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A coupled calibration and modelling approach to the understanding of dry-land lake oxygen isotope records.

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Abstract

Comparisons between climate proxies and instrumental records through the last two centuries are often used to understand better the controls on palaeoarchives and to find relationships that can be used to quantify changes in pre-instrumental climate. Here we compare an 80-year-long annually resolved oxygen isotope record from Nar Gölü, Turkey, a varved lake sequence, with instrumental records of temperature, precipitation, wind speed, relative humidity and calculated values of evaporation, all of which are known to be possible controls on lake oxygen isotope systems. Significant relationships are found between the isotope record and summer temperatures and evaporation suggesting these are dominant controls on the isotope hydrology of this non-outlet lake. Modelling the stable isotope hydrology of the lake system allows these relationships to be tested independently. We show that the isotope record follows the same trends in the temperature and evaporation records but that, even when combined, these two climatic factors cannot fully explain the magnitude of change observed in the isotope record. The models show the lake system is much less sensitive to changes in evaporation and temperature than the climate calibration suggests. Additional factors, including changes in the amount of precipitation, are required to amplify the isotope change. It is concluded that proxy-climate calibrations may incorrectly estimate the amplitude of past changes in individual climate parameters, unless validated independently.

Introduction

For records of palaeoclimatic change to be useful to climate modellers, and to put the recorded climate variability through the last century into context, it is necessary to try and quantify records of past change. This requires understanding of the modern processes controlling the palaeoarchive. It also requires testing of the assumption that these controls have not changed significantly through time or, if they have, that changes can be sufficiently accounted for. This is particularly important when trying to quantify changes in climate from lake oxygen isotope records as there are many factors that control lake stable-isotope values (Leng and Marshall 2004). Even a single climate variable can have many effects on the lake isotope system. Temperature, for example, not only controls the isotopic fractionation between lake waters and precipitated carbonates but also affects the stable isotope value of precipitation entering the lake. Temperature also plays a large role in controlling the amount of evaporation and this may be the dominant effect in some lake systems (e.g., Jones et al. 2002; Wick et al. 2003).

A number of methods have been used to try and quantify environmental change from lake archives. For example, recent palaeoarchives have been compared to instrumentally-recorded environmental variability thereby allowing the sensitivity of the proxy to be established and a relationship observed that can be extrapolated back in time (e.g., Laird et al. 1998; Verschuren et al. 2000). Previous studies have increased understanding of isotope controls in lake systems by comparisons with climate (e.g., Smith and Hollander 1999) and lake level data (Kirby et al. 2004). In order that such reconstructions are robust, they should ideally be cross-validated against other independent sources of data. For lake stable isotope data, one way to do this is via a modelling-calibration inter-comparison (Teranes and McKenzie 2001).

Lake isotope models are based on mass balance equations for lake water balance and also for stable isotope mass balance. The hydrological budget of a lake can be explained by the sum of the inflows and outflows from the system (e.g., Ricketts and Johnson 1996; Gibson et al. 1999), for example

$$dV/dt = P + S_i + G_i - E - S_o - G_o. \qquad\qquad (1)$$

where V is lake volume; t, time ; P, precipitation on the lake surface; S_i, surface inflow from rivers and/ or overland runoff; G_i, groundwater inflow; E, evaporation from the lake surface; S_o, surface outflow; G_o,

groundwater outflow. (V, P, S_i, G_i, E, S_o and G_o are measured in the same units). Some of these values can

be measured directly e.g., P, S_i, others often have to be calculated e.g., E or estimated e.g., G_o.

As well as the water balance equation above (equation 1) the stable isotope values of the lake

hydrological system must also balance such that

$$dV\delta_l/dt = P\delta_P + S_i\delta_{Si} + G_i\delta_{Gi} - E\delta_E - S_o\delta_{So} - G_o\delta_{Go} \qquad (2)$$

where the values δ_l, δ_p, δ_{Si}, δ_{Gi}, δ_E, δ_{So}, δ_{Go} are the isotope values (either $\delta^{18}O$ or δD) of the lake waters, lake

surface precipitation, surface inflow, groundwater inflow, lake surface evaporation, surface outflow and

groundwater outflow respectively (e.g., Gibson et al.1999; Benson and Paillet, 2002).

δ_l, δ_{Si}, δ_{So} can be measured directly from water samples collected from the lake and catchment. δ_P

can be measured if rainwaters are sampled near the lake site although it is preferable to use the long term

averaged data from the GNIP database (IAEA/WMO, 2001) if there is a station near the site. Samples for

measurement of δ_{Gi} and δ_{Go} are difficult to collect, however δ_{Gi} can be estimated from measurement of

groundwaters in or near the catchment e.g., from wells or springs, and δ_{Go} is the same as lake water and

therefore is usually taken to equal δ_l.

δ_E is very difficult to measure directly and is usually calculated. Many authors (e.g., Gonfiantini

1986; Gibson et al. 1999; Kebede et al. 2002) use equations based on the Craig-Gordon model of evaporation

(Craig and Gordon 1965) such as:

$$\delta_E = ((\alpha \cdot \delta_l) - h\delta_A - \epsilon)/(1 - h + \epsilon_k) \qquad (3)$$

where α^* is the equilibrium isotopic fractionation factor dependent on the temperature at the evaporating

surface, h is the relative humidity normalised to the saturation vapour pressure at the temperature of the air

water interface (lake surface temperature); $\epsilon = \epsilon^* + \epsilon_k$; $\epsilon^* = 1000 (1 - \alpha^*)$ and ϵ_k is the kinetic fraction factor.

δ_A is the isotopic value of the air vapour over the lake.

An alternative equation for calculating δ_E was obtained from observations at Pyramid Lake, Nevada (Benson and White 1994), and has been used by other authors for modelling other sites (e.g., Ricketts and Johnson 1996). It states that

$$R_{evap} = [(R_{lake}/\alpha_{eq})-(RH f_{ad} R_{ad})]/ [((1-RH)/\alpha_{kin})+RH(1-f_{ad})] \qquad (4)$$

where $\delta_i = (R_i - 1)10^3$, R_{ad} is the isotope ratio of the free atmospheric water vapour, f_{ad} is the fraction of atmospheric water vapour in the boundary layer over the lake (if all the atmospheric water overlying the lake is derived from evaporation then $f_{ad} = 0$), RH is the relative humidity, and α_{eq} $(1/\alpha^*)$ and α_{kin} are fractionation factors.

Many of the hydrological components in equations 1 to 4 are dependent on climatic parameters (Table 1) and all these factors therefore control recorded values of δ_l and in turn values of $\delta^{18}O$ recorded from carbonates precipitated in the lake water. It is clear that controls on lake isotope systems are complex, they are also unique for every lake. If robust interpretations are to be made about past climate and environmental change from lacustrine stable isotope records the contemporary system must therefore be fully understood.

Previous climate-isotope proxy comparisons have largely been from open lakes in temperate climates (e.g. von Grafenstein et al. 1996). Controls on Mediterranean lake isotope records have not been previously investigated in this way and there is debate as to the dominant controls on these systems (Roberts and Jones 2002). Here we investigate changes in $\delta^{18}O$ values recorded in a varved sedimentary archive from Nar Gölü, central Turkey, through the last 80 years, over which time annual isotopic change recorded within the lake archive can be compared with instrumental records of climate variability. The lake isotope hydrology is modelled to understand better the lake system and test independently the accuracy of proxy data-climate relationships.

Methods

Nar Gölü (Fig. 1 and described below) was chosen as a suitable site for this study. The varved nature of the sediment record removes uncertainties with proxy data-climate calibrations that would occur if only a radiometric chronology were available. The hydrology of this crater lake is relatively simple and the small catchment has undergone minimal human disturbance which could mask any relationships between the proxy record and climate. Finally, there is no detrital carbonate in the watershed so that the stable isotope record will not have been affected by allogenic input.

Water samples for oxygen and hydrogen isotope analysis were taken in leak proof plastic bottles, initially washed 3 times in the sample, and sealed with plastic insulating tape. Bottles were filled completely to prevent isotopic exchange with air bubbles. The samples were refrigerated at the earliest opportunity, until analysis could be undertaken. Samples were analysed using the equilibration method for oxygen (Epstein & Mayeda 1953), and Zn-reduction method for hydrogen (Coleman et al. 1982; Heaton & Chenery 1990), using a VG SIRA mass spectrometer. Isotopic ratios are defined in relation to the international standard, V-SMOW (Vienna Standard Mean Ocean Water). Analytical precision is typically +/-0.05‰ for δ^{18}O and +/-2.0‰ for δD.

Simple sediment traps, comprising long thin tubes below funnels, were secured at two depths in the water column, at two separate locations in the deepest part of the basin, in July 2001 and collected in August 2002.

A laminated core sequence (NAR01) was obtained from overlapping core samples from Glew (Glew 1991) and Livingstone (Livingstone 1955) samplers (Fig. 1). It was important that the sediments were as undisturbed as possible so sampling could be carried out at an inter-laminae resolution. Glew cores were therefore returned to the laboratory within the core tubes. In some cases drying and shrinking of the sediment occurred, however no stratigraphic integrity was lost. Livingstone core sections were extruded into lengths of half guttering, and covered in non-PVC cling film to prevent the sediments reacting with the plastic. At the earliest opportunity the core sections were refrigerated at 4ºC. The number of laminae were counted independently by two counters in 6cm sections along the core. If the difference between two counts for any given section was greater than 3 laminae both counters recounted the section until agreement was reached.

Dried sediment samples from the top 50cm of the sequence were analysed for ^{210}Pb, ^{226}Ra and ^{137}Cs by direct gamma assay in the Liverpool University Environmental Radioactivity Laboratory, using Ortec HPGe GWL series well-type coaxial low background intrinsic germanium detectors (Appleby et al. 1986).

[210]Pb was determined via its gamma emissions at 46.5keV, and [226]Ra by the 295keV and 352keV γ-rays emitted by its daughter isotope [214]Pb, following 3 weeks storage in sealed containers to allow radioactive equilibration. [137]Cs was measured by its emissions at 662keV. The absolute efficiencies of the detectors were determined using calibrated sources and sediment samples of known activity. Corrections were made for the effect of self-absorption of low energy γ-rays within the sample (Appleby et al. 1992).

For the isotope analysis carbonate samples were disaggregated in sodium hypochlorite solution overnight to remove any organic material. Samples were then sieved at 75μm to remove biogenic carbonates. The <75μm fraction was filtered through micro fibre filter paper (Whatman 41) and washed three times with deionised water to remove any remaining sodium hypochlorite solution and oxidised organic matter before drying at 40ºC. The dried sediment was ground in an agate pestle and mortar to homogenise the sample. Samples containing ~10mg of calcium carbonate were reacted with anhydrous phosphoric acid under vacuum and left at 25ºC overnight (Craig 1957). Stable isotope measurements were then made from the liberated CO_2 on a dual inlet (VG Optima) mass spectrometer. Values are reported using the δ notation against the VPDB standard with analytical errors of ± 0.1 ‰ for $δ^{18}O$.

The study site

Nar Gölü (38°27'30''E; 38°22'30''N; 1363 masl) is a small crater lake located in the Cappadocian region of central Anatolia (Fig. 1). The lake has an area of 556,500 m^2 and a maximum depth of 26 m. Lake volume is 7,692,360 m^3 (July 2001). The crater walls form a small catchment with an area of 2,408,000 m^2 (Fig. 1).

The north, east and west sides of the crater are steep scree slopes with a faulted basalt intrusion visible on the eastern and western sides. The southern side of the crater is less steep, and comprises volcanic ash. At the southern end of the crater is an alluvial fan-delta that extends into the lake. There is no stream activity in this fan system at present. Two springs are found in the southern wall of the crater (Fig. 1) as well as hot springs at the southern edge of the lake.

On the steep north, east, and western slopes there is little vegetation apart from limited deciduous oak (*Quercus cerris* type) which is also found at higher elevations in the south of the catchment. A few small fields around the edge of the lake have been cultivated with a variety of crop plants (including lentils and

chickpeas). Lines of pine tree saplings have recently been planted in the crater on the alluvial fan complex. A fringe of *Phragmites* sp and other emergent macrophytes (including *Scirpus maritimus*) surrounds the lake.

pH and conductivity values indicate that the lake is alkaline and oligosaline, becoming slightly more alkaline and more saline between August 1999 and August 2002 (Table 2). $\delta^{18}O$ values have become more positive at the same time (from -3.0 ‰ to -2.4 ‰) suggesting that evaporation is currently exceeding inflow, which accounts for the increase in salinity (3.1 mS in August 1999 to 4.0 mS in August 2002). The major ion water chemistry in the summer of 1999 was dominated by Na^+ (380 mg/l), Mg^{2+} (103.4 mg/l) and Cl^- (970 mg/l) (Jane Reed, pers. com.).

The lake waters are stratified (Table 2) and the lake bottom waters are anoxic. There is no evidence of benthic life in any of the surface sediment cores taken, and sediments change colour from black to green and yellow on contact with the air due to oxidation.

Climate

The nearest meteorological station to Nar Gölü was at Derinkuyu (25km to the west), although data were only recorded there between 1965 and 1990. From these records the warmest months of the year are July and August (average maximum temperatures 27.6 ºC and 27.8 ºC respectively), with January and February the coldest (average maximum temperatures of 2.9 ºC and 4.9 ºC). Through the 26 years of recording there was greater variability in winter temperatures (1σ = 3.3°C; 3.2°C for January and February respectively), compared with summer values (1σ = 1.0°C, 1.3°C for July and August).

Total annual precipitation averaged 320 ± 68 mm (mean value ± 1 σ) between 1966 and 1990. April and May were the wettest months, accounting for 27 % of the total annual precipitation, with the driest two months, August and September, accounting for only 2 % of the annual total.

Values of evaporation are measured at only a few of the meteorological stations in Turkey, with the nearest station to Nar at Niğde (1208 masl; 37°59' N, 34°40' E), where the average value between 1935 and 1970 was 1547.6 mm yr^{-1} (Meteoroloji Bulteni 1974). Ankara (894 masl; 37°57' N, 32°53' E) had an average value of 1307.6 mm yr^{-1}.

Evaporation can be calculated using the equation of Penman (1948) and simplified by Linacre (1992)

$$E = [0.015 + 4 \times 10^{-4} T_a + 10^{-6}z] \times [480 (T_a + 0.006z)/ (84 - A) - 40 + 2.3 u (T_a - T_d)] \tag{5}$$

E (mm/day), T_a (°C), z = altitude (m), A = latitude, T_d = dew point temperature = $0.52 T_{a\ min} + 0.60 T_{a\ max} - 0.009 (T_{a\ max})^2 - 2$ °C.

Using equation 5 evaporation at Nar Gölü is calculated as 1140 mm yr^{-1} (using annual average values from Derinkuyu for temperature, and Niğde, the nearest station with available data, for wind speed). However calculated values using the same equation for Ankara and Niğde (1209 and 1315 mm yr^{-1} respectively) are below recorded values (Meteoroloji Bulteni 1974). There is a systematic difference between recorded E and calculated E ($E_{recorded} = 1.53 E_{calculated} + 512$; n=11; $r^2 = 0.66$) and using this relationship, evaporation at Nar Gölü is the equivalent of a recorded value of 1025 ± 121 mm yr^{-1}. This latter value results in an Aridity Index of 0.31 (P/E) which is consistent with previously calculated regional values (Türkeş 2003).

Stable isotope hydrology

Lake waters from Nar have more positive δ^{18}O and δD values compared to rainfall and lie on a Local Evaporation Line (LEL) with other lakes in Central Anatolia (Fig. 2). This suggests that evaporation is an important driver of δ_l in these lakes (Leng et al. 1999). Values are more positive in the dry, warm summer months than they are in the wet spring (Table 2) again suggesting that evaporation: precipitation (E: P) ratios are the dominant control on δ_l. Groundwaters from the catchment and the hot springs at the edge of the lake lie on the Ankara meteoric water line showing there is no change in the isotope chemistry of groundwaters following precipitation.

Hydrological budget

Two methods, a water balance model (from equation 1) and an isotope mass balance model (from equation 2), were used to investigate the modern hydrological budget of Nar Gölü. The results, giving values for all components in equations 1 and 2 are shown in Table 3 with the explanation of how these figure are derived given in the appendix. Two sets of results are described as the two different equations (3 and 4) for calculating δ_E give different results.

Direct precipitation only accounts for around 15 to 20 % of the input to the lake system (Table 3). Runoff could be underestimated as k was not experimentally derived, however it would need to increase considerably to have any impact on the amount of groundwater inflow required for equations 1 and 2 to balance. Nar lies at the "fresh" end of the local evaporation line (Fig. 2) suggesting that, although evaporation is still an important control on the hydrological budget, other lakes in the Central Anatolian region are losing more water through evaporation and are likely to have longer residence times. The fact that groundwater lies on the present meteoric water line suggests precipitation is still the biggest influence on input into the lake. The response time of the lake to changes in precipitation may be slower than if the majority of rainfall fell directly into the lake rather than entering via groundwaters.

Lithology and chronology

The sediment core sequence is laminated throughout, with occasional 0.1-5.0 cm thick grey clastic layers, which often show a fining upwards sequence that are thought to be turbidite type events. Laminations are formed in couplets comprising light, often white, carbonate layers and dark layers, made up of organic material and diatoms. The visual appearance of the dark laminae remains more or less constant through the sequence, however there are marked changes in the carbonate laminae. In some cases the carbonate laminae comprise fine grained aragonite crystals (< 5 μm in size) whereas other laminae are made up of a fining upward sequence of calcite polyhedra (10-40 μm in size). Changes in the calcium carbonate mineralogy are probably due to changes in the Mg/Ca ratio within the lake. Kelts and Hsu (1978) describe lake waters with Mg/Ca < 2 where calcite is precipitated, as in the current lake. With increased evaporation lakes become enriched in Mg and the Mg/Ca ratio increases such that aragonite may precipitate.

Laminated sediments composed of alternating carbonate and organic laminae with fining upward calcite crystals have been described elsewhere (e.g., Anderson 1993; Kelts and Hsu 1978). In general the

calcium carbonate layer is precipitated in warm waters during the spring and summer and the dark organic

layer represents the autumn and winter periods (Saarnisto 1986). Contemporary sedimentation recorded in

sediment traps at Nar would support this. The initial sediment in the traps was dominated by organic material

which must have been deposited after July 2001, when the traps were placed in the lake, and the latter

sedimentation was dominated by calcite which must have precipitated prior to August 2002.

Taking values for lake temperatures to be the average of the mean and maximum temperatures for

each month (all lake water temperature measurements lie between these values for all the months measured),

and by estimating monthly lake water values by fitting a sine curve to the lake water values from April and

August 2002, calculations of $\delta^{18}O_{carbonate(PDB)}$ can be made for each month (Fig. 3). Compared to the measured

values of carbonate from 2002 this suggests that carbonate was precipitated between April and June, most

likely during May, or between September and November. The latter scenario is unlikely as the sediment trap

evidence suggests carbonates are precipitated in the lake prior to July. It therefore seems most likely that

carbonate sedimentation at Nar takes place in late spring and early summer. It is possible that this calcium

carbonate deposition is associated with times of increased photosynthesis caused by increased productivity

during times of mixing of the water column (Lamb et al. 2002) although it is not known if the Nar waters

mix at any time during the year.

Contemporary trap data suggests that each lamina couplet represent 1 year of sedimentation (e.g.,

one period of organic deposition and one period of carbonate precipitation occurred between July 2001 and

August 2002). Comparison of the laminae counts with the radiometric chronology (Fig. 4) confirms this. The

[137]Cs activity profile has two well resolved peaks, at 6 cm and 30.5 cm that record fallout from the 1986

Chernobyl reactor accident and the 1963 fallout maximum from the atmospheric testing of nuclear weapons

respectively. The 1963 date is supported by the detection of significant concentrations of [241]Am at that depth

(Appleby et al. 1991). [210]Pb dates calculated using the CRS dating model (Appleby and Oldfield 1978) show

different age-depth relationships compared to the ^{137}Cs peaks, placing 1986 at a depth of ~10.5 cm and 1963

at a depth of ~15.5 cm. The most likely causes of this discrepancy are incomplete recovery of the [210]Pb

record in the older sections of the core, due to very low activities caused by rapid accumulation rates. By

fitting the ^{210}Pb age model to the ^{137}Cs dates, following methods outlined in Appleby (2001), a composite

model has been used to construct a chronology that best fits all of the radiometric data. The two chronologies

(varve and radiometric) are fixed at 1963 and 1986 and they fit very well over most of the sequence except

for a divergence at the very top of the core. The radiometric chronology suggests that the top of the sequence has been undercounted and under sampled with only 5 laminae counted and sampled for a period that covers 1992 to 2001 (11 years). However, it is possible that the loss of laminae in the very top of the core may be due to disturbance or non-recovery of the soft surface sediments during coring.

Isotope results

The carbonate isotope data from the sediment cores show values between 0 and 1 ‰ between 1926 and 1960 (Fig. 5). Between 1960 and 1986 there is a trend to more negative values, declining to -2.8 ‰ in 1986. There is then a sharp negative shift in the isotope values to -4.8 ‰ between 1986 and 1987. This shift is much larger than any other shift in the record and occurs at the same time as a change in carbonate mineralogy from aragonite to calcite. Oxygen isotope values are 0.6 ‰ more positive in aragonite in comparison to calcite precipitated under the same conditions due to different mineral-water fractionation effects but are considered to be constant at all temperatures so an offset can be applied (Kim and O'Neil 1997). A shift in the isotope record of this amount at times of changing mineralogy explains some of the variation in the isotope record. However it is likely that the calcite here is high-Mg calcite which, like aragonite, has more positive $\delta^{18}O$ values compared to low-Mg calcite precipitated under the same conditions by 0.06 ‰ per mole per cent of $MgCO_3$ (Tarutini et al. 1969). The shift here of 2 ‰ is therefore likely to be a minimum shift and cannot be explained solely by a change in mineralogy. Raw isotope data (rather than corrected to calcite) are therefore described throughout this study.

Calibration against meteorological records

Temperature, precipitation, wind speed and relative humidity are known to control lake oxygen isotope values (Table 1). Because there is evidence to show that the $\delta^{18}O_{carbonate}$ record from Nar is annual it can be compared directly to instrumentally recorded values of these climatic parameters. Evaporation can also be calculated from these climate values (e.g., from equation 5) and also compared to the $\delta^{18}O$ record. Because only a short period of data, 1965-1990, is available from the climate station nearest the site (Derinkuyu), climate data from stations with longer records have therefore also been used to compare to the Nar record, as

in most cases trends could be shown to be coherent with the Derinkuyu record. The $\delta^{18}O$ record was

compared to seasonal (three month average) meteorological data using cumulative sum of squared difference

(CSSD) analysis (Taylor 2002). The significance of the relationships (p) are show by the rank of the data in

1000 random perturbations of the isotope data run against the climate data. The gradient of the CSSD line

indicates the strength of this relationship with a gradient of 2 signifying no relationship, a gradient of 0

signifying a strong positive relationship and a gradient of 4 signifying a strong negative relationship.

Traditional Pearson correlation techniques cannot be used here as the $\delta^{18}O$ record is not truly independent

i.e., latter values are dependent on the previous isotopic values of the lake waters. Comparisons were made

with the full record and also with the record between 1926 and 1986 and between 1987 and 2001, to remove

the effect of the large shift in the isotope record between 1986 and 1987 and to check that any relationships

hold through time.

Precipitation

Year-to-year variability in rain and snowfall events in the region is very localised. Comparisons between

Derinkuyu and Nevşehir, 25km to the North, showed relatively weak relationships between rainfall

variations ($r^2 = 0.51$, $p = 0.00$). Only rainfall data from Derinkuyu were therefore used for calibration

against the Nar isotope results, although it is likely that precipitation variability at Nar was, in turn, different

to that at Derinkuyu. From the CSSD analysis no strong or significant trends (significance value $p < 0.05$)

were found between Derinkuyu precipitation variability and the Nar isotope record.

Wind speed

Records of wind speed between 1940 and 2001 from Niğde were used for comparisons with the Nar record

as it was the closest site with available data, although it is possible that wind speed variability was not

coherent between Nar and Niğde. There are no data available for this to be tested. From the CSSD analysis

no strong or significant relationships were found between wind speed and the Nar isotope record.

Temperature

Comparisons of monthly values of temperature between 1965 and 1990 showed variations in temperature to

be coherent between Ankara, the station with the longest records on the Anatolian plateau, and Derinkuyu

(e.g., for average monthly temperatures $r^2 = 0.99$, p = 0.00). The isotope record from Nar was therefore

compared to minimum, average and maximum temperatures from Ankara between 1926 and 2001. From

CSSD analysis no significant relationships were found with minimum temperatures. Significant relationships

were found between the Nar record and average summer temperature variability between 1926 and 2001 (p =

0.01; Fig. 5) and also between 1926 and 1986 (p = 0.01). Significant relationships were also found with

maximum summer temperature variability between 1926 and 2001 (p = 0.04), and maximum annual (p =

0.04; Fig. 5) and summer (p = 0.04) temperature variability between 1926 and 1986. Trends in both the

isotope records and the temperature records are similar between 1987 and 2001, however due to the low

number of data points (n=10) it is difficult to observe whether these relationships are significant or not.

Relative humidity

Comparisons of monthly values between 1965 and 1990 showed trends in relative humidity (RH) to be

coherent between Ankara and Derinkuyu. The isotope record from Nar was therefore compared to RH values

from Ankara between 1926 and 2001. No clear relationships were found between annual variability in RH

and the annual Nar isotope record using CSSD analysis. However there is a strong relationship between

temperature and relative humidity at Derinkuyu (RH = 77.21 − 1.24T_{max} ($r^2 = 0.81$)). It would therefore seem

likely that there is some relationship between RH and $\delta^{18}O$ values if there is such a significant relationship

with temperature.

Evaporation

Values of evaporation were calculated from equation 5 between 1940 and 2001 using wind speed data from

Niğde and values of temperature from Ankara converted to Derinkuyu values based on the relationships

between 1965 and 1990. Strong and significant relationships were found between summer evaporation and

the Nar isotope record between 1940 and 2001 (p = 0.00; Fig. 6), and between 1940 and 1986 (p = 0.02).

Climate comparison summary

From these analyses the only statistically significant relationships (at a 95% confidence limit) between

climate variables and the Nar $\delta^{18}O$ record were found with summer (JJA) maximum and average

temperatures between 1926 and 2001, with summer average and maximum temperatures and also annual and autumn maximum temperatures between 1926 and 1986, and with summer evaporation between 1940 and 2001 and also between 1940 and 1986. Decreasing summer temperatures and evaporation from 1960 to the present day are marked by a trend to more negative isotope values. These are relationships that would be expected as decreasing evaporation leads to less ^{16}O being removed from the lake system and lake waters therefore becoming isotopically more negative. Temperature also has a strong control on evaporation with decreasing temperatures leading to less evaporation; additionally decreasing temperatures will also tend to lead to more negative isotope values in lake waters due to the relationship with δ_P (Leng and Marshall, 2004).

It is clear from the comparison of the isotope and temperature and evaporation records (Fig. 5 and 6) that the lake δ^{18}O record represents a smoothed record of climate variability. Annual variability in the temperature data is not registered in the lake system. This can be explained as annual variability in climate is damped by the ~8 year residence time of the lake.

The sensitivity of the lake isotope system to changes in summer temperature and evaporation can be established from regression relationships between the data sets. These result in the following relationships:

$$\delta^{18}O\ (\text{‰}) = 1.34\ E\ (\text{mm/day}) - 12.42 \quad (r^2 = 0.21;\ p = 0.00) \quad \text{between 1940 and 2001} \quad (6)$$

$$\delta^{18}O\ (\text{‰}) = 0.23\ T_{max}\ (\text{ºC}) - 6.74 \quad (r^2 = 0.03;\ p = 0.10) \quad \text{between 1926 and 1986} \quad (7)$$

The r^2 values for these relationships are very low because the climate data are more variable than the δ^{18}O record. Taking an 8 year running mean of the meteorological data to take into account the lake residence time much clearer relationships can be observed and result in the following relationships:

$$\delta^{18}O\ (\text{‰}) = 3.80\ E\ (\text{mm/day}) - 32.44 \quad (r^2 = 0.79) \quad \text{between 1947 and 2001} \quad (8)$$

$$\delta^{18}O\ (\text{‰}) = 2.08\ T_{max}\ (\text{ºC}) - 60.216 \quad (r^2 = 0.57) \quad \text{between 1933 and 1986} \quad (9)$$

These relationships are statistically less significant due to the smoothing of the meteorological data.

Modelling

Modelling lake response to climatic change can help interpretation of isotopic changes through the sediment record (e.g., Benson and Paillet 2001; Ricketts and Johnson 1996). The simplest form of lake isotope model takes a well mixed lake system in equilibrium and then forces one of the variables, keeping all others constant, to observe the lake system response. A steady state model for Nar can be built from the initial hydrological variables calculated and estimated for the contemporary lake system. The sensitivity of the modelled lake system can then be compared to the relationships found with the climatic variables.

Model assumptions

As the model results in changes in volume the lake area will also change, thus changing the amount of evaporation and rainfall directly from and into the lake respectively. It is assumed that G_o is proportional to the surface area of the lakebed in contact with the lake water, which will also therefore depend on volume. A relationship between volume, lake area and lakebed surface area is therefore required. From the basin bathymetry (Fig. 1) these variables can be calculated for different lake depths. However, there is no simple relationship between these values that can be transferred to the model, especially at values when the lake is deeper than currently observed, and at shallow depths. In the model the lake basin is therefore taken to be an oblate ellipsoid ($a = b > c$), as this is the nearest regular shape to the real lake basin. Changes in lake volume can then be used to calculate changes in lake area and lake bed surface area.

As groundwater inflow is largely meteoric water, δ_{Gi} is taken to equal δ_P. Groundwater inflow to a lake tends to be concentrated at the lake margin (Almendinger 1990) and the amount of inflow is therefore also dependent on the area of the lake. G_i is therefore taken to be a function of lake area and the amount of precipitation. In the model δ_P and RH are both taken to be proportional to temperature as the relationships between these variables and temperature are known to be strong. The amount of evaporation is also dependent on temperature. Equation 4 is used to calculate values of evaporation.

Changes in evaporation

From the climate comparisons evaporation and temperature are the most significant controlling factors

behind stable isotope change in the Nar system. To investigate the role of evaporation a simple model can be

set up where only the amount of evaporation is changed, however δ_E will also change as it is dependent on δ_l.

From equation 2 it can be shown that

$$d\delta_l/dt = [(P\delta_P + S_i\delta_{Si} + G_i\delta_{Gi} - E\delta_E - G_o\delta_{Go}) - (\delta_l dV/dt)]/V \qquad (10)$$

So for a simple model where all parameters remain constant, at the initial values calculated for Nar

in section 3.3, apart from E, δ_l, and δ_E the amount of evaporation can be changed and the response of the lake

water isotope values observed (Fig. 8). From the model curves 3 values, the initial, maximum and

equilibrium values were recorded for different changes in evaporation in the Nar model. The following

relationships result:

for equilibrium values $\qquad \delta^{18}O\ (‰) = 1.02\ E\ (mm/day) - 5.65 \qquad\qquad (11)$

for initial values $\qquad\qquad \delta^{18}O = 0.35\ E - 3.75 \qquad\qquad\qquad\qquad (12)$

for maximum values $\qquad\qquad \delta^{18}O = 1.68\ E - 7.51 \qquad\qquad\qquad\qquad (13)$

Changes in temperature

Changes in temperature lead to much more complex changes in the system as E, δ_p, δ_E, and h are all

dependent on temperature. Again the model does not go straight to equilibrium but first reaches a maximum

value (Fig. 9). From changing the temperature in the model to different values the following relationships

result:

for equilibrium values $\qquad\qquad \delta^{18}O\ (‰) = 0.87\ T\ (^{o}C) - 10.59 \qquad\qquad (14)$

for maximum values $\delta^{18}O = 1.1\ T - 12.62$ (15)

for initial values $\delta^{18}O = 0.17\ T - 4.30$ (16)

Model-calibration comparison

The climate calibration and the models suggest different sensitivities of the lake isotope system to climatic change. The models suggest the lake system is much less sensitive to changes in evaporation and temperature than the climate calibration. However equations 6 and 13 give similar values for changes in evaporation and equations 7 and 16 give similar values for temperature. This might imply that the true relationships are in the order of 0.7 mm/day (summer evaporation) and 5 ˚C (summer temperature) per 1 ‰ change in $\delta^{18}O$. Such a temperature relationship would clearly be incorrect as this would require a shift of approximately 30ºC in summer temperature to account for the observed isotope shift, an order of magnitude too high. Although the value for evaporation change is at the right order of magnitude none of these relationships can fully explain the observed shifts in isotope values (Table 4).

It is therefore likely that neither the calibration nor the modelling relationships is fully correct. This would be expected as the model equations show the number of variables that control a lake isotope system, highlighted in Table 1, suggesting that one climate variable cannot account for the entire shift in the isotope system. The stronger relationships found with evaporation, which includes changes in temperature and wind speed compared to temperature alone would support this. Additionally the increase in relative humidity with decreasing temperature would also lead to decreasing evaporation and possibly increased precipitation at the site even though no significant relationships were found with the Nar isotope record. The model includes changing RH, E, and δ_P with temperature and still does not fully explain the shift in recorded isotope values. In addition the climate calibration relationships are based on summer climate values whereas the models are based on annual shifts and the calibration relationship is based on carbonate isotope values whereas the model relates to lake-water isotope values.

To take into account these differences the model is run between 1926 and 2001 using real temperature data (Fig. 10) and compared to the carbonate values recorded in the sediments. Carbonate values

are calculated from modelled lake water isotope values and summer temperatures for each year. The model does not simulate the amplitude of change recorded in the sediments.

The biggest unknowns in the model equation are the amount of groundwater inflow and outflow, particularly at the beginning of the model run. The models discussed above use lake values from 2001 not from 1940 where lake waters must have been considerable more enriched in ^{18}O than at present. By changing the amount of groundwater flux such that the lake has a residence time of 11 years, compared to 9 years used in the models discussed previously, the model values shift nearer to those recorded between 1940 and 1960 in the sediments. This suggests that there was a change in lake residence time between 1960 and 1986.

The models use constant precipitation amounts as rain and snowfall patterns are very localised in the region, wind speed values are also held constant as they are unknown prior to 1940. If precipitation values are also increased by approximately 25% this leads to changes in the model at the same magnitude as those recorded in the sediments (Fig. 11), suggesting that changes in both P and E are required to force the Nar isotope record. An increase in precipitation may be expected as relative humidity is known to increase with the fall in temperature at Nar between 1960 and 1986 and there are significant relationships between P and both T_{max} and RH at Niğde, Nevşehir and Derinkuyu (Table 5), with decreasing T and increasing RH leading to increases in precipitation of the order needed for the shift described in Fig. 10. In Ankara precipitation rises slowly through this time period, in particular between 1920 and 1970, by a similar amount (35%) to that suggested for Nar by the model.

The model therefore suggest that both evaporation and precipitation changes control the isotope system at Nar although as values of precipitation, relative humidity and temperature are all linked in this region changes in E will automatically lead to changes in P that will enhance isotope shifts. Although not linked to an individual climate parameter, isotope changes in Nar, and therefore probably other lakes in the region, are recording overall changes in the thermo-hydrological climate state.

Conclusions

This study has shown statistically significant relationships between a non-outlet lake δ^{18}O record measured on annual carbonate varves and summer drought intensity derived from historic meteorological records, specifically summer evaporation and temperature. In addition, the lake isotope hydrology has been

successfully modelled using the same climate variables. However, the climate calibration and modelling show differences in the amplitude of relationships between climate and $\delta^{18}O$, and neither can fully explain the observed variance in the isotope record. The calibration overestimates the sensitivity of the $\delta^{18}O$ system to changes in evaporation and the model cannot explain the recorded carbonate $\delta^{18}O$ by using changes in temperature and evaporation alone.

There are a number of possibilities why these relationships do not fully explain the sediment isotope record;

- there may be non-linear lake isotope responses to climate forcing, possibly linked to within-lake dynamics; as illustrated by the large isotopic shift that occurred around 1986.

- the lack of historic rain- and snow-fall data from the lake site itself, and the fact that precipitation is less coherent spatially than temperature across central Anatolia, means it is difficult to interpret the system fully without also employing a modelling approach. The modelling allows understanding of unexplained parameters such as precipitation, which probably increased at Nar by around 25% during the time period of this study, which cannot be explained by a calibration approach alone.

- isotopes respond to several climate parameters that have varied simultaneously and which are themselves linked. This latter point is important for the interpretation of all stable isotope records from the Mediterranean region.

It is therefore important to have a sound understanding of the archive systematics e.g., via modelling, before a calibration relationship can be used to fully interpret proxy records.

The modelling experiments for the contemporary hydrological budget of Nar Gölü used δ_E equations derived from both the Craig-Gordon relationship (e.g., Gibson et al. 1999, equation 3) and from Benson and White (1994, equation 4). Although based on the same theory equation 4 is additionally taken from field based experiments. Both equations have been used in the past to understand better contemporary and palaeolake hydrology. However our modelling results have shown different values for δ_E from equations 3 and 4, and further investigation into these relationships may be required if lake isotope modelling is to be taken further.

Finally, the relationships – based on both empirical calibration and isotope mass-balance modelling – that are developed here can be used to interpret longer records of carbonate $\delta^{18}O$ change from Nar Gölü, to provide high-resolution palaeoclimate reconstructions for the pre-instrumental period.

Acknowledgements

Appendix

Water balance model

For a given period of time it can be assumed that the lake volume is constant so that

$$P + S_i + Gi = E + S_o + G_o. \tag{a1}$$

P, E and S_o are known (Table 3).

S_i: runoff from the catchment is the only factor to take into account as there are no rivers or permanent streams that currently enter the lake. A runoff coefficient (k) of 0.25 is assumed. Although k is unknown, if it is assumed that $G_i > G_o$ (see further discussion below), then k < 0.4, from the extreme values of P and E. A value of 0.25 has been used elsewhere for crater lakes in evaporatively dominated regions (Telford and Lamb 1999).

Using these values ($G_i - G_o$) can be calculated. G_i and G_o are both unknown and difficult to estimate. G_o is assumed to be low, relative to E as the lake has been shown to be isotopically enriched and saline.

Isotope mass balance model

At Nar there is no surface outflow and lake waters leaving the lake through ground water are assumed to be the same value as all other waters within the lake such that $\delta_{Go} = \delta_l$. As the only surface input into the lake is from runoff directly from precipitation $\delta_{Si} = \delta_P$. As in equation 3 it is assumed that the lake is in a steady state such that $dV\delta_l/dt = 0$. This relationship also assumes that the lake is well mixed although in reality it is stratified for at least part of the year. Taking these assumptions into account

$$P\delta_P + S_i\delta_P + G_i\delta_{Gi} = E\delta_E + G_o\delta_l \qquad (a2)$$

P, S_i, E and δ_l are known (Tables 2 and 3).

δ_{Gi}: spring waters from both the catchment and the hot springs in the lake lie at approximately $\delta^{18}O = -10.6 \pm 0.1$, $\delta D = -74.4 \pm 0.7$ (Table 2).

δ_p: in a given year the average isotope ratio of rainwaters entering the lake will be a weighted average of the value for each month. The weighted average of annual rainfall at Ankara is ($\delta^{18}O = -8.8$, $\delta D = -58.6$) from 1996, 1997, 2000 and 2001 monthly values (IAEA/WMO 2002; A. Dirican, pers. com. 2002). It is unlikely that isotope values of precipitation are the same at Nar. Various relationships are known for differences in precipitation isotope values between locations (e.g., Dansgaard 1964; Rozanski et al. 1993) and these relationships can be used to estimate values for Nar Gölü based on those recorded in Ankara. Temperatures in Derinkuyu are consistently colder than those in Ankara (Derinkuyu $T_{av} = 1.01$ Ankara $T_{av} - 2.55$; $r^2 = 0.99$). Using the relationship between precipitation isotope values and temperature from Ankara ($\delta^{18}O_p = 0.31 T_{max} - 13.18$; $r^2 = 0.64$) $\delta^{18}O$ values would be 0.89‰ lower at Derinkuyu and δD values 6.38‰ lower. There is also a standard relationship for changes in isotope values due to height, a large part of which is due

to the corresponding change in temperature (Gat et al. 2001). Nar lake lies 460 m above the recording station in Ankara and this corresponds to a change of -0.92 ‰ in oxygen isotope values (-0.2‰/100m) and -6.9‰ in δD (-1.5‰/100m). Additionally Nar lies further along rain tracks than Ankara and δ_p values are therefore likely to be more negative due to the rain out effect although this is difficult to quantify with the data currently available. δ_P has also been estimated from the intercept of local evaporation lines (LEL) with meteoric water lines (MWL), i.e., the value of the waters from which lakes in a given region evaporate (e.g., Ricketts and Johnson 1996). The Nar evaporation line intercepts the Ankara meteoric water line at $\delta^{18}O = -11.0$ ‰. Groundwater values, which may also represent the weighted average value of mean annual rainfall, lie close to the LEL / MWL intercept at -10.6 ‰ (Fig. 3). A value of -10.6 ‰ is therefore used for the precipitation at Nar.

δ_E: from equation 3

α^*: the equilibrium fractionation factor can be calculated

$$1/\alpha^* = \alpha_{eq} = \exp(1137\ T^{-2} - 0.4156\ T^{-1} - 2.0667 \times 10^{-3}) \tag{a3}$$

where T is the temperature of the lake surface water in degrees Kelvin (Majoube 1971).

T: lake surface temperatures measured over a number of field seasons lay between the mean and maximum monthly temperatures from Derinkuyu. Lake surface temperatures will therefore be taken to be the average of the mean and maximum air temperatures. This gives a value of 12.6°C for the mean annual lake surface temperature.

h: for a relative humidity of 0.57, the average at Derinkuyu through the recorded time frame, at a mean air temperature of 9.2°C, the normalised value of h at 12.6°C is 0.45.

ϵ_k: for $\delta^{18}O$ has been shown to approximate 14.2(1-h) ‰ (Gonfiantini 1986).

δ_A: the isotope value of atmospheric moisture is difficult to quantify unless measured at the site, although is usually assumed to be proportional to δ_P. The relationship between isotope values of atmospheric moisture and precipitation recorded at Ankara (A. Dirican pers. com., 2002) gives a value of -21 ‰ for $\delta_{atmosphere}$ over Nar Gölü. δ_A has often been calculated as δ_P- ϵ^*, and been shown experimentally to be a correct approximation to with in 5% (Gibson et al. 1999). For the Nar data this method gives a δ_a value of -20.2‰, close to that calculated from the moisture precipitation relationship in Ankara. A value of -21 ‰ will be used here.

For equation 4:

R_{lake}, α_{eq}, RH and R_{ad} are known (Tables 2 and 3).

f_{ad}: Benson and White (1994) suggest that f_{ad} should always be taken as 0 as the water vapour immediately overlying the liquid surface will be the dominant control on δ_E and will be almost entirely made up of evaporated water. This is likely to be the case at Nar as evaporation id known to be one of the largest fluxes in the system and the crater provides some shelter winds which would replace the air overlying the lake more often. A value of 0 for f_{ad} is therefore used here.

α_{kin}: the kinetic fractionation factor, is dependent on wind speed and for wind speeds less than 6.8 ms^{-1}, $\alpha_{kin} = 0.994$. This value is appropriate for Nar as average wind speeds in the region are 3.2 ms^{-1}.

Simultaneously solving equations 1 and 2 using the known figures above therefore leads to values for G_i and G_o (Table 3).

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Table 1 Controls on parameters determining δ_l.

Parameter	Controls
V	see equation 1
P	p, A_l
S_i	p, A_c, k
G_i	A_l, p
E	T_a, T_l, u, RH, I
S_o	V
G_o	V
δ_P	T_a, source area etc.
δ_{Si}	δ_P, E_c
δ_{Gi}	δ_P, E_c, t
δ_E	See equation 3
$\alpha^{\cdot}$	T_l
h	RH, T_l, T_a
ϵ_k	RH
δ_A	δ_P
f_{ad}	u, E, RH
α_{kin}	u

p = precipitation; A_l = lake area; A_c = catchment area; k = runoff coefficient; T_a = air temperature; T_l = lake surface temperature; u = wind speed; I = insolation; E_c = catchment evaporation. All other symbols are defined in the text.

Table 2 Lake water data from Nar Gölü.

Sample Location	Depth (m)	Temperature (°C)	Conductivity (mS)	pH	$\delta^{18}O$ (‰)	δD (‰)
Lake Waters (08/1999)						
Lake Edge		22.0	3.1	7.1		
Lake Centre		21.0	2.5	7.4	-3.0	-37.3
Lake Waters (07/2000)						
Lake Edge		26.8	2.9	6.4	-3.2	-36.8
Lake Waters (07/2001)						
Lake Centre	0.0	24.2	3.3	7.9	-2.6	-34.6
	4.0	24.1	3.4	7.2		
	9.0	22.0	3.4	7.2	-2.8	-36.1
	12.0	20.7	3.2	6.6		
	14.0	17.2	3.2	6.2	-2.9	-35.8
	19.0	16.7	3.1	6.6		
	24.0	15.7	3.0	6.5	-3.2	-38.9
Lake Waters (04/2002)						
Lake Edge		8.9			-3.1	-37.3
Lake Waters (08/2002)						
Lake Centre	0.0	23.2	4.0	8.0	-2.4	-33.3
	5.0	22.8	4.2	7.9	-2.4	-33.1
	7.5	20.3	4.2	7.6	-2.8	-35.5
	10.0	14.3	4.0	7.1	-3.3	-39.0
	12.5	10.7	4.0	7.1	-3.3	-39.3
	15.0	9.9	4.0	6.9	-3.3	-39.0
	20.0	10.1	4.0	6.8	-3.3	-38.2
Hot Springs (2000)						
					-10.6	-73.7
Springs (07/2001)						
Spring 1		15.7	0.1	6.0	-9.3	-64.7
Spring 2		12.6	0.2	7.1	-10.5	-73.7
Springs (04/2002)						
Spring 2					-10.6	-75.0
Springs (08/2002)						
Spring 2					-10.7	-75.0

Table 3 Summary of Nar Gölü hydrology. Lake residence time is calculated by dividing lake volume by the flux through the system.

	Value (m^3 yr^{-1})		Percentage	
	$\delta_E = -20.58$	$\delta_E = -15.68$	$\delta_E = -20.58$	$\delta_E = -15.68$
Inputs				
Precipitation	178,080		14	19
Runoff	148,120		11	16
Groundwater	974, 048	615,712	75	65
Outputs				
Evaporation	570,412		44	60
Groundwater	729, 836	371,499	56	40
Total Flux	1,300,248	941,922		
Residence Time	5.92 yrs	8.17 yrs		

Lake Area	556,500 m^2	**Lake Volume**	7,692,360 m^3
Catchment Area	2,408,000 m^2	**Precipitation** (p)	0.320 myr^{-1}
Evaporation (E)	1.025 myr^{-1}	$\delta_P, \delta_{Si}, \delta_{Gi}$	- 10.6 ‰
Run off coefficient (k)	0.25	δ_l, δ_{Go}	-2.8 ‰
Surface runoff (S$_o$)	0		

Table 4 Predicted change in $\delta^{18}O$ between 1960 and 1986 from climate-isotope relationships found from climate calibrations (equations 6 to 9) and models (equations 11 to 13 and 14 to 16) compared to the recorded shift of -3.6 ‰. Predicted changes are based on instrumented shifts in evaporation and summer maximum temperature over this time.

	Equation	Predicted change in $\delta^{18}O$	Percentage of recorded shift
Evaporation			
Calibration	6	-0.76	21.13
	8	-1.72	47.67
Model	11	-0.39	10.73
	12	-0.13	3.58
	13	-0.63	17.55
Temperature			
Calibration	7	-0.05	1.28
	9	-0.41	11.56
Model	14	-0.17	4.83
	15	-0.22	6.11
	16	-0.03	0.94

Table 5 Percentage of precipitation variability described by temperature and relative humidity at three meteorological stations close to Nar Gölü.

P v.	Derinkuyu		Niğde		Nevşehir	
	r^2	p	r^2	p	r^2	p
T_{min}	0.10	0.00	0.14	0.00	0.16	0.00
T_{av}	0.09	0.00	0.17	0.00	0.20	0.00
T_{max}	0.12	0.00	0.19	0.00	0.21	0.00
RH	0.16	0.00	0.28	0.00	0.29	0.00

Figures
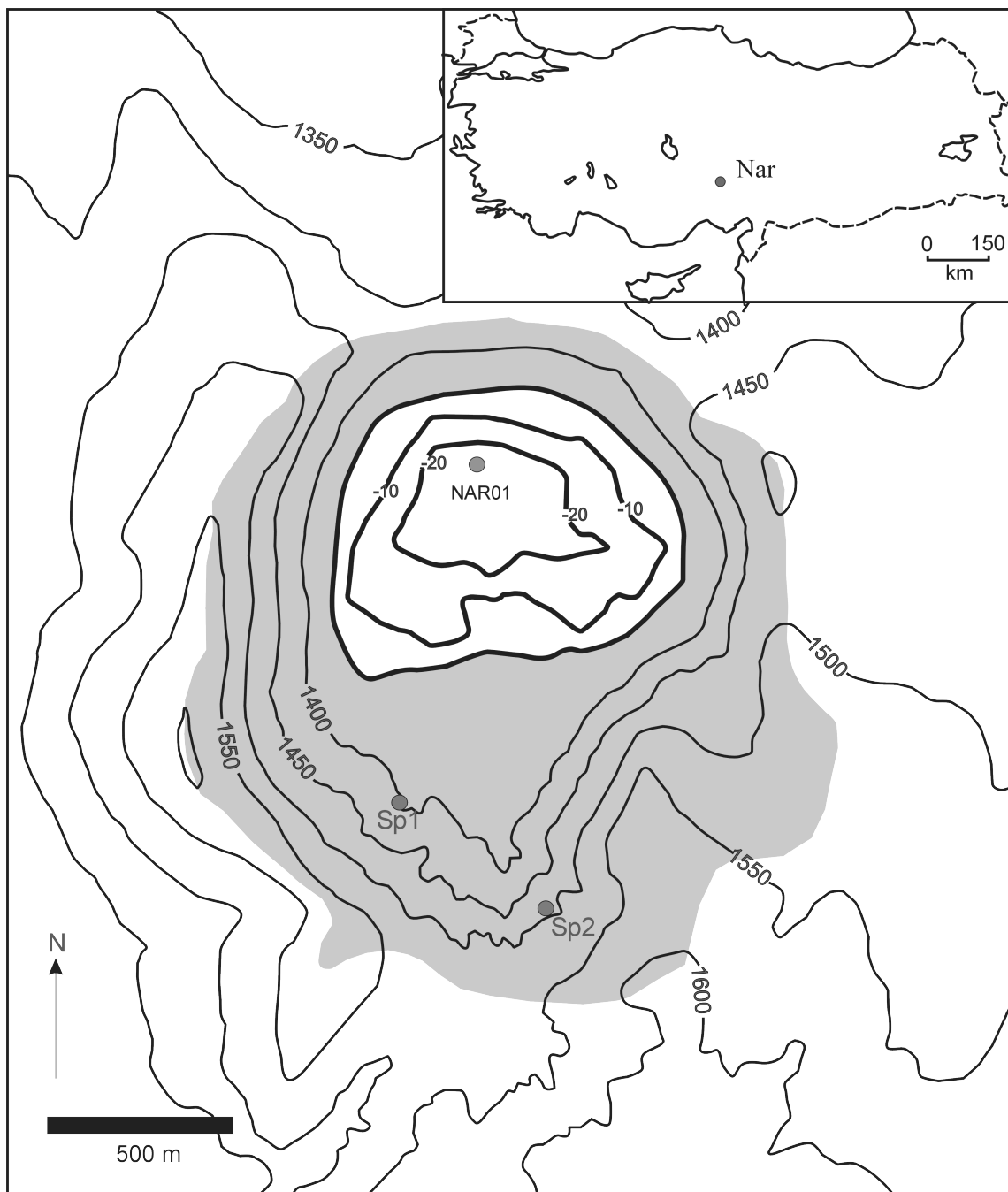
Fig.1 Location of Nar Gölü and catchment morphology. Location of core samples and springs (Sp) are shown and the lake catchment is shaded. Contours in masl, lake depths in m.
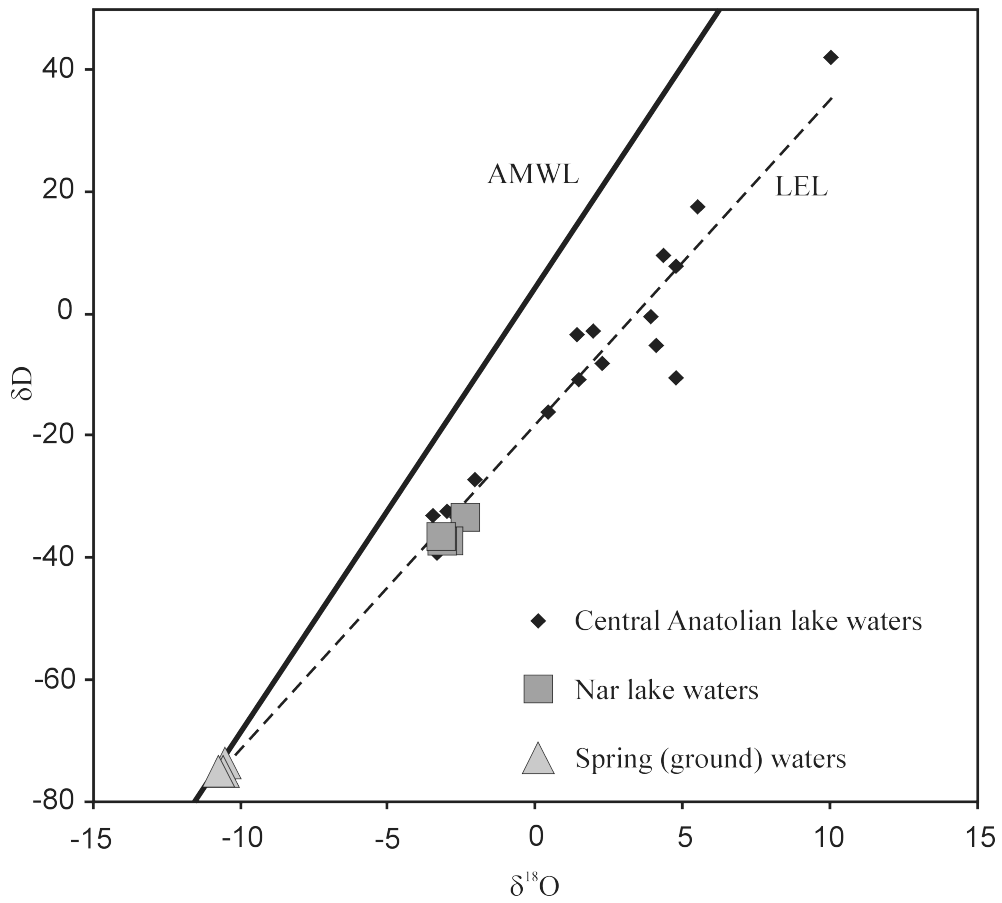
Fig. 2 Nar isotope hydrology in the context of other lakes in the region (Continental Central Anatolia) as defined on the basis on rainfall patterns by Türkeş (1996, 1998). Solid line is the Ankara meteoric water line (AMWL) and dashed line the central Anatolian local evaporation line (LEL). Data from Leng et al. (1999) and unpublished.
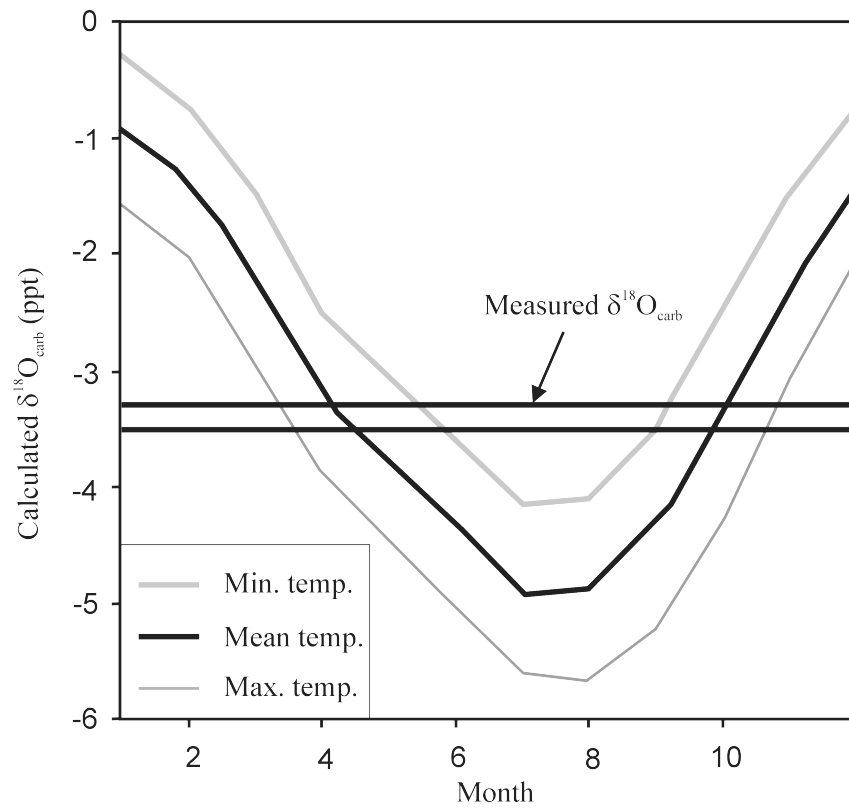
Fig. 3 Theoretical values of calcite $\delta^{18}O$ from Nar Gölü calculated from mean, minimum and maximum air temperatures. Horizontal lines show values of surface sediments and carbonates from sediment traps collected in 2002.
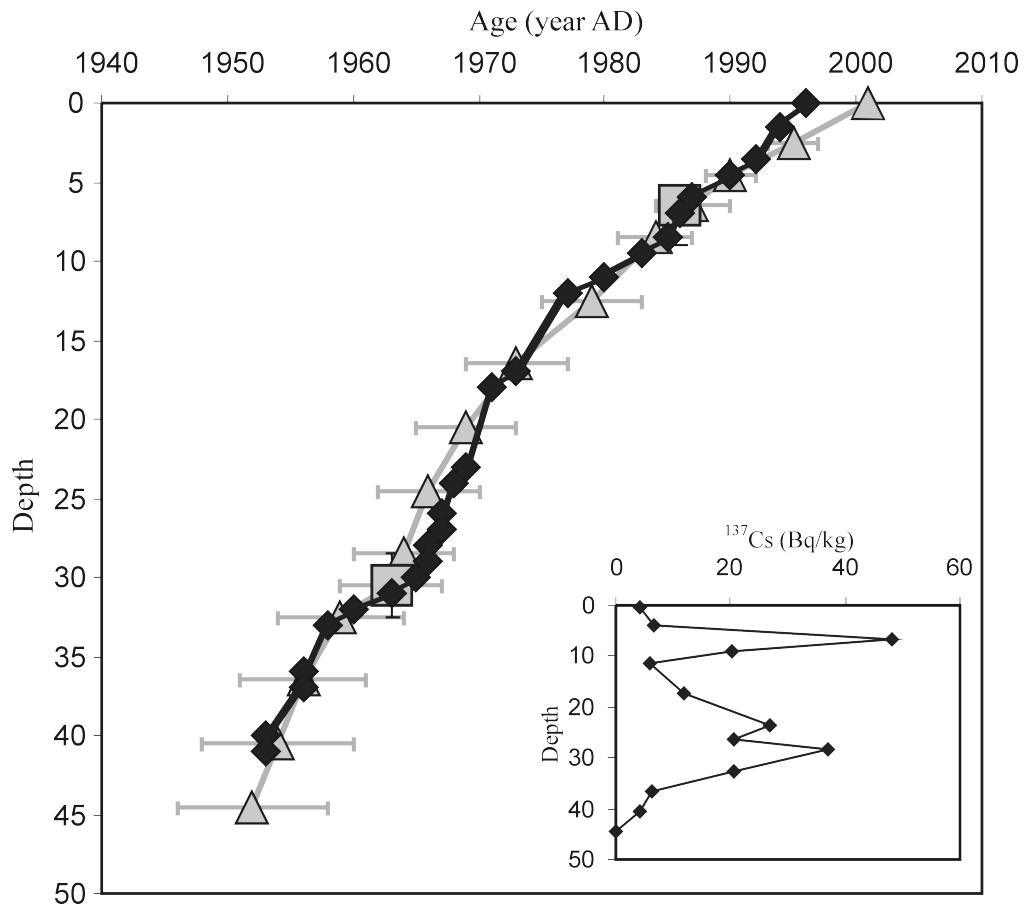
Fig. 4 Comparison of radionuclide (grey triangles ^{210}Pb dates, squares ^{137}Cs peaks) and varve chronology (black diamonds) from the top 50cm of NAR01. Inset shows ^{137}Cs counts.
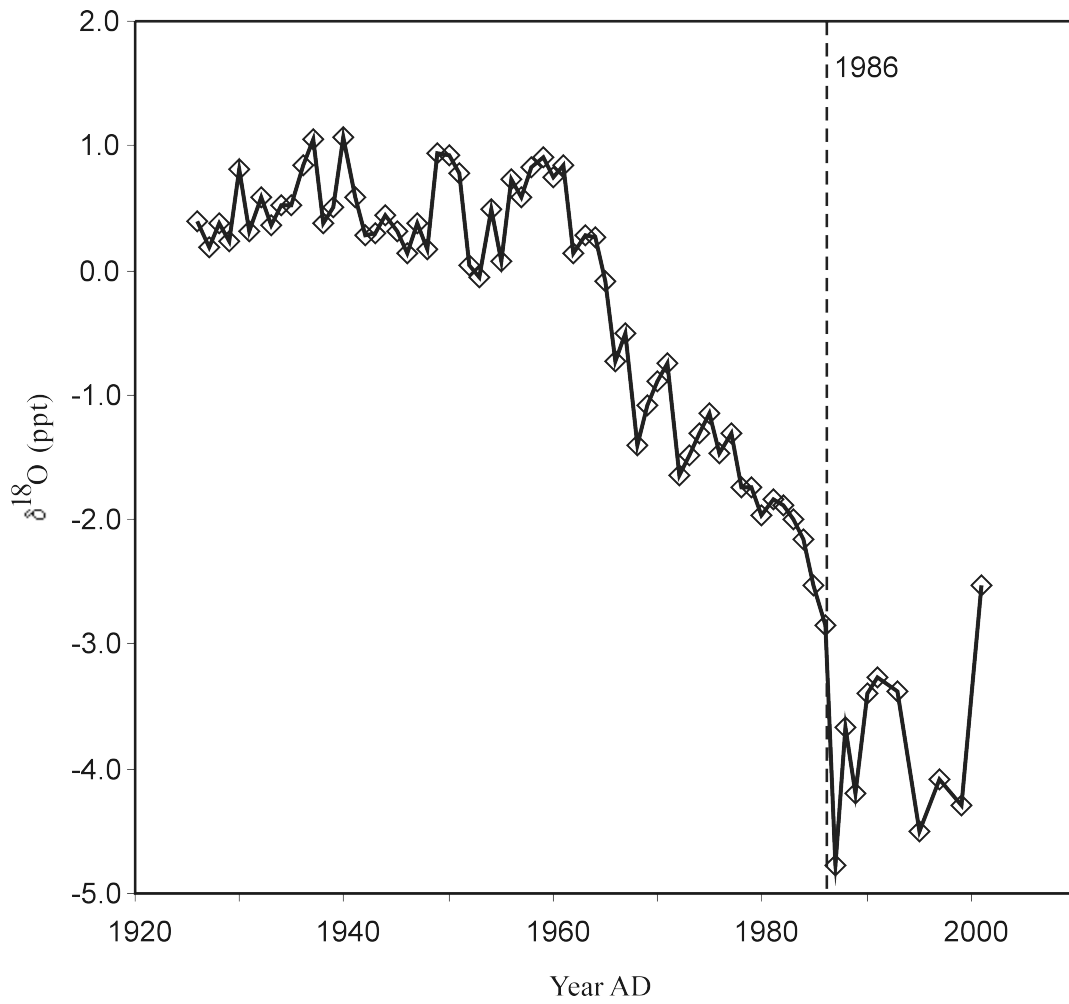
Fig. 5 Raw carbonate $\delta^{18}O$ results from the last 80 years from Nar Gölü showing trend to more negative isotope values between 1960 and 1987. Mineralogy changes in 1986 from aragonite (pre. 1986) to calcite. Data is available at ………….
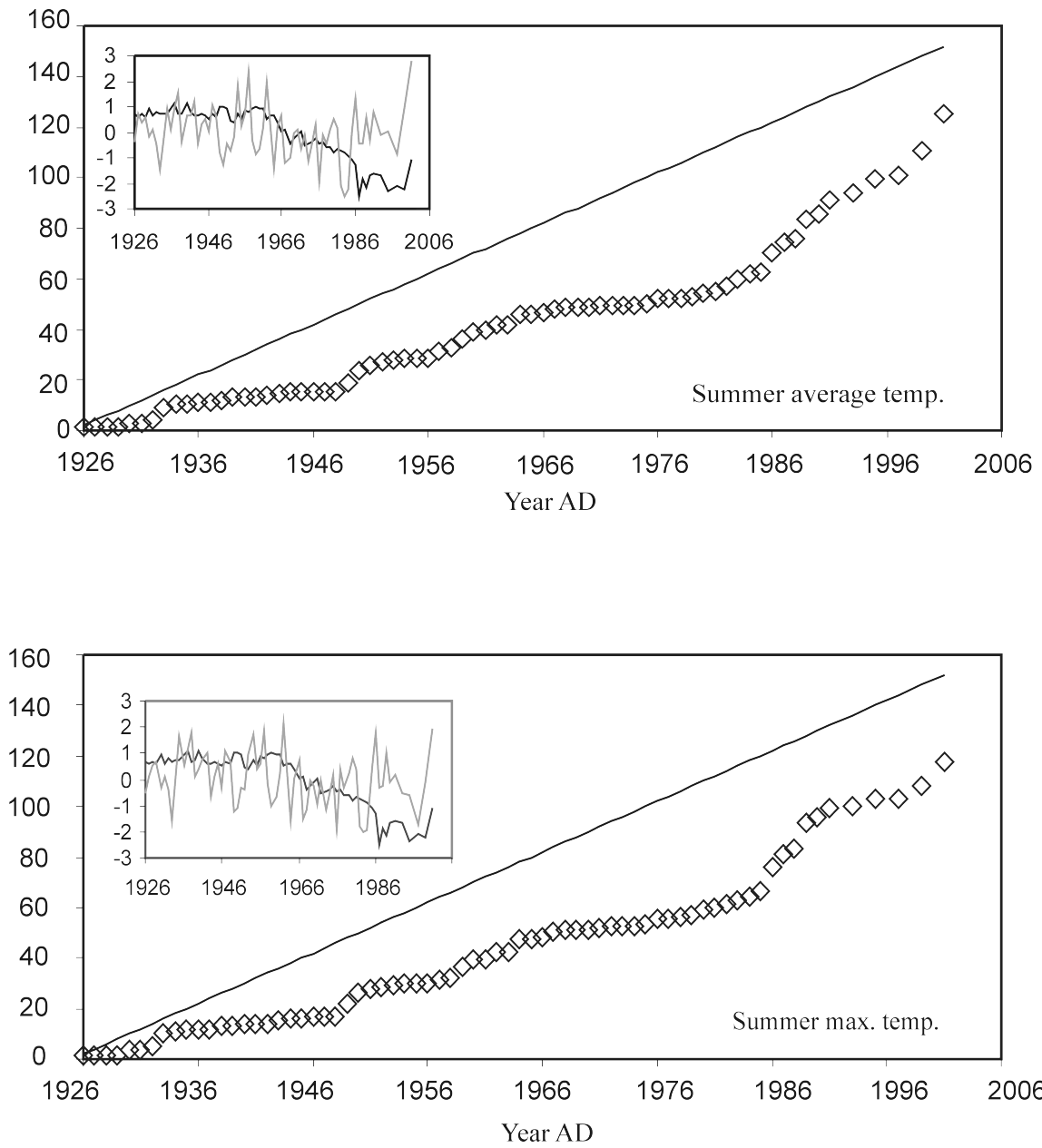
Fig. 6 CSSD plots showing correlation between $\delta^{18}O$ and temperature. The open symbols show the cumulative sum of the squared difference between summer mean and maximum temperatures compared to the Nar $\delta^{18}O$ record. The thin line has a gradient of 2 to show the relative strengths of the relationships. Insets show the normalised temperature (grey line) and isotope data (black line).
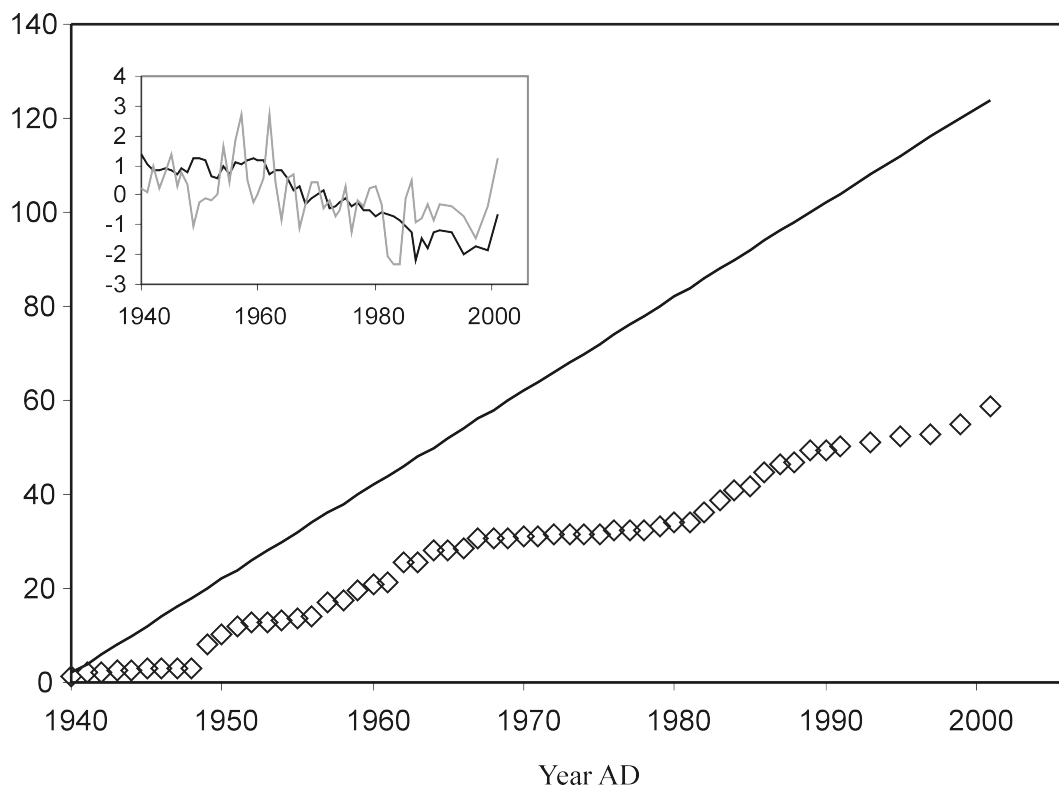
Fig. 7 CSSD plot showing correlation between δ^{18}O and evaporation. Inset grey values show normalised calculated values of evaporation.
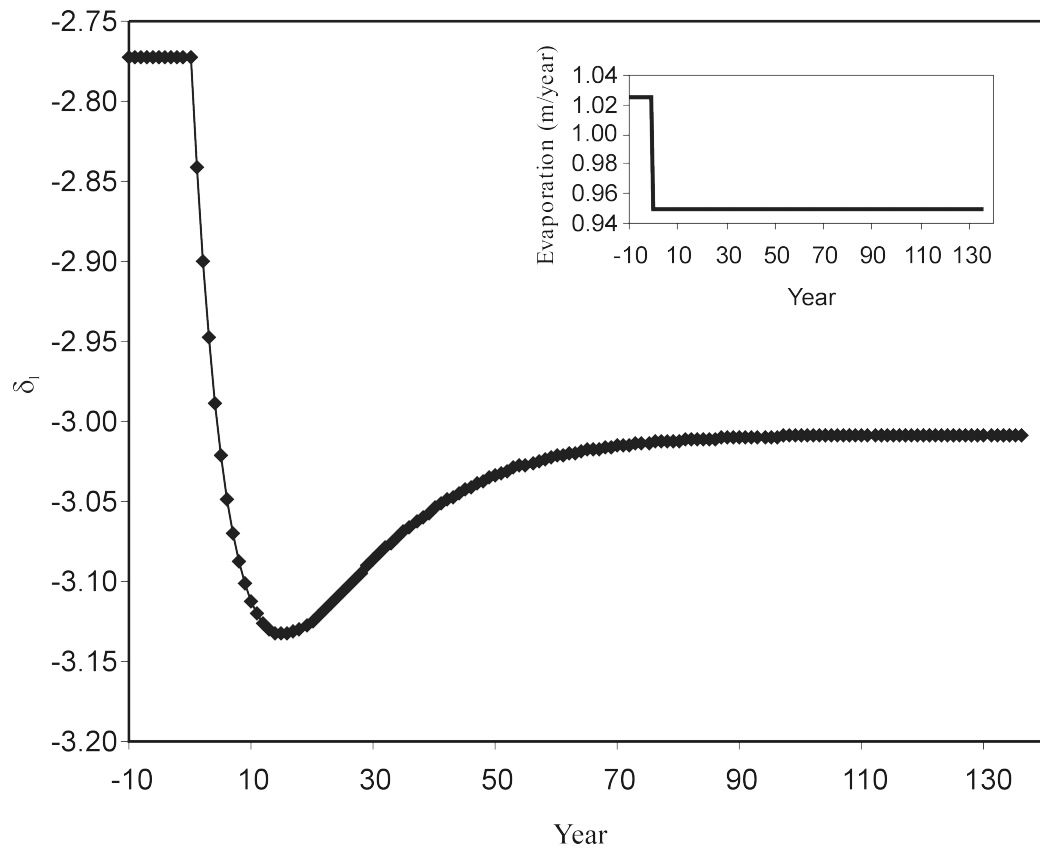
Fig. 8 Change in model lake water $\delta^{18}O$ with a change in evaporation from 1.025 myr^{-1} to 0.95 myr^{-1}.
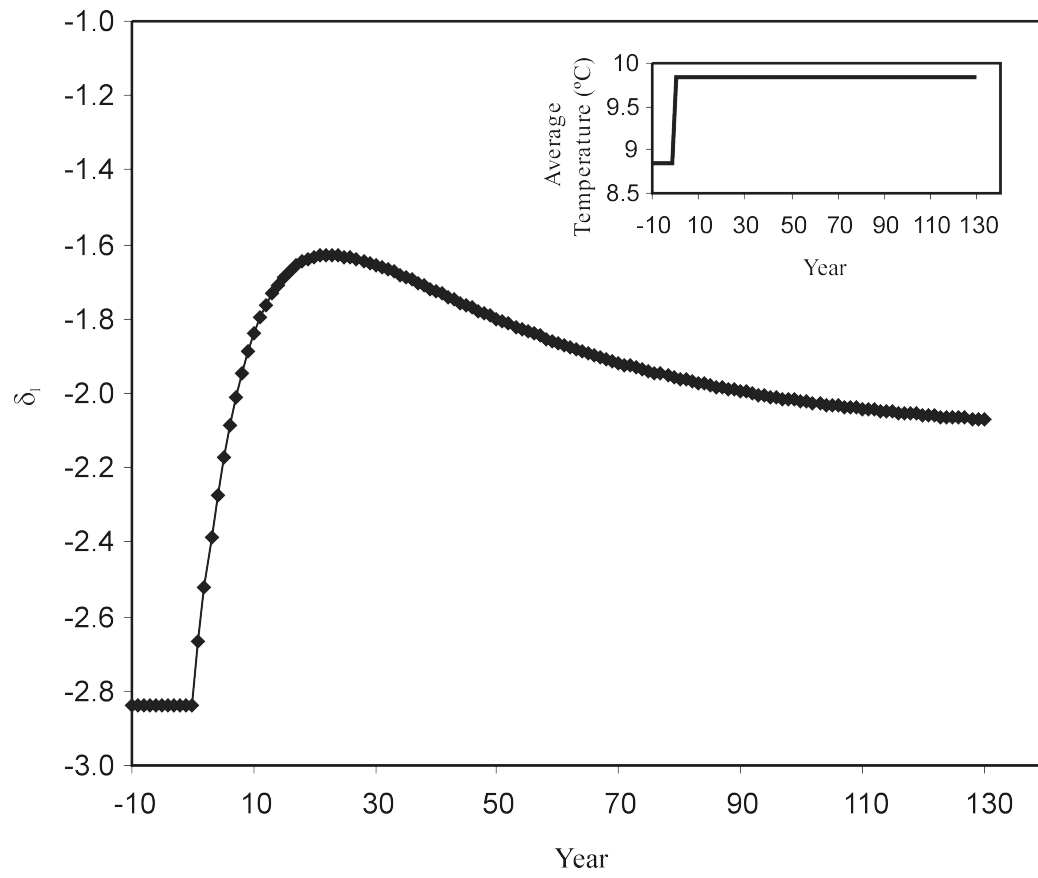
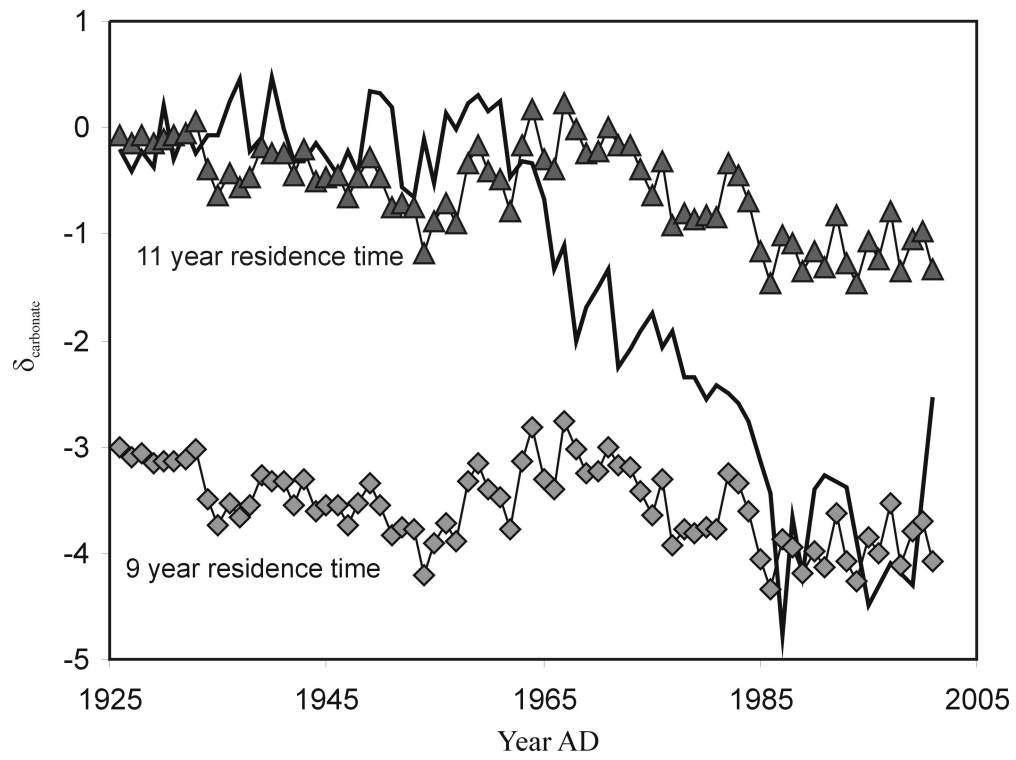**Fig. 9** Change in model lake water δ¹⁸O with a change in mean annual temperature from 8.85 ºC to 9.85 ºC.

Fig. 10 Modelled (grey symbols) and recorded (black line) carbonate isotope values between 1926 and 2001

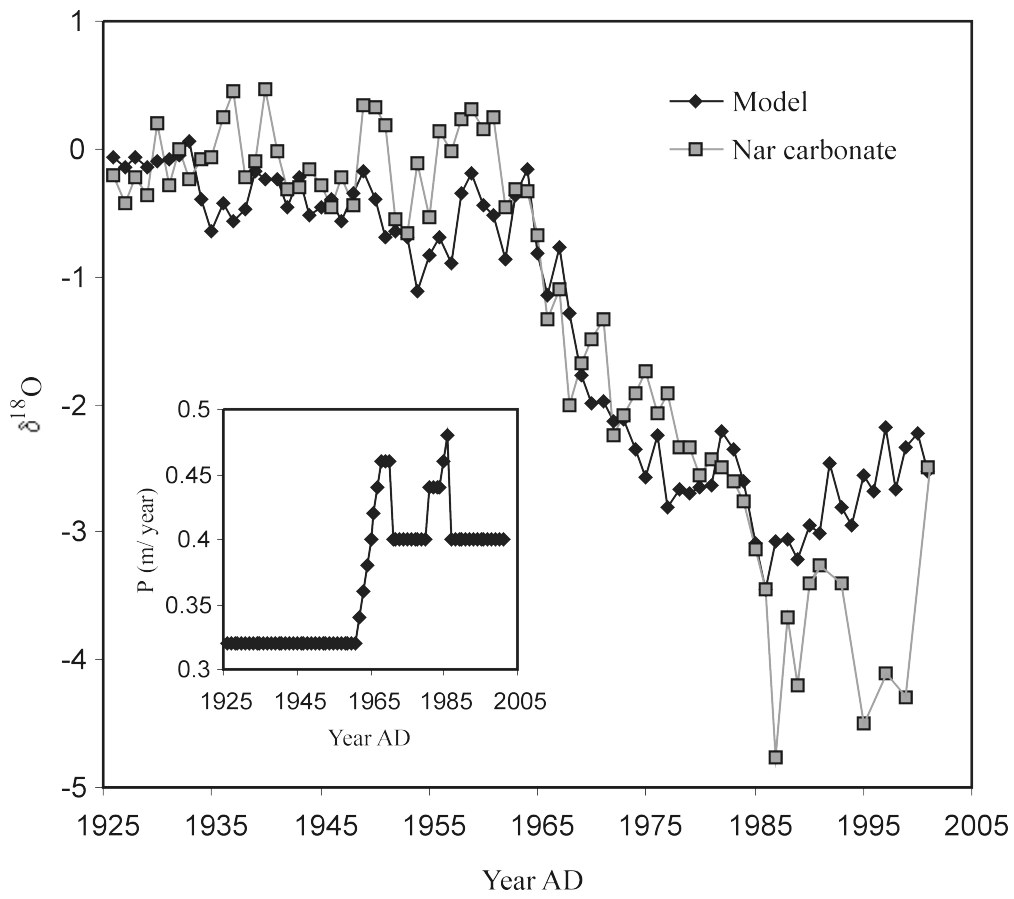with constant precipitation. Models were run with varying initial residence times.

Fig. 11 Model output from the 11 year residence time model with required change in precipitation (inset) to force observed change in Nar Gölü carbonate δ^{18}O.