Climatically induced lake level changes at Lake Van, Turkey, during the Pleistocene/Holocene transition

Günter Landmann and Andreas Reimer
Institute for Biogeochemistry and Marine Chemistry, Centre for Marine and Climate Research
University of Hamburg, Hamburg, Germany

Stephan Kempe
Geological Palaeontological Institute, Technische Hochschule Darmstadt, Darmstadt, Germany

Abstract. Sediment core K10 from Lake Van (eastern Turkey) provides a continuous varve record back to 14,570 calendar years B.P. (before present, 1950), the longest unbroken and non-floating lake varve sequence yet described. The underlying sediment is unvarved and hard. Changes in the aragonite/calcite ratio, the presence of prismatic dolomite and magnesite in certain profile sections, the annual record of the sedimentation rate, the water content of the sediment, the concentrations of organic carbon and opal, and the texture of the sediments from this core provide a record of the lake level history. The new chronology enabled us to redate the old pollen profile [van Zeist and Woldring, 1978a, b] and to establish an accurate timescale for the reconstructed lake level change. Carbon 14 dates show that the highest lake terrace corresponds to high lake level at around 19,600 years B.P. during the Last Glacial, >70 m above its present level. Before 15,000 years B.P. the lake must have been completely dry, marking a reduction of lake level by 500 m in maximum 4000 years. Beginning at 14,600 years B.P. and ending at 12,040 years B.P., the lake level recovered by 250 m to fall again during the next 1400 years. By 10,600 years B.P. the lake began to rise and reached, following another regression between 9000 and 8100 years B.P., the Holocene highstand by about 7500 years B.P., dropping to today's level at about 3000 years B.P.

1. Introduction

Several authors have tried to reconstruct Near East palaeoclimatic events [e.g., Roberts, 1983; Prentice et al., 1992; Roberts and Wright, 1993]. Terminal deep lakes, such as Lake Van, react sensitively to changes in regional evaporation and precipitation and offer the best possibilities to study uninterrupted sedimentary sequences.

Sieger [1894] reviewed the records of Lake Van level changes reported by European travelers for the period of 1800-1888 A.D. These changes had amplitudes of a few meters at most. Levels much higher than at present (1648 m above sea level (asl)) have left conspicuous accumulation and abrasion terraces around Lake Van at 1726 m, 1701 m, 1676 m, and 1658 m asl [e.g., Schweizer, 1975]. Indicators of much lower levels than the present one also exist. Erosion structures and delta bodies were recorded acoustically at a depth of 100 to 200 m in the eastern shelf area of Lake Van [Wong and Degens, 1978; Wong et al., 1978; Wong and Finckh, 1978; Degens et al., 1984]. Chaotic beds were noted below well-stratified layers, interpreted to record drastic lake level changes.

Kempe [1977] used nine sediment cores, recovered from the lake in 1974, to establish the first varve chronology for Lake Van. He suggested a lowering of lake level of up to 300 m for the period 10,000 to 6000 years B.P. This dry period correlates with high concentrations of Chenopodiaceae, Ephedra, and Artemisia pollen [van Zeist and Woldring, 1978a, b]. The onset of more humid conditions is heralded by an increase in tree pollen (mainly Quercus) beginning in 6400 years B.P.

In Lake Urmia, Iran, located 200 km SE of Lake Van, Kelts and Shahrabi [1986] used acoustic profiling to document morphological relicts of a desiccation playa phase, buried below a few meters of lacustrine sediments correlating with gypsum layers in sediment cores. The dry phases are characterized by maxima of Artemisia and minima of Graminaceae pollen, indicating steppe vegetation of a cold, dry climate. Wetter conditions with the Quercus increase appear at 9500 14C years B.P. [Kelts and Shahrabi, 1986].

The general pattern of the pollen record of Lake Zeribar, Iran, located 450 km SE of Lake Van, is very similar to that of Lake Van with regard to the ratio of tree to Gramineae pollen and to the increase in Quercus pollen [van Zeist and Woldering, 1978a]. There is a difference of as much as 3000-4000 years in the timing of the dry phase and of the Quercus increase in the three sites. Kelts and Shahrabi [1986] explain this deviation by differences in dating methods. Lake Zeribar dated by 14C on organic matter, Lake Urmia dated by 14C on aragonitic Artemisia fecal pellets, and Lake Van dated by varve counts, while van Zeist and Woldering [1978b] also discuss the possibility that the difference reflects regional climate factors. More likely, the dating method was
erroneous as suggested by the comparison of the pollen profiles of Söğütlü and Lake Van, which still show a difference of 1100-2200 years in Quercus onset even though they are only 40 km apart [Bottema, 1995].

We now can resolve the time difference between the three records. In 1990, additional cores were recovered from the deep basin of Lake Van, and we established a new varve chronology yielding older dates [Landmann et al., 1996]. We reevaluated the old pollen counts of van Zeist and Woldering [1978a] and investigated the new cores mineralogically and geochemically in great detail. In the present paper these data will be used to reconstruct the water level history of Lake Van allowing us to gain insight into important climatic changes in the Near East since the last glacial maximum (LGM).

2. Geographical Setting and Water Chemistry of Lake Van

Lake Van is located in the mountains of eastern Anatolia, Turkey, at about 43°E and 38.5°N, close to the border to Iran (Figure 1). Its water level stands at 1648 m asl. The lake measures 130 km WSW-ENE and has a surface area of 3522 km². It is 450 m deep and has a volume of 576 km³, making it the fourth largest of all terminal lakes in the world. The depth/area and depth/volume functions are listed in Table 1.

The drainage basin covers 16,096 km² and encompasses the eastern part of the Mus Basin, a large depression structure stretching W-E over a distance of 250 km. The steep southern shore is formed by the Bitlis massif rising to more than 3500 m asl. It is presumably of Paleozoic age, consisting of metamorphic rocks, partly covered by Permo-Carboniferous limestone. Upper Cretaceous limestones and conglomerates locally overlie by

Table 1. Decrease of Area and Volume With Depth for Lake Van for a Water Level at 1648 m asl

<table>
<thead>
<tr>
<th>Depth, m</th>
<th>Area, km²</th>
<th>Area, %</th>
<th>Volume, km³</th>
<th>Volume, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>+100</td>
<td>4930</td>
<td>138.0</td>
<td>1007.7</td>
<td>175.0</td>
</tr>
<tr>
<td>0</td>
<td>3522</td>
<td>100.0</td>
<td>575.9</td>
<td>100.0</td>
</tr>
<tr>
<td>10</td>
<td>3318</td>
<td>94.2</td>
<td>541.3</td>
<td>94.0</td>
</tr>
<tr>
<td>20</td>
<td>2765</td>
<td>83.3</td>
<td>509.4</td>
<td>88.5</td>
</tr>
<tr>
<td>30</td>
<td>2550</td>
<td>73.6</td>
<td>481.7</td>
<td>83.6</td>
</tr>
<tr>
<td>40</td>
<td>2446</td>
<td>70.5</td>
<td>456.0</td>
<td>79.2</td>
</tr>
<tr>
<td>50</td>
<td>2330</td>
<td>67.0</td>
<td>431.4</td>
<td>74.9</td>
</tr>
<tr>
<td>100</td>
<td>1865</td>
<td>53.0</td>
<td>324.2</td>
<td>56.3</td>
</tr>
<tr>
<td>150</td>
<td>1506</td>
<td>42.8</td>
<td>238.2</td>
<td>41.4</td>
</tr>
<tr>
<td>200</td>
<td>1189</td>
<td>33.8</td>
<td>168.9</td>
<td>29.3</td>
</tr>
<tr>
<td>250</td>
<td>889</td>
<td>25.3</td>
<td>115.8</td>
<td>20.1</td>
</tr>
<tr>
<td>300</td>
<td>762</td>
<td>21.6</td>
<td>73.8</td>
<td>12.8</td>
</tr>
<tr>
<td>350</td>
<td>620</td>
<td>17.6</td>
<td>38.5</td>
<td>6.7</td>
</tr>
<tr>
<td>400</td>
<td>405</td>
<td>11.5</td>
<td>11.7</td>
<td>2.0</td>
</tr>
<tr>
<td>440</td>
<td>76</td>
<td>2.2</td>
<td>0.2</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Data are based on the bathymetric map of Lake Van (sea chart no. 9008, department of hydrology and oceanography, Istanbul, 1983). The first line of data provides the area and volume of the 1750 m asl isohypse.
Palaeocene to Eocene marls and carbonates as well as Miocene continental lacustrine deposits dominate the mountains to the east of the lake and occur as isolated outcrops along the southern shore. The area north and west of the lake is almost completely capped by volcanics of Pliocene to Quaternary age, erupted from the stratovolcanoes Nemrut, (3000 m high) and Süphan (4400 m high). Between 100,000 and 200,000 years ago the Nemrut built a lava dam across the Mus Basin, cutting off part of the former Euphrates headwaters (Bendimahi, Zilan). In addition, subsidence helped to form a large internal drainage basin [Degens et al., 1984].

Eastern Anatolia has a continental climate, with warm sum-
mers and long, cold winters. Prevailing southwesterly winds from the Mediterranean bring winter and spring precipitation. Summers are dry with winds from the north and air temperatures of above 20°C in July and August. Along the north and east side of the lake the mean annual precipitation amounts to 300-400 mm, while 600-800 mm fall to the south of the lake, increasing to 1000 mm (mostly as snow) in the southwestern mountains. Most precipitation falls in March and April. The snow melt in the mountains is retarded by 1 to 2 months, resulting in maximum river runoff from late April to June which causes a marked rise of the lake. From July to August, evaporation is higher than inflow, and the lake level starts to drop. Seasonal lake level variations are of the order of 0.5 m. Water balance calculations suggest that rivers add 2.1 km³ yr⁻¹ and that direct precipitation adds another 1.7 km³ yr⁻¹. Evaporation amounts to 3.8 km³ yr⁻¹ [Reimer, 1995]. In years with unusually high or low precipitation the lake level can change by as much as 1 m [Kempe, 1977; Reimer, 1995].

Lake Van water is characterized by a very high alkalinity (153 meq L⁻¹), high pH (9.8), a salinity of 22 parts per million, a Mg concentration of 110 mg L⁻¹ and a very low Ca concentration of 4 mg L⁻¹ [Kempe, 1977; Reimer, 1995]. It is highly saturated with regard to all alkaline earth carbonate minerals; both calcite (CaCO₃) and aragonite (CaCO₃) precipitate at Ca inputs. At river mouths, extensive whitings can be observed, and at sublacustrine groundwater springs and seeps there are large columns (up to 40 m high) of inorganically precipitated calcite tufa and chrombolitic crusts of aragonite, induced by coccolidal cyanobacteria [Kempe et al., 1991].

3. Sedimentary Record From Lake Van

3.1. Cores

In summer 1990, 10 sediment cores were retrieved from depths between 115 m and 446 m in Lake Van. Seven of these cores, recovered up to 30 km apart in the main lake basin, yielded sediment sequences which correlate well, not only with respect to ash layers and prominent color changes, but also lamina for lamina. The varved sequences are interrupted by turbidites occurring with increasing frequency during the periods 11,900-10,600 and after 2600 years B.P. The core that penetrated the oldest layers (K10, length 8.55 m, diameter 58 mm, water depth 420 m, and geographical position 38°32.4'N, 42°48.0'E) (see Figure 2) contains 12 ash layers and is varved down to a depth of 8.22 m. Below, an unvarved hard sediment sequence was encountered which contains large (1-3 cm), rounded pumice pieces terminating with an oolitic layer on top. Because of the high alkalinity of the water, diatoms normally dissolve rapidly and are only pre-

Table 2. Mean Grain Size Distribution of the Sediment Cores K6 and K9

<table>
<thead>
<tr>
<th>Grain Size, µm</th>
<th>K6, wt %</th>
<th>K9, wt %</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt; 0.6</td>
<td>15</td>
<td>15</td>
</tr>
<tr>
<td>0.6-2</td>
<td>38</td>
<td>47</td>
</tr>
<tr>
<td>2-6.3</td>
<td>33</td>
<td>27</td>
</tr>
<tr>
<td>&gt; 6.3</td>
<td>14</td>
<td>11</td>
</tr>
</tbody>
</table>

K6 water depth is 396 m; K9 water depth is 115 m.

Figure 3. Comparison between the new varve count [Landmann et al., 1996] and the old varve count [Kempe, 1977]. Deviations are labeled with numbers and explained in the text.

served in units 3a and 3b. The sediment columns of the northern cores (K7, V8, and K9) are not varved throughout. Partly, core K9 can be correlated with the other cores with the help of the ash layers.

Grain size analyses were conducted on two cores (K6, 27 samples, oldest sample 13,000 years B.P. and K9, 16 samples, oldest sample about 11,000 years B.P.) [Landmann, 1996]. The results of both cores are similar and show a rather homogeneous grain size distribution throughout the Holocene (Table 2). The time period 10,900-13,000 years B.P. is characterized by a 12% higher weight in the size class <0.6 µm which is balanced by a corresponding decrease in the fraction 2-6.3 µm.

3.2. Varve Chronology

K10 was selected for the preparation of overlapping polished plates and thin sections of the entire core. Varve counting and determination of the annual sediment accumulation rates were carried out using a scanner and image analyzing programs, followed by visual correction [Landmann, 1996, Landmann et al., 1996]. The counting error of these combined methods was estimated to amount to ±0.6-1.4%.

Correlation between this chronology and the one obtained by Kempe [1977] was possible comparing the distinctive banding in the new cores with archived photographs of the old cores showing the same banding. The old and new cores could be correlated to better than 50 years. Figure 3 shows that the old count yielded lower dates throughout the record. Even in the time period 3500-14,000 years B.P. the slope of the correlation curve deviates slightly from the 1:1 correlation, indicating that by the optical counting procedure [Kempe, 1977] some years were missed. The largest deviation occurs in the early part of the record (indicated by "1" on Figure 3) where 800 years in the old count are registered as 3500 years in the new count. Kempe [1977] already
Figure 4. Pore water (PW) content, sediment deposition rate (SDR), and mineralogy according to X-ray diffraction analyses of core K10 plotted versus varve years. The intensity of the strongest reflection is used as a proxy of concentration. Additionally, in the dolomite curve the occurrence of protodolomite (squares) and magnesite (triangles) and their Ca/Mg ratios is given. Note the break of the x axis. Dotted vertical lines represent mean values.

pointed out that the largest counting error may be found in this section. Between 200 and 600 years were lost because no pilot core was available in 1974. The other years were probably missed because of a 50% compression of the upper core segment in the piston coring process. Therefore the laminae were much thinner and easily misinterpreted for multilayered annual varves. At "2" the counting in 1974 apparently included a short segment of allochthonous origin, and at "3" an unvarved section was interpolated by using a mean varve thickness. The overall difference in counting adds to a difference of 3000-4000 years in the deeper part of the record.

To compute the annual sediment deposition rate (SDR) from the measured accumulation rate, pore water (PW) content must be taken into account. It was measured on 72 samples in K10 (Figure 4), and cross checks were made on about 300 samples from the other cores. These data and an average dry sediment density of 2.4 g cm⁻³ (mean value of 29 samples from various depths) were used to compute the SDR and plotted versus varve years in Figure 4. There are abrupt changes of SDR at 13,910, 13,380, and 10,920 years B.P., probably reflecting environmental changes [Landmann et al., 1996].

3.3. Redating of the Pollen Record From Lake Van

Pollen counts were made by van Zeist and Woldering [1978a, b] from cores recovered in 1974. This record was redated according to the new varve chronology (Figure 5). Zones 1-3 are characterized by high herbaceous pollen percentages (gray-black), indicating a steppe or desert-steppe vegetation. The increase of Ephedra pollen (zone 2) corresponds to a decrease of tree pollen and reflects extremely arid conditions [van Zeist and Woldering, 1978a]. The maximum of Artemisia pollen (zone 3) coincides with an occurrence of magnesite and with low pore water contents (see Figure 4). The development of the tree pollen is very similar in the redated Lake Van profile to those of Lake Urmiya and Lake Zeribar (Table 3). From 6500 to 3500 years B.P. Quercus dominates the pollen spectrum at Lake Van. A decrease of Quercus appears in the record at around 3500 years B.P. (Figure 5) and has also been reported from Lake Gölbaşı (37°45'N, 37°33'E) [van Zeist et al., 1968].

3.4. Water Content of the Core Sediments

The wet weight of the sample was determined immediately after subsampling. The samples were then freeze-dried and
In the periods 0-2800 and 3400-8000 years B.P., linear downcore decreases in water content are interpreted as caused by compaction. Short-term high water contents in the period 0-3000 years B.P. correlate well with brown sediment bands which contain a higher concentration of organic material. The interpretation of increases in water content at around 3000, 10,500, and 14,500 years B.P. is not as straightforward.

The water content of freshly deposited sediments is determined by grain size, mineralogy, salinity, and sedimentation rate [e.g., Meade, 1964]. As clay mineral composition (see section 3.5) and grain size distribution (see section 3.1) are rather stable throughout the cores, the large water content variations must have a different cause. Neither changes in the sedimentation rate (comparable to the SDR of Figure 4), known to influence water content [see Füchslbauer and Reineck, 1963; Meade, 1966], nor the concentrations of aragonite, calcite, and quartz (Figure 4) show significant correlation with water content in the Lake Van cores. This suggests that the salinity of the lake water during deposition influences the sediment water content. Van der Waal bonds which govern repulsion between mineral surfaces determine the water content of phyllolites, which is reduced by increasing salinities [Meade, 1964]. Similar results were obtained in lab experiments with montmorillonite and fine-grained illite [Hoffmann and Hausdorf, 1945; Bolt, 1956; Mitchell, 1960]. This suggests that water content may be used to reconstruct palaeosalinities and hence lake level changes.

Water content is lowest around 14,600 and 10,700 years B.P. In these sections we also find minerals and sedimentary textures which indicate strong evaporation (see sections 3.4 and 4.1 and Figure 4). The water content curve suggests that lake level recessions may have occurred from 12,800 to 10,700 and from 3400 to 2800 years B.P. and transgressions may have occurred between 14,600 and 12,800 and between 10,600 and 9,000 years B.P. Less pronounced decreases of water content occur at 8700, 6200, and 4800 years B.P.

3.5. Mineral Components of the Core Sediments

X-ray diffraction (XRD) analyses were made on samples of cores K2 and K10; both cores yielded similar results (Figure 4). Micritic carbonates, mostly aragonite and calcite, account for roughly 40% of the sediments [Landmann et al., 1996]. They are mostly authigenic in origin. Satellite images show that whitens are present throughout the lake from July to October. Carbonate precipitation seems to proceed when the Ca-rich surface waters are concentrated because of summer evaporation. Correlation coefficients of -0.1 (K10) and -0.3 (K2) between Corg and Ccalc concentrations in the sediments show that biological productivity does not govern carbonate precipitation. The d values (spacing of crystal lattice) of the calcite vary without a recognizable pattern.

**Table 3.** Tree Pollen Concentrations in Sediment Cores From Lake Van and Lake Zeribar [van Zeist and Woldring, 1978b] and From Lake Urmia [Kelts and Shahrabi, 1986]

<table>
<thead>
<tr>
<th>Tree Pollen</th>
<th>Van Varves</th>
<th>Zeribar [(^{14}C) on (^{13}C_{\text{org}})]</th>
<th>Urmia [(^{14}C) on (^{13}C_{\text{calc}})]</th>
</tr>
</thead>
<tbody>
<tr>
<td>first increase</td>
<td>13.4 - 12.0</td>
<td>13.5 - 12.0</td>
<td>13.5 - 11.8</td>
</tr>
<tr>
<td>low</td>
<td>12.0 - 10.6</td>
<td>12.0 - 10.6</td>
<td>11.8 - 9.6</td>
</tr>
<tr>
<td>strong increase</td>
<td>10.6 - 6.4</td>
<td>10.6 - 5.6</td>
<td>9.6 - 6.4</td>
</tr>
</tbody>
</table>

The ages (kiloyears B.P.) between \(^{14}C\) dating points were linearly interpolated.
between 3.032 and 3.027 Å. This is indicative of a MgCO₃fraction of 2-3.5 mol % [Langbein et al., 1982]. The formation of low-Mg calcite (<4 mol % Mg) from water with a Mg/Ca ratio of >40 [Reimer, 1995] is extremely unusual [e.g., Muller et al., 1972]. Therefore Irion [1973] assumed an early diagenetic formation of the micritic calcite by sulfate reduction in the presence of organic material. This hypothesis is in accordance with the observation that calcite is predominantly found in the dark bands while the light bands have larger proportions of aragonite. However, samples from river mouth whittings and from a sediment trap deployed at a depth of 120 m in the center of the lake (Figure 1) also show similar Ar/Ca ratios as determined for the
sedi\ntents. The carbonates of these samples certainly did not form under sulfate reducing conditions. Probably the lower Mg/Ca ratios in river plume whittings that are visible both on satellite images and from the lake shore allow the formation of calcite. The finer portion of this material is then advected into the lake center where it settles together with the aragonite.

In Figure 4 the intensity of the dolomite (Dol) reflections is plotted versus age. The wide, poorly defined peaks (Figure 6a, 14,520 years B.P.), with maxima between 2° values of 30.8-31.0, indicate a low crystallinity of these dolomites. Comparing the Dol peak intensities with the CaCO$_3$ concentration, a Dol content of 1-3 wt% can be estimated. While calcite (Cc) (r = 0.08) and aragonite (Ar) (r = -0.26) contents do not show any significant correlation with SDR, the correlation coefficient of Dol with SDR reaches 0.65. Since we also found Dol in small amounts in the sediment trap, an allochthonous input of Dol or the current authigenic formation in lake surface waters cannot be ruled out.

Within the upper 15 cm of the sediment column, Mg and Ca concentrations in the pore water decrease from 4.5 to 2.0 and from 0.1 to 0.05 mmol L$^{-1}$, respectively. Additionally, a decrease of pH is observed [Reimer, 1985]. These gradients are consistent with the assumption of an early diagenetic formation of carbonates in the topmost sediments. Many authors assume that dolomite is formed by recrystallization of precursor carbonates in the sediment [e.g., Fuchtbauer and Müller, 1970; Dean, 1993].

However, this can only proceed in pore waters of high Mg/Ca ratios and is aided by high pH values [Müller et al., 1972; Keits and Hsu, 1978]. Apart from low-Mg calcite as the most common carbonate mineral in 46 lakes sampled in North America, Dean and Gorham [1976] also found high-Mg Cc and Dol in waters with high Mg/Ca ratios. In south Australian and Canadian lakes, Mg-Cc, protodolomite, dolomite, and magnesite form with increasing salinity [Alderman and van der Borch, 1963; Nesbitt, 1974, 1990]. Alderman [1965] assumes that the increase of Mg/Ca ratios during evaporation causes the recrystallization of Ar or Cc and the formation of phases richer in Mg. On Sugarloaf Key, Florida and in West Andros, Bahamas a recent Dol formation from Ar can be observed [Shinn and Ginsburg, 1964]. The Mg/Ca ratios in the pore waters of these sediments reach 40.

The Dol reflections in the periods 15,100-14,700 and 9000-8000 years B.P. (Figures 6a and 6c) differ from the usual pattern. The maxima of the more intense peaks are found at 2° values of 30.7 (d = 2.91 A). Theoretically, a surplus of Ca or the inclusion of Fe could cause this shift. However, Fe is removed in the pore waters as Fe-mono sulfide, formed by an excess of HS$^{-}$ generated by sulfate reduction. Hence Fe is not available for carbonate recrystallization [Reimer, 1995]. Also, Fe does not show increased values during the presumed evaporative time periods: FeOtot total amounts to 5 wt % 15,100-14,300 years B.P. and amounts to 2.5 wt % 10,600-7000 years B.P. [Landmann, 1996]. Hence the reason for the Dol peak shift may be the presence of a Ca surplus in the Dol lattice. According to Fuchtbauer and Goldschmidt [1965] the stoichiometric composition of this protodolomite is Ca$_3$SiMg$_4$.

The XRD diagrams of the period 11,160-10,660 years B.P. (Figure 6b) also show conspicuous alterations such as diffuse reflections following the Dol peak, which increases with decreasing age from 31.6 to 32.3 (2° values). These reflections are caused by poorly crystallized magnesite whose lattice parameters are smaller than in stoichiometric, coarse-grained, and well-


crystallized magnesite. Goldschmidt et al. [1955] conclude that the decreasing 2° values correspond to a reduction of the Ca content from Ca$_3$Mg$_7$O$_{12}$. Pronounced Mg concentrations at 10,700 years B.P. were also detected in in X ray fluorescence analyses [Landmann, 1996]. The formation of magnesite and protodolomite in Lake Van sediments must be interpreted as the consequence of an increased Mg/Ca ratio in the lake water. While today's conditions (molar ratio Mg/Ca ~50) would favor the formation of dolomite, the Mg/Ca ratio would be increased further in evaporative periods suitable for the formation of magnesite. This enhances the extraction of Mg from the lake waters and, in the next transgressive phase, leads to a short-term decrease of the ratio, favoring the formation of protodolomite.

On a long-term basis the ratio increases again by Ca extraction caused by the continuous Cc and Ar formation.

The magnesite formation in the second dry phase of the lake is a consequence of the first and is itself the cause for the protodolomite formation in the transgression 9000-8000 years B.P. The older dry phase prior to 15,000 years B.P. is only poorly documented in K10 because the lake apparently dried up completely and left a rather hard layer which could not be penetrated with the corer. The bottom layer is unvarved, shows greatly altered geochemistry, and contains an oolitic layer. Magnesite was not found here, most likely because the initial brine which developed by redissolution of the precipitated salts was not very Mg-rich. This is due to the fact that all Mg had been removed in carbonates which do not redissolve.

The clay minerals of Lake Van sediments were investigated semiquantitatively on 19 samples by Khoo et al. [1978]. They report a relatively stable composition of 12-25% mixed-layer clays, mainly illite-montmorillonite with small contributions of illite-chlorite and chlorite-montmorillonite and 2-10% montmorillonite. Illite, kaolinite, and chlorite were not found in all samples but reach 2-10% in some samples. The average clay mineral content reached 32-34%.

On the basis of microscopic investigations by Khoo et al. [1978] the Lake Van sediments contain about 15% quartz (Figure 4) and 1-2% feldspars (not shown, mostly Na-rich plagioclases, 2° values of 27·6-28·). The relatively low correlation of the concentrations of these minerals with SDR (Qz 0.5 and Fsp 0.4) demonstrates that the SDR is only partly governed by detrital input. Even though feldspar concentrations in some samples seem to indicate higher volcanic input, the correlation coefficient between quartz and feldspar concentrations is at 0.7, suggesting that they are mostly of clastic origin. In addition, muscovite, chlorite and, exclusively in ash layers, analcime (NaAlSi$_2$O$_6$H$_2$O) are present.

4. Discussion

4.1. Reconstructed Lake Level Changes at Lake Van

The oldest point for the reconstruction of the lake level curve (Figure 7) is given by new 14C dates on plant residues from the highest lake accumulation terrace. The samples were recovered from a terrace outcrop near the mouth of the river Engil (Figure 1) between 1674.3 and 1675.6 m a.s.l. and yielded 14C dates of 17,250 ± 115, 17,480 ± 220, and 17,550 ± 220 years B.P. (M.A. Geyh, Niedersächsisches Landesamt für Bodenforschung, Hannover, Germany, personal communication, 1991). According to
the Bard et al. [1990] U/Th 14C calibration these dates translate to a mean calendar age of 19,400 years B.P. The terrace was sampled continuously from 1670 to 1682 m asl and textural analysis and varve counts [Müller, 1994] show that the lake level must have stood at >1720 m asl for at least 500 years, that is, until <18,900 years B.P. These results corroborate earlier assumptions [e.g., Schweizer, 1975; Kempe, 1977; Degens et al., 1984] that the highest erosion terrace at 1725 m asl is contemporary with the LGM.

The lowest layers in K10, recovered from a depth of 420 m, are unvarved hard sediments, containing dolomite, protodolomite, iron oxide schlieren, and rounded pumice pieces, strongly suggesting that the lake was completely dry at around 15,000 years B.P. An oolith layer occurs at a depth of 8.22-8.24 m. Oolith formation requires moving water, indicating a water depth of less than 10 m at this time. Noticeable annual sediment cycles developed in Lake Van immediately above the oolith layer. However, the contrast between light and dark laminae is too weak to detect individual varves. Continuous counting was possible from 14,230 years B.P. upward, yielding an initial accumulation rate of 0.5 mm yr⁻¹. This rate was used to derive an age for the oolith layer (i.e., 14,570 years B.P.) and for the unvarved section by extrapolation.

Desiccation should have left the salt once dissolved in the lake water (today 1.3*10⁹ metric tons total) as a deposit several tens of meters thick on the floor of the lake, forming a playa at the depth of about -410 m. Today the lake would start precipitating NaCl if evaporated down to -390 m. In an alternative scenario one could assume that during desiccation the increasingly dense brine would convect into the pore spaces of the older sediments below the lake bottom, leaving only small amounts of salt and soda to be crystalized on the playa surface. According to this scenario, little salt would be available to be redissolved in early refilling; more would reenter the water column by slow upward diffusion from the pore space over time. Our pore water profiles indicate that this process is still going on: in the cores from the central basin of the lake we found that the pore water salinity doubles from top to bottom of the cores. No such increase is seen in the cores taken from shallower depths [Reimer, 1995]. This scenario would also explain the presence of ostracods in parts of the unvarved section, indicating that the initial water in the lake following its desiccation was not very saline.

For varves to be preserved, a lake must have a certain minimum depth. At Lake Van, judging from cores taken much higher in the basin, varve preservation is only possible in water depths greater than 40-50 m. When the first fully developed varves appear at 14,230 years B.P., the lake level must already have risen to >370 m (Figure 7). Sediments at 350 m depth start to be varved at 13,900 years B.P. [Kempe, 1977, core station 3], that is, the lake level had risen to about -300 m at this time.

The high water content, the decrease in dolomite concentration, the low SDR (Figure 4), and the high Corg concentrations all suggest that this transgression persisted until 12,700 years B.P. Our conclusion that a low SDR is evidence for a high lake level is based on the following: most of the clastic material enters the lake basin with the rivers Bendimahi, Zilan, and Engil from the east. If the wide terraces at a depth of 150-200 m in the eastern part of the lake basin [acoustic investigations by Wong and Degens, 1978] are flooded, then much of the river clastics would deposit there without entering the central basin. Consequently, a low SDR indicates that part of the lake was partly inundated; that is, the water level may have increased to at least -150 m.
At about 12,000 years B.P. the SDR and the dolomite concentrations increase, the pore water content decreases, and turbidites, particularly in the cores from the deeper basin, become more frequent. Magnesite formation commences (Figure 6b), and at 10,920 years B.P. the SDR increases once more abruptly. We interpret this as a further, rather rapid reduction in lake level. There is no interruption in the deposition of varved sediments in K10 (and also in the old station 3 core from 350 m depth), indicating that the lake level did not drop below -310 m. The low stand is also marked by a decreasing aragonite formation, a fact which is also observed in the playa phases 1 and 2 of Lake Urmia [Keiss and Shahrobi, 1986], which can now be correlated with the Lake Van low stand at 15,000 and 10,700 years B.P.

Magnesite formation reached the maximum at 10,660 years B.P. (Figure 6b), and about 100 years later, at the beginning of another strong increase in SDR (Figure 4), a 1-cm-thick, clear-cut layer is deposited, which has twice as much organic carbon (2-3 wt%) as usual. It contains an unusual concentration of long chained alkenones, possibly the trace of intense plankton blooms caused by an eutrophication event triggered by complete overturn of the lake following longer stagnation [Thiel, 1993] [Thiel et al., Unusual distribution of long-chain alkenones and tetrahydronanol from the higly alkaline Lake Van (Turkey), submitted to Geochimica et Cosmochimica Acta, 1996]. This layer may mark the end of the evaporative period. Strong increase of water content (Figure 4) suggests a fast rise of the lake level. A reworked sediment layer in core K8 indicates that the depth of -115 m (water depth of K9) was flooded at 9500 years B.P.

From 9000 to 8100 years B.P. this increase is interrupted once more, denoted by the occurrence of protodolomite and a decrease of water content in the sediments (Figure 4). The sediments also contain a large number of fecal pellets, which are, with regard to their size, structure, and mineral content, similar to the fecal pellets of the halophilic crustacean Artemia, as described from Lake Urmia [Keis and Shahrobi, 1986]. Artemia forms resting eggs that hatch at a salinity of 80-100 parts per million [Gruner, 1993]. If, indeed, these are Artemia pellets, then the lake level must have dropped to about -200 m, the depth corresponding to such a salinity (Figure 7).

At 8100 years B.P. another reworked layer in K9 indicated that the site was flooded again (-115 m). Both the Corg concentrations and Quercus pollen increase. From 6500 to 3500 years B.P., Quercus dominates the pollen spectrum (Figure 5), and the opal concentrations reach values of up to 30 wt% [Landmann et al., 1996]. A more humid environment leads to a stable vegetation cover, an increase in chemical weathering, and hence in the availability of SiO2 and PO4 to the lake ecosystem and to a reduction of the elastic input. Because of this low elastic input, annual varves become very thin in coastal stations. There winnowing of sediments apparently leads to a reduction in SDR by transporting the sediments into the central basin, where somewhat thicker varves are formed. The lake level was probably higher than today, possibly responsible for the pronounced terrace at 1700 m asl. A short regression at about 4800 years B.P. is indicated by a drop in opal concentration and water content and an increase in the dolomite concentration.

The recession of the lake to modern levels seems to have taken place at about 3000 years B.P., as shown by a renewed decrease in the water content of the sediments and an increase in the number of turbidites in all cores. At about 2700 years B.P., the lake level stood about 2-3 m higher than today as can be concluded from the existence of Urartu stone piers at the foot of the Van rock.

4.2. Reconstructed Climate Changes at Lake Van

Similar to Lake Van, a high LGM lake level followed by a dramatic drop after 17,000 14C years B.P. and an elevated level from 12,000 to 10,800 14C years B.P. is also reported from the Konya Basin in southern central Turkey [Roberts, 1983]. Calculations of the water budget for the Konya paleolake indicate that, for the existence of the lake with modern precipitation values, a glacial annual mean temperature 11°-12°C lower than today was necessary [Roberts, 1980]. An integrated palaeoecological study of Lake Zeribar, van Zeist and Bottema [1977] showed that high water levels existed contemporarily with an open stepped vegetation rather than a forest vegetation. LGM lake level maxima have been reported for other lakes in the Mediterranean region as well, for example, for Lake Ioannina, Greece [Higgs and Vita-Finzi, 1966, 1967]. Prentice et al. [1992] were able to show that high lake levels and a vegetation indicative of a dry climate are not mutually exclusive. Their model reveals that 3-6°C lower annual temperatures and a shift of the precipitation maxima to the winter could explain such a situation. Numerical modeling of the general atmospheric circulation at LGM shows for the month of January temperatures of at least 10°C lower than at present in the Near East, an enhanced Siberian high-pressure area, a strengthening and southward displacement of the westerly jet stream and associated storm tracks, and a decrease in precipitation [Karbach et al., 1993]. Regarding vegetation, the winter conditions of extreme cold and precipitation mainly in the form of snow would have been the dominant influences preventing tree growth [Roberts and Wright, 1993]. For July, temperatures were 1°C-2°C lower than today, the subtropical high-pressure area was reduced, and storm tracks were displaced south so that the Mediterranean region experienced some precipitation. The associated cloudiness in the summer may have been especially important in reducing evaporation. Water budget calculations for Lake Van by keeping the runoff factor at the modern level (R = 0.35) reveal that a decrease of evaporation by 21% (today E = 1079 mm yr⁻¹) or an increase of precipitation by 25% (today P = 478 mm yr⁻¹) would lead to the observed increase in lake level at LGM [Landmann, 1996]. The lake drop from 18,900 to 15,000 years B.P. can, for example, be simulated by an increase in evaporation to E = 1500 mm yr⁻¹ and a decrease in precipitation to P = 250 mm yr⁻¹ and a runoff factor R = 0.1.

The lake level curve and the tree pollen distribution show similar trends: during the lake level highstand from 13,400 to 12,000 years B.P. the tree pollen rises slightly as well, while during the lake level recession from 12,000 to 10,600 years B.P. a decrease in tree pollen is noted. This correlation suggests that tree pollen was governed by moisture changes during the late Pleistocene. These periods seem to correlate with the Late Glacial warming period 'known throughout Europe as the Alleröd chro- nozone and with the cold period of the Younger Dryas.

Quercus pollen start to increase rapidly at about 10,600 years B.P. and reach their highest concentration at around 6400 years B.P. This is 2500 years later than the Quercus maximum monitored in northern and northwestern Greece [Bottema, 1974], and 800 years younger than the maximum reported from Lake Zeribar.
and Lake Mirabad, Iran [van Zeist, 1967; van Zeist and Bottema, 1977]. According to the simulation of the climate at 9000 years B.P. [Kuzbach et al., 1993], July temperatures were 2°-3°C higher and winter temperatures about 1.5°C lower than today over the North African-Eurasian landmass. This shift in the timing of the Quercus maximum from west to east appears to indicate a shift in the moisture transport. Robert and Wright [1993] speculate that the depressions of westerly origin, which today advect the moisture during winter to the eastern Mediterranean, may have been less pronounced in the early Holocene because of a stronger transport of winter air masses southward from the Eurasian landmass, blocking the eastward moisture transport.

At 3500 years B.P. a decrease of the Quercus concentration in the Lake Van record was observed. Simultaneously, the sedimentological evidence indicates a recession in lake level. At this time, we can only speculate on the cause of this decrease in effective moisture. It could, for example, indicate a gradual weakening of the westerlies and/or an enhanced cold air transport out of the Eurasian continent.

The reconstructed lake level curve of Lake Van shows three major regressions. Between the LGM highstand, 14C-dated to have lasted until at least 18,900 years B.P. (calendar years) and the first minimum at 15,000 years B.P., the level dropped by almost 500 m, leaving the lake completely dry. The next minimum occurred at around 10,700 years B.P. and again at 8700 years B.P. Apparently, further minor recessions occurred at 6300, 4800, and 3000 years B.P. The two oldest minima correlate with the playa phases 1 and 2 at Lake Urmia [Kels and Shahrabi, 1986], showing that these evaporative periods are not only local events. A high energy environment, that is, shallow conditions, was recorded in Lake Urmia contemporary with the third minimum in Lake Van.

Regressive lake level phases were also reported for the Black Sea at around 15,000 14C years B.P. [Degens and Ross, 1974] and for Lake Tigliamamine, Morocco at 10,000, 6900, 4200, 2900, and 2000 14C years B.P. [Lamb et al., 1995]. In these regions the dry phases lasted for 150 to 400 years and show, similar to Lake Van, decreasing intensity and duration toward the present. Lamb et al. [1995] assumed that reduced winter precipitation was responsible for these dry phases.

The observed minima in Lake Van have time intervals of between 2000 and 4300 years. Cycles of this periodicity (about 2500 years) are common in climate proxy records of the late Quaternary, such as glacial advances in the northern hemisphere [Denton and Karlen, 1973], oxygen isotopes in ice cores of Greenland [Dansgaard et al., 1984] and Antarctica [Benoist et al., 1982], atmospheric 14C variations [Suess, 1980], and lichic grains contents (>150 μm) in deep sea sediments from the North Atlantic [Bond and Louti, 1995]. The causes of these climatic cycles may be a variation of solar activity [Anderson et al., 1990], but the exact mechanism of how they govern the water balance in Lake Van remains to be revealed.

5. Conclusions

The records of the Late Glacial and Holocene climate and environment from Europe and Africa are detailed and internally consistent for each of the two regions [see Huntley and Prentice, 1993; Street-Perrott and Perrott, 1993] but differ from each other. The transition between these two climate regimes must occur around 30°-40° northern latitude. The Near East, with its position at the junction of Europe, Africa, and Asia, is potentially of great significance for the reconstruction of Late Glacial and Holocene climate changes. Lake Van is situated in this critical area. Because of its large depth (450 m) and its seasonally changing CaCO3 precipitation yielding varves, the lake offers the best potential to study regional water balances. Its water level curve, spanning the last 20,000 years, is a valuable addition to the moisture history in this region which can be used to validate climate models.

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