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Paleoclimate and vegetation of the last glacial cycles in Jerusalem from a speleothem record

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Abstract. A speleothem isotopic record taken from Jerusalem is used to reconstruct regional climate over the last 170,000 years. Glacial periods in Jerusalem were generally cooler and wetter than the present climate. Stage 5e in the desert margin of Jerusalem was extremely unstable, dry, and warm, and instability persisted throughout the transition to glacial conditions. The climate after stage 5e became gradually cooler and wetter over a 20,000-year interval and did not recover to interglacial conditions in stage 5c and 5a. The $\delta^{13}C$ varied by up to 12‰, from glacial (stage 6, 4, 3, 2) values of -10 to -12‰ that reflect dense C_3 vegetation above the studied cave, and up to 0‰ in early stage 5 when there was probably complete loss of vegetation. The climatic instability during interglacial periods is much larger than during glacial periods, and glacial/interglacial transitions do not behave the same in each climatic cycle in this region.

Keywords: Speleothem; Karst record; stable isotopes; east Mediterranean paleoclimate; climatic instability; Late Quaternary climate.

Index terms: Isotopic composition/chemistry; biogeochemical processes; hydroclimatology; unsaturated zone.

1. Introduction

A satisfactory paleoclimatic reconstruction over a complete glacial cycle in the climatic transition zone of Israel (Figure 1) has not yet been achieved. The principal problems have been the lack of continuous records with independent time calibration and the difficulty of interpreting the available proxies in terms of the basic climatic parameters, temperature and precipitation. Evidence pertaining to the last interglacial period (isotope stage 5e) in the region is especially scarce, as radiometric dating of continuous paleoclimatic proxies beyond the radiocarbon limit is hardly available. Here we discuss the paleoclimatic implications of the carbon isotopic record of the stalagmite AF12 (Figure 2) which grew in Jerusalem West Cave, in Jerusalem (Figure 1).

Calcite cave deposits (speleothems) provide a datable isotopic record, rare in other terrestrial environments. The feedwater residence times of speleothems precipitated in the vadose zone are brief, and thus the isotopic composition of calcite responds quickly to local environmental changes. In Jerusalem, there are some small, sealed caves that retain high humidity, providing ideal conditions for long-term, continuous speleothem deposition in isotopic equilibrium. Stalagmite AF12 from Jerusalem West Cave was most suitable for paleoclimatic inferences [*Frumkin et al.*, 1994]

AF12 grew in a small cave located at 730 m mean sea level (msl), in the Upper Cretaceous dolomitic Amminadav Formation. Today the region experiences a Mediterranean climate according to the Köppen classification, with rainy winters and dry summers. Mean annual precipitation is 700 mm, mean temperature is 17°C, and potential annual evaporation is $\sim$1600 mm. Thin patches of terra rossa soil above the cave sustain remnants of the local Mediterranean vegetation, all of which use the C_3 photosynthetic pathway. A few kilometers to the east is the semi-arid zone (Figure 1), where the vegetation is comprised of both C_3 and C_4 plants [*Shomer-Ilan et al.*, 1981; *Vogel et al.*, 1986]. The border between the Mediterranean and semi-arid climatic zones in Israel has shifted during the last 70 years [*Potchter and Saaroni*, 1998], as well as during earlier parts of the Holocene [*Goodfriend*, 1990]. The close proximity of the study site to the region where C_4 vegetation is abundant and the climatic sensitivity of C_4 plants allow us to use their existence as an indication of previous warmer and drier climate in places which presently sustain only C_3 plants. The large difference of $\delta^{13}C$ between C_3 and C_4 plants serves as a proxy to detect changes in vegetation type and quantity [*Lloyd*, 1980]. However, as will be shown, climate has influenced $\delta^{13}C$ of speleothem in other ways as well.

2. Methods

AF12 consists of dense laminated calcite, confirmed by X-ray diffraction and electron probe microanalyzer. Visual inspection was used to identify portions of large crystals and lack of growth laminations. Such portions along the growth axis were avoided because of suspected recrystallization. The dating and the oxygen isotopic record were discussed in an earlier paper [*Frumkin et al.*, 1999b]. The $\delta^{13}C$ and $\delta^{13}O_c$ (carbon and oxygen isotopic composition of calcite), reported in ‰ versus PDB (VPDB) ±0.08‰, were measured for 395 points along the growth axis, 100 of them with

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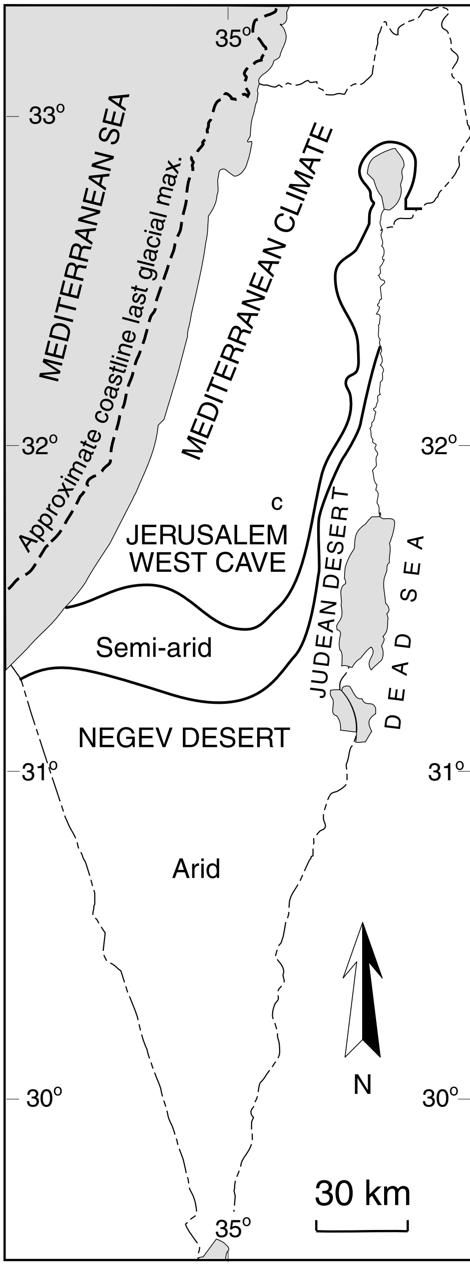

Figure 1. A map of Israel showing the studied cave (C) and main climatic regions according to Köppen, using meteorological mean data for 1961–1990 [*Potchter and Saaroni*, 1998].

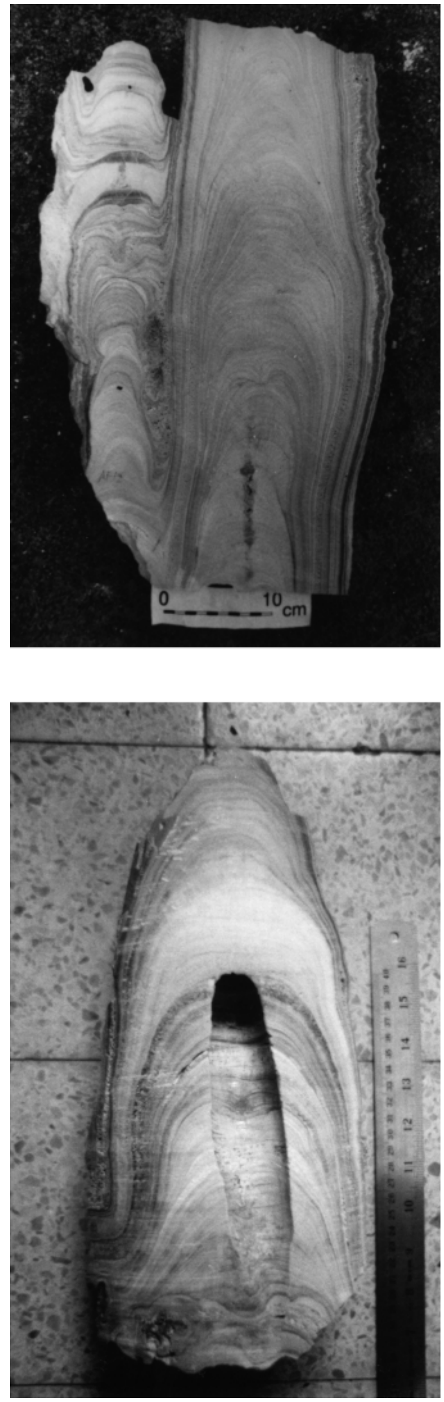

Figure 2. Longitudinal section of the studied speleothem AF12. The stalagmite was taken out of the cave in several parts, and a core drill was used to extract the lower part from the floor of the cave. The lower half of the lower part (age 220,000–170,000 years) was not used in this study. The uppermost left part was deposited up to the present time. Locations of samples for dating are given by *Frumkin et al.* [1999b].

duplicates, using a VG SIRA mass spectrometer with an automated carbonate analyzer. Chronology of AF12 is based on $^{230}Th/^{234}U$ thermal ionization mass spectrometry (TIMS) using a VG 354 mass spectrometer. The TIMS U-series dating indicated that AF12 has grown from 170,000 B.P. to the present, with one hiatus [*Frumkin et al.*, 1999b]. The deposition of this stalagmite through almost two glacial cycles up to the present allows us to detect some secular climatic variations and to calibrate them against present conditions.

3. Carbon Speleothem Record

The $\delta^{13}C$ values (VPDB) in AF12 range between -12 and 0‰ (Figure 3].), ~1 order of magnitude larger than the glacial/interglacial variations reported in $\delta^{13}C$ of atmospheric CO_2 [*Leuenberger et al.*, 1992; *Marino et al.*, 1992]. Values from −12 to −9‰ are most common (Figure 4) during glacial periods and early Holocene. About half of the samples are even more constrained, falling between −11 and −10‰. Isotopic stage 6 was relatively stable with $\delta^{13}C$ values of −10 to −11‰. At its close around 135,000 years, $\delta^{18}C$ values decreased to −12‰ followed by an extremely sharp rise to 0‰ to mark the beginning of stage 5e. Note that $\delta^{18}O$ of the calcite records a sharp depletion at the same time. A gradual decrease of $\delta^{18}C$ values with large fluctuations followed, reaching −11‰ at stage 5d. The $\delta^{13}C$ values remained between −10 and −12‰ until stage 3. The $\delta^{13}C$ values were approximately −9‰ during stage 2, with a sharp rise up to −5‰ at the onset of deglaciation. During later deglaciation, values decreased again to about −10‰ at the early Holocene. The $\delta^{13}C$ values increased at the late Holocene up to the its present value of −7‰.

4. Discussion

The dissolution of carbonate rock (such as limestone) by water and CO^2 is generally expressed by

$$CaCO_3 + CO_2 + H_2O \ \emptyset \ 2HCO_3{}^- + Ca^{2+}$$

and calcite speleothem deposition is given by the same equation running from right to left. This implies that under "normal" conditions, half of the carbon in speleothems will be derived from bedrock and the other half will be derived mainly from the vegetation above the cave which supplies most of the CO^2 in the

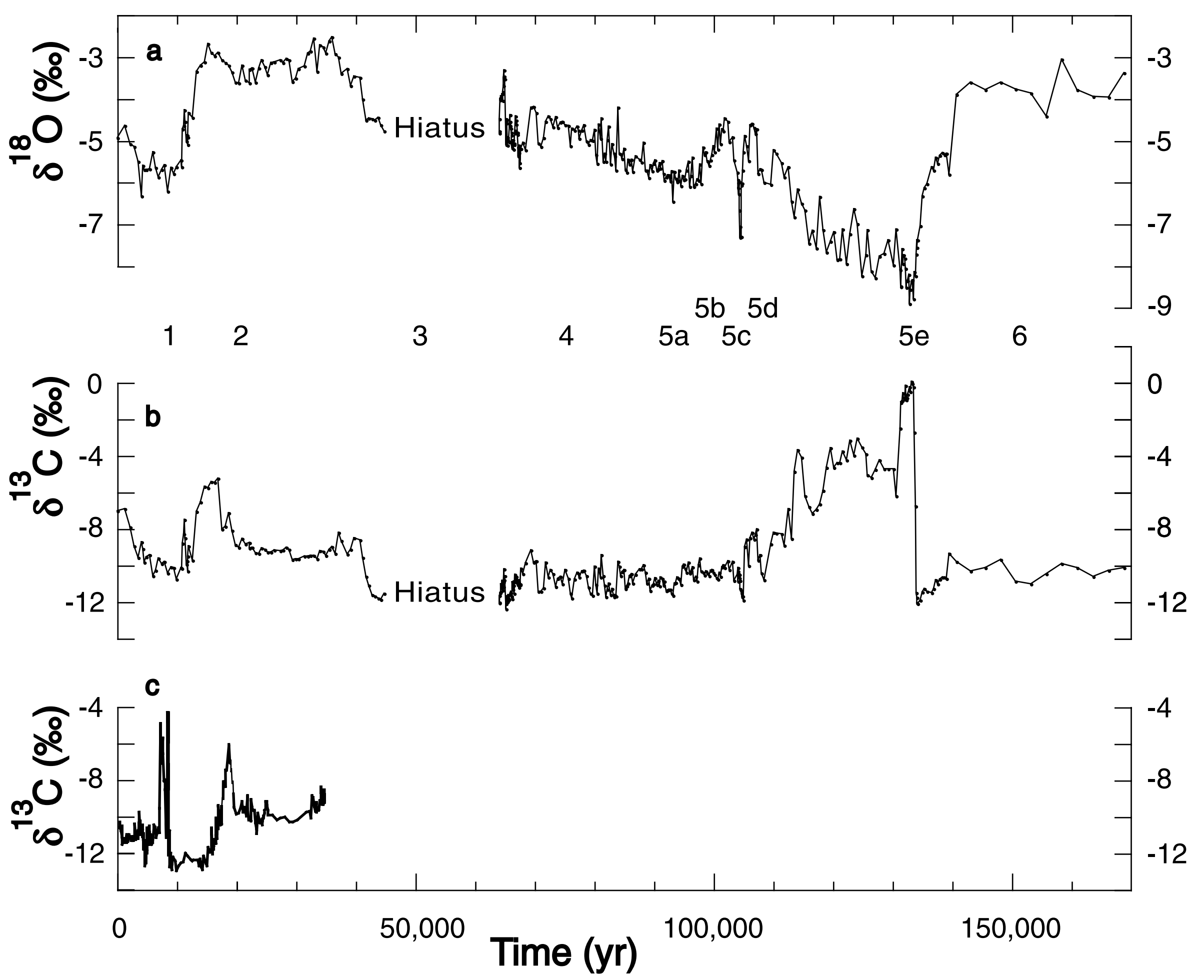

Figure 3. The isotopic record, sampled along the growth axis: (a) The $\delta^{18}O_c$ profile of the stalagmite AF12 (in ‰ versus PDB) [*Frumkin et al.*, 1999b]. (b) The $\delta^{13}C$ profile of AF12 (in ‰ VPDB). Numbers are in isotope stages. (C) Composite $\delta^{13}C$ from Soreq Cave (14 km west of Jerusalem), for comparison [*Bar-Matthews et al.*, 1998; *Bar-Matthews et al.*, 1999].

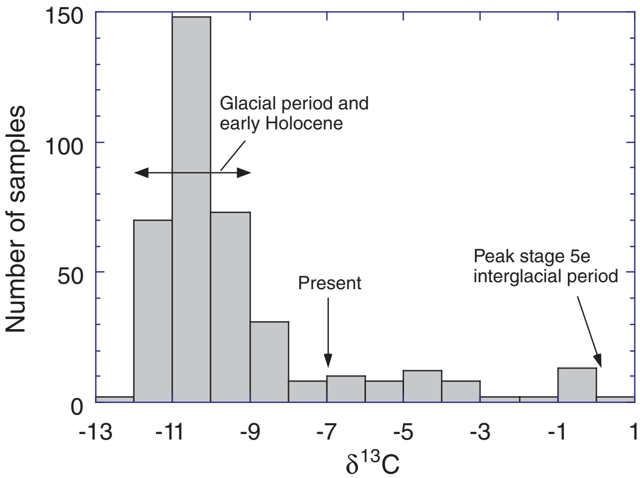

Figure 4. Distribution of $\delta^{13}C$ values in samples from AF12 profile.

grass species) increases from 0 at the cave (700 mm annual precipitation) through 18% in the Judean Desert (~220 mm rain), rising to 50% at the Dead Sea Valley (~75 mm rain) [*Vogel et al.*, 1986].

Bedrock $\delta^{13}C$ of the Amminadav Formation is 2‰ PDB [*Magaritz*, 1973]. Combined with the above considerations, we deduce that under conditions of entire C_3 vegetation, having $\delta^{13}C$ of about −26‰ [*Cerling and Quade*, 1993], we should expect speleothem $\delta^{13}C$ values to be approximately −10 to −12‰. This is indeed the case for much of our record (Figure 3) and is typically observed in many other speleothems in the region [*Bar-Matthews et al.*, 1999; *Geyh*, 1994].

Further enrichment in $\delta^{13}C$ in speleothem may be caused by (1) an increase in the fraction of C_4 plants having $\delta^{13}C$ value of about −14‰: The past dominance of C_4 plants in the vegetation over the cave could have raised the $\delta^{13}C$ value of the CO_2 component by no more than a few per mil. Today, an interglacial period, we do not see C_4 vegetation close to the cave. However, past periods of high $\delta^{13}C$ in the speleothem may be because C_4 plants entered this area due to a climate hotter and drier than today. Further enrichment in $\delta^{13}C$ in speleothem may also be caused by (2) a decrease in total amount of biogenic activity: This would also cause increase in $\delta^{13}C$ of speleothem. In the absence of vegetation, the isotope ratio is determined by mixing of the open atmospheric CO_2 (−7‰) and local bedrock (2‰). Considering equilibrium calcite-HCO_3 carbon

soil atmosphere. Our range of $\delta^{13}C$ values (Figure 4) is very close to the global range of −13 to +2‰ for soil carbonates, derived from direct measurements and a soil CO_2 diffusion model [*Quade et al.*, 1989; *Cerling and Quade*, 1993], indicating that the cave and soil carbon systems act similarly. The large amplitude shifts of $\delta^{13}C$ values observed in AF12 are most likely related to shifts in the $\delta^{13}C$ value of soil atmosphere.

Caution must be implemented, however, to avoid elevated $\delta^{13}C$ values resulting from short residence time of the water in the soil or fractionation due to degassing or evaporation above the cave and inside it [*Baker*, 1997]. We therefore selected AF12 carefully out of three tested caves and seven speleothems [*Frumkin et al.*, 1994]: The soil above the cave is terra rossa with low infiltration rates, containing almost 100% clay [*Yaalon*, 1955]. When soil is removed, higher $\delta^{13}C$ values are expected, as described below in this section. The cave is only 5 m below surface, so degassing between the soil and the cave is unlikely. Kinetic fractionation in the cave is avoided by selection of a small sealed cave, a stalagmite standing just a few centimeters below the dripping point, and careful checking of equilibrium deposition, using 25 samples along growth layers [*Frumkin et al.*, 1994]. Finally, by using a single speleothem deposited during a long period, we avoid the variability that may exist when using composite records whose parts may have experienced different fractionation histories. We conclude that our record reflects mainly temporal shifts in $\delta^{13}C$ values of soil CO_2 which in turn varies inversely with vegetation density and directly with the ratio of C_4/C_3 plants above the cave, both functions of climate.

C_4 plants in Israel belong mainly to *Gramineae* grasses and *Chenopodiaceae* shrubs, which have $\delta^{13}C$ values of approximately −14‰. These plants are common in terms of biomass and coverage where precipitation is low and temperature is high throughout the year [*Shomer-Ilan*, 1983; *Vogel et al.*, 1986]. A good correlation exists between the frequency of grasses and annual precipitation (Figure 5), although other factors, such as temperature, rock type, and fissure density cause deviations from the main trend [*Vogel et al.*, 1986; *Yair and Danin*, 1980]. More specifically, in the Jerusalem region, the C_4 grasses percentage (of the total number of

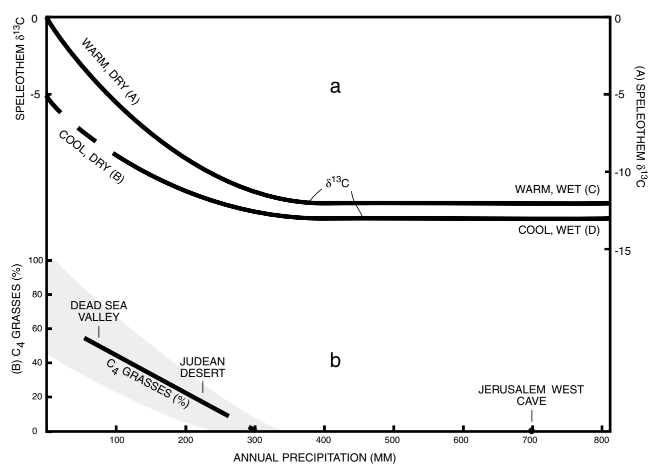

Figure 5. (a) Suggested schematic relation between $\delta^{13}C$ value of speleothem, mean annual precipitation, and temperature. The $\delta^{13}C$ value is apparently sensitive to changes of precipitation at the transition between arid and Mediterranean climate but reaches a saturation value at about −13‰ and remains constant during wetter conditions. Higher temperature tends to raise the saturation value: The upper line commonly represents interglacial periods, and the lower line commonly represents glacial periods (although deviations are not rare). Letters in parentheses denote quadrants of Figure 6. (b) Frequency of C_4 grass species as a percentage of total number of grass species in a climatic gradient section from Jerusalem eastward toward the Dead Sea, as a function of precipitation. The gray area indicates the scatter of data in Israel and Sinai Desert, due to additional controlling factors (see text); on the basis of data from *Vogel et al.* [1986].

isotope fractionation and diffusion fractionation effects, the theoretical $\delta^{13}C$ can be ~2‰, supported by measurements [*Quade et al.*, 1989; *Romanek et al.*, 1992]. The highest $\delta^{13}C$ values in AF12, ~0‰ (Figure 3), can be accounted for by carbonate rock outcrops with hardly any vegetation. Under such conditions, common in the deserts of Israel today, the moisture from the rain equilibrates with CO_2 of air and CO_2 released by lichens and cyanobacteria, followed by slight solute enrichment from aeolian dust and the epikarst zone (the top ~1 m of bedrock above the cave) [*Danin*, 1983; *Danin et al.*, 1983]. We would expect rather slow growth rates for speleothems because the rate of dissolution of calcite is slow under these conditions. Without biogenic CO_2, calcite can also be deposited when the temperature inside the cave is warmer than above ground [*Dreybrodt*, 1982], i.e., during the winter, which is also the wet season today. Unfortunately, the resolution of U-series dates for AF12 over stage 5e does not allow us to estimate the growth rate.

The potential influence of biogenic activity and C_4/C_3 ratio is also seen in $\delta^{13}C$ values of soil calcic horizons, which are generally similar to that of speleothems. Soil calcite $\delta^{13}C$ increases by 10‰ from about −11‰ in the latitude of Jerusalem to about −1‰ 80 km southward toward the Negev Desert [*Goodfriend and Magaritz*, 1988].

Changes in atmospheric CO_2 concentration over the glacial/interglacial cycle could also influence the $\delta^{13}C$ record by changing vegetational primary production. However, in the warm conditions of Jerusalem, the main factor limiting biogenic activity is the amount of available moisture [*Zohary*, 1982]. We therefore believe that our $\delta^{13}C$ record primarily reflects changes in precipitation, with periods of low rainfall and consequent sparse vegetation producing high $\delta^{13}C$ values up to a limit of 0‰, while periods of high rainfall lead to development of a thicker soil profile, and precipitation of speleothem with $\delta^{13}C$ values approaching −12‰.

Unlike $\delta^{18}O^c$, we should not expect speleothem $\delta^{13}C$ values at a given site to be identical to other sites in the region because all depend critically on the local density and type of vegetation above the caves, which will vary on small spatial scales. Nevertheless, a general agreement is apparent between the $\delta^{13}C$ records of AF12 and the record from Soreq Cave, 30 km to the west (Figure 3). The main exception is a short period in the early Holocene, when a strong increase in $\delta^{13}C$ values was observed only in Soreq Cave. The absence of this feature from both the Jerusalem and Galilee isotope records [*Geyh*, 1994] indicates that local factors also affected early Holocene $\delta^{13}C$ in Soreq Cave.

The total range of $\delta^{13}C$ in AF12 is 12‰ (Figure 3b), suggesting large shifts in biogenic activity. As noted, the most common values fall between −12 and −9‰ (Figure 4) and about half of the samples are even more constrained, −11 to −10‰. This range is typical of the glacial periods as well as late stage 5 and the early Holocene. These low $\delta^{13}C$ values must reflect the presence of extensive C_3 vegetation and probably more intense biogenic activity than today. This suggests conditions favorable to C_3 vegetation, higher precipitation and/or lower temperatures and/or summer rains. Such conditions may have prevailed, with some fluctuations, in stages 6, 4, 3, and most of stage 2 (Figure 3b).

When climate becomes wetter and cooler, $\delta^{13}C$ values decrease to about −12 to −13‰ (Figure 5). Some 15 caves have been studied in Israel but $\delta^{13}C$ values lower than −13‰ have almost never been observed [*Frumkin et al.*, 1994; *Geyh*, 1994; *Bar-Matthews et al.*, 1999; *Frumkin et al.*, 1999a, 1999b]. This seems to be a saturation value, representing C_3 vegetation of relatively high density, beyond which there is no further depletion in response to increasing precipitation and decreasing temperatures. This suggests that $\delta^{13}C$ is useful for identifying shifts from arid to normally wet conditions but cannot resolve detailed changes in precipitation and temperature within generally wet, cool periods. Shifts between arid and Mediterranean conditions are readily observed in the $\delta^{13}C$ record because of the major changes both in vegetation density and in C_4/C_3 biomass ratio.

Very high and fluctuating $\delta^{13}C$ values during stage 5e suggest extremely dry and unstable conditions. The short (3000 years) period of depletion in ^{13}C at the end of stage 6 is followed by sharp enrichment in ^{13}C by 12‰ up to 0‰ in Termination II. These events are coeval with an 8000 year long decrease in $\delta^{13}O_c$ at the end of the penultimate ice age. This isotopic threshold represents the most intense environmental event in the whole record. It is probably a consequence of rapid warming and drying, causing destruction of C^3 vegetation that had flourished during the penultimate glacial period. Soil cover was probably lost when the vegetation disappeared abruptly, possibly with the aid of forest fires. Action of fire alone may be discounted as the prime cause of the isotopic shift, however, since the duration of the period of change and of very high $\delta^{13}C$ values amounts to many centuries. A later decrease of $\delta^{13}C$ values to approximately −5‰ indicates that there was an increase in plant cover, possibly involving some C_4 vegetation. About 12,000 years later, the start of gradual decrease in ???C values with fluctuations suggests reintroduction of abundant C3 vegetation. The $\delta^{13}C$ values continued to fluctuate until relative stability was restored in late stage 5. Low $\delta^{13}C$ values from then until stage 3 indicate a stable cover dominated by C_3 vegetation. The $\delta^{13}C$ value is generally less stable in the last interglaciation (stage 5e–5c) than in either of the glacial periods represented here. This could be because during interglacial periods, Jerusalem is located (as today) near the boundary of two climatic zones, desert and Mediterranean, whereas during full glacial periods, this boundary shifts to the southeast. Fluctuations in the position of the boundary across Jerusalem could lead to larger oscillations in $\delta^{13}C$ values.

Another distinct enrichment of $\delta^{13}C$ values occurred just after the end of the last glacial maximum (~17,000 years). This event suggests that the last glacial period culminated in an intense, short dry event with reduced biogenic activity, apparently the converse of the climatic conditions just prior to Termination II. Rapid recovery to wetter (not drier as in Termination II) conditions is indicated by a depletion of ~5‰ at stage 2/1 transition (Termination I). This rapid change is interrupted by a short increase in $\delta^{13}C$ values that may represent the ~12,000 year B.P. "Younger Dryas" event [*Roberts et al.*, 1993; *Taylor et al.*, 1997]. The jumps in $\delta^{13}C$ values during the last glacial maximum-Holocene transition were generally smaller compared with Termination II. During both Terminations II and I the region experienced an extreme dry event. However, the Termination I dry event occurred at the onset of deglaciation and was shorter and milder, with rapid return to intermediate values at the first half of the Holocene. This is unlike Termination II, when the dry climate occurred after deglaciation and then persisted throughout stage 5e. For the early Holocene the combination of high temperatures (indicated by $\delta^{18}O_c$) and lower $\delta^{13}C$ values suggests a warmer, wetter climate than today. The gradual enrichment in $\delta^{13}C$ from the early Holocene to today probably reflects a shift to lower vegetation cover, as no C_4 plants are present in the region today. Such degradation could have been caused by a natural drying trend but is probably also a result of the increasing impact of grazing and deforestation. This degradation probably caused loss of some soil from the slopes above the cave.

5. Climate Reflected in $\delta^{13}C$ – $\delta^{18}O_c$ Interrelationship

AF12 shows no increase of $\delta^{18}O_c$ and δ^{13} values nor correlation between the two isotope values along growth layers [*Frumkin et al.*, 1994]. This indicates that there are no kinetic fractionation effects during deposition while the water flows along the stalagmite face [*Hendy*, 1971; *Gascoyne*, 1992].

Under such conditions the long-term interrelationships between the $\delta^{13}C$ and $\delta^{18}O_c$ measured across the layers may provide further insight into paleoclimatic variability. Some positive correlation through time between the two variables is apparent during the glacial periods, while there is negative correlation during stage 5e. This pattern may indicate some kinetic effects. Alternatively, it can be associated with a change of precipitation season: summer rains may favor positive $\delta^{13}C$–$\delta^{18}O_c$ correlation, while winter growing season typical of interglacials may induce a different relationship [*Cerling and Quade*, 1993]. The $\delta^{13}C$ record reflects predominantly local environmental changes, while the $\delta^{18}O_c$ record mainly reflects the isotopic signature of the water source and trajectory. Therefore one universal correlation between the two variables that applies to all periods is unlikely. Today, both $\delta^{13}C$ and $\delta^{18}O_c$ values fall in the midrange of the late Quaternary values. The $\delta^{13}C$–$\delta^{18}O_c$ space may be divided into four quadrants with respect to the modern values (Figure 6). The data points form a broad asymmetric U-shaped pattern. We divide the last 170,000 years into subperiods so that each subperiod falls into one or two quadrants that can be distinguished climatically from the present. A simplified model is suggested here to identify the main climatic variables. In this model, $\delta^{18}O_c$ is assumed to reflect mainly temperature changes, in concert with the global glacial/interglacial cycle [*Frumkin et al.*, 1999b]. The $\delta^{13}C$ record is assumed to reflect changes in precipitation (amount, season, and intensity), the limiting factor for vegetation development in the region. The terms dry, wet, warm, and cool are used qualitatively for comparison with the "present conditions" relative reference point, noting that the present environment is influenced also by human factors.

During late stage 6 (170,000–140,000 years B.P.), all data points fall in quadrant D, suggesting cool, wet conditions with dense vegetation. This cool and wet climate, repeated in most of stage 2, seems to be typical of glacial periods. The short-term depletion of both $\delta^{13}C$ and $\delta^{18}O_c$ values at 140,000–134,000 years B.P. (quadrant C) indicates rising temperatures ($\delta^{18}O_c$ depletion) and possibly increasing precipitation (decreasing $\delta^{13}C$ values) at the beginning of Termination II. This slight drop in $\delta^{13}C$ values could also be due to some change in the isotopic balance of the soil water CO_2^-–HCO_3^- system because of the rising temperature. Increase in rate of oxidation of plant debris would increase soil pCO_2 and shift the HCO_3^-–H_2CO_3 partitioning in drip waters, leading to lower $\delta^{13}C$ values of HCO_3^-. Then, at the culmination of stage 6/5 deglaciation, vegetation changed dramatically, indicated by the 12‰ rise in $\delta^{13}C$, shifting isotopic values to quadrant A from 134,000 to 114,000 years B.P. The extreme isotopic values in this stage suggest that it was the driest (lowest $\delta^{13}C$ values) and warmest (lowest $\delta^{18}O_c$ values) period in the record. Rain was probably limited to rare high-intensity winter

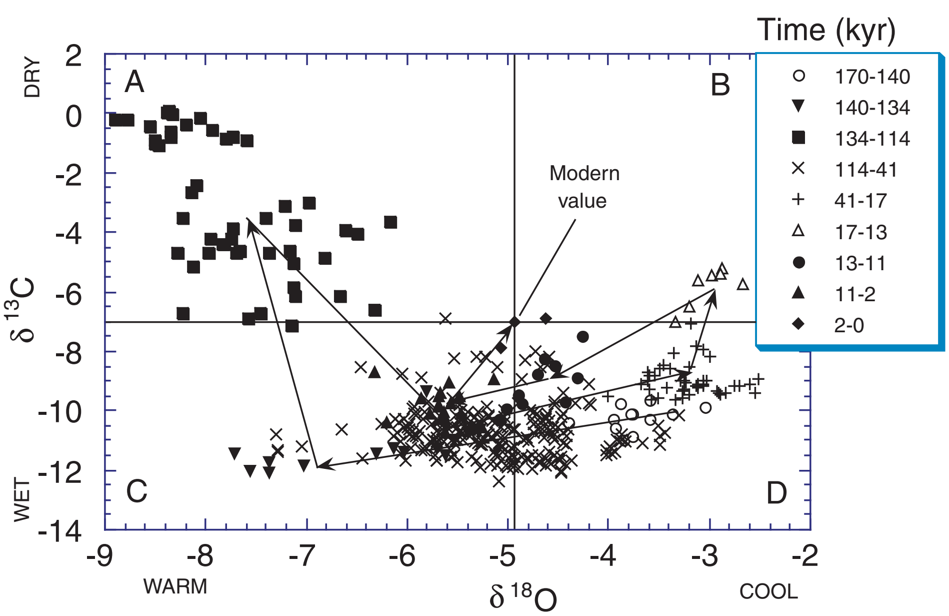

Figure 6. The $\delta^{13}C$–$\delta^{18}O_c$ relationship in AF12 samples. Arrows indicate the general changing trend of isotopic values through time. A simplified model relates $\delta^{18}O_c$ mainly to temperature change and $\delta^{13}C$ mainly to precipitation change. This model does not consider other variables which are also reflected in the isotopic values [see, also, *Frumkin et al.*, 1999b].

showers with rapid discharge of surface runoff, similar to the present arid conditions in the Negev and Judean Desert. Destruction of the previous vegetation probably allowed the stripping of soil cover. Vegetation of high C_4/C_3 ratio was probably the first to colonize the stripped surface, while the climate remained arid. Large fluctuations in both $\delta^{18}O_c$ and $\delta^{13}C$ are common at this period, possibly due to migration of the regional climatic zone boundary back and forth across the site. Climate became gradually wetter and cooler with an increasing proportion of C_3 vegetation, finally switching to a wetter and cooler mode at 114,000–41,000 years B.P. (quadrants C and D), with dense C_3 vegetation. Temperatures decreased gradually, still with some fluctuations, mainly in the first part of the period. A short event (stage 5c?) of 3‰ decrease in $\delta^{18}O_c$ values $\sim$105,000 years B.P. is not reflected in the $\delta^{18}C$ record. This supports our observation that $\delta^{18}C$ reflects mainly local environmental changes, while $\delta^{18}O_c$ may represent geographically remote events, whose $\delta^{18}O$ values are carried over long distances by precipitation [*Frumkin et al.*, 1999b]. Such an event could possibly be an arrival of water in the east Mediterranean with low-$\delta^{18}O$ values, associated with water stratification and formation of sapropel 4 at $\sim$105,000 years B.P [*Fontugne and Calvert*, 1992; *Jenkins and Williams*, 1983/1984]. The constant $\delta^{13}C$ values during this $\delta^{18}O_c$ event indicate that no distinct environmental change took place in Jerusalem at that time.

Climate became cooler gradually (with some fluctuations) from 41,000 to 17,000 years B.P. (quadrant D). Fossil traces of shrub and tree roots a few kilometers east of Jerusalem indicate that at the beginning of this period Mediterranean vegetation still extended toward the presently semiarid zone [*Danin et al.*, 1987]. The coldest and driest conditions were reached around 17,000 years B.P. (quadrant B). Gradual warming and increasing precipitation are observed from 17,000 years B.P. until $\sim$13,000 years B.P. Between 13,000 and 11,000 years B.P. (quadrants C and D), precipitation and temperature were relatively high, with a short cool and dry "Younger Dryas" spell $\sim$12,000 years B.P. These Younger Dryas conditions are in agreement with pollen and archaeological evidence from Israel [*Bar-Yosef and Valla*, 1991; *Baruch and Bottema*, 1991]. Warmer and wetter conditions during the early Holocene (quadrant C) gave place to gradual desiccation during the middle to late Holocene. The late Holocene brought about drier conditions but was overprinted by anthropogenic effects.

6. Conclusion

We find that speleothem $\delta^{13}C$ is an excellent proxy for determining the climatic fluctuations between Mediterranean and desert conditions, but it is not so sensitive to changes between Mediterranean and more humid climate. The last interglacial period, stage 5e, in Jerusalem appears to have been warmer, drier, and more unstable than the Holocene. These conditions allowed the northward expansion of the African and Saharo-Arabian biotic zone in Israel [*Tchernov*, 1994]. This was also a period when anatomically modern humans first entered this region, also possibly emerging from Africa [*Bar-Yosef and Vandermeersch*, 1981; *Schwarcz*, 1994]. Only when cooler glacial period conditions returned at the end of stage 5 were the Saharo-Arabian species pushed southward and palearctic biota able to invade the region [*Tchernov*, 1994]. Our record of climatic instability during stage 5 is in agreement with records from Greenland ice [*Dansgaard et al.*, 1993] and European pollen [*Field et al.*, 1994; *Thouveny et al.*, 1994]. The warmer than present temperatures of stage 5e also agree with similar indications from other records [*Burckle*, 1993; *LIGA Members*, 1991; *Sejrup and Larsen*, 1991]. The climatic change of late stage 5e may serve as an analog for the future of the present interglacial period, although its general features are not similar to the present interglacial. The isotopic record shows that after an initial hot and dry period, full interglacial conditions in Jerusalem ended with a gradual, fluctuating transition to glacial conditions. Glacial conditions in Jerusalem were typically wetter and cooler than today. These conditions are observed during most of stages 6, 4, 3, 2, and also the later part of stage 5. The association of wet regional climate with glacial periods is evident also in other proxies in Israel, particularly lake levels and pollen [e.g. *Begin et al.*, 1985; *Horowitz*, 1979; *Issar et al.*, 1985; *Schramm et al.*, 1997; *Weinstein*, 1976], indicating that the glacial climate was commonly wet throughout the entire region. This was probably caused by a southward migration of the westerlies and Saharan climatic belts, associated with the high-pressure zone over the north European ice sheet.

There has been a claim that interglacial climate of the Negev Desert was wetter than glacial climate, based on dating of aquatic sediments [*Livnat and Kronfeld*, 1985]. However, the scatter of the given dates shows that the Negev travertines were deposited not only during interglacials but also during isotope stages 6 and 4. The glacial (stage 2) high values of speleothem $\delta^{18}O$ in Soreq Cave were interpreted as indicating less rainfall than today [*Bar-Matthews et al.*, 1997]. However, further study showed that these high-$\delta^{18}O$ values mainly reflect the isotopic signal of the Mediterranean source of the rainfall, which has a higher amplitude than local rain signal [*Frumkin et al.*, 1999b; *Bar-Matthews et al.*, 1999]. Other continental evidence, such as geomorphology, soil carbonates, and sand dunes [*Goodfriend and Magaritz*, 1988; *Goldberg*, 1994] are fragmentary in their time span but generally agree with the paleoclimatic framework presented here, within the dating error.

Stage 2 in Jerusalem was terminated by cold, dry conditions and increasing instability, with the highest $\delta^{13}C$ values of the last 100,000 years. This dry period, observed in our record $\sim$19,000–14,000 years B.P. caused the drying of Lake Lisan [*Begin et al.*, 1985; *Kaufman et al.*, 1992; *Schramm et al.*, 1997]. The intense dry event was also observed in other proxies in Israel, such as soil carbonates [*Gat and Magaritz*, 1980; *Goodfriend and Magaritz*, 1988] and pollen [*Horowitz*, 1971; *Baruch and Bottema*, 1991], but some records are slightly shifted in phase owing to dating uncertainties.

The early Holocene was wetter and warmer but deteriorated again to the present drier conditions. Regional conditions during the present interglacial period are unlike the previous one, which was drier and warmer. Apparently, during stage 5e, the desert advanced to Jerusalem, associated with a northward shift of the Saharan high-pressure system. Conversely, the early Holocene orbitally induced expansion of the African monsoon [*COHMAP Members*, 1988; *Dolly et al.* 1998] could have slightly affected Jerusalem, bringing about wetter climate.

Our results show that climatic instability during interglacial periods is greater than during glacial periods in this continental region. The instability in $\delta^{18}O_c$ suggests that the interglacial environment is very sensitive to changes in atmospheric conditions, whereas the oscillations in $\delta^{13}C$ may be more a result of shifting climatic boundaries.

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