

Salinity and Hydrology of Closed Lakes

GEOLOGICAL SURVEY PROFESSIONAL PAPER 412



Salinity and Hydrology of Closed Lakes

By WALTER B. LANGBEIN

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*A study of the long-term balance
between input and loss of salts
in closed lakes*



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SALINITY AND HYDROLOGY OF CLOSED LAKES

By WALTER B. LANGBEIN

ABSTRACT

Lakes without outlets, called closed lakes, are exclusively features of the arid and semiarid zones where annual evaporation exceeds rainfall. The number of closed lakes increases with aridity, so there are relatively few perennial closed lakes, but "dry" lakes that rarely contain water are numerous.

Closed lakes fluctuate in level to a much greater degree than the open lakes of the humid zone, because variations in inflow can be compensated only by changes in surface area. Since the variability of inflow increases with aridity, it is possible to derive an approximate relationship for the coefficient of variation of lake area in terms of data on rates of evaporation, lake area, lake depth, and drainage area.

The salinity of closed lakes is highly variable, ranging from less than 1 percent to over 25 percent by weight of salts. Some evidence suggests that the tonnage of salts in a lake solution is substantially less than the total input of salts into the lake over the period of existence of the closed lake. This evidence suggests further that the salts in a lake solution represent a kind of long-term balance between factors of gain and loss of salts from the solution.

Possible mechanisms for the loss of salts dissolved in the lake include deposition in marginal bays, entrapment in sediments, and removal by wind. Transport of salt from the lake surface in wind spray is also a contributing, but seemingly not major, factor.

The hypothesis of a long-term balance between input to and losses from the lake solution is checked by deriving a formula for the equilibrium concentration and comparing the results with the salinity data. The results indicate that the reported salinities seemingly can be explained in terms of their geometric properties and hydrologic environment.

The time for accumulation of salts in the lake solution—the ratio between mass of salts in the solution and the annual input—may also be estimated from the geometric and hydrologic factors, in the absence of data on the salt content of the lake or of the inflow.

INTRODUCTION

Closed lakes have no outlet and are therefore salty to various degrees. Some, like Walker Lake in Nevada, contain less than 1 percent salt in solution; others such as Great Salt Lake, contain more than 25 percent salt. These differences in salt content need to be explained.

The classic explanation of the salt content of closed lakes is that the salt load continuously accumulates, except for the precipitation of the less soluble salts, the total content is presumed to be the total input of salts during the time since the lake last overflowed

as an open lake. This theory has often been applied to determination of the age of lakes. Indeed, Edmund Halley (1715), suggested that the "saltiness" of the ocean is a measure of its age, observing rather cautiously that "though perhaps by it the world may be found much older than many have hitherto imagined."

The principle presumes a steady accumulation of one or more of the influent ions. But if the influent salts were accumulative, then many closed lakes ought to be saturated with respect to one or more of the common salts. A count by Hutchinson (1957) showed that most closed lakes contain less than 5 percent dissolved matter; only a few have over 10 percent in solution. The saturation point for common salt is over 30 percent. The relatively low salinity suggests that salt accumulation may be offset by some process of salt wastage.

In reviewing this evidence in an earlier paper, Hutchinson (1937) concluded:

The relatively high frequency of such brackish as opposed to highly saline waters may be explained in part by the opportunity given by extremes of climatic fluctuations for the loss of salt from closed basins as indicated above. That the total duration of the lake basin is of less importance is abundantly demonstrated by the probable history of the Lahontan Lakes.

Hutchinson therefore concluded that

The present salinity of the lakes [Pyramid and Walker Lakes in Nevada] is not directly related to their age.

The question had intrigued Gilbert (1890, p. 229) when, in 1880, he visited Sevier and Rush Lakes in Utah. Like Great Salt Lake, these lakes are remnants of Pleistocene Lake Bonneville. Sevier Lake had over twice the salinity of sea water, but nearby Rush Lake was "drinkable though too brackish to be palatable." Although he observed that these two lakes existed under nearly identical conditions, he concluded that the relative freshness of Rush Lake, the smaller of the two, was to be accounted for by the fact that it had once "evaporated to dryness, and its saline matter being thus precipitated has become buried under mechanical sediment."

The Gilbert and Hutchinson hypothesis is that loss of salts occurred during the extremes of lake overflow

and dessication. The water in the lake is freshened during periods of overflow; at the other extreme, salts are lost from the dry lakebed by burial under aeolian or waterborne sediment or by removal by wind. The present investigation explores an enlarged hypothesis that each fluctuation in lake level can be an agent for wastage of salts, and that there is therefore a sort of long-term balance between input and loss of salts about which the salt content of the water varies in the same way that lake levels vary about a long-term balance between the inflow of water and discharge by evaporation. Whether the salts are removed by the wind or endure in the lake sediments is not known so that only the lake water and its contained salts are discussed herein.

In a hydraulic sense, lakes are classified as open or closed, depending on whether they have an outlet or not. There are two kinds of open lakes—drainage lakes and seepage lakes (Hutchinson, 1957). Drainage lakes are the usual lakes with stream inflow and outflow. Seepage lakes are those where the outflow is due to seepage; the input also may be through seepage. Seepage lakes topographically may appear to be closed lakes.

This paper deals only with those closed lakes whose chief source of inflow is surface water—the lake is fed by surface streams. All closed lakes are saline and their salinity is greater than that of the influent water. However, not all saline lakes are closed lakes in the sense of this paper. For example, excluded are the salt basins described by Meinzer (1911) that are wind excavated in the floors of arid valleys. The basins are occupied by highly saline water derived from saline ground water.

Lakes considered in this report contain relatively well-mixed water; that is, there is no marked saline stratification. Excluded therefore are meromictic lakes which, as defined by Hutchinson (1957, p. 480), contain saline water at depth which remains unmixed with the main water mass.

Closed lakes that meet these definitions vary in their permanence (see fig. 1). Pyramid Lake, for example, occupies a deep basin and contains water even during rigorous droughts. Great Salt Lake is shallow, and its fluctuations in stage and area are relatively great; but it is a perennial lake. On the other hand, there are many closed lakes, as, for example, Lake Abert in Oregon, that are nearly dry on rare occasions, and those, like Lake Eyre in Australia, that contain some water on rare occasions.

The number of closed lakes in an area increases with the impermanence of lakes in the area. Few closed lakes are like Pyramid Lake or the Dead Sea. Many more are like Lake Abert, whereas playa lakes (or the

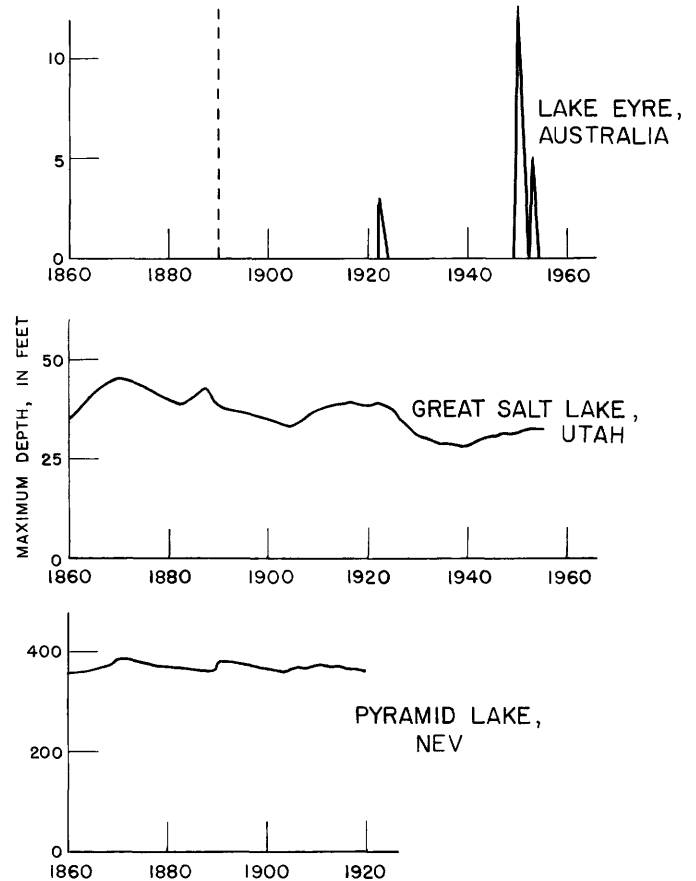


FIGURE 1.—Fluctuations in depth of three types of closed lakes.

paradoxical but descriptive term “dry lakes”) must be counted in the thousands. Lake Eyre, graphed on figure 1, is one of the few of this type for which some records are available (Bonython, 1955).

Pertinent data assembled for this study are given in table 1 for a number of closed lakes.

CLIMATOLOGIC AND HYDROLOGIC FACTORS

The hydrologic imbalance between evaporation and input is itself a fascinating aspect of closed lakes. A consideration of these factors enables us to speculate upon the climatic controls on the existence of closed lakes and upon the stability and fluctuation of water level, a subject of direct concern to the salt balance.

CLIMATOLOGIC LIMITS FOR CLOSED LAKES

The one universal control on the existence of closed lakes is evaporation. Evaporation from lake surfaces has been studied extensively and intensively (Harding, 1935; Harbeck, Kohler, and Koberg, 1958) so that close estimates of it can be made quite directly, given sufficient pertinent data on temperature, radiation, wind, and humidity. Kohler and others (1959) show a map of annual evaporation from water surfaces that

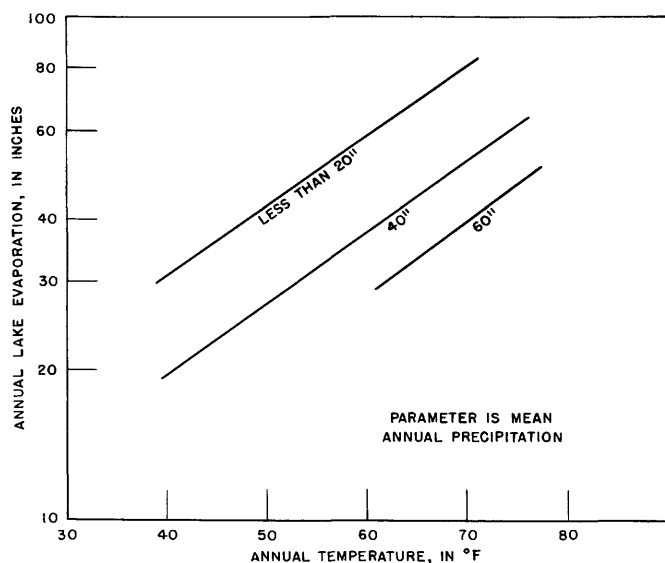


FIGURE 2.—Relation of annual lake evaporation to annual temperature and annual precipitation.

summarizes the variation in the rate of evaporation in the United States.

Estimates of evaporation from lakes abroad must be made on scant information. The writer hesitates to introduce another relation for evaporation to the hundred or more that have already been proposed; but strange as it may seem, none appears to be in just the form needed for estimating evaporation if only data on precipitation and temperature are given. Hence, available data on lake evaporation in the United States (Kohler and others, 1959) have been correlated with pertinent data on precipitation and temperature. The resulting correlation is graphed on figure 2. Precipitation does not affect evaporation directly, but it is an indirect measure of humidity and radiation. In continental climates, greater precipitation usually means greater humidity and greater cloudiness, each a factor that tends to decrease the rate of evaporation. Since precipitation is not generally an adequate measure of these factors in coastal regions, figure 2 is not intended to apply to such places.

The evaporation rate from saline water is less than from fresh water in the same climatic environment. The decrease is due to a lowering of the vapor pressure of a liquid by the solute. Although the net effect on evaporation depends on the relative humidity of the atmosphere and on the kind of salt in solution, the lowering in rate of evaporation in percent is roughly equal to the percentage of salt in solution, under a wide range of atmospheric conditions and with the salts usually found in natural waters. Thus, the evaporation rate from a solution containing 20 percent by weight of salts (200,000 ppm) will approximate 80 per-

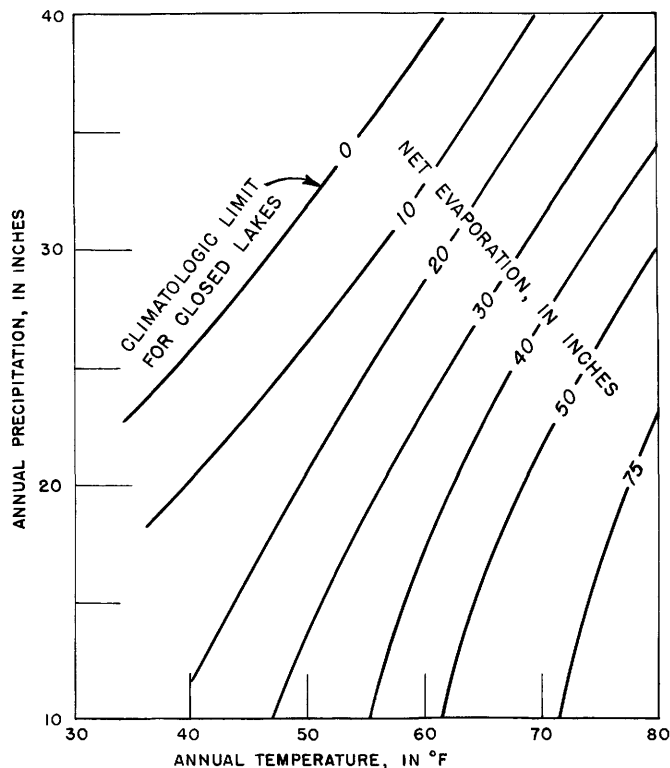


FIGURE 3.—Net evaporation in terms of temperature and precipitation.

cent of that from fresh water, similarly exposed. The effect is important only for lakes with a salinity greater than about 50,000 ppm.

Figure 2 shows the gross rate of evaporation from a lake. However, the difference between gross evaporation and precipitation on the lake is the factor that represents the net rate of removal of water from a closed lake. Figure 3, prepared from figure 2, shows the values of the net evaporation in terms of the annual temperature and precipitation.

Since closed lakes cannot exist where precipitation on the lake surface exceeds the gross rate of evaporation, the zero line of figure 3 defines the climatologic limit of closed lakes. In other words, closed lakes cannot exist where the combinations of precipitation and temperature lie above the zero line on figure 3. There is a further restriction. As pointed out by Eville Gorham (written communication, Sept. 1960), annual fluctuations in precipitation would cause frequent overflow of lakes near this theoretical limit so that such lakes would not be saline nor considered closed. Moreover, a closed lake at the theoretical limit cannot have any tributary area, otherwise the inflow would cause it to overflow. Since all closed lakes have tributary drainage areas, no closed lakes can occur near the theoretical limit. These two factors restrict the occurrence of closed lakes to regions where gross evaporation is appreciably in excess of precipitation.

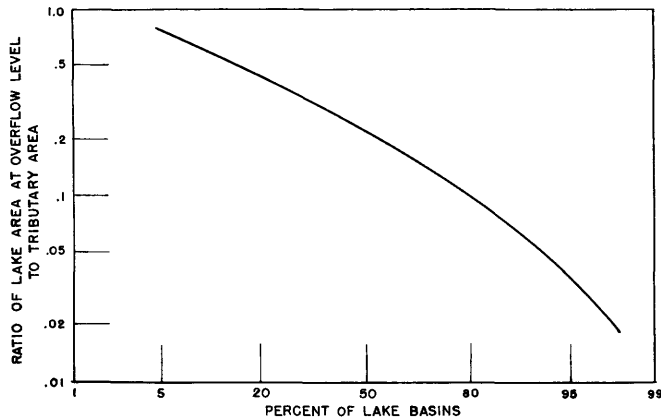


FIGURE 4.—Distribution of ratios of lake areas at overflow level to tributary area.

The easternmost closed lakes in the United States are those in North Dakota at the approximate boundary of the dry climates as defined by Thornthwaite (1948). This boundary runs southward from the eastern part of the Dakotas, through the central part of Oklahoma, and through central Texas.

The closed lake is dominantly a feature of the arid zone. As long ago as 1896, Russell (p. 131-132) concluded that "The study of the present geography of the earth shows that, in regions where the mean annual precipitation exceeds 20 or perhaps 25 inches, inclosed lakes do not occur, although the topographic conditions may be favorable." The explanation seems to be as follows: The greater the aridity, the greater is the rate of net evaporation. The greater the net evaporation, the smaller the area of a closed lake. The smaller the lake area, the more numerous are the topographic opportunities for a closed lake.

Figure 4 shows that the increase in the number of closed basins increases as the ratio of the lake area at overflow level to tributary area becomes smaller. This diagram was constructed by computing the ratios for a number of closed basins. These ratios were listed in order of magnitude, beginning with the lowest. Order numbers were assigned beginning with the lowest as number 1, each higher ratio receiving the next higher number. The order numbers were transformed into percent by the formula $100m/N+1$ where m is the order number and N the total number of lakes (11 in this sample).

In regions where net evaporation is low, the ratio of lake area to tributary area is high, decreasing the likelihood of closed lakes. The reverse occurs in arid regions where net evaporation is high and lake area correspondingly small. Thus, one may explain the increasing frequency of closed lakes with increasing net evaporation, or, in more general terms, with increasing aridity.

WATER BALANCE AND FLUCTUATIONS IN WATER LEVEL

The hydrologic factors affecting the water balance of a closed lake are simple and well known. In Halley's words (1715),

Now I conceive that as all these lakes receive rivers, and have no exit or discharge, so it will be necessary that their waters rise and cover the land, until such time as their surfaces are sufficiently extended, so as to exhale in vapour that water which is poured in by the rivers; and consequently that lakes must be larger or smaller, according to the quantity of the fresh water they receive.

The balance is expressed by the equation

$$I + PA_L = E'A_L$$

where I is inflow from the tributary area in acre-feet per year, P is precipitation on the lake surface in feet per year, A_L is area of the lake in acres, and E' is gross evaporation rate from the lake in feet per year. This equation represents a long-term balance about which a lake fluctuates. Year-to-year variations in I and P produce fluctuations in water level with corresponding variations in surface area. Hence

$$A_L \Delta H = I - (E' - P)A_L$$

or

$$\Delta H = \frac{I}{A_L} - E$$

where E is the so-called net-evaporation rate, the gross evaporation minus precipitation on the lake surface. Since year-to-year changes in the rate of evaporation per unit of lake area are comparatively small, the actual change in stage, ΔH , during a period of time depends almost entirely on the volume of inflow per unit of lake area. But since lake area depends on the stage or water level, the details of change in stage are rather complex.

The stage of a closed lake at any time represents the net sum of all previous inputs and outputs. The following analysis illustrates this point quantitatively:

1. Rate of inflow, I .

2. Rate of discharge, Q , to be the net-evaporation rate, E , times the lake area, A_L , which, in turn, is taken as a linear function of the lake volume, V . Thus

$$Q = EA_L = \frac{(V+b)}{k}$$

where b and k are constants for a particular lake.

The validity of the assumed linear relation between lake area and volume is tested on figure 5. The curves for Lake Eyre and Great Salt Lake show that the relation can be quite irregular. As for the four other lakes shown, the area appears to vary between the 0.4

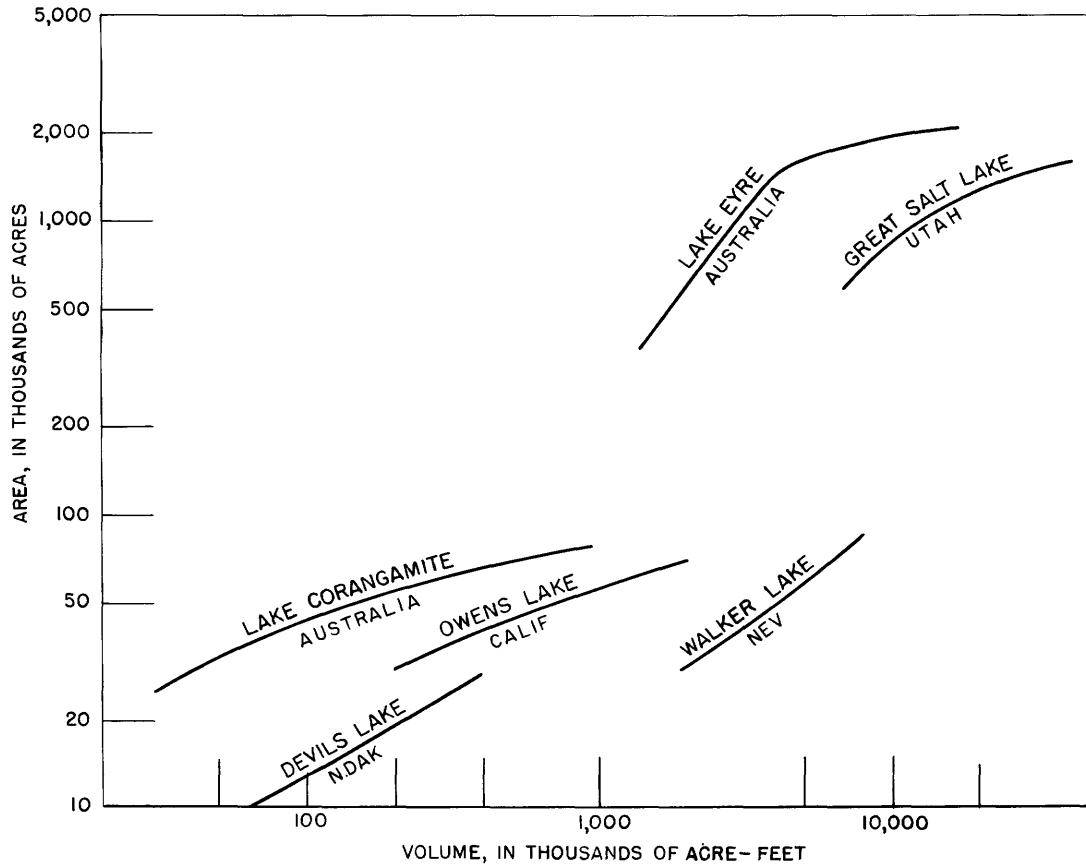


FIGURE 5.—Typical relations between surface area and volume of closed lakes.

and the 0.75 power of the volume. However, the mathematics to follow become intractable under any assumption other than that of a linear relation between evaporation and lake volume. Nevertheless, as can be shown, a linear equation can fit the variation between area and volume over a wide range. It is instructive to follow through to a conclusion under this simplifying assumption.

A change in lake volume, ΔV , occurs whenever there is a difference between rates of inflow and discharge (evaporation). Hence

$$\Delta V = I - Q \quad (1)$$

in which I is rate of inflow and Q is rate of discharge. But

$$Q = \frac{(V+b)}{k};$$

hence, for a unit time,

$$\Delta V = I - \frac{V}{k} - \frac{b}{k}$$

If time intervals are sufficiently close,

$$V_2 - V_1 = I_2 - \frac{V_2}{k} - \frac{b}{k} \quad (2)$$

where subscripts 1 and 2 refer to successive time intervals.

From equation 2 we obtain

$$V_2 = \frac{k}{1+k} \left(I_2 + V_1 - \frac{b}{k} \right) \quad (3)$$

We may also write

$$V_3 = \frac{k}{1+k} \left(I_3 + V_2 - \frac{b}{k} \right) \quad (4a)$$

$$V_4 = \frac{k}{1+k} \left(I_4 + V_3 - \frac{b}{k} \right) \quad (4b)$$

etc.

Inserting the value of V from equation 3 into equation 4a and for V from equation 4a into 4b, etc., we obtain

$$V_n = \frac{k}{1+k} I_n + \left(\frac{k}{1+k} \right)^2 I_{n-1} + \left(\frac{k}{1+k} \right)^3 I_{n-2} + \dots = b. \quad (5)$$

Hence, the volume at any time n is a weighted average of the inflows during the preceding intervals of time, the weights diminishing in geometric ratio with time. Lake

stages can be computed from lake volumes by use of the proper stage-volume curve. and

$$\sigma_i = 0.26 \sqrt{EA_L A_T}$$

The die-away coefficients depend entirely on the value of k , which is the ratio of a change in lake volume to the corresponding change in rate of discharge, and has the dimension of time. The value of k can, therefore, explain a good deal about the nature of fluctuations in the level of closed lakes. Considering the three lakes shown on figure 1, for example, Lake Eyre has a value of k of 1.5 years, Great Salt Lake, 9 years, and Pyramid Lake, 65 years. A lake with a low value of k , near 1 year, is a playa lake. It fills and dries up in a year. It responds to the current year's rainfall and virtually not at all to that of preceding years. A lake with a high value of k , on the other hand, reacts slowly, and may be at a high level during a period of low rainfall and vice versa. The value of k is the response time, in years, of a closed lake, an important characteristic. Response times for closed lakes are listed in table 1.

Since the lake value is made up of successive proportions of inflows, the standard deviation of the lake volumes, σ_v , can be computed from the sum of the squares of the standard deviations of the components with the following result.

$$\sigma_v = \sigma_i \sqrt{\frac{k}{2 + \frac{1}{k}}} \quad (6)$$

where σ_i is the standard deviation of the annual inflows. The standard deviation of the inflows is equal to the coefficient of variation (U_r) times the mean flow. The coefficient of variation of inflow, in turn, varies regionally. It is greater in arid regions than in humid regions. Neglecting local deviations from the general climatic pattern, streamflow records show that the coefficient of variation is related to the mean annual depth of the runoff per unit of drainage area, about as shown on figure 6. Taking a straight line approximation on the logarithmic chart,

$$U_r = \frac{0.9}{\sqrt{r}}$$

where U_r is the coefficient of variation of r , the annual runoff in inches from the tributary area to the lake. But

$$r = 12 \frac{EA_L}{A_T}$$

where E is net evaporation in feet per year, A_L is lake area in square miles, and A_T is the tributary area in square miles. Therefore,

$$U_r = 0.26 \sqrt{\frac{A_T}{EA_L}}$$

Since $V = A_L D$, where D is the mean depth, in feet, it follows that

$$\frac{\sigma_v}{V} = \frac{0.26}{D} \sqrt{\frac{E \left(\frac{A_T}{A_L}\right) k}{\left(2 + \frac{1}{k}\right)}} \quad (7)$$

Equation 7 gives the coefficient of year-to-year variation in lake volume, U_v , in terms of readily obtainable data. The relative variation in lake area can be computed from equation 7 by multiplying the righthand side of the equation by the slope of the area-volume curve as graphed in figure 5. Thus, the coefficient of variation of lake area is

$$U = \frac{\sigma_v}{A} = \frac{0.26n}{D} \sqrt{\frac{E(A_T/A_L)k}{(2+1/k)}}, \quad (8)$$

where n is the exponent in the relationship $A_L \propto V^n$. A rough estimate of 0.6 may be taken for those lakes for which the value of n is unknown. Since the discharge from a closed lake varies with its surface area, the value of U corresponds to the coefficient of variation of discharge from the lake.

In applying equation 7 or 8, the value of k , in years, is defined as

$$k = \frac{V'' - V'}{E(A'' - A')} \quad (9)$$

where V'' refers to a high lake stage and V' a low stage. The value of k so computed is usually greater than the ratio D/E , the excess being greater for shallow than for deep lakes.

It is of interest to compare some results of equation 8 with the direct computation of U for the few closed

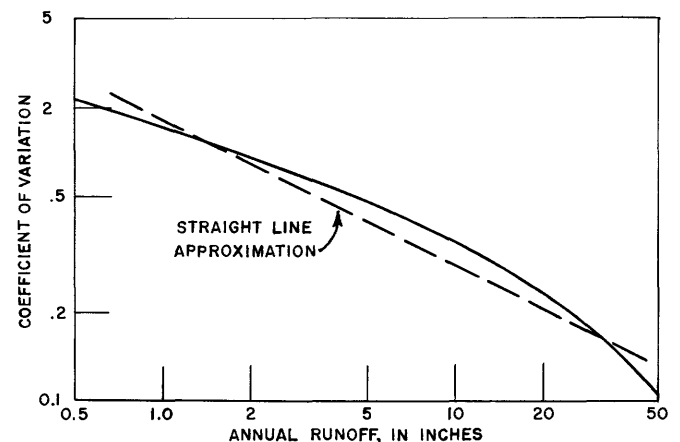


FIGURE 6.—Generalized relation between coefficient of variation of annual runoff and average annual runoff.

lakes for which records of fluctuations in lake level are available.

| | Coefficient of variation of lake area | |
|----------------------------------|--|--------------|
| | By equation 8 | From records |
| Pyramid Lake, Nev..... | 0.040 | 0.025 |
| Mono Lake, Calif..... | .043 | .05 |
| Walker Lake, Nev..... | .075 | .05 |
| Great Salt Lake, Utah..... | .125 | .15 |
| Lake Corangamite, Australia..... | .21 | .30 |
| Devils Lake, N. Dak..... | .40 | .6 |
| Lake Eyre, Australia..... | 2.5 | 3.6 |

The comparison seems satisfactory and gives some confidence that the variability of lake area can be computed from available information on areas and depths. Computed values are listed in table 1. However, because of the simplifying generalizations made in the derivation of equation 8, it is not to be expected that equally good checks can be made with all closed lakes. Ground-water inflow to a lake can be an important stabilizing factor. The greater the amount of ground-water inflow, the smaller the value of U .

HISTORICAL FLUCTUATIONS IN LAKE LEVELS

Existing, as they do, in balance between factors of input and output, the fluctuations in water level of closed lakes are often viewed as indexes of climatic variation. The interpretation of the lake levels depends on the value of the response time, k . The longer the response time, the greater is the possibility that climate and lake levels may be out of phase. Even so, in the United States their usefulness as a measure of current trends seems ended. As Harding (1942) points out, many lakes in the Western United States are declining in level owing not so much to climate as to the effect of diversion of river water for irrigation. For some lakes, the lowering in water level seems to be greater than can be explained by the combined effect of climate and the measured diversion of water. Devils Lake is a noted case (Swenson and Colby, 1955). The level of this lake has dropped about 35 feet in the period from 1867 to 1940. The recession appears far greater than can be accounted for either on the basis of irrigation, which is negligible, or on the basis of a decline in precipitation. Intensive farming that began in the 1880's may have reduced the runoff to the lake. Also, a rising trend in atmospheric temperature may have contributed to the decline in lake levels, directly by increasing lake evaporation and indirectly by increasing the land evapotranspiration and thereby reducing the runoff from the tributary area.

COMPARISON WITH DRAINAGE LAKES

The fluctuations in the level of drainage lakes differ considerably from those of closed lakes. The difference

is due largely to the method of discharge. The discharge from closed lakes varies with the surface area. The fluctuations in level of drainage lakes are controlled by the hydraulics of the outlet.

As previously defined, the response time of a lake is the ratio of a change in volume in the lake to the corresponding change in the rate of discharge from the lake. Thus

$$k = \frac{\Delta V}{\Delta Q} = \frac{A_L \Delta H}{\Delta Q} \quad (10)$$

where A_L is the lake area in square feet, ΔH is change in lake level in feet, and ΔQ is change in discharge in cubic feet per second. For a drainage lake, the ratio $\Delta H/\Delta Q$ is the slope of the stage-discharge rating curve of the outlet. The rating curve is curvilinear, so its slope varies. However, taking the slope at the mean discharge, it is found from an analysis of several lake-discharge rating curves that the ratio $\Delta H/\Delta Q$ approximates $0.07/\bar{Q}^{0.6}$, where \bar{Q} is the mean rate of discharge in cubic feet per second. Hence, the approximate expression for the response time of a drainage lake may be written

$$k = 0.07 A_L / \bar{Q}^{0.6} \quad (11)$$

where A_L is the lake area in square feet, \bar{Q} is mean discharge in cubic feet per second, and k is in seconds; and

$$k = 0.06 A_L / \bar{Q}^{0.6} \quad (12)$$

where A_L is the lake area in square miles, \bar{Q} is mean discharge in cubic feet per second, and k is in years.

Applying equation 12 to a few unregulated lakes (without control works at their outlets), the following results are obtained:

| | Lake area (sq mi) | Mean discharge (cfs) | Response time (years) |
|--|----------------------|----------------------------|-----------------------------|
| Lake Victoria, Africa ¹ | 67,000 | 20,000 | 4.0 |
| Lakes Huron and Michigan.... | 45,000 | 180,000 | 1.9 |
| Lake Baikal, U.S.S.R..... | 12,000 | 100,000 | .72 |
| Lake Ontario..... | 7,520 | 230,000 | .3 |
| Lake Champlain..... | 435 | 11,000 | .10 |
| Yellowstone Lake..... | 139 | 1,300 | .085 |
| Smalley Lake, Ind..... | .1 | 16 | .0011 |

¹ Before regulating works were installed.

The main point is that drainage lakes and closed lakes are of different hydraulic species. Drainage lakes have response periods that are much shorter than those of closed lakes. Hydraulically, most drainage lakes are just "wide places in a river," and their fluctuations in level are markedly less than those of closed lakes.

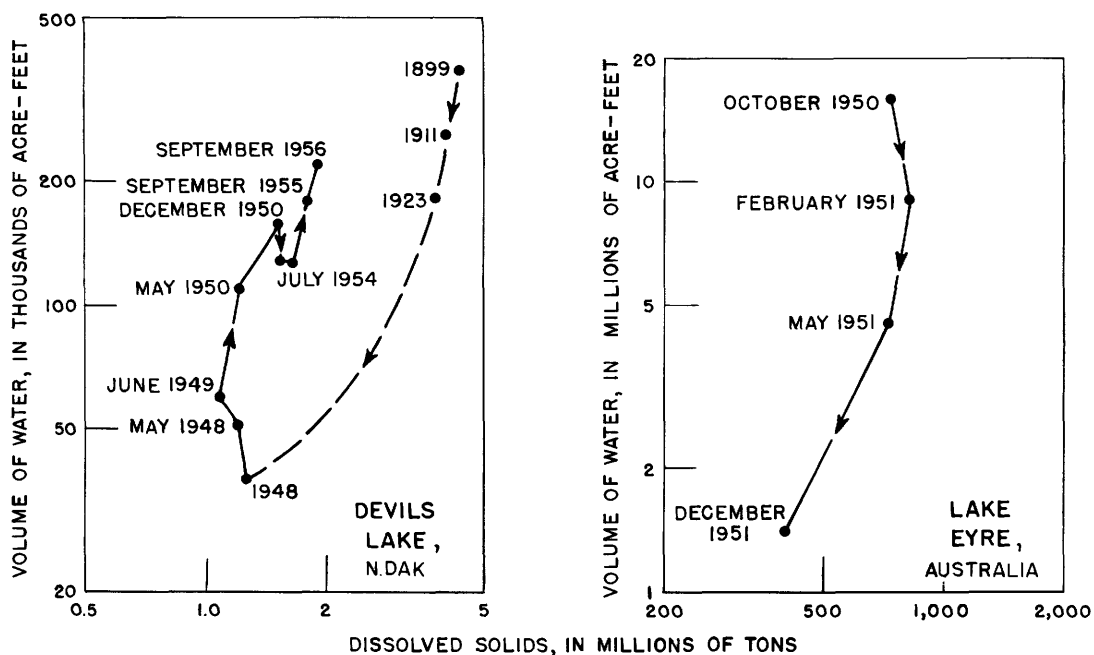


FIGURE 7.—Variation of salt content with water volume.

SALT CONTENT OF CLOSED LAKES

LOSS OF SALTS WITH VARIATION IN LAKE LEVEL

Fluctuations in levels of closed lakes have been studied because, as indicated in the introduction, these fluctuations have a bearing on the salt economy of the lakes. The change in amount of salts in solution during a single fluctuation in lake level must be small indeed, generally too small to be detectable by an ordinary sequence of random sampling of water quality. For example, assuming long-term balance between loss and input, the loss in salts during a lowering in level of the order of magnitude that occurs once in 50 years would equal a 50-year input to the lake. For most lakes, this would be only 1 to 5 percent of the salts in solution, too small to be measurable except where accurate samplings and lake surveys are available. Such data are rare. Nevertheless, some evidence is available for two lakes that have fluctuated considerably in lake level: Devils Lake (Swenson and Colby, 1955) in North Dakota, and Lake Eyre (Bonython, 1955) in South Australia.

LAKE DATA

Figure 7 shows the variation in salt content with lake volume for Devils Lake. This lake receded steadily after 1899, when the water was first sampled, reaching a low stage in 1948 when a series of intensive samplings was begun. During this 49-year period, according to a table published by Swenson and Colby (1955, p. 61), the weight of dissolved solids decreased by two-thirds. The salt tonnage¹ remained substan-

¹ Salt tonnage is the product of the factor 0.00135, the lake volume in acre-feet, the concentration in parts per million by weight, and the density.

tially constant during the increase in lake volume from 1948 to 1950. Seemingly, it would take a state of hydrologic equilibrium for a great number of years to restore the tonnage to the amount in the lake in 1899.

Lake Eyre, a large Australian playa, received water in 1950 that covered an area of 3,100 square miles. By the end of 1952 the lake was dry. Calculations by C. W. Bonython (1955, p. 66) based on samples collected during the receding phase showed that tonnage increased slightly after the first observation in October 1950, presumably reflecting the continuing solution of lakebed saline deposits, and reached a maximum in February 1951. Thereafter, as shown in figure 7, the tonnage in solution apparently decreased markedly as the lake continued to shrink. Bonython points out that his calculations are based on salinity measured at only one place on the shore of an extensive shallow lake and that his results may not be accurate. Although he regards as accurate only the tonnage figure for December 1951, when the lake covered a relatively small area, the figure may be sufficiently representative to suggest the pattern of variation in tonnage with lake volume.

These two examples, the first of a perennial lake and the second of a playa, show a marked decrease in the amount of salt in solution during contractions in lake volume. There is evidence, admittedly fragmentary, that the variation of salinity with lake volume is not directly a simple matter of concentration or dilution of a fixed mass of salts in a changing volume of lake water. The mass of salts in solution appears to decrease with contraction in lake volume. Although the

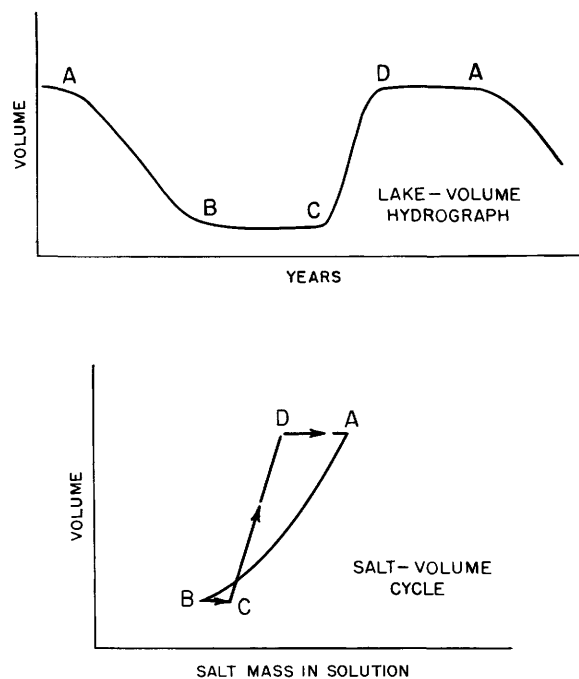


FIGURE 8.—Schematic variation of salt tonnage in solution with fluctuations in lake volume.

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

A SCHEMATIC CYCLE

On the assumption that the loss of salts is linked to fluctuations in lake level, one may follow through to a schematic description of the variation of the salt content of a closed lake. The top diagram of figure 8 represents a generalized pattern of fluctuations in lake volume produced by the year-to-year variations in inflow. The lower diagram suggests an exaggerated pattern of variations in the amount of salt in solution as the lake fluctuates in volume.

During the receding phase, A-B, there is postulated a net loss of salts from solution. The annual input of salts is insufficient to counterbalance the loss of salts as the lake recedes.

The limb B-C of the schematic cycle pictured on figure 8 represents a period of relatively stable low level of a lake during which time input of salts exceeds the losses.

The rising phase, C-D, represents a relatively stable tonnage of salts in solution, if the rise is rapid, or perhaps an increasing tonnage if the rise is prolonged. Increases in lake volume may be more rapid than decreases, because the latter are limited by the annual rate of evaporation from the lake surface.

The limb D-A represents a stable high-level phase of a lake during which the annual input of salts exceeds the losses.

No records are available to demonstrate this idealized cycle in full. The salt-tonnage graph for Devils Lake on figure 7 illustrates a receding phase A-B and a rising phase C-D. Stable limb B-C is absent. The graph for Lake Eyre in figure 7 is entirely during the receding phase A-B.

There are no data to illustrate that part of the schematic cycle that shows an increase in the amount of salt during a long period of relatively stable lake stages.

The actual volume-tonnage curves over a long period of time may never describe the idealized cycle diagrammed on figure 8, but may consist of a series of zigzag curves of which each part is related to one of the limbs of the hypothetical cycle.

Figure 9 illustrates a suggested pattern of fluctuations in the salt content of a playa. In its usual dry state, the lowest part of the playa contains a salt crust that generally thickens toward the central lowest point of the lake bed. The upper edge of the salt crust marks the "salting" level, that level where the dominant salts in the shrinking lake reach saturation. As the lake continues to shrink, salts are precipitated, so that as at Lake Eyre the thickness of the salt crust is about 25 percent of the depth of the water at the salting level. As Bonython (1956) points out, however, local rains may dissolve and carry salts from the upper edges of the salt crust to the lower parts where the salts are reprecipitated. The valley sediments increase in fineness toward the center of the playa, and the sediment under the salt crust is usually a fine-grained mud which, in turn, may overlie deeper salt crusts, as diagrammed on figure 9. Salt crystals grow in the mother liquor in the mud, often forming crystals of "textbook" perfection.

Following a storm sufficiently heavy to cause runoff, water enters the playa and the salt crust begins to dissolve. Although the concentration reaches its maximum at the salting level, due to the rapidity of the storm inflow, the whole of the crust may not be dissolved until the water stands at some height and for some time above the salting level (see fig. 9 B and C). As the lake rises beyond the salting point, the concentration decreases but the amount in solution increases, and, indeed, as depicted in figure 9C, the salt crust may not be entirely dissolved until the lake has begun to recede. This is the point of maximum salt content. Gradually, the amount in solution decreases as the shoreline dries. But loss of salts is not sufficient to retard the increase in concentration with decreasing lake volume. The concentration reaches saturation at the salting level when the lake is saturated with respect to the dominant

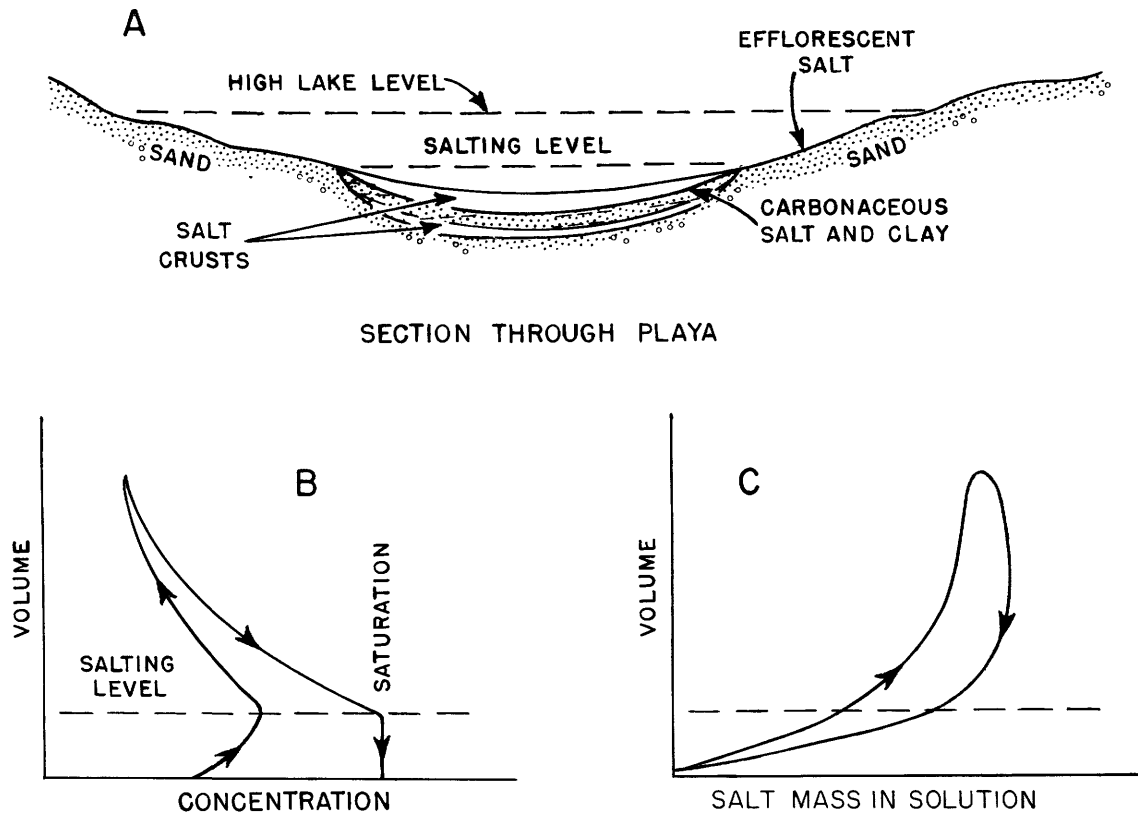


FIGURE 9.—Salt regime of a playa.

salts in the solution. Concentration thereafter remains essentially constant; as the volume of water shrinks, dissolved salts are precipitated, and the lake returns to its usual dry state. In this idealized example, the amount of salt in the crust remains as before; over a period of years, the salt added by inflow equals the loss on evaporation to dryness.

MECHANISMS FOR THE LOSS OF SALTS

The possible mechanisms that cause a decrease in dissolved salts during fluctuations in lake volume must, in the absence of direct evidence, remain somewhat conjectural. A recession in lake level leaves water in marginal bays separated from the main body of water. The entrapped water in the bordering alluvium, or in puddles, dries and the dissolved salts are precipitated. Deposited salts that effloresce are subject to wind transport. Or, as Grabau (1920, p. 175) points out, surficial salt deposits that are hygroscopic readily become powdered over with dust. Grabau adds that a surface dust layer, even if only of slight thickness, will prevent the underlying salt from dissolving when the lake level rises.

As mentioned in the previous section, the beds of closed lakes consist of elastic sediments and a large proportion of carbonates precipitated from the solution. Soluble salts trapped in the muds are insulated from

re-solution and there may also be a downward migration of ions in the bottom sediments from places of high concentration to places of low concentration. Thus, the lake sediments may contain not only precipitated carbonates, but some more soluble constituents as well.

Wind must be a major agent for the loss of salts from a closed lake. Most travelers in the arid regions have observed dust clouds whipped up by the wind from the fine sediments and salts in playa bottoms. L. B. Laird, chemist, U.S. Geological Survey, reports observations of clouds of "soda dust" being blown from the dry parts of the bed of Summer Lake in Oregon. (written communication). Feth and others (1960) report that snow in the Wasatch Mountains in Utah is more mineralized than that in the Sierra Nevada in California. Every indication, including maximum values for various constituents, is that the air masses producing snow in the Wasatch Mountains accumulate appreciable amounts of mineral particles while passing over the arid lands of the Great Basin. Despite this direct evidence to substantiate the reality of wind action, data are lacking on its quantitative importance as an agent in the salt balance of closed lakes.

Wind may also be an agent for the loss of salts by the removal of small droplets of water (called aerosols) with dissolved salt from the surface of the lake. The small particles of water are carried aloft and evaporate

in the air, and the fine particulate salt remaining may then be transported by wind for considerable distances and may constitute nuclei for cloud formation. Thus entrained in subsequent precipitation, the salt returns to the earth.

It has been argued by many and for many years (for example, see Ackroyd, 1904) that the chloride of river waters is derived principally from rainfall, which carries the salt that originated from the oceans. The rough order of magnitude of removal of salts from the oceans by wind may be estimated if one is permitted the liberty of several broad assumptions. The first of these suppositions is that the salt ($\text{Cl}^- + \text{Na}^+$) balance of the oceans is maintained by the above means; that is, the net removal of these ions from the oceans in aerosols is equal to the input by the rivers. The annual input of chloride ions to the oceans from the continents, according to data given by Clarke (1924), is about 1.1×10^8 tons. Taken together with the equivalent sodium, the input of sodium chloride is about $58/35 \times 1.1 \times 10^8$ tons per year, or 1.8×10^8 tons per year.

According to the cyclic hypothesis, this input would be derived from fallout of ocean-produced aerosols. However, this quantity would represent only that part of the airborne salt that is carried over and deposited on the continents. Compounding the suppositions with the assumption that twice as much salt is deposited per unit area on the ocean as on the land, the total amount of ocean salts carried in aerosols becomes about 1×10^9 tons per year, or 7.5 tons per year per square mile, of which a fourth is deposited on the continents to be returned to the oceans by streams, and three-fourths directly on the oceans. A further assumption that the removal of salts in aerosols is directly proportional to the concentration yields a factor of 2.1×10^{-4} tons per square mile per year per part per million of concentration, or about 1/300 inch of ocean water per year. This quantity does not differ significantly from an estimate by Erickson (1958) who assumes that production of airborne particles of sea salt varies with latitude and with the transport of water vapor from the sea surface; that is, with the evaporation. He suggests that annual production of sea-salt particles increases from near 0 at the poles to about 4,500 mg per square meter at lat. 20° N. and S., or a mean of 2,800 mg of salt per square meter. This figure is equivalent to 1/300 inch of water containing 36,000 ppm of salt in solution.

Applying this depth of 1/300 inch of water per year to some salt lakes gives rates of transport from the lakes that appear far short of those necessary to balance the river-carried input of salts. For Great Salt Lake, using a concentration of 200,000 ppm and an area of 1,800

square miles, an annual removal of 75,000 tons is obtained. Applied to the Caspian Sea, the largest of the closed lakes, and using an areal mean concentration of 12,000 ppm, an annual removal of 300,000 tons is calculated. Both of these results appear much smaller than the input of salts to the lakes.

This extrapolation from the oceans to closed lakes may be inappropriate because of the great disparity in area and depth. Further, the smaller area and shorter wind sweep would lead to lesser wave action and hence reduce the production of aerosols from lakes.

Moreover, geometrical considerations strongly suggest that although areal influences may be dominant for the ocean, marginal or peripheral influences would be dominant for closed lakes. The losses of salts in aerosols occur from the lake surface, but the effects of fluctuations in lake level on the loss of salts operate around the periphery of a lake. The relative importance between areal and peripheral influences would tend to change with the square root of the area. The ratio of area in square miles to periphery in miles for the ocean is of the order of 600 to 1. For a small lake the ratio is reversed and of the order of 1 to 5.

Still another reason for discounting the influence of the loss of salts in aerosols from closed lakes is that this action is continuous, whereas the evidence seems to indicate that the loss of salts from lakes is associated with fluctuations in lake level.

EFFECT OF CHANGES IN LAKE VOLUME ON COMPOSITION

Closed lakes retain in solution the more soluble of the salts brought to it by the influent river water, while the less soluble salts, such as calcium and magnesium carbonates, are precipitated. The precipitates form much of the mud deposits on the lakebeds. Blinov (1956, p. 87), for example, reports that the gray mud in the bottom of the Aral Sea is composed of 69 percent calcium carbonate, primarily of chemogenic origin. Calcitic and dolomitic precipitates account for a large part of the more or less permanent losses from the lake solution as they do not easily reenter the solution. The remaining constituents form a lake solution of generally stable composition. Although river waters are considerably altered in composition by the ready precipitation of the less soluble salts, the present data indicate that the chemical composition of lake solutions varies but slightly with changes in lake volume. For example, the water volume of Devils Lake decreased by 90 percent during the period 1899 to 1948 and the concentration of salts increased threefold. The total amount of salt in solution decreased by 67 percent, yet the composition remained essentially uniform. The same was observed during the recession in water

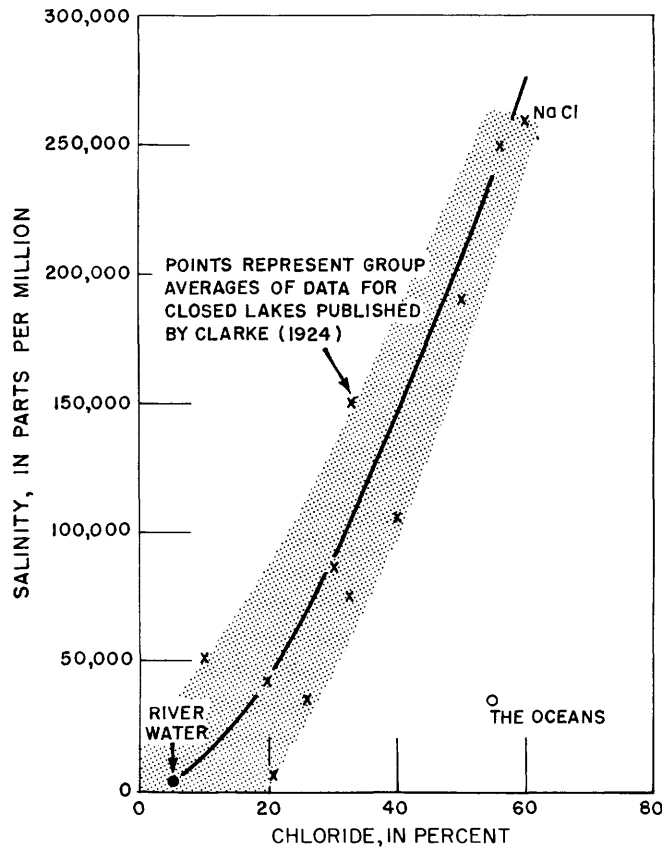


FIGURE 10.—Generalized graph showing variation in chloride with salinity.

level of Lake Eyre in 1950–52. The deposition of salts seems to affect all constituents alike. Therefore, the observation follows that the loss of salts results from complete evaporation, for otherwise there would have been a relative change in composition of the remaining water through fractional precipitation.

PERCENTAGE OF CHLORIDE IONS

Although the composition of a given lake may be stable, the composition among different lakes is variable. To demonstrate these differences, the data on total salinity of closed lakes assembled by Clarke (1924) were classified according to percentage of chloride and salinity. Group averages were computed and the results plotted on figure 10.

The line on figure 10 begins with a point representing river water (5.7 percent chloride, 150 ppm average salinity; after Clarke, 1924, p. 119). Among different lakes, the greater the salinity, the greater the percentage of chloride. With concentration of salts in closed lakes, those less soluble, such as the carbonates, tend to precipitate, increasing the relative fraction of chloride salts which tend to remain in solution. The band on figure 10 gives a qualitative indication of the decreasing variability in the percentage of chloride with increasing

total salinity. River waters are most variable in this respect, highly concentrated lakes are least variable. However, the point for the oceans on figure 10 shows a higher percentage of chloride than lakes of corresponding salinity. The generalized relationship on figure 10 appears to express the intensity of the selective precipitation of salts in favor of the chlorides.

APPLICATION OF LAKE DATA

SALINITY IN RELATION TO HYDROLOGIC PROPERTIES

If wastage of salts occurs, then there is a means whereby a lake can acquire a kind of stability between input and output of salts.

In reviewing the salt budget of Lake Eyre, Bonython (1956, p. 89) reaches a similar conclusion:

The theory here presented of the occurrence of the salt deposits in Lake Eyre is one in which a stream of what is principally airborne and surface waterborne oceanic salts constantly enters the lake, and another stream, of different composition to the first, constantly escapes from it, while in the lake itself lies a certain 'stock' of salts reflecting the equilibrium between the two streams. The composition of the stock is likely to have little resemblance to those of the incoming and outgoing streams as regards the proportions of the specific constituents, although the same constituents in greatly differing proportions are likely to be found in all three.

To verify the theory in the absence of direct observations on the input and output of salts requires at least a consistent relation between the salt content of closed lakes, their hydrographic properties, and their hydrologic environment. For this purpose, data on 25 closed lakes have been assembled from diverse sources. The data are listed in table 1. The entries are explained in notes that follow the table.

The hydrographic and hydrologic character of closed lakes can be described by four parameters: the ratio of annual net evaporation, E , to mean depth, D , or E/D ; the coefficient of variation of discharge from the lake, U ; the geometric shape as expressed by the ratio of the square root of the surface area, A_L , to mean depth, D , or $\sqrt{A_L}/D$; and the volume, V_o , required to fill the lake basin to the overflow level.

The hydrologic balance is a direct factor in the salinity of a closed lake. This balance is determined by the ratio of lake area to tributary area, which, in turn, determines the relative proportions of supply of relatively dilute rainwater to the more highly saline river water. Moreover, the greater this ratio, the greater is the volume of the lake for dilution of the salt input. The net effect can be expressed by the ratio of the annual salt input to the lake volume.

Thus

$$\frac{crA_T + c_pPA_L}{A_LD}$$

where c is the concentration of salts in the runoff, r ; c_p is the concentration of salts in the precipitation, P ; and the other terms are as before. But at hydrologic equilibrium $rA_T = EA_L$, hence the above ratio becomes

$$\frac{cE + c_p P}{D}$$

and if $c_p P$ be taken as negligible compared to cE , then the overall effect is expressed by the simple ratio

$$cE/D.$$

The ratio E/D is the ratio of the annual removal of water vapor from the lake surface to the mean lake depth, and thus may be viewed as a measure of the annual rate of increase in salinity.

The coefficient of variation of lake area, U , is related to the annual rate of loss of the salts, which for equilibrium balances the annual rate of increase in salinity by evaporation.

The geometric-shape factor describes the relative depth of the lake: a high value of the ratio $\sqrt{A_L}/D$ is indicative of a saucer-shaped lake, a low value indicates a deep lake.

Thus the salinity, C , can be tentatively expressed as

$$C \propto \left(\frac{E}{D}\right) \frac{1}{U} \left(\frac{\sqrt{A_L}}{D}\right)^x.$$

The ratio E/D is directly related to the salinity, whereas the variability, U , has a reciprocal relationship. However, the influence of the shape factor is less clear and is shown to a power x , which probably is less than unity but must be evaluated.

Evaluation of this equation using the data listed in table 1 gives

$$C = 180 \frac{E}{D} \frac{1}{U} \left(\frac{\sqrt{A_L}}{D}\right)^{\frac{2}{3}} \quad (13)$$

where E and D are in feet and A_L is in square feet.

The constant 180 represents a median concentration of lake constituents in river water in parts per million, as obtained from the Geological Survey annual reports on "Quality of Surface Waters for Irrigation, Western United States." As a basis for these computations, "lake constituents" in river waters were defined by the following arbitrary procedure.

Carbonate (as reported in equivalents per million) is first linked with calcium and magnesium, in that order. Any excess of the alkaline earths is combined with sulfate. Remaining sulfate is paired with sodium (plus potassium), and then excess sodium is linked to the chloride. So computed, the combinations of the

| | | | |
|----|-----------------|-----------------|-----|
| Na | .4 | .3 | 1.0 |
| Mg | .0004 | .9 | 1.3 |
| Ca | .00005 | .006 | 1.5 |
| | CO ₃ | SO ₄ | Cl |

FIGURE 11.—Relative solubilities of salts in distilled water. (NaCl=1.0 at 50° F.)

alkaline earths with carbonate and of calcium with sulfate are considered insoluble, as compared with the solubility of the remaining combinations. To illustrate the sharp contrast, compare the relative solubilities as indicated on the diagram of figure 11. There is more than a 100 to 1 contrast in the solubilities of compounds inside and outside of the shaded zone. Therefore all pairs within the shaded zone are considered "lake constituents."

As applied to the composition of water of rivers in the Western United States, the region where closed lakes occur, these calculations give a wide range of results, varying sharply with the aridity and lithology of the terrain with a median value of 180 ppm.

Equation 13 is easier to apply if modified so that lake area is in square miles:

$$C = 55,000 \frac{E}{D} \frac{1}{U} \left(\frac{\sqrt{A_L}}{D}\right)^{\frac{2}{3}} \quad (13a)$$

where C is in parts per million by weight, E and D are in feet, and A_L is in square miles.

The results can be improved further by including a factor representing the long-term stability of the closed lake. A closed lake ceases to exist when it dries up, and it is no longer a closed lake when it overflows. The shallower the lake, the greater the probability of its dessication. On the other hand, the deeper the lake, the greater is the probability of its overflow. The greater distance a lake lies below its overflow rim, the less the probability of its freshening by overflow. Although none of these probabilities can be determined, the long-term stability of a closed lake in a given basin

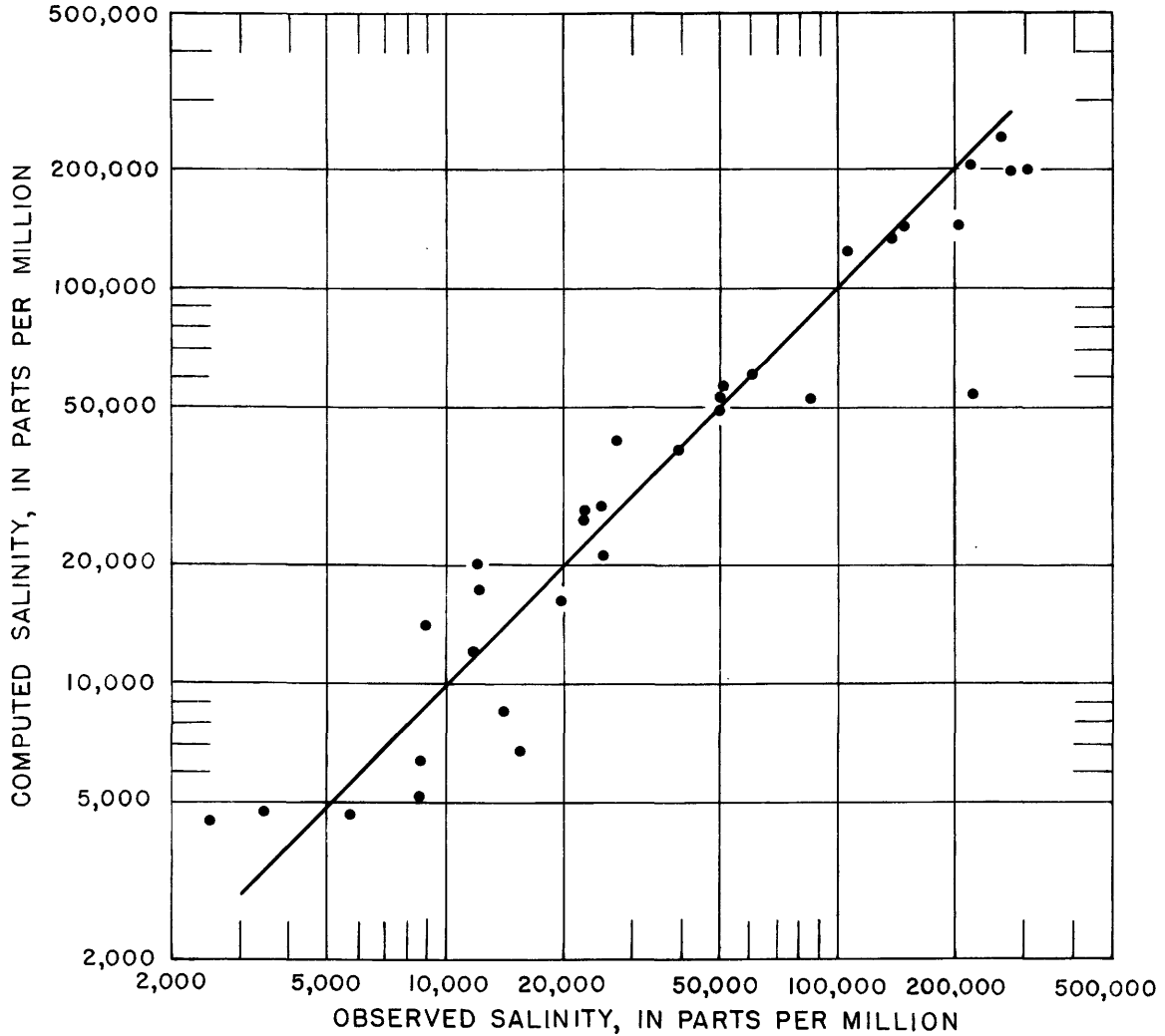


FIGURE 12.—Comparison between observed and computed salinity of closed lakes.

can be estimated in respect to its mean depth and to the overflow volume as follows:

$$\frac{D \frac{V_o}{A_T}}{D + \frac{V_o}{A_T}} = P \tag{14}$$

where D is the mean lake depth in feet, V_o is the volume above the lake to the level of overflow in cubic feet, and A_T is the tributary drainage area in square feet. Values of V_o/A_T as well as D are given in table 1. For any given closed basin, the value of the above function is 0 where either D or V_o is 0. In the first case there would be no lake, in the second the lake would be fresh. For any given basin the function reaches a maximum where the lake occupies somewhat more than half the volume of the closed basin below the level of overflow.

The effect of the function represented by P upon lake salinity was determined by investigation of the departures

of salinities as computed by equation 13a from the observed salinities. These departures indicated that the effect varies as follows:

$$0.55P^{1/6}10^{P/100} = W. \tag{14a}$$

The final formula for salinity therefore becomes

$$C = 55,000 \frac{E W}{D U} \left(\frac{\sqrt{A_L}}{D} \right)^{2/3}, \tag{15}$$

where C is in parts per million by weight, E and D are in feet, A_L in square miles, and P is the product defined by equation 14 where both D and V_o/A_T are in feet. A comparison between computed and observed salinities is plotted on figure 12. The coefficient of correlation is about 0.9+.

The correlation shown on figure 12 appears sufficient to justify the reasoning to this point. The results were obtained without any data on influent salts or reference to the composition of the lake solution.

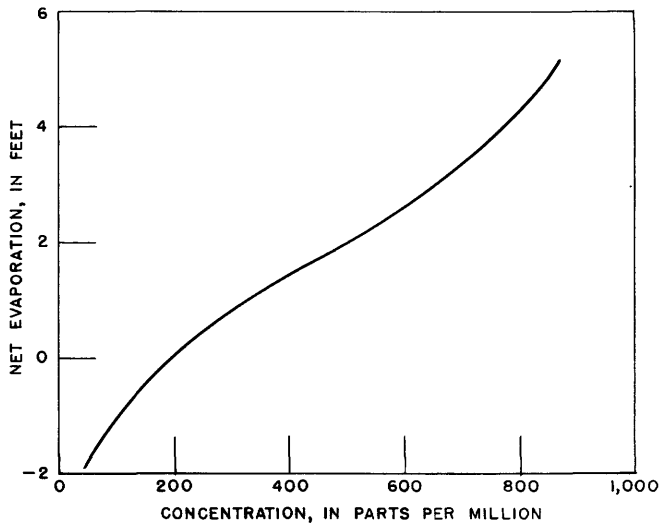


FIGURE 13.—Generalized association between concentration of dissolved matter in river water and regional net evaporation.

The scatter of points on figure 12 may be due in part to (1) inadequacies of the theory advanced in this paper, and (2) to necessary generalized assumptions made in the absence of specific data such as the lack of information on salinity of the inflow to the lakes, lack of records of fluctuations in lake levels, and, for most of the lakes, the lack of concurrent data on depths, area, and salinities.

Despite its seeming conformance with many of the lakes for which scant and often haphazard data are available, equation 15 should not be interpreted as explaining the salinity of any particular lake, owing to the chain of assumptions introduced into its derivation. The equation is intended only to express the possible effects of the hydrologic factors contained in that formula upon the salinity of closed lakes in general. Imperfections of the theory as well as other influences may dominate in any particular situation.

In this analysis no direct consideration has been given to the concentration of lake constituents in the influent water, despite the obvious influence of the salinity of the inflow on the concentration of salts in the lake. That the lake salinities seemingly can be explained even with this omission might be ascribed to the fact that the salinity of the influent water is itself related to the climatic environment and thus indirectly taken into account in the correlation expressed by equation 15. It is well known, for example, that the concentration of dissolved matter in river waters varies in a general way with the aridity of the tributary basin. Net evaporation is a measure of aridity. Figure 13, based on records of chemical quality, shows the association between concentration and net evaporation. Equation 14 therefore reflects the concentration of salts in river waters to the extent that the concentration is associated with net

evaporation as a measure of the climatic environment.

There are, of course, large variations due to other factors, among which the kind of rocks that compose the tributary basin is most important. The effects of such local variations on the salinity of river waters are, in turn, masked by the imperfections of the theory here advanced, except where such variations are very great. The River Jordan, the main affluent of the Dead Sea, for example, has a chloride content far above that normal for its climatic environment, because it cuts through deep and extensive salt deposits. The salinity of the Dead Sea greatly exceeds that indicated by equation 15. In the absence of data on the salinity of river waters that feed closed lakes, nothing beyond generalized concepts can be included in equation 15, which, as explained, contains factors that take into account the variation in salinity with the climatic factors.

Upstream lakes also tend to increase the concentration of salts in water that feeds saline lakes that occupy the lowest part of closed basins. The increase in concentration is proportional to the ratio of the net evaporation from the upstream lake to that from the closed lake. For example, the net evaporation (gross evaporation minus precipitation) from Lake Tiberias, about 70 miles upstream from the Dead Sea, averages about 220,000 acre-feet per year, or about 18 percent of that from the Dead Sea. Although no adjustment for this percentage increase in salinity has been made in the calculations, it illustrates another of the many factors that may influence the salinity of a closed lake in addition to those contained in equation 15.

So far as is known, rivers are the dominant source of water for the lakes listed in table 1. But ground-water inflow is doubtless present to some degree in all closed lakes. The salinity of ground water is apt to be higher than that of river water, a fact that is not taken directly into account by equation 14. Since inflow from ground water fluctuates in annual volume to a lesser degree than river flow, lakes fed in significant part by ground water should have relatively stable water levels, making the value of U less. According to equation 14, the lower value of U would be reflected in a higher value for the concentration of lake water. Thus, in indirect ways, the equation makes some allowance for the difference in the salinity of the water influent to the lakes. This allowance must, however, be deemed imperfect and only justified by the absence of adequate data upon which to base a better theory.

TIME FOR ACCUMULATION OF SALTS

As already made clear, the salt content of a closed lake is no guide to its age. It has also been suggested that the calculation of the age as the quotient of the mass of salts (or of one or more of the ions in solution)

divided by the annual input usually leads to a figure that is far short of a reasonable time since the time when the lake last overflowed. It would be instructive to derive several such figures—were data on salt input more readily available. But such data are still among the rarities of hydrologic statistics, for there appears to be little economic interest in salt transport at the mouths of rivers debouching into salt lakes. Some recent investigations of closed lakes will doubtless produce significant data. In the meantime, some of the theory presented in this report can be applied to the calculation of the period of years, T , represented by the total dissolved salts content of a salt lake, although it must be repeated that this figure does not represent the age of the lake. Thus

$$T = \frac{CDA_L}{cE\bar{A}_L} \quad (16)$$

where CDA_L represents the tonnage in solution and $cE\bar{A}_L$ the annual tonnage input of those constituents in the lake solution. The input concentration, c , is usually unknown, but it can be estimated by writing equation 14 in the form

$$C = c \frac{EW}{DU} \left(\frac{\sqrt{A_L}}{D} \right)^{\frac{2}{3}}, \quad (17)$$

where the concentration of lake constituents in the influent waters is written in lieu of the constant. Substituting the value of c defined by equation 17 into equation 16,

$$T = \frac{A_L W}{\bar{A}_L U} \left(\frac{\sqrt{A_L}}{D} \right)^{\frac{2}{3}}, \quad (18)$$

where A_L is in square feet and D in feet. Lake area, A_L , and mean depth, D , are to be taken at some high lake level, whereas \bar{A}_L is the average area over a period of years.

Equation 18 yields the following results for some representative closed lakes.

| | T (years) |
|----------------------|-------------|
| Devils Lake..... | 800 |
| Great Salt Lake..... | 6, 500 |
| Pyramid Lake..... | 1, 400 |
| Walker Lake..... | 900 |
| Owens Lake..... | 1, 700 |
| Lake Abert..... | 500 |
| Lake Eyre..... | 5, 500 |
| Caspian Sea..... | 20, 000 |
| Dead Sea..... | 30, 000 |
| Lake Van..... | 17, 000 |

There are few checks available that can be based on accurate figures of salt input, but some rough comparisons may be made. On the basis of nine samples of inflow to Devils Lake, as reported by Swenson and Colby (1955), the concentration of lake constituents in the influent waters is about 96 ppm. Using this figure in equation 16 yields a time period for the accumulation of the salts in solution of 2,200 years. Using the concentrations of sulfate, the dominant ion in the lake, the result is 2,100 years. In both cases, the result is nearly three times that computed above from equation 18.

Eardley and others (1957) computed the mass of salt ($\text{NaCl} + \text{Na}_2\text{SO}_4$) dissolved in Great Salt Lake and contiguous salt crusts at 5.9×10^9 tons. They also calculated the salt input to the lake as 1.1×10^6 tons per year, giving an accumulation time of 5,300 years, compared with 6,500 years as obtained from equation 18.

Some other rough comparisons with previous estimates can be made. Jones (1925, p. 31) calculated that the chloride content of Pyramid Lake could have accumulated in less than 4,000 years, a statement that is consistent with the 1,400 years derived by formula 18. Russell (1885, p. 226) estimated that the salt in Walker Lake accumulated in less than 500 years; the above estimate is 900 years.

Gale (1915, p. 264) concluded that the length of time taken to accumulate the mass of salts in Owens Lake "is in very general terms about 4,000 years, or possibly considerably less." Equation 18 gives 1,700 years as the accumulation time.

Van Winkle (1914, p. 123) computed that the salt contained in Lake Abert represents an accumulation time of 4,000 years, which Matthes (1942, p. 211) used to substantiate the estimated date as the beginning of the present "little ice age"—a moist, cool period when the lakes reformed after a long period of dessication. However, Van Winkle's calculations appear defective. A recomputation based on his data gives a period of about 2,000 years. The computed time would be even shorter if salinity records on Chewaucan River, the principal affluent, were available closer to the lake than 20 miles upstream at Paisley, Oreg., where Van Winkle's records were obtained. Field measurements of electrical conductivity in May 1959 indicated that there is a substantial increase in salt load between Paisley and the lake. The short period represented by the salt content of Lake Abert (500 years as computed from

equation 18) does not necessarily affect the date of the onset of the "little ice age" and the rejuvenation of the lake, but it does greatly affect the interpretation of the significance of the salt content of lakes in relation to their history.

Chilingar (1959), in a review of a Russian symposium on the formation of sediments in recent basins, reports an interesting calculation of the "age" of the Caspian Sea. On the theory that the Caspian Sea is a relict from the ocean, the time is computed for the influent river water to alter the ionic composition from that of the ocean to that of the contemporary lake. Using the ratio of chloride to sulfate, the "age of isolation" is computed as 10,600 years; using the chloride to magnesium ratio and the chloride to calcium ratio, the time period is 13,700 years. Both of these results are somewhat less than the time computed by equation 18, but since the theories are at variance with one another, correspondence may lack physical significance. However, sufficient data are available to compute the time period represented by sulfate ion which appears as a larger fraction of the salts dissolved in the Caspian Sea than in the ocean. Data indicate that the lake content of sulfate ions (including the Karabogaz Gulf) is about 240×10^9 metric tons and the annual input is about 18.2×10^6 metric tons. According to these figures, the time for accumulation of the salts in solution is 13,000 years, compared with 20,000 years obtained by equation 18.

HYDROLOGIC DATA

The lakes listed in table 1 are presumed to be closed basins—that is, there is no liquid discharge in surface streams or by seepage. Because all occupy topographic

sinks, it is assumed that the ground-water gradient is toward the lake.

Except for Devils Lake, Great Salt Lake, Owens Lake, Lake Eyre, and Lake Corangamite, for which survey results and frequent samplings were available, the data in table 1 have been obtained from diverse sources. The salinity data are generally from Clark (1924) who, however, gives no other hydrologic data. Therefore, information on areas, depths, and drainage areas was obtained from another source and may not correspond to the salinity data. For the deeper lakes, the changes in salinity with time are slow, but for the shallow lakes, the lack of correspondence of estimates or measurements of depth and lake area with the observation of salinity can introduce serious error. Concentrations of salt in the lake water at a low stage of a shallow lake can be much higher than at a high stage of the lake. Water temperature is another unknown factor.

Except as noted, salinity is reported in parts per million; that is, in milligrams per kilogram of solution. Salinity in milligrams of dissolved matter per liter of solution can be approximated from the following formula:

$$\text{mg per liter} = \text{ppm} (1 + 0.8 \text{ ppm}/10^{-6})$$

The coefficient of 0.8 is based on a comparison of salinities reported in milligrams per liter and in parts per million.

The values of the coefficient of variation of lake area are computed from equation 8, except as indicated in the notes to table 1.

TABLE 1.—Hydrologic data for closed lakes

| Lake | Drainage area (sq. mi.) | Evaporation (ft per year) | | Coefficient of variation of lake area | Response time (years) | Overflow expressed as depth over tributary area (feet) | Salinity | | Mean depth (ft.) | Lake area (sq. mi.) |
|--------------------------------|-------------------------|---------------------------|-----|---------------------------------------|-----------------------|--|------------------|----------------------|------------------|---------------------|
| | | Gross | Net | | | | Date | Ppm | | |
| Devils Lake, N. Dak.----- | 3,000 | 2.5 | 1.2 | 0.40 | 14 | 2.5 | 1899 | 8,470 | 13 | 45 |
| | | | | | | | 1923 | 15,210 | 10 | 26 |
| | | | | | | | 1948 | 25,000 | 4.5 | 14 |
| | | | | | | | 1952 | 8,680 | 10 | 20 |
| Basin Lake, Saskatchewan----- | 105 | 2.25 | 1.0 | .07 | 25 | ----- | 1938-41 | 11,900 | 20 | 16 |
| Quill Lakes, Saskatchewan----- | 2,700 | 2.0 | .75 | .15 | 20 | ----- | 1938-41 | 25,000 | 10 | 230 |
| Redberry Lake, Saskatchewan-- | 120 | 2.25 | 1.0 | .038 | 50 | ----- | 1938-41 | 14,000 | 43 | 27 |
| Great Salt Lake, Utah----- | 21,000 | 3.3 | 2.7 | .125 | 9 | 1.0 | 1877 | 138,000 | 18 | 2,200 |
| | | | | | | | 1932 | 276,000 | 13 | 1,300 |
| Sevier Lake, Utah----- | 16,000 | 3.7 | 3.2 | .35 | 3 | 4 | 1872 | 86,400 | 8 | 188 |
| Pyramid Lake, Nev----- | 2,650 | 4.2 | 3.7 | .04 | 65 | 9 | 1882 | 3,486 | 167 | 200 |
| Walker Lake, Nev----- | 3,500 | 4.2 | 3.8 | .075 | 45 | 13 | 1882 | 2,500 | 120 | 110 |
| Mono Lake, Calif----- | 600 | 4.1 | 3.3 | .043 | 35 | 200 | 1882 | 51,170 | 61 | 85 |
| Elsinore Lake, Calif----- | 717 | 4.5 | 3.2 | .68 | 3.0 | .3 | 1949 | 8,880 | 5 | 5 |
| Owens Lake, Calif----- | 2,900 | 5.5 | 5.0 | .10 | 10 | 12 | 1876 | 60,000 | 24 | 105 |
| | | | | | | | 1905 | 213,700 | 11 | 76 |
| Omak Lake, Wash----- | 100 | 3.2 | 2.2 | .067 | 30 | 19 | (¹) | 5,704 | 50 | 5.5 |
| Lake Abert, Oreg----- | 900 | 3.5 | 2.5 | .5 | 6 | 11 | 1902 | 76,000 | 5 | 50 |
| | | | | | | | 1912 | 30,000 | | |
| | | | | | | | 1956-59 | 20,000 | 10 | 60 |
| Summer Lake, Oreg----- | 330 | 3.5 | 2.5 | 1.0 | 2 | 61 | 1901 | 36,000 | 3 | 30 |
| | | | | | | | 1912 | 18,000 | | |
| Harney Lake, Oreg----- | 5,300 | 3.3 | 2.5 | .8 | 2 | 1.0 | 1912 | 22,380 | 4.8 | 47 |
| Lake Eyre, Australia----- | 550,000 | 7.5 | 7.0 | 2.5 | 1.5 | 10 | 1950 | ² 40,000 | 8.5 | 3,100 |
| | | | | | | | 1951 | ² 240,000 | 2.8 | 740 |
| Lake Corangamite, Australia--- | 1,300 | 4.0 | 2.0 | .30 | 10 | 3.8 | 1933 | 105,000 | 3.5 | 74 |
| | | | | | | | 1950 | 50,000 | 6.0 | 88 |
| | | | | | | | 1956 | 12,000 | 12 | 140 |
| Aral Sea, U.S.S.R.----- | 625,000 | 3.0 | 2.6 | .10 | 35 | 10 | ----- | 10,700 | 52 | 25,000 |
| Caspian Sea, Asia----- | 1,400,000 | 3.3 | 2.8 | .015 | 300 | 22 | ----- | 11,000 | 600 | 170,000 |
| Dead Sea, Palestine----- | 12,000 | 5.1 | 4.8 | .03 | 40 | 175 | ----- | 220,000 | 460 | 390 |
| Lake of Urmia, Iran----- | 20,000 | 3.0 | 2.5 | .11 | 9 | ----- | ----- | 148,000 | 16 | 1,800 |
| Lake Van, Turkey----- | 6,000 | 3.3 | 2.0 | .02 | 150 | 53 | 1944 | ² 22,400 | 175 | 1,450 |
| Tuz Golu, Turkey----- | 4,400 | 3.4 | 2.4 | .5 | 1 | 86 | 1959 | 250,000 | 2 | 650 |
| Elton Lake, U.S.S.R.----- | ----- | ----- | 3.0 | 1.0 | 1.0 | ----- | ----- | 300,000 | 2.3 | 110 |
| Baskuntschak Lake, U.S.S.R.--- | ----- | ----- | 3.0 | 2.0 | .5 | ----- | ----- | 260,000 | 1.15 | 50 |

¹ Before 1924.² Milligrams per liter.

SOURCE OF DATA

- Devils Lake, N. Dak.: Swenson and Colby (1955).
 Basin, Quill, and Redberry Lakes, Saskatchewan: Rawson and Moore (1944). The three lakes listed were selected from about 25 saline lakes described. Drainage area estimated by author from 1:500,000 topographic maps.
 Great Salt Lake, Utah: Woolley (1944), and Clarke (1924).
 Sevier Lake, Utah: Gilbert (1890, p. 225).
 Pyramid Lake, Utah: Depth and area from Harbeck (1951). Salinity from Clarke (1924).
 Walker Lake, Nevada: Depth and area from Harbeck (1951). Salinity from Clarke (1924).
 Mono Lake, Calif.: Depth from Hoyt and Wood (1913). Salinity from Clarke (1924).
 Elsinore Lake, Calif.: Harbeck (1951).
 Owens Lake, Calif.: Gale (1915).
 Omak Lake, Wash.: Depth and area from Pardee (1918). Salinity from Clarke (1924).
 Lake Abert, Oreg.: Waring (1908), Van Winkle (1914), and reconnaissance survey made in May 1959. Coefficient of variation of lake area computed from records of lake level.
 Summer Lake, Oreg.: Waring (1908), and Van Winkle (1914). Coefficient of variation of lake area is on basis of reported shifts in lake due to wind.
 Harney Lake, Oreg.: Waring (1909) and Piper, and others (1939).
 Lake Eyre, Australia.: Bonython (1955).
 Lake Corangamite, Australia.: Alexander and Sutcliffe (1956). Coefficient of variation of lake area computed from records of lake levels.

Aral Sea, U.S.S.R.: Depth and area from Woeikof (1909).

Value of U computed from lake levels reported by Blinov (1956, p. 19).

Blinov (1956, p. 99) states that the Aral Sea loses salt by seepage out of the lake through the sand deposits along the northern, eastern, and southern coasts of the lake. He adds "The multitude of shallow-water lakes along the eastern coast do not have any visible drainage into the sea as they are separated by more or less wide sandbars; but they do not dry up and do maintain a constant level during the entire warm period. The water in these lakes becomes highly saline as autumn approaches."

Aral Sea is therefore stated in effect to be a leaking lake, and Blinov further computes that outdrainage and subsequent evaporation of a quantity of water equivalent to 20 mm over the lake surface is sufficient to account for the loss of salts from the lake. But he does not explain the ultimate disposition of the order of several million tons of salts per year. Presumably the salts accumulate in and are subsequently lost from these offshore salinas.

The following equation can be written to describe the salt equilibrium in a leaking lake:

$$cEA_L = CwEA_L$$

where w is the part of the inflow that seeps out of the lake, and other terms are as before. The leakage, w , equals c/C . If $c=180$ ppm and $C=10,700$ ppm, a leakage rate of only 1.7 percent would account for the salinity of the lake. Leakage of 1.7 percent does not appear sufficient to be detected in an accounting of the water balance of the lake as Blinov attempts to do, because none of the items of that balance can be measured with the necessary degree of precision.

Caspian Sea, Asia: Depth and area from various encyclopedias, salinity from Clarke (1924). The Caspian is considered

a compound lake, the average salinity being a composite of that of the main sea and that of the Karaboghaz Gulf.

Dead Sea, Palestine: Depth and area from Lynch (1849) and by correspondence with Dov Nir of the Israel Institute of Technology. Salinity from Clarke (1924). Value of U computed from records of fluctuations in lake levels.

According to Ackroyd (1904), the high chloride load of the River Jordan is due to cyclic—that is, wind-carried—salts from the Mediterranean Sea. However, there is a material increase in chloride load downstream. Upstream at a point where the drainage area is 600 square miles, the chloride load is of the order of 20 tons per square mile per year. Downstream near its mouth (drainage area 6,300 square miles), the chloride load averages 60 tons per square mile per year. The downstream decrease in rainfall, and the opposite increase in chloride, does not appear to suggest rain-carried salts as the dominant source of the chloride. The source may rather be the extensive beds of rock salt and other chemical sediments in the lower River Jordan basin.

Lake of Urmia, Iran: Salinity from Clarke (1924). Depth and area from Encyclopedia Britannica. This Iranian lake appears to be almost a twin of Great Salt Lake. They have about the same drainage area, lake area, depth, rate of evaporation, and salinity. The composition of the lake water is also closely the same.

Lake Van, Turkey: Data furnished by H. H. Contürk, Ankara, Turkey.

Tuz Golu, Turkey: Data derived by author.

Elton Lake, U.S.S.R.: Grabau (1920, p. 200). The value of U is estimated on basis of Grabau's account. The lake has a high magnesium chloride content.

Baskuntschak Lake, U.S.S.R.: Grabau (1920, p. 202). Value of U is estimated on basis of Grabau's statement that it is a typical playa.

REFERENCES

- Ackroyd, Wm., 1904, On a principal cause of the saltiness of the Dead Sea: Chem. News, v. 89, p. 13.
- Alexander, G. N., and Sutcliffe, A. J., 1956, Hydrologic investigation of Lake Corangamite floodings: State Rivers and Water Supply Commission, Melbourne, Victoria. [Mimeographed]
- Blinov, L. K., 1956, Gidrokimiya Aralskogo Morya [Hydrochemistry of the Aral Sea]: Gidrometeorozdat, 251 p. [Only Chap. 3, "Salt Balance of the Aral Sea," was translated]
- Bonython, C. W., 1955, Lake Eyre, South Australia; The great flooding of 1949-50: Royal Geog. Soc. Australasia, South Australia Branch.
- 1956, The salt of Lake Eyre—its occurrence in Madigan Gulf and its possible origin: Royal Soc. South Australia Trans., v. 9, p. 66-92.
- Chilingar, G. V., 1959, Formation of sediments in recent basins: Internat. Geology Rev., v. 1, p. 105-111. [A review of a Russian symposium]
- Clarke, F. W., 1924, The data of geochemistry: U.S. Geol. Survey Bull. 770.
- Eardley, A. J., Gvosdetsky, Vasyl, and Marsell, R. E., 1957, Hydrology of Lake Bonneville and sediments and soils of its basin: Geol. Soc. America Bull., v. 68, p. 1141-1202.
- Erickson, E., 1958, The chemical climate and saline soils in the arid zone, in Arid zone research, climatology: Reviews of Research, U.N. Educ. and Cultural Organization, Paris, France.
- Gale, H. S., 1915, Salines in southeastern California: U.S. Geol. Survey Bull. 580, p. 251-323.
- Gilbert, G. K., 1890, Lake Bonneville: U.S. Geol. Survey Mon. 1.
- Grabau, A. W., 1920, Principles of salt deposition: 1st ed. New York, McGraw-Hill.
- Halley, Edmund, 1715, On the cause of the saltiness of the ocean, and of the several lakes that emit no rivers: Royal Soc. London Philos. Trans., v. 6, p. 169.
- Harbeck, G. E., Jr., 1951, Utility of selected western lakes and reservoirs for water-loss studies: U.S. Geol. Survey Circ. 103.
- Harbeck, G. E., Jr., Kohler, M. A., and Koberg, G. E., 1958, Water-loss investigations, Lake Mead studies: U.S. Geol. Survey Prof. Paper 298.
- Harding, S. T., 1935, Evaporation from large water surfaces based on records in California and Nevada: Am. Geophys. Union Trans., v. 16, pt. 2, p. 507-511.
- Harding, S. T., 1942, Lakes, in Meinzer, O. E., ed., Physics of the earth, pt. 9, Hydrology: New York, McGraw-Hill, p. 220-243.
- Hoyt, J. C., and Wood, B. D., 1913, Gazetteer of surface waters of California: U.S. Geol. Survey Water-Supply Paper 297.
- Hutchinson, G. E., 1937, A contribution to the limnology of arid regions: Connecticut Acad. Arts and Sciences Trans., v. 33, p. 47-132.
- 1957, A treatise on limnology: New York, John Wiley & Sons.

- Jones, J. C., 1925, *Geologic history of Lake Lahontan*: Carnegie Inst. Washington Pub. 352, p. 3-50.
- Kohler, M. A., Nordenson, T. J., and Baker, D. R., 1959, *Evaporation maps for the United States*: U.S. Weather Bureau Tech. Rept. 37.
- Lynch, W. F., 1849, *in* Report of the Secretary of the Navy transmitting official report on examination of the Dead Sea: U.S. 30th Cong., 2d sess., Senate Exec. Doc. 34.
- Lynch, W. F., 1849, *Narrative of the U.S. Expedition to the River Jordan and the Dead Sea*: Philadelphia, Pa., Lea and Blanchard.
- Matthes, F. E., 1942, *Glaciers*, *in* Meinzer, O. E., Ed., *Physics of the earth*, Pt. 9: New York, McGraw-Hill.
- Pardee, J. T., 1918, *Geology and mineral deposits of the Colville Indian Reservation*, Wash.: U.S. Geol. Survey Bull. 677.
- Meinzer, O. E., 1911, *Geology and water resources of Estancia Valley, New Mexico*: U.S. Geol. Survey Water-Supply Paper 275.
- Piper, A. M., Robinson, T. W., and Park, C. F., 1939, *Geology and ground-water resources of the Harney Basin, Oregon*: U.S. Geol. Survey Water-Supply Paper 841.
- Rawson, D. S., and Moore, J. E., 1944, *The saline lakes of Saskatchewan*: Canadian Jour., Research, v. 22, p. 141-201.
- Russell, I. C., 1885, *Geological history of Lake Lahontan*: U.S. Geol. Survey Mon. 11.
- 1896, *Present and extinct lakes of Nevada*, *in* Natl. Geog. Soc., *The Physiography of the United States*: New York, American Book Co.
- Swenson, H. A., and Colby, B. R., 1955, *Chemical quality of surface waters in Devils Lake Basin, North Dakota*: U.S. Geol. Survey Water-Supply Paper 1295.
- Thorntwaite, C. W., 1948, *An approach toward a rational classification of climate*: Geog. Rev., v. 38, p. 55-94.
- Van Winkle, Walton, 1914, *Quality of the surface waters of Oregon*: U.S. Geol. Survey Water-Supply Paper 363.
- Waring, G. A., 1908, *Geology and water resources of a portion of south-central Oregon*: U.S. Geol. Survey Water-Supply Paper 220.
- 1909, *Geology and water resources of the Harney Basin region, Oregon*: U.S. Geol. Survey Water-Supply Paper 231.
- Woeikof, A., 1909, *Der Aralsee und sein Gebiet nach den neuesten Forschungen*: Petermanns Mitteilungen, v. 55, p. 82-86.
- Woolley, R. R., 1944, *An engineer looks at Great Salt Lake*: Civil Eng., v. 14, p. 332-334.

