4. Hydrochemical regulation of the lake

Jean-Pierre Carmouze

4.1 Introduction

Chad basin, situated between 5° and 25° latitude north (Fig. 1, Chapter 2) has a sub-arid climate, mainly characterized by strong variations in rainfall. During the course of this study, the rainfall was about 1500 mm year^{-1} in the south and less than 200 mm in the north. This rainfall gradient was associated with a topographic gradient; tropical waters flowed from south to north towards arid zones. Collected by the river Shari, these waters spread into a large, shallow and closed basin located between 12°5 and 14°5 latitude north to form Lake Chad (Fig. 1).

This water carries dissolved materials which have been picked up into the lake. They are mainly Ca^{++}, Mg^{++}, Na^{+}, K^{+}, HCO_3^- and SiO_4H_4 arising from monosialitic alteration of crystalline rocks (Carmouze 1976).

Due to its endorheic nature and its climate, the lake appears to be a basin that concentrates salts. However the water salinity is low and relatively unchangeable in time. Our purpose is to explain these apparent contradictions by analyzing the different factors that control the concentration of dissolved constituents in the lake waters.

Although Lake Chad is geographically closed, it is, like every aquatic environment, an open system in which material transfers take place. Therefore, the problem is to determine the different flows of water and dissolved elements which control the various lake stocks of dissolved salts, and to determine the volume of water at all times and in each chemically homogeneous region of the lake, since the salt concentration is the ratio of these two parameters.

Generally, flows vary in time and space. Therefore, especially in Lake Chad, local heterogeneities and temporal fluctuations in the environmental hydrochemistry occur.

For this reason, two simplifications were introduced:

— in order to reduce spatial variations, the lake was divided into two less heterogeneous parts representing the south and north basins of the lake;

— over a long period of time the hydrochemical state of a lake oscillates around

J.-P. Carmouze et al. (eds.) Lake Chad

ISBN 90 6193 106 1. Printed in the Netherlands

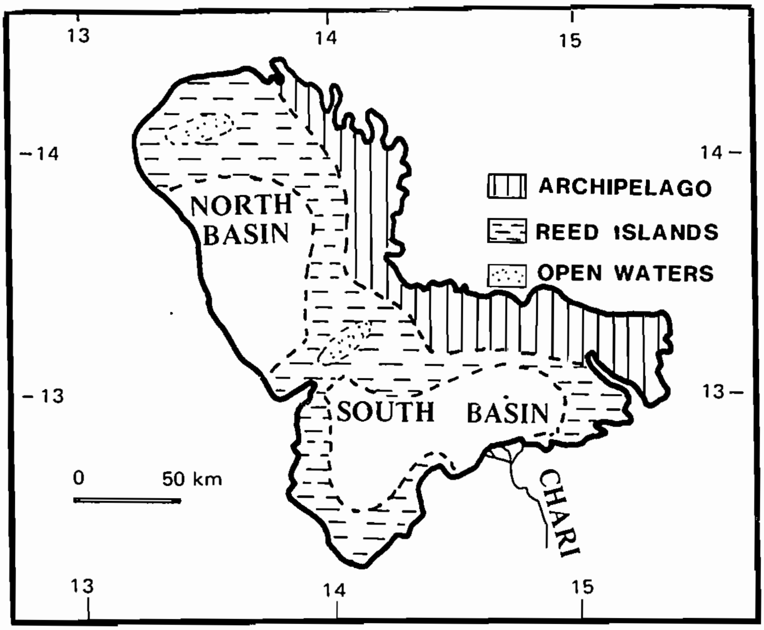

Fig. 1 Lake Chad and its different ecological zones.

an average condition because material input and output balance. If one ignores these fluctuations, the environment can be considered as a steady-state system. So the chemical characteristics of water become independent of time. To compare Lake Chad to such a system, a period of 18 years was examined (1954–1972).

Depending upon their nature the material flows may be classified in two groups: those associated with hydrological currents and those associated with physical, chemical, geochemical or biochemical reactions (Carmouze and Pedro 1977). The water currents primarily control the 'climato-geographical' dissolved salt currents (climato-geographical regulation). Depending upon this pre-regulation, biogeochemical dissolved salt currents occur (biogeochemical regulation) that control the total.

Choosing this approach, the lake will first be considered as a steady-state system. The average water and dissolved salt currents and the corresponding dynamic equilibrium for the period 1954–72 will be examined successively for the whole lake and its north and south basins. The contribution of the two hydrochemical reactions representing the main aspect of the biogeochemical reactions, i.e. mineralogical changes inside the lake and the influence of molluscs and macrophytes will be also examined.

Then, the hydrochemical regulation of the lake will be analyzed during flood and recession periods.

4.2 The hydrology of the lake and its contribution to hydrochemical regulation during a permanent state

Lake levels have been estimated from 1895 to 1954 (Toucheboeuf et al. 1969) and measured since 1954. Complete data about climatology and hydrology are also available after this date. Therefore the period 1954–1972 has been chosen to describe the average hydrological state of the lake. This state was similar to that of the longer period 1895–1972, according to the lake level fluctuations (Fig. 2).

4.2.1 *The total hydrology: mean hydrological balance of the lake*

4.2.1.1 *Input regime. (a) River supplies*: almost all of the river supplies come from the Shari and El Beïd. The annual Shari supplies are between 19.5×10^9 and $54.5 \times 10^9 m^3$, the average being $40 \times 10^9 m^3$. The seasonal distribution is similar to a transition tropical regime with the lake receiving 73% of the average annual supply between September and December.

The river El Beïd flows from October to March. The annual supply varies from 0.6×10^9 to $2.7 \times 10^9 m^3$ with a mean value of $1.35 \times 10^9 m^3$. Over the period under consideration, the lake received $41.5 \times 10^9 m^3$ water from rivers.

(b) *Rainfall*: the rainy season lasts from May to October. In fact the lake receives 50% of the annual supply in August with an average of 300–310 mm $year^{-1}$. It progressively decreases from south (600 mm) to north (250 mm). The annual volume ranges between 2.7×10^9 and $8.7 \times 10^9 m^3$, with a mean value, $(\bar{v})_M$ of $6.3 \times 10^9 m^3$.

Overall, the annual supply of water varies from $23 \times 10^9 m^3$ to $61.4 \times 10^9 m^3$ with an average of $48 \times 10^9 m^3$. 87% of this is from the rivers and 13% from rainfall.

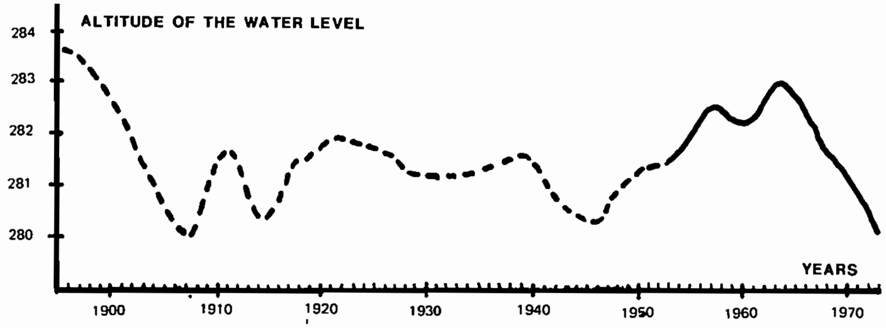

Fig. 2 Annual variations in the water level of Lake Chad from 1895 to 1973 (------ estimated; ——— measured).

4.2.1.2 *Output regime*. (a) *Evaporation losses*: with an annual evaporation rate of 2.11 m, the volume of water evaporated varies between approximately 47 and $38 \times 10^9 \mathrm{m}^3\ \mathrm{year}^{-1}$ with a mean value of $44 \times 10^9 \mathrm{m}^3$.

(b) *Infiltration losses*: the lake is directly connected with a 50–70 m thick groundwater bed. According to the piezometric gradients the lake is above. This is a situation favouring infiltrations. Moreover they are facilitated because of the sandy nature of the shores and a part of the lake bottom (10 to 15%) (Dieleman and De Ridder 1963; Pirard 1963; Schneider 1967, Fontes et al. 1969; Roche 1973; Maglione 1974).

The infiltrations have been indirectly estimated, using the Na budget and assuming that this element is only eliminated by infiltration. During the period 1954–1972, the Na output of 5.6×10^{12}mM year^{-1} has balanced the Na input. This quantity $(\overline{\mathrm{Na}})_I$ is eliminated in a volume of infiltration $(\bar{v})_I$ equal to $(\overline{\mathrm{Na}})_I/[\overline{\mathrm{Na}}]_I$, where $[\overline{\mathrm{Na}}]_I$ is the mean Na concentration in the infiltrated water. Assuming that the infiltrations occur throughout the lake, $[\overline{\mathrm{Na}}]_L$, whose value is 1.46mM l^{-1} (Carmouze 1976). Thus the infiltration volume can be calculated as:

$$(\bar{v})_I = 5.6 \times 10^{12}/1.46 \times 10^3 = 3.85 \times 10^9 \mathrm{m}^3\ \mathrm{year}^{-1}$$

The annual evaporation and infiltration losses representing a water height of 2.29 m have been approximately calculated for each year of the period 1954–1972, by allowing the total losses to be roughly proportional to the lake surface. The total losses varied from 41.3×10^9 to $51.8 \times 10^9 \mathrm{m}^3$, of which evaporation represented 92% and infiltration 8%.

4.2.1.3 *Hydrological characteristics of the lake*. The volume, area and depth are controlled by the basin morphology, and the input/output regimes of water previously described. These parameters were calculated from morphometric curves of the lake (Carmouze 1976; Lemoalle 1978).

The lake levels registered on the northern shore of the south basin from 1954 to 1972 changed from 281.5 to 280.5 m with a maximum of 283.3 m; and was always modified by seasonal changes of 0.85 to 0.80 m.

The lake volume varies between $42.5 \times 10^9 \mathrm{m}^3$ and $91 \times 10^9 \mathrm{m}^3$, with an average value, $(\bar{v})_L$, of $72 \times 10^9 \mathrm{m}^3$. The considerable volume fluctuations are caused by very variable inputs: from 1954 to 1972 the water supply varied from 23 to $61.4 \times 10^9 \mathrm{m}^3\ \mathrm{year}^{-1}$ while the annual output variations were less important, the losses ranging from 41.3 to $51.8 \times 10^9 \mathrm{m}^3\ \mathrm{year}^{-1}$. The small volume of the lake poorly matches the output–input inegality: on average 66% of the lake water is renewed each year.

The seasonal fluctuations in volume represent about 15% of the total volume and result from 80% of the water supplies being received over 5 months while the losses are distributed more evenly through the year.

The lake area varies between 22 600 km^2 (1962) and 1 800 km^2 (1972) with a mean of 20 900 km^2. The seasonal fluctuation in area during an average hydrological regime of the lake is $\pm 3\%$.

4.2.1.4 *The mean hydrological regulation and its contribution to hydrochemical regulation.* From the data just presented, it is possible to evaluate the annual mean hydrological balance of the lake (Fig. 3) and point out the following features:

- the input mainly comes from the river (86.5% against 13.5% from rainfall) while the losses due to evaporation are greater than those due to infiltration (92% against 8%);
- the input and output regimes and the shape of the basin are such that the lake volume remains small. Two thirds of the lake volume is renewed each year with a turnover time of about 1.5 years.

Because of the first characteristic, the lake is a salt concentration basin whose concentration factor is easy to calculate. Let us consider a dissolved element, i, whose circulation is only controlled by water flows. This means that the input of i is annually balanced by infiltration.

$$[i]_F \cdot (\bar{v})_F = [\bar{i}]_I \cdot (\bar{v})_I = [\bar{i}]_I \times [(\bar{v})_F + (\bar{v})_M - (\bar{v})_E] \quad (1);$$

$[\bar{i}]_F$ and $[i]_I$ = mean value of respectively, river and infiltrated water. $(\bar{v})_F$, $(\bar{v})_I$, $(\bar{v})_M$ and $(\bar{v})_E$ = annual mean volume respectively, of river water, infiltrations, rain and evaporation.

Let us assume that the infiltration occurs approximately uniformly all over the lake. Then, $[i]_I = [i]_L$ = mean lake concentration of i. According to (1)

$$\frac{[\bar{i}]_L}{[\bar{i}]_F} = \frac{(\bar{v})_F}{(\bar{v})_I} = \frac{(\bar{v})_F}{(\bar{v})_F + (\bar{v})_M - (\bar{v})_E} = 10.8$$

This dissolved salt concentration factor controlled by the 'climato-geographical' regulation is close to 11. It is not very high for a closed lake because of the relative importance of infiltrations.

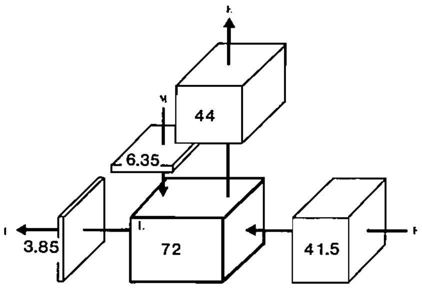

Fig. 3 Mean annual hydrological balance of the whole lake. F = fluvial input; E = evaporation; M = meteoric input; I = infiltration. All the flows and the lake volume are expressed in m^3 × 10^9.

The second characteristic of the hydrological regulation denotes the lake instability in regard to the input irregularities. The effects on the hydrochemistry are examined in para. 5.1.

For example, an annual excess of the river input of about 33% produces a 15% decrease in the lake concentration of i. On the contrary, for a deficit of 33% the increase in concentration of i is 21.5%. So, when only geographical regulation takes place, the variation in saline concentration corresponding to the input–output imbalances (frequently having these extents), is relatively important.

4.2.2 *Regional hydrology: mean hydrological balance of the south and north basins and its influence on hydrochemistry*

The south and the north basins are two distinct geographical entities with different characteristics. So, to obtain a better hydrological knowledge of the lake the mean annual hydrological balance of the two basins has been determined in a way similar to that for the whole lake (Carmouze 1976).

From the results shown in Fig. 4 the following features can be pointed out:

- the north basin contains twice as much water as the south basin, while having roughly the same area;
- the losses due to evaporation are almost equal in both environments because the evaporation rates are similar;
- the infiltration losses are twice as important in the north basin as in the south (65% of the total infiltration against 32%);

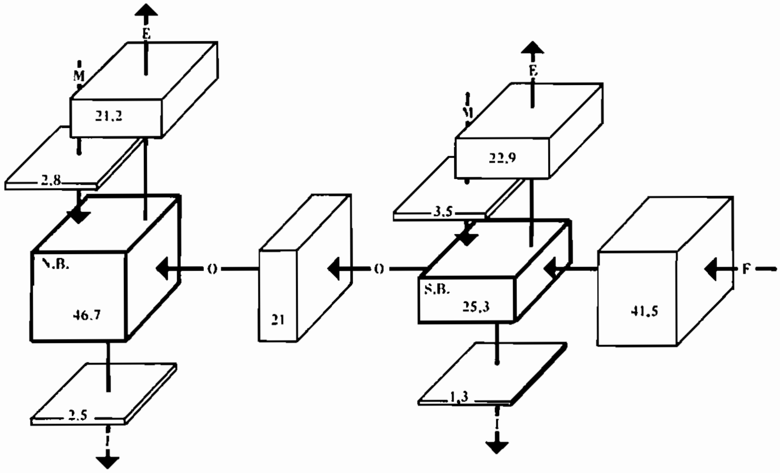

Fig. 4 Mean annual hydrological balance of the south and the north basin. F = fluvial input; E = evaporation; M = meteoric input; O = north basin input coming from south basin. All the flows and volumes of the north (N.B.) and south (S.B.) basins are expressed in $m^3 \times 10^9$.

— the turnover rate of the water in the south basin is very high and the residence time of the water is 6 months. 50% of the river input flows into the north basin where the average residence time of the water is about 2 years.

These hydrological regulation differences cause some hydrochemical modifications in the two basins by determining the rate of increase of the different salinities.

For an element i, whose regulation is only climato-geographical (a conservative element) 8.2% of its annual input remains in the south basin, while 91.8% flows into the north (Carmouze 1976). A rough calculation indicates that the rate of increase of the concentration of i, due to the geographical regulation is about 2.6 in the south basin and 8.5 in the north basin. Consequently, the climato-geographical increase of salinity is 3.25 times higher in the latter basin (Carmouze 1976).

4.3 Contribution of biogeochemical sedimentation to the regulation of steady-state hydrochemistry of the lake

The mean annual dynamic equilibrium of each constitutent of the water salts corresponding to 1954–1973 can be established for the lake and its north and south basins. The constituents are: Na^+, Ca^{++}, Mg^{++}, K^+, HCO_3^- and SiO_4H_4. To achieve this, some approximations must be made because hydrochemical data are only available for part of this period.

The river salinity data are only available from 1967 (Carmouze 1969, 1972, 1976; Roche 1969, 1973), but they point out that, for a given season, the salt concentrations do not change from one year to another, whatever the water flow may be. So, knowing the mean monthly concentrations, it is possible to calculate the different salt inputs if the monthly water flows are available.

The mean hydrochemical characteristics of the lake can be approximately obtained, using the fact that lake salt stocks depend mainly on lake volume. In 1967, the lake had reached a balanced situation and so, for this year the average stock values can be extrapolated from the data of 1968 to 1972.

The total mean annual losses can be estimated as being equal to the inputs for the entire period 1954–1972. Moreover, supposing that Na is only eliminated by infiltration, it becomes possible to evaluate the fraction of the other constituents that are biogeochemically eliminated.

4.3.1 *The hydrochemistry: mean annual balances of the dissolved salts*

4.3.1.1 *River input of dissolved salts.* (a) *The lower Shari supplies*: The ionized salinity of the Shari varies between 0.38 and 0.78 mé l^{-1} with a minimum at the beginning of the flood (August–September) of 0.38–0.42 mé l^{-1} and a maximum at the time of low water (May–June) of 0.68–0.78mé l^{-1}.

The mean monthly values of the dissolved components vary between 0.41 and 0.72mM l^{-1} for HCO_3^-, 0.071 and 0.137mM l^{-1} for Ca^{++}, 0.067 and 0.115mM l^{-1} for Mg^{++}, 0.105 and 0.155mM l^{-1} for Na^+, 0.04 and 0.06mM l^{-1} for K^+ and 0.300 and 0.425mM l^{-1} for SiO_4H_4. The salinity of these waters is half that of water in continental rivers (Davies and De Viest 1966). However the dissolved silica concentration of the Shari river is 1.5 times higher than the average.

The salt inputs of each constituent have been calculated on a monthly basis for the period 1954–1972 using monthly concentration values and the corresponding hydrological data (Billon et al. 1969). The annual inputs generally vary between 75 and 125% of the mean values which are:

$$(\overline{Ca^{++}})_{CH} = 4.01 \times 10^{12}\text{mM};\ (\overline{Mg^{++}})_{CH} = 3.1 \times 10^{12}\text{mM};\ (\overline{Na^{+}})_{CH} = 5.25 \times 10^{12}\text{mM};$$

$$(\overline{K^{+}})_{CH} = 1.9 \times 10^{12}\text{mM};\ (\overline{HCO_3})^- = 20.8 \times 10^{12}\text{mM};\ (\overline{SiO_4H_4})_{CH} = 15 \times 10^{12}\text{mM}$$

Generally speaking, the supply regimes for dissolved salts parallel those for water supplies.

(b) *The El Beïd supplies*: the mean chemical composition of the El Beïd water is:

$$(\overline{Ca^{++}})_{EB} = 0.215\text{mM } l^{-1};\ (\overline{Mg^{++}})_{EB} = 0.15\text{mM } l^{-1};\ (\overline{Na^{+}})_{EB} = 0.35\text{mM } l^{-1};$$

$$(\overline{K^{+}})_{EB} = 0.10\text{mM } l^{-1};\ (\overline{HCO_3^-})_{EB} = 1.15\text{mM } l^{-1};\ (\overline{SiO_4H_4})_{EB} = 0.71\text{mM } l^{-1} \text{ (Roche 1973).}$$

From these chemical data and the hydrological data given by Billon et al. (1974) the annual inputs for 1954–1972 have been calculated:

$$(\overline{Ca^{++}})_{EB} = 0.3 \times 10^{12}\text{mM};\ (\overline{Mg^{++}})_{EB} = 0.2 \times 10^{12}\text{mM};\ (\overline{Na^{+}})_{EB} = 0.5 \times 10^{12}\text{mM};$$

$$(\overline{K^{+}})_{EB} = 0.15 \times 10^{12}\text{mM};\ (\overline{HCO_3^-})_{EB} = 1.6 \times 10^{12}\text{mM};\ (\overline{SiO_4H_4})_{EB} = 1.1 \times 10^{12}\text{mM}.$$

Hence, the mean annual values of the total dissolved salt inputs which include the Shari and El Beïd supplies are:

$$(\overline{Ca^{++}})_{F} = 4.35 \times 10^{12}\text{mM};\ (\overline{Mg^{++}})_{F} = 3.3 \times 10^{12}\text{mM};\ (\overline{Na^{+}})_{F} = 5.6 \times 10^{12}\text{mM};$$

$$(\overline{K^{+}})_{F} = 2.05 \times 10^{12}\text{mM};\ (\overline{HCO_3^-})_{F} = 22.4 \times 10^{12}\text{mM};\ (\overline{SiO_4H_4})_{F} = 16.1 \times 10^{12}\text{mM}.$$

4.3.1.2 *Output rates of dissolved salts.* The dissolved salt losses arise from infiltration, biogeochemical elimination and marginal deposits. The latter loss occurs when the lake decreases in size, isolating near-shore ponds that dry up, and the less soluble dissolved salts precipitate. On the other hand, when the lake size increases, these salts are partially recovered. We admit that, for a long period, the precipitation–dissolution phenomena are balanced and, consequently, the only effective losses are due to infiltration and biogeochemical sedimentations.*

Thus for the period 1954–1972, the mean annual losses, $(\bar{i})_p$, were similar to the mean annual input.

$$(\overline{Ca^{++}})_P = 4.3 \times 10^{12}\text{mM};\ (\overline{Mg^{++}})_P = 3.3 \times 10^{12}\text{mM};\ (\overline{Na^{+}})_P = 5.6 \times 10^{12}\text{mM};$$

$$(\overline{K^{+}})_P = 2.05 \times 10^{12}\text{mM};\ (\overline{HCO_3^{-}})_P = 22.4 \times 10^{12}\text{mM}:\ (\overline{SiO_4H_4})_P = 16.1 \times 10^{12}\text{mM}.$$

Due to its physico-chemical properties and the hydrochemical environment, Na sedimentation may be unimportant. Therefore the Na supplies from the river $(\overline{Na})_F$ are mainly eliminated by infiltration: $(\overline{Na})_F = (\overline{Na})_I = 5.60 \times 10^{12}$mM per year. The infiltration volume has been calculated by this hypothesis in para. 2.1.2 to be $(v)_I = 3.85 \times 10^9 m^3$. Also calculated was the increasing salinity factor of 10.8 (para. 2.1.4) in the lake caused by the water currents.

This factor is lower for dissolved constituents other than Na^+ because they are subject to biogeochemical sedimentations. It is evaluated as follows: without sedimentation, the 'climato-geographical' lake concentration of i, $[i]_{L\ cg}$, would be: $[i]_{L\ cg} = 10.8[i]_F$. The annual loss by infiltration would be equal to $[i]_{L\ cg} \times (\bar{v})_I$. But as the lake concentration of i is equal to $[i]_{L\ cg}$, the effective infiltration of i is equal to $[i]_L \times (\bar{v})_I$. The difference represents the sedimentation, $(\bar{i})_S = \{[i]_{L\ cg} - [i]_L\}\ (\bar{v})_I$.

The mean annual losses by infiltration can then be calculated:

$$(\overline{Ca^{++}})_I = 2.4 \times 10^{12}\text{mM};\ (\overline{Mg^{++}})_I = 1.95 \times 10^{12}\text{mM};\ (\overline{Na^{+}})_I = 5.6 \times 10^{12}\text{mM};$$

$$(\overline{K^{+}})_I = 1.65 \times 10^{12}\text{mM};\ (\overline{HCO_3^{-}})_I = 15.6 \times 10^{12}\text{mM};\ (\overline{SiO_4H_4})_I = 3.0 \times 10^{12}\text{mM}$$

and those by sedimentation:

$$(\overline{Ca^{++}})_S = 1.95 \times 10^{12}\text{mM};\ (\overline{Mg^{++}})_S = 1.35 \times 10^{12}\text{mM};\ (\overline{Na^{+}})_S = 0.0;$$

---

* By neglecting the border precipitation–dissolution of salts, the losses by infiltration and biogeochemical pathways are overestimated because the redissolution of salts deposited on the border is certainly not complete.

$(\overline{K^+})_S = 0.40 \times 10^{12}$mM; $(\overline{HCO_3^-})_S = 6.8 \times 10^{12}$mM; $(SiO_4H_4)_S$
$= 13.1 \times 10^{12}$mM.

4.3.1.3 *Mean hydrochemical characteristics of the lake.* Let us recall that our purpose is to describe the mean hydrochemical state of the lake during 1954–1972, but the chemical data are only available for 1968–1972. Fortunately, the hydrochemistry depends on the hydrology and the dissolved salt content is largely proportional to the volume (Carmouze 1976). Although 1967 represented the average state of the lake, there are no data available for it. thus the calculations are made by extrapolation from the 1968–72 data.

Using this procedure, the mean salt stocks of the different constituents have been calculated:

$(\overline{Na^+})_L = 105.2 \times 10^9$mM; $(\overline{Mg^{++}})_L = 36.65 \times 10^{12}$mM; $(\overline{Ca^{++}})_L$
$= 45.05 \times 10^{12}$mM;
$(\overline{K^+})_L = 31.10 \times 10^9$mM; $(\overline{HCO_3^-})_L = 292.15 \times 10^{12}$mM; $(SiO_4H_4)_L$
$= 56.1 \times 10^{12}$mM.

The absolute concentrations were calculated using $(\bar{v})_L = 72 \times 10^9 m^3$:

$[\overline{Na^+}]_L = 1.46$mM l^{-1}; $[\overline{Mg^{++}}]_L = 0.509$mM l^{-1}; $[\overline{Ca^{++}}]_L$
$= 0.625$mM l^{-1};
$[\overline{K^+}]_L = 0.432$mM l^{-1}; $[\overline{HCO_3^-}]_L = 4.057$mM l^{-1}; $[\overline{SiO_4H_4}]_L$
$= 0.779$mM l^{-1}.

4.3.1.4 *Conclusion: the hydrochemical regulation and biogeochemical sedimentation of dissolved salts.* The mean annual dynamic equilibria of the different mineral salts are shown in Figs. 5A and 5B, considering the lake as a homogeneous, single environment.

Comparing these equilibria, we note that each component has its own regulation. The geographical regulation alone would have the same concentration factor of 10.8 for each. So the concentration of Ca, Mg, K, HCO_3 and SiO_4H_4 would have been respectively equal to 1.14, 0.862, 0.527, 5.85 and 4.18mM l^{-1}, but their actual values are 0.625, 0.509, 0.432, 4.06 and 0.779mM l^{-1}. These observed decreases are caused by intervention of biogeochemical sedimentations.

Considering only the two extremes of Na and SiO_4H_4, we note that the turnover of the Na stock represents 5.3% while the turnover of the SiO_4H_4 stock is 78.7%. In other words the lake is a transit place for Na where the regimes of river supplies and losses by infiltration maintain a relatively high stock of this element. However for SiO_4H_4 the lake is primarily a place of chemical sedimentation where the regime of river supplies and losses by biogeochemical means and to a lesser degree by infiltration maintain it in a relatively low stock.

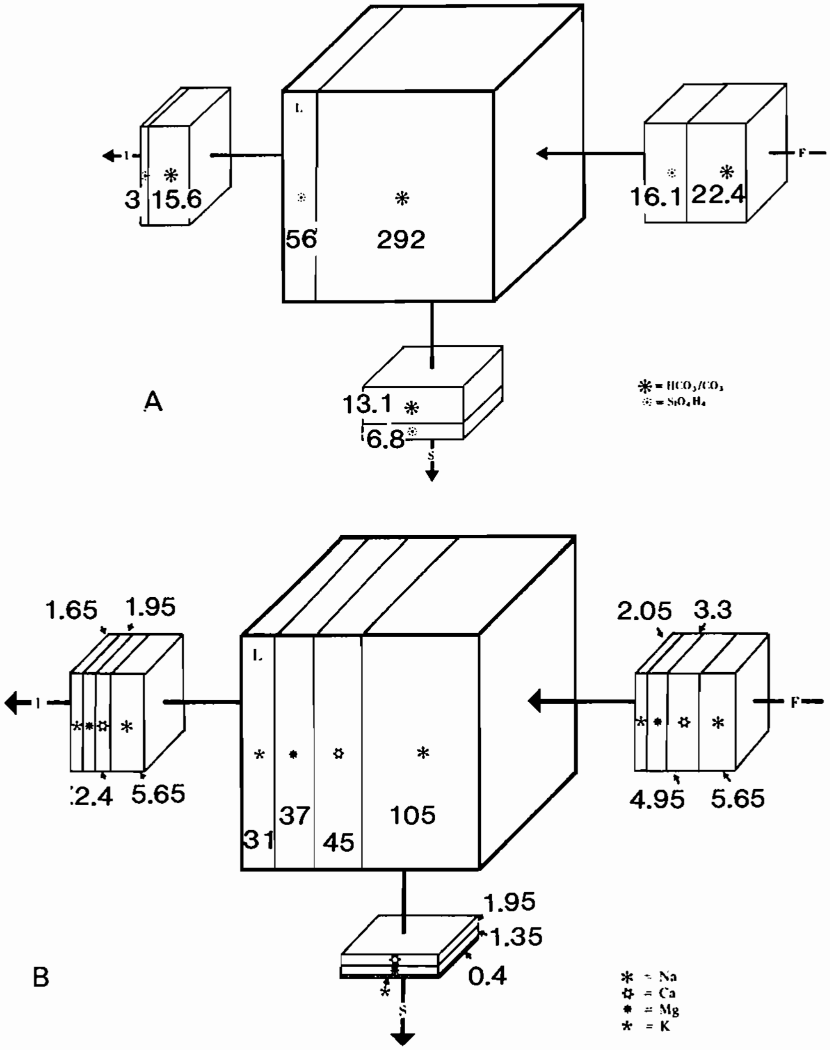

Fig. 5 Mean annual salt balances in the whole lake. A) HCO_3^-/CO_3^{--} and SiO_4H_4; B) Na^+, Ca^{++}, Mg^{++} and K^+.
F = fluvial input; S = sedimentation; I = infiltration. The flows and the lake stocks (L) are expressed in moles × 10^9.

In summary, there are three main reasons for the low salinity of the water:

- the salinity concentration factor in this arid closed lake is only 10.8, a low value due to the considerable infiltrations;
- the low river salinity is on average twice as low as the mean salinity of continental waters;

— considerable biogeochemical sedimentation occurs in the lake, which reduces the salinity by 45%.

The varying importance of biogeochemical sedimentation according to the constituents causes the relative chemical composition of the river to tend towards the Na–HCO_3/CO_2 side.

The mean relative chemical composition of the rivers and the lake is:

$[\overline{Ca^{++}}]^r_F = 29.5\%$ $[\overline{Ca^{++}}]^r_L = 20.8\%$; $[\overline{Mg^{++}}]^r_F = 22.3\%$, $[\overline{Mg^{++}}]^r_L = 16.8\%$
$[\overline{Na^{+}}]^r_F = 34.3\%$, $[Na^{+}]^r_L = 47.8\%$; $[\overline{K}^{+}]^r_F = 13.9\%$, $[\overline{K}^{+}]^r_L = 14.5\%$
$[\overline{HCO_3^-}]^r_F = 58.7\%$, $[\overline{HCO_3^-}]^r_L = 84\%$; $[\overline{SiO_4H_4}]^r_F = 41.3\%$, $[\overline{SiO_4H_4}]^r_L = 16\%$.*

4.3.2 *Regional hydrochemistry: average salt balance in the north and south basins*

The mean hydrochemical balance of the dissolved constituents has also been established for the two basins (Carmouze 1976). They provide good information on the salt transfers on a regional scale. The results are shown in Table 1 and Figs. 6A and 6B.

The south basin is mainly a transit zone for dissolved salts, with very short residence times (18 months for Na, 11 months for SiO_4H_4). About 73% of the river input flows into the north basin where the residence time ranges from 19 years for Na to 4.5 years for SiO_4H_4. Consequently the north basin is the principal reservoir of dissolved salts with 88.5% of the total stock compared with 11.5% in the other basin.

Sedimentation is equally important in both basins. In the south, for SiO_4H_4, the sedimentation is 48% compared with 52% in the north: for Ca 51% S/49% N; for Mg 44.5% S/55.5% N for K 37% S/63% N and for HCo$_3^-$ 47.5% S/52.5% N.

On the contrary the salt loss by infiltration is much less important in the south basin being 11% of the total as compared with 89% in the north.

As a consequence of these different hydrological and hydrochemical balances between the basins, the salinity of the north basin is 4 times higher than that of the south basin (10.7 mM l^{-1} against 2.55 mM l^{-1}). This increase is accompanied by a higher relative concentration of Na and HCO_3/CO_3 to the detriment of other elements.

$$^{*}[\text{Cation i}]^r = \frac{[\text{Cation i}]}{[\Sigma\ \text{cations}]};\ [HCO_3^-]^r = \frac{[HCO_3^-]}{[HCO_3^-] + [SiO_4H_4]};[SiO_4H_4]^r = \frac{[SiO_4H_4]}{[HCO_3^-] + [SiO_4H_4]}.$$

Table 1 Characteristics of the mean salt balances in the north and south basins (N.B. and S.B.); M = moles.

	Na^+		Ca^{++}		Mg^{++}		K^+		HCO_3^-/CO_3^-		SiO_4H_4	
	S.B.	N.B.	S.B.	N.B.	S.B.	N.B.	S.B.	N.B.	S.B.	N.B.	S.B.	N.B.
River inputs $M \times 10^9$	5.60	5.14	4.35	3.09	3.30	2.50	2.05	1.76	22.35	17.6	16.1	9.03
Infiltrations $M \times 10^9$	0.46	5.15	0.27	2.13	0.20	1.75	0.14	1.51	1.50	14.05	0.77	2.19
Sedimentations $M \times 10^9$	—	—	0.99	0.96	0.60	0.75	0.15	0.25	3.23	3.60	3.30	6.83
Lake stocks $M \times 10^9$	8.95	96.2	5.22	39.8	3.86	32.8	2.82	28.3	29.2	2.83	15.0	41.1
Absolute concentrations $mM \times l$	0.355	2.05	2.205	0.85	0.15	0.70	0.110	0.605	1.15	5.65	0.595	0.88
Relative concentrations %	42.9	48.8	25	20.2	18.5	16.6	13.5	14.3	66	86,5	34	13.5
Turnover %	62.5	5.35	83.9	8.0	85.5	7.6	88.4	7.20	76.7	6.2	107	22

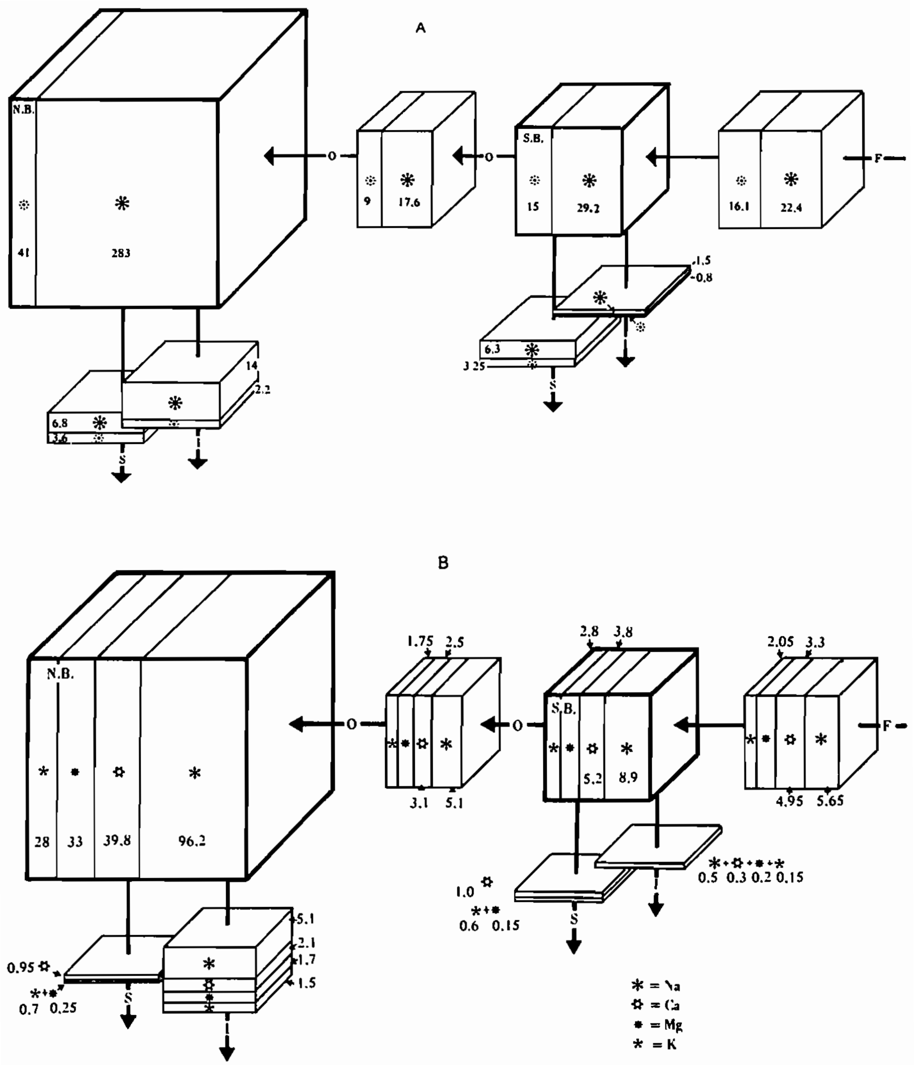

Fig. 6 Mean annual salt balances in the south and the north basins. A) HCO_3^-/CO_3^{--} and SiO_4H_4; B) Na^+, Ca^{++}, Mg^{++} and K^+.

F = fluvial input; S = sedimentation; I = infiltration; O = north basin input coming from the south basin. All the flows and the north (N.B.) and south (S.B.) basin stocks are expressed in moles × 10^9.

4.4 Nature of the biogeochemical sedimentation

4.4.1 *The geochemical sedimentation: neoformation of smectites and calcite precipitation*

4.4.1.1 *Geochemical changes in the lake.* Each year the Shari transports an average of 2.8×10^6t of suspended material to the lake, 69% of which is discharged between June and September. Its average mineralogical composition is given in Table 4.2. Kaolinite is the major component and the smectite is a ferriferous beidellite. $[Si_{3.47}\,Al_{0.53}]\,O_{10}\,[Al_{0.33}\,Fe^{II}{}_{1.17}\,Ti_{0.29}\,Mg_{0.21}]\,(OH)_2\,Mg_{0.06}\,Ca_{0.16}\,Na_{0.025}$.

The lake sediments contain some minerals similar to those transported by the Shari but with different composition. The relative percentage of smectite is greater in the sediments.

To evaluate these modifications, the sum of the relative percentages of kaolinite, illite and feldspars was taken as a reference. These minerals have the advantage of being uniformly distributed throughout the lake, without mineralogical modifications and all showing identical composition in the solid load of the Shari and in the sediments (Carmouze 1976). Thus we calculated the gains and losses of their mineralogical constituents after sedimentation in relative and absolute values based on a mean annual transport for the period 1954–1972.

Thus 30% and 80% of Fe_2O_3 and Al_2O_3, supplied annually are recombined while there is an enrichment of sediments by amorphous silica and smectite in quantities representing 200% and 45% of the annual supplies of these constituents. One such enrichment of smectite comes largely if not completely from neoformation in the middle of the lake. As the mineralogical supplies of the Shari are less varied over the period of surface sediment formation, 35% of the smectites in the sediments can be estimated to be of allochthonous origin against 65% of autochthonous origin (Carmouze 1976).

4.4.1.2 *Characterization of neoformed smectites: formation, reaction and resulting hydrochemical modifications.* Three kinds of smectites were identified (Carmouze 1976; Carmouze et al. 1976–1977; Pedro et al. 1978). They were:

— a nontronite in the pseudo-sand sediments near the delta;
— a ferrous beidellite in the central region of the lake;
— a magnesium montmorillonite in the northern region of the lake.

In addition, sediments located in the north contained 5% to 10% calcite.

(a) *The nontronites.* The sediments near the Shari delta are composed of granules (so-called pseudo-sands) whose sizes range from 0.125 to 0.500 mm. They present different forms, colours and hardness, changing from black, manually unfriable and spherical granules to greenish-yellow, manually friable and oval pellets. They are constituted of goethite Fe O(OH) and nontronite whose average composition is $[Si_{3.83}\ Al_{0.06}\ Fe^{3+}_{0.11}]\ 0_{10}\ [Fe^{3+}_{1.76}\ Mg_{0.20}]\ (OH)_2\ Ca_{0.22}\ Na_{0.02}\ K_{0.02}$.

Black granules which derive from an agglomeration of monocristalline microparticles of ferric oxide of fluvial origin change into greenish-yellow ones in the delta. This change corresponds to a progressive silicification of goethite into nontronite: $1.87\ Fe\ O(OH)+0.06\ Al(OH)_3+3.83\ SiO_4H_4+0.20\ Mg^{++}+0.22\ Ca^{++}+0.02\ Na^{+}+0.02\ K^{+} \rightarrow$ nontronite $+\,0.88\ H^{+}+7.2\ H_2O$, followed by: $0.88\ H^{+}\ 0.88\ HCO_3^{-} \rightarrow 0.88\ CO_2+0.88\ H_2O$.

The dissolved silica and the magnesium in the nontronite structure come from the dissolved phase. The aluminium is the amorphous aluminium of fluvial origin or that contained in the goethite. This reaction produces an important hydrochemical elimination of silica (4 moles are necessary to form a mole of nontronite) and to a lesser extent, an elimination of Ca^{++} and Mg^{++} (0.20–0.22 moles). Besides, the reaction transforms a HCO_3^{-} fraction into Co_2 which is liberated, and thus the pH of the water decreases. Geochemically, there is an impoverishment of the environment, mainly of ferric oxide.

(b) *The ferriferous beidellite.* 65% of the smectite contained in the sediments located in the central zone is of autochthonous origin and 35% of fluvial origin, the formula of the former is: $[5Si_{4.0}]\ 0_{10}\ [Al_{1.32}\ Fe^{3+}_{0.43}\ Ti_{0.02}\ Mg_{0.18}]\ (OH)_2\ Ca_{0.14}\ Na_{0.01}$. The iron and the aluminium constituting this smectite come respectively from microcrystalline amorphous ferric hydrate and amorphous aluminium from the Shari. The titanium originates from the hydrate of titanium of fluvial origin in a free form. Finally the silica and the magnesium of the system and the exchangeable cations are extracted from the dissolved phase.

The formation reaction is: $4.0\ SiO_4H_4+1.32\ Al(OH)_3+0.43\ FeO(OH)+0.02\ Ti\ O(OH)_{2S}+0.18\ Mg^{++}+0.13\ Ca^{++}+0.01\ Na^{+} \rightarrow$ ferriferous beidellite $+\,0.63H^{+}+8.9\ H_2O$ followed by $0.63\ H^{+}+0.63\ HCO_3^{-} \rightarrow 0.63\ CO_2+0.63\ H_2O$. Thus, the formation of a mole of beidellite causes an impoverishment in the sediment of $Al\ (OH)_3$ (1.32 moles) and of $Fe\ O(OH)$ (0.43 moles) and in the aqeuous solution of SiO_4H_4 (4.0 moles), Mg^{++} (0.18 moles) and Ca^{++} (0.13 moles). CO_2, is also liberated, decreasing the pH of water.

(c) *The magneseous montmorillonite.* In the sediments of the north there is a mixing of smectites of fluvial origin, and smectites neoformed in the central zone with a locally formed magneseous smectite which formula is $[Si_4]O_{10}\ [Al_{0.24}\ Fe^{3+}_{0.12}\ Ti_{0.11}\ Mg_{1.99}]\ (OH)_2\ Ca_{0.195}\ Na_{0.07}$.

The aluminium component is derived from amorphous aluminium and the iron and the titanium from their corresponding oxides in free form. The silica and magnesium as well as their exchangeable cations come from the solution. Schematically the reaction is: $4.0\ SiO_4H_4+0.24\ Al\ (OH)_3+0.12\ Fe\ O(OH)_2+0.11\ Ti\ O(OH)_2+1.99\ Mg^{++}+0.195\ Ca^{++}+0.07\ Na^{+} \rightarrow$ magneseous montmorillonite $+\,4.44\ H^{+}+4.4\ H_2O$, followed by $4.44\ H^{+}+4.44\ HCO_3^{-} \rightarrow 4.44\ CO_2+4.44\ H_2O$.

The neoformation of a mole of this smectite causes a major elimination of

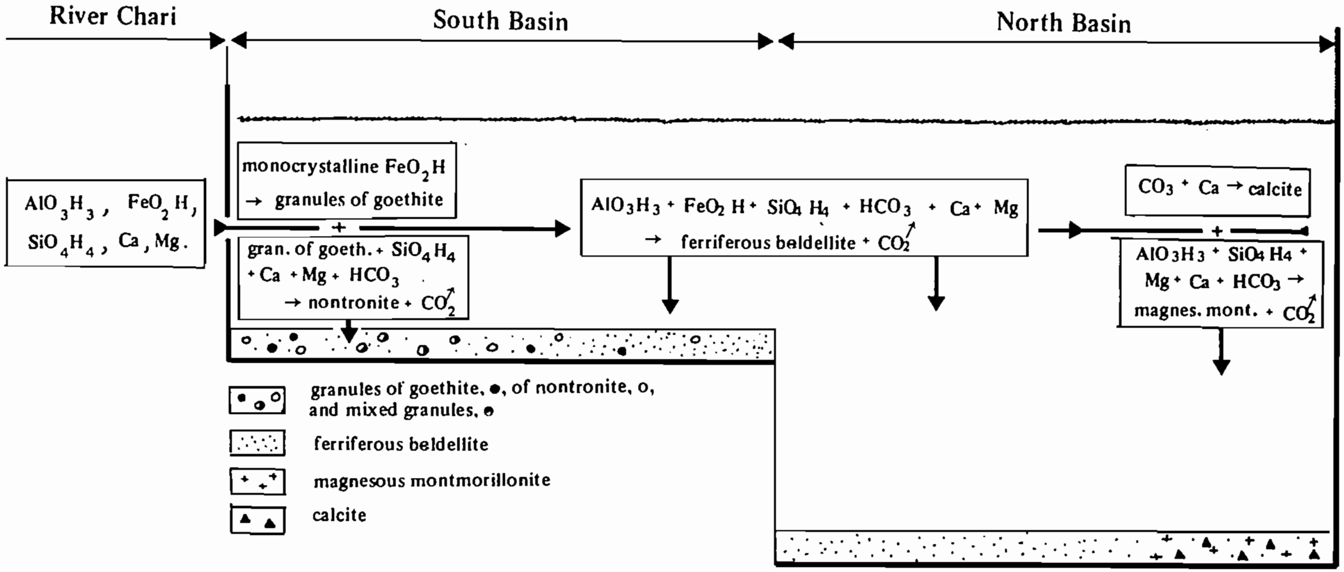

Fig. 7 Simplified geochemical reactions occurring in the lake and their localizations.

SiO_4H_4 (4 moles) and of Mg^{++} (2 moles) and to a lesser extent Ca^{++} (0.20 moles) from the aqueous solution. The large quantity of CO_2 liberated (4.44 moles) attenuates the pH increase due to evaporation.

(d) *In addition* to the neoformation of clays, precipitation of calcium carbonate occurs in the north, where the sediments contain about 5–10% calcite. A detailed study of this phenomenon has been made (Carmouze 1976) showing that the precipitation occurs when the product of the activities of free Ca^{++} and free CO_3^{--}, $\{Ca^{++}\}.\{CO_3^{--}\}$ exceeds 18 times the solubility product of calcite.

The calcite formation results in a reduction in the environment of Ca^{++} and CO_3^{--} and either in a decrease in alkalinity of the water or a lessening restraining its increase due to evaporation.

Figure 7 shows in a simplified form, these four geochemical reactions.

4.4.2 *The biochemical sedimentation*

The biological activity may profoundly affect the hydrochemistry. A given element is more affected when it participates largely in biological cycles of organisms whose biomass and production are high and when its release rate from dead organisms is low. It is indirectly affected when it is in direct relation with one or more elements which are modified by biological activity. With this general consideration, molluscs must have an influence on Ca regulation, the macrophytes on K and SiO_4H_4 and to a lesser extent on Mg and Ca and the diatoms on SiO_4H_4 regulation.

A few observations can be made on the biochemical sedimentation by the molluscs and the macrophytes. Although the sedimentation caused by the diatoms has been studied locally (Lemoalle 1978), it is difficult to draw any conclusion from it.

4.4.2.1 *Influence of molluscs.* Lévêque (1972) has shown the importance of molluscs on Ca regulation by estimating that a population equal to that of 1970 needs 700 000 t of Ca each year. This amount represents 4 times the annual river input of Ca or half the dissolved stock of Ca in the lake. Evidently, without recycling the Ca concentration would have been rapidly diminished in the lake and become zero in the zones furthest from the delta. If the biomass and production of molluscs remains constant and if the liberation of Ca from dying organisms is rapid, the regulation of Ca is not affected by the molluscs. In fact, the populations and the biomass change with lake fluctuations and the dissolution rate of shells varies regionally with the water pH.

The Ca transfers relative to variations in mollusc biomass are difficult to describe because we have little data on this subject. The total biomass changes from 560 000 t to 745 000 t in fresh weight from 1968 to 1970 while the lake level decreased from 281.8 to 281.1 m. In 1972 it fell from a mean of 280.7 m to a value 5 to 10 times lower than that of 1970. At this time the calcium demand must have strongly decreased otherwise the dissolution rate of shells would have changed. This imbalance probably enriched the environment in one or two years from 100 000 to 125 000 tons of calcium which corresponded to 9.3 to 11.7% of the dissolved calcium stock in 1971. There was no increase in Ca stock at this time but as noted in para. 5.3 during the subsidence the increase of Mg and K sedimentation was greater than that of Ca. It may be explained by the influence of molluscs.

The rate of Ca liberation is indicated by the percentage of 'dead shells/live shells'. Lévêque (1972) showed that this percentage is low in the south basin as well as in the southern half of the north basin, proving that the recycling of Ca is largely assured during the period of relative stability of population or of the water level. In the northern half of the north basin the dead shells are relatively abundant and less eroded. They tend to be buried by the actual deposition and then the calcium loss is permanent. But this loss stays low because the mollusc biomass is low in this region. Moreover, at the beginning of their burial, the shells are in contact with the interstitial solutions aggressive against the aragonite, and are then dissolved. These shell levels are not found in recent sediments.

In summary, the permanent calcium losses are greater with increasing distance from the delta; they decrease with a decrease of biomass and increase with an increase in biomass. While the Ca stock tends to increase when the biomass decreases sharply (the fixation rate of Ca by the organisms decreases faster than the rate of Ca release by dissolution of dead shells) it tends to decrease when the biomass grows sharply for the opposite reasons.

4.4.2.2 *Influence of macrophytes.* The particulate materials of fluvial origin are enriched by organic matter during their sedimentation by a factor of 5. The organic matter largely derives from the degradation of macrophytes (Chevery 1974).

The principal macrophytes, *Phragmites australis, Cyperus papyrus, Vossia cuspidata* and *Typha angustiflora* represent 85%, 10%, 4% and 1% of the occupied surface and mostly hold dissolved silica and potassium.

The approximate quantities of K^+, Ca^{++}, Mg^{++}, Na^+ and SiO_4H_4 contained in the macrophytes represent respectively 138×10^3t, 7.4×10^3t, 6.6×10^3t, 2.3×10^3t and 230×10^3t or 170%, 4%, 8.5%, 2.2% and 23.5% of the mean annual inputs of these elements. This represents 11.2%, 0.4%, 0.8%, 0.1% and 6.8% of their corresponding dissolved lake stocks. The amount of K stocked in the plants is relatively higher than that supplied each year by them. Considering that the ratio of production/biomass is about one, these quantities are similar to those which are annually taken up by the macrophytes themselves.

As organic accumulations are registered in some parts of the lake we can infer, from these data, that substantial amounts of K and, to a lesser extent, of

SiO_4H_4 are not recycled and thus represent permanent losses.

The annual amount of dissolved salts fixed by the macrophytes change in the same proportion as the biomass. It may vary greatly because the vegetation area which is about 2500 km^2 at the lake level of 281.5 m can double or be reduced by half respectively at lake levels of 280 and 283 m.

In conclusion, during the flood the biomass is reduced, the mineral salt requirements decrease, the rate of mineralization surpasses that of production and there is a partial restoration of mineral salts to the aqueous phase from the sedimented materials. The chemical sedimentation dependent on the activity of macrophytes must decrease.

During the lake recession the opposite phenomenon occurs. However, during a severe drought, some zones dry up causing the disappearance of macrophytic vegetation. When the waters reflood these regions they tend to be enriched in K and to a lesser degree in Mg (Chantraine 1978).

4.5 The hydrochemical regulation during the period of flood and subsidence

4.5.1 *Environmental instability from hydrological regulation*

(a) Instability of the entire lake against the annual supply–loss imbalance.

Due to the characteristics of its hydrological regulation, Lake Chad at first appears to be very unstable hydrochemically in the face of the irregularity of supplies. This aspect can be shown by a simple example:

Supposing that in an 'average' state, the lake is subjected to an annual supply–loss imbalance of water equal to:

$$\Delta(v)_A = \Delta(v)_F + \Delta(v)_M = \Delta(v)_F$$

For simplification

$$\Delta(v)_M = 0$$

Thus

$$\Delta(v)_A = \Delta(v)_F = \Delta(v)_L$$

Placing $\Delta(v)_F = x(\bar{v})_F$, we know that $(\bar{v})_F = 0.576\ (\bar{v})_L$.
Thus

$$\Delta(v)_F = \Delta(v)_L = 0.576\ x(\bar{v})_L$$

The new lake volume is thus

$$(v)'_L = (\bar{v})_L\ (1 + 0.576\ x)$$

The lacustrine stock of i also changes:

$$\Delta(i)'_L = \Delta(v)_F \times [i]_F = 0.576\ x(\bar{v})_L \times [i]_F$$

Now

$$[\bar{i}]_F = 1/10.8\ [\bar{i}]_L$$

Thus

$$\Delta(i)'_L = 0.053\ x(\bar{v})_L \times [i]_F \text{ and } (i)'_L = (\bar{v})_L \times [\bar{i}]_L\ (1 + 0.053\ x)$$

The new value of i

$$[i]'_L \text{ is} = \frac{(i)'_L}{(v)'_L} = [\bar{i}]_L \frac{(1 + 0.053\ x)}{(1 + 0.576\ x)}$$

Based on this formula when there is an annual excess of 33% in the fluvial supplies ($x = 1/3$), the salt concentration of the lake decreases by 15%. However, when there is a deficit of 33% ($x = 1/3$) the salt concentration increases by 21.5%. Thus considering only the geographical regulation, the variations in salt concentration, corresponding to frequent supply–loss imbalances, are relatively important.

(b) Instability of the south and north basins against the annual supply–loss disequilibria imbalance.

The previous example was used to calculate the instability of the two basins: 50.6% of the total supplies enter the north basin. Thus the volume variation of this basin is:

$$\Delta(v)_{CN} = 0.506\Delta(v)_F = 0.506\ x(\bar{v})_F$$

We know that

$$(\bar{v})_F = 0.888\ (\bar{v})_{CN}$$

Thus

$$\Delta(v)_{CN} = 0.506x \times 0.888\ (\bar{v})_{CN} = 0.449x\ (\bar{v})_{CN}$$

The new volume is:

$$(v)'_{CN} = (\bar{v})_{CN}\ (1 + 0.449\ x)$$

The lacustrine variations of i in this basin are:

$$\Delta(i)_{CN} = 0.918\ \Delta(v)_F\ x[i]_F$$

In fact 91.8% of the fluvial supplies arrive in the north basin (cf. para. 2.2). Thus

$$\Delta(i)_{CN} = 0.918\ x(\bar{v})_F \cdot [i]_F$$

Given that $(\bar{v})_F = 0.888\ (\bar{v})_{CN}$ and that $[i]_F = \frac{1}{15.24}[i]_{CN}$

$$(i)_{CN} = 0.918x \times 0.888\ (\bar{v})_{CN} \times \frac{1}{15.24}\ [\bar{i}]_{CN} = 0.053\ x\ (\bar{v})_{CN}\ [\bar{i}]_{CN}$$

From this

$$(\mathrm{i})'_{\mathrm{CN}}=(\bar{\mathrm{i}})_{\mathrm{CN}}+\Delta(\mathrm{i})_{\mathrm{CN}}=(\bar{\mathrm{v}})_{\mathrm{CN}}[\bar{\mathrm{i}}]_{\mathrm{BN}}\ (1+0.053\ \mathrm{x})$$

Finally,

$$[\mathrm{i}]'_{\mathrm{CN}}=\frac{(\mathrm{i})'_{\mathrm{CN}}}{(\mathrm{v})'_{\mathrm{CN}}}=\frac{[\mathrm{i}]_{\mathrm{CN}}\ (1+0.053\ \mathrm{x})}{(1+0.449\ \mathrm{x})}$$

The remaining 49.4% of the annual water suplies are in the south basin. The volume variation is:

$$(\mathrm{v})_{\mathrm{CS}}=0.494\ \Delta(\mathrm{v})_{\mathrm{F}}=0.494\ \mathrm{x}(\bar{\mathrm{v}})_{\mathrm{F}}$$

We know that

$$(\bar{\mathrm{v}})_{\mathrm{F}}=1.64\ (\bar{\mathrm{v}})_{\mathrm{CS}}$$

Then,

$$\Delta(\mathrm{v})_{\mathrm{CS}}=0.494\mathrm{x}\times 1.64\ (\bar{\mathrm{v}})_{\mathrm{CS}}=0.81\ \mathrm{x}\ (\bar{\mathrm{v}})_{\mathrm{CS}}$$

and

$$(\mathrm{v})'_{\mathrm{CS}}=(\bar{\mathrm{v}})_{\mathrm{CS}}+\Delta(\mathrm{v})_{\mathrm{CS}}=(\bar{\mathrm{v}})_{\mathrm{CS}}\ (1+0.81\ \mathrm{x})$$

The variation in the stock of i in the south basin $\Delta(\mathrm{i})_{\mathrm{CS}}$ is:

$$\Delta(\mathrm{i})_{\mathrm{CS}}=0.082\ \mathrm{x}(\bar{\mathrm{v}})_{\mathrm{F}}\times[\mathrm{i}]_{\mathrm{F}}$$

Being given that $(\mathrm{v})_{\mathrm{F}}=1.64\ (\bar{\mathrm{v}})_{\mathrm{CS}}$ and that $[\mathrm{i}]_{\mathrm{F}}=\frac{1}{2.61}[\bar{\mathrm{i}}]_{\mathrm{CS}}$, we obtain:

$$\Delta(\mathrm{i})_{\mathrm{CS}}=0.082\ \mathrm{x}\times 1.64\ (\bar{\mathrm{v}})_{\mathrm{CS}}\times\frac{1}{2.61}[\bar{\mathrm{i}}]_{\mathrm{CS}}=0.051\times(\bar{\mathrm{v}})_{\mathrm{CS}}\times[\mathrm{i}]_{\mathrm{CS}}$$

and

$$(\mathrm{i})'_{\mathrm{CS}}=(\bar{\mathrm{i}})_{\mathrm{CS}}+\Delta(\mathrm{i})_{\mathrm{CS}}=(\mathrm{v})_{\mathrm{CS}}=(\mathrm{v})_{\mathrm{CS}}\ [\mathrm{i}]_{\mathrm{CS}}\ (1+0.051\ \mathrm{x})$$

Finally,

$$[\mathrm{i}]'_{\mathrm{CS}}=\frac{(\mathrm{i})'_{\mathrm{CS}}}{(\mathrm{v})'_{\mathrm{CS}}}=\frac{[\mathrm{i}]_{\mathrm{CS}}\ (1+0.051\ \mathrm{x})}{(1+0.81\ \mathrm{x})}$$

Based on these two statements, when there is an excess of 33% (x = 1/3) of the fluvial supplies, the salt concentration decreases by 11.5% in the north basin and by 20% in the south basin. However, for a deficit of 33% (x = 1/3) the salt concentration increases by 15% in the north basin and by 34% in the south basin. The instability of the south basin is thus much greater.

4.5.2 *Hydrochemical regulation during a lake reduction*

From 1967 to 1972, the annual supplies were lower than the annual losses and the lake level has diminished by 1.35 m. In such a case the lake volume with a higher turnover rate than that of the saline stocks decreased faster; therefore the salt concentrations as well as the dissolved elements increased. This increase was more marked if the element decreased a little or if the removal coefficient of this stock was low.

The salinity changes can be calculated from the water and salt input deficits. During the period under consideration, the salt input deficit, $\Delta(i)_F$ which is equal to the difference between the supplies of an average regime (see para. 3.1.1) and those of 1967 to 1972, is as follows:

$\Delta(Ca^{++})_F = 5 \times 10^{12}$mM; $\Delta(Mg^{++})_F = 3.8 \times 10^{12}$mM;
$\Delta(Na^{+})_F = 6.7 \times 10^{12}$mM;
$\Delta(K^{+})_F = 2.4 \times 10^{12}$mM; $\Delta(HCO_3^{-})_F = 26 \times 10^{12}$mM;
$\Delta(SiO_4H_4)_F = 19.4 \times 10^{12}$mM

or, in % of the corresponding initial stocks

$\Delta(Ca^{++})_F = 11.1\%$; $\Delta(Mg^{++})_F = 10.5\%$; $\Delta(Na^{+})_F = 6.5\%$

$\Delta(K^{+})_F = 7.7\%$; $\Delta(HCO_3^{-})_F = 9\%$; $\Delta(SiO_4H_4)_F = 34.6\%$

These values indicate that the stock changes are proportional to the stock turnovers.

Assuming that annual outputs have not changed and that the salt stock variations arise only from input deficits, the salt stocks of 1972 could be calculated:

$(i)'_L = (i)_L - \Delta(i)_F$
$(Ca^{++})'_L = 40 \times 10^{12}$mM; $(Mg^{++})'_L = 32.8 \times 10^{12}$mM;
$(Na^{+})'_L = 98.5 \times 10^{12}$mM;
$(K^{+})'_L = 28.8 \times 10^{12}$mM; $(HCO_3^{-})'_L = 266 \times 10^{12}$mM;
$(SiO_4H_4)'_L = 36 \times 10^{12}$mM.

The lake volume has changed from 72.10^9 in 1967 to $45{,}5.10^6 m^3$ in 1972, so the constituent concentration would be:
$(Ca^{+})'_L = 0.88$mM 1^{-1}; $(Mg^{++})'_L = 0.72$mM 1^{-1};
$(Na^{+})'_L = 2.16$mM 1^{-1};
$(K^{+})'_L = 0.63$mM 1^{-1}; $(HCO_3)'_L = 5.84$mM 1^{-1};
$(SiO_4H_4)'_L = 0.80$ mM 1^{-1}.

In fact, the actual total salinity represents 67% of the forecasted value. The values in 1972 were lower than those we have just calculated:

$(Ca^{++})''_L = 18.8 \times 10^{12}$mM; $(Mg^{++})''_L = 20.2 \times 10^{12}$mM;
$(Na^{+})''_L = 78.6 \times 10^{12}$mM;

$$(K^+)''_L = 19.7 \times 10^{12}\,mM;$$
$$(CO_3H^-/CO_3^{--})''_L = 172 \times 10^{12}\,mM;$$
$$(SiO_4H_4)''_L = 25.85 \times 10^{12}\,mM.$$

Initially this difference $(i)' - (i)''$ can be explained by an increase in the infiltration and sedimentation losses, taking as reference the outputs which occur when the lake is in the steady-state. But the infiltration can hardly increase because, although the waters which infiltrate are more concentrated during the subsidence, the infiltration volume diminishes because the infiltration zones are reduced and more impermeable when they are located toward the middle of the lake. In the case of sodium, the increased loss by sedimentation cannot be taken into account.

A very flat relief of the lake basin actually favours the isolation of bordering ponds. A small quantity of salt is then trapped and deposited when these ponds dry up.

Thus, for sodium, we assume that the increase in losses $\Delta(Na)_p$ is due only to these marginal losses $\Delta(Na)_{Pm}$ which represent 18.7% of the initial stock of 1967. The fraction corresponding to the marginal losses of the other elements is slightly different, depending on the stock distributions in space. For Ca^{++}, Mg^{++}, K^+, Na^+, HCO_3^- and SiO_4H_4 the marginal losses are respectively equal to 17.3, 18.1, 18.6, 18.7, 18.2 and 16.4% of the initial stocks.

Thus:

$$\Delta(Ca^{++})_{Pm} = 7.8 \times 10^{12}\,mM;\ \Delta(Mg^{++})_{Pm} = 6.65 \times 10^{12}\,mM;\ \Delta(Na^+)_{Pm} = 19.7 \times 10^{12}\,mM;$$
$$\Delta(K^+)_{Pm} = 5.8 \times 10^{12}\,mM;\ \Delta(HCO_3^-/CO_3^{--})_{Pm} = 53 \times 10^{12}\,mM;\ \Delta(SiO_4H_4)_{Pm} = 9.2 \times 10^{12}\,mM.$$

The higher the stocks are, the greater the losses.

The losses by chemical sedimentation $\Delta(i)_{Ps}$ are calculated from the difference: $(i)_{Ps} = \{(i)' - (i)''\} - (i)_{Pm}$.

$$\Delta(Ca^{++})_{Ps} = 13.45 \times 10^{12}\,mM;$$
$$\Delta(Mg^{++})_{Ps} = 6.0 \times 10^{12}\,mM;\ \Delta(Na^+)_{Ps} = 0;$$
$$\Delta(K^+)_{Ps} = 3.2 \times 10^{12}\,mM;$$
$$\Delta(HCO_3^-/CO_3^{--})_{Ps} = 41.0 \times 10^{12}\,mM;\ \Delta(SiO_4H_4)_{Ps} = 1.70 \times 10^{12}\,mM$$

or in % of the corresponding initial stocks:

$$\Delta(Ca^{++})_{Ps} = 28.9\%;\ \Delta(Mg^{++})_{Ps} = 16.3\%;$$
$$\Delta(K^+)_{Ps} = 10.3\%;\ \Delta(HCO_3^-/CO_3^{--})_{Ps} = 13.9\%;\ \Delta(SiO_4H_4)_{Ps} = 2.9\%$$

Charges in the hydrochemistry from 1967 to 1972 are shown in Fig. 8.

In summary:

— the decrease of the dissolved silica stock is due to a deficit of fluvial supplies (65% of the total decrease of SiO_4H_4 and to marginal deposits (30.5%),

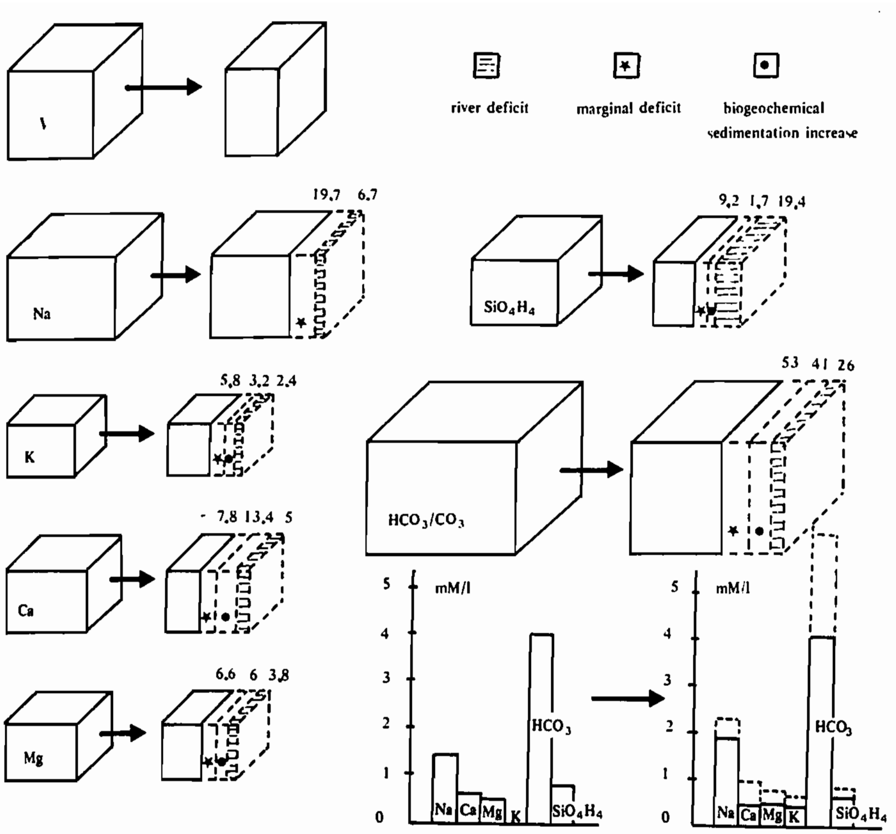

Fig. 8 Changes in the lake salt stocks and the chemical composition of the lake waters during the 1967–1972 lake contractions. The salt stocks variations are expressed in moles × 10^9.

while the increase in losses by biogeochemical sedimentation is only 5.5%;

- the decrease in the sodium stock is caused by marginal deposits (14%) and decrease in fluvial supplies (25.5%);
- the decrease of the calcium stock is mainly caused by the increase in chemical sedimentation (51.2%). Marginal losses and those caused by a decrease in fluvial supplies are respectively 29.7% and 19%.

Finally, the decrease of magnesium, potassium and carbonate-bicarbonate stocks are due to marginal losses (of 40%, 50.9% and 44.1%) and to the increase in chemical sedimentation (respectively 36.5%, 28% and 34.0%). Deficits from fluvial supplies are 21 to 23%.

Thus the decrease in stocks in proportions similar to those of the volume is caused by marginal losses and an increase in chemical sedimentation. By comparison with the values of 1967 Na^+ increased by a third; [K^+] [HCO_3^-/CO_3^{--}] and [Mg^{++}] were less varied while [Ca^{++}] and [SiO_4H_4]

decreased significantly. The total salt content of the waters did not change with the sum of dissolved salts was 7.90mM l^{-1} in 1967 and 7.86mM l^{-1} in 1972.

The chemical composition changed following the flow of matter towards the sodium-bicarbonate pole, which was:

— in 1972

$$[Ca^{++}]_r = 13.7\%;\ [Mg^{++}]_r = 14.6\%;\ [Na^{+}]_r = 57.3\%;$$
$$[HCO_3^-/CO_3^{--}]_r = 87.1\%;\ [SiO_4H_4]_r = 12.9\%.$$

— in 1967

$$[Ca^{++}]_r = 20.8\%;\ [Mg^{++}]_r = 16.8\%;\ [Na^{+}]_r = 47.8\%;$$
$$[HCO_3^-/CO_3^{--}]_r = 89.4\%;\ [SiO_4H_4]_r = 10.6\%.$$

To understand the rhythm of the decrease in stocks from one basin to another, and for a given basin, of one element to another, the ratio $(i)_r/(i)_s$ is supposed to represent the stocks of i in the north and south basins. The relative change of the two basins from 1967 to 1972 is shown by Fig. 9.

The Na^+ and SiO_4H_4 stocks of the south basin decreased more quickly than those of the north basin. The HCO_3^-/CO_3^{--} and K^+ stocks decreased according to a rhythm similar to that of their homologues in the north basin. Finally the

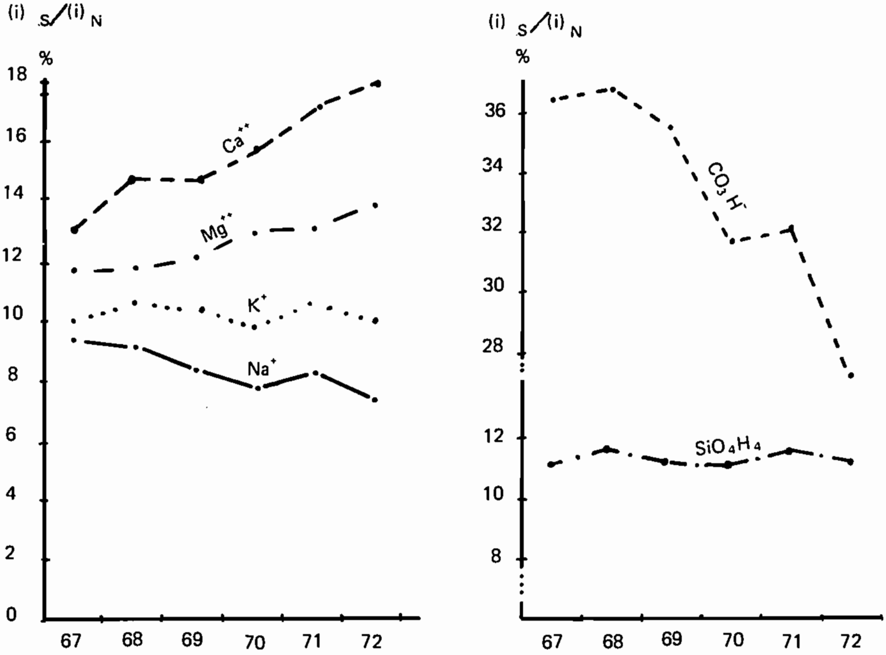

Fig. 9 Relative changes in the salt stocks of the north $(i)_N$ and the south $(i)_S$ basins from 1967 to 1972.

Ca^{++} and Mg^{++} stocks of the south basin decreased less quickly than the corresponding ones in the north basin.

These results show that the factors of the stock decline (deficit of fluvial supplies, increase of sedimentation, marginal losses) have a variable importance from one basin to another not only for a given element but also from one element to another. Unfortunately, it is not possible to calculate satisfactorily the contribution of each of these factors in each of the basins as has been done for the whole lake.

The following changes occurred in the chemical composition of the waters:

- in the south basin, ionic concentrations increased by a third, and the SiO_4H_4 concentrations did not change;
- in the north basin, only the Na^+ concentration increased by 30%; the other elements remained relatively constant (K^+, HCO_3^-/CO_3^{--} and SiO_4H_4) or decreased (Ca^{++} and Mg^{++}).

If relative concentrations are considered, the waters of the south basin developed towards the sodium-bicarbonate pole and those of the north basin towards the sodium pole, ($[HCO_3^-/CO_3^{--}]$ and $[SiO_4H_4]$ remained relatively constant between them).

4.5.3 *Hydrochemical regulation during a flood*

Only two reference points are available for the hydrochemical development of the lake when the water level was increasing: one from 1945 and the other from 1957.

The lake volume in 1945 was similar to that of 1972, attained after a similar lake decrease. Therefore we can assume that the dissolved stocks of 1945 were also similar to those of 1972, because they changed in approximatively the same proportions as volume. Consequently, the total ionized salt stock of the lake in 1945, evaluated with the data of 1972, was equal to 175×10^{12}mé. In 1957 the electrolyte stock, estimated from conductivity data (Guichard 1957) was about 265×10^{12}mé. So, the increase was 90×10^{12}mé, while the lake level rose 2.20m.

This increase arose from excess annual supplies and from partial redissolution of salt deposited on the marginal zone during the preceding contraction of the lake and possibility from a decrease in the annual chemical sedimentation rate. Losses by infiltration were not diminished; on the contrary they may have increased.

The increase of the electrolyte stock caused by excess river supplies between 1945 and 1957 was about 45×10^{12}mé (Carmouze 1976). This represents 50% of the stock increase. Consequently, the remainder, i.e. 45×10^{12}mé, was attributed to the redissolution of the marginal salt deposits* when the lake

---

* The redissolution of salt deposits is about 60% (Carmouze 1976).

increased, and, to a lesser extent, to a diminution of the chemical sedimentation rate.

During the period 1945–1957, the volume of the lake doubled: from 42×10^9 to $84 \times 10^9 m^3$, while the ionized salt stock increase was 50%. Consequently, the mean electrolyte concentration diminished from 4.15mé l^{-1} to 3.25mé l^{-1} (or from 365 $\mu S\ cm^{-1}$ to 290 $\mu S\ cm^{-1}$). However, if the salt stock had only been dependent on the fluctuations of inputs during this period it would have been equal to about 2,60mé l^{-1} or 235 $\mu S\ cm^{-1}$ (80% of the actual value).

This difference points out the importance of the dissolution process of the marginal salt deposits and may be, to a lesser extent, the diminution of the chemical sedimentation rate. They both help to reduce the decrease in salinity caused by geographical factors when the lake increases in size.

4.6 Conclusion

In spite of its endorheic nature and its subarid environment *Lake Chad is not a lake of high salinity*. Three main reasons are responsible for this:

- the river salinity is *low* (about 60 mg l^{-1} or half of the mean salinity of continental waters);
- the climato-geographical regulation of the salinity determines a concentration factor of the river water of about 10.8. This value is not very high for a closed lake situated in an arid zone because the infiltrations are considerable. This factor for the south basin, which is a transit zone, is only 2.60 against 8.45 for the north one;
- the biogeochemical regulation is characterized by major sedimentation of SiO_4H_4, Ca, Mg, HCO_3/CO_3 and to a lesser extent K. The geochemical sedimentation is due to neoformations of smectites and precipitation of calcite. The neoformations of smectites are facilitated by a relatively high concentration of SiO_4H_4 in the river Shari. The salinity of water where calcite precipitation takes place is not very high because the anions are only CO_3H and CO_3. The mollusc and macrophyte contributions to the biochemical sedimentation are surely important because the biomasses and the productions of these two groups are substantial. Then the 'climato-geographical' value of the salinity is reduced by 45%.

The salinity is relatively stable in spite of the high instability of the lake caused by the turnover of water and the dissolved salts. This is mainly due to the existence of marginal salt deposits formed when the lake decreases and partial redissolution of these deposits when the lake increases. These phenomena produce in the first case an additional output of salts, in the second case an additional input, in such a way that dissolved salt stocks vary in approximatively the same proportion as the volume. They are favoured by the morphology of the lake, which is flat and marked by a dune system. For this reason when the lake contracts many ponds are isolated and rapidly dry up. Also, the biochemical sedimentation rate increases when the lake decreases because the biomasses of molluscs and macrophytes increase (this is true only when the decrease is not very strong). The inverse phenomenon takes place when the lake increases.

References

Billon, B. et al., 1969. Monographie hydrologique du Chari. ORSTROM Paris, 5 parts in 6 vols., 610 pp., mimeo.

Carmouze, J. P., 1969. Salures globales et spécifiques des eaux du lac Tchad en 1968. Cah. ORSTOM Sér. Hydrobiol. 3: 3–14.

Carmouze, J. P., 1972. Originalité de la régulation saline du Lac Tchad. C. r. Acad. Sci. Paris 275: 1871–1874.

Carmouze, J. P., 1976. La régulation hydrogéochimique du Lac Tchad. Trav. Doc. ORSTOM No. 58, 418 pp.

Carmouze, J. P., Golterman, H. P. and Pedro, G., 1976. The neoformation of sediments in Lake Chad. Their influence on the salinity control. Interaction between sediments and freshwater. Amsterdam, September 6–10: 33–39.

Carmouze, J. P., Pedro, G. and Berrier, J., 1977. Sur la nature des smectites de néoformation du lac Tchad et leur distribution spatiale en fonction des conditions hydrogéochimiques. C. r. Acad. Paris 284 Sér. D: 615–618.

Carmouze J. P. and Pedro, G., 1977. Contribution des facteurs géographiques et sédimentologiques à la régulation saline d'un milieu lacustre. Cah. ORSTOM Sér. Hydrobiol. 11: 231–237.

Chantraine, J. M., 1978. Evolution hydrochimique du lac Tchad de septembre 1973 à septembre 1975 lors d'une phase de décrue. Cah. ORSTOM Sér. Hydrobiol. 12: 3–18.

Dieleman, J. P. and De Ridder, N. A., 1963. Expertise sur les mouvements des eaux et du sel dans le polder de Bol-Guini. Int. Inst. Land. Rech. Wageningen, 50 pp.

Fontes, J., Maglione, G. and Roche, M. A., Données isotopiques préliminaires sur les rapports du lac Tchad avec les nappes de la bordure nord-est. Cah. ORSTOM Sér. Hydrol. 6: 17–34.

Guichard, E., 1957. Sédimentation du lac Tchad. ORSTOM Fort-Lamy, 46 pp., mimeo.

Lemoalle, J., 1978a. Application des images Landsat à la courbe bathymétrique du lac Tchad. Cah. ORSTOM Sér. Hydrobiol. 12: 83–87.

Lemoalle, J., 1978b. Relations silice-diatomées dans le lac Tchad. Cah. ORSTOM Sér. Hydrobiol. 12: 135–142.

Lévêque, C., 1972. Mollusques benthiques du lac Tchad; écologie, production et bilans énergétiques. Thèse d'Etat, Univ. Paris VI. 225 pp., mimeo.

Maglione, G., 1974. Géochimie des évaporites et silicates néoformés en milieu continental confiné. Thèse d'Etat Univ. Paris VI, 331 pp., mimeo.

Pedro, G., Carmouze, J. P. and Velde, B., 1978. Peloïdal nontronite formation in recent sediments of the lake Chad. Chem. Geol. 23: 139–149.

Pirard, F., 1963. Reconnaissance hydrogéologique du Niger Oriental. BRGM, Orléans.

Roche, M. A., 1969. Evolution dans l'espace et le temps de la conductivité électrique des eaux du Tchad. Cah. ORSTOM Sér. Hydrol. 6: 35–78.

Roche, M. A., 1973. Traçage naturel salin et isotopique des eaux du système hydrologique du lac Tchad. Thèse d'Etat, Univ. Paris VI, 385 pp., mimeo.

Schneider, J. L., 1967. Relation entre le lac Tchad et la nappe phréatique. A.I.H.S. Symposium de Garda, Publ. No. 70: 122–131.

Toucheboeuf de Lussigny, P., 1969. Monographie hydrologique du lac Tchad. ORSTOM Paris, 169 pp., mimeo.

LAKE CHAD

ecology and productivity
of a shallow tropical ecosystem

edited by J.-P. Carmouze J.-R. Durand C. Lévêque

W. Junk Publishers

LAKE CHAD

Ecology and Productivity of a Shallow Tropical Ecosystem

Edited by

J.-P. CARMOUZE, J.-R. DURAND and C. LÉVÊQUE

1983 **Dr W. JUNK PUBLISHERS**
a member of the KLUWER ACADEMIC PUBLISHERS GROUP
THE HAGUE / BOSTON / LANCASTER

Distributors

for the United States and Canada: Kluwer Boston, Inc., 190 Old Derby Street, Hingham, MA 02043, USA
for all other countries: Kluwer Academic Publishers Group, Distribution Center, P.O.Box 322, 3300 AH Dordrecht, The Netherlands

Library of Congress Cataloging in Publication Data

Main entry under title:

Lake Chad.

(Monographiae biologicae ; v. 53)
Includes bibliographies and index.
1. Lake ecology--Chad, Lake. 2. Biological
productivity--Chad, Lake. 3. Chad, Lake. I. Carmouze,
Jean-Pierre. II. Durand, Jean René. III. Lévêque, C.
IV. Series.
QP1.P37 vol.53 574s [574.5'26322'096743] 83-4288
[QH195.C46]
ISBN 90-6193-106-1

ISBN 90-6193-106-1 (this volume)

Cover design: Max Velthuijs

Copyright

PRINTED IN THE NETHERLANDS