

CHEMICAL AND MINERALOGICAL EFFECTS OF SALINE WATER MOVEMENT THROUGH A SOIL DURING EVAPORATION

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We conducted a column experiment in the field, for 1 yr, to determine the chemical effects of the evaporation of a shallow aquifer through a homogeneous sandy clay loam soil. The experimental arrangement simulated conditions found in the polders surrounding Lake Chad in Central Africa. To characterize the nature of the water transfer, oxygen isotopes were determined in the soil solution. Major soil elements, as well as soluble species, were also measured as a function of depth. The data obtained indicated that the water, moving upward by capillarity through the soil profile induced the transfer of soluble ions from the lower column depths. Evaporation took place just below the soil surface, where soluble salts tended to accumulate: the concentration of the soil solution increased drastically, while some soluble species tended to precipitate. Thermodynamic simulation of direct evaporation of the aquifer water and mass balance calculations of the mass transfer during experimentation suggested the formation of Mg-rich clay minerals at the surface of the soil. Detailed transmission electron microscope observations confirmed these results in showing tiny fibers around detrital particles and longer rigid palygorskite-like laths. Thus, even after a short time, water evaporation of a nonsaline shallow aquifer affected strongly the chemistry of the soil above.

Arid and semiarid regions have great potential as food-producing lands, but a frequent problem with their development is the accumulation of soluble salts by both natural and man-induced processes. Deterioration of soils by irrigation of saline waters has been studied extensively (Kelly and Cummins 1921; Bresler 1967;

Rhoades 1972; Rieu and Cheverry 1976; Ayers and Westcot 1985). In addition to irrigation effects, other problems, like upward movement of low-quality groundwater by capillarity and evaporation processes, can also occur (Gardner and Fireman 1958; Hassan and Ghaibeh 1977; Anderson and Bouma 1977).

To gain practical experience in the domain of intensive evaporation, we began a large program of investigation on the polders around Lake Chad, Republic of Chad, in which the water table may fluctuate up to a depth of 10 cm below the soil surface (Cheverry 1974; Riou 1975; Gac 1979). The aim of the present study of evaporation is to focus on the transport of soluble species through a soil column as a function of depth, to examine the interactions of aquifer water with soil minerals, and to establish mass-balance relationships. The experiment was designed to reproduce the evaporation behavior of the water in a shallow aquifer as closely as possible to that in the natural environment of the polders surrounding Lake Chad. For this reason, the experiment was conducted in the field at N'Djamena, Republic of Chad.

MATERIALS AND METHODS

The soil used for this experiment was sampled to a depth of 30 to 40 cm from the B horizon of an hydromorphic soil near N'Djamena. Particles' sizes were determined according to the USDA limits, by sieving down to a particle diameter of 0.05 mm, then by beaker sedimentation, where settling times were calculated by means of Stock's law. Size fractionation of this soil gave 20% clay, 22% silt, and 58% sand (sandy clay loam). After air-drying, the soil was passed through a 2-mm screen, thoroughly mixed, and packed in 10-cm-diameter plastic cylinders to an average bulk density of 1.24 g cm⁻³. The bottom of each column was placed in a tank filled with water in order that the lower 2 cm of each column was permanently submerged. The water sampled in the Chari River, Chad, had a low EC and SAR, the predominant dissolved species being bicarbonates, calcium, magnesium, and silica (Table 1). The water sur-

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Received for publication 5 May 1986; revised 17 July 1987.

TABLE 1
Chemical composition of the Chari River water

pH	EC	Soluble ions, mmol(+) kg ⁻¹						Si	SAR	
		Na	K	Ca	Mg	HCO ₃	Cl			SO ₄
7.36	0.056	0.13	0.05	0.25	0.17	0.54	.005	.02	.353	0.28

face was covered with a thin layer of paraffin oil to prevent direct evaporation and its volume was maintained constant. The soil columns were placed outside in the same semiarid climatic conditions as a field soil. The columns utilized were of different length, because the rate of evaporation of soil water is influenced strongly by the relative depth of the water table (Gardner and Fireman 1958; Johns 1982). The columns were also exposed for different periods to evaporation: two, A and B, of 11-cm length and a third one, C, of 51-cm length were exposed to evaporation for 75 d (A) or 210 d (B and C). At the end of the experiment, the soil columns were sectioned into 1.7-cm slices for A, 1.6 ± 0.5 -cm slices for B and 4.2 cm-thick slices for C. The soil water was extracted by fractional distillation under vacuum. Oxygen isotopic composition of this water was measured by mass spectrometry; laboratory procedures and chemical data are reported in detail in the published work of Gouvea (1980). The soil samples collected were kept frozen until further analysis.

Both before and after each experiment, the soil samples were studied for mineralogy and morphology. The bulk soil (randomly oriented) and the clay fraction (parallel-oriented) were submitted to x-ray diffraction (Philips diffractometer, CuK α radiation, terminated at 30° 2 θ). Clay slides were glycolated, heated at 490°C for 4 h, or treated with monohydrate hydrazine to determine smectites, chlorites, or kaolinites (Weber and Larqué 1978). Soil aggregates were observed with a scanning electron microscope (resolution of 70 Å, 100 000 \times). They were simply attached to specimen mounts and coated with conducting silver paint. The morphology of the clay fraction was determined by transmission electron microscopy (resolution of 15 Å, 200 000 \times). Samples were prepared according to the method described in detail by McKee and Brown (1977). The soil consisted mainly of quartz, potassic feldspars, and muscovites coated by clays. The clay fraction was mainly smectites (55%) and kaolinites (40%), with some

illites (5%). The semiquantitative percentages were measured by comparing the surface area of the x-ray diffraction picks. Smectites occurred either as flakes with irregular edges or as sheets with curly and rolled up borders. The chemical composition of each bulk soil sample was determined (atomic emission spectrometer). The soluble salts, which precipitated in the soil during the water extraction procedure, were redissolved by mixing the soil for 24 h with distilled water in a ratio of 1:10 and extracting the solutions by centrifugation. To dissolve carbonates, soil samples were washed with a 0.1 N HCl solution, according to the same procedure. This last washing might have partially extracted exchangeable cations. The concentrations of the soluble salts were determined using standard colorimetric and atomic adsorption methods (Krempp 1982).

RESULTS AND DISCUSSION

The arid-zone environmental conditions induced a high evaporation rate when the water table was close to the soil surface: 6.5 and 15 L of water evaporated, respectively, through columns A and B. Only 3.5 L evaporated through the longer column C, however, wherein the evaporation rate was expected to be much more influenced by soil characteristics since the water table was located deeper in the soil (Todd and Kemper 1972).

A separate study on the oxygen isotopic composition of the soil solution was undertaken to characterize the nature of the water transfer. This study is based on results obtained by Gouvea (1980), but because the report was published in French, the essential details are presented here. Oxygen-18 measurements provide a powerful method for studying water movement under temperature gradients in soils. As is well known, an oxygen isotopic fractionation occurs during water-phase changes, due to slight differences in chemical potentials between isotopic species (Fritz and Fontes 1980). In an evaporation process, for example, the liquid phase becomes enriched in heavy isotopes relative to the

vapor phase. Allison and Barnes (1983) described successfully a method for the estimation of evaporation from a bare surface based on the enrichment of natural deuterium. Then, Barnes and Allison (1984) developed a model that predicts the shape of $H_2^{18}O$ depth profiles, resulting from evaporation of water from a bare soil under nonisothermal and quasi-steady-state conditions, which is so for this column experiment. In this model, the liquid phase is assumed to be submitted to a process of convective-diffusion, whereas the vapor-phase movement is regarded only as a result of molecular diffusion through the soil atmosphere (Allison et al. 1983). The depth profiles of oxygen composition of the water contained in the soil are shown in Fig. 1. The $^{18}O/^{16}O$ isotope data are reported relative to the SMOW standard (Craig 1961). Those profiles are slightly different from one column to the other, indicating that the increase in ^{18}O is somewhat dependent on specific characteristics in each column. In column A, the $\delta^{18}O$ values remained virtually constant with depth and close to the aquifer water value ($\delta^{18}O = -2.0\%$); therefore it can be assumed that the vapor flux

was negligible and that all the water flux was in the liquid phase. Below the soil surface, the $\delta^{18}O$ value increased drastically; this isotopic enrichment was most likely a result of evaporation and indicated clearly the boundary between the liquid phase and the vapor phase. According to Barnes and Allison (1984), these conditions apply when the water content is high and restricts the pore space available for vapor movement. In column C, on the other hand, groundwater moving up became isotopically heavier, reflecting progressive fractionation from evaporation starting about 20 cm below the soil surface. In this region, both liquid and vapor movements occurred, the vapor transport region was much thicker compared with column A, and the evaporating front was deeper. Column B seems to follow a more complex scenario because, as in column A, evaporation should have taken place just below the surface, but an enrichment in ^{18}O can be noticed starting at the 6-cm depth. This column was exposed to the wet season, which was not true for column A. Although columns were protected from rainfall, temperature variations might have induced a temporary reverse water flow in the upper part of the column; therefore a progressive mixing between the isotopically heavier water going down and the lightest one moving up could have occurred in between the 2- and 6-cm depths. Similar profiles were reported by Allison and Barnes (1983) as a result of a balance between the upward evaporation flux and the downward diffusion flux occurring in soils subjected to evaporation. This oxygen isotopic study of the soil solution points out clearly the location in the soil columns where evaporation took place and should give information about the zone where soluble salts tended to accumulate.

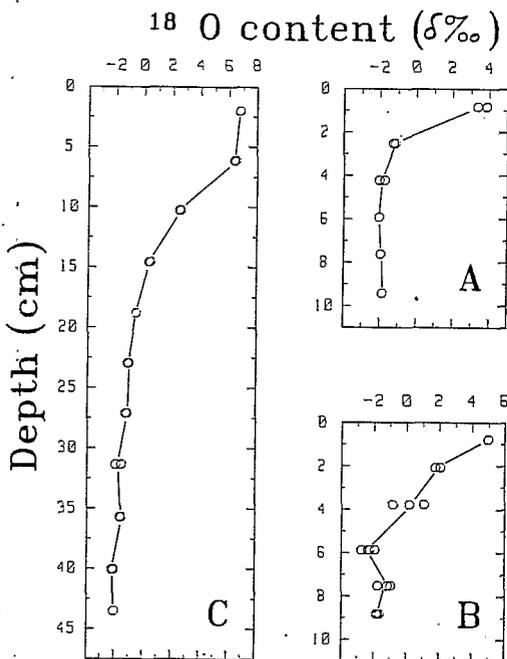


FIG. 1. Oxygen-18 content of the soil solution versus the column depths. The initial content is equal to $-2\% \pm 0.4\%$.

The concentrations of major ions (Si, Al, Fe, Ca, K, Na, Mg, and Ti) were determined for each sample. They did not change significantly according to depth, except for Ca, Mg, and Na (Fig. 2). Those three elements were strongly concentrated in the top section of all three columns (1.7 cm for A and B, and 4.1 cm for C), as a result of water removal from evaporation. To quantify this accumulation process and to determine its origin, simple mass balance calculations were made based on initial and final concentrations and on the exact amount of soluble species carried by the water from the simulated aquifer (Table 2). Mass balances are determined from

the following equation

$$C_{C1} M_1 + C_{C2} M_2 = C_{C0} (M_1 + M_2) + CG MG$$

where

C_C = cation concentration in the soil before (C_{C0}) and after the experiment, in the

upper section (C_{C1}) and the remaining levels (C_{C2});

CG = cation concentration in the experimental aquifer;

M = mass of the upper level (M_1) and lower sections (M_2); and

MG = mass of evaporated water, assuming that the density is equal to 1.

We can also write that

$$E = C_{C1} M_1 - C_{C0} M_1$$

where E = excess of cation in the upper section, after the experiment.

If we assume, according to the oxygen-18 profiles, that most of the soluble species that came from the aquifer accumulated in the upper column section

$$G = CG MG$$

and

$$T = E - G$$

where G = total amount of cations coming from the aquifer and accumulated in the upper section, and T = total amount of cations carried away by the aquifer water, from the lowest levels to the upper section. If we assume that each liter of water carried up the same amount of soluble species from the lower levels to the surface

$$T = C_T M_G$$

with C_T being equal to the cation concentration carried away by 1 L of aquifer water toward the soil surface.

The accumulation of Na, Ca, and Mg was specially high in the top section of the smallest columns A and B (Table 2, C_{C0} compared with C_{C1}) for which the volume of evaporated water was larger, but it did not depend solely on the

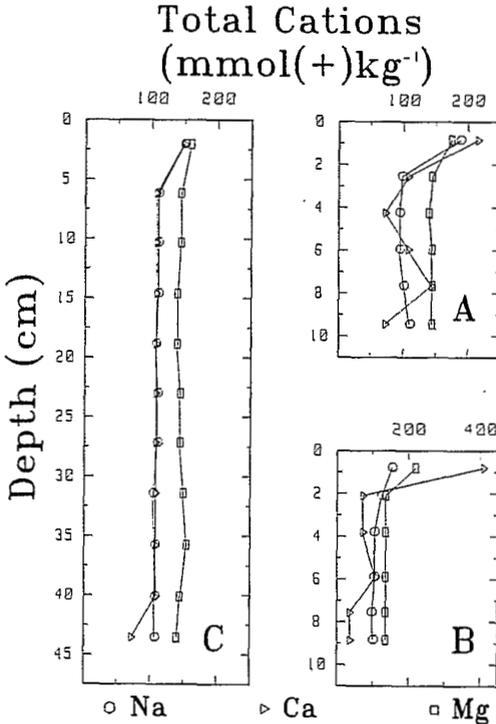


FIG. 2. Sodium Ca, and Mg concentrations of the bulk soil samples after the experiment, as a function of depth. The concentrations in the soil prior to the experiment are equal, respectively, to 111.3, 115.6, and 145.8 mmol(+) kg^{-1} for Na, Ca, and Mg.

TABLE 2

Mass concentrations and mass balance data for Na, Ca, and Mg^a

Columns	M_{C1}	M_{C2} kg 10^{-3}	M_G	C_{C0}	C_{C1} mmol(+) kg^{-1}	C_{C2}	E	G	T mmol(+) kg^{-1}	P	C_T mmol(+) kg^{-1}
Na	A	166.9	854.1	6500	111.3	187.2	12.7	0.8	11.9	4.9	1.8
	B	157.1	785.5	15000	111.3	151.7	104.2	6.3	2.0	4.3	0.3
	C	402.5	3995.6	3500	111.3	148.4	107.8	14.9	0.5	14.4	2.7
Ca	A	166.9	854.1	6500	115.6	214.0	99.2	16.4	1.6	14.8	15.8
	B	157.1	785.5	15000	115.6	407.2	69.5	45.8	3.7	42.1	45.1
	C	402.5	3995.6	3500	115.6	148.0	104.8	13.0	0.9	12.1	2.9
Mg	A	166.9	854.1	6500	145.8	173.6	142.9	4.6	1.1	3.5	4.6
	B	157.1	785.5	15000	145.8	218.2	133.9	11.4	2.5	8.9	11.4
	C	402.5	3995.6	3500	145.8	158.8	144.1	5.2	0.6	4.6	2.1

^aThe column headings are defined in the text.

input of aquifer water, since the lowest levels were depleted in Na, Ca, and Mg, compared with the initial soil (Table 2, C_{C0} versus C_{C1}). The excess of cations determined in the upper section of each column was always much larger than the total input from the aquifer (Table 2, E versus G) and was related to the volume of water that migrated through the columns, as well as to the length (Table 2, T). The amounts of Ca and Mg, carried away by 1 L of water, were similar in columns A and B, while they were much higher in column C, as a consequence of its larger size (Table 2, C_T). The low sodium concentration, observed at the surface of column B, compared with columns A and C, probably reflected the downward flux, pointed out by the oxygen isotopes study, and a redistribution of Na in the lower levels, because it is more mobile than Ca and Mg. Those cations, carried away from the lower levels toward the surface, were probably in the initial soil solution, but they could have been released by cation exchange or by the weathering of highly soluble materials.

Soluble species recovered during the washing procedure have been plotted as a function of depth in Figs. 3 and 4. The concentrations of cations and anions in the water leachates were

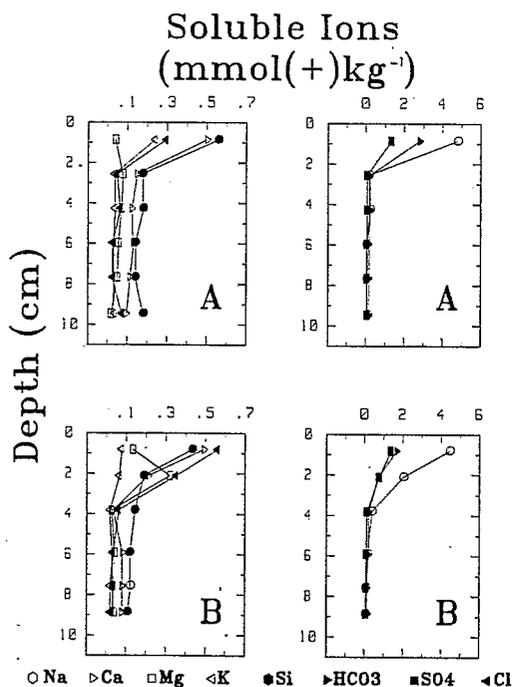


FIG. 3. Ion concentrations in the water leachates versus the depths of columns A and B.

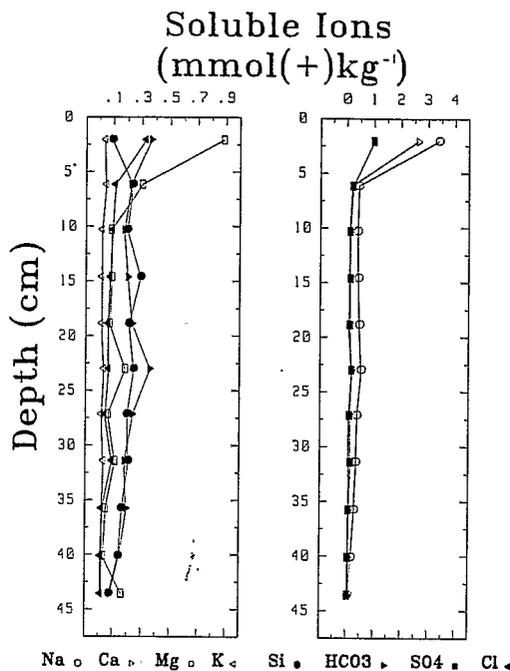


FIG. 4. Ion concentrations in the water leachates versus the depth of column C.

fairly constant with respect to depth and were systematically greater than those in the aquifer water (Table 1), suggesting that the composition of the soil solution was not in equilibrium with the aquifer. Almost all the concentrations increased in the upper level of columns A and C and in the two last sections of column B, except the magnesium concentration, which tended to decrease below the surface of A and B while the potassium one did not increase. On the other hand, at the surface of column C, K, Si, and HCO_3 concentrations were not significantly different from the rest of the column.

Mass balances were performed for Na, Ca, and Mg with

$$P = E - C_w M_{C1}$$

where C_w = excess of cations in the leachate of the upper section, compared with the lower ones; and P = excess of cations in the upper columns' level, which is not removed during washing. Though almost all the sodium in excess below the surface was removed during washing (Table 2, E compared to P), relatively large amounts of Ca and Mg remained in the soil surface of columns A and B. The acid leaching procedure

extracted almost the same amounts of Na, Ca, Mg, and K from the lower levels of the columns as from the surface or the untreated soil.

Because it is unlikely that the cation exchange capacity would increase drastically below the surface, and because we did not observe a large release of Na and K, the amounts of Ca and Mg retained in the soil were probably not involved in cation exchange processes. Ionic strengths, ion activities, and the distribution of complex species were calculated for each washing solution with the computer program EQUIL (Fritz 1975). They were tested for saturation in respect of various minerals, including calcite, dolomite, and gypsum, and they all appeared unsaturated with respect to those minerals. Nevertheless, the washing solutions were much more dilute than the true soil solutions.

As a further step to the present study, evaporation of the water from the Chari River was simulated by a computer program, and the results of the simulation of water-free evaporation were compared to analytical data, where this same water evaporated through soil columns. The computer program EVAPOR (Fritz 1975) calculates chemical evolution in the composition of waters subjected to evaporation on the basis of equilibrium reactions proposed by Garrels and MacKenzie (1967). This program was applied earlier to the evaporation of water from the Chari River, and the results showed good agreement with analytical data of a pan evaporation (Gac et al. 1977). Starting from the chemical composition of the water (Table 1), soluble species are concentrated by successive increments. At each step, the program calculates the ionic strengths, the activity coefficients, and the activities of the free and complex species and tests the state of the solution saturation with respect to minerals included in a bank. As soon as saturation point is reached, the mineral involved precipitates, and the number of moles precipitated is calculated, as are the changes of chemical composition of the solution.

Results of the simulation are reported in Fig. 5, where the logs of total molality of elements are shown as functions of log concentration factors. During the first steps, all molalities increased with the concentration factor; then the molalities of Mg and Ca decreased while that of Si increased more slowly. According to stability constants, the first mineral to reach saturation is a magnesium silicate, having a talclike

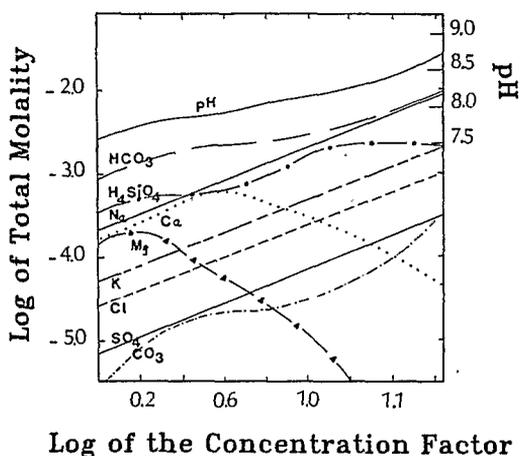
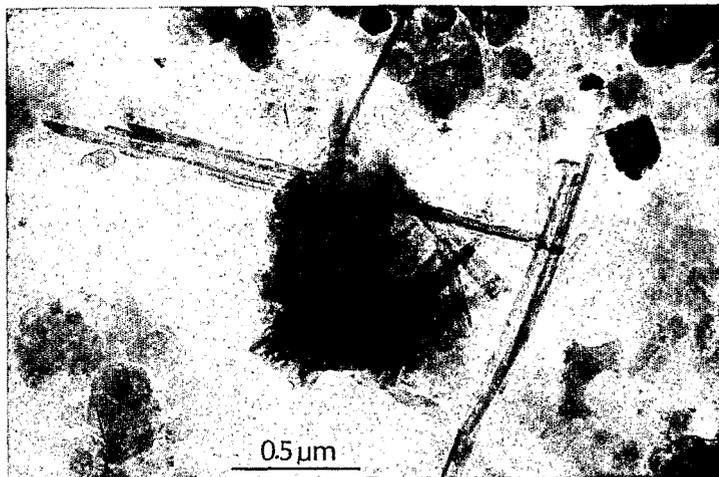


FIG. 5. Simulated evaporation of Chari River water.

formula $[(\text{Si}_4\text{Mg}_3\text{O}_{10}(\text{OH})_2)]$, followed by precipitation of calcite.

Analytical data and mass balance calculations showed that some of the ions, like Mg and Ca, brought to the top of columns A and B during the evaporation experiment were not completely removed during the water and dilute acid leachings. In column B, for example, where 15 L of water evaporated, the observed Mg and Ca in excess would allow the formation of about 1.44 g of Mg-silicate and 4.51 g of calcite, which represents, respectively, 0.9 and 2.9% of the soil of the upper level. The amounts of newly formed clay minerals were too small to be detected by routine methods like x-ray diffraction, as was the calcite by chemical methods (Holtzclaw et al. 1986). Also it was possible that the first phases obtained were amorphous. Each sample, as well as the untreated soil, was examined by transmission electron microscopy. The upper sections of columns A and B revealed that in some cases, the flakelike smectites, characteristic of this soil, appeared to be associated with dense aggregates of very thin laths, 0.2 to 0.3 μm thick (Fig. 6). These laths were not observed in the original material nor in dust that could have contaminated the surface of the columns. Palygorskite-like minerals have been described in various environments. Neof ormation of such minerals was observed in calcareous soils (Paquet and Millot 1973; Eswaran and Barzanji 1974; Singer and Norrish 1974; Yaalon and Wieder 1976; Millot et al. 1977; El Prince et al. 1979), in arid and semiarid climates. Gardner (1972) suggested their formation by direct pre-

FIG. 6. Transmission electron photomicrograph of the surface soil sample of column B.



precipitation from concentrated solutions during evaporation. Millot et al. (1977) and Paquet (1983), on the other hand, thought that they might be derived from transformation of smectites in contact with Mg- and Si-rich soil solutions. The small amounts of newly formed minerals, at the surface of the smaller columns, did not allow any precise chemical analysis. Nevertheless, the highly concentrated soil solution, with large amounts of available soluble Mg resulting from the high evaporation rate, could have favored a recrystallization based on a smectite to palygorskite-like lath mineral conversion.

CONCLUSIONS

An evaporation experiment involving water of known chemical composition moving through homogeneous soil columns of different lengths led to the following conclusions.

1. The determination of oxygen-18 in the soil solutions pointed out exactly the depth where evaporation took place: at the surface for the smaller columns, lower in the profile when the depth increased.

2. The solution moved by capillarity from the aquifer toward the soil surface, where the evaporation took place, therefore, where soluble salts accumulated. Those soluble species were supplied by the aquifer water, but mostly by the lower levels that had been leached by the migrating solution.

3. Leaching of the upper section of the columns with distilled HCl did not remove all the accumulated ions. Specially Ca and Mg stayed in the soil in large quantities.

4. A thermodynamic computer simulation of direct evaporation of the aquifer water suggested that an Mg-rich clay mineral could form. Mass-balance calculations favored this product, which was, in fact, observed in a detailed TEM study as palygorskite-like laths around detrital smectites.

ACKNOWLEDGMENTS

We wish to thank Dr. G. Spósito for his review in preparing the manuscript and R. B. Gouvea for access to oxygen isotopes data.

REFERENCES

- Allison, G. B., and C. J. Barnes. 1983. Estimation of evaporation from non-vegetated surfaces using natural deuterium. *Nature* 301:143-144.
- Allison, G. B., C. J. Barnes, and M. W. Hughes. 1983. The distribution of deuterium and ^{18}O in dry soils: 2. Experimental. *J. Hydrol.* 64:377-397.
- Anderson, J. L., and J. Bouma. 1977. Water movement through pedol soils: 1. Saturated flow. 2. Unsaturated flow. *Soil Sci. Soc. Am. J.* 41:413-423.
- Ayers, R. S., and D. W. Westcot. 1985. Water quality for agriculture. *FAO Irrigation and Drainage Paper* 29, p. 174.
- Barnes, C. J., and G. B. Allison. 1984. The distribution of deuterium and ^{18}O in dry soils: 3. Theory for non-isothermal water movement. *J. Hydrol.* 74:119-135.
- Bresler, E. 1967. A model for tracing salt distribution in a soil profile and estimating the efficient combination of water quality and quantity under varying field conditions. *Soil Sci.* 104:227-233.
- Chevry, C. 1974. Contribution à l'étude pédologique des polders du lac Tchad: Dynamique des sols en milieu continental subaride dans les sédiments argileux et organiques. Thèse ULP, Strasbourg, 275.

- Craig, H. 1961. Standard for reporting concentrations of deuterium and oxygen-18 in natural waters. *Science* 133:1833-1834.
- El Prince, A. M., A. S. Mashhady, and M. M. Aba-Husayn. 1979. The occurrence of pedogenic palygorskite (attapulgite) in Saudi Arabia. *Soil Sci.* 128:211-218.
- Eswaran, H., and A. F. Barzanji. 1974. Evidence of neof ormation of attapulgite in some soils of Iraq. *Trans. 10th Int. Congr. Soil Sci. Moscow* 7:154-161.
- Fritz B. 1975. Etude thermodynamique et simulation des réactions entre minéraux et solutions: Application à la géochimie des altérations et des eaux continentales. *Sci. Geol. Mem.* 41:152.
- Fritz, P., and J. Ch. Fontes. 1980. Handbook of environmental isotope geochemistry—The terrestrial environment. Elsevier, p. 545.
- Gac, J. Y. 1979. Géochimie du bassin du lac Tchad. Bilan de l'altération, de l'érosion et de la sédimentation. Thèse ULP, Strasbourg, 249.
- Gac, J. Y., A. Droubi, B. Fritz, and Y. Tardy. 1977. Geochemical behaviour of silica and magnesium during the evaporation of waters in Chad. *Chem. Geol.* 19:215-228.
- Gardner, L. R. 1972. Origin of the Morman Mesa Caliche, Clark County, Nevada. *Geol. Soc. Am. Bull.* 83:143-156.
- Gardner, W. R., and M. Fireman. 1958. Laboratory studies of evaporation from soil columns in the presence of a water table. *Soil Sci.* 85:244-249.
- Garrels, R. M., and F. T. MacKenzie. 1967. Origin of the chemical compositions of some springs and lakes. *In Am. Chem. Soc. Ser.* 67, pp. 222-242.
- Gouvea Da Silva, R. B. 1980. Migration des sels et des isotopes lourds à travers des colonnes de sédiment non saturé sous climat aride. Thèse, Paris 6, 116.
- Hassan, F. A., and A. S. Ghaibeh. 1977. Evaporation and salt movement in soils in the presence of water table. *Soil Sci. Soc. Am. J.* 41:470-478.
- Johns, G. G. 1982. Measurement and simulation of evaporation from a red earth. *Aust. J. Soil Res.* 20:165-191.
- Kelley, W. D., and A. B. Cummins. 1921. Chemical effect of salts on soils. *Soil Sci.* 11:139-159.
- Krempp, G. 1982. Techniques de prélèvement des eaux naturelles et des gaz associés—Methodes d'analyse des eaux et des roches. *Notes Tech. Inst. Géol. Strasbourg*, 11:59.
- McKee, T. R., and J. L. Brown. 1977. Preparation of specimens for electron microscopic examination. *In Minerals in soil environments.* J. B. Dixon and S. B. Weed (eds.). *Soil Sci. Soc. Am.*, pp. 809-846.
- Millot, G., D. Nahon, H. Paquet, A. Ruellan, and Y. Tardy. 1977. L'épigenèse calcaire des roches silicatées dans les encroûtements carbonatés en pays sub-aride, Anti-Atlas, Maroc. *Sci. Geol. Bull.* 30:129-152.
- Paquet, H., and G. Millot. 1973. Geochemical evolution of clay minerals in the weathered products and soils of Mediterranean climates. *Proc. 4th Int. Clay Congr. Madrid*, pp. 199-206.
- Paquet, H. 1983. Stability, instability and significance of attapulgite in calccrete of mediterranean and tropical areas with marked dry season. *Coll. Int. CNRS, Petrologie des Alterations et des Sols* 2:131-140.
- Rhoades, J. D. 1972. Quality of water for irrigation. *Soil Sci.* 113:277-284.
- Rieu, M., and C. Cheverry. 1976. Mise au point bibliographique sur quelques recherches récentes en matière de sols salés. *Cah. ORSTOM Ser. Pedol.* 14:39-61.
- Riou, C. 1975. La détermination pratique de l'évaporation—Application à l'Afrique Centrale. *Mém. ORSTOM* 80:236.
- Singer, A., and K. Norrish. 1974. Pedogenic palygorskite occurrences in Australia. *Am. Mineral.* 59:508-517.
- Todd, R. M., and W. D. Kemper. 1972. Salt dispersion coefficients near an evaporating surface. *Soil Sci. Soc. Am. Proc.* 36:539-543.
- Weber, and P. Larqué. 1978. Technique de préparation des minéraux argileux en vue de l'analyse par diffraction des rayons X. *Notes Techn. Inst. Geol., Strasbourg*, 34.
- Yaalon, D. H., and M. Wieder. 1976. Pedogenic palygorskite in some brown (calciorthid) soils of Israel. *Clay Miner.* 11:73-79.