available data; more shallow groundwater data points at both 1 m and 3 m might reveal similar dynamics in the northeast section of the lake bed. The model of groundwater convection is most likely appropriate only for the groundwater that flows through the sandy sediments near the lake margin, generally less than 2 m below lake bed surface, where evaporation is more likely to produce sufficient differences in groundwater density, and hydraulic conductivities are high enough to promote instability. This location of relatively lower solute concentrations (0.35 – 1.5 km from lake margin) may be an appropriate location for establishing vegetation.
Isotopic variation in shallow groundwater

Deuterium versus oxygen-18

A plot of the δD and 18O values from all the piezometer samples (Figure 27) follows a line of slope 4.33 ($r^2 = 0.97$). This suggests the evaporation of shallow groundwater, since an ideal evaporation line originating from the Meteoric Water Line has a slope approximately equal to 5.0 (Craig, 1961b). However, this simple plot does not indicate whether groundwater at different depths has been subjected to evaporation at different times or what other mechanisms may have produced the observed isotopic signatures. The range of isotopic values is great ($δD = -120 \%_o$ to $-45 \%_o$), which also suggests that more than one process controls the isotopic signatures of the shallow groundwater.

If the values from various depths are plotted separately, each group of piezometers plots on a different slope from the meteoric water line. Deuterium and oxygen data from the 1 m piezometers plot on a line of slope 4.19 (standard error $= 0.093$). Data from the 3 m piezometers plot on a line of slope 4.58 (standard error $= 0.12$). The 10 m piezometers, although the correlation is slightly lower based on only six data points, plot on a line of 4.79 (standard error $= 0.75$). The slopes of the 1 m and 3 m piezometers are not statistically distinguishable if three standard errors are added to each of the regression slopes (within a 99% confidence interval). The two slopes (1 m and 3 m) barely overlap if only two standard errors (95% confidence) are added to the regression slopes (highest slope of 1 m data is 4.38, and lowest slope of 3 m data is 4.35). Because of the small number of sample points in the 10 m data set, the slope of the regression cannot be statistically differentiated from the other two slopes.

Slight slope differences with depth may indicate that evaporative effects are only partially responsible for the isotopic values observed in the groundwater. Slope differences may be
Figure 27: Stable oxygen and δD isotopes from all shallow piezometers plotted with Craig's (1961a, 1961b) meteoric water line (MWL).
due to the effect of variable groundwater mixing, or evaporation under different climatic conditions. Each of these scenarios is examined in this section.

**Deuterium versus chloride**

If equilibrium evaporation is the sole mechanism controlling groundwater chemistry, δD and Cl concentrations would be linearly related since both δD and Cl ideally behave as conservative substances, prior to halite precipitation. All the piezometer groundwater samples are undersaturated with respect to halite.

The δD–Cl relationship in the shallow groundwater compared with similar data from Friedman et al. (1976) (Figure 28) demonstrates again that evaporation does not fully explain groundwater geochemical variation. Data from Friedman et al. (1976) are from the 1969 flooding and subsequent desiccation of Owens Dry Lake. The authors assume that the D evaporating from the lake surface water is in isotopic equilibrium with the D remaining in the liquid reservoir. This assumption is validated by a linear relation between the log of the evolving D/H ratios (R/R₀) and the log of the fraction of water remaining in the lake (θ). where R is the D/H ratio in the sample, R₀ is the same ratio in the initial input, and θ is the fraction of water remaining in lake relative to original volume (represented by Li_initial/Li_measured). Friedman et al. (1976) used Li as a conservative tracer since Li is not involved in any precipitation or other removal mechanisms.

Because no Li data are available from the piezometer groundwater, comparisons are made between δD and Cl data of Friedman et al. (1976) and similar data from this study. The use of Cl in this case is valid because none of the samples of the shallow groundwater are near halite saturation, and the Friedman et al. (1976) data that are used are near the initiation of the evaporation process. A plot of the piezometer chemistry compared to the line generated from Friedman et al. (1976) is shown in Figure 28. In this case, θ = Cl_initial/Cl_measured, where the
Figure 28: Comparison between evolution of piezometer $D$ and $Cl$ with that from Friedman et al. (1976) data from the equilibrium evaporation of the surface brine pool. $R$ is the measured $D/H$ ratio, $R_0$ is the initial $D/H$ ratio (Owens River). $\theta$ is the fraction of water remaining in the reservoir, calculated from $Cl_{\text{initial}}/Cl_{\text{measured}}$, where the initial $Cl$ concentration is that of the Owens River.
initial Cl concentration is taken to be that of the Owens River and the measured Cl concentrations are from each of the piezometers.

Both the slope and the correlation of the two data sets (Friedman et al., 1976 and the shallow groundwater) are visibly and statistically different. Evaporation of the shallow groundwater evidently does not follow an equilibrium process, or the ionic constituents (and hence Cl) are controlled by other factors in addition to evaporation that lead to scatter in the Cl data.

Description of different types of isotopic/geochemical signatures

A plot of δD versus Cl (Figure 29) including piezometer samples from each of the three depths indicates only a moderate correlation ($r^2 = 0.73$) between these two parameters. The scatter observed in both Figures 28 and 29 is the result of samples enriched in δD with relatively low Cl concentrations (designated as type A, generally samples from the 3 m and 10 m piezometers and approximately 50 percent of the 1 m piezometers), and relatively isotopically light samples with high Cl concentrations (designated as type B, the remaining 50 percent of 1 m piezometer samples). A distinction was made between Type A and B samples initially because a noticeable correlation with depth was observed. Those 1 m samples that plot in the lower portion of the graph (Type B) tend to be located nearer to the lake bed margin than those 1 m samples that plot in the upper portion of the graph (Type A). The influence of depth and lithology on the chemical and isotopic signature of the shallow groundwater is discussed further in the following section.

There is a cluster of 41 samples with low Cl (<500 mmol/L) and low δD (most <−100‰ ) that are not considered with those samples that evolve along a line of increasing δD and Cl concentrations. Most of these samples are located near the margins of the lake bed where there is a higher percentage of fresh water from spring discharge.
Figure 29: Deuterium and Cl values for all shallow piezometers. Type A samples are those with relatively enriched δD values and relatively low Cl. Type B samples are relatively depleted in δD and enriched in Cl.
In order to quantify the geochemistry of different sites, saturation indices are compared between type A and B groundwater. A saturation index represents how close a solution is to equilibrium with a particular mineral:

\[ SI = \frac{IAP}{k_{sp}} \]  

where \( SI \) is the saturation index, \( IAP \) is the ion activity product of the mineral-forming ions measured in the solution, and \( k_{sp} \) is the solubility product (determined by experiment) of the mineral. A saturation index of 1.0 represents thermodynamic equilibrium, greater than 1.0 represents supersaturation, and less than 1.0, undersaturation. Saturation indices are usually reported on a logarithmic scale: in this case equilibrium is 0.0, negative numbers indicate undersaturation, and positive numbers indicate supersaturation. Figures 30a and 30b picture the distribution of saturation indices for calcite, mirabilite, trona, and halite (chosen to represent a range of compositions), compared between type A and B samples. The histogram plots were constructed by computing the frequency of a range of saturation indices using data calculated by PHRQPITZ.

Type B samples have only slightly higher average saturation indices with halite and calcite, although calcite is the only mineral that is supersaturated in these waters. Current mineral saturation states do not appear to be a controlling factor in distinguishing Type A from Type B groundwater.
Figure 3A: Saturation indices for calcite and mirabilite compared between Type A and B.
Figure 30b: Saturation indices for trona and halite compared between Type A and Type B piezometer groundwater.
Explanati on of Type A waters

Salt morphology changes

Change in salt morphology with temperature has been documented in the surface crust (Saint Amand et al., 1986) and in salts crystallizing in the vicinity of the brine pool (Smith and Friedman, 1986). Trona, thenardite, burkeite and thermonatrite tend to be more stable at higher temperatures and less humidity (summer); natron and mirabilite are more common in the winter months when temperatures are less than 18°C. Halite solubility is relatively insensitive to temperature change and is therefore stable in both the winter and summer months. There have been no studies on the effect of temperature changes on salts precipitated in the subsurface, although the groundwater temperature in the piezometers at the time of sampling (April) ranged between 16°C and 24°C.

The argument presented in this section is that water molecules released when subsurface salts change with temperature have the potential to dilute the existing groundwater. Dilution could explain the sites with enriched δD but relatively low Cl concentrations (Type A groundwater). Variation in the groundwater chemistry depends on the amount of salt in the subsurface at a particular site, the composition of those salts, and the depth to which temperature variations penetrate the subsurface. The piezometers were sampled in April, a time when subsurface temperature would be increasing.

One scenario was modeled which included the results of a hypothetical salt morphology change with temperature (natron to trona) that releases eight moles of water. The starting composition chosen is a brine sampled by Friedman et al. (1976) from the brine pool on February 26, 1970. This brine composition is nearly identical to the composition of the lake in 1912 (prior to initiation of desiccation) as cited in Gale (1915, p. 258), and may be representative of the initial concentration of groundwater in the 3 m piezometers, prior to salt morphology changes. PHRQPITZ was used to simulate the addition of eight moles of water
(released from natron as temperature increases above 18°C), by mixing the brine with pure water in a ratio of 87.3/12.7 percent. This ratio was calculated by adding 8 mols water to 1 kg water (55.09 mols). Eight mols is 12.7 percent of the total mols (63.09 mols). The resulting Cl concentration is 0.67 mol/kg H₂O (the concentration units used in PHRQPITZ). The resulting δD concentration is calculated assuming the δD value of the water in natron is -10 ‰ (Smith and Friedman, 1986). The mixture results in a final δD concentration of -72.9 ‰. A groundwater with a Cl concentration approximately equal to 670 mmol/kg H₂O and a δD value close to -73 ‰ is close to an average concentration of several Type A (3 m) groundwater samples with relatively low Cl (780 mmol/L) concentrations (compared to other shallow groundwater sites) and relatively enriched δD values (-72 ‰).

**Groundwater mixing – convection**

Another process that may explain solute and isotopic concentrations in some of the 1 m Type A groundwater is convective mixing between groundwaters of different densities, as described in a previous discussion of solute distribution on the lake bed (page 59). The goal is to generate a water that has a moderate to low Cl concentration and an enriched δD concentration relative to other shallow groundwater samples. As demonstrated with the example of the 1 m N7 transect (Figures 25 and 26), mixing between a hypothetical brine and less saline water (similar in concentration to groundwater in the 3 m piezometers) will generate a groundwater with isotopic and solute concentrations similar to the 1 m Type A groundwater (N7(6)1 is used in the example).

**Explanation Type B waters**

**Groundwater mixing – convection**

Groundwater mixing generated by convective processes may also explain the geochemistry and isotopic characteristics of some Type B groundwater, but the use of this model to explain Type B groundwater requires that the proportions of mixing be different from the Type A
model. A smaller contribution from the brine (~10%) will result in a disproportionate increase in Cl compared to δD. A mixture of 10 percent brine (Cl = 2000 mmol/L, δD = -50 ‰) with 90 percent 3 m groundwater (110(4)3) results in a Cl increase of 90 percent, but a δD increase of only 5.8 percent. This may explain the Type B samples up to approximately 1,000 mmol/L Cl, but does not explain those waters (such as 110(3,4,5)1) with Cl concentrations greater than 1,000 mmol/L with depleted δD values (<-100 ‰). These higher Cl samples require greater than a 50% contribution from a hypothetical brine, which enriches the δD values more than is observed (to values greater than -75‰). Explanation of highly enriched Cl samples is the subject of the next section.

Partial desiccation of lake – solute memory

Groundwater that is enriched in Cl (>1,000 mmol/L) but depleted in δD (<-100 ‰) requires a process that increases Cl without an increase in the δD value above that observed in many of the Type B samples. Dissolution of salt could explain this situation, but the lithologic characteristics of these type B samples do not indicate a greater proportion of precipitated salts in sediments than around the Type A piezometers. A more likely explanation, especially considering the location of many of these anomalous samples (Figure 6) (11,10), and Sulfate transects near the margin in the east-central portion of the lake bed, P(5) and 6 transects at the southern and northern margins of the lake, respectively) is the desiccation or near-desiccation of the lake in these regions, followed by replenishment with isotopically lighter groundwater from springs. These areas have the steepest concentration gradients with distance from the lake margin, presumably due to higher spring discharge that mixes with shallow groundwater. In drier periods, the lake would desiccate near its margins first and most frequently. Solutes would precipitate or concentrate with desiccation, but the δD signature would not be retained since all or most of the water would evaporate. When recharge from springs replenished these areas, precipitated salts would dissolve, raising the
Cl concentration, but the predominant δD signature would be that of the recharging water, which was light. This may explain some of the Type B waters with relatively high Cl concentrations that are depleted in δD relative to most of the other shallow groundwater samples.

**Explanation of groundwater from the 10 m piezometers**

None of the processes discussed to this point has yielded a satisfactory explanation for the composition of the groundwater in the 10 m piezometers. The 10 m piezometer groundwater is anomalously enriched in δD (−69 ‰) and depleted in Cl concentrations (372 mmol/L) compared to other shallow groundwater sites. However, the groundwater at 10 m is highly saline compared to Owens River input. The origin of the solutes in the 10 m sites must be identified in order to decipher the processes that have determined the chemistry at this depth. Have the solutes diffused to this depth from the surface or is the solute concentration a result of an in-situ process which progressed over time?

The age of the sediments in the OL-92 core at 10 m of depth is approximately 15,000 years based on 14C dating of humate at this depth (Bischoff et al., 1993). At 15,000 years ago and for at least 15,000 years prior to that, the lake was not a terminal lake (Benson and Bischoff, 1993). Therefore the original pore waters in these sediments would have been relatively fresh, similar in concentration to the Owens River. In order to explain both the solute and isotopic concentrations currently observed in the 10 m groundwater, diffusion must have occurred. The processes invoked to explain the solute chemistry observed at 10 m of depth must also support the isotopic values observed in the 10 m groundwater.
Diffusion of brine from most recent lake desiccation

Diffusion of evapo-concentrated solutes from the surface to 10 m of depth over the time period since the lake was most recently desiccated (approximately 90 years) is not theoretically possible. This is demonstrated by the solution to the diffusion equation:

\[ C_{(x,t)} = C_o \text{erfc} \frac{x}{2 \sqrt{D^* t}} \]  

(Freeze and Cherry, 1980, p. 104) (10)

where \( C_o \) is a constant concentration at the surface. If an initial concentration of solute in the groundwater is taken into account:

\[ \frac{C_{(x,t)}-C_i}{C_o-C_i} = \text{erfc} \frac{x}{2 \sqrt{D^* t}} \]

where \( C(x,t) \) is the concentration at a certain depth and time, \( C_o \) is the concentration at \( x = 0 \) for all times, \( C_i \) is an initial concentration, \( \text{erfc} \) is the complementary error function, \( x \) is the distance of interest (depth), \( D^* \) is the diffusion coefficient, and \( t = \) time.

The initial Cl concentration \( (C_i) \) is that of the Owens River (0.395 mmol/L), and the concentration of high salinity water at the surface \( (C_o) \) is 3.000 mmol/L Cl, the approximate concentration of the most desiccated brine sampled by Friedman et al. (1976). A range of \( D^* \) values realistic for Owens Dry Lake sediments (Kranz, 1994), and \( t = 90 \) years is input into the diffusion equation. The results are pictured in Figure 31, which indicates that even if sediments were characterized by the highest diffusion coefficient, Cl at a concentration of 372 mmol (concentration in the 10 m piezometers) would not have diffused beyond 4 m of depth in 90 years.

Diffusion of brine from desiccation 5,000 years ago

According to the \( \delta D \) and salinity content of the pore waters extracted from the OL–92 core (Friedman et al., 1993), the lake underwent a partial desiccation approximately 5,000 years ago.
Figure 31: Theoretical diffusion profiles for a range of diffusion coefficients empirically determined (Kranz, 1994) for Owens Dry Lake sediments. Initial Cl concentration is approximately equal to most evaporated brine in Friedman et al. (1976).
Figure 32 illustrates the diffusion profiles calculated with the same data as Figure 31, but with time allowed for diffusion equal to 5,000 years. This figure demonstrates that given 5,000 years, solutes could diffuse from the surface to result in the concentrations observed in the 10 m piezometers. However, this model requires a constant concentration of solutes at the surface (3,000 mmol/L), which there is no evidence to support. After the partial lake desiccation 5,000 years ago, the solute concentration in the lake decreased again according to the OL-92 pore water chemistry.

The depth of the sediments that would have been deposited around the time period of the partial desiccation is 5 m (Bischoff et al., 1993). Therefore concentrated solutes and heavy isotopes would only need to diffuse 5 m (rather than 10 m) to affect the concentration of the current 10 m groundwater. Recall that after approximately 8,900 years ago, Owens Lake became a terminal lake, and consequently solute concentrations would have continued to increase from that time. Therefore, a steep chloride concentration gradient would have always existed between the surface and deeper groundwater after 8,900 years ago.

Throughout this time period, however, the lake continued to receive fresh water input from the Owens River. The isotopic composition of the 10 m groundwater may be explained by two processes, steady-state evaporation and diffusion. Evaporation is necessary to increase the isotopic value from the initial value of the Owens River (δD = -120 ‰) to that observed (δD = -69.5 ‰). Diffusion is then necessary to explain the presence of the isotopically heavy water at 10 m depth. The steady-state solution is described below.
Figure 32: Theoretical diffusion profiles for a range of diffusion coefficients empirically determined (Kranz, 1994) for Owens Dry Lake sediments. Initial Cl concentration is approximately equal to most evaporated brine in Friedman et al. (1976) and the time allowed for diffusion to take place is 5,000 years.
Evaporation with continuous river input

Provided with a supply of a low Cl input water (from the Owens River) and evaporation, the 10 m groundwater isotopic composition could be generated. Evaporative effects can be examined by addressing the isotopic exchange of lake water with water vapor in the atmosphere, or by neglecting the effects of this exchange. Friedman et al. (written communication, 1994) derive an expression assuming that all inflow leaves as evaporation (terminal lake):

\[ R_L = \frac{R_i}{\alpha} \]  

where, \( R_L \) is the D/H ratio in the lake water, \( R_i \) is the D/H ratio in the Owens River, and \( \alpha \) is the fractionation factor determined to be 0.935 (Friedman et al., 1976, with lake water between temperatures of 5°C and 28°C). Using one value to represent the δD values of input surface water is not realistic. Smith et al. (1992) considers waters more than 10‰ lighter than present winter precipitation to be Pleistocene waters that were recharged when temperatures were 4°C to 8°C cooler than today. The following calculations were made considering the range of values from \( \delta D = -120‰ \) (current value, treated as a maximum) to \( \delta D = -157.5‰ \).

Using the simple model of inflow=evaporation, δD values displayed in Table 8 are generated.

Table 8: Deuterium value of lake water when lake is modeled as a terminal lake, under various input conditions.

<table>
<thead>
<tr>
<th>( \delta D_{\text{inflow}} )</th>
<th>( R_i )</th>
<th>( R_L )</th>
<th>( \delta D_{\text{lake}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>-157.5‰</td>
<td>0.8425</td>
<td>0.901</td>
<td>-99‰</td>
</tr>
<tr>
<td>-145‰</td>
<td>0.855</td>
<td>0.914</td>
<td>-86‰</td>
</tr>
<tr>
<td>-130‰</td>
<td>0.870</td>
<td>0.930</td>
<td>-69.5‰</td>
</tr>
<tr>
<td>-120‰</td>
<td>0.880</td>
<td>0.941</td>
<td>-58.8‰</td>
</tr>
</tbody>
</table>

With no outflow, the Cl will increase until halite saturation occurs, so it is possible to attain any Cl value until this point is reached. From these calculations, only δD values equal to or
heavier than \(-130\%\) could evaporate to a steady state value \((-69.5\%\) \() close to the \(\delta D\) values in the 10 m groundwater. The conclusion to be drawn from these calculations is that recharge water to the lake in which \(\delta D\) was lighter than \(-130\%\) is unlikely.

A second model considers the effect of outflow, but still neglects atmospheric isotopic exchange (Friedman, written communication, 1994):

\[
R_L = \frac{R_I}{\alpha - \alpha \phi + \phi}.
\]

(13)

where \(\phi\) is the percent of inflow escaping as outflow. The salinity attained is also related to the percent of inflow escaping as outflow:

\[
S_L = \frac{S_I}{\phi}.
\]

(14)

where \(S\) is the salinity of the lake, and \(S_I\) is the salinity of the input (Owens River). In order for the salinity (Cl) to reach 372 mmol/L, from a starting Cl concentration of 0.395 mmol/L, outflow cannot exceed 0.1 percent of inflow. Such a small percentage of outflow does not alter the calculated final values, as presented in Table 9.

Table 9: Deuterium value of lake water if the lake is modeled as an overflowing lake with 0.1 percent of inflow escaping as outflow.

<table>
<thead>
<tr>
<th>(\phi)</th>
<th>(\delta D_{\text{inflow}})</th>
<th>(R_I)</th>
<th>(R_{L})</th>
<th>(\delta D_{\text{lake}})</th>
</tr>
</thead>
<tbody>
<tr>
<td>.001</td>
<td>-157.5%</td>
<td>0.842</td>
<td>0.901</td>
<td>-99%</td>
</tr>
<tr>
<td>.001</td>
<td>-145%</td>
<td>0.855</td>
<td>0.914</td>
<td>-86%</td>
</tr>
<tr>
<td>.001</td>
<td>-130%</td>
<td>0.870</td>
<td>0.930</td>
<td>-69.5%</td>
</tr>
<tr>
<td>.001</td>
<td>-120%</td>
<td>0.880</td>
<td>0.941</td>
<td>-58.8%</td>
</tr>
</tbody>
</table>

A different approach (Gat and Levy, 1972) is to consider the effects of isotopic exchange with the atmosphere. The following expression represents the steady state isotopic composition of a surface water body with inflow:
\[
\delta_{ss} = \frac{\varepsilon + \delta_a + \frac{I}{E} \frac{1-h}{h} \delta_i}{1 + \frac{1-h}{h} \frac{I}{E}}.
\]

where \(\delta_{ss}\) is the steady state isotopic composition of the lake, \(\varepsilon\) is the fractionation factor, \(E\) is the evaporation rate, \(I\) is the amount of inflow to the lake, \(\delta_a\) is the isotopic composition of atmospheric water vapor, \(h\) is the humidity, and \(\delta_i\) is the isotopic composition of the input water.

The steady state isotopic composition of the lake is calculated for several scenarios. By varying \(I/E\), the lake can be adjusted from a terminal lake, where all input is evaporated (\(I/E = 1\)), to a lake with increasing amounts of outflow. The isotopic composition of the inflow is taken to be \(-120 \%e\) (Owens River). \(\delta_a\) is the isotopic composition of atmospheric water vapor in equilibrium with a surface water value of \(-120 \%e\) (which is \(-180 \%e\)) at a temperature of 25°C. Humidity is varied from 10\% to 80\%, as is outflow. In no scenario did the isotopic composition increase above \(-120 \%e\). In most cases, the steady state isotopic composition would be significantly less than \(-120 \%e\) (lighter isotopically) because of feedback from the atmospheric water vapor, which is very light. Therefore groundwater now at a depth of 10 m could not have undergone a steady state evaporation with isotopic feedback from atmospheric water vapor if continual input from the Owens River is assumed. The neglect of exchange with atmospheric water vapor may be justified (as in Friedman’s model, written communication, 1994) if humidities are low enough such that contribution from atmospheric water vapor is insignificant. Removal of water vapor by winds for example, may prevent the attainment of a steady state equilibrium between the atmosphere and the lake water. Both of these scenarios, low humidity and strong winds, are likely scenarios for the Owens Lake area, even if cooler Pleistocene temperatures are assumed.

The proposed model to explain the groundwater composition at 10 m of depth, therefore, is as follows. The original pore water in 10 m sediments was fresh. Approximately 8,900 years
ago, the lake became terminal and solute concentrations began to increase. From that time onward, solutes continued to diffuse downward because of a continual supply of high salinity water near the surface. Evaporation of the lake water with continual supply of input from the Owens River proceeded from 8,900 years ago when outflow ceased to be significant. Deuterium which had increased to a steady state value around $-69.5\%$ diffused downward also. The conclusion, then, is that groundwater at 10 m is representative of originally fresh lake water which has been altered by solute and heavy isotope diffusion over the last 8,900 years. Groundwater at 10 m of depth, however, has not been affected by the lake's most recent desiccation.

**Summary of shallow system**

One process alone cannot be invoked to explain the shallow groundwater chemistry. Evaporation is certainly a major mechanism of ionic enrichment in the shallow groundwater, but must be accompanied by other processes in order to explain deviation from ideal behavior.

Dissolution of Na–CO$_3$ salts, Na–SO$_4$ salts, and halite are important, based on the correlation of Na concentrations with SO$_4$, CO$_3$, and Cl concentrations. The average proportion of Na compared to HCO$_3$ in the surface salt crust agrees with dissolved proportions. There is less agreement of precipitated and dissolved proportions of Na to Cl and Na to SO$_4$, probably reflecting the irregular distribution of these salts compared to that of the more ubiquitous Na–CO$_3$ salts. Sodium to CO$_3$ and Na to SO$_4$ ratios are higher in current groundwater than in an analysis of the 1866 lake water, indicating that the current groundwater has less CO$_3$ and SO$_4$ in comparison to Na than the 1866 lake. Removal of Na–CO$_3$ and Na–SO$_4$ salts by wind and mining has apparently affected the shallow groundwater geochemistry.
Distribution of solutes in the groundwater is affected by the presence of precipitated salts in the sediments available for dissolution, variation in sedimentologic and mineralogic characteristics of the subsurface, and convective groundwater mixing. Peaks in solute concentrations followed by decrease in solute concentrations in the direction of groundwater flow within 1.5 km of the lake bed margin may be attributed to these processes. Density instability that promotes convective groundwater mixing is brought about by progressive evaporation as groundwater flows from the lake bed margins toward the center of the lake bed, and by high hydraulic conductivity sandy sediments near the lake margins. Decrease in the magnitude of the vertical gradients between 3 m and 1 m groundwater indicates an increased permeability in the areas where the solute concentrations demonstrate greatest variation. This area, in the central portion within 0.35 km to 1.5 km of the eastern lake bed margin may be an appropriate location to establish vegetation. A localized decrease in solute concentration in the shallow groundwater may occur in other areas (such as the northeast portion of the lake bed) but definition of this pattern requires more data points (at both 1 m and 3 m) than are currently available.

Two types of groundwater are distinguished based on the $\delta$D and Cl concentrations. Samples that are enriched in $\delta$D, but relatively depleted in Cl (Type A) may be explained by two mechanisms: convective groundwater mixing and salt morphology changes. Groundwater mixing promoted by density instability may contribute to geochemical variation observed in shallow groundwater near the lake margins by mixing evaporated concentrated brines with lower salinity groundwater. Change in salt morphology, with subsequent release of water from the salt structure, may also influence the chemistry of groundwater affected by temperature variations and generate low Cl, high $\delta$D groundwater. Variations due to this mechanism are more likely at the beginning or end of seasons, when average temperatures are changing.
Samples that are enriched in Cl and relatively depleted in δD (Type B) may also be generated by convective mixing, but the proportion of evaporated brines that are mixed are less than that required to generate a Type A groundwater. In order to generate some of the marginal shallow groundwater with the most enriched Cl concentrations and the most depleted δD values (compared to the other shallow groundwater sites), desiccation or near-desiccation of the lake in these areas would be required. The Cl concentrations observed in excess of 1000 mmol/L could not be generated by groundwater mixing while still maintaining the δD concentrations below −100 ‰. The lake water in marginal areas of the lake would need to evaporate almost completely in order to concentrate the solutes while eliminating the enriched isotopic water from the system.

Composition of groundwater from the 10 m piezometers cannot be explained by any of the above mechanisms. The water is saline (Cl concentration is 372 mmol/L), but not as concentrated as other samples of shallow groundwater with enriched δD concentrations (−69 ‰). Connate groundwater in sediments at 10 m of depth was incorporated approximately 15,000 years ago, when Owens Lake was overflowing and lake water was fresh. Both evaporation of the terminal lake water (after 8,900 years before present) and diffusion of solutes and heavy isotopes from the partial desiccation of the lake 5,000 years ago can account for the solute and isotopic composition currently observed in the 10 m groundwater.

Most of the processes determined to affect the shallow groundwater operate exclusively in the shallow system. However, some conclusions from observations of the shallow system have implications for the deeper groundwater system. The shallow groundwater record indicates that in the last 100+ years, salts have been lost from the system (by wind). There is no record of total lake desiccation prior to the 5,000 year event, but the implication is that salts precipitated in drier periods (partial desiccation) would not necessarily be preserved in the sedimentological or geochemical record.
Second, the lack of high correlation between theoretically conservative constituents (Cl, D) suggests that neither is an absolutely reliable indicator of groundwater mixing or evaporation. Because enriched δD can be removed from the system (by total evaporation), δD does not always provide an accurate record of past processes.

Finally, an understanding of the chemical processes that control salt crust composition and distribution must be addressed if there is any hope for long-term control of dust production. This initial study of the shallow system processes may aid in achieving that goal.
CHAPTER V. DEEP GROUNDWATER SYSTEM

Vertical and lateral aquifer connections in the deep groundwater system are not well understood at Owens Dry Lake, mainly due to widely spaced and limited sampling locations and a lack of information about well construction. The goal of this chapter is to explain the different hydrologic, geologic, or geochemical processes that may determine the composition of the deep groundwater in different parts of the lake, and to identify connections between deep aquifers. The development of a conceptual groundwater flow model and a water balance will be aided by an understanding of the relative importance of various hydrologic processes such as vertical groundwater movement or groundwater flow out of the basin. The sampling locations referred to in the following chapter are displayed in Figure 33. The samples considered in the deep groundwater system are from wells greater than 30 m in depth, margin springs (sampled by Lopes, 1988), and spring mounds (described later in this chapter). Unless otherwise specified, groundwater samples mentioned in the text and figures were collected by this author. A few sites previously sampled by Lopes (1988) were re-sampled for additional analyses. Appendix C summarizes the dates, collectors, and chemical analyses of all the deep groundwater samples mentioned in this chapter. When the site was sampled numerous times, generally the first sampling date is listed.

Hypothesis

Deep groundwater should not be treated as one homogeneous system, or a system in which groundwater in different locations has necessarily been subjected to similar physical processes over similar time periods. Evaporation has been much less of a universal controlling mechanism on the deep groundwater than on the shallow groundwater, probably due to variation in the physiographic conditions of the lake (terminal versus overflowing). Isotopic variation of groundwater across the lake bed may reflect distinct periods of groundwater recharge, groundwater mixing of fresher water with saline pore water in lake sediments, or distinct source areas. Near the lake margins, groundwater flow may be
Figure 33: Deep groundwater sample locations. Lopes (1988) samples indicated by "L." Distances to sample locations are air distance, not driving distance.
fault–controlled, and vertical flow may have occurred between aquifers interlayered with lake sediments. Vertically downward groundwater movement may affirm the potential for groundwater flow out of the basin to the south or the east.

This chapter evaluates the various aspects of the hypothesis in the order that they are described above. The heterogeneous nature of the deep groundwater system is verified by examining hydraulic data, $\delta^{18}$O and $\delta^{D}$ isotopes, $^{14}$C ages, and geochemical indicators of different source areas. The effects of evaporation are assessed by examining the $\delta^{18}$O and $\delta^{D}$ data in conjunction with concentrations of assumed conservative ions. Different ages of groundwater and groundwater mixing are investigated with evidence from $^{14}$C and $\delta^{13}$C analyses. Different areas of recharge are suggested by patterns of $^{87}$Sr/$^{86}$Sr and As distribution. The hypothesis that groundwater discharge may be controlled by the presence of faulting is evaluated by examining the northeastern part of the lake, where spring mounds appear to be aligned on fault traces. Vertical groundwater movement is suggested by similarities of, and reasonable geochemical interactions between, shallower and deeper aquifers. Vertical groundwater flow is also substantiated with hydraulic data from three locations. The potential for basin outflow is evaluated by examining the chemistry of wells and springs south and east of the Owens Dry Lake basin.

**Heterogeneity of deep groundwater**

**Hydraulic measurements**

Potentiometric measurements across Owens Dry Lake (Figure 34) suggest multiple aquifer systems. Flow paths are not always evident between head measurement locations, and measurements are often only approximations of the system. Measurements were compiled from field work and data provided by various sources. Field elevation measurements were based on Thommen® altimeters accurate to ±10 m with a 1 m resolution. The actual elevations of a well is not as important as the elevation differences between the wells. Two
Figure 34: Potentiometric elevations in Owens Dry Lake wells rounded to nearest meter above sea level. Well elevations were determined with a Thommen® altimeter (accurate to ±10 m, with 1 m resolution), from topographic maps, or from survey data. The derivation of the piezometric elevation, the reliability of the data, and the date of water table measurement are presented in Table 10.
altimeters were used, one taken to the field, and a second left connected to a computer which recorded barometric fluctuations throughout the day. These fluctuations were then considered when interpreting the field measurements. Elevation data were often estimated from topographic maps. Table 10 summarizes the hydraulic data available, data sources, and dates of collection. Some potential flow paths are illustrated in Figure 34. Based on the estimated hydraulic data flow paths may exist between the Aqueduct Well and CW-MW2S and/or CW-MW2D, from the River Wells southward toward OL-92, from locations on the upper to the lower eastern alluvial fans, from the lake bed margin near Mill Site to the Sulfate Well, and from the the area south of Owens Dry Lake to Owens Dry Lake. Multiple aquifer systems are suggested by similar and possibly higher (given the error of the elevation measurements) potentiometric heads in wells near the lake margin, such as Keeler-Swansea and CW-PW1, compared to heads in wells on the alluvial fans.

In general, groundwater flow comes from the north with some inflow components from both the east and west. Water levels are lowest in the vicinity of the Cottonwood alluvial fan, suggesting higher permeability sediments in this area. The more gentle topography on the eastern side of the lake bed allows confining units to be more extensive and to exert more control on hydraulic heads than do sediments along the western shore.

<table>
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<th>Well</th>
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<th>date of piezometric measurement</th>
<th>screened interval (m above sea level)</th>
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<tr>
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<tr>
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Oxygen and deuterium isotopes

The stable isotopic values from deep groundwater samples are plotted in Figure 35 and exhibit a wide range in values. Deuterium values, for example, range from -130 ‰ to -90 ‰. The 90 ‰ sample is from a spring sampled by Lopes (1988) which he states may have been subjected to evaporation. The remaining samples still vary over a wide range.

Although the isotopic values plot close to an ideal evaporation line (slope = 5.4, r² = 0.94), no other geochemical observations (¹⁴C, ⁸⁷Sr/⁸⁶Sr) indicate that samples represent one homogeneous reservoir of groundwater.
Figure 35: $\delta^{18}$O and $\delta$D values from deep groundwater sample locations.
Some of the lighter isotopic values can be generated by the steady state solution (equation 15) for the isotopic evolution of a lake with inflow and outflow (Gat and Levy, 1972), using an initial δD value of −120 ‰ and an atmospheric δD value equal to −143 ‰ (in equilibrium with average modern precipitation at 25 °C). However, δD values more enriched than the input water (−120 ‰), which are found in OL-92, Sulfate well, and the west side wells, cannot be generated from this steady state model. If the input water (Owens River) were isotopically lighter than observed today (as may have been the case during the Pleistocene), this scenario would not produce any of the observed isotopic values. Friedman's model (1994, written communication) that does not include the effects of isotopic exchange with atmospheric water vapor generates isotopic values observed in the groundwater, but Cl concentrations observed in some of the wells require that outflow be limited to less than 0.1 percent of inflow, which was not the case over much of Owens Lake history (Benson and Bischoff, 1993).

To assess the influence of evaporation on deep groundwater chemistry, both Cl and Li concentrations were plotted versus δD. Figures 36 and 37 indicate no discernable correlation between the concentrations of assumed conservative ions (and theoretically the degree of evaporation) and enrichment with δD. Neither is there a correlation between Li and Cl, as illustrated in Figure 38. This does not necessarily eliminate evaporation as an important controlling mechanism, but may indicate that the deep groundwater originates in several source areas and the chemistry is complicated by other hydrologic or geologic factors.

In general, the isotopically lightest groundwater comes from wells and springs in the northern and eastern portions of the lake (−130 ‰ to −122 ‰). More δD enriched groundwater is found in wells and springs on the west side of the lake (−114 ‰ − 104 ‰). There are numerous explanations for the diversity in isotopic values. Groundwater from different aquifers may have been recharged under different climatic conditions. Deuterium values of water
Figures 36 and 37: Chloride (36) and Li (37) plotted with δD from groundwater samples (springs and wells). Correlation is not evident in either case, and neither Li nor Cl may be treated as a conservative ion. Figure 38: Li plotted with Cl, demonstrating no evident correlation.
which precipitated during the Pleistocene are generally at least 10% lighter than modern surface water (Smith et al., 1992). Another explanation for isotopic variation is groundwater mixing at lake margins between fresh mountain recharge and older groundwater incorporated in lake sediments. The degree of mixing, however, may not correlate as expected with the amount of recharge supplied by the mountain ranges. For example, the more enriched δD values on the west (Sierra) side of the lake could indicate more mixing with lake sediment groundwater despite a higher recharge rate than exists on the east side of the lake. Higher recharge on the west side of the basin may also supply younger and isotopically heavier water than that recharged on the east side. Notable differences between east and west side groundwater that may be attributed to different time periods since recharge are discussed in greater detail below.

Carbon-14 analysis

Carbon-14 data are consulted to further explore the influence of different periods of recharge, the degree of mixing between recharge and saline pore water in lake sediments, and various geochemical reactions on groundwater chemistry. Carbon-14 analysis was performed on samples of groundwater from four wells around the margin of the lake: Cottonwood Production Well (CW-PW1) and the Aqueduct Well on the west side of the lake, River Well Lower in the north part of the lake, and Keeler-Swansea Lower on the eastern margin of the lake bed. These wells were chosen for 14C analysis in order to compare waters of different δ18O and δD composition that are potentially from different source areas or were recharged at different times. With the exception of the Aqueduct well, the uncorrected data from analysis of these waters (percent modern carbon – pmc) indicates relatively old water (>20,000 years). In contrast travel times between alluvial fans and the center of the valley calculated from Hollett et al. (1991) are on the order of hundreds to thousands of years. These calculations will be discussed in more detail later in the text. The two possibilities to consider are that the groundwater is truly as old as the uncorrected data
indicate, or that geochemical and biogeochemical processes have altered the carbon (and hence the $^{14}$C and $\delta^{13}$C) composition.

Correction of the apparent $^{14}$C ages is conducted invoking only assumptions that can be justified based on the known geology and hydrology. The process most often used to explain groundwater depleted in $^{14}$C is dissolution of a carbonate mineral in which all the $^{14}$C has decayed and left the rock with a $^{14}$C signature of 0 pmc. Many of the models that adjust the initial $^{14}$C activity to account for dilution with $^{14}$C-depleted ("dead") carbon assume equilibrium of the groundwater with carbonate minerals (Ingerson and Pearson, 1964; Mook, 1972; Fontes and Garnier, 1979). On the eastern margin of Owens Dry Lake, marine formations in the Inyo Mountains that include calcites and dolomites are likely sources of $\text{CO}_3$ to the groundwater. To the north and west, however, CO$_3$ sources are probably limited to pedogenic CO$_3$ or lacustrine CO$_3$ (Na–CO$_3$ salts and CaCO$_3$) in or near the lake sediments. The geochemical/biogeochemical reactions to which the groundwater has been subjected are often reflected in the $\delta^{13}$C value of the groundwater. Carbon–13 is a stable isotope of carbon and therefore records only mixing or geochemical reactions.

When $^{14}$C data from two wells along a proposed flow path are available, NETPATH may be used to calculate a corrected $^{14}$C age. Several geochemical scenarios that may result in the observed $\delta^{13}$C composition are modeled. With information or logical guesses about the $\delta^{13}$C and $^{14}$C values of $\text{CO}_3$ and soil–gas CO$_2$ in the system, NETPATH calculates the final $\delta^{13}$C composition. Carbon–13 may also be used as a constraint in the mass balance model that NETPATH calculates, which assures that the mass balance reaction results in the observed $\delta^{13}$C composition in the down-gradient water. Based on reactions that NETPATH invokes to explain the down-gradient chemistry, a final $^{14}$C composition is calculated not including decay. The difference between this non-decay value compared to the observed value is the adjusted $^{14}$C age of the groundwater. The adjusted $^{14}$C age may be significantly
different (up to an order of magnitude) than the apparent $^{14}$C age and should not be viewed as more than an approximation of the time period since the groundwater was isolated from the atmosphere.

**Western portion of lake basin**

Carbon–14 analysis was performed on groundwater samples from two wells on the west side of the lake, the Aqueduct Well and the Cottonwood Production well (CW–PW1). Figure 39 is a schematic cross section of the area of interest indicating lithology, well construction, and approximate water table elevations. Hydrologic studies conducted in the Cottonwood Springs area indicate that at the lake margin, two aquifer systems exist. Based on lithologic logs and water levels in monitoring wells and piezometers surrounding CW–PW1 (described in detail in a following section), the subsurface is divided into two aquifers separated by a one-meter thick sandy clay layer (MHA Environmental Consulting, Inc., 1994; Luhdorff and Scalmanini Consulting Engineers, 1993a and 1993b; CH2M Hill, 1991). The production well CW–PW1 is located in the deeper aquifer. Water levels in the two aquifers are different, as are the gradient magnitudes and directions. In the shallower aquifer groundwater flows to the east, and in the deeper aquifer groundwater flows from north to south. These gradient magnitudes and directions were determined from water levels in wells and piezometers within 500 m of CW–PW1.

Former studies do not indicate, however, the hydraulic relationship between the groundwater on the alluvial fan, 2.5 km to the west, and the shallow and deep aquifer systems at the lake margin. Since accurate surveyed elevation data is not available for the Aqueduct Well, the elevation of the water table at this location (indicated in Figure 39) is an approximation based on information from a topographic map, and may vary by more than a meter from the elevation indicated in Figure 39. Water table elevation at CW–PW1 is based on surveyed data and is accurate. The hydraulic data do not confirm or refute that the
Figure 39: Schematic cross section of west side of lake from alluvial fan (Aqueduct Well) to lake margin (PW-1).
Aqueduct Well and CW–PW1 lie along the same flow path. The purpose of the following discussion is to elucidate the relationship between these two locations using $^{14}$C data. The initial discussion assumes that the two wells lie on a flow path which connects the recharge area on the alluvial fans and the discharge area at the margins of the lake sediments.

The uncorrected data indicate that the percent modern carbon (pmc) from the Aqueduct well is greater than 100 (114.6 pmc), and that from CW–PW1 is less than 0.5 (<0.5 pmc). Groundwater from the Aqueduct well has apparently been affected by input of $^{14}$C from bomb testing in the mid-twentieth century. The difference in $^{14}$C content between the two wells is informative, and with a few logical assumptions, the data may be used to give a more realistic picture of the groundwater travel time between the Aqueduct Well and CW–PW1.

None of the models available in NETPATH are used to adjust the initial $^{14}$C activity in the Aqueduct Well. The reasons for this choice are as follows. The $\delta^{13}$C signature of the Aqueduct well is $-14.6 \text{‰}$, which would represent an approximate 50:50 mixture of $\delta^{13}$C from soil zone CO$_2$ ($\delta^{13}$C = $-20 \text{‰}$) (Amundson and Donner, 1986) and atmospheric CO$_2$ ($\delta^{13}$C = $-9 \text{‰}$). The Aqueduct Well is in an unconfined alluvial fan aquifer in which both of these sources would be available. There is no indication that additional geochemical reactions have altered the isotopic carbon content in the Aqueduct Well groundwater. The second reason not to adjust the initial $^{14}$C composition is that the geochemical processes that will affect the $^{14}$C composition occur (as modeled) between the Aqueduct Well and CW–PW1 or (realistically) at CW–PW1 when that groundwater comes into contact with more saline groundwater and lacustrine CO$_3$ precipitates. The desired result from NETPATH is a scenario of specified geochemical reactions, justified based on the geology, which produce the observed isotopic compositions in the assumed down-gradient well (CW–PW1).
As mentioned, the most reasonable reactions that may alter the carbon composition and isotopic signature between the Aqueduct well and CW-PW1 include the dissolution of CO$_3$ (predominantly Na–CO$_3$ salts rather than CaCO$_3$, based on the differences in major ion chemistry between the two wells), or mixing with older, more saline groundwater (in the lake sediments) in which the $^{14}$C (in HCO$_3$) has entirely decayed. The possibility of mixing with a more saline water is also suggested by the increase in Cl between the two wells. This increase may also be attributed to the dissolution of halite, but neither case (mixing or dissolution) affects the isotopic calculations. A halite phase or Na–Cl groundwater should be included in the model scenarios to avoid NETPATH's use of Na–CO$_3$ salt to account for all of the Na increase between the two wells. This would affect the isotopic composition because of the excess CO$_3$ ions this scenario would generate.

The results of some modeled scenarios are presented in Table 11. The main differences in the model scenarios arise with the choice of including dissolution of organic carbon or dissolution of soil zone CO$_2$ gas. One of these choices is necessary in order to bring the calculated $\delta^{13}$C composition down to the observed value (both organic carbon and soil zone CO$_2$ have depleted $\delta^{13}$C values of $-25\%e$ and $-20\%e$, respectively). The assumed $^{14}$C values, however, are different. Soil zone CO$_2$ is assumed modern (100 pmc) because CO$_2$ is being produced currently and is in contact with the atmosphere. Organic carbon is assumed dead (0 pmc) based on $^{14}$C dates of OL–92 sediments (Bischoff et al., 1993). As expected from this assumption, incorporation of soil zone CO$_2$ (modern), as in scenario 1, results in a higher final calculated pmc and hence much older age than incorporation of the same amount of organic carbon, as in scenarios 3 and 5.

The model is also quite sensitive to the choice of the $\delta^{13}$C for the Na–CO$_3$ mineral, a value for which there is significant uncertainty. A value of $+4.0\%e$ is input into the model based on the $\delta^{13}$C analyses of OL–92 sediments (Benson and Bischoff, 1993) at depths that correspond to
the well locations being modeled. Carbon-13 may be included as a constraint in the model, in which case NETPATH performs a mass balance on $\delta^{13}$C and the final calculated value matches the observed. In scenarios 1 and 3, $\delta^{13}$C is included as a constraint. In scenarios 2, 4, and 5, $\delta^{13}$C is not included as a constraint in order to test the validity of the geochemical reactions in producing the observed $\delta^{13}$C value in the down-gradient well. In scenarios 2 and 4, all the Na comes from dissolving Na-CO$_3$ salt and halite, and this affects the $\delta^{13}$C composition by making it too enriched (2.7%o and 3.9%o versus observed value of -8.6%o). In scenario 5 some of the Na is allowed to come from dissolution of a Na-SO$_4$ salt, requiring less isotopically heavy Na-CO$_3$ salt. Hence, the calculated $\delta^{13}$C value (-3.7%o) is closer to that observed (-8.6%o). This value can be brought much closer to the observed value of -8.6%o by reducing the $\delta^{13}$C value in the Na-CO$_3$ salt to 0.0%o (final calculated $\delta^{13}$C of -6.5%o). There is an acknowledged uncertainty in the $\delta^{13}$C value in the Na-CO$_3$ salt, because the phase of the Na-CO$_3$ has not been specified, i.e. Na and CO$_3$ ions may be supplied by dissolving a Na-CO$_3$ salt or by mixing with a saline groundwater. The $\delta^{13}$C in the HCO$_3$ in each case could be different due to fractionation that occurs during carbonate mineral precipitation (Dienes et al., 1974). Note that despite the reactions modeled to balance the $\delta^{13}$C concentrations, the percent modern carbon is unaffected or only slightly affected. With the exception of scenario 1, explained above, the $^{14}$C age of the groundwater remains relatively constant. In other words, the final calculated percent modern carbon is not sensitive to slight changes in the mixing scenario.
Table 11: NETPATH models which are realistic based on given mineralogy to explain ionic and $\delta^{13}C$ compositions of CW–PW1. All concentrations in molality (mmol/kg H$_2$O). Negative values indicate precipitation or degassing. In scenario 5, if the $\delta^{13}C$ value for NaHCO$_3$ is 0.0% instead of 4.0%, the calculated $\delta^{13}C$ value is -6.5%, much closer to the observed value of -8.6%.

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<td>7.69</td>
<td>5.49</td>
<td>7.69</td>
</tr>
</tbody>
</table>

Of the four scenarios for incorporating a phase depleted in $\delta^{13}C$ (scenarios 1, 3, 4, and 5), those which invoke the dissolution of organic carbon are more likely than those which incorporate soil zone CO$_2$ because the pCO$_2$ decreases between the Aqueduct well and CW–PW1. Also the addition of CO$_2$ results in a down-gradient water with a calculated pH which is much lower (6.6) than is observed at CW–PW1 (7.7). This scenario is confirmed by entering the NETPATH reactions into PHRPITZ, using the Aqueduct Well as an initial groundwater. What NETPATH models as the "dissolution" of organic carbon is more likely the respiration of organisms which utilize organic matter and produce HCO$_3^-$ with $\delta^{13}C = -25$ %o. The general details of the complete respiration process are presented below:

'organic carbon' + 138O$_2$ $\rightarrow$ 106HCO$_3^-$ + 16NO$_3^-$ + HPO$_4^{2-}$ + 16H$_2$O + 124H$^+$ (Morel and Hering, 1993, p. 205).

If the amount of HCO$_3^-$ equal to the amount of organic carbon that NETPATH dissolves is added as a reaction in PHRPITZ, the final pH and alkalinity closely agree with the observed values in CW–PW1. The comparison between the values calculated in PHRPITZ (based on NETPATH mass and isotope balance) and observed values is presented in Table 12.
Table 12: Comparison of hypothetical groundwater that has undergone a series of reactions in PHRQPITZ, to the concentrations observed at CW-PW1. All concentrations as molality.

<table>
<thead>
<tr>
<th>components</th>
<th>calculated in PHRQPITZ (scenario 3)</th>
<th>observed in CW-PW1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na</td>
<td>$5.00 \times 10^{-2}$</td>
<td>$5.23 \times 10^{-2}$</td>
</tr>
<tr>
<td>Cl</td>
<td>$3.11 \times 10^{-2}$</td>
<td>$3.11 \times 10^{-2}$</td>
</tr>
<tr>
<td>Ca</td>
<td>$6.02 \times 10^{-2}$</td>
<td>$6.01 \times 10^{-2}$</td>
</tr>
<tr>
<td>tot Alkalinity</td>
<td>$2.03 \times 10^{-2}$</td>
<td>$1.97 \times 10^{-2}$</td>
</tr>
<tr>
<td>pH</td>
<td>$7.78$</td>
<td>$7.70$</td>
</tr>
</tbody>
</table>

Based on the pmc values in Table 11 for the reactions proven reasonable, the adjusted $^{14}$C age of the groundwater at CW-PW1 is no younger than 20,000 years (pmc has been reduced from 7.69 to less than 0.5) if a flow path between the Aqueduct Well and CW-PW1 is assumed.

A travel time of 20,000 years between the Aqueduct well on the alluvial fan, and CW-PW1 on the lake margin (a distance of two kilometers) is not realistic if horizontal flow is assumed. Hollett et al. (1991) have calculated hydraulic conductivity values for two aquifers in the Owens Valley (referred to as Units 1 and 3). Although these aquifers are continuous with alluvial fans on the margins of the valley, there are no hydraulic conductivity data for the alluvial fans. They speculate that due to poor sorting, the hydraulic conductivity of the alluvial aquifers will be less than the hydraulic conductivity of the fluvial and lacustrine aquifers in the center of the valley.

A range of travel times is calculated assuming horizontal flow from the Aqueduct Well to CW-PW1 using the lowest hydraulic conductivity value from Hollet et al. (1991) for Units 1 and 3 ($K = 3.658 \text{ m/day}$).

$$\text{traveltime} = \frac{\text{distance}}{v},$$

$$v = -\frac{K \Delta h}{n \frac{\partial l}{\partial t}} \quad \text{(Freeze and Cherry, 1979, p.71)},$$

where $v$ is the average linear velocity, $K$ is the hydraulic conductivity, $n$ is the porosity, and
\( \frac{\delta h}{\delta l} \) is the hydraulic gradient between the wells. Since a surveyed elevation is not available for the Aqueduct Well, elevation range from a topographic map is used in the calculations. The gradient between the Aqueduct Well and CW-PW1 may range between zero and 2.0 m over a distance of 2,500 m. This range of gradients with a range in porosity values (0.20 - 0.35 from Fetter, p. 68) for mixed sand and gravel is used with the hydraulic conductivity value to calculate a range of average linear velocities and travel times between the Aqueduct Well and CW-PW1. Calculated travel times range between 450 and 3,780 years. The \(^{14}\text{C} \) age, then, is one to two orders of magnitude greater than calculated travel times.

There are two possibilities to consider in interpreting this \(^{14}\text{C} \) data in light of the hydraulic data. A flow path may exist between the Aqueduct Well and CW-PW1, and the gradient between the two wells may be so slight (which the water table elevation data do not refute) that the corrected \(^{14}\text{C} \) age at CW-PW1 is approximately correct. If the head difference between the two wells were only 0.1 m, with a porosity of 0.35, the travel time would be 16,000 years. A head difference of 0.05 m results in a 32,000 year travel time. However, the possibility of a low gradient between the two wells impeding rapid groundwater flow is unlikely given the presence of relatively extensive springs discharging at the lake margin.

A more likely possibility is that the two wells are not connected by a horizontal flow path. A hydraulic connection between the two wells by a curved, deeper flow path is also unlikely as a travel distance greater than 15 km is required in order to increase the age of Aqueduct Well groundwater to that at CW-PW1. CW-PW1 could be recharged at a higher elevation in the Sierra Nevada Mountains, and lie on a completely different flow path than the Aqueduct Well. Shallower groundwater (penetrated by the Aqueduct Well) may be hydrodynamically separated from deeper groundwater and discharge in the lake margin springs. Deeper infiltrated groundwater from alluvial fans to the west does not discharge at the lake margins, but may diverge toward the south upon approaching the lake deposits. Near the lake margins, flow from the west may mix with flow from the north, as well as saline pore waters in lake
sediments. The low $^{14}$C content of the deep groundwater indicates that a combination of converging flow directions, low gradients, and decreasing sediment permeability may result in relatively slow moving groundwater at depth.

**Northern portion of lake basin**

Another area of the lake where $^{14}$C analyses are consulted is the northern portion, using data from the lower River Well and wells near Independence, California (Lyons *et al.*, in press). Approximate locations of the wells discussed are presented in Figure 40. The δD values in the River well aquifers (−126 ‰) are similar to values from some wells located to the north in the Owens Valley (−122 ‰ to −127 ‰) (Gleason *et al.*, 1992). Similarity in isotopic composition as well as an assumed north to south flow path in the center of the valley (Hollett *et al.*, 1991) imply a connection between wells to the north in the Owens Valley and the River Wells. Hollett *et al.* (1991) estimates a horizontal hydraulic conductivity in what they term the Unit 3 aquifer between 3.5 and 46 m/day. Their Unit 3 aquifer is a confined system of alluvial, fluvial and lacustrine deposits that occupy depths similar to those bracketed by the lower River well (148 m - 170 m bgs). The hydraulic conductivity of the lower River Well aquifer is approximately 34 m/day (Jacobson *et al.*, 1990).

Using these hydraulic conductivities, estimating the hydraulic gradient from the potentiometric map of Hollett *et al.* (1991), and assuming a porosity of 0.3, a range of average linear velocities is calculated and applied over the 29 kilometers between Independence and Owens Dry Lake. Calculated travel times range between 280 years and 3,515 years. The $^{14}$C ages in the two wells near Independence (Lyons *et al.*, in press) are 439 and 3,460 years. Addition of the age of the initial water (Independence wells) to the travel time between the initial water and the lower River Well gives a maximum age of 6,975 years for the final water.
Figure 40: Location of Owens Valley wells, LADWP 368, 357, and 57, in relation to the lower River Well (RWL) at Owens Dry Lake.

Analysis of lower River well groundwater reveals 2.9 pmc and $\delta^{13}C = -4.9\%$. Although the apparent pmc indicates relatively old groundwater, the heavy $\delta^{13}C$ value represents significant geochemical reaction in addition to radioactive decay. A NETPATH simulation is created with the up-gradient well being LADWP well 368, which Lyons et al. (in press) determined to have a $^{14}C$ age of 3,460 years (62.6 pmc, adjusted by Fontes and Garnier, 1979), and $\delta^{13}C = -16.7\%$. Since a complete chemical analysis was not available for this well, chemical compositions from two nearby wells (LADWP 357 and 57) are obtained from Hollet et al. (1991) and averaged. The down-gradient well in the simulation is the lower River well. Geochemical reactions similar to those modeled on the west side of the lake are
invoked in order to explain the difference in major ion chemistry, as well as to achieve $\delta^{13}$C isotopic mass balance. The main difference in the geochemical reactions modeled in this case versus the scenario modeled on the west side of the lake, is the inclusion of dolomite, instead of calcite, to supply the necessary acquisition of cations and alkalinity. This choice is considered reasonable based on proximity to the Inyo mountains (not the case in the west side scenario), which contain significant amounts of dolomite. Dissolution of organic carbon is again included in the model, as is dissolution of Na–CO$_3$ salt and halite (or perhaps mixing with a Na–Cl–HCO$_3$ water) and the precipitation of calcite and a Na–SO$_4$ salt. The molar quantities required for mass balance calculated by NETPATH are listed in Table 13. PHRQPITZ is again used to calculate the effects of adding the reactions suggested by NETPATH to the initial well. A comparison between the PHRQPITZ calculations and the observed conditions at the down-gradient well (lower River Well) are displayed in Table 14.

Table 13: Amounts of various minerals that need to dissolve in order to generate the ionic and $\delta^{13}$C composition of the lower River Well groundwater, starting with groundwater composition from wells near Independence. All concentrations in molality (mmol/kg H$_2$O).

<table>
<thead>
<tr>
<th>Input</th>
<th>scenario 1</th>
</tr>
</thead>
<tbody>
<tr>
<td>NaHCO$_3$</td>
<td>5.24</td>
</tr>
<tr>
<td>CaMg(CO$_3$)$_2$</td>
<td>2.39</td>
</tr>
<tr>
<td>NaCl</td>
<td>2.30</td>
</tr>
<tr>
<td>CH$_2$O</td>
<td>2.43</td>
</tr>
<tr>
<td>CaCO$_3$ precipitation</td>
<td>−1.31</td>
</tr>
<tr>
<td>Na$_2$SO$_4$ precipitation</td>
<td>−1.70</td>
</tr>
<tr>
<td>$\delta^{13}$C calc.</td>
<td>−4.4</td>
</tr>
<tr>
<td>pmc</td>
<td>4.8</td>
</tr>
</tbody>
</table>
Table 14: Comparison of calculated concentrations from PHRQPITZ and observed values in the lower River Well groundwater (RWL). PHRQPITZ is used to simulate the effects of adding the reactions specified by NETPATH. All concentrations are molality. A range of pH values was obtained at the lower River Well.

<table>
<thead>
<tr>
<th>components</th>
<th>calculated in PHRQPITZ</th>
<th>observed in RWL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na</td>
<td>$4.83 \times 10^{-3}$</td>
<td>$4.83 \times 10^{-3}$</td>
</tr>
<tr>
<td>Cl</td>
<td>$2.55 \times 10^{-3}$</td>
<td>$2.55 \times 10^{-3}$</td>
</tr>
<tr>
<td>Ca</td>
<td>$1.53 \times 10^{-3}$</td>
<td>$1.53 \times 10^{-3}$</td>
</tr>
<tr>
<td>tot Alkalinity</td>
<td>$1.84 \times 10^{-2}$</td>
<td>$1.15 \times 10^{-2}$</td>
</tr>
<tr>
<td>pH</td>
<td>7.98</td>
<td>7.4 - 7.9</td>
</tr>
</tbody>
</table>

Using the molar amounts calculated by NETPATH, the calculated pmc in the River Well is approximately 4.8. Compared to the observed pmc (2.9 pmc), this indicates an adjusted age of 4,200 years older than the initial water. This adjusted age is similar to the maximum calculated travel time (from Hollet et al., 1991 K values) of 3,515 years between Independence and Owens Dry Lake. Since the $^{14}$C age of the groundwater from well 368 (Lyons et al., in press) is approximately 3,460 years, the groundwater at the site of the River well could be as young as 7,660 years (adding the age of initial water and the travel time). This age seems likely based on the reasonable geologic situations invoked, and agreement with hydraulic data. The maximum travel time between Independence and Owens Dry Lake assumes the lowest hydraulic conductivity value from Hollet et al. (1990) of 3.5 m/day. The hydraulic conductivity value from the lower River Well aquifer is higher (approximately 34 m/day), and this discrepancy may be attributed to aquifer heterogeneities. Since there is reasonable agreement between the $^{14}$C age of the groundwater and the maximum travel time calculated from Hollet et al. (1990), their data may be used to calculate recharge to the Owens Lake basin from the north.
Eastern portion lake basin

On the east side of the lake bed, groundwater from one well (Keeler–Swansea Lower) has been analyzed for $^{14}$C. There are no $^{14}$C data available for wells up-gradient from the Keeler–Swansea well, so a flow path scenario like those previously discussed cannot be created for this well. Since the isotopic carbon data from Keeler–Swansea groundwater cannot be attributed to geochemical changes from a known starting value, and because a reasonable model can be chosen based on the geology, the $\delta^{13}$C data ($-6.1\%$) is input into the Ingerson and Pearson model (1964) to adjust the initial quantity of modern carbon ($A_0$). This model assumes equilibrium with a carbonate mineral ($0$ ppm, $\delta^{13}$C = $0.0\%$), which is a logical assumption based on the carbonate formations in the Inyo Mountains, as well as the calcite saturation index in Keeler–Swansea groundwater (0.8). Under these assumptions and the Ingerson–Pearson model, the adjusted age of the Keeler–Swansea groundwater is approximately 22,500 years.

Assembling a picture based on the four $^{14}$C dates that are available, suggests that groundwater being recharged to the lake from the north may be attributed to horizontal (and perhaps vertical, discussed in later section) flow within alluvial and fluvial aquifers that occupy the central portion of the Owens Valley. Groundwater flow to the east and west portions of the lake (presumably from the alluvial fans), however, may not be as straightforward. Even after accounting for likely geochemical reactions that would falsely increase the apparent $^{14}$C age of the groundwater, these groundwaters appear to be at least 20,000 years old. Data suggest that a hydrodynamic separation of groundwater occurs with flow from alluvial fans toward the lake margin. Shallower, modern groundwater discharges in the margin springs, but deeper groundwater moves more slowly as low permeability lake sediments are approached.
This discrepancy in the age of groundwater in different locations around the lake substantiates the heterogeneous nature of the deep groundwater system. The waters have not undergone the same geochemical reactions, nor have they been subjected to geochemical reactions for the same amount of time.

**Other indicators of deep groundwater heterogeneity**

Influence from different periods of recharge is suggested by δ¹⁸O and δD variation in the deep groundwater, as well as different ¹⁴C ages between groundwater on the lake margin and in the northern portion of the lake. Variation of other geochemical indicators may reflect different recharge areas or flow paths.

For example, $^{87}\text{Sr}/^{86}\text{Sr}$ tends to be slightly lower in groundwater from the west side of the lake. Figure 41 displays the Sr ratios from the groundwater. The Sr ratios are plotted with K concentration because high K may correlate with higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios due to the geochemical similarity of K to Rb, which decays to $^{87}\text{Sr}$ (McNutt et al., 1987; Faure and Powell, 1972, p. 63). No obvious correlation is evident, but the plot indicates a slight separation in values between wells on the east side versus wells on the west side. Well CW-MW2D, the deep monitoring well at the Cottonwood site, is anomalous in that the Sr ratio does not resemble that from CW-PW1, completed in the same aquifer. The reason for this discrepancy is unexplained at this time, unless caused by contamination during sampling. Blum et al. (1994) determine that the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in Yosemite National Park, CA streams is determined by the weathering rates of minerals having different Sr isotope ratios. Biotite, for example, has a high $^{87}\text{Sr}/^{86}\text{Sr}$ value (0.862 – 0.878) and increased weathering of biotite results in a higher $^{87}\text{Sr}/^{86}\text{Sr}$ signature in the surface water. All of Blum et al. (1994) $^{87}\text{Sr}/^{86}\text{Sr}$ values in streams (0.7076 – 0.7088) as well as the Sr ratios in this study are well above the whole rock ratios for surrounding Sierra granites and granodiorites (avg. <0.707, Bateman et al., 1991).
Figure 41: Strontium isotopic ratios of deep groundwater locations sampled. Values from the west side of the lake (except CW-MW2D) are slightly less than those from the east side, indicated with the dashed line. OR = Owens River. The spring mounds are labeled SMA#, RWU and RWL are the upper and lower River Wells, respectively. KSI = Keeler Swansea Intermediate. HSpring = Horseshoe Spring. LL = Little Lake Spring.
Strontium ratios in groundwater rarely reflect the whole rock ratios of their recharge area because of incomplete rock–water interactions. The variation in the Sr ratios among the locations sampled in this study probably reflects the minerals present and different weathering rates. One would expect higher weathering rates on the west side of the lake (given more relief and higher recharge). Perhaps more rapid weathering leads to low Sr ratios as observed in west side groundwater. Strontium ratios in west side groundwater may be closer to the whole rock Sr ratios in granites than the Sr ratios in east side groundwater are to whole rock Sr ratios of marine carbonates.

Another distinct difference between the east side and west side groundwater is the concentration of As (Figure 42). Detectable concentrations were found only from groundwater on the west side of the lake. This discrepancy may be due to differences in the redox state of the groundwater or the recharge areas of the groundwater.

Data on the redox state of groundwater is limited. Redox chemistry is available from two wells in which Fe$^{2+}$ and total iron concentrations were obtained colorimetrically (Standard Methods, 1992, 3500-FeD). Total iron concentrations are available from ten wells and a spring, also determined colorimetrically. Figure 43 is a plot of the total iron concentrations determined, plotted against nitrate concentration. Both of these constituents are affected by the redox state of the groundwater. Only two samples display detectable nitrate concentrations (>6 x 10^{-4} mmol/L), the Crystal Geyser well (CGR–2) and Little Lake spring. The remaining samples are under reducing conditions or have low levels of nitrate (Horseshoe springs, Southeast Spring). The iron concentrations range over two orders of magnitude, with the highest concentrations (10^{-2} mmol/L) detected in the River wells and Keeler–Swansea. The Fe$^{2+}$ and total iron concentrations for the River wells and Sulfate Well (the only wells for which this analysis was performed) is tabulated in Table 15.
Figure 42: Arsenic concentrations in some deep groundwater locations. Detectable concentrations are found only in wells from the west side of the lake bed. CGR-2 = Crystal Geyser well. KSI = Keeler-Swansea Intermediate.

Figure 43: Total iron concentrations and nitrate concentrations for some deep groundwater locations. Dashed line indicates approximate detection limits of iron and nitrate. OL-92 and CW-MW2S are below detection with both total iron and nitrate. Little Lake spring abbreviated LL. HSspring = Horseshoe Springs; SEspring = Southeast spring (L9); RWU and RWL are the upper and lower River Well, respectively; KSL = Keeler Swansea Lower. CGR-2 = Crystal Geyser well.
Table 15: Iron concentrations in mmol/L for three wells.

<table>
<thead>
<tr>
<th>Well</th>
<th>$\text{Fe}^{2+}$</th>
<th>Total iron</th>
</tr>
</thead>
<tbody>
<tr>
<td>River Well Upper</td>
<td>$2.68 \times 10^{-3}$</td>
<td>$3.94 \times 10^{-1}$</td>
</tr>
<tr>
<td>River Well Lower</td>
<td>$4.92 \times 10^{-2}$</td>
<td>$4.92 \times 10^{-2}$</td>
</tr>
<tr>
<td>Sulfate Well</td>
<td>$5.37 \times 10^{-4}$</td>
<td>$5.37 \times 10^{-4}$</td>
</tr>
</tbody>
</table>

Samples with the highest iron concentrations (upper and lower River Well, Keeler–Swansea Intermediate) are interpreted as being under reducing conditions but prior to the precipitation of pyrite. In this case, iron is soluble:

$$2\text{SO}_4^- + \text{Fe}^{2+} + 16\text{H}^+ + 14\text{e}^- \rightarrow 2\text{Fe}^{2+} \cdot 3\text{H}_2\text{O} \quad \text{[Drever, 1988, p. 292].}$$

Samples from deep wells with low or non-detectable iron (CW–MW2D, Sulfate Well, OL–92) are interpreted to be under more strongly reducing conditions, to the point of pyrite precipitation. In this case, iron is removed from solution:

$$2\text{SO}_4^- + \text{Fe}^{2+} + 16\text{H}^+ + 14\text{e}^- \rightarrow \text{FeS}_2 + 8\text{H}_2\text{O} \quad \text{[Drever, 1988, p. 301].}$$

Based on available data, redox chemistry does not appear to control As distribution. For example, Keeler–Swansea Intermediate has one of the highest iron concentrations (presumably a function of redox), but non-detectable As. Conversely, in CW–MW2D, iron concentration is low relative to the other wells, but displays the highest As concentrations of those locations sampled. Also, CGR–2 is an assumed oxic environment (based on detectable nitrate and non-detectable iron), yet still has a detectable As concentration. Therefore, the As distribution is assumed to be an indication of different source areas, rather than redox chemistry. This conclusion is supported by the work of Cherry et al. (1979). They find that the As(III)/As(V) ratio varies with changes in redox conditions, and hence is a good indicator of redox state. The total As concentration, however, is only redox dependent at low pH and pe (pe = 0 to -6; pH = 4 to 8) in the presence of sulfur, when there is the potential for As–sulfide precipitation. They state that the presence of total dissolved As in groundwater is more dependent on As abundance in geologic materials than redox condition.
Average As concentrations in granites are twice as high as those in limestones (Drever, p. 328, Table 15–1). The dominant rocks of the Owens basin are granites on the west and limestones on the east, which suggests that source rocks may control As distribution.

The preceding examples are intended to point out geochemical and isotopic differences between groundwater from different areas on and around the lake. The variation in stable isotopes, $^{14}$C dates, Sr isotopes and As distribution suggests that the deep groundwater in the Owens basin is not a homogeneous system and that different recharge times and recharge areas are represented.

The next two sections of this chapter use geochemical evidence to reveal information about the movement of groundwater in the basin, hence contributing information to a conceptual groundwater model.
Evidence of fault–controlled groundwater flow

The goals of this section are to provide evidence that groundwater flow may be controlled by the presence of faulting, and to explain how faulting affects the mixing of different groundwaters.

In the northeastern portion of the lake, there is an opportunity to investigate the possible effects of faulting on groundwater flow. Numerous fault–zones on and around the lake bed are delineated by Pakiser et al. (1964), one of which is shown in Figure 44. Figure 44 also shows the locations of several spring mounds in the northeast section of the lake which discharge at least 0.4 km out onto the lake bed and follow distinctly linear patterns in the vicinity of the mapped fault zones. Spring mounds are unusual because most springs around Owens Dry Lake discharge at the lake margins where topography and permeability change. The mounds vary in size (from several centimeters in height and diameter to more than a meter), moisture content, and vegetative cover. Often small pools of water are present at the crest of the mounds, but generally those that possess a cover of salt grass do not have surface water. The spring mounds from which groundwater is analyzed are identified as SMA001, SMA004, SMA007, SMA019, and SMA069 in Figure 44, and are shown relative to the locations of the wells and shallow piezometers in the vicinity.

In order to prove the hypothesis of fault–controlled groundwater flow, the geochemical and isotopic characteristics of groundwater from the spring mounds are compared to surrounding shallow and deep groundwater. Processes that could result in the observed chemistry are explored, including direct evolution from groundwater on the alluvial fans to the east, and groundwater mixing between alluvial fan groundwater and deeper groundwater. Mixing between spring mound groundwater and shallow, more saline groundwater is also discussed. Geochemical evolutionary paths are one indication of groundwater flow paths. If spring mound groundwater can be shown to evolve from groundwater on the alluvial fans to the
Figure 44: Detail of northeast portion of lake bed. Approximate spring mound locations from global positioning survey (GPS) conducted by Great Basin UAPCD. Approximate location of fault zones from Pakiser et al. (1964).
east, then the possibility exists that the fault zone intercepts groundwater flow from the east and discharges this groundwater in the spring mounds. If the spring mound groundwater can be evolved from groundwater in the northern part of the lake bed (River wells, for example) then faulting may have created a zone of preferential flow from the north to the northeast section of the lake, along the N30W fault trace mapped by Pakiser et al. (1964).

**Groundwater chemistry**

Spring mound groundwater is similar to groundwater from deep wells (>30 m) on the east side of the lake and distinct from shallow groundwater in the vicinity. Major ions in the spring mound groundwater are Na, HCO₃, and Cl, and they contribute to salinity (defined as total dissolved solids) values between 2.0% and 4.5% (Table 16). The greatest variation among the spring mound water samples occurs in the relative concentration of Mg. SMA019 is distinguished by a Mg concentration significantly higher than the other spring mound samples (1.3 meq/L compared to <0.14 meq/L). The spring mound samples are supersaturated with respect to calcite and dolomite.

Salinity of spring mound groundwater is two orders of magnitude less than average salinity in shallow groundwater. In shallow groundwater, however, SO₄ is more significant compared to the three dominant ions (Na, Cl, HCO₃). The SO₄ concentration in the shallow groundwater (after normalization to Cl) is on average two orders of magnitude higher than SO₄ concentrations in the spring mounds (also normalized to Cl). Shallow groundwater is more likely to be oxidizing than reducing, and may not be capable of reducing SO₄ to H₂S. This implies that the spring mound groundwater originates at some point deeper than the piezometer groundwater, where anaerobic conditions dominate.
Table 16: Chemistry of spring mound groundwater. All concentrations in meq/L.

<table>
<thead>
<tr>
<th></th>
<th>SMA001</th>
<th>SMA004</th>
<th>SMA007</th>
<th>SMA019</th>
<th>SMA069</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na</td>
<td>27.7</td>
<td>30.8</td>
<td>30.8</td>
<td>24.6</td>
<td>51.7</td>
</tr>
<tr>
<td>K</td>
<td>0.862</td>
<td>0.926</td>
<td>0.862</td>
<td>1.37</td>
<td>1.55</td>
</tr>
<tr>
<td>Ca</td>
<td>0.115</td>
<td>0.085</td>
<td>0.140</td>
<td>0.209</td>
<td>0.110</td>
</tr>
<tr>
<td>Mg</td>
<td>0.063</td>
<td>0.072</td>
<td>0.086</td>
<td>1.31</td>
<td>0.135</td>
</tr>
<tr>
<td>Cl</td>
<td>8.35</td>
<td>9.926</td>
<td>10.1</td>
<td>6.63</td>
<td>18.9</td>
</tr>
<tr>
<td>HCO₃</td>
<td>21.1</td>
<td>24.2</td>
<td>24.1</td>
<td>23.3</td>
<td>38.8</td>
</tr>
<tr>
<td>SO₄</td>
<td>0.065</td>
<td>0.046</td>
<td>0.094</td>
<td>0.002</td>
<td>0.025</td>
</tr>
<tr>
<td>pH</td>
<td>8.96</td>
<td>8.92</td>
<td>9.12</td>
<td>8.35</td>
<td>8.79</td>
</tr>
<tr>
<td>δDsnow%ο</td>
<td>−124</td>
<td>−124</td>
<td>−124</td>
<td>−130</td>
<td>−125</td>
</tr>
<tr>
<td>⁸⁷Sr/⁸⁶Sr</td>
<td>−</td>
<td>0.70938</td>
<td>−</td>
<td>0.70915</td>
<td>0.70928</td>
</tr>
</tbody>
</table>

The chemistry of wells and springs in the vicinity of the spring mounds is presented in Table 17. Salinity of groundwater collected from deep wells and springs ranges from 0.6% to 3.0%o. Groundwater from Keeler town well (KTW) and the Keeler Fan well (KFW) may be characterized as Na–Mg–HCO₃–Cl water with a significant concentration of SO₄. Sulfate concentrations decrease significantly to the west (Keeler–Swansea Intermediate (KSI) and Sulfate Well (Sulf)), and Mg decreases in groundwater west of the lake bed margin. The River Wells (upper–RWU, lower–RWL), 9.6 kilometers north–northwest of the spring mounds, at the Owens River delta, are Ca(Mg)–HCO₃–Cl waters. All of the spring and well waters are supersaturated with calcite and dolomite.

Spring mound similarity to groundwater in nearby wells is also expressed by the signature of ⁸⁷Sr/⁸⁶Sr. Three of the spring mound samples were analyzed for ⁸⁷Sr/⁸⁶Sr. The reported ratios (±10 in the last two significant figures) are 0.70915, 0.70928, and 0.70938 for spring mounds SMA019, SMA069, and SMA004, respectively. The ratios are similar to Keeler–Swansea Intermediate (0.70901) and Sulfate well (0.70909) within the accuracy of the analysis.
The spring mounds are similar in geochemistry to nearby margin springs (L3, L4) in terms of salinity and major ion categories (Na–HCO₃–Cl). However, the concentrations of Ca and Mg are significantly higher in L3 and L4 indicating that if the springs are on the same flow path as the spring mounds, then the springs are geochemically at an earlier evolutionary stage. Calcium and Mg will precipitate from groundwater as CO₃ concentrations increase along a flow path.

Table 17: Comparison of major ion chemistry of wells and springs in vicinity of northeast spring mounds. All concentrations in meq/L.

<table>
<thead>
<tr>
<th></th>
<th>KTW</th>
<th>KFW</th>
<th>KSI</th>
<th>L3</th>
<th>L4</th>
<th>Sulf</th>
<th>RWU</th>
<th>RWL</th>
<th>Mill</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na</td>
<td>6.70</td>
<td>12.7</td>
<td>18.4</td>
<td>22.4</td>
<td>39.7</td>
<td>36.4</td>
<td>4.87</td>
<td>4.74</td>
<td>39.19</td>
</tr>
<tr>
<td>K</td>
<td>0.588</td>
<td>0.317</td>
<td>1.23</td>
<td>1.47</td>
<td>0.089</td>
<td>0.647</td>
<td>0.683</td>
<td>0.396</td>
<td>1.95</td>
</tr>
<tr>
<td>Ca</td>
<td>1.87</td>
<td>0.823</td>
<td>1.11</td>
<td>0.873</td>
<td>1.56</td>
<td>0.090</td>
<td>0.634</td>
<td>3.15</td>
<td>0.091</td>
</tr>
<tr>
<td>Mg</td>
<td>5.38</td>
<td>1.13</td>
<td>8.10</td>
<td>6.12</td>
<td>0.413</td>
<td>0.133</td>
<td>1.20</td>
<td>4.96</td>
<td>0.15</td>
</tr>
<tr>
<td>Cl</td>
<td>2.68</td>
<td>2.88</td>
<td>7.05</td>
<td>8.46</td>
<td>13.2</td>
<td>7.45</td>
<td>0.773</td>
<td>2.51</td>
<td>7.73</td>
</tr>
<tr>
<td>SO₄</td>
<td>2.24</td>
<td>2.40</td>
<td>0.062</td>
<td>&lt;0.01</td>
<td>0.77</td>
<td>0.035</td>
<td>0.052</td>
<td>0.042</td>
<td>1.89</td>
</tr>
<tr>
<td>HCO₃</td>
<td>9.77</td>
<td>10.8</td>
<td>23.3</td>
<td>23.3</td>
<td>29.7</td>
<td>28.5</td>
<td>6.80</td>
<td>11.3</td>
<td>27.58</td>
</tr>
<tr>
<td>CO₃</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>1.43</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>2.22</td>
</tr>
<tr>
<td>pH</td>
<td>7.96</td>
<td>7.98</td>
<td>8.06</td>
<td>7.49</td>
<td>8.45</td>
<td>8.34</td>
<td>8.37</td>
<td>7.4</td>
<td>7.2</td>
</tr>
<tr>
<td>⁸⁷Sr/⁸⁶Sr</td>
<td>–</td>
<td>–</td>
<td>0.70901</td>
<td>–</td>
<td>–</td>
<td>0.70909</td>
<td>0.709486</td>
<td>0.709426</td>
<td>0.70873</td>
</tr>
</tbody>
</table>

Ratios of various major ions are compared between the spring mounds, springs, and wells in Table 18. The ratios of the spring mounds compared to the wells to the east, especially Mg/Ca, SO₄/Cl, and Ca+Mg/Cl, indicate that geochemical processes may link the two locations. Ratios in the spring mounds are most similar to those in L4, a spring slightly west of the lake bed margin, and the Sulfate Well. The ratios in Table 18 also indicate that the chemistry of the upper and lower River Wells is distinct from the chemistry of the groundwater in the northeastern section of the lake.
Table 18: Ratios of major ions in spring mounds, springs, and wells in northeast section of lake.

<table>
<thead>
<tr>
<th>ratio</th>
<th>SMA001</th>
<th>SMA004</th>
<th>SMA007</th>
<th>SMA019</th>
<th>SMA069</th>
<th>KTW</th>
<th>KFW</th>
<th>KSI</th>
<th>L3</th>
<th>L4</th>
<th>Sulf</th>
<th>RWU</th>
<th>RWL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Na/HCO₃</td>
<td>1.31</td>
<td>1.27</td>
<td>1.28</td>
<td>1.05</td>
<td>1.33</td>
<td>0.686</td>
<td>1.17</td>
<td>0.790</td>
<td>0.961</td>
<td>1.34</td>
<td>1.28</td>
<td>0.716</td>
<td>0.419</td>
</tr>
<tr>
<td>Na/Cl</td>
<td>3.32</td>
<td>3.10</td>
<td>3.05</td>
<td>3.71</td>
<td>2.73</td>
<td>2.50</td>
<td>4.51</td>
<td>2.61</td>
<td>2.64</td>
<td>3.01</td>
<td>4.88</td>
<td>8.93</td>
<td>2.19</td>
</tr>
<tr>
<td>HCO₃/Cl</td>
<td>2.53</td>
<td>2.44</td>
<td>2.39</td>
<td>3.51</td>
<td>2.05</td>
<td>3.64</td>
<td>3.75</td>
<td>3.30</td>
<td>2.75</td>
<td>2.25</td>
<td>3.82</td>
<td>8.80</td>
<td>4.50</td>
</tr>
<tr>
<td>Mg/Ca</td>
<td>0.548</td>
<td>0.847</td>
<td>0.614</td>
<td>6.27</td>
<td>1.23</td>
<td>2.88</td>
<td>1.37</td>
<td>7.30</td>
<td>7.01</td>
<td>0.265</td>
<td>1.48</td>
<td>1.89</td>
<td>1.57</td>
</tr>
<tr>
<td>SO₄²⁻/Cl</td>
<td>0.00778</td>
<td>0.00463</td>
<td>0.00928</td>
<td>0.00030</td>
<td>0.00132</td>
<td>0.838</td>
<td>0.850</td>
<td>0.00879</td>
<td>&lt;0.001</td>
<td>0.00589</td>
<td>0.00470</td>
<td>0.235</td>
<td>0.0148</td>
</tr>
<tr>
<td>Ca+Mg/Cl</td>
<td>0.0213</td>
<td>0.0158</td>
<td>0.0224</td>
<td>0.0241</td>
<td>0.0130</td>
<td>2.70</td>
<td>1.21</td>
<td>1.31</td>
<td>0.826</td>
<td>0.149</td>
<td>0.0299</td>
<td>2.37</td>
<td>3.23</td>
</tr>
</tbody>
</table>
Hydraulic connections based on groundwater geochemistry

Groundwater from wells located east of the spring mounds (Keeler Swansea Well, Keeler Town Well) is geochemically similar to the groundwater collected from the spring mounds. However, there are distinguishable characteristics that imply geochemical evolutionary trends from the alluvial fans to the spring mounds. Lopes (1988) notes a disappearance of \( \text{SO}_4 \) between wells on the alluvial fans and springs on the lake margin. This trend is repeated between the alluvial fan wells and the spring mounds. Sulfate concentrations in groundwater collected from wells on the alluvial fans is significantly greater than \( \text{SO}_4 \) concentrations in water collected from the spring mounds (Figure 45). Calcium concentrations also decrease significantly between the alluvial fan well samples and the spring mound samples (Figure 46). Low Ca concentration in Keeler Fan Well is attributed to a lack of well purging prior to sample collection. Calcite may have precipitated in the casing. The corresponding decrease of Ca and \( \text{SO}_4 \) concentrations suggests gypsum precipitation \((\text{CaSO}_4)\), but undersaturation with gypsum in wells, springs, and spring mounds does not indicate that this occurs.

Disappearance of \( \text{SO}_4 \) may result from groundwater flow from alluvial sediments to lake sediments with higher organic content. Microbially mediated \( \text{SO}_4 \) reduction to hydrogen sulfide could account for the \( \text{SO}_4 \) loss. This process may be accompanied by a rise in pH by the following reaction:

\[
\text{SO}_4^{2-} + 2\text{C}_{\text{org}} + \text{H}_2\text{O} = \text{H}_2\text{S} + \text{HCO}_3^- \quad \text{(Denver, 1988, p. 60)}.
\]

This scenario is supported by a rise in pH (7.96–9.12) from the alluvial fan to the spring mounds (see Tables 16 and 17), and the smell of hydrogen sulfide in groundwater from Keeler Swansea Intermediate and the spring mound waters. This geochemical evolutionary trend implies a flow path from the alluvial fan to the spring mounds.