

Stratified lake and oceanic brines: Salt movement and time limits of existence¹

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Abstract

Pronounced salt concentration gradients in five antarctic, arctic, and Pacific coastal lakes can be accounted for by diffusional transport of salt out of the deeper saline water layers. The computed values of the mean salt diffusion coefficients, based on the ages of salinity stratification, agree to within an order of magnitude with molecular diffusivities for four out of five lakes. This agreement suggests that no major mixing events occurred in the water column during the late historical stages of the lakes. Upper limit time estimates for the removal of most of the salt from the saline bottom layers range from 5,000 to 35,000 years, depending on lake depth. Historical records of deepening of the Great Bitter Lake owing to dissolution of a salt layer on the bottom suggest that dissolution was a diffusion controlled process. For the saline brines in the Red Sea Deeps, an assumption that they are transient structures leads to the following estimates of the time to mixing with Red Sea water: 10^3 – 10^4 years, if mixing takes place by diffusional transport of salt between the heavier and lighter brines, and 10^4 – 10^5 years, if salt diffuses from the brines upward. The geologically short range of times suggests that the possible recycling of evaporative brines through the deeper ocean could not affect the ocean water salinity for any significant time interval.

A stratified water column consisting of a denser saline layer below a fresher water layer, with neither external input nor removal of salt, would eventually become homogeneous owing to dispersal of salt by molecular or eddy diffusion. Thus, many stratified brine lakes and ocean basins, where composition or salinity gradients are pronounced, are likely to be in a transient state with respect to the distribution of dissolved species. The differences in salt concentration between the denser and lighter layers in natural bodies of water range from a few hundred milligrams per liter to the upper limits of hundreds of grams per liter in salt-saturated brines overlain by freshwater.

Here we examine the vertical distribution of salt in a number of coastal brine lakes, arriving at the rates of salt transport through the water column, as well as the limiting estimates of the longevity of brine stratification. In some of the lakes we examine, the higher salinities of the deeper waters are related to entrapment of ocean water during the postglacial rise of the

land. We also consider one historically documented case of salt dissolution on the lake floor—the Great Bitter Lake on the Suez Canal. For an oceanic environment, we estimate the time limits of the possible existence of stratified oceanic brines, such as the Red Sea Deep brines. A corollary of this analysis is its contribution to the evaluation of possible changes of the oceanic salinity in the geologic past.

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Salt transport and time of mixing model

In a hydrodynamically stable water column, where the higher salt concentration (density) is near the bottom, the net transport of salt is from the bottom up. Our approach is to determine the rates of salt migration from the bottom up, using the reported vertical salinity distributions and ages of stratification in some saline lakes (Tables 1 and 2). In the lakes we consider, salt concentration profiles have been reported from single locations, so that the nature of the data dictates a one-dimensional approach to the problem. The validity of this approach rests on the horizontal dispersal being significantly stronger

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Table 1. Physical data summary for five saline lakes (Figs. 1-5).

Lake and Reference	Descriptive data
Lake Bonney, Antarctica (Shirtcliffe 1964; Shirtcliffe and Benseman 1964; Hoare et al. 1964)	Lake Bonney occupies the lowest part of Taylor Valley, Victoria Land, Antarctica (McMurdo Sound). The lake consists of two lobes connected by a narrow strait, 39 m wide and 8 m deep. Taylor Glacier at the western end of the lake is the main source of inflow. Inflow is apparently balanced by evaporation. The eastern lobe of the lake is 5.6 km long, 0.8 km wide, and it comprises the main body of the lake. Water temperature is a maximum of 7.5°C at 13 m depth, decreasing to -2.5°C at the bottom. The lake is permanently ice-covered, and possesses a dense saline layer at depth. Lake area: 4.5 km ² . Mean depth: 31.5 m. Mean water temperature: 2.5°C.
Lake Vanda, Antarctica (Wilson 1964)	Lake Vanda occupies a depression in the lowest part of Wright Dry Valley, Victoria Land, Antarctica. The lake is 8 km long and 1.6 km wide and has no surface outflow. It is supplied from the east by melt water from the Wilson Piedmont Glacier, via the 29-km-long Onyx River, which flows for about 60 days in the summer. Although the lake is permanently ice-covered, bottom water temperature is 25°C. Lake area: 12.8 km ² . Mean depth: 66.5 m. Mean water temperature: 9°C.
Lake Powell, B.C., Canada (Williams et al. 1961)	Lake Powell is a fjord lake, about 50 km long, 2 km wide, with a maximum depth of 358 m. The southern end of the lake is separated from the adjacent Strait of Georgia by a rocky sill. The lake consists of the upper and lower (south) basins. The lower basin is separated from the upper basin by an island and two narrow straits, appreciably shallower than the two basins. Nearly all of the important stream tributaries drain into the upper basin. Turbidity currents which may develop from stream discharge are likely to be trapped in the upper basin, leaving the lower basin free of influx of turbid waters at depth. The lower basin contains warm (9°C) saline water beneath 275 m of freshwater. The basin is 10 km long and 350 m deep, with a nearly horizontal floor. Temperature increases toward the bottom, from 2°C at 150 m to approximately 9.3°C at 350 m. Lake area: 100 km ² . Mean depth: 350 m. Mean water temperature: 6°C.
Northern Ellesmere Island lakes (Hattersley-Smith et al. 1970)	These coastal lakes, situated between fjords, have formed by post-glacial uplift that followed glacial deepening, allowing seawater to be trapped behind. The lakes contain an upper freshwater layer, underlain by a saline layer, with salinities of 30 to 32 g/kg near the bottom, which is close to the salinities in nearby fjords connected with the Arctic Ocean. The lakes are covered by 1.6 to 2 m of canded ice. They are drained for only two months during the summer. Temperature maxima occur in the two lakes: 7°C in lake A and 10°C in lake C at 16 m depth.
Lake A:	Area: 4.9 km ² . Mean depth: 57 m. Mean water temperature: 5°C. Uplift rate: 5.6×10 ⁻³ m yr ⁻¹ .
Lake C:	Area: 0.9 km ² . Mean depth: 57 m. Mean water temperature: 6°C. Uplift rate: 5.6×10 ⁻³ m yr ⁻¹ .

than the vertical dispersal in natural bodies of water in general (Stommel 1949; Pritchard et al. 1971), including chemically stratified and ice-covered lakes (Edmondson and Anderson 1965; Ragotzkie and Likens 1964).

If molecular diffusion is responsible for the migration of salt along a concentration gradient in water, the duration of the transport process would be long and comparable to the geologic ages of the lakes. Other,

faster, transport mechanisms (such as thermal convection and eddy diffusional dispersal) can accelerate the migration of salt and mixing of the lake water column. The existence of salinity gradients in the lakes where density stratification is an old feature (Tables 1 and 2) requires that the rates of salt migration be evaluated first with reference to the slowest possible transport mechanism—molecular diffusion. The question of whether the reported salt con-

Table 2. Parameters used in computation of concentration vs. distance curves in Figs. 1-5, and of the time limits for obliteration of salinity differences between the bottom and the top in five stratified lakes. Ages of stratification from references in Table 1. Concentrations in Lakes Bonney and Vanda refer to the chloride ion; in other lakes, total salts. Salt diffusion coefficients (D) evaluated from the concentration, age, and saline layer parameters as explained in the text.

Lake	Age of stratification (years)	Thickness of saline layer h (m)	Salt diffusion coefficient, D ($10^{-5} \text{ cm}^2 \text{ s}^{-1}$)	Concentration (g liter ⁻¹)			Time limit to removal of 95% of initial salinity difference, T (years)
				C_0	C_h	C_T (at $z=0$)	
Bonney	60	17.5	4.0±0.7 *	180	85	9.0	5,060
Vanda	1,000	12	0.63 †	84	27	4.2	11,500
Powell	10,000	58	1.26	18	12	1.6	29,000
A, Ellesmere Is.	600	41	2.8±0.3	33	11	1.6	34,800
C	400	47	22.2	33	6.3	1.6	5,100

* Shirtcliffe (1964) has used $1.25 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$, based on diffusion coefficients reported by Tyrrell (1961).

† Wilson (1964) has used $0.79 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$, based on the value for CaCl_2 .

centration profiles can be accounted for by molecular diffusion of dissolved species from the bottom up can be answered, and the presence of additional transport processes, as indicated by the results, will be discussed below.

The salt transport model for molecular diffusion of dissolved species applies to the following conditions:

1. Initially, a homogeneous saline layer was covered by a fresher water layer. Some mixing occurred at the upper boundary of the original saline layer, resulting in a concentration lower than in the deeper water.

2. Salt concentration at the upper boundary of the saline layer remained constant, after the initial flooding and mixing event.

3. There was neither input nor removal of salt through the lake floor.

4. Migration of salt in the lower saline layer takes place by molecular (ionic) diffusion.

Condition 2 can generally apply to a stratified lake if the less saline upper water layer is open to inflow and outflow on a long term time scale. Inflow and outflow may be either surface or subterranean, the latter difficult to observe directly (see Table 1 for summary of physical data on the

lakes studied). Condition 3 is supported if the salt concentration gradients near the lake bottom are close to zero. Condition 4 is, obviously, the one to be tested by the model.

The vertical transport of salt in the bottom layer will be represented by the conventional Fickian equation, with a constant diffusion coefficient (D):

$$\partial C / \partial t = D (\partial^2 C / \partial z^2), \quad (1)$$

where C is the salt concentration (mass vol⁻¹), z is the vertical distance coordinate increasing upward from $z = 0$ at the bottom to $z = h$ at the top of the saline layer, and t is time.

For Eq. 1, the initial and boundary conditions are

$$\text{at } t = 0: C = C_0 \text{ in } 0 < z < h, \quad (2)$$

$$\text{at } t > 0: C = C_h \text{ at } z = h, \quad (3)$$

$$\text{and } dC/dz = 0 \text{ at } z = 0. \quad (4)$$

Boundary conditions 2 and 3 are mathematical forms of conditions 1 and 2, given above. Boundary condition 4 is a mathematical form of condition 3.

The solution of 1, obtained by the

method of Laplace transformation using the boundary conditions 2-4, is

$$C = C_0 + (C_h - C_0)\phi_1, \quad (5)$$

where ϕ_1 is a function of time (t), position (z), and diffusivity (D); its complete form is

$$\phi_1 = \sum_{n=0}^{\infty} (-1)^n \left[\operatorname{erfc} \frac{(2n+1)h-z}{2(Dt)^{1/2}} + \operatorname{erfc} \frac{(2n+1)h+z}{2(Dt)^{1/2}} \right]. \quad (5a)$$

In 5a, the function erfc is defined as $\operatorname{erfc} x = 1 - \operatorname{erf} x$, where erf is the error function. Algorithms for computation of the error function and its complement (erfc) are given in many handbooks of mathematical tables. Also, algorithms for different ranges of the argument of $\operatorname{erfc} x$ have been summarized by Lerman (1971). The infinite series in 5a converges rapidly, such that only a few terms are needed for computation.

To compute salt concentration profiles using 5 and 5a, the time elapsed (t) and the constant parameters h , C_h , C_0 , and D must be known. (Reference to Figs. 1-5 and Tables 1 and 2 will facilitate reading the following section.)

The time t was taken as the reported (or estimated) age of stratification, given in Table 2. The thickness of the saline layer (h) was obtained by plotting the gradient of concentration against distance from the bottom. Such a plot passes through a maximum, and the distance from the bottom of this maximum gradient was chosen as the top of the lower water layer ($z = h$). This method of determining the effective height of the lower saline layer has been discussed and substantiated by measurements in Lake Bonney, a permanently ice-covered antarctic lake (Shirtcliffe 1964). Salt concentration at the top of the saline layer (C_h) was estimated from the reported concentration profiles. The initial salt concentration in the saline layer (C_0) was chosen to coincide with the highest value of the concentration near the bottom as observed in the lakes.

With the values of t , h , C_h , and C_0 , listed for each of the five lakes in Table 2, the value of diffusion coefficient D was estimated by trial and error computation using Eq. 5 and 5a. The values of D giving concentration profiles reasonably close to the reported profiles are indicated in Figs. 1-5. The validity and limitations of the model, as well as the effects that the choice of time (t) and initial concentration (C_0) have on the estimates of D , will be discussed below.

If migration of dissolved species from the lower saline layer proceeds without interruption, Eq. 5 can be used to estimate the length of time required to remove some large fraction of the salt initially present in the lower layer, or the time required to virtually obliterate the density stratification by diffusion. Estimates of these times for the five lakes given in Table 2 are subject to the same limitations as those involved in the estimates of D .

Discussion

It is intuitively clear that for a given salt concentration profile developed within a given period of time t , a higher initial concentration (C_0) would require faster diffusion to bring it down to the observed values or, in other words, a higher value of D . Alternatively, for a given initial concentration (C_0), a longer time (t) is compatible with a slower diffusion and a lower value of D . Given the approximate nature of the estimates of the age of stratification and the initial and boundary salt concentrations as listed in Table 2, one should not assign a better than order of magnitude significance to the values of salt diffusion coefficients derived from the model. Within an order of magnitude range, however, the values of D for four lakes listed in Table 2 fall in the range of 10^{-6} to 10^{-5} $\text{cm}^2 \text{ s}^{-1}$, characteristic of diffusion of salts in dilute aqueous solutions and seawater (Robinson and Stokes 1965; Li and Gregory 1974).

The value of D computed for Lake Bonney (Fig. 1), $(4.0 \pm 0.7) \times 10^{-5}$ $\text{cm}^2 \text{ s}^{-1}$, is somewhat higher than the diffusion coefficients of dissolved ionic species in water,

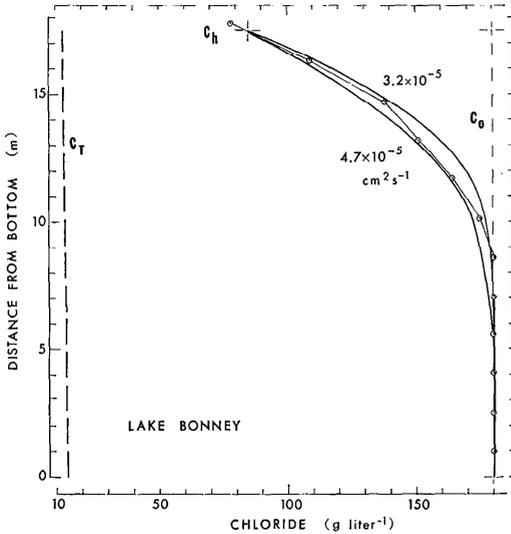


Fig. 1. Lake Bonney, Antarctica. Circles are measured chloride concentrations (Shirtcliffe and Benseman 1964). Concentration curves computed using Eq. 5 with the diffusion coefficient values shown in the figure. C_T is the concentration profile at the time T , when salt concentration near the bottom has been reduced to 5% of the initial difference between C_0 and concentration in the upper water layer. Additional data in Tables 1 and 2.

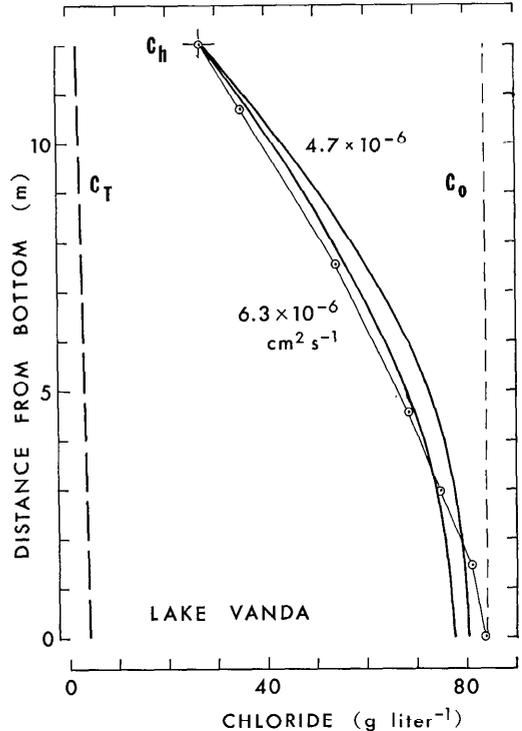


Fig. 2. Lake Vanda, Antarctica. Circles are measured chloride concentrations (Wilson 1964). Computed curves and other notation as in Fig. 1.

at room temperature and below. The computed value of D suggests that other transport mechanisms may have contributed to the salt migration process. However, since the calculated value differs from the reported values of D in aqueous salt solutions only by a factor of 3, the implication is that contributions from other processes were not significant. Alternatively, the large calculated values of D may be due to uncertainties in the parameters in Eq. 5 used to compute the salt profiles.

The way in which uncertainties in the age of stratification (t) and initial salt concentration (C_0) bear on the computed value of D is demonstrated in the following two cases. For Lake Bonney (Fig. 1, Table 2), the reported age of stratification $t = 60$ yr, used in the model, is compatible with a value of $D \approx 4 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$. If the age of stratification were twice as long, $t = 120$ yr, the computed diffusion coefficient would have been $D \approx 2 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$. The

latter value is close to $1.25 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$ used by other investigators (Table 2, footnote). For Lake Powell (Fig. 3, Table 2), the value of $D = 1.26 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$ is based on an age of stratification $t = 10,000$ yr and an assumed initial salt concentration of $C_0 = 18 \text{ g liter}^{-1}$. If the initial concentration in the saline layer was 20 g liter^{-1} , the diffusion coefficient describing the computed profile would be $D = 2.5 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$, and with the initial concentration closer to the salinity of ocean water, $C_0 = 31 \text{ g liter}^{-1}$, the diffusion coefficient would be higher, $D = 6.3 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$.

In the antarctic lakes Bonney and Vanda, migration of salt from the bottom up has been treated as a diffusional process by Wilson (1964) and Shirtcliffe (1964). Their models differ from ours insofar as they consider diffusion of salt as taking place from a thin saline layer near the bottom into an "infinitely" long column of wa-

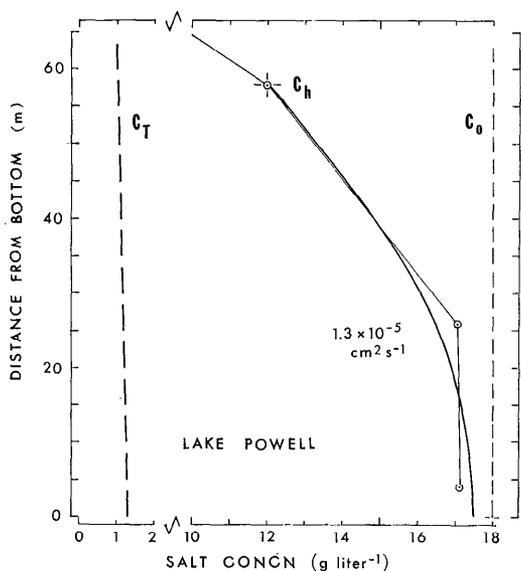


Fig. 3. Lake Powell, British Columbia. Circles are measured salt concentrations (Williams et al. 1961). Computed curves and other notation as in Fig. 1.

ter. The similarity of their and our values of the diffusion coefficients (Table 2) is because when the salt concentration difference between the bottom and top of the lake brine layer is large, the models produce approximately the same concentration against depth profiles, for a given time (t) and diffusion coefficient (D). But when the concentration gradients are not as strong as in the two antarctic lakes, the model for diffusion in a layer given by Eq. 5 is easier to rationalize in terms of the historical development of the lakes.

Irrespective of the uncertainties in the ages of stratification and initial salt concentrations in the lakes, an additional physical process may be responsible for the computed values of the diffusion coefficients that are higher than the molecular diffusivities in aqueous solutions, and this mechanism requires an explanation.

A stable water column in which the solute concentration decreases from the bottom up can be destabilized by superposition of a temperature gradient, with the temperature also decreasing from the bot-

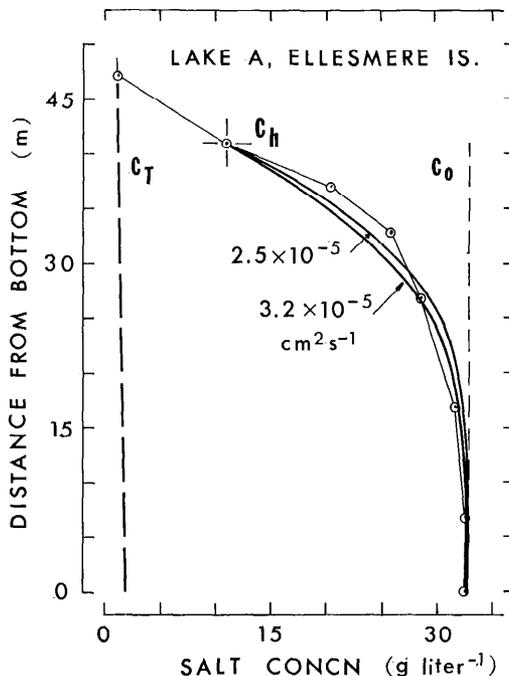


Fig. 4. Lake A, northern Ellesmere Island. Circles are measured salt concentrations (Hattersley-Smith et al. 1970). Computed curves and other notation as in Fig. 1.

tom up. The hydrodynamic relationships between temperature and salinity gradients in a water layer (Turner 1973) and physical characteristics of seawater solutions (e.g. Millero and Lepple 1972; compilation of data in Horne 1969) lead to the following relationship between critical salinity and temperature differences:

$$\Delta T \approx 7 \Delta S, \quad (6)$$

where ΔS is the concentration difference in units of grams per liter between the bottom and top of a brine layer, and ΔT is the temperature difference between the bottom and the top. Hydrodynamically, unstable conditions set in if the temperature difference between the bottom and top exceeds the ΔT value given by relationship 6. It also indicates that a salinity increase of 1 g liter⁻¹ can be offset by a temperature increase of about 7°C. In 6, $\Delta T > 0$ and $\Delta S > 0$. A water column is unstable if $\Delta T > 0$ but $\Delta S < 0$. If, however, $\Delta T < 0$ and $\Delta S > 0$, the water column is always

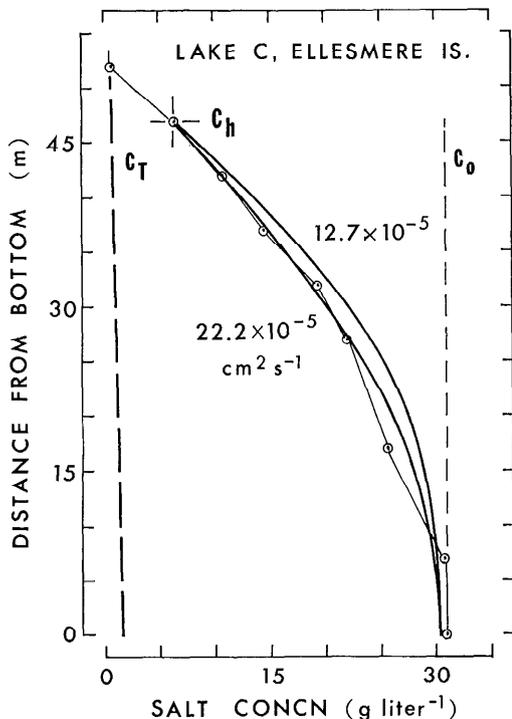


Fig. 5. Lake C, northern Ellesmere Island. Circles are measured salt concentrations (Hattersley-Smith et al. 1970). Computed curves and other notation as in Fig. 1.

stable because both the salinity and temperature gradients stabilize it.

In the five lakes dealt with in this paper (Figs. 1-5), the salinity gradients in the upper part of the saline layers control the density gradients: the reported temperature differences (Table 1) in the upper 5 to 10 m of the saline layers do not exceed 3°C, and these are too small to destabilize the upper part of the brine column. Near the lake bottom, however, salt concentration gradients are near zero, so that even weak heating from below may induce instability and turbulent mixing of the lower part of the brine layer. If this process takes place, the thickness of the layer in which salt migrates by molecular diffusion is smaller than the thicknesses used in our model (Table 2). Therefore, the computed values of the diffusion coefficients may be too high: in terms of Eq. 5, a smaller value of h requires a smaller value of D if other

parameters remain unchanged. In general, any perturbation or partial mixing of the water column (cf. Turner 1973) would lead to redistribution of salt in the lower brine layer at rates higher than molecular diffusional fluxes. The fact that the diffusion coefficients computed from the model are in the molecular domain or within one order of magnitude above it (Fig. 5) indicates that perturbations of the water column in the five lakes considered were not extensive and not frequent on the time scale of the lakes' lifetime.

As transport of salt from the lower water layers continues, the decrease in concentration gradients results in a lower hydrodynamic stability that may lead to onset of turbulent mixing. Such a process would accelerate salt transport from the lower to upper layer (cf. Lerman 1971), and it would obliterate the salinity gradients within the lake in a shorter time than if diffusion alone were responsible for salt transport. An upper limit estimate of the time required to reduce the initial concentration difference between the lower and upper layers by 95% can be obtained from the diffusional transport model, as given in Eq. 5. The times required to reduce the salt concentration from C_0 to C_T (Figs. 1-5) range, in round numbers, from 5,000 to 30,000 years (Table 2: T). These times are significantly longer than the ages of stratification reported for the lakes, and they are upper limit estimates, as mentioned above. Note that salt concentration decreases to the value lower than C_h in each of the lakes: the lower value is the reported salt concentration in the upper layer (Table 1) which was assumed to remain constant according to the conditions of the model.

Great Bitter Lake: salt dissolution—The Great Bitter Lake lies in the southern part of the Suez Canal connecting the Red and Mediterranean Seas (Fig. 6). At the time of construction of the canal (opened in 1869), a layer of salt existed on the bottom of what was to become the Great Bitter Lake. The saltbed covered about a third of the present area of the lake and con-



Fig. 6. View of the Great Bitter Lake, part of the Suez Canal, and Gulf of Suez. A freshwater canal extending east-west from the Nile is marked by vegetation. Photograph by Earth Resources Technology Satellite, No. 81074075415A200, taken 5 October 1972.

sisted of several-centimeter-thick layers of halite, interbedded with calcium sulfate and sand. From the published historical records of the rate of deepening of the Great Bitter Lake due to salt dissolution and concomitant changes in salt concentration in the lake (Wüst 1951; Morcos 1960), we obtain order of magnitude estimates of the rates of salt transport from the lake bottom to the overlying brine.

Distribution of salinities in the Great Bitter Lake reported by Morcos (1960) indicates that the lake is well mixed vertically. After the opening of the Suez Canal, salt concentration in the Great Bitter Lake, as well as in the canal as a whole, continuously declined from 1870 through 1935 (Wüst 1951); Krauss (1958) concluded that in the 1950s salinities had already become stable. Dissolution of the saltbed must have resulted in the removal of halite and, possibly, calcium sulfate minerals, leaving the sand behind. What has been left of the original bed thickness, stated by Ferdinand de Lesseps (quoted in Wüst 1951) as 13 m in 1868, is not known. However, the depth of the lake (presumably, the maximum depth) increased by 4 m over a period of 52 years, from 7.7 m in 1869 to 11.7 m in 1921, the decrease being linear with time (Wüst 1951).

Above the sediment-water interface, the possible existence of a thin brine layer within which molecular processes dominate is suggested by the analysis of current velocity and temperature gradients near the ocean bottom (Wimbush and Munk 1971). The thickness of such a viscous layer near a smooth bottom has been estimated as between 1 and 10 cm, depending on the current velocity and temperature regimes. Transport of salt through this boundary layer is by diffusion (Morse 1974). The insoluble mineral residue of the saltbed is likely to provide a porous cover layer, through which the migration of salt is also by molecular diffusion. A reasonable volume fraction of insoluble minerals (sand and silt) in a saltbed can be taken as 20%. Then, in the course of dissolution of 4 m of salt, the thickness of the sediment cover

Table 3. Physical data summary for Atlantis II and Discovery Deep brines, Red Sea.

Brine and references	Descriptive data
<p><u>Atlantis II</u> Deep (Bischoff 1969; Ross 1969; Turner 1969; Brooks et al. 1969; Erickson and Simmons 1969)</p>	<p>An elongate basin trending slightly NW-SE with an irregular bottom. The water column consists of the bottom brine, a pycnocline, intermediate brine, pycnocline, Red Sea bottom water. The salinity and temperature of pore water in the sediments of the Deep floor, beneath the brine, are essentially the same as in the bottom brine, although both positive and negative thermal gradients have been reported. Bottom brine layer: 50 m; 257 g kg⁻¹; 56°C. Lower pycnocline: 5 m. Intermediate brine layer: 28 m; 135 g kg⁻¹; 44°C. Upper pycnocline: 25 m. Red Sea water: 1,984 m; 41 g kg⁻¹; 22°C.</p>
<p><u>Discovery</u> Deep (Bischoff 1969; Ross 1969; Turner 1969)</p>	<p>A circular flat-floored depression, approximately 4 km SW of Atlantis II Deep. The lower brine layer in Discovery Deep is essentially identical to that of Atlantis II Deep, except for its lower temperature of 45°C. Both the temperature and salinity of pore waters beneath the Deep decrease. It has been suggested that the brine is not in a steady state (Turner 1969; Craig 1969; Pugh 1969). Bottom brine layer: 93 m; 257 g kg⁻¹; 45°C. Lower pycnocline: 15 m. Intermediate brine layer: 4 m; 127 g kg⁻¹; 36°C. Upper pycnocline: 37 m. Red Sea water: 1,986 m; 41 g kg⁻¹; 22°C.</p>

could increase from almost nil at the beginning to $0.2 \times 4/0.8 = 1$ m at the end, giving an average cover of 50 cm for the duration of salt dissolution. The mean rate of salt dissolution, using the values given in this section, is

$$(400 \text{ cm}) / (0.8 \times 52 \text{ yr}) \approx 10 \text{ cm yr}^{-1}.$$

The mean dissolution rate, equivalent to the flux of salt out of the sediment, can be used to obtain an order of magnitude estimate of the diffusion coefficient of salt within the water layer through which it migrates. Taking the layer thickness as 50 cm and the sediment porosity of 0.8 within the layer, the diffusion coefficient of salt is

$$D \approx (10 \text{ cm yr}^{-1})(50 \times 0.8 \text{ cm}) / (3.16 \times 10^7 \text{ s yr}^{-1}) \approx 1.3 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}.$$

This value is comparable to the diffusion coefficient of NaCl in its concentrated solution near room temperature, where the effect of sediment porosity on the diffusion coefficient is small (cf. Robinson and Stokes 1965; Li and Gregory 1974).

The computation of the rate of salt trans-

port given here suggests that dissolution of old saltbeds coming in contact with fresh or seawater can be a diffusion controlled process. The water layer through which diffusion takes place may consist of the viscous boundary layer near the bottom (Wimbush and Munk 1971), as well as the pore water layer in the sediments overlying the dissolving salt.

Red Sea brines—The *Atlantis II* and *Discovery* Deep hot brines are situated along the central Red Sea rift valley, about midway between the Strait of Bab El Mandeb and Port Sucoz, the southern and northern extremities of the sea. The deeps contain two homogeneous brine layers separated by sharp and well defined pycnoclines. The brines are chemically and thermally stratified, with both the higher salinity and temperature occurring in the lower brine layer. (A brief summary of the relevant data on the brines and the deeps they occupy is given in Table 3.) Stratification and distribution of chemical species in the Red Sea Deep brines have been discussed in terms of a steady state model by Craig

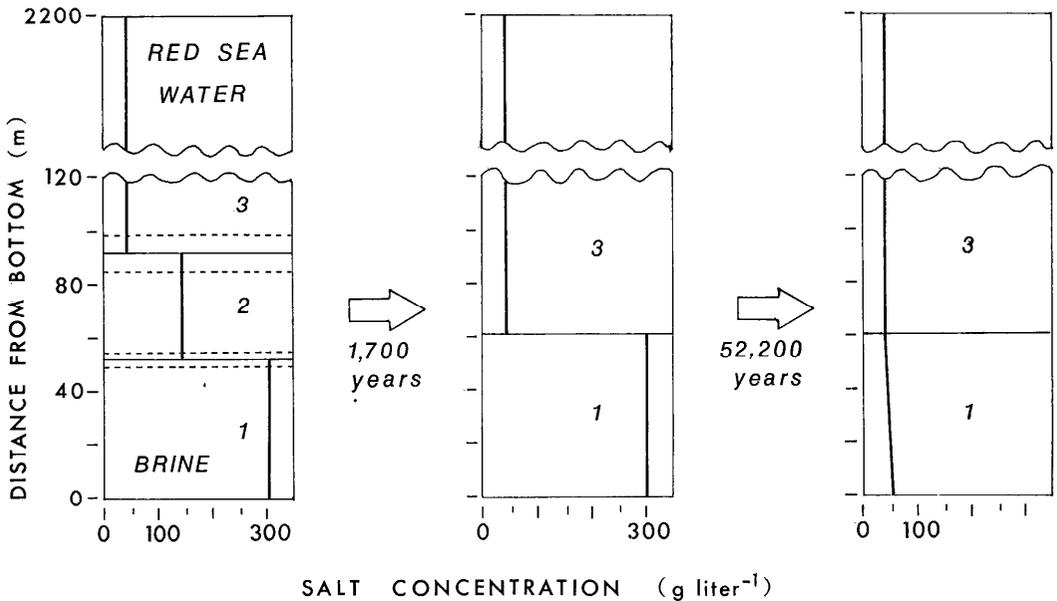


Fig. 7. Schematic representation of a Red Sea Deep brine. Layer 1—bottom brine; layer 2—intermediate brine; layer 3—Red Sea water. Dashed lines denote the pycnocline layers. Transition time from the three-layer to two-layer model (1,700 years) and time limit of further reduction in salinity of the bottom brine are as given in Table 4 (T_1 and T_2 for *Atlantis II* Deep).

(1969), who also gave the brines a post-glacial age of about 12,000 years.

Following the discussion in the preceding sections that dealt with migration of salt by diffusion, we present here upper limit time estimates for mixing of the saline brines with the overlying Red Sea water. This approach implies that the brines are transient features and continuously lose their salt to the Red Sea waters. Whether the deep brines are in fact steady state or transient features is of secondary significance, insofar as the computation of the time limits of their existence is aimed at the broader problem of salt storage and cycling in the ocean.

Schematically, the vertical structure of the *Atlantis II* Deep brine is: bottom brine—pycnocline—intermediate brine—pycnocline—Red Sea bottom water. This layered structure was approximated to three layers by apportioning the pycnoclines between the adjacent brine layers, as shown in Fig. 7; the resulting model includes bottom brine—intermediate brine—Red Sea water.

If each of the three layers is well mixed, a denser layer may be gradually eroded by a lighter layer from above, by analogy with the experiments showing descent of an interface between a less saline well mixed layer and a more saline layer below it (Turner 1973). In such a process, the salt concentration in layer 3 (Fig. 7) will remain unchanged, because the mass of the Red Sea is so large in comparison with that of the brine layers. Also, the salt concentration in layer 1 can be assumed to be constant because its erosion from above is likely to be much faster than any diffusional supply of salt from the pore waters of the sediments (pore water salinity data given by Brooks et al. 1969). The salt concentration in layer 2, however, and the thickness of all layers will be changing. Their values can be obtained by solution of the following system of three simultaneous equations:

$$dH_1/dt = -k_1(C_1^* - C_2)/C_1^*, \quad (7)$$

$$dH_3/dt = k_2(C_2 - C_3^*)/C_3^*, \quad (8)$$

Table 4. Red Sea brines. Estimates of the time limits of brine existence according to the three- and two-layer models (T_1 and T_2). The nature of the brine layers and other parameters are discussed in the text and Fig. 7. For the hot brines (45° – 55°C), the salt diffusion coefficient was taken as $D = 3.2 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$ in the three- and two-layer models.

	Brine layer thickness (m) and salt conc. (g liter^{-1})			Pycnocline thickness (m)		Rate constants (cm yr^{-1})		Time limit for intermediate brine, T_1 (years)
	(H_1) C_1	(H_2) C_2	(H_3) C_3	ΔH_{12}	ΔH_{23}	$k_1 =$ $D/\Delta H_{12}$	$k_2 =$ $D/\Delta H_{23}$	
Three-layer brine model (diffusional fluxes between well mixed layers)								
<u>Atlantis II Deep</u>	(52) 307	(40) 147	(2,000) 43	5	25	2.0	0.4	1,700
<u>Discovery Deep</u>	(100) 291	(30) 120	(2,000) 43	15	37	0.7	0.3	2,900
Two-layer brine model (diffusion of salt through lower layer)								
	(62)		(2,000)					Time limit for removal of 95% salinity difference between bottom and top layers, T_2 (years) 50,500
<u>Atlantis II Deep</u>	$C_{z=0} = 307$ $C_h = 43$ (106)		43 (2,000)					147,000
<u>Discovery Deep</u>	$C_{z=0} = 291$ $C_h = 43$		43					

$$dC_2/dt = (1/H_2) \{ [k_1(C_1^* - C_2)^2/C_1^*] + [k_2(C_3^* - C_2)^2/C_3^*] \}, \quad (9)$$

and

$$H_T = H_1 + H_2 + H_3, \quad (10)$$

where H_T is the total thickness of the water column (constant), H_1 , H_2 , and H_3 are the thicknesses of the three brine layers, C_1^* and C_3^* are constant concentrations in the bottom and top layers, C_2 is concentration in the intermediate layer, and k_1 and k_2 are rate constants (dimensions of length/time) of salt fluxes from layer 1 to 2, and from 2 to 3, respectively.

The solution of Eq. 7–10 is conveniently done by numerical integration on a computer. In the model, the fluxes of salt between the adjacent layers are proportional to the differences in salt concentration. The values of the transport coefficients k_1 and

k_2 , as well as the other parameters, are given in Table 4. k_1 and k_2 are based on a molecular diffusion coefficient of NaCl in water at 55°C ($D = 3.2 \times 10^{-5} \text{ cm}^2 \text{ s}^{-1}$), as calculated from the limiting ionic conductivities (Robinson and Stokes 1965), and pycnocline layer thicknesses shown in Table 4. As diffusion of ionic species in solution is much more sensitive to temperature than variations in concentration, the salinity differences between the brine layers would have only a small effect on the estimated value of D .

The salt flux from the bottom to the intermediate layer, $k_1(C_1^* - C_2)$, is greater than the flux from the intermediate to the upper layer, $k_2(C_2 - C_3^*)$. Therefore, the salt concentration in layer 2 increases while its interfaces descend at the expense of layers 3 and 1. Taking the present state of the

brine as $t = 0$, at time $t = 1,700$ yr concentration in layer 2 becomes equal to the concentration in layer 1, such that the two layers coalesce (Fig. 7). At that time, the combined remainders of layers 1 and 2 form a brine layer thicker than the original layer 1. Whatever the shortcomings of the assumptions underlying the model, the brief existence of the three-layer brine structure, as computed from the model, underscores the transient nature of such brines in the longer term evolution of brine basins in general.

After the water column has been reduced to two major units—a brine layer and less saline seawater on top—two limiting time estimates of its longevity can be obtained by the following procedures.

One, shorter, estimate, is based on the rate of salt transport by diffusional flux from the lower to the upper layer, when each layer is considered well mixed. For an order of magnitude estimate, with the brine layer thickness (H_1) and the rate constant of salt flux (k_1) from Table 4, the time to obliteration of the brine in *Atlantis II* Deep is $H_1/k_1 = 60/0.02 = 3,000$ yr. A similar estimate for *Discovery* Deep is $100/0.007 = 14,000$ yr.

Another, longer, estimate can be based on the diffusion model as applied to the stratified saline lakes earlier in this paper. In contrast to the first alternative, in this diffusion model the saline brine is not well mixed. From the diffusion model, the time required to reduce the salt concentration near the bottom to the value of 5% of the initial concentration difference ($C^*_{1} - C^*_{3}$) is of the order of 50,000 yr, as shown in Fig. 7 for the *Atlantis II* Deep. For *Discovery* Deep, with a brine layer about 1.7 times as thick, the length of time to removal of 95% of the original salinity difference is about 150,000 yr. These time estimates are based on the value of $D = 3 \times 10^{-5}$ cm² s⁻¹ that applies to diffusion of salt in a brine at ca. 55°C. Had a smaller value of the diffusion coefficient ($D \approx 1 \times 10^{-5}$ cm² s⁻¹, characteristic of ca. 20°C brine) been used, a longer time to mixing (by as much as a factor of 3) would have been obtained. It

should also be noted that a temperature difference of 20° to 35°C between the bottom and top of the saline layer is not great enough to destabilize the effects of the 10 g kg⁻¹ salinity difference that would exist between the bottom and the top after the initial salinity difference has been reduced by diffusion to 5% of its initial value.

Conclusions

Vertical migration of salt in a number of salinity-stratified lakes can be accounted for by molecular diffusion. An order of magnitude similarity between the computed salt diffusion coefficients in lake waters and those reported for aqueous solutions indicates that the model of diffusional transport is compatible with the geological ages of some of the lakes (Figs. 1–4). The diffusional transport of salt, as derived from the data reported for the lakes, underscores the continuity of physical processes and the absence of *extensive* intermittent mixing during the lifetime of the lakes.

In two of the five lakes studied (Figs. 1 and 5), the computed values of the diffusion coefficients exceed to a lesser or greater extent the values characteristic of molecular diffusion. Such departures from molecular transport generally indicate the effects of turbulent mixing owing to partial destabilization of the water column by temperature gradients or other modes of thermally induced convection.

The rates of salt migration from the lake bottom up, derived from a diffusion model, are the slowest possible; consequently, the time required to reduce salt concentration in the deeper waters of a lake by some amount, based on the diffusion model, is an upper limit estimate. For the five lakes considered, the times required to obliterate the salinity differences between the deeper and surface waters range from 5,000 to 35,000 years (Table 2), and these times are significantly longer than the geologic ages of salinity stratification in the lakes.

The diffusional transport model was also applied to the saline brines of the Red Sea occupying the *Atlantis II* and *Discovery* Deeps. If the brines are transient features,

losing their salt content to the overlying Red Sea water, then the upper limit estimates of time required to reduce the salinity differences by 95% are of the order of 50,000 and 150,000 years (Table 4). A lower bracket estimate of time, based on diffusional fluxes of salt between the well mixed saline brine and Red Sea water reservoirs, gives the times of 5,000–16,000 years.

Even the longest estimates, based on the diffusional transport model, indicate that mixing of brine pools on the ocean floor with the overlying ocean water is a fast process on the geologic time scale. For example, a brine layer 50 m thick, of salinity comparable to the Red Sea brines (257 g kg⁻¹) and overlain by normal seawater (35 g kg⁻¹), would lose 95% of the salinity difference in about 10⁵ yr ($D = 1 \times 10^{-5}$ cm² s⁻¹). Doubling the brine thickness increases the time by a factor of 4. It has been occasionally suggested (e.g. Fischer 1964) that the salinity of the world ocean could be temporarily reduced by the process of strong evaporation in coastal areas and subsequent flow and storage of saline brines in depressions on the ocean floor. A reduction of the ocean water salinity by 5 g kg⁻¹ over a period of 20 million years [a value used by Fischer (1964) as an example] is equivalent to storing salt in a brine layer of salinity 257 g kg⁻¹, 75 m thick over the entire ocean floor. This mechanism would be possible if the rates of mixing of saline brines and ocean water were effective on the time scale of 10⁷ yr. But, as discussed here, even the slow diffusional migration of salt operates on a shorter time scale of 10⁵ yr, and this makes the possibility of prolonged brine storage doubtful.

Historical records of the deepening of the Great Bitter Lake, owing to salt dissolution on the lake floor, can be accounted for by dissolution of a salt layer in contact with seawater and diffusional transport of salt through a layer ca. 50 cm thick near the top of the lakebed. The rate of dissolution of a salt layer, of the order of several centimeters per year, can serve as a first-approximation value to the dissolution of

older saltbeds exposed to marine transgressions.

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