

# Geochemical Modeling of the Effects of the Proposed Red Sea-Dead Sea Canal

Nizar Abu-Jaber\*

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## Abstract

Expected changes in the chemical composition of Dead Sea (DS) water as a result of the implementation of the Red Sea-Dead Sea canal are modeled. Using historical data as well as knowledge of the chemistry of the canal water, thermodynamic equilibrium models using PHRQPITZ are constructed. The results show that massive deposition of gypsum is to be expected. Moreover, the reestablishment of a stable pycnocline as expected by other researchers will not occur. The DS will reach supersaturation with respect to halite in about 50 years.

## Introduction.

Population growth and social and economic development in the Jordan River and Dead Sea (DS) basins in the twentieth century has led to serious environmental problems in the area. In particular, the need for fresh water has led to near complete utilization of these resources in Jordan, Palestine and Israel, and in many cases over utilization.

The hydrological balance of the DS used to depend on surface water inflows that were roughly equal to evaporation over the water body. This balance has now been disrupted because of the headwater diversions that are taking place in the basin. It has been estimated that about 1500 Million Cubic Meters of water/year (MCM/a) have been diverted, causing a significant drop in the water level in the lake as well as in the adjacent aquifers.

Because it is unfeasible to re-establish the hydrological regime to its original state, other initiatives have been proposed in order to restore the hydrological balance of the DS. Specifically, a canal or a pipeline connecting either the Mediterranean or the Red

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\* Department of Earth and Environmental Sciences, Yarmouk University, Irbid, Jordan.

Seas with the DS has been suggested (Beyth, 2002), with the Red Sea–Dead Sea (RS-DS) canal project now favored from both political and environmental perspectives. Such a project would both help restore the DS to its original level and would utilize the 400m difference in water level for hydroelectric generation or desalination. Naturally, there is a strong economic imperative to implement this project.

The purpose of this paper is to explore the hydrochemical implications of this proposed project. While the hydrological balance issues are reasonably well understood, the chemical issues related to the mixing of two brines in the DS have received little attention. The chemistry of the DS is the main resource on which potash and other chemical extracts are based on both shores of the sea. Moreover, the tourism industry relies on the aesthetic value of the sea, which in turn partially relies on the clarity of the water.

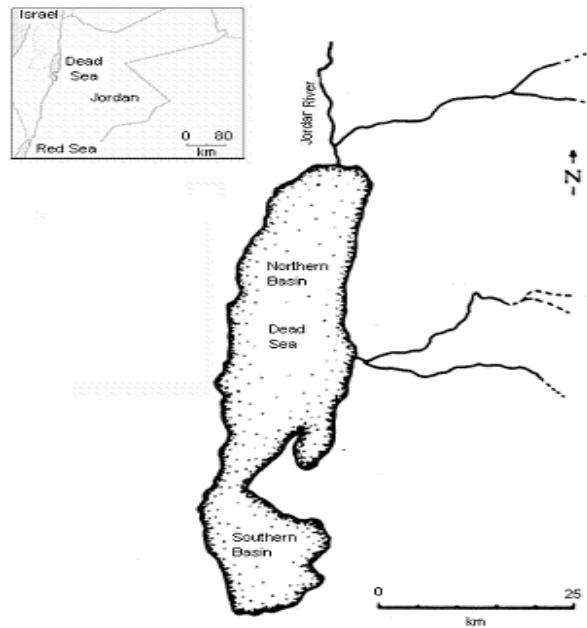
Thus, it would be useful to conduct some geochemical modeling in order to assess the nature and volume of precipitates that might be formed, and potential changes to the water chemistry that might be caused by such a massive mixing experiment. In this paper a numerical modeling approach will be applied in order to assess the effects of the brine mixing on future evolution of the chemistry of the DS.

### **Geology and Limnology of the DS**

The DS lies at the lowest continental topographic point within the East African-Red Sea-Jordan Valley rift system (Fig.1). This system defines the boundary between the African and the Arabian tectonic plates. In the DS area, the rift system is a left-lateral strike-slip fault system characterized by a series of rhomb-shaped grabens (Garfunkel, 1981). Total lateral displacement is estimated to be about 106 km, with vertical displacements defined by the topography and post-rift sediment thickness (Manspeizer, 1985). In the DS area, the lowest point is about 800 m below sea level. Post-rift sediment thickness in the DS area is estimated to be about 12 km at some locations.

This rift system was initiated during the Miocene, resulting in a thick sequence of sediments that range from that age to the present (Bender, 1974). These sediments vary from the thick salt deposits of the Usdum formation (Miocene) to fresh water lacustrine sediments and conglomerates (Shaghour, Ghor el Katar, Kufranja and Abu Habil formations)/(Bender, 1974). The variations seen in the types of sediments reflect changes in tectonism and climate during the history of the DS (Manspeizer, 1985). During the Holocene, the modern Dead Sea evolved.

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**Figure 1.** Location map of the Dead Sea.

The DS consists of two sub-basins; a deep (400m) northern basin and a shallow (currently exposed) southern basin. The southern basin was exposed due to the drop in sea level, but it is now used for solar ponds in which DS water is further concentrated for the chemical plants of the Arab Potash Company (Jordan) and the Dead Sea Works (Israel). The total area of the historical DS was about 1000 km<sup>2</sup>, and is now about 700km<sup>2</sup> (Hall, 1997). The volume of water in the sea currently stands at 128km<sup>3</sup>, down from an initial volume of about 155km<sup>3</sup> at the turn of the twentieth century (Neev and Emery, 1967). This can be seen in the continuous drop in the level of the DS, which has accelerated to a net rate of about 80 cm/a at the end of the twentieth century (Anati and Shasha, 1989). During historical times, the hydrological balance in the DS was maintained through input from the Jordan (about 1207 MCM/a), rivers east of the DS (about 219 MCM/a) rivers west of the DS (163 MCM/a) and from intermittent streams from the south of the DS (81 MCM/a; Salameh and El-Naser, 1999). Currently, it is estimated that submarine stream discharge contributes between 200 to 370 MCM/a to the current lake (Abu-Jaber, 1998; Salameh and El-Naser, 2000).

The flow of these fresh water sources into the DS led to the formation of a stratified lake, with a deep, anoxic dense fossil water layer and an upper, less dense renewable mixed layer. Initially, the upper layer was about 40m deep, but as the imbalance between inflow and evaporation grew, this layer became deeper but less pronounced, until the final mixing event which occurred in 1979 (Carmi et al., 1984). The physical parameters controlling the relationship between the upper and lower layers and the stability of the pycnocline have been discussed by Anati and Stiller (1991).

In the event of the implementation of the RS-DS canal, it has been estimated that a rapid reestablishment of the pycnocline will occur (Asmar and Ergenzinger, 2002). This is in line with the rapid reestablishment in layered conditions following unusually rainy seasons (Anati et al., 1987; and Anati and Stiller, 1991).

The DS is characterized by a unique chemical composition. It is basically a Ca-Cl<sub>2</sub> brine which is highly enriched in K<sup>+</sup>, Mg<sup>2+</sup> and Br<sup>-</sup>. SO<sub>4</sub><sup>2-</sup> and HCO<sub>3</sub><sup>-</sup> are relatively low (Neev and Emery, 1967). This has led to speculation as to the sources of the salts. Starinsky (1974), Vengosh et al. (1991), Stein et al. (1997) and others consider the source of the salts to be of marine origin, with appropriate modifications to explain the discrepancy between real evaporated seawater and the actual DS water. Others have suggested a deep continental origin for the brines (Hardie, 1990 and Abu-Jaber, 1998). This controversy is of little direct relevance in the context of this work. Abu-Jaber (1998) suggests a continuous input of brine into the DS is needed to reach the currently observed chemical conditions, whereas the marine brine model would suggest that the sources of salts have long been terminated. However, since it is expected that while the canal input will affect the upper layer (because of its supposed relatively low density), the supposed brine seeping to the DS will stay in the deep layer owing to its higher density.

### **Rationale and Approach**

The geochemical modeling of the expected processes related to the mixing involves estimations of the mixing ratios, input water chemistry, and evaporation. After these parameters are reasonably quantified, thermodynamic equilibrium models are constructed in order to determine what mineral phases are expected to become supersaturated in the resulting new brine, and the mass of the phases precipitating from the brine. This is done using the brine equilibrium software package PHRQPITZ (Plummer et al., 1988). PHRQPITZ is a computer code specifically designed to evaluate thermodynamic states of various minerals in brines. This is done using the Pitzer equations to evaluate activity coefficients of various solutes under high ionic strengths, as was demonstrated by Harvie and Weare (1980). The Pitzer equations were integrated into the old thermodynamic code known as PHREEQE. While PHREEQE has been superseded by more user-friendly codes, PHRQPITZ is still valid and ideal for the study of thermodynamic conditions in brines, given the lack of similar modifications on the newer codes for dealing with brines.

Most researchers expect that the mixing between the canal water and the DS water will not be a complete mix. It is expected that a new upper layer defined by a clear pycnocline will develop (Asmar and Ergenzinger, 2002). Historical layering of the DS did not preclude the presence of similar chemistries between the upper and lower layers (See Neev and Emery, 1967). Thus, while the fresher waters tend to float above the deeper water, some mixing is expected between the two water types, if only in the initial stages of the formation of the pycnocline. This can be surmised from historical analysis of how fresh water flowing into the DS behaves. Analysis of the physical stability of the established pycnocline shows that little flux of water, solutes or heat can be expected across the pycnocline interface (Anati and Stiller, 1991). In the short term, the ratio of

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the canal water is expected to be high at the discharge point. However, the mixing ratio between the canal water and the DS water is not obvious. In order to make reasonable assumptions about the ratios of the different waters in the expected upper layer, it is necessary to study the variation in chemistry between the upper and lower layers in the historical lake. Thus, the modeling will include a number of steps defined by assumptions that are derived from historical data.

The first step, however, will be to determine the chemistry of the canal water. As the project is currently envisioned, this canal water will consist of reject brine from the reverse osmosis extraction of fresh water from the marine water taken up at the Red Sea (Beyth, 2002; Asmar and Ergenzinger, 2002). As it is currently estimated, the extracted water will have a salinity of 200-300mg/l, and the extraction efficiency of about 45%. Thus, input from the Red Sea will be about 2000MCM/a, and reject water flowing into the DS will be about 1100MCM/a (Beyth, 2002). In order to determine the chemistry of the reject water, a simple mass balance approach is needed:

$$\Sigma_{RSi} = \Sigma_{CWi} + \Sigma_{FWi} \dots\dots\dots(1)$$

$$\text{and } \Sigma_{RSi} = RSi * vRS; \Sigma_{CWi} = CWi * vCW; \text{ and } \Sigma_{FWi} = FWi * vFW \dots\dots\dots(2)$$

Where RSi, CWi and FWi are the concentrations of any chemical species (i) in the Red Sea, canal and fresh waters, respectively. The lower case v refers to the volume of each type of water. Thus, assuming that there is no fractionation between the various solutes during the reverse osmosis process, and that salinity is reduced to less than 1% of its original state (Red Sea water has about 41000 mg/l TDS compared with the expected fresh water containing 300mg/l TDS), the above relationship is solved as follows:

$$CWi = [(RSi) - (FWi * 0.45)] / 0.55 \dots\dots\dots(3)$$

Since FWi is a very small figure and approaches zero, the above relationship is reduced to the following:

$$CWi = RSi / 0.55 \dots\dots\dots(4)$$

Such concentration of solutes in the effluent water leads to supersaturation with respect to calcite and aragonite, while the solution is very nearly saturated with respect to gypsum (Log IAP/K = -0.0262). The carbonates will thus be expected to precipitate in the RO facility, on the canal walls or in the canal water leading to eventual discharge into the DS.

The evaluation of the hydrological and chemical balances in the DS as a result of the introduction of the effluent water is more difficult because of the lack of knowledge about mixing ratios in the upper layer. Generally, this was evaluated by using the known historical analogs. Specifically, the mixing ratios which can be estimated from data collected prior to the disappearance of the old pycnocline.

In general, a simple mass balance approach can be used to determine the various parameters that existed prior to the overturn. The parameters that are known are the chemistry of the upper and deep layers, the volume of the upper layer, the volume of fresh water entering the DS and its chemistry, the age of the upper layer and

evaporation rates. Thus, a mass balance equation can be written to determine the dilution of the various constituents in the upper layer, and thus the mixing rates between the fresh water and the deeper water. Because this exercise requires simultaneous solution for the various constituents, especially given the possibility of mineral precipitation, two approaches can be adopted. The first would be the use of mass balance software which would attempt to take into consideration all of the variables mentioned above for all of the major solutes, and the second would require simple modeling of a “conservative” ion which would be used as an indicator for mixing ratios. Both approaches were attempted in this study.

Estimations of evaporation rates will take into account changes in the surface area as well as changes in the specific gravity of the water. Historical data will also be taken into consideration.

After the determination of the mixing status that existed in historical times, the model for the canal water mimicking the fresh water input of the past was constructed. This model involved the evaluation of the saturation states of various minerals which were deemed relevant to the situation. In particular, the saturation state and volume of gypsum was evaluated, because it is clear from the thermodynamic modeling that the brines would be near to saturation to supersaturated with respect to this phase. Gypsum precipitation is expected to be a significant side effect of this project (Beyth, 2002). DS water is notably poor in sulfate, whereas the effluent water which is to be introduced is high in sulfate and near saturation with gypsum.

**Red Sea Canal water**

The canal discharge into the DS was calculated using the above mass balance approach (equation 4) assuming a start with normal marine water. The resulting chemical composition was evaluated using PHRQPITZ (Plummer et al., 1988). The evaluation revealed slight supersaturation with respect to calcite, and undersaturation with respect to gypsum. Table 1 lists the starting chemistry, the chemistry after RO but prior to equilibration with the relevant mineral phases, and the final chemistry after equilibrium with calcite.

**Table (1):**

Sea water, canal water after concentration and canal water after equilibrium with calcite (molal units). Sea water chemistry from Vengosh and Rosenthal (1984).

	Sea water	RO discharge	Canal discharge
Ca <sup>2+</sup>	0.011	0.021	0.021
Mg <sup>2+</sup>	0.055	0.12	0.12
Na <sup>+</sup>	0.483	1.09	1.09
K <sup>+</sup>	0.010	0.02	0.02
Cl <sup>-</sup>	0.581	1.27	1.27
SO <sub>4</sub> <sup>2-</sup>	0.029	0.065	0.065
HCO <sub>3</sub> <sup>-</sup>	0.003	0.005	0.004
Br <sup>-</sup>	0.001	0.002	0.002

### The upper layer of the DS

In historical times, the thickness of the upper layer of the DS was about 40m when the elevation of the surface was at 390m below sea level (Neev and Emery, 1967). According to the hypsometric curves compiled by Hall (1997), the volume of the DS at that elevation is about 155 km<sup>3</sup>. The volume of the deep water layer (elevation 430mbsl) is 135 km<sup>3</sup>. Thus the volume of the upper layer was about 20,000 MCM. Assuming a historical annual input into the DS of about 2,000 MCM/a, then the water added annually amounted to about 10% of the reservoir. A similar volume was removed by evaporation (Salameh and El-Naser, 1999). The significance of this on the chemistry of the upper layer was small, as might be calculated from a conservative indicator such as Br<sup>-</sup>. In the deeper layer the Br<sup>-</sup> content was about 5.3g/l, and in the upper layer it was 4.6g/l (Neev and Emery, 1967). On the other hand, the major surface water input (the Jordan River) only has 0.009g/l. Thus, if a simple mass balance mixing approach is used:

$$SWBr = fdw * DWBr + frw * RWBr \dots \dots \dots (5)$$

Where SWBr, DWBr and RWBr are the Br<sup>-</sup> contents in the surface water, deep water and river water, respectively. Likewise, fdw and frw are the fractions of the contributions of the deep water and river water, respectively. If this equation is solved using the above-mentioned variables, then the river water contribution to the chemistry is about 14.3%. Applying the same approach using other conservative tracers, such as K<sup>+</sup> and Mg<sup>2+</sup>, yield river water contributions of 15.9 and 16.4%, respectively. The discrepancy between the volumetric contribution of river water (10%) and the chemical contribution of the river water (~15%) can be partially explained by the difference between standard units of measuring chemical content and volume. Chemical content is measured using mass per volume (g/l), whereas it is more appropriate in this case to express it in mass per mass of water (g/kg). This is because the specific gravity of water varies between fresh water (1.0), deep water (1.2305) and surface water (1.205). Recalculating the above relationship on a g/kg basis yields river water contributions of 10.9, 13.0 and 12.9% for Br<sup>-</sup>, Mg<sup>2+</sup> and K<sup>+</sup>, respectively. The remainder of the discrepancy can be attributed to factors such as the contributions of submarine springs to the balance and chemical reactions in the sediments of the basin.

It is important to point out here that the contribution of fresh water to the chemistry of the upper layer is mainly a dilution effect, and for the most part the chemical variations in the chemistry of fresh water inputs have insignificant effects. Steinhorn (1997) estimated that the lower historical evaporation on the DS was about 166cm/a, amounting to a volume of about 1,660 MCM/a, whereas Salameh and El-Naser (1999) estimated evaporation to be about 2,000 MCM/a, to be comparable to the estimated fresh water input. Given the relatively stable levels of the DS in the early 20<sup>th</sup> century, it is reasonable to assume that evaporation was stable at about 2m/a. If the volume of evaporation is nearly equal to the volume of fresh water input, there is no obvious reason why the surface water should form. Thus, it is plausible to assume that input exceeded evaporation during that period. The alternative explanation for this phenomenon is that the deeper layer is continuously becoming more saline, allowing for the maintenance of

the pycnocline. The continuous input of salts by deep brine springs into the DS has been proposed to be a major factor in the chemical evolution of the DS (Abu-Jaber, 1998). Thus, in the event of a new pycnocline developing as the result of the RS-DS canal, then the two water bodies would evolve separately, beginning for the current water that is relatively homogeneous throughout the water body.

### **Evaporation**

Evaporation off of the DS has been estimated to range from 1.2m by Oroud (1995) to 2m/a by Salameh and El-Naser (1999). An important reason for the discrepancies is probably related to uncertainties as to the volume of water entering the lake from spring discharges. Salameh and El-Naser (1999) estimate these to have contributed about 200MCM/a. Asmar and Ergenzinger (2002) conclude that evaporation rates over the lake will increase from a current level of 1.23m/a to a final level of about 1.54m/a.

Thus, these variations in estimates make it difficult to model how much evaporation will actually take place. The canal project is largely designed to maintain the DS at a level of about 390m bsl. Thus, in the final analysis flow rates will probably be controlled in such a way that total input (from the canal and all the other natural sources) will equal evaporation at that level. In historic terms, this was probably about 1.7m/a. At that level, the surface area of the DS will be about 1000 km<sup>2</sup>, yielding an evaporation rate of 1,700 MCM/a. This would account for the canal water (1,100 MCM/a), as well as the other sources feeding into the DS (about 617 MCM/a; Salameh and El-Naser, 1999). The plausible alternative is that both input and evaporation are closer to 2,000MCM/a, as suggested by Salameh and El-Naser (1999). The modeling of the changes in either case is similar, as are the results, because the discrepancy is largely a function of the fresh water input, which has little minimal bearing on the solute load. In this paper, the upper limit of evaporation is modeled.

Thus, the model for the evolution of the new upper layer would start with the current water chemistry of the DS with the discharge water from the canal. The mixing ratio will be the most conservative plausible number of 90% DS water and 10% canal water by volume, which is subsequently diluted to account for fresh water inputs. The modeling will assume an evaporation rate of about 2 m/a. The progressive changes in the surface water will be modeled through a ten year period.



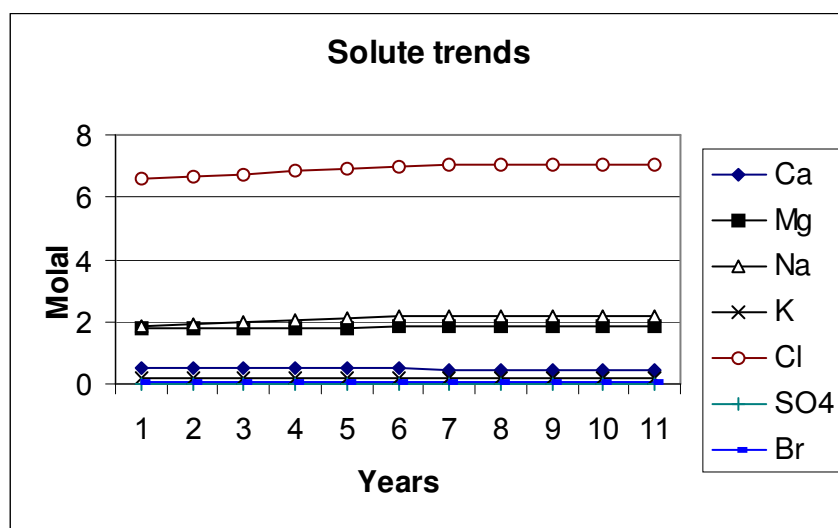
**Results**

The results of the modeling of the water chemistry change over the first ten years are shown in table 2 and in figure 2. These solutes can be grouped into three categories: those increasing due to evaporative concentration, those increasing due to canal water input as well as evaporative concentration, and those decreasing with time.

**(Table 2):**

Modeling results of chemical changes expected in the upper water mass over the first ten years. Concentrations of individual ions are expressed in molalities, whereas total concentrations are expressed in g/kg.

Solute	DS (molal)	Year 1	Year 2	Year 3	Year 4	Year 5	Year 6	Year 7	Year 8	Year 9	Year 10
Ca	0.50	0.49	0.49	0.49	0.48	0.48	0.48	0.47	0.47	0.47	0.46
Mg	1.79	1.80	1.81	1.81	1.82	1.83	1.84	1.85	1.86	1.86	1.87
Na	1.83	1.90	1.96	2.02	2.08	2.14	2.18	2.17	2.16	2.15	2.15
K	0.19	0.19	0.19	0.19	0.19	0.19	0.20	0.20	0.20	0.20	0.20
Cl	6.60	6.67	6.75	6.83	6.90	6.98	7.03	7.04	7.04	7.05	7.06
SO4	0.0093	0.0041	0.0041	0.0040	0.0039	0.0039	0.0040	0.0040	0.0040	0.0039	0.0040
Br	0.086	0.086	0.086	0.087	0.087	0.087	0.087	0.087	0.088	0.088	0.088
Total	354.9	358.4	362.7	367.0	371.2	375.5	378.3	378.5	378.6	378.8	379.0

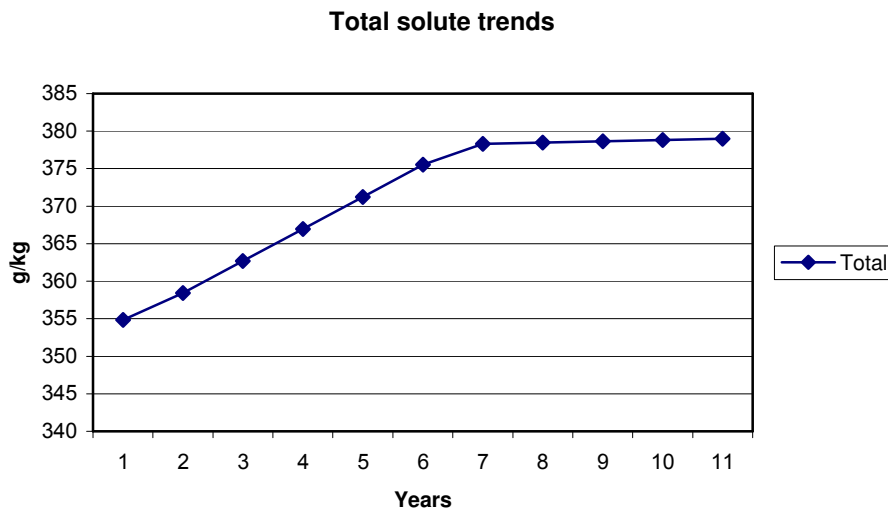


**Figure 2.** Changes in ion concentration trends in the upper water mass through the first ten years following the implementation of the canal assuming no mixing.

Examples of evaporative increase due can be seen in the trends of  $Mg^{2+}$ , and  $Br^-$  which are expected to increase at rates of 4.7% and 2.2%, respectively over the first ten years of the canal. These can be contrasted with  $Na^+$ ,  $K^+$  and  $Cl^-$ , which are expected to increase 16.9, 8.1% and 6.9%, respectively. The differences seen are obviously the result of variations in the chemical inputs of the canal water, which are highest for  $Cl^-$  (see table 1). The concentration of  $Na^+$  and  $Cl^-$  peak after six years because the water reaches supersaturation with respect to halite.

On the other hand,  $Ca^{2+}$  and  $SO_4^{2-}$  are expected to decrease in over the ten year modeling period. It is expected as the results of the modeling that  $Ca^{2+}$  will decrease by 7.7%, and that  $SO_4^{2-}$  will decrease by 56.7%. Carbon species are not modeled because they have very low concentrations in the DS, and thus are of marginal interest to the overall geochemical picture of the DS. Generally, input by surface water floods result in rapid precipitation of aragonite, producing notable whitening events. It has been estimated that  $1.4 \text{ mol/m}^2/\text{a}$  is deposited in the DS, amounting to about 140000 metric tons (Barkan et al., 2001). The dynamics of carbon in the DS are not expected to change significantly in the event of the building of the canal.

Total salinity is expected to increase from 355 g/kg to over 380 g/kg over ten years (Figure 3). This is largely due to the expected increase in the  $Cl^-$  content. Again, the total salinity peaks after six years because halite begins to precipitate out of the water.



**Figure 3.** Changes in the total salt concentration in the upper water mass (g/kg) over the ten year modeling period assuming no mixing with deep water.

## Discussion

The results of the modeling suggest that the precipitation of gypsum will result from the proposed mixing. More specifically, 0.05 moles of sulfate will be added to the water with each liter of canal discharge. All of this will be removed along with some of the existing sulfate in the DS water. This adds about 94000 metric tons of gypsum that can be expected to precipitate annually in the DS. This will probably result in numerous whitening events of DS water at a scale similar to the aragonite events which currently occur (Barkan et al., 2001).

While the DS is currently undersaturated with respect to halite according to the modeling results, the mineral currently precipitates along the shores of the lake as well in some of the salt pans of the Arab Potash Company (Anati, 1993 and Talbot et al., 1996). This is the result of local enrichment due to weak circulation or intentional impounding. The modeling suggests that this phenomenon will continue as both  $\text{Na}^+$  and  $\text{Cl}^-$  concentrations increase through time.

It is doubtful that the supposed pycnocline will be a very stable feature. The presence of a pycnocline is dependent on density stratification. Water density is a function of NaCl precipitation and other salt contents (Anati et al., 1987; Anati, 1993). The modeling clearly shows that any upper water body mass will quickly become saltier and thus denser than the lower body mass. This is obvious because about 80 grams of dissolved salt will be added with each liter of canal water. Dividing this over the presumed surface water layer containing 20000 MCM of brine will result in adding four grams of salt per year per kg water (see figure 3). Thus, while it is possible that a pycnocline will develop, as suggested by Asmar and Ergenzinger (2002), this will not be a stable feature and will experience frequent overturning.

Thus, chemical changes in the DS water due to the proposed canal will be less pronounced than is modeled herein. The frequent mixing will dampen the modeled effects shown in table 2 and in figures 2 and 3 by a factor of 1/7.75, which is the ratio between the volume of the upper water body and the total water body. The only exception is the change in the  $\text{SO}_4^{2-}$ , which changes almost instantaneously, and will continue to cause gypsum precipitation despite stability or overturning of the upper water body. The DS water will become supersaturated with respect to halite about 45-50 years after the canal begins operation.

## Conclusions

The results clearly show that the short term effects of gypsum precipitation are of relatively minor importance. However, the input of large amounts of salt water will have a profound effect on the longer term, drastically increasing the salinity of the DS until supersaturation with halite is achieved in four to five decades. The delay in the onset of halite supersaturation will be caused by the lack of development of a pycnocline.

## نمذجة جيوكيميائية لتأثيرات قناة البحرين الاحمر والميت المقترح

نزار ابو جابر

### ملخص

تمت نمذجة النتائج الكيميائية المتوقعة من تنفيذ قناة البحرين الاحمر والميت. أستخدمت البيانات التاريخية بالإضافة إلى كيميائية مياه القناة، لبناء نماذج جيوكيميائية للتوازن الترموديناميكي باستخدام حزمة PHRQPITZ. وقد بينت النتائج توقع ترسيب كميات كبيرة من الجبس. كما بينت عدم إمكانية حدوث تطبق ملحي كما توقع باحثون آخرون. كما يتوقع أيضا أن تصل المياه مرحلة الإشباع بالنسبة للهالايث بعد حوالي 50 سنة من إنشاء القناة.

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