

CALIBRATION OF LACUSTRINE SEDIMENT AGES USING THE RELATIONSHIP BETWEEN ^{14}C LEVELS IN LAKE WATERS AND IN THE ATMOSPHERE: THE CASE OF LAKE KINNERET

Mariana Stiller¹ • Aaron Kaufman • Israel Carmi • Genia Mintz

Department of Environmental Sciences and Energy Research, Weizmann Institute of Science, Rehovot 76100, Israel

ABSTRACT. The source of endogenic organic and inorganic carbon in lacustrine sediments is the dissolved inorganic carbon (DIC) in the lake water. The relation between the radiocarbon levels of DIC in Lake Kinneret and of CO_2 in the atmosphere has been investigated. The ratio of the former to the latter was found to be 0.814 ± 0.013 . This ratio is used for calibrating the age of the sediment according to the natural fluctuations in the atmospheric levels of ^{14}C that occurred during the past 10,000 years.

INTRODUCTION

In many hard-water lakes, the dissolved inorganic carbon (DIC) does not achieve isotopic equilibrium with atmospheric CO_2 because the residence time of the water is too short. Therefore the initial radiocarbon specific activity of organisms and of carbonate minerals formed in these lakes is lower than that of the contemporaneous atmosphere causing the apparent ^{14}C ages of endogenic sediment fractions (i.e. those produced within the lake waters) to be higher than their true ages. This age difference is the well known, ubiquitous “hard water effect” which was recognized long ago by Deevey et al. (1954).

The magnitude of the hard water effect remains constant only during periods of stable atmospheric ^{14}C levels and under stable hydrologic and isotopic conditions in the lake system. Variations in atmospheric ^{14}C levels, such as have occurred during the late Pleistocene and the Holocene (Stuiver et al. 1993) must have induced concomitant variations in the ^{14}C content of hardwater lakes, and in the magnitude of the hardwater effect in their sediments.

During the past 50 years, the atmospheric testing of nuclear weapons followed by the almost complete cessation of these tests, induced a sudden rise and a subsequent, more subdued, decline in atmospheric $^{14}\text{CO}_2$ levels (Levin and Kromer 1997:205 and references therein). A unique opportunity was thus provided to determine the response of the ^{14}C level of DIC in any particular hard-water lake to these global fluctuations in atmospheric ^{14}C levels. Based on this response, one can deduce past changes in the magnitude of the hard-water effect of that lake as far back as the known atmospheric record, or about 10 millennia.

In this report we will describe the changes in the ^{14}C levels of the DIC of Lake Kinneret as they occurred during the last three decades, in response to nuclear testing induced changes in the ^{14}C levels of atmospheric CO_2 . We will then determine a quantitative relationship between these two sets of changes. Finally, this relationship will be used to construct a table that permits us to determine the magnitude of the hard water effect in Lake Kinneret and the true age of any sedimentary horizon.

Lake Kinneret

Lake Kinneret (Figure 1), also known as the Sea of Galilee, is a hard-water lake (52 mg Ca L^{-1} and $0.65 \text{ g total dissolved salts L}^{-1}$), situated at $32^\circ 48' \text{N}$, $35^\circ 35' \text{E}$. The lake is monomictic, i.e. its water column overturns once a year at the onset of the cooler, rainy season (December to February). When the lake level is at -211 m MSL , it has a surface of 166.7 km^2 , a volume of $4066 \times 10^6 \text{ m}^3$ and a max-

¹Corresponding author. Email: cistiler@wis.weizmann.ac.il.

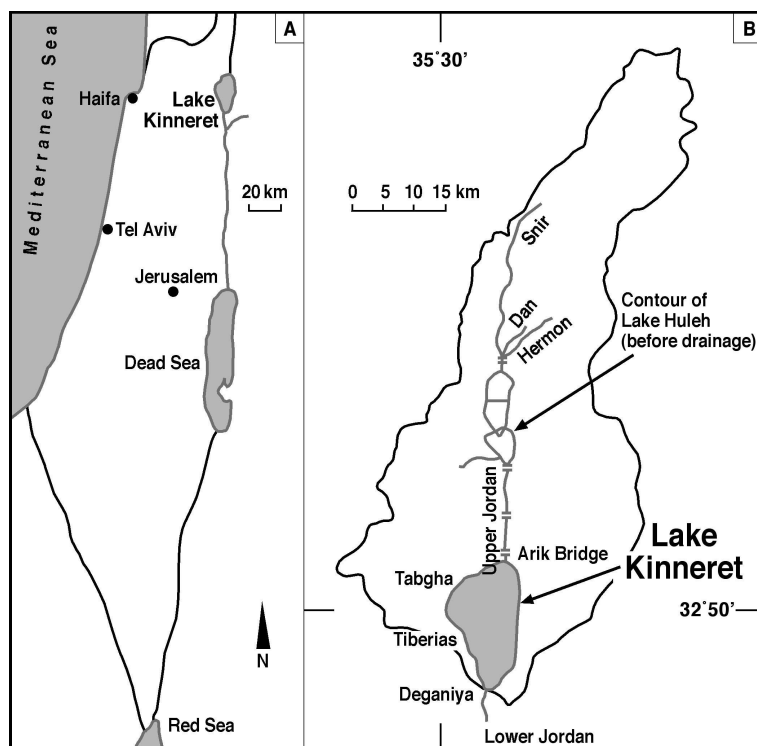


Figure 1 Map of the study area. A) Location of Lake Kinneret, B) The drainage basin of Lake Kinneret.

imal depth of 43 m. Lake Kinneret receives most (66%), of its water from the Jordan River: $477 \times 10^6 \text{ m}^3 \text{ yr}^{-1}$. Other streams and seasonal floods contribute about 16% and underwater springs about 8% of the water influx while direct rainfall constitutes only 9% of the water input. Evaporation is about $250 \times 10^6 \text{ m}^3 \text{ yr}^{-1}$ ($\pm 10\%$). During 1970–1995 the mean residence time of water in Lake Kinneret was about 5.5 years (Nishri et al. 1999). Since 1964, 250 to $500 \times 10^6 \text{ m}^3$ are being diverted annually to the National Water Carrier (NWC); this quantity is similar to the previous natural out-flow (through Deganyia, Figure 1) which has been dammed almost completely.

The concentration of DIC in the Jordan River (Carmi et al. 1985) is about 3.4 mole m^{-3} and in inflowing floodwaters and streams (Nishri et al. 1999) about 3.1 mole m^{-3} . In Lake Kinneret there is less DIC than in its inflows: dissolved carbon uptake by photosynthetic activity of plankton and endogenic precipitation of CaCO_3 reduce the DIC concentration of Lake Kinneret to about 2.5 mole m^{-3} (Stiller and Nissenbaum 1999).

SAMPLING AND MEASUREMENT

In the 1970s and in 1985, water samples were collected from Lake Kinneret and from the Jordan River which were 150 L in volume. DIC was precipitated as BaCO_3 immediately after collection and CO_2 was evolved from the precipitate in the laboratory with HCl (dilution 1:3). Plankton samples were washed with HCl (dilution 1:5) and rinsed with water before the organic carbon was recovered by burning at 1000°C . The resulting CO_2 was transformed into ethane in which form it was counted in a proportional gas counter (Carmi et al. 1971).

In the early 1980s, Talma et al. (1997) extracted CO_2 from smaller volumes (about 10 L) of Lake Kinneret water, and performed the counting in a minicounter. In 1998, CO_2 was extracted from 60 L of water and trapped in 250 mL of 2.5N NaOH solution. The CO_2 was then released from the NaOH solution, transformed into benzene (Gupta and Polach 1985), and its ^{14}C level determined by liquid scintillation counting.

The measured ^{14}C specific activities of samples ($= {}^{14}\text{A}$ per g of carbon) are expressed as ratios relative to the ^{14}C standard specific activity, i.e. $0.95 \times$ the specific activity of the NBS Oxalic Acid Standard nr 1. The symbol for this ratio is ^{14}a , as recently proposed by Mook and van der Plicht (1999); ^{14}a , expressed in %, is commonly called percent Modern Carbon (pMC). The ^{14}a values are then corrected for isotope fractionation by normalizing them for deviations of the measured $^{13}\delta$ from -25‰ ; the symbol for the normalized value is $^{14}\text{a}_\text{N}$, as recently proposed by Mook and van der Plicht (1999). Instead of using the relative activities, ^{14}a and $^{14}\text{a}_\text{N}$, it is also common practice to report them as deviations (of the relative activities) from unity, by: $^{14}\delta = {}^{14}\text{a} - 1$ and $\Delta^{14}\text{C} = {}^{14}\text{a}_\text{N} - 1$, respectively, and express them in ‰ (i.e. multiplied by 1000). In our previous studies (Carmi et al. 1985; Stiller et al. 1988), we have used the symbol X to denote the ratio of the normalized activities (for isotope fractionation) of the sample to the standard. This symbol X is in fact $^{14}\text{a}_\text{N}$.

^{14}C IN LAKE KINNERET FROM 1971 TO 1998

The results of the ^{14}C measurements are reported in Table 1, as $\Delta^{14}\text{C}$ values and as X values as well. After the sudden rise of atmospheric $^{14}\text{CO}_2$ in the early 1960s, its levels started to decline slowly. Our measurements document the response of the lake waters and of the Jordan River to the atmospheric decline.

In the stratified lake of autumn 1971, the ^{14}C levels measured in the upper water layer (2/3 of the lake volume) were higher than in the lower layer (Table 1) by about 40‰. At that time, atmospheric $\Delta^{14}\text{C}$ was +480‰ (Levin and Kromer 1997) and the higher ^{14}C level in the surface waters was probably due to lake-atmosphere exchange of CO_2 during the preceding 8 month long stratification. In late December 1971 the lake overturned and the water column remained well mixed until mid March. Because plankton assimilates DIC, the plankton samples collected in spring 1972 represent the ^{14}C level of DIC during the winter months in the vertically homogeneous water column. The ^{14}C level which probably existed in the mixed water column of spring 1972 has been estimated (from the respective volumes and ^{14}C concentrations of the stratified lake in autumn 1971) to have been $\Delta^{14}\text{C} = +53\text{‰}$. This value is practically identical to that measured in the plankton samples: $\Delta^{14}\text{C} = +49\text{‰}$ and $\Delta^{14}\text{C} = +53\text{‰}$ (Table 1) and is consistent with our assumption that ^{14}C levels of plankton are a good proxy for ^{14}C levels of the lake's DIC.

In 1972 the ratio between $X_{\text{DIC lake}}$ (represented by X_{plankton}) and X_{atm} was 0.688 (Table 2) indicating that during the water residence time of about 5 years, the lake DIC did not achieve isotopic equilibrium with atmospheric CO_2 . It should be noted that each X_{atm} value given in Table 2 is the average of the 5 years which preceded the respective ^{14}C measurement in the lake waters (mean annual X_{atm} values in the Northern Hemisphere were taken from Levin et al. 1985 and Levin and Kromer 1997).

About ten years later, in the early 1980s, $X_{\text{DIC lake}}$ was only a little lower, by about 25‰, than it had been in 1972, though X_{atm} had dropped substantially, by about 250‰ (Figure 2). The ratio $X_{\text{DIC lake}}/X_{\text{atm}}$ was about 0.8 and did not change much until 1998 (Table 2). The larger ratio of $X_{\text{DIC lake}}/X_{\text{atm}}$ during the 1980s and 1990s, indicates that during that period the lake DIC was closer to isotopic equilibrium with the atmosphere than it had been in 1972 (Figure 2).

Table 1 ^{14}C measurements in Lake Kinneret and in the Jordan River^a

| Date of sampling | Description | $\delta^{13}\text{C}$ ‰ PDB | $\Delta^{14}\text{C}$ ‰ | X^b | Date of sampling | Sample description | $\delta^{13}\text{C}$ ‰ PDB | $\Delta^{14}\text{C}$ ‰ | X^b |
|---------------------|---|-----------------------------------|----------------------------|--------------------------------|---------------------|--------------------------|-----------------------------------|----------------------------|---------------|
| 2 Sep 1971 | Center of lake, 25 m depth | -7.6 | +25 ± 45 | 1.028 ± 0.045 | | | | | |
| 2 Dec 1971 | Center of lake, 5 m depth | -5.0 | +67 ± 41 | 1.067 ± 0.041 | | | | | |
| 12 Mar 1972 | Plankton (P) and Zooplankton (Z), P/Z=3.5 | -21.9 | +49 ± 49 | 1.049 ± 0.049 | | | | | |
| 26 Mar 1972 | Plankton | -20.7 | +53 ± 21 | 1.053 ± 0.021 | | | | | |
| 18 Feb 1982 | Deganya, surface ^d | | +28 ± 23 | 1.028 ± 0.023 | 22 Jul 1972 | Arik Bridge ^c | -6.3 | -23 ± 23 | 0.977 ± 0.023 |
| 19 Oct 1983 | Deganya, surface ^d | | +22 ± 16 | 1.022 ± 0.016 | 19 Oct 1983 | Arik Bridge ^c | -7.4 | -166 ± 17 | 0.834 ± 0.017 |
| 8 Apr 1985 | Deganya, surface | -5.5 -5.6 | +18 ± 26 -15 ± 12 | 1.018 ± 0.026 0.985 ± 0.012 | | | | | |
| 15 Jun 1998 | Center of lake, surface | -5.1 | -77 ± 5 | 0.923 ± 0.005 | 15 Jun 1998 | Arik Bridge | -9.8 | -219 ± 4 | 0.781 ± 0.004 |
| 15 Jun 1998 | Center of lake, 30 m depth | -7.9 | -83 ± 4 | 0.917 ± 0.004 | | | | | |

^aAll samples are of water unless otherwise specified^b X values are $^{14}\text{a}_\text{N}$ in the notation of Mook and van der Plicht (1999)^cCarmi et al. (1985)^dTalma et al. (1997). (The symbol δ in table 17.1 of this reference should read Δ).

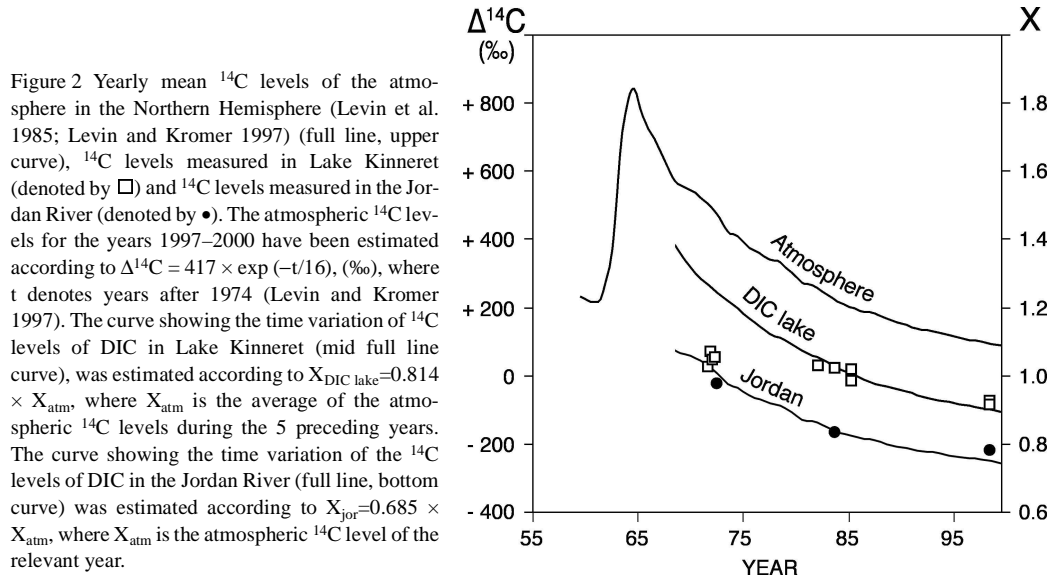


Figure 2 Yearly mean ^{14}C levels of the atmosphere in the Northern Hemisphere (Levin et al. 1985; Levin and Kromer 1997) (full line, upper curve), ^{14}C levels measured in Lake Kinneret (denoted by \square) and ^{14}C levels measured in the Jordan River (denoted by \bullet). The atmospheric ^{14}C levels for the years 1997–2000 have been estimated according to $\Delta^{14}\text{C} = 417 \times \exp(-t/16)$, (‰), where t denotes years after 1974 (Levin and Kromer 1997). The curve showing the time variation of ^{14}C levels of DIC in Lake Kinneret (mid full line curve), was estimated according to $X_{\text{DIC lake}} = 0.814 \times X_{\text{atm}}$, where X_{atm} is the average of the atmospheric ^{14}C levels during the 5 preceding years. The curve showing the time variation of the ^{14}C levels of DIC in the Jordan River (full line, bottom curve) was estimated according to $X_{\text{jor}} = 0.685 \times X_{\text{atm}}$, where X_{atm} is the atmospheric ^{14}C level of the relevant year.

Table 2 Ratios of ^{14}C levels: Lake Kinneret/atmosphere and Jordan River/atmosphere

| Date | $X_{\text{DIC lake}}$ | X_{atm}^a | $X_{\text{DIC lake}}/X_{\text{atm}}$ | X_j | X_{atm}^b | X_j/X_{atm} |
|----------|-----------------------|--------------------|--------------------------------------|-------|--------------------|----------------------|
| Mar 1972 | 1.051 | 1.527 | 0.688 | 0.977 | 1.472 | 0.664 |
| Feb 1982 | 1.028 | 1.283 | 0.801 | | | |
| Oct 1983 | 1.022 | 1.262 | 0.810 | 0.834 | 1.234 | 0.676 |
| Apr 1985 | 1.001 | 1.230 | 0.814 | | | |
| Jun 1998 | 0.920 | 1.105 | 0.832 | 0.781 | 1.093 | 0.715 |
| 1982–98 | | | 0.814 ± 0.013 | | | |
| 1972–98 | | | | | | 0.685 ± 0.027 |

^aAverage of preceding 5 yr, calculated from the data of Levin and Kromer (1997). For 1997 and 1998, X_{atm} was estimated from the relationship (Levin and Kromer 1997) $\Delta^{14}\text{C} = 417 \times \exp(-t/16)$, (‰), where t is years after 1974. To convert from $\Delta^{14}\text{C}$ to X values we used the relation $X = (\Delta^{14}\text{C}/1000) + 1$.

^bCorresponding to the year of measurement

There are two questions that arise from the data in Table 2: 1) why is the ratio $X_{\text{DIC lake}}/X_{\text{atm}}$ lower in 1972 than at all other sampling dates, and 2) why is this ratio practically constant during the 1980s and the 1990s?

We propose the following answer to the first question:

In the very rainy year 1969, the discharge of the Jordan River and of the floods into Lake Kinneret was twice as large as in an average year (Nishri, pers.com.). To avoid flooding the lake's shorelines, $916 \times 10^6 \text{ m}^3$ of the lake's water were released by opening the Deganya dam (Figure 1) and $385 \times 10^6 \text{ m}^3$ were pumped out into the NWC. Therefore the outflow of water in 1969 amounted to about one third (32%) of the lake volume, while in normal years (when the residence time of water is about 5.5 years) it is only about 11.5%. Because one third of the lake volume was replaced in 1969, the lake DIC of the early 1970s also had a shorter residence time (less than 5 years), during which relatively less exchange with the atmosphere could take place.

As regards the second of the above questions: the almost constant $X_{\text{DIC lake}}/X_{\text{atm}}$ ratio (Table 2) calculated for the 1980s and 1990s (0.814 ± 0.013) suggests that both, the hydrological budgets and the exchange rate of CO_2 varied very little during this period. The simple relationship, $X_{\text{DIC lake}} = 0.814 \times X_{\text{atm}}$, which emerges from this constant ratio can be (and will be) used to reconstruct the variations of $X_{\text{DIC lake}}$ which occurred during the past millennia in response to the natural fluctuations in X_{atm} . But before doing this, we will attempt in the next section to get more insight into the above relationship, by examining the ^{14}C budget of Lake Kinneret.

The Components of the ^{14}C Budget in Lake Kinneret

We assume that the lake is at steady state with respect to both, its hydrology and its DIC concentration. Also, we assume that any changes in the atmospheric levels of ^{14}C during the relatively short residence time of the water are small. At this (quasi) steady state, the sum of the inflows of ^{14}C are equal to that of the outflows of ^{14}C and the ^{14}C level of lake water, $X_{\text{DIC lake}}$, is then given by:

$$X_{\text{DIC lake}} = (F_j X_j + F_s X_s + F_m X_m + F_r X_r + F_{us} X_{us} + F_{ex} X_{atm}) / (F_o + F_{ex}) \quad (1)$$

The F 's in the numerator of Equation (1) are the influxes of carbon: F_j by the inflowing Jordan River, F_s by floods and streams, F_m by a group of miscellaneous minor sources, F_r by direct rainfall, F_{us} by underwater springs and F_{ex} by lake-atmosphere exchange. The F 's in the denominator are outfluxes of carbon: F_o by outflow of water and by sedimentation and F_{ex} by lake-atmosphere exchange. The X 's are the corresponding ^{14}C levels of the influxes and outfluxes. Our X notation is identical with the $^{14}\text{a}_\text{N}$ notation of Mook and van der Plicht (1999).

In units of $\text{mole m}^{-2} \text{ yr}^{-1}$: the influxes F_j , F_s and F_m are 9.73, 2.18, and 0.46, respectively (Nishri et al. 1999); F_o (the total outflux) is 13.37 and includes outflow (7.05), sedimentation of CaCO_3 (3.12) and sedimentation of organic matter (3.2) (Nishri et al. 1999; Nishri personal communication). At a steady state concentration of DIC in the lake's water column, F_o equals the sum of the influxes ($F_j + F_s + F_m + F_r + F_{us}$) and therefore $F_r + F_{us}$ turns out to be about $1 \text{ mole m}^{-2} \text{ yr}^{-1}$. This is a minimal estimate of $F_r + F_{us}$ because the possible loss of CO_2 from the lake to the atmosphere at winter overturn (Stiller and Nissenbaum 1999) has not been accounted for in the calculation of F_o .

The three results of ^{14}C in Jordan River DIC (Table 1) are very different from each other with $\Delta^{14}\text{C}$ decreasing from -23‰ in 1972 to -166‰ in 1983 and to -219‰ in 1998. However, their ratios to the corresponding ^{14}C levels in the atmosphere, X_j / X_{atm} , are quite similar, 0.685 ± 0.027 (Table 2; the values of X_{atm} taken for the calculation of this ratio correspond to the year of sampling of the Jordan River). The similarity of the three X_j / X_{atm} ratios suggests that the variations of X_j are generated by the variations in X_{atm} and that the simple relationship $X_j = 0.685 \times X_{\text{atm}}$ may be used to deduce the approximate value of X_j for any year that we know X_{atm} . Because the Jordan River is the major source of DIC to the lake it is important for our further discussion that the ^{14}C level of this source, X_j , can be so easily expressed as a constant times X_{atm} .

We do not have a direct measurement of X_s , i.e. of the ^{14}C level of floods discharging into Lake Kinneret. But, data are available for 4 floods sampled in winter 1982/1983 in the Dead Sea catchment area (Talma et al. 1997). The $X_{\text{flood}} / X_{\text{atm}}$ ratios for the ephemeral floods of Zohar, Hever, Arugot and Kidron turn out to be 0.780, 0.887, 0.709 and 0.761, respectively (average 0.784 ± 0.075 , as in the case of the X_j / X_{atm} ratios of the Jordan River, in the case of floods too, it is appropriate to use X_{atm} values which correspond to the year of the flood sampling). X_{flood} is lower than X_{atm} probably because of interaction between topsoil carbonate particles (Stiller et al. 1988) and rainwater. On the other hand, all of these $X_{\text{flood}} / X_{\text{atm}}$ ratios are higher than the X_j / X_{atm} ratio of 0.676, (Table 2),

observed in the Jordan River in 1983. By analogy with the Dead Sea floods, it is reasonable to assume that the ^{14}C level of the floods entering Lake Kinneret, X_s , is lower than X_{atm} but higher than X_j and can be tentatively expressed as $X_s = B \times X_{\text{atm}}$, where $B = 0.780 \pm 0.078$.

Finally, the magnitudes of X_m , X_r and X_{us} should also be discussed. X_m cannot influence the ^{14}C budget very much because F_m is very small; it will be assigned, as a first approximation, the same magnitude as X_s . The ^{14}C level of direct rainfall, X_r , should be higher than that of the Jordan and that of the floods, while X_{us} may be lower or similar to that of the Jordan; it will be tentatively assumed that their combined value is similar to X_s . We are thus assigning to all of the non-Jordan inflows the common ^{14}C level X_s .

At this point we take the known F 's and rewrite Equation (1), which becomes:

$$X_{\text{DIClake}} = \{(9.73 \times X_j) + (3.64 \times X_s) + (F_{\text{ex}} \times X_{\text{atm}})\} / (13.37 + F_{\text{ex}}) \quad (2a)$$

Finally, we turn to the lake - atmosphere exchange flux, F_{ex} . Though it is not known, we would expect it to be a more or less constant value, as a characteristic feature of Lake Kinneret. Now if we substitute $X_j = 0.685 \times X_{\text{atm}}$ and $X_s = B \times X_{\text{atm}} = 0.78 \times X_{\text{atm}}$, (where B expresses the relationship of ^{14}C levels in Lake Kinneret floods and other minor DIC inflows, to X_{atm}) and rearrange Equation (2a) we get:

$$X_{\text{DIClake}} = [\{(9.73 \times 0.685) + (3.64 \times 0.78) + F_{\text{ex}}\} / (13.37 + F_{\text{ex}})] \times X_{\text{atm}} \quad (2b)$$

Equation (2b) shows that X_{DIClake} can be expressed by a constant (the term within the brackets) times X_{atm} . This constant has already been evaluated empirically (Table 2) to be 0.814 ± 0.013 . Thus, we can now also calculate that the lake-atmosphere exchange flux of CO_2 , F_{ex} , is $7.4 \pm 1.5 \text{ mole m}^{-2} \text{ yr}^{-1}$ (the uncertainty derives from the uncertainty in B).

The simple relationship $X_{\text{DIClake}} = 0.814 \times X_{\text{atm}}$ will be used in the next section for reconstructing historical variations of X_{DIClake} in Lake Kinneret, i.e. the variations of the hard water effect back in time.

Variations of the Hard Water Effect and the Calendar Ages of Lacustrine Endogenic Sediments

Variations of X_{atm} during the past millennia have been deduced from the comparison between dendro ages and ^{14}C ages (Klein et al. 1980). These variations are now routinely taken into account when converting ^{14}C ages into calendar ages (Stuiver and Pearson 1986; Pearson and Stuiver 1993).

Since, as shown above, X_{DIClake} is dependent on X_{atm} , variations in X_{DIClake} must have occurred in response to past variations in X_{atm} . In fact, X_{DIClake} is also the initial ^{14}C activity ($X_{0,\text{sed}}$) of endogenic carbonate and of endogenic organic matter which settle at the lake bottom, and the hard water effect is the age equivalent of $X_{0,\text{sed}}$ (and/or of X_{DIClake}). It follows that the magnitude of the hard water effect was not constant back in time and therefore straightforward dating (by exponential decay of $X_{0,\text{sed}}$) is not feasible because the relevant $X_{0,\text{sed}}$ of a measured lacustrine sample ($X_{t,\text{sed}}$) is not known until we know its age, t_{sed} .

A similar problem exists in dating marine samples; the ^{14}C activity of the ocean, and therefore also the initial ^{14}C of marine samples, varied back in time in response to the atmospheric fluctuations. Stuiver and Braziunas (1993) have addressed this problem by modeling the response of the global ocean to atmospheric ^{14}C changes. Then, they constructed curves and tables back to 10,000 BC,

which relate the modeled, conventional ^{14}C ages of marine samples to the corresponding calibrated ages.

We briefly describe below how to overcome the difficulty in dating endogenic lacustrine sediments while taking into account the time dependent hard water effect. It should be mentioned that in our earlier attempt to apply this method to Lake Kinneret sediments (where we have also dealt with the contribution of allogenic material; Thompson et al. 1985) we have used the Calibration Table (for past variations of X_{atm} and the respective dendroages) of Clark (1975) and a relationship of $X_{\text{DIC lake}}$ to X_{atm} which was derived from measurements performed in 1971/72 in Lake Kinneret and in the Jordan River. This relationship has now been revised according to our measurements from the 1980s and the 1990s. In the meantime, more accurate Calibration Tables also became available (Stuiver and Pearson 1993; Pearson and Stuiver 1993; Stuiver et al. 1998).

$X_{t,\text{sed}}$ is the measured ^{14}C level of the endogenic sediment sample and $X_{0,\text{sed}}$ was the initial ^{14}C level of that sample at the time of deposition, t_{sed} years ago. (As mentioned above, the age equivalent of $X_{0,\text{sed}}$ is the hard water effect). Because $X_{0,\text{sed}} = X_{\text{DIC lake}} = 0.814 \times X_{0,\text{atm}}$, it follows that:

$$X_{t,\text{sed}} = 0.814 \times X_{0,\text{atm}} \times \exp(-t_{\text{sed}}/8267) \quad (3)$$

where t_{sed} is the calibrated, calendar age of the sediment. But, Equation (3) emphasizes clearly that t_{sed} cannot be calculated from the measured $X_{t,\text{sed}}$ before we know the relevant, corresponding $X_{0,\text{atm}}$ at the time of sediment deposition, t_{sed} years ago. We have therefore calculated for each $X_{0,\text{atm}}$ value and corresponding calibrated age, (both given in calibration tables), the value of $X_{0,\text{sed}}$ by using $X_{0,\text{sed}} = 0.814 \times X_{0,\text{atm}}$ and that of $X_{t,\text{sed}}$ by Equation (3). A sediment calibration table for Lake Kinneret has thus been generated. An excerpt of our table is shown below (Table 3). To illustrate our method, we have chosen to use the Bidecadal Calibration Tables of Stuiver and Pearson (1993) and of Pearson and Stuiver (1993). Since the t_{sed} that we are looking for is in fact the calibrated age (from the calibration tables), we can find in our sediment calibration table a corresponding t_{sed} for each measured $X_{t,\text{sed}}$.

The mineralogical, chemical and isotopic composition of lake sediments provide us with a record of the past variations which have occurred in lake productivity and species diversity as well as in climate. Therefore, obtaining the calibrated age of a sediment horizon, t_{sed} , is important not only as a mere improvement in dating accuracy; it is an essential tool in reconstructing the history of the lake and its environment.

The time variation of the hard water effect in Lake Kinneret has been computed by expressing the age equivalent of $X_{0,\text{sed}}$ in conventional ^{14}C years and is shown in Figure 3. It can be clearly seen that the hard water effect increased from about 1000 conventional ^{14}C years at 7000 BP, to about 1800 conventional ^{14}C years at 2000 BP. It should be mentioned that our method of calculating the hard water effect and of correcting the ages of endogenic sediments assumes implicitly that the hydrological and ^{14}C budgets did not change much along this time period.

Similar calculations, ^{14}C budgets and Sediment Calibration Tables, are needed for any hard-water lake, if a more precise dating of their sediments is sought.

Table 3 Calibration Table for Lake Kinneret Endogenic Sediment, back to 8000 BP
(an excerpt from a complete bidecadal table)

| Calibrated age ^a and also t_{sed} (BP) | $X_{0,\text{atm}}$ ^a | $X_{0,\text{sed}}$ | $X_{t,\text{sed}}$ |
|---|---------------------------------|--------------------|--------------------|
| 50 | 0.995 | 0.810 | 0.805 |
| 500 | 1.008 | 0.820 | 0.772 |
| 1010 | 0.981 | 0.798 | 0.706 |
| 1500 | 0.985 | 0.801 | 0.668 |
| 2019 | 0.984 | 0.801 | 0.627 |
| 2519 | 0.993 | 0.808 | 0.596 |
| 3019 | 1.003 | 0.816 | 0.556 |
| 3519 | 1.014 | 0.825 | 0.539 |
| 4019 | 1.022 | 0.832 | 0.511 |
| 4519 | 1.044 | 0.850 | 0.492 |
| 5019 | 1.058 | 0.862 | 0.469 |
| 5519 | 1.076 | 0.876 | 0.449 |
| 6019 | 1.078 | 0.878 | 0.424 |
| 6519 | 1.076 | 0.876 | 0.398 |
| 7019 | 1.085 | 0.883 | 0.378 |
| 7509 | 1.089 | 0.887 | 0.358 |
| 7949 | 1.069 | 0.870 | 0.333 |

^aThe calibrated ages and the corresponding atmospheric ^{14}C levels are taken from Stuiver and Pearson (1993) and Pearson and Stuiver (1993); the data are given in Table 1 of both references, in columns four and two, respectively. The atmospheric ^{14}C levels given in $\Delta^{14}\text{C}$ (‰) were converted into X values (or $^{14}\text{a}_\text{N}$ values according to the notation of Mook and van der Plicht [1999]) using the relationship $X = (\Delta^{14}\text{C}/1000) + 1$.

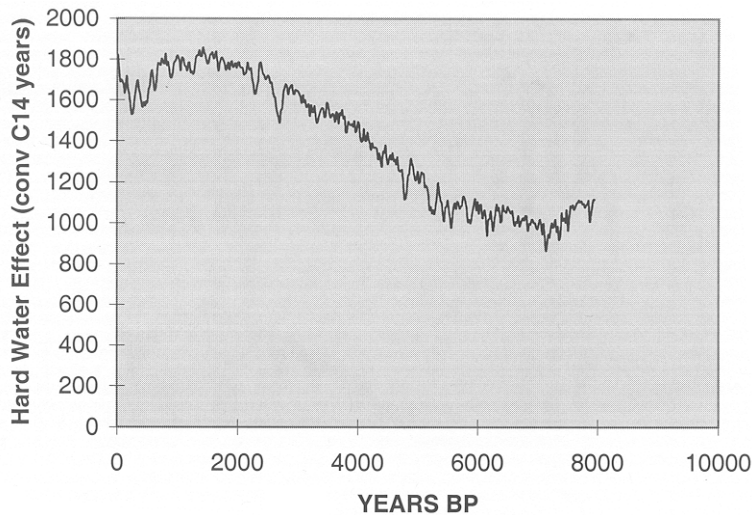


Figure 3 The time variation of the hard water effect in Lake Kinneret back to 8000 BP

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