



THE MINISTRY OF NATIONAL
INFRASTRUCTURES
GEOLOGICAL SURVEY OF ISRAEL

Quaternary paleoclimate of the north-eastern boundary of the Saharan Desert: reconstruction from speleothems of Negev Desert, Israel

Vaks Anton



© Published by the Geological Survey of Israel
30 Malkhe Israel St., Jerusalem 95501, Israel

Cover design: Nili Almog

Front cover: A stalactite grove in Ashalim Cave, central Negev Desert, Israel



THE MINISTRY OF NATIONAL
INFRASTRUCTURES
GEOLOGICAL SURVEY OF ISRAEL

Quaternary paleoclimate of the north-eastern boundary of the Saharan Desert: reconstruction from speleothems of Negev Desert, Israel

Vaks Anton

This work was submitted for the degree "Doctor of Philosophy"
to the Senate of the Hebrew University, Jerusalem.

The study was carried out under the supervision of:

Dr. Miryam Bar-Matthews, Geological Survey of Israel

Prof. Alan Matthews, Institute of Earth Sciences, Hebrew University of Jerusalem

Prof. Amos Frumkin, Department of Geography, Hebrew University of Jerusalem

Acknowledgements:

I would like to thank everyone who assisted me to bring this work to its successful completion.

I had an opportunity to learn from excellent advisors: Dr. Miryam Bar-Matthews, Prof. Alan Matthews and Prof. Amos Frumkin, and I thank them very much for guidance, advising, fruitful discussions, encouragement, being ready to help every time, even during personal vacations, or in 90-th moment before the deadlines.

I would like to very much thank Dr. Avner Ayalon for guidance and advice, and for his generous help with the stable isotope analyses. I would like to thank: Dr. Bettina Shilman and Efrat Eliani for help with performing the stable isotope analyses; Natalya Teplakov, Dr. Ludwik Halicz, Dr. Irena Segal, Dr. Olga Yoffe, Dr. Sarah Ehrlich, Dina Stiber and Galit Sharabi for guidance and help with chemical and mass spectrometric analyses; to Dr. Ahuva Almogi-Labin for fruitful discussions and help with understanding of the marine records, and students Liraz Laor, Ruthi Kiro, Sharona Shlomi, Neta Shalev, Asaf Wunsch, Gilad Garber and Yael Neumieir for the assistance in the laboratory and in the field.

I would like to thank to the previous Director of the Geological Survey of Israel Dr. Amos Bein and the present Director, Dr. Benjamin-Ze'ev Begin for help and encouragement, to Onn Crouvi, Dr. Yoav Avni, Dr. Rivka Amit, Dr. Ezra Zilberman, Dr. Naomi Porat and Dr. Rani Calvo for help with understanding soil and sedimentary records of the Negev climate, to Michael Dvoracheck for guidance and help with SEM analyses, to Dr. Amir Sandler, Yoetz Deutsch, Uri Simchai and Shlomo Leyzenbah for help with XRD analyses, and to Michael Kitin for help with sieving the soil samples. I would like to thank Dr. Henry Foner for help with understanding of the dust fall patterns in the region.

Many thanks to Prof. Jon Woodhead from University of Melbourne, Australia, and to Prof. Robert Cliff from University of Leeds, United Kingdom, for U-Pb dating of speleothems which are older than U-Th dating limit.

I would especially like to thank Shlomo Ashkenazi for his great help in the field and laboratory and especially for being good friend, to Yehuda Peled, Eli Ram, Yaakov Mizrachi, Ariel Gai, Robert Knafo, Yaakov Rafael for the help in field work,

to Moshe Peri for preparation of the thin sections, to Camille Lousqui and Ariel Belmas for help with purchasing an equipment, to Channa Netzer-Cohen, Bat-Sheva Cohen, and Nili Almog for help with graphics and publications, to vice manager of Geological Survey of Israel Shmuel Zitronblat, to Ira Peer, Anna Feigin, Masha Kahalani, Valery Doudin, David Sidi, Oksana Zaripov and to other people of Geological Survey of Israel that was for me a good home during many years.

Special thanks to Dr. Mordechai Stein and Prof. Yehouda Enzel for fruitful discussions. Although we had disagreements (and maybe because of them), these discussions contributed a lot to improve the quality of this work.

I would like to thank Prof. Uri Dayan and Prof. Baruch Ziv for help with understanding of meteorological systems in the region. Many thanks to Dr. Lior Grossman, Prof. Erella Hovers, Dr. Rivka Rabinovich, Prof. Na'ama Goren-Inbar from the Institute of Archeology of the Hebrew University for help with understanding the archaeological issues, to Dafna Kadosh, Nicolas Waldman, Adi Torfshtein, and Itai Haviv for fruitful discussions, to Yael Yakobi, Surin Lisker, Roi Porat, Azriel Raikin and other members of Israeli Cave Research center for the assistance in the field work.

Many thanks to my parents Evgenya Vaks-Reznik and Evgeny Vaks from Be'er-Sheva, Nadav Bloch, Smadar Bloch-Wiener and Zvi Wiener from Arad, Rakefet Dar from Neot-Smadar and Dr. Georgy Shenbrot from Ramon Science center for the assistance with the rain sampling. Many thanks to the secretaries of the Institute of Earth Sciences, Hebrew University, Maggi Parkin, Carmela Lev, Batia Moshe, Mali Shisha-Halevi.

I would like to thank to anonymous reviewers whose remarks helped to increase the quality of this work.

Thanks a lot to my friends Dan Asael, Borianna Kalderon, Liraz Laor and Dina Vachtman for their support, help and friendly good word of encouragement.

For my dear parents Evgenya Vaks-Reznik and Evgeny Vaks, and my brother David Vaks many thanks for their love and support.

I would like to thank and apologize to all people that I maybe inadvertently omitted from these acknowledgements.

Abstract

Carbonate cave deposits (speleothems) from caves of the Israeli deserts were used to reconstruct the paleoclimate conditions of the northern boundary of Saharan-Arabian Desert. The deserts studied in this work are of two types: the Negev Desert in southern Israel, which is part of the Saharan-Arabian desert belt; and the "rain shadow" desert located east of Central Mountain Ridge (CMR) in the rift valleys of the Jordan River and Dead-Sea (Judean Desert). The presence of speleothems in numerous caves in these present-day arid regions indicates that humid climatic conditions (i.e., periods with positive effective precipitation/infiltration index) occurred in the past. In this study, the timing of the speleothem growth, as an indicator of periods of increased effective precipitation, was obtained by precise U-Th dating; the origin of rainfall, its amounts and the types of vegetation were examined from studies of the speleothem $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values and the δD values of their fluid inclusions, and the sources of dust were studied through the Sr isotopic composition of the speleothems.

The minimum precipitation amounts necessary to deposit speleothems are estimated to be: 200-275 mm/year during glacial periods and 300-350 mm/year during interglacial periods. Speleothem deposition in the Jordan Valley mainly occurred during the three last glacial periods, with minor deposition occurring during Termination II (~135 ka) and MIS-7 interglacial (225-205 ka), and no speleothem deposition taking place during the Holocene, the peak of last glacial maximum (~19 ka), and for most of the previous interglacial MIS-5 (130-75 ka). The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profiles of speleothems deposited between 67 ka and 25 ka in the Jordan Valley match the general isotopic trends of previously studied speleothems from central and northern Israel; suggesting a similar Eastern Mediterranean (EM) Sea source for the precipitation and similar climatic conditions. Decrease in temperature and evaporation, and the consequent increase in effective precipitation, were probably the major factors controlling the decay of the "rain shadow" effect during glacial periods.

In caves of the northern Negev located on the present-day 150-160 mm isohyet, most of the speleothem deposition occurred during glacial periods (as in the Jordan Valley), and only restricted, episodic deposition occurred during the interglacial periods and glacial maxima. No speleothem deposition occurred at 117-96 ka and 92-84 ka (MIS-5(3-2)) and during the Holocene, indicating that the climate in these periods was similar to present or even more arid. The $\delta^{18}\text{O}$ values of the speleothems mostly show the same trends as speleothems from central and northern Israel; thus pointing to the EM Sea as the precipitation source. This source is also indicated by the δD - $\delta^{18}\text{O}$ relations derived from fluid inclusions. Speleothem $\delta^{13}\text{C}$ values in the northern Negev show that the vegetation was usually C3+C4 steppe, with rare invasions of Mediterranean C3 species during the most humid events, or that the area experienced episodes of desertification. Generally the conditions were drier than in the Jordan Valley. The maximum southward shift of the boundary between the Mediterranean climate and the semi-desert zone during humid periods was 20-25 km.

Most speleothem deposition in the Judean Desert also occurred during the three last glacial periods, with minor deposition during the two last interglacials. Climate conditions there were generally drier than in the northern Negev and Jordan Valley.

Major humid periods in the central and southern Negev Desert (Negev Humid Periods – NHP) during the last 350 ka occurred at: 350-310 ka (NHP-4), 310-290 ka (NHP-3), 220-190 ka (NHP-2), and 142-109 ka (NHP-1). NHP-4 was most intense in the southern Negev and probably occurred as a result of rainfall of tropical origin, whereas NHP-3, 2 and 1 were dominant in the central Negev, and only NHP-1 reached the southern Negev. Only NHP-3 was associated with a glacial period (MIS-8). The rainfall episodes during NHP were short-lived, as evident from very thin speleothem laminae and the very minor development of semi-desert C4/steppe C3+C4 indicated by $\delta^{13}\text{C}$ values. Precipitation during NHP 3, 2 and 1 originated in the EM Sea (Atlantic-Mediterranean Cyclones), as evident from the δD - $\delta^{18}\text{O}$ relations in the fluid inclusions and the systematic decrease in speleothem numbers from north to south. The $\delta^{18}\text{O}$ values of all Negev speleothems are, however, systematically lower

than in contemporaneous speleothems from the Jordan Valley, central and northern Israel. This in part is attributed to the increased rainout of the heavy isotopes by Rayleigh fractionation processes.

The dominant speleothem deposition during interglacial periods in the central and southern Negev contrasts with the northern Negev, Jordan Valley and Judean Deserts, where most deposition occurred during glacial periods. Glacial speleothem deposition in the latter areas is associated with high infiltration coefficients during the cold glacial periods. During humid interglacial episodes the intensity of the Cyprus cyclones was higher than during the glacials, bringing precipitation high as 300-350 mm as far south as the present-day ~50 mm isohyet. However, higher temperatures and lower effective precipitation only allowed limited water infiltration to the caves and the deposition of thin speleothem laminae.

The major decreases of speleothem $\delta^{18}\text{O}$ in Soreq Cave at 132 ka and 127-126 ka and in Peqi'in Cave at 200 ka occurred simultaneously with the highest peaks of speleothem deposition in the Negev Desert during the NHP-1 and NHP-2. These events were probably associated with most intensive humid episodes above the EM Sea.

NHP events were contemporaneous with periods where the monsoon index ≥ 51 cal/cm²×day and sapropel formation occurred in the Mediterranean Sea. Simultaneous intensification of the monsoon and Atlantic-Mediterranean cyclones probably relates to weakening of the high pressure cell above the sub-tropical Atlantic Ocean. Absence of speleothem deposition in the Negev during the Holocene and MIS-11 (430-400 ka) interglacials could be explained by lower Northern Hemisphere insolation and lower African Monsoon index than during the NHPs. Thus, the high pressure cell above subtropical Atlantic Ocean was probably stronger, leading to more northern trajectories of Atlantic-Mediterranean cyclones and preventing from Mediterranean precipitation to reach the Negev.

Speleothem $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the “rain shadow” and the Negev deserts mostly vary between 0.7082 and 0.7085, and show only slight changes between glacial and interglacial periods. Compared to the wider variation observed in central Israel, this range probably reflects lower host rock dissolution rates, or higher dust supply from the proximal Sinai-Nile Delta source characterized by $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of ~ 0.7084 .

Most speleothems of the central and southern Negev Desert are older than the limit of uranium series dating (*ca* 550 ka). Preliminary U-Pb studies give ages of 3.3-2.7 Ma, with a few speleothems deposited at ~ 1.3 Ma. These ages correspond to periods when lakes were known to exist in the region. A reduction in the rate of speleothem deposition, together with an increase in speleothem $^{87}\text{Sr}/^{86}\text{Sr}$ ratios after ~ 2.7 Ma, probably reflects increased desertification of the region and of the entire Saharan-Arabian Desert, which increased the dust supply from Sinai and Saharan-Arabian sources.

Multiple growth / hiatus cycles in a single speleothem over >3 Ma show that in arid environment the fractures conducting water from the surface to the caves remain open over long times.

The Sinai-Negev land bridge was the major, and possibly the only land-bridge, connecting the Africa with Asia since the Miocene, and the Saharan-Arabian Desert was a significant obstacle for hominid and animal migration from Africa to other parts of the world. The occurrence of humid phases in the Sinai-Negev deserts contemporaneous with humid phases in the southern and central Saharan-Arabian Desert probably opened climatic “windows” for migration of hominids and animals out of the African continent. Migration of the early modern humans from Africa to the Levant was probably associated with NHP-1 at 142-109 ka.

Table of contents

1. Introduction	1
2. Geography, climate and geology of the Israeli deserts	3
2.1 Geographical definition and climatic zoning	3
2.2. Rainfall gradients in the northern Negev Desert and Jordan/Dead Sea Valleys	5
2.3 Geology	6
2.4 Speleothems as paleoclimatic records	8
2.5. Speleothem climate record from the Mediterranean climate zone of Israel	11
3. Research goals	14
3.1 The major aims of the study	14
3.2 Structure of the thesis	15
4. Methodology	17
4.1 The studied caves and sampling	17
4.2 U-Th and U-Pb dating	18
4.3 Stable isotope analyses of speleothems and present day rainwater	20
4.4 D/H analyses of speleothem fluid inclusions	22
4.5 Sr isotope analyses	23
5. Vaks, A., Bar-Matthews, Ayalon., A., Schilman, B., M., Gilmour, M., Hawkesworth, C. J., Frumkin, A., Kaufman, A., and Matthews, A. (2003), Paleoclimate reconstruction based on the timing of speleothem growth, oxygen and carbon isotope composition from a cave located in the “rain shadow” in Israel., <i>Quaternary Research</i>, 59, 2, pp 182-193.	25
6. Vaks , A., Bar-Matthews, M., Ayalon, A., Matthews, A., Frumkin, A., Dayan, U., Halicz, L., Almogi-Labin, A. and Schilman, B. (2006), Paleoclimate and location of the border between Mediterranean climate region and the Saharo-Arabian Desert as revealed by speleothems from the northern Negev Desert, Israel., <i>Earth and Planetary Science Letters</i>, 249, 3-4, 384-399.	37
7. Vaks, A., Bar-Matthews, M., Ayalon, A., Matthews, A., Halicz, L. and Frumkin, A., (2007), Desert speleothems reveal climatic window for African exodus of early modern humans, <i>Geology</i>, 35, 9, 831-834.	53
8. Paleoclimate of northern Saharan-Arabian Desert from speleothem record in central and southern Negev and Judea Desert, Israel	57
8.1 Introduction	57
8.2 Results	57

8.2.1 Field and petrographic characterization of the speleothems and the cave sediments	57
8.2.1.1 <i>Vadose speleothems and preliminary U-Pb dating results</i>	57
8.2.1.2 <i>Phreatic speleothems</i>	60
8.2.1.3 <i>Cave sediments and paleosols washed into the caves</i>	61
8.2.2 Uranium concentrations of central and southern Negev speleothems	62
8.2.3 U-Th ages and depositional periods	63
8.2.3.1 <i>Kanaim Cave, Judea Desert and the modified record of Ma'ale-Efrayim Cave, Jordan Valley</i>	63
8.2.3.2 <i>Caves from the central and southern Negev Desert</i>	63
8.2.4 Oxygen and carbon isotopic compositions of the central and southern Negev speleothems	65
8.2.4.1 <i>Oxygen isotopic compositions</i>	66
8.2.4.2 <i>Carbon isotopic compositions</i>	68
8.2.5 Present-day rainfall in the Negev Desert: amounts and hydrogen and oxygen isotopic compositions	69
8.2.6 δD values of speleothem fluid inclusions from northern, central and southern Negev	70
8.2.7 Sr concentrations and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in speleothems, cave host rocks and soils for the Jordan Valley and the Negev Desert	71
8.3 Discussion	73
8.3.1 Humid episodes in central /southern Negev, and Judea deserts	73
8.3.1.1 <i>Speleothem deposition as recorder of effective precipitation</i>	73
8.3.1.2 <i>Humid periods in central and southern Negev Desert</i>	76
8.3.1.3 <i>Humid periods in Judea Desert</i>	79
8.3.2 Origin of precipitation in the Negev Desert under present-day conditions and during the past 550 ka	79
8.3.2.1 <i>Present-day rainfall</i>	79
8.3.2.2 <i>Origin of precipitation in the past</i>	81
8.3.3 Vegetation in the central and southern Negev Desert determined from the $\delta^{13}\text{C}$ of the speleothems and its relation to regional climatic change	85
8.3.4 Correlation between the Negev Humid Periods and the regional and global climate	88
8.3.5 "North – south paradox" of the Negev paleoclimate	90
8.3.6 Pliocene-Pleistocene humid periods in the central and southern Negev Desert as evident from speleothems older than 550 ka	93

8.3.7 Phreatic speleothems in Makhtesh-ha-Qatan as markers of paleo-groundwater levels	95
8.3.8 – Humid periods in the northern Saharan-Arabian Desert as climatic “windows” for Out of Africa dispersals of hominids and animals	95
8.3.9 Sr isotope ratios from Israeli desert speleothems as tracers of the origin of dust and host rock weathering	97
9. Summary and conclusions	101
Bibliography	107
Appendix 1: Figures	119
Figure 1: Geographical location of the study <i>area</i> , annual precipitation and sample sites	121
Figure 2A: Geological stratigraphic section of the research area	122
Figure 2B: Geologic map of the research area.	123
Figure 3: Typical cross sections of the Ashalim Cave speleothems as observed in the field, hand specimens and under SEM	124
Figure 4: Cross section of the flowstone KTO(1)-1 from Ktora Cracks	125
Figure 5: The mineralogy and petrography of the calcite and fine laminae representing hiatuses within the Young member of the stalactite ASH-11	126
Figure 6: Flowstone MMR-7(2) from Ma’ale-ha-Meyshar Cave	127
Figure 7: SEM analysis of the minerals	128
Figure 8: Petrography of stalactite KN-8 from the Kanaïm Cave	129
Figure 9: Phreatic speleothems in Ma’ale-ha-Meyshar Cave	130
Figure 10: The relative frequencies of the speleothem ages in “rain shadow” desert	131
Figure 11: Dating of central and southern Negev speleothems	132
Figure 12: Plots of the relative age frequencies	133
Figure 13: Sample ASH-33 from the Ashalim Cave	134
Figure 14: $\delta^{18}\text{O}$ values of central and southern Negev speleothems during the last ~3 Ma	135
Figure 15: Oxygen isotope profiles of central and southern Negev speleothems during the last 350 ka	136
Figure 16: Oxygen isotope profiles in central Negev during the two major periods of speleothem deposition	137
Figure 17: $\delta^{13}\text{C}$ values of central and southern Negev speleothems during the last ~3 Ma	138
Figure 18: Carbon isotope profiles of central and southern Negev speleothems during the last 350 ka.	139
Figure 19: Carbon isotope profiles in central Negev during the two major periods of speleothem deposition.	140
Figure 20: Northern Negev rainfall isotopic composition	141

Figure 21: Central and southern Negev rainfall isotopic composition	142
Figure 22: δD values of Negev Desert speleothem fluid inclusions	143
Figure 23: Sr concentrations in the Negev and Jordan Valley speleothems	144
Figure 24: $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios of Ma'ale-Efrayim Cave speleothems	145
Figure 25: $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Negev Desert vadose speleothems	146
Figure 26: $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in Negev Desert vadose speleothems compared with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of dolomitic host rock, of bulk soil above the caves and of its silicate fraction	147
Figure 27: Hydrogen and oxygen isotopic composition of paleo-cave waters based of speleothem fluid inclusion analysis	148
Figure 28: d-excess values from speleothem fluid inclusions in central and southern Negev Desert during the last 220 ka	149
Figure 29: Thinning of the speleothem sequences from the north to the south	150
Figure 29 (A-G): Width of the NHP-1 and NHP-2 sequences in the Negev speleothems	150
Figure 29 (H, I) Thinning of the speleothem laminae from the north to the south in the Negev Desert during the NHP-1 and during the NHP-2	151
Figure 29J: Shortening of the speleothem deposition periods from north to south.	152
Figure 30: Comparison of the $\delta^{18}\text{O}$ profiles and age frequencies of central Negev speleothems with $\delta^{18}\text{O}$ values of Tzavoa, Soreq and Peqi'in speleothems	153
Figure 31: Comparison of the $\delta^{13}\text{C}$ profiles and age frequencies of central Negev speleothems with $\delta^{13}\text{C}$ values of Tzavoa, Soreq and Peqi'in speleothems.	154
Figure 32: Correlation between the peaks of relative frequencies of the speleothem ages during the last 550 ka and the solar radiation energy	155
Figure 33: Relative frequencies of the speleothem ages compared with African monsoon index and timing of Mediterranean sapropels	156
Figure 34: Correlation between the relative frequencies of the speleothem ages during the NHP-1 and NHP-2 and the: $\delta^{18}\text{O}$ profiles of Soreq and Peqi'in Caves, monsoon index, parameters of solar radiation, and timing of Mediterranean sapropels	157
Figure 35: Correlation between the relative frequencies of the speleothem ages and sapropel events during the period between 550 ka and 250 ka with African monsoon index values and solar radiation	158
Figure 36: Location of the caves where Sr isotopic composition of speleothems was studied	159
Figure 37: $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of Ma'ale-Efrayim Cave speleothems, Jordan Valley and Negev Desert speleothems compared with those of the speleothems of Soreq and Jerusalem caves	160
Figure 38: Comparison between Sr concentrations in central Israel caves (Soreq and Jerusalem) and caves of northern Negev (Tzavoa and Ma'ale-Dragot), Jordan Valley (Ma'ale-Efrayim), and central and southern Negev	161
Figure 39: $^{87}\text{Sr}/^{86}\text{Sr}$ vs $1/\text{Sr}$ plot in speleothems, host rocks and bulk soils	162

Figure 40: Annual dust fall (gr/m²) map of Ganor and Foner (1996, 2001) and the location of the studied caves and the caves of central Israel 163

Appendix 2: The studied caves 164

Appendix 3: Analytical procedures of U-Th and U-Pb dating methods 169

A.3.1. U-Th dating: chemical separation between U and Th 169

A.3.2. U-Th dating: measurement of U and Th isotopic ratios by MC-ICP-MS 170

A.3.3 U-Pb dating method 171

A3.3.1 Chemical purification and extraction of U and Pb 171

A3.3.2 Measurement using MC-ICP-MS 172

A3.3.3 Isochrone formalization 173

A3.3.4 Correction for U series disequilibrium 173

References of Appendix 3 174

Appendix 4: Tables of results 177

Table 1: U-Pb and U-Th ages of speleothems 178

Table 2: δD and $\delta^{18}O$ values of present-day rainfall in the Negev Desert 183

Table 3: δD values and calculated $\delta^{18}O$ values of the Negev Desert speleothem fluid inclusions 186

Table 4: Table 4: Sr concentrations and $^{87}Sr/^{86}Sr$ ratios: speleothems, host rocks, bulk soils and their silicate fractions in the Jordan Valley and the Negev Desert. 188

1. Introduction

The Saharan-Arabian Desert (Fig 1A)¹ is the largest and the most arid desert in the world, but its history is punctuated by many humid periods and the formation of this desert is connected with the glaciation of the northern hemisphere at about 2.7 Ma were (deMenocal, 1995; 2004). Numerous studies have shown that the paleoclimate of southern and central part of Saharan-Arabian Desert is strongly related to global climate changes (Street and Grove, 1979; Miller et al., 1991; Szabo et al., 1995; Crombie et al., 1997; Rohling et al., 2002; Fleitmann et al., 2003; Almogi-Labin et al., 2004; Osmond and Dabous, 2004; Drake and Bristow, 2006). Middle-Late Quaternary monsoon activity increased during periods of highest northern hemisphere insolation, mainly associated with precession (19-23 ka) and obliquity (42 ka) cycles of Earth's orbit, bringing about wet conditions and the formation of mega-lakes and rivers in the southern and middle parts of present day hyper-arid Saharan-Arabian Desert. Derricourt (2005) suggested that humid phases in the Saharan Arabian Desert may have controlled the dispersal of hominids from the tropical Africa to the others parts of the world, via the Sinai-Negev land bridge.

The northern part of the Saharan-Arabian Desert mainly receives its precipitation from mid-latitude cyclones, but the data on its paleoclimate and the relationship with the global climate change is still fragmented and controversial. Evidence from Tunisia shows that rainfall increased during Holocene African Humid period, resulting in the formation of large lakes (Causse et al., 2003). Evidence for wetter conditions during the early Holocene in the eastern parts of Saharo-Arabian Desert was given by the $\delta^{13}\text{C}$ values of the organic matter of the land snails in the northern Negev Desert, southern Israel (Magaritz and Goodfriend, 1987; Goodfriend, 1990; 1991; 1999). Evidence for clustering of flood events during glacial and late Holocene is given by the presence of tufa and slack water deposits in caves near the wadi channels (Kuperman, 2005; Greenbaum et al.,

¹ All figures in this work are given in Appendix 1

2006). However, no evidence of Holocene humidity has been found in the deserts of northern Egypt or the southern Jordan (Petit-Maire et al., 2002; Brook et al., 2002). Older humid periods at 8-15 ka, 25-60 ka, ~75-145 ka, ~150 ka, ~170-250 ka and ≥ 280 ka, have been interpreted from studies on: soil carbonate nodules (Goodfriend and Magaritz, 1988), spring tufa (Schwarcz et al., 1979; Livnat and Kronfeld, 1985; 1990; Enmar, 1999), soils on alluvial terraces (Plakht, 1995; 2000), lacustrine sediments from southern Israel (Avni, 1997; Avni et al., 2001-a; 2001-b; Ginat, 1997; Ginat et al., 2003) and southern Jordan (Petit-Maire et al., 2002) and speleothems from Djara Cave in Egypt (Brook et al., 2002). However, the dating precision of these humid periods is far from being accurate.

In the present work, carbonate cave deposits (speleothems) from caves of Israeli deserts were used to reconstruct the paleoclimate conditions of the northern boundary of Saharan-Arabian Desert. The Israeli deserts can be classified in two types: the Negev Desert in southern Israel, which is part of the Saharan-Arabian desert belt; and the "rain shadow" desert of eastern Israel, located to the east of Central Mountain Ridge (CMR) along the Dead-Sea Transform in the rift valleys of the Jordan River and Dead-Sea, (Fig. 1B). These regions are ideal for paleoclimate research because of: a) the very sharp precipitation gradients between the Mediterranean climate zone (>350 mm of rainfall/year, with cool rainy winters and hot and dry summers), the mildly arid steppe and semi-desert zone (350-150 mm), and the arid to hyper arid desert (<150 mm) (Fig. 1C); and b) the presence of numerous caves containing speleothems in the three climate zones. Speleothems grow in caves when water reaches the unsaturated zone and vegetation is present on the surface to supply the CO_2 necessary for limestone dissolution (Hendy, 1971; Schwarcz, 1986), but they do not grow in water and soil- CO_2 depleted conditions of arid and hyper-arid deserts (Holmgren et al., 1995; Fleitmann et al., 2003). At present, the caves in the semi-desert and desert regions of Israel are dry with no speleothem formation. However, the presence of speleothems in these caves indicates

that wet conditions prevailed in the past. The focus of this study is to reconstruct (1) the timing of the speleothem growth, i.e., reconstruction of the periods of increased precipitation in present-day desert by precise dating using U-Th method, and (2) the origin of precipitation, eolian dust sources, and vegetation above the caves, using the oxygen, carbon and Sr isotopic composition of the speleothems, and the hydrogen isotopic composition of their fluid inclusions.

2. Geography, climate and geology of the Israeli deserts

2.1 Geographical definition and climatic zoning

The Negev Desert (Fig. 1B, C) occupies the southern half of Israel between latitudes of 31°30'N - 29°30'N and longitudes 34°30'E - 35°30'E. Its eastern geographical boundary is marked by the Dead Sea Transform (Dead Sea - Arava Valley), the western geographical boundary by the Israeli-Egyptian border, and the southern limit is the Gulf of Elat (Gulf of Aqaba). The northern boundary is defined by the Mediterranean Sea coast of Gaza Strip and by a line ~20 km to the north of Be'er-Sheva Arad Valley. The Be'er-Sheva Arad Valley dissects southern Israel from west to the east along latitude ~31°15'N, and separates between the CMR to the north, and the highlands of the central Negev to the south. The CMR creates strong "rain shadow" effect in Dead Sea and Jordan Valleys, because of their depressed position relatively to westerly winds. Thus, a 15-30 km strip of desert and semi-desert continues to the north along the Dead Sea Transform till ~32°20'N: the arid to hyper-arid Judea Desert near the Dead Sea and the mildly-arid semi-desert zone in the Jordan Valley (Fig. 1C).

Conventionally, the Negev Desert is divided to three geographical regions (Fig. 1C): the northern Negev to the north of Wadi Zin, the central Negev from Wadi Zin to Wadi Paran, and the southern Negev from the Wadi Paran to the Gulf of Elat. In this study the division of the Negev desert is different and based mainly on climate and vegetation zoning as follows:

1) The mildly arid *northern Negev Desert*, bordering the south-east corner of the Eastern Mediterranean (EM) Sea, forms a ~40 km wide belt that includes the coastal plain near Gaza, the southern edge of the CMR, and a few smaller ridges to the east near Arad. The northern border of the area is defined by the 350 mm isohyet (the southern boundary of Mediterranean climate zone). The Be'er-Sheva Arad Valley dissects the area from west to east and the southern border of the northern Negev is defined by the ~150 mm isohyet along the southern margins of this valley. Vegetation in this region changes southward from C3 Mediterranean steppe forest to a mix of C3 and C4 semi-desert Irano-Turanian vegetation (Vogel, 1986; Cerling, 1993; Feinbrun-Dothan, 1998; Goodfriend, 1990; 1999).

2) The arid *central Negev Desert* or *Negev Highlands*, located south of Be'er-Sheva Arad Valley consists of several small NE-SW trending ridges with elevations of 500-1033m asl. Rainfall varies from 150 mm in the north to 50 mm in the south. The vegetation changes southward from semi-desert Irano-Turanian vegetation to Saharan-Arabian desert type, both comprising mixed C3 and C4 vegetation (Vogel, 1986; Cerling, 1993; Feinbrun-Dothan, 1998; Goodfriend, 1990; 1999).

3) The hyper-arid *southern Negev Desert*, located south of the Negev Highlands, presently receives ~30-50 mm average annual rainfall and is characterized by Saharan-Arabian desert flora (C3+C4) (Vogel, 1986). Whereas the northern parts of the Negev mainly receive winter rainfall associated with mid-latitude Atlantic-Mediterranean (Cyprus) cyclones, the southern Negev receives its rainfall mostly at the beginning (October-November) and end of the winter rainy season (March-May) in sporadic short storms, usually accompanied by local floods. Some of this rainfall is associated with synoptic systems that originate in the tropical Atlantic Ocean, pass over Africa and approach the region from the south-southwest (Kahana et al., 2002). These synoptic conditions infrequently occur also in the northern parts of the Negev.

The climate zoning in the rain shadow deserts includes: the mildly arid steppe and semi-desert of northern and central Jordan Valley receiving 350-150 mm annual rainfall, and the arid deserts in southern Jordan Valley and the Judea Desert near the Dead Sea receiving 150-50 mm annual rainfall. These regions mainly receive winter rainfall from Cyprus cyclones, although rare thunderstorms bringing tropical moisture occur at the beginning or at the end of the rainy season, similar to the southern Negev (Kahana et al., 2002). South of the Dead Sea in the Arava Valley, the climate is hyper-arid with less than 50 mm rainfall (Fig. 1B, C). The relative abundance of C4 vegetation is higher in these regions compared with the Negev Desert, owing to the higher temperatures (Vogel, 1986).

2.2. Rainfall gradients in the northern Negev Desert and Jordan/Dead Sea Valleys

At present the northern and central Negev Desert and the Jordan Valley region receive most of their rainfall during November to March from mid-latitude Cyprus cyclones moving eastwards above the EM Sea. The typical Cyprus cyclones (the major contributor of rainfall in Israel) correspond to the majority of long fetch maritime air masses crossing the Mediterranean (Dayan, 1986). Summers, between May and September, are hot and dry and result from the sinking air of subtropical highs, which develop over the Mediterranean Sea as strong high-pressure ridges pushed eastwards from the Azores subtropical high.

The Mediterranean Sea is the snow/rainfall moisture source and supplies the latent heat from its warm waters to the air masses entering the southern Levant through the passage of the Cyprus Lows. When passing over the land, the supply of moisture and latent heat becomes dramatically reduced to the south, and more gradually to the east (Shay-El and Alpert, 1991; Enzel et al., 2007). Due to the orographic increase in precipitation above CMR and adiabatic heating of descending air masses on its eastern flanks, the rainfall decreases by 400-500 mm over ~25 km from the CMR crest to the

Jordan and Dead Sea Valleys, bringing about the formation of the "rain shadow" desert. Consequently, isohyets run from north to south in northern and central Israel parallel to the coastline and CMR, but then abruptly change their orientation and run to the west parallel to the northern coastline of the Sinai Peninsula. In the Negev Desert the precipitation decrease from north to the south becomes very sharp; even though there isn't any significant topographic barrier in the 50 km belt inland of the coastline (Fig. 1B, C). This sharp precipitation gradient occurs in the zone where the Cyprus cyclones cross the Levant from the west to the east, because the northern Sinai coastline forms the southern limit at which rain clouds can form (Zangvil and Druian, 1990). Consequently, the rainfall amount north of Be'er-Sheva Arad Valley is much higher than to its south (Fig. 1C).

2.3 Geology

During the Mesozoic and Cenozoic epochs Israel was a marginal region between the land and sea. Thus, most of exposed sediments in the Negev are of Cretaceous-Pleistocene age, and are composed of limestones, dolomites, chalks and marls of marine origin, with fewer exposures of sedimentary rocks of terrestrial origin, such as sandstones, conglomerates, lacustrine marls, and loess (Fig. 2A, B). Triassic – Early - Cretaceous sandstones, dolomites, limestone and gypsum are exposed in erosional craters and deep canyons, and the deepest Pre-Cambrian crystalline basement rocks and Cambrian sandstones and dolomites are exposed in Elat Mountains in southern Negev.

Karstic caves are mainly associated with dolomites and limestones of Cenomanian and Turonian Judea Group, with few caves in Eocene limestone rocks (Appendix 2). These marine sediments are usually exposed at the ridges. In the valleys they are locally overlain by sandstones, conglomerates, and lacustrine marls of Miocene Hazeva Group, by lacustrine marls and conglomerates of Pliocene Arava Formation and Early Pleistocene Zhiha Formation, and by loess deposits and alluvium of Middle-Late

Pleistocene. Last major period of loess deposition in the northern and central Negev occurred between 90 ka and 14 ka (Crouvi et al., 2007). Partial erosion of the loess occurred during the Holocene (Avni et al., 2006).

The present landscape of the Negev resulted from the two major geological processes:

1) The formation of the Syrian Arc, which started during the Senonian period about 70 Ma ago, and probably was associated with the plate movements during the beginning of the Alpine orogenic processes in the north. The NW-SE convergence stress re-activated Jurassic normal faults in a reverse direction, causing the formation of anticline chains extending from northern Sinai through the Negev and central Israel to Syria. The folding of the anticlines was syn-sedimentary, with few pulses continuing until the Neogene (Garfunkel, 1982; Garfunkel, 1988-a; Solomon and Garfunkel, 1989; Calvo, 2002). As a result of these processes, thick sequence of marine chalks, marls, cherts and phosphates were deposited in the synclines, and a thin sequence at the tops of the anticlines (Soudry et al., 1985). Senonian-Paleocene soft sediments overlying the hard dolomites and limestones of Judea Group, known as Mount Scopus Group, contain a high concentration of uranium (~120 ppm) (Soudry et al., 2002). The anticlines of the Syrian Arc form the present-day SW-NE ridges of the central Negev, small ridges to the east of Be'er-Sheva-Arad Valley and the CMR.

2) The formation and evolution of the Dead Sea Transform during the Neogene and Quaternary. Following a major marine regression at the end of the Eocene, the southern Levant was relatively flat until the middle Miocene, when the region started to experience left lateral strike slip movement. During the late Miocene – Early Pliocene, a first major tectonic phase shaped the present outlines of the long and narrow depression (rift) along the Dead Sea transform and controlled the uplifting and tilting of its shoulders, which included the Negev Desert (Picard, 1943; Garfunkel and Horowitz, 1966; Garfunkel, 1970; Garfunkel, 1978; Garfunkel, 1988-b; Steinitz and Yosef, 1991; Zilberman, 1991; Avni et al., 2001-a). A second major tectonic phase during the Early Pleistocene caused

the central Arava Valley to subside by 500-700 m relative to central Negev, and the western part of the southern Negev to be uplifted by ~200 m relative to central Sinai. The central and southern Negev were tilted 1-2° eastwards creating a new drainage system flowing towards Dead-Sea Valley (Avni, 1991; Zilberman, 1991; Avni, 1997; Avni et al., 2001-a). As a result, erosion removed most of the Tertiary and Senonian soft sediments, exposing the hard carbonate rocks of the Judea Group on the axes of the Syrian Arc anticlines.

Conglomerates and lacustrine marls of Pliocene Arava Formation were deposited in the southern Negev and the Arava Valley between the two major tectonic phases of the Dead Sea transform (Avni et al., 2001-a). Early Pleistocene conglomerates and lacustrine sediments of Zhiha Formation were deposited at the beginning of the second tectonic phase (Ginat, 1997; Ginat et al., 2003). Both these formations are associated with lakes.

The caves were most probably formed between Late Eocene and Miocene when the Judea Group was below groundwater table. Some caves (mainly those on the anticline crests) were formed in open aquifer conditions, whereas others formed further from the anticline crests in confined conditions below the aquitards of Mount Scopus marls (Frumkin and Fischhendler, 2005). The caves were uplifted above groundwater table during the two late major tectonic phases, thus enabling the formation of vadose speleothems.

2.4 Speleothems as paleoclimatic records

Vadose carbonate speleothems (stalactites, stalagmites and flowstone) form when meteoric waters react with soil CO₂, dissolve the carbonate rocks in the unsaturated zone, and deposit secondary CaCO₃ deposits in the caves as a result of CO₂ degassing. The process of speleothem formation depends on the presence of water in unsaturated zone, which in turn depends of the effective precipitation (i.e. precipitation – (evaporation + runoff)); and the presence of vegetation above the cave to supply the soil CO₂ necessary

for the dissolution of carbonate host rock (Hendy, 1971; Buchmann and Dreybrodt, 1985; Schwarcz, 1986; Ford and Williams, 1989). Availability of water is a particularly important factor controlling the speleothem growth rates in regions where climate fluctuates between arid and semi arid (Holmgren et al., 1995; Vaks 2001; Fleitmann et al., 2003). Chemical models also show that factors such as Ca concentrations in the seepage water, cave temperature, CO₂ pressure and water flow patterns influence speleothem growth rates (Baker and Smart, 1995; Dreybrodt et al., 1996).

Speleothems are excellent recorders of climate. They are protected from weathering and alteration; and can be dated precisely by the uranium-series disequilibrium methods, with a precision approaching $\pm 1\text{-}2\%$ (2σ), using Thermal Ionization Mass Spectrometry (TIMS) and Multi-Collector Inductively Coupled Plasma Mass Spectrometry (MC-ICP-MS) techniques. The high ionization efficiency of the MC-ICP-MS relative to TIMS significantly improves the ionization for thorium, which make this method particularly efficient for low-uranium young speleothems containing relatively little radiogenic ²³⁰Th (Robinson et al., 2002; Fleitmann et al., 2003). Speleothems usually contain low amounts of detrital material, which make their ages more reliable than carbonate sediments with higher detrital contents (i.e., tufa and travertine, lacustrine carbonates and soil carbonate nodules).

When deposited in isotopic equilibrium with the cave waters, the $\delta^{18}\text{O}$ values of speleothems and the δD values of their fluid inclusions depend on cave temperature and the $\delta^{18}\text{O}$ and δD values of the rainfall. The cave temperature usually approximates the mean annual surface temperature because it is buffered by large mass of rock (Hendy and Wilson, 1968; Schwarcz, 1986). Rainfall $\delta^{18}\text{O}$ and δD depend on the vapor source (the sea surface water $\delta^{18}\text{O}$ and δD), rainfall amounts, the latitude, altitude and distance from the source (Dansgaard, 1964; Frumkin et al., 1999; Bar-Matthews et al., 1997, 2003-a). Thus, speleothem $\delta^{18}\text{O}$ values and the fluid inclusion δD values are excellent proxies of the paleo-temperatures, source of water vapor, and rainfall amounts (Hendy and Wilson,

1968; Schwarcz, 1986; Bar-Matthews et al., 1996, 1997, 1999; Frumkin et al., 1999; Matthews et al., 2000; Dennis et al., 2001; Burns et al., 2003; McDermott, 2004; McGarry et al., 2004).

The independent chronology of speleothem data offers opportunities to critically assess leads and lags in the climate system (timing and duration of major O isotope-defined climatic events – e.g., Marine Isotope Stages (MIS) and smaller scale climatic changes), which in turn can provide important insights into global forcing and feedback mechanisms (McDermott, 2004). In this respect, speleothems can often offer advantages over many other paleoclimate records.

Speleothem $\delta^{13}\text{C}$ values depend on soil CO_2 and the carbonate host rock. Under closed system conditions the contribution of the two sources is approximately equal (Hendy, 1971; Bar-Matthews et al., 1996; McDermott, 2004). The $\delta^{13}\text{C}$ values of the marine carbonate host rock is considered to be constant (-1‰ - $+1\text{‰}$), while the $\delta^{13}\text{C}$ of soil CO_2 depends on the vegetation type above the cave. Organic carbon originating from C3 vegetation type (Calvin photosynthesis cycle) has $\delta^{13}\text{C}$ values ranging from -26‰ to -20‰ , whereas C4 vegetation type (Hatch-Slack cycle) has values from -16‰ to -10‰ . These differences are preserved as distinctive ranges of $\delta^{13}\text{C}$ in speleothem carbonates: from -14‰ to -6‰ for C3 plants and from -6‰ to $+2\text{‰}$ for C4 plants. However, speleothem $\delta^{13}\text{C}$ values can become higher due to deposition in partially open system conditions, resulting in a kinetic fractionation during rapid CO_2 degassing, and incomplete equilibration of water with soil CO_2 during periods of heavy rain because of short residence time of the water in the soil zone (Schwarcz, 1986; Bar-Matthews et al., 1996, 1997, 2000; Fleitmann et al., 2003; McDermott, 2004).

Trace elements such as Sr and U are incorporated into the calcite lattice. Their isotopic ratios ($^{87}\text{Sr}/^{86}\text{Sr}$ and $^{234}\text{U}/^{238}\text{U}$) in speleothem calcite can provide information on the upper vadose zone water chemistry, water-soil-rock interactions, weathering and origin of

olian dust (Banner, 1995; Goede et al., 1998; Ayalon et al., 1999; Frumkin and Stein, 2004; Li et al., 2005).

Growth rate, growth periods and depositional hiatuses (growth breaks) are important indicators of the presence and amounts of water in the unsaturated zone, and are thus an indication of desertification processes (Holmgren et al., 1995; Vaks 2001; Fleitmann et al., 2003), or permafrost formation and melting in cold climates (Harmon et al., 1977; Schwarcz, 1986; Lauritzen, 1995). Laminae thickness, crystal morphology, mineralogy and petrography can be indicators of dripping rates, degree of super saturation of the precipitating water, fluid chemistry and the nature of the fluid transport (Bar-Matthews et al., 1991, 1997; Ayalon et al., 1999; Frisia et al., 2000; 2002). Large columnar crystals are usually indicative of slow and constantly dripping water where the speleothem is continuously wet and growing at near equilibrium saturation conditions with respect to the CaCO_3 . A fabric consisting of small crystals with a high density of crystal defects is indicative of deposition under similar low super saturation conditions, but with variable water discharge and presence of growth inhibitors. Dendritic fabrics with the highest density of crystal defects are indicative of speleothem deposition under high super saturation disequilibrium conditions and prolonged periods of low fluid flow regime that resulted in prolonged out gassing (Frisia et al., 2000).

2.5. Speleothem climate record from the Mediterranean climate zone of Israel

Accurate high resolution U-Th dating of speleothems from the Soreq, Jerusalem and Peqi'in caves, in central and northern Israel (Fig. 1B, C), show that speleothem deposition was continuous over several glacial-interglacial cycles, thus providing a powerful tool for understanding the EM paleoclimate (Ayalon et al., 1999; 2004; Bar-Matthews et al., 1997, 1999, 2000, 2003-a; Frumkin et al., 1999, 2000). The $\delta^{18}\text{O}$ isotopic profile of the EM speleothems shows a good correspondence with the stacked oxygen isotopes (Martinson et al., 1987) and with the marine record of the EM Sea, suggesting that their

isotopic composition reflects global and regional climatic changes, and that climatic events on sea and land were linked (Frumkin et al., 1999; Bar-Matthews et al., 2000, 2003-a).

Periods of low $\delta^{18}\text{O}$ values usually coincide with low $\delta^{18}\text{O}$ values of the planktonic foraminifera *G. ruber* (i.e., sea surface water), with the times of increased atmospheric CO_2 , warming, sea level rise and the formation of Mediterranean sapropels. Sapropels are black, organic rich layers found in sediments throughout the Mediterranean Sea. They are interspersed within the normal, whitish, oxidized hemi pelagic sediments and reflect oxygen deficient conditions in the deep-sea due to stratification of the water column resulting from a high input of the fresh water. Periods of sapropel formation in the EM Sea are associated with the minimum speleothem $\delta^{18}\text{O}$ both during interglacials (Early Holocene, MIS-5.1, 5.3, 5.5, 7.1, 7.3, 7.5), but also during glacial interstadial MIS-6.4 (Bar-Matthews et al., 2003-a; Ayalon et al., 2004), suggesting increased discharge of freshwater to the Mediterranean Sea via the Nile (Olausson, 1961; Ryan, 1972; Rossignol-Strick and Paterne, 1999), via other rivers from present-day Sahara Desert region (Rohling et al., 2002) and via increase in rainfall over the entire Mediterranean basin (Kallel et al., 2000). The highest speleothem $\delta^{18}\text{O}$ values coincide with higher $\delta^{18}\text{O}$ values of the *G. ruber* (i.e. colder EM Sea surface temperatures), corresponding to extreme global cooling episodes, such as Heinrich events and the Last Glacial Maximum (LGM) (Ayalon et al., 2002; Bar-Matthews et al., 1999, 2000, 2003-a). The strong links between sea and land $\delta^{18}\text{O}$ records shows that the $\delta^{18}\text{O}$ of the rainfall above the caves was mainly controlled by the EM Sea precipitation source (Frumkin et al., 1999; Bar-Matthews et al., 2003-a; Kolodny et al., 2005). Superimposed on this, rainfall amount is important control on the $\delta^{18}\text{O}$ of precipitation and therefore on the $\delta^{18}\text{O}$ of the speleothems (Bar-Matthews et al., 1997; 2003-a). The influence of the “amount effect” (as originally defined by Dansgaard (1964)) on speleothem $\delta^{18}\text{O}$ in EM Sea region is still under debate. Frumkin et al. (1999, 2000) and Kolodny et al. (2005) argue for a

negligible contribution whereas Bar-Matthews et al., (2000; 2003-a) argue that periods with very little $\delta^{18}\text{O}$ change in EM Sea surface water, but larger contemporaneous $\delta^{18}\text{O}$ changes in speleothems, reflect the amount effect.

The EM Sea as a precipitation source is also evident by δD - $\delta^{18}\text{O}$ relations based on δD values of fluid inclusions (FI) trapped in EM speleothems. Present-day precipitation originating in the EM Sea is characterized by d-excess values ($\delta\text{D}-8\times\delta^{18}\text{O}$) of $\sim+22\text{‰}$, defining the local Mediterranean Meteoric Water Line (MMWL) ($\delta\text{D}=8\times\delta^{18}\text{O}+22$) (Gat and Carmi, 1987; Ayalon et al., 1998), which differs from the global Meteoric Water Line (MWL) relationship ($\delta\text{D}=8\times\delta^{18}\text{O}+10$) (Craig, 1961). The δD - $\delta^{18}\text{O}$ relationships of paleo cave waters determined from fluid inclusions show that interglacial waters follow the MMWL, but most glacial waters are slightly offset from the MMWL, becoming closer to the MWL during the last glacial maximum (LGM) (Matthews et al., 2000; McGarry et al., 2004). Temperature changes on land and sea are very similar as evident from alkenone sea surface temperatures from EM sediments (Emeis et al., 2000; Bar-Matthews et al., 2003-b; Kolosovsky, 2003) and calculated land temperatures inferred from the δD of trapped fluid inclusions (McGarry et al., 2004).

Speleothem $\delta^{13}\text{C}$ values in central and northern Israel are generally in the range from -9‰ to -13‰ and are indicative of the dominance of C3 type Mediterranean vegetation during interglacial periods. During global cooling episodes like the glacial maxima, $\delta^{13}\text{C}$ values are higher ($\sim -7\text{‰}$ - -9‰), indicative of higher contribution of C4 type vegetation and drier conditions (Bar-Matthews et al., 1997, 1999, 2000, 2003-a). During interglacial intervals with low $\delta^{18}\text{O}$ minima at 127-121 ka and 9-7 ka $\delta^{13}\text{C}$ values are anomalously high. Bar-Matthews et al. (1997, 2000, 2003-a) suggested by them that these high values represent deluge episodes that stripped the soil cover, and/or the rainwater residence time in the soil zone was too short to equilibrate with the soil CO_2 . Frumkin et al., (2000) on the other hand had suggested that these periods of high $\delta^{13}\text{C}$ are indicative of dry conditions.

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of speleothems is a proxy for paleo-dust input and weathering conditions of the cave host rock. Ayalon et al., (1999) and Bar-Matthews et al., (2000-b) found that concentrations of Sr and U and $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{234}\text{U}/^{238}\text{U}$ ratios in Soreq Cave speleothems are lower (less radiogenic) during interglacials, and interpreted this to indicate wetter conditions causing enhanced leaching of soil and rock. They also found a higher values of these parameters during the last glacial period, attributing it to an increase in the contribution of salts derived from exogenic sources, (sea spray and aeolian dust), with less leaching of the host rock, owing to drier conditions. Frumkin and Stein (2004) found the similar patterns in Jerusalem Cave speleothems, and suggested that Sr and U in speleothems were mainly derived from the soils above the caves. The increase in $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{234}\text{U}/^{238}\text{U}$ ratios of the Jerusalem cave speleothems were interpreted by Frumkin and Stein (2004) to represent enhanced supply of Saharan dust during the glacial periods, due to the increased aridity of the Sahara Desert. The decrease of these values during interglacials was attributed to shut-down of Saharan dust source as a result of wetter conditions in the source regions and an increase in host rock contribution of Sr and U.

3. Research goals

3.1 The major aims of the study

The objective of this study is to determine using the speleothems the paleoclimate of the Israeli deserts: the Negev Desert, southern Israel, and the rain shadow deserts in the Jordan and Dead Sea Valleys, located between present day isohyets of 300 mm and 30 mm, in the following terms:

- 1) Reconstruction of the periodicity of wet and dry intervals in: (a) the mildly arid and arid zones of the Jordan and Dead Sea Valleys and northern Negev and (b) the arid and hyper arid zones of the central and southern Negev Desert. To use this data to determine: (c) how the northern boundary of Saharan-**

Arabian Desert has shifted in time and space and (d) how these shifts may have influenced the migration of hominids and animals between Africa and Asia via Sinai-Negev land bridge. These goals are mainly achieved by U-Th dating of speleothems.

- 2) Reconstruction of rain sources, rainfall amounts and paleo-vegetation type above the caves.** These goals were achieved by the determination of high resolution $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records of calcite laminae and the δD values of trapped fluid inclusions and comparison of these records with other marine and terrestrial paleoclimate proxies in the region, especially with the well-dated high-resolution records of speleothems in the Mediterranean climate area of Israel (Soreq, Jerusalem and Peqi'in caves).
- 3) Determination of the origins of speleothem Sr and dust provenance in the Negev Desert and Jordan Valley.** This was done by analysis of Sr concentrations and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in speleothems, host rocks and dust-born soils; the relationship between them and the speleothems in the Mediterranean climate region (Soreq and Jerusalem Caves).

3.2 Structure of the thesis

The thesis is composed of three published articles and one chapter. The research topics in each part are detailed below:

The first article is: “Vaks et al. (2003), Paleoclimate reconstruction based on the timing of speleothem growth, oxygen and carbon isotope composition from a cave located in the “rain shadow”, Israel., *Quaternary Research*, 59, 2, pp 182-193”. The aim of this study was a paleoclimate reconstruction in the mildly-arid semi-desert of the Jordan Valley, located in the “rain shadow” of CMR, using the speleothem record for Ma’ale-Efrayim Cave (#1, Fig. 1C, Appendix 2).

The second article is: “Vaks et al. (2006), Paleoclimate and location of the border between Mediterranean climate region and the Saharo-Arabian Desert as revealed by speleothems from the northern Negev Desert, Israel., *Earth and Planetary Science Letters*, 249, 3-4, pp. 384-399”. The aim of this study was the paleoclimate reconstruction in the mildly-arid semi-desert of the northern Negev using the speleothem record from Ma’ale-Dragot and Tzavoa Caves (#3 and #4 respectively, Fig. 1C, Appendix 2). The focus was mostly on the Tzavoa Cave, being the southernmost cave in this region containing thick speleothems (few tens of cm) younger than 550 ka (U-Th method limit).

The third article is: “Vaks et al. (2007), Desert speleothems reveal climatic window for African exodus of early modern humans, *Geology*, 35, 9, pp. 831-834”. This study reconstructs humid periods in arid and hyper arid central and southern Negev Desert during the last 180 ka using the speleothem record for five caves: Hol-Zakh, Ashalim, Even-Sid-Ramon, Ma’ale-ha-Meyshar, and Ktora Cracks (#6, #8, #9, #12, and #14 respectively, Fig. 1C, Appendix 2). The study examines the possible link between the desert humid periods and migration of hominids and animals out of Africa through the Saharan-Arabian Desert and Sinai-Negev land bridge.

The last chapter of the thesis is: “Paleoclimate of northern Saharan-Arabian Desert from speleothem record in central and southern Negev and Judea Desert, Israel”. This chapter deals with a detailed paleoclimate reconstruction in the central and southern Negev and Judea deserts, for longer time intervals (Pliocene – present) than were discussed in articles 1-3. The speleothem record of the following caves was used: Kanaim, Izzim, Hol-Zakh, Makhtesh-ha-Qatan, Ashalim, Even-Sid-Ramon, Wadi-Lotz, Ma’ale-ha-Meyshar, Shizafon and Ktora Cracks (#2, #5 - #9, #11 - #14 respectively, Fig. 1C, Appendix 2). The record of Ma’ale-Efrayim Cave, Jordan Valley was expanded for this study. This chapter also studies the origins of the speleothem Sr and the dust provenance using Sr isotopic compositions from the major studied caves.

4. Methodology

4.1 The studied caves and sampling

Fourteen Caves were studied: twelve caves from the Negev Desert, one cave in the Judea Desert near Dead Sea Valley and one cave from the Jordan Valley. Twelve caves are located within Cretaceous (Cenomanian and Turonian) host dolomite and limestone and one within Eocene limestone. The cave names and their numbers in Fig. 1C are: Ma'ale-Efrayim Cave (#1), Kanaïm Cave (#2) Ma'ale-Dragot Caves (#3), Tzavoa Cave (#4), Izzim Cave (#5), Hol-Zakh Cave (#6), Makhtesh-ha-Qatan Cave (#7), Ashalim Cave (#8), Even-Sid-Ramon mini-caves (#9), Mitzpe-Ramon Cave (#10), Wadi-Lotz Cave (#11), Ma'ale-ha-Meyshar Cave (#12), Shizafon mini-caves (#13), Ktora Cracks (#14). The coordinates and geological and geographical background of each cave are given in Appendix 2. Ma'ale-Efrayim Cave and Kanaïm Cave are located in the “rain shadow” desert: Ma'ale-Efrayim cave in mildly-arid semi-desert in Jordan Valley located between 250 mm and 300 mm isohyets, and Kanaïm Cave in arid Judea Desert located at ~100 mm isohyet. The twelve Negev caves are located along north-south climatic transect from ~ 300 mm isohyet in the north to ~30 mm isohyet in the south (Fig. 1C and in Appendix 2).

123 in-situ or broken vadose speleothems (flowstones, stalactites and stalagmites) and phreatic speleothems were collected from the fourteen caves. The speleothems vary in size from few mm thick flowstone crusts to 60 cm length 20 cm width stalagmites, stalactites, and flowstones. The samples were sectioned using a diamond saw to expose their internal structure and to eliminate diagenetically altered samples (Bar-Matthews et al., 1997). The mineralogy and petrography was determined using petrographic microscope, Jeol 840 scanning electron microscope equipped with Oxford ISIS EDS system, and Philips PW 3020 X-ray diffractometer. Preliminary petrographical studies and U-Th dating using alpha-spectrometry indicated that the speleothems older than the U-Th method limit were composed of large columnar crystals a few cm in length,

forming massive layers ranging from a few cm to 40 cm thick, whereas younger speleothems were thinly laminated. Based on their petrography sixty five speleothems including thinly laminated sections from the thirteen caves (all caves except Mitzpe-Ramon Cave) were chosen for high precision U-Th dating: eight speleothems from Ma'ale-Efrayim Cave and other small caves in its surroundings (Vaks et al., 2003); five speleothems from Kanaim Cave, Judea Desert; six speleothems from Ma'ale-Dragot Caves and fifteen speleothems from Tzavoa Cave, northern Negev (Vaks et al., 2006), and thirty one speleothems from other ten caves of central and southern Negev Desert (Vaks et al. (2007) and , Table 1)².

4.2 U-Th and U-Pb dating

37 speleothem samples from Ma'ale-Efrayim Cave were dated using thermal ionization mass spectrometers (TIMS) at the Open University, Milton Keynes, UK, and six other samples from this cave were dated by alpha-spectrometry in Geological Survey of Israel (Vaks et al., 2003) using methods described in Bar-Matthews et al. (1997), Kaufman et al. (1998), McDermott et al. (1999) and Vaks (2001). Most of the U-Th dating (295 speleothem samples from thirteen caves – (Vaks et al., 2006) and Table 1) was performed using the MC-ICP-MS in the Geological Survey of Israel.

For dating purposes up to 1 g material was drilled using 0.8 - 4 mm diameter drill bits along the growth axis for stalagmites (Fig. 2A), and across the growth axis or cross-section for stalactites and flowstone (Fig. 2B). A few laminae were sampled two times or more by drilling in few different locations along the same lamina (these samples with duplicate or triplicate dating are designated by the roman numerals I and II in Table 1). Depending on the uranium concentration, 10-800 mg calcite powder was dissolved in 7N HNO₃ (Table 1). The sample was loaded onto mini-columns contained 2ml Bio-Rad AG 1X8 200-400 mesh resin. U was eluted by 1N HBr and Th with 6N HCl. The U and Th

² All tables in this work are given in Appendix 4

solutions were evaporated to dryness and dissolved in 2ml and 5ml of 0.1N HNO₃ respectively (McDermott, 2001; Kuperman, 2005).

High precision U-Th dating was performed using a Nu Instruments MC-ICP-MS, equipped with 12 Faraday cups and 3 ion counters. The sample was introduced to the MC-ICP-MS through an Aridus[®] micro-concentric desolvating nebuliser sample introduction system. The instrumental mass bias was corrected using an exponential equation by measuring the ²³⁵U/²³⁸U atomic ratio and correcting with the natural ²³⁵U/²³⁸U ratio (0.0072). The calibration of ion-counters relatively to Faraday cups was performed using several cycles of measurement with different collector configurations in each particular analysis. The age determination was possible due to the accurate determination of ²³⁴U and ²³⁰Th concentrations by isotope dilution analysis using the ²³⁶U-²²⁹Th spike. U-Th ages were corrected for detrital ²³⁰Th (Kaufman et al., 1998), assuming a ²³²Th/²³⁸U isotope atomic ratio of 3.8 (the mean crustal value) in the detrital components. However, less than 10% of the samples had a ²³⁰Th/²³²Th activity ratio less than 30 and needed this correction (Table 1). The reproducibility of ²³⁴U/²³⁸U ratio was 0.11% (2σ). For a detailed description of the chemical preparation of the samples and the dating by MC-ICP-MS see the Appendix 3 and Kuperman (2005).

Speleothem deposition in the arid zone of Israel was intermittent and depositional hiatuses with distinct petrographic characteristics separated the different calcite laminae (Chapter 8.4.2, Vaks et al., (2003, 2006, 2007)). Each calcite lamina represents a growth period. In order to determine the duration of the speleothem growth periods, the top and base of growth laminae were dated. Where possible, several age determinations were performed across the lamina cross section. This enabled calculation of the growth rates during the periods of speleothem deposition, with the assumption that the age represents the center of the drilled area, and that there is a constant growth rate between two dated points. When growth laminae were very thin it was possible to obtain only one U-Th age. In such situations, or if the ages at the top and at the bottom of the lamina were within

each other's 2σ uncertainty range, it was assumed that the entire growth period was not longer than 1000 years.

Five samples of speleothems older than the U-Th dating limit, ASH-15-C, ASH-15-D and ASH-15-E from Ashalim Cave and KTO(1)-J and KTO(1)-1-K from Ktora Cracks, were dated using U-Pb method in University of Melbourne, Australia, according to the method described in Woodhead et al., (2006), and summarized Appendix 3, section A3.3. The sample ASH-15-D+E from Ashalim Cave (including both the ASH-15-D and ASH-15-E in one piece), was also dated using U-Pb method in University of Leeds, UK, according to the method of (Walker et al., 2007).

4.3 Stable isotope analyses of speleothems and present day rainwater

For $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ analyses, samples of 1-2 mg material were drilled using a 0.8-1 mm diameter drill, either along or across the growth axis. Hendy tests (55 measurements of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) were performed on several stalagmites and flowstone samples from Ma'ale-Efrayim and Tzavoa Caves, Ashalim Cave, Hol-Zakh Cave and Ma'ale-ha-Meyshar Cave to ensure that speleothem calcite deposition occurred in isotopic equilibrium with the dripping water (Hendy, 1971). These tests were most important in Tzavoa and Ashalim Caves because these caves have natural openings. At all locations, the speleothems were indicated to have deposited in equilibrium. More than a thousand measurements of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ were performed in high resolution on 5 stalagmites, 7 flowstone samples and 9 stalactites from Ma'ale-Efrayim, Tzavoa, Hol-Zakh, Ashalim, Even-Sid-Ramon, Ma'ale-ha-Meyshar and Shizafon caves using VG SIRA-II Mass Spectrometer with ISOCARB system for carbonate analysis (Bar-Matthews et al., 1997). $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of calcite are reported in per-mil (‰) relative to the PDB standard.

$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profiles were measured in speleothems having dateable laminae thicker than 4 mm, which enabled sampling of at least 3-4 samples across the lamina. The isotope profiles together with the U-Th ages enabled wiggle-matching between the $\delta^{18}\text{O}$

profiles of different speleothems that grew in the same time period (from the same cave or the adjacent caves) within 2σ dating uncertainty of each sample.

Data on the present day rainfall was obtained by collecting rainwater samples during two rainy seasons, winters of 1998-99 and 1999-2000 (September to May) in village of Ma'ale-Efrayim, Jordan Valley; and winters of 2004-5 and 2005-6, from the five stations located on a north-south transect of the Negev Desert (Fig. 1C and Table 2):

- 1) The city of Be'er-Sheva, northern Negev, 380 m asl, and ~40 km from the EM Sea coast;
- 2) The town of Arad, northern Negev, 600 m asl, and ~80 km from the EM Sea coast;
- 3) Makhtesh-ha-Qatan, on the western flank of the Dead-Sea Valley and in rain shadow of the hills to the west, 1 km from the cave 3, ~ 0 m asl (on sea level), and 98 km from the EM coast;
- 4) The town of Mitzpe-Ramon in the highlands of the central Negev, 2 km from cave #5; 820 m asl, ~100 km from the EM Sea coast);
- 5) The village of Neot-Smadar, at the southern Negev, located between the caves 9 and 10; 400 m asl, ~165 km from the EM coast).

In order to estimate $\delta^{18}\text{O}$ and δD of meteoric water in southern Negev during the years preceding the program of rainfall sampling, the water of Ein-Netafim spring near Eilat was sampled in June 2004. This spring water represents the mean $\delta^{18}\text{O}$ and δD values of major rainfall events during the preceding years.

Each rain event was sampled in Ma'ale-Efrayim, Be'er-Sheva and in Neot-Smadar during the first year, whereas at the other sites the rain was allowed to accumulate in plastic bucket with a 1 cm thick oil layer added to prevent evaporation. Rainwater that accumulated below the oil was sampled at 20-45 days intervals and transferred to sealed plastic bottles. $\delta^{18}\text{O}$ and δD measurements of the cave and rain water were performed using methods described in Bar-Matthews et al. (1996) and Ayalon et al. (1998), and are reported in the SMOW scale.

4.4 D/H analyses of speleothem fluid inclusions

The thermal decrepitation technique was used for extraction of the fluid inclusions (FI) water from the calcite. The extracted water was reduced to H₂. The measurement of δD followed the procedures of Yonge (1982), modified by Matthews et al. (2000) and McGarry et al. (2004). This technique has been found most applicable to the speleothems containing small amounts of water (McGarry et al., 2004). Nine speleothems, 2 stalagmites, 2 stalactites and 5 flowstone samples from Tzavoa, Hol-Zakh, Ashalim and Ma'ale-Ha-Meyshar caves were studied for D/H analyses of FI. Speleothem FI δD of Ma'ale-Efrayim Cave were studied by McGarry et al. (2004). δD analyses were made on 32 samples, each representing a growth lamina from a particular speleothem. The laminae were chosen using the two following criteria: 1) they represent the main depositional periods in northern, central and southern Negev Desert based on U-Th dating; and 2) they are composed of pure calcite with typical columnar fabric with no evidence of alteration and contain very low detrital material such as clay, to avoid extraction of water of non-calcitic origin.

The thermal decrepitation technique involves an isotopic fractionation accompanying the extraction of calcite bound water. This fractionation needs to be calibrated. For Soreq Cave speleothems Matthews et al. (2000) and McGarry et al. (2004) calibrated for the fractionation: $(\delta D_{FI\ water} = \delta D_{extracted\ water} - \delta D_{cave\ water}) = -30\text{‰}$. In the present study it was found that vacuum line blank has increased relatively to the previous studies. The contribution of the vacuum line blank was estimated by the analysis of samples with different water amounts giving the following relation, which was used to correct the results in this study:

$$(1) \delta D_{FI\ water} = \delta D_{extracted\ water} - (6.6 \times V_S - 40)$$

where V_S amount of the extracted water in ml measured by the mass spectrometer.

4.5 Sr isotope analyses

Strontium concentrations and isotopic ratios were studied on 115 speleothem samples from the following nine caves: Ma'ale-Efrayim, Ma'ale-Dragot, Tzavoa, Hol-Zakh, Makhtesh-ha-Qatan, Ashalim, Even-Sid-Ramon, Ma'ale-ha-Meyshar, Shizafon and Ktora Cracks. In order to determine the Sr contents and isotopic composition of soil and rock, which are the main contributors of Sr to the speleothems (Ayalon et al., 1999; Bar-Matthews et al., 1999; Frumkin and Stein, 2004), 9 samples of Cenomanian, Turonian, Senonian and Eocene rocks adjacent to the caves (Fig. 2) were analyzed as well as 11 samples of the soils above the caves. 3 samples of paleosols washed to the Ma'ale-ha-Meyshar and Even-Sid-Ramon caves (See section 8.2.7) were also analyzed. Strontium contents of samples were measured using Perkin Elmer Optima 3300 ICP-AES. For speleothem and host rock samples, 100 mg of the powdered calcite was dissolved in 3N HNO₃. For silicate soils, 0.25 g samples of the sieved < 60 µm fraction dried in 110°C, were reacted with 1.25 g LiBO₃ in 900°C in platinum-gold crucibles. The glass formed after cooling was dissolved in 4% HNO₃ and 500 ml of distilled water was added. Sc was added as an internal standard prior to chemical analysis of samples.

In order to separate silicate fraction from the carbonates in the soil samples, 10-40 mg of sample containing 1-2 µg Sr was dissolved in 1.25M HNO₃ for 24 hours in order to dissolve the carbonate fraction, then centrifuged to deposit the silicates. The aliquots were removed and remaining material was washed in double distilled H₂O to remove the remains of aliquots, and evaporated to dryness. For extraction of Sr from the silicates 5 ml HF+1ml HNO₃ was added to 0.1-0.5 g of the sample (according to the Sr content) and evaporated to dryness. The residues were dissolved in 3.5M HNO₃, and the solution was then loaded on a column containing the resin Eichrom-Sr-spec (50-100 mesh), and the matrix was rinsed away with 3 portions of 1 ml and one portion of 0.5 ml of 3.5M HNO₃. Strontium was eluted from the column by 3X1ml 0.05M HNO₃.

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios were measured using the MC-ICP-MS following the procedure of (Ehrlich et al., 2001). Five Faraday collectors were used for measurements of masses 83 and 85-88 for isotopes of Kr (83), Rb (85) and Sr (86, 87 and 88). ^{83}Kr was measured in order to correct for ^{86}Kr interference ($^{86}\text{Kr}=1.52\times^{83}\text{Kr}$), due to the Kr contamination of the argon gas. ^{85}Rb was measured in order to correct for ^{87}Rb interference. ^{87}Rb is calculated from the ratio $^{87}\text{Rb}=0.3860\times^{85}\text{Rb}$, after mass discrimination correction with the exponential law. The correction factor is calculated from the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, employing the exponential law mass bias correction, using the $^{87}\text{Sr}/^{86}\text{Sr}$ natural ratio of 0.1194.



ACADEMIC
PRESS

Available online at www.sciencedirect.com

SCIENCE @ DIRECT®

Quaternary Research 59 (2003) 182–193

QUATERNARY
RESEARCH

www.elsevier.com/locate/yqres

Paleoclimate reconstruction based on the timing of speleothem growth and oxygen and carbon isotope composition in a cave located in the rain shadow in Israel

Anton Vaks,^a Miryam Bar-Matthews,^{a,*} Avner Ayalon,^a Bettina Schilman,^a Mabs Gilmour,^b Chris J. Hawkesworth,^c Amos Frumkin,^d Aaron Kaufman,^e and Alan Matthews^f

^a Geological Survey of Israel, 30 Malchei Israel Street, Jerusalem 95501, Israel

^b Department of Earth Sciences, The Open University, Walton Hall, Milton Keynes, MK7 6AA, UK

^c Department of Earth Sciences, The University of Bristol, Bristol, UK

^d Department of Physical Geography, The Hebrew University of Jerusalem, Jerusalem 91905, Israel

^e Department of Environmental Sciences, The Weizmann Institute of Science, Rehovot 76100, Israel

^f Department of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem 91905, Israel

Received 29 July 2002

Abstract

High-resolution $^{230}\text{Th}/^{234}\text{U}$ ages and $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ compositions of speleothems in Ma'ale Efrayim Cave located to the east of the central mountain ridge of Israel enable us to examine the nature of the rain shadow aridity during glacial and interglacial intervals. Speleothem growth occurred during marine glacial isotopic periods, with no growth during the two last marine isotope interglacial intervals and during the peak of the Last Glacial Maximum. This contrasts with speleothem growth in caves located on the western flank of the central mountain ridge, in the Eastern Mediterranean semiarid climatic zone, which continued throughout the last 240,000 yr. Thus, during glacial periods water reached both sides of the central mountain ridge. A comparison of the present-day rain and cave water isotopic compositions and amounts at the Ma'ale Efrayim Cave site with those on the western flank shows that evaporation and higher temperatures on the eastern flank are major influences on isotopic composition and the lack of rainfall. The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profiles of the speleothems deposited between 67,000 and 25,000 yr B.P. match the general trends of the isotopic profiles of Soreq Cave speleothems, suggesting a similar source (eastern Mediterranean Sea) and similar climatic conditions. Thus, during glacial periods the desert boundary effectively migrated further south or east from its present-day location on the eastern flank, whereas interglacial periods appear to have been similar to the present, with the desert boundary at the same position. The decrease in overall temperature and a consequent reduction in the evaporation to precipitation ratios on the eastern flank are viewed as the major factors controlling the decay of the rain shadow effect during glacial periods.

© 2003 Elsevier Science (USA). All rights reserved.

Keywords: Speleothems; $\delta^{18}\text{O}$; $\delta^{13}\text{C}$; Paleoclimate; Eastern Mediterranean; Rain shadow; Effective precipitation

Introduction

Israel is located in the eastern Mediterranean region in the transition zone between a humid climate in the north and an extremely arid climate in the south and southeast. The arid areas of the Levant can be divided into two types. One type includes the Negev Desert in southern Israel and the

deserts in Sinai and southern Jordan which are parts of the Saharo-Arabian desert belt formed by the subtropical high-pressure zone located mainly to the south of latitudes 30–31°N. The other type is the local rain shadow desert which is located on the eastern flanks of the mountain ridges, along the Dead Sea transform. The temperatures on the eastern side of the central ridge of Israel (Fig. 1) are usually warmer by several degrees than those on the western side, and the rain clouds coming from the west (i.e., from the eastern Mediterranean Sea) are adiabatically cooled by the moun-

* Corresponding author. Fax: +00-972-2-5380688.

E-mail address: matthews@mail.gsi.gov.il (M. Bar-Matthews).

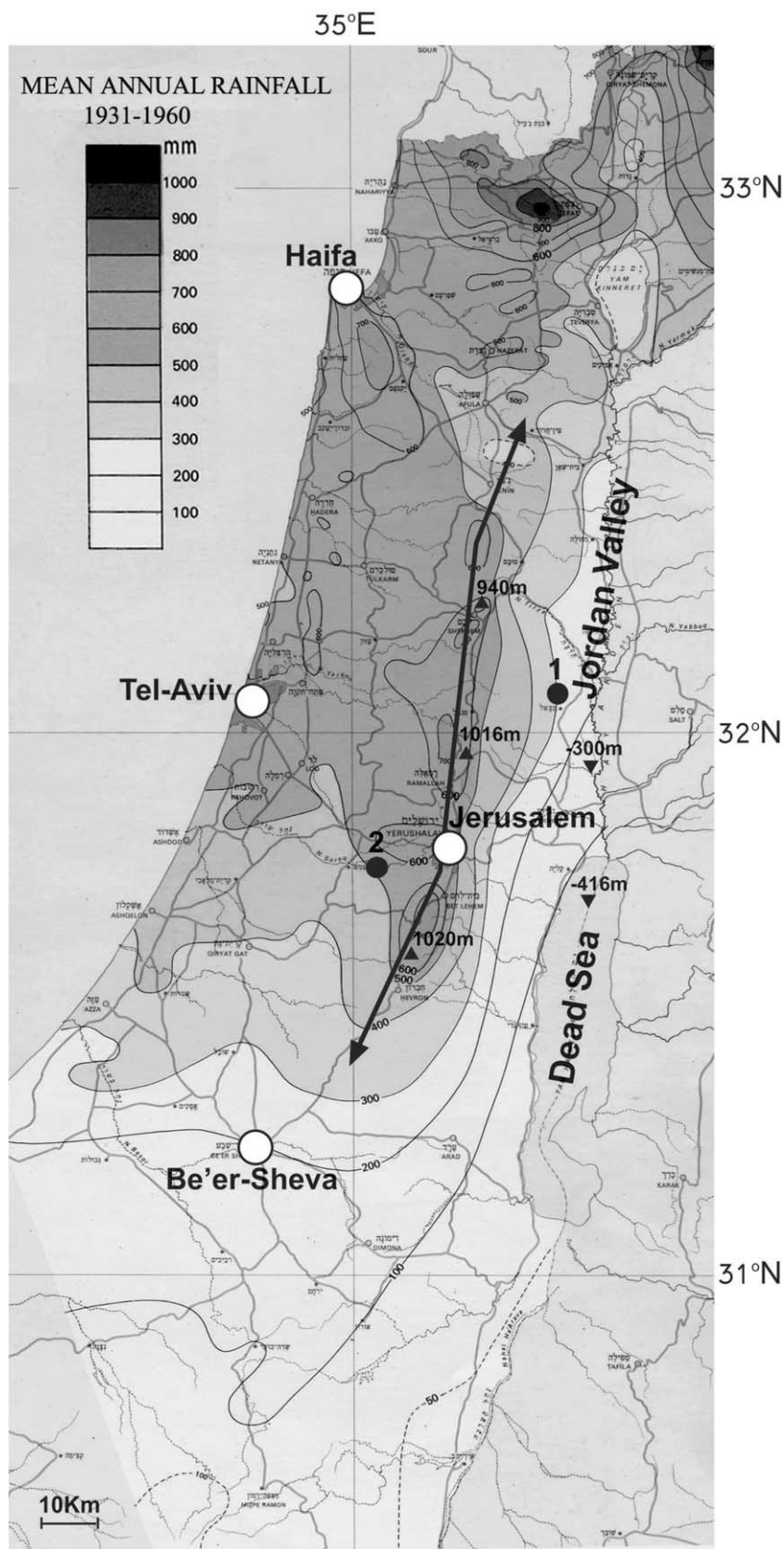


Fig. 1. Location map, showing the studied area. The arrow shows the direction of the central mountain ridge of Israel with Ma'ale Efrayim Cave (1) to the east and Soreq Cave (2) to the west. The gray levels represent the average annual rainfall. The topographic heights are given in meters.

tain ridge, which rises to 800–1000 m (Fig. 1), producing orographic precipitation. On the rain shadow side of the ridge the air undergoes compressional warming and drying.

Most storm tracks reaching the area originate in the Atlantic Ocean and pass over the eastern Mediterranean Sea. The interaction of cold and relatively dry continental air masses with the warm Mediterranean Sea during the winter results in rainfall with a unique isotopic signature in which $\delta^{18}\text{O}$ – δD relationships define the local Mediterranean Meteoric Water Line (MMWL) with $\delta\text{D} = 8 \cdot \delta^{18}\text{O} + 20$ – 30 (e.g., Gat and Carmi, 1970, 1987; Gat, 1996). The $\delta^{18}\text{O}$ and δD values of rainfall become more depleted as the rain moves inland away from the eastern Mediterranean Sea and as elevation increases (e.g., Dansgaard, 1964; Gat, 1996). As a result, under present-day conditions the $\delta^{18}\text{O}$ values of rain falling near the shoreline are about -4 to -5‰ , whereas on the central mountain ridge of Israel (Fig. 1) the $\delta^{18}\text{O}$ values decrease to about -5 to -7‰ as a function of elevation (Bar-Matthews and Ayalon, 2001; Bar-Matthews et al., 2003).

The present research deals with the timing and isotopic composition of speleothem deposition from Ma'ale Efrayim Cave. The cave is located in the rain shadow on the eastern side of the central mountain ridge. The timing of speleothem growth provides an ideal indication of the climatic conditions, because the formation and growth rates of speleothems depend on the availability of water. In colder climates, speleothems do not grow during glacial periods because of the freezing conditions (e.g., Gascoyne et al., 1982; Gordon et al., 1989; Lauritzen, 1995; Schwarcz, 1986). Thus, in warmer areas, such as the site of this study in the Levant region of the eastern Mediterranean, the timing of speleothem growth can serve as an indicator for when the amount of rainfall is less than the amount of water lost through evaporation.

The purpose of the present study was to enhance our understanding of the climatic effect of the rain shadow in three major aspects: (1) to determine the present-day effect of the rain shadow and evaporation on the rainfall amount and its isotopic composition; this was carried out through measurements of the rainfall amount and the oxygen and hydrogen isotopic composition of present-day rain and cave water in Ma'ale Efrayim and through comparison with the rain and cave water composition from Soreq Cave, located in the rainier region on the western side of the ridge; (2) to evaluate the effect of the rain shadow on speleothem growth during interglacial and glacial intervals; this was carried out through detailed ^{230}Th – U TIMS dating; and (3) to evaluate the effect of the rain shadow on the paleoclimatic conditions on both sides of the central mountain ridge during the past 80,000 years; This was achieved through measurements of the carbon and oxygen isotope composition of the Ma'ale Efrayim speleothems and through comparison of this record with the well-studied record of the Soreq Cave (e.g., Ayalon et al., 2002; Bar-Matthews et al., 2000, 2003).

Cave location and research methods

Study area

Ma'ale Efrayim Cave is located in the rain shadow on the eastern side of the central ridge of Israel, 250 m above sea level, 60 km inland from the Mediterranean Sea, and near the Jordan Rift Valley (Fig. 1). The present-day average annual rainfall near the cave is 250–300 mm, the mean annual temperature is 21–22°C, and the annual potential evaporation is 1800 mm. The area is karstified Cretaceous dolomite with scarce Irano-Turanic vegetation located mainly in soil pockets. The cave was formed along a tectonic fracture in the dolomitic Cenomanian Amminadav Formation and is presently 2–20 m below the surface. The cave width varies from 3 m in the lower level to about 8 m in the upper level and the length is about 50 m. Until its discovery during road construction about 12 years ago, the cave was closed with no natural opening. Conditions today are arid and the interior of the cave is almost dry, with only a little water dripping mainly during the winter months. No recent speleothem deposition has occurred. However, the large number of speleothems inside the cave indicates that deposition was intensive in the past.

Analytical methods

For the purpose of ^{230}Th – ^{234}U dating, 15 speleothems from Ma'ale Efrayim Cave were sampled, including both stalagmites and stalactites, and ages of only 9 speleothems were within the range of the dating method. The speleothems were sectioned perpendicular and parallel to their growth axes to expose their growth layers. Series of laminae about 0.5–2 cm thick were separated from each other by a diamond saw, resulting in subsamples of 2–20 g each. An effort was made to prevent the contamination of the subsamples by nearby laminae. Age determinations were performed by alpha spectroscopy at the Geological Survey of Israel and by TIMS at the Open University (UK), following the procedures described by Bar-Matthews et al. (1997), Kaufman et al. (1998), and McDermott et al. (1999). The dating results are summarized in Table 1.

Two speleothems were chosen for detailed carbon and oxygen isotopic studies. Four growth layers of the stalagmite were isotopically analyzed: for each layer, four or five points were analyzed, and it was evident that they were deposited in isotopic equilibrium according to Hendy's (1971) criteria. The carbon and oxygen isotopic analyses were performed on samples from a transect along the growth axis, with sampling every 1.0 mm. Isotopic analyses of calcite speleothems were performed on a SIRA-II mass spectrometer at the Geological Survey of Israel and are given the usual δ notation. All $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values were calibrated against the international standard NBS-19 and are reported in per-mil (‰), relative to the Pee Dee Belemnite standard. Instrument precision was better than $\pm 0.1\text{‰}$, and

Table 1
U-Th analytical results: Ma'ale Efrayim cave

Sample	Lamina	^{238}U ppb	$^{234}\text{U}/^{238}\text{U}$	$^{230}\text{Th}/^{234}\text{U}$	$^{230}\text{Th}/^{232}\text{Th}$	Age (10^3 yr)
Stalactite ME-2	ME-2-A1	484	1.1215 ± 0.0520	0.2840 ± 0.0120	245	36.0 ± 1.9
	ME-2-A2	409	1.0701 ± 0.0178	0.3526 ± 0.0072	22	47.0 ± 1.3
	ME-2-A3	317	1.0763 ± 0.0096	0.4857 ± 0.0058	320	71.6 ± 1.3
	ME-2-B	501	1.0627 ± 0.0088	0.7517 ± 0.0067	12876	148.0 ± 3.3
	ME-2-C1	453	1.0723 ± 0.0110	0.8170 ± 0.0150	3344	177.9 ± 9.1
Stalactite ME-3	ME-3-A1	519	1.1141 ± 0.0055	0.2596 ± 0.0019	319	32.5 ± 0.3
	ME-3-A2	505	1.0771 ± 0.0041	0.3202 ± 0.0050	292	41.8 ± 0.8
	ME-3-B	444	1.0846 ± 0.0042	0.3457 ± 0.0025	283	45.9 ± 0.4
	ME-3-C	366	1.1042 ± 0.0074	0.3653 ± 0.0027	132	49.1 ± 0.5
	ME-3-D1	479	1.0865 ± 0.0040	0.3980 ± 0.0016	139	54.8 ± 0.3
	ME-3-D2	415	1.0797 ± 0.0052	0.4471 ± 0.0026	274	63.9 ± 0.5
	ME-3-E	426	1.0843 ± 0.0035	0.7198 ± 0.0034	27	134.7 ± 1.4
	ME-3-F	380	1.0968 ± 0.0049	0.7796 ± 0.0042	260	158.0 ± 2.2
	ME-5-A1	289	1.0768 ± 0.0106	0.4563 ± 0.0091	31	65.7 ± 1.9
Stalactite ME-5	ME-5-A2	276	1.0918 ± 0.0102	0.5182 ± 0.0109	41	78.4 ± 2.2
	ME-5-B	300	1.0599 ± 0.0100	0.7741 ± 0.0167	19	157.8 ± 8.6
Stalactite ME-7	single lamina	408	1.1117 ± 0.0010	0.3016 ± 0.0021	1862	37.3 ± 0.3
Stalactite ME-8	single lamina	793	1.1036 ± 0.0064	0.2915 ± 0.0020	4513	38.8 ± 0.3
Stalagmite ME-12	ME-12-A	690	1.0966 ± 0.0193	0.1438 ± 0.0026	175	16.9 ± 0.3
	ME-12-B	366	1.1166 ± 0.0160	0.2186 ± 0.0033	467	26.7 ± 0.5
	ME-12-C	249	1.0682 ± 0.0044	0.2476 ± 0.0024	147	30.8 ± 0.3
	ME-12-D	235	1.0696 ± 0.0032	0.2536 ± 0.0016	245	31.6 ± 0.4
	ME-12-E	293	1.1439 ± 0.0062	0.2314 ± 0.0020	439	28.5 ± 0.5
	ME-12-F	186	1.0866 ± 0.0035	0.2528 ± 0.0025	395	31.6 ± 0.4
	ME-12-G1	202	1.0827 ± 0.0041	0.2538 ± 0.0011	209	31.9 ± 0.2
	ME-12-I1	226	1.0965 ± 0.0503	0.2914 ± 0.0154	98	37.3 ± 2.4
	ME-12-I2	270	1.0877 ± 0.0051	0.2813 ± 0.0033	213	35.8 ± 0.5
	ME-12-I3	242	1.1013 ± 0.0078	0.2967 ± 0.0031	151	38.1 ± 0.5
	ME-12-I4	146	1.0935 ± 0.0044	0.3018 ± 0.0022	160	38.7 ± 0.2
	ME-12-J1	169	1.1052 ± 0.0076	0.3592 ± 0.0043	42	48.0 ± 0.7
	ME-12-J2	175	1.1028 ± 0.0063	0.3424 ± 0.0095	45	45.3 ± 1.6
	ME-12-J4	146	1.0816 ± 0.0062	0.3931 ± 0.0028	89	53.9 ± 0.5
	ME-12-J5	234	1.0720 ± 0.0062	0.4214 ± 0.0050	223	59.1 ± 1.0
	ME-12-K3	182	1.0969 ± 0.0049	0.4355 ± 0.0058	114	61.6 ± 1.1
	ME-12-K4	198	1.1321 ± 0.0070	0.4138 ± 0.0030	71	57.4 ± 0.6
	ME-12-K5	163	1.0998 ± 0.0052	0.4429 ± 0.0047	147	63.0 ± 0.9
	ME-12-K7	232	1.0967 ± 0.0046	0.4363 ± 0.0033	343	61.7 ± 0.7
Alpha ages						
Stalactite ME-9	ME-9-A	2546	1.1942 ± 0.0108	0.2954 ± 0.0218	25	37.7 ± 3.3
	ME-9-B	1804	1.1401 ± 0.0110	0.7693 ± 0.0121	1008	152.0 ± 5.5
Stalagmite ME-12	ME-12-G2	210	1.1249 ± 0.0563	0.2440 ± 0.0275	35	33.4 ± 4.0
	ME-12-H1	217	1.1071 ± 0.0465	0.2954 ± 0.0249	17	37.9 ± 3.9
	ME-12-H2	207	1.0833 ± 0.0387	0.3130 ± 0.0203	15	40.6 ± 3.0
Stalagmite MME-1	MME-1-A	481	1.4444 ± 0.0420	0.7527 ± 0.0241	327	136.4 ± 8.9

the external reproducibility based on duplicate measurements of reference standard was $\leq 0.05\%$.

A continuous isotopic record was obtained under the assumptions that the measured age represents the center of the dated lamina and that the growth rate from the center of the dated lamina to the center of the adjacent dated lamina is constant. Consequently the age of an isotopic event is more precise for time intervals with a larger number of dated laminae.

All rain events were collected in Ma'ale Efrayim village, 800 m from the cave, during two rainy seasons, from October 1998 to May 2000. The collection of each rain event enables the precise determination of the amount and distribution of the rain. The rainwater amount was measured by

allowing water to accumulate in a large funnel and drip into a narrow-headed bottle according to the method described in Ayalon et al. (1998). Cave water was sampled from April 1999 to May 2000 at 10 different sites within the cave. Oxygen isotopic compositions of rain and cave water were determined after equilibration with CO_2 by shaking a 2-ml water sample for 4–6 h at 25°C (Epstein and Mayeda, 1953). Hydrogen isotopic compositions were determined after the water samples reacted with Zn reagent at 500°C for 50 min (Coleman et al., 1982; Tanweer et al., 1988). Analytical reproducibility of duplicates measured on different days is better than 0.1% for $\delta^{18}\text{O}$ and 1.0% for δD . $^{18}\text{O}/^{16}\text{O}$ and D/H measurements were made on the VG SIRA-II mass spectrometer at the Geological Survey of Israel and are

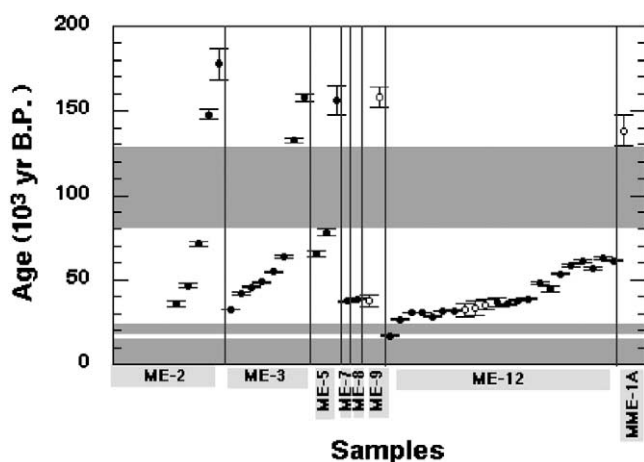


Fig. 2. Ages with error bars of eight speleothems from Ma'ale Efrayim Cave in stratigraphic order (from the youngest to the oldest lamina). Solid circles mark the TIMS ages; open circles show the alpha spectrometric ages. The shaded areas mark hiatuses.

given in per-mil notation relative to SMOW (Craig, 1961), calibrated using the Vienna SMOW standard.

The average cumulative annual rainfall and the average cumulative isotopic composition were calculated after weighing to allow for the proportion of the annual precipitation contributed by each rain event Ayalon et al., 1998; (Bar-Matthews et al., 1996). The isotopic composition and the amount of rainfall in the Ma'ale Efrayim region and cave water were compared to the rain and cave water of Soreq Cave collected during the same time periods.

Results

Growth periods

The results of the dating are shown in Table 1 and are plotted in Fig. 2 for each speleothem, in order of the youngest to the oldest lamina. The distribution shows that almost all the ages are in accord with the stratigraphic order, except for several laminae from one stalagmite where some slight age overturns were measured (Fig. 2, sample ME-12). These overturns may have been caused during the sawing of the sample by contamination from the nearby outer younger laminae. The age distribution shows that there are periods during which speleothems grew in Ma'ale Efrayim Cave and periods during which no growth occurred, i.e., periods when no ages were determined in any of the speleothems.

The age distributions for all eight speleothems (Fig. 2) show that speleothem growth occurred during the following periods: glacial marine oxygen isotope stage 6, from 185,000 to 130,000 yr B.P.; glacial stages 4 and 3, from 80,000 to 25,000 yr B.P.; and the short interval near the end of the stage 2 Last Glacial Maximum from about 19,000 to 16,000 yr B.P. Some speleothems are older than 185,000 yr

B.P. but the magnitude of the uncertainties precludes their confident interpretation.

Hiatuses (periods of no speleothem deposition) are indicated for the following time intervals: the last interglacial, i.e., marine isotope stage 5 from 130,000 to 80,000 yr B.P.; the Last Glacial Maximum at about 25,000 to 19,000 yr B.P.; and during the end of stage 2 and throughout the Holocene from about 16,000 yr B.P. until the present day. The hiatuses in speleothem growth inferred from the ages are also defined in the petrography of the speleothems by the occurrence of very sharp changes in color and the presence of thin (~1-mm) white micritic laminae, red detrital-rich laminae, and paleo-corallites.

The ceiling of Ma'ale Efrayim Cave is characterized by numerous small conical stalactites (ME-7, ME-8, ME-9), whose tips range in age between 39,000 and 37,000 yr B.P. No small conical stalactites younger than these were found, suggesting that this was a period when the cave was highly active. In contrast, ages of similar small conical stalactites from caves on the western flanks of the central ridge are of the Holocene time interval, showing that those caves were active for at least part of the period during which Ma'ale-Efrayim Cave was dry.

Isotopic composition of present-day rain and cave waters

$\delta^{18}\text{O}$ – δD relationships of individual rainfall events are plotted in Fig. 3 and show that the rain water in Ma'ale-Efrayim Cave falls between the MMWL and the global Meteoric Water Line (MWL). For comparison, the $\delta^{18}\text{O}$ and δD values of the rainfall and cave water at the Soreq site follow the MMWL trend (Ayalon et al., 1998; Bar-Matthews et al., 2003) and show significantly lower $\delta^{18}\text{O}$ and the δD values than those of Ma'ale Efrayim rain and cave

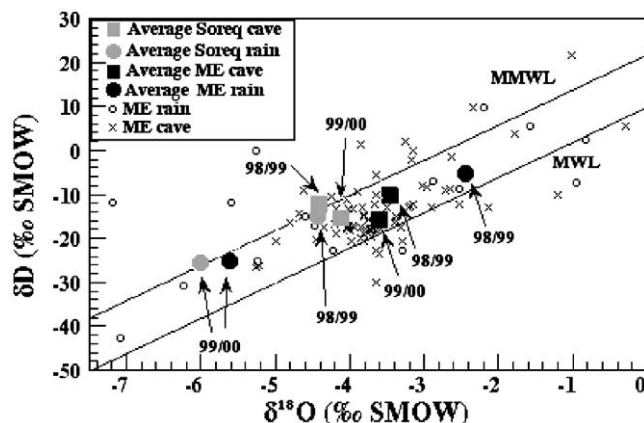


Fig. 3. $\delta^{18}\text{O}$ – δD relationships of all rain events in the Ma'ale Efrayim site and their weighted average for the hydrological years 1998–1999 and 1999–2000. Also shown is Ma'ale Efrayim Cave waters and the weighted average. The weighted average Soreq rain and cave waters for 1998–2000 are also shown (after Bar-Matthews et al., 2003). The Mediterranean Meteoric Water Line (MMWL) and the global Meteoric Water Line (MWL) are shown for reference.

Table 2

Weighted average $\delta^{18}\text{O}$ and δD values of all rainfall and cave water during 1998–2000 in the Ma'ale Efrayim and Soreq caves

	Years	Rainfall (mm)	Rain $\delta^{18}\text{O}$ (‰ SMOW)	Rain δD (‰ SMOW)	Cave $\delta^{18}\text{O}$ (‰ SMOW)	Cave δD (‰ SMOW)
Ma'ale Efrayim	1998–1999	100	−2.45	−5.1	−3.45	−10.3
	1999–2000	250	−5.41	−25.1	−3.6	−15.8
Soreq	1998–1999	198	−4.43	−15.0	−4.1	−12.1
	1999–2000	450	−6.00	−25.6	−4.1	−15.4

water (Fig. 3). The average isotopic composition of the rain and cave water in each site is presented in Table 2.

During the dry year of 1998–1999 the annual rainfall amount was much below average, and high average $\delta^{18}\text{O}$ and δD values characterize the rainwater. During 1999–2000, when the amount of annual rainfall was closer to the mean annual average, $\delta^{18}\text{O}$ and δD values of the rainfall were lower. However, the isotopic composition of Ma'ale-Efrayim Cave water during both years was similar (Table 2, Fig. 3) despite the large differences of the rainwater composition in the 2 years. This suggests that the cave water $\delta^{18}\text{O}$ and δD values are the averages of several years of rainfall, because of the water storage in the epikarst.

The rainfall $\delta^{18}\text{O}$ values of individual events can be divided into two distinct groups (Fig. 4): the light rain events (<5 mm) ranging from +14 to −5‰ and the relatively more massive rain events (>5 mm) with $\delta^{18}\text{O}$ values ranging from 0 to −10‰. This is similar to the trend observed for rains in the western flank of the ridge (Ayalon et al., 1998; Gat, 1996).

Isotopic composition of Ma'ale Efrayim speleothems

Stalagmite ME-12, about 42 cm long, was dated with high resolution, showing continuous growth from 67,000 to ~24,000 yr B.P. with no petrographic evidence for hiatuses during this period. Therefore this sample was considered to

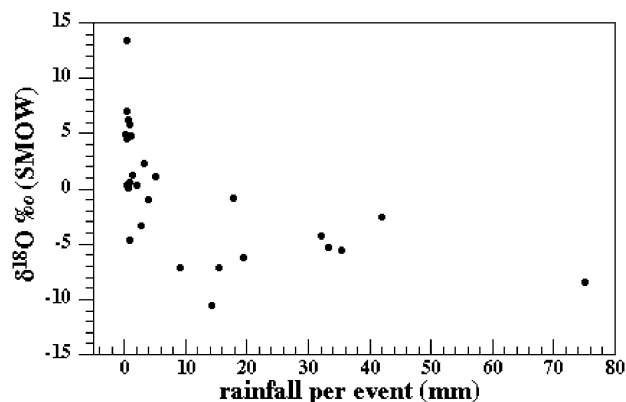


Fig. 4. $\delta^{18}\text{O}$ of rain events vs rainfall amount at the Ma'ale Efrayim site. The rainfall $\delta^{18}\text{O}$ values of individual events can be divided into two distinct groups: the light rain events (<5 mm) ranging from +14 to −5‰ and the relatively more massive rain events ranging from 0 to −10‰.

be the best for high-resolution oxygen and carbon isotopic studies. A second growth period from 19,000 to 16,000 yr B.P. is observed in this sample. Between these two growth periods there is a petrographically distinct white lamina and a sharp change in the color of the laminae from white to brown. Stalactite ME-5, smaller in size, was deposited between 80,000 and 72,000 yr B.P.

The oxygen isotopic composition diagram shown in Fig. 5a is a composite profile based on stalagmite ME-12 and stalactite ME-5. The overall pattern of the $\delta^{18}\text{O}$ profile is of a general increase in $\delta^{18}\text{O}$ values from about −5.5‰ in the oldest lamina to approximately −2‰ in the youngest. Superimposed on the general trend are frequent isotopic oscil-

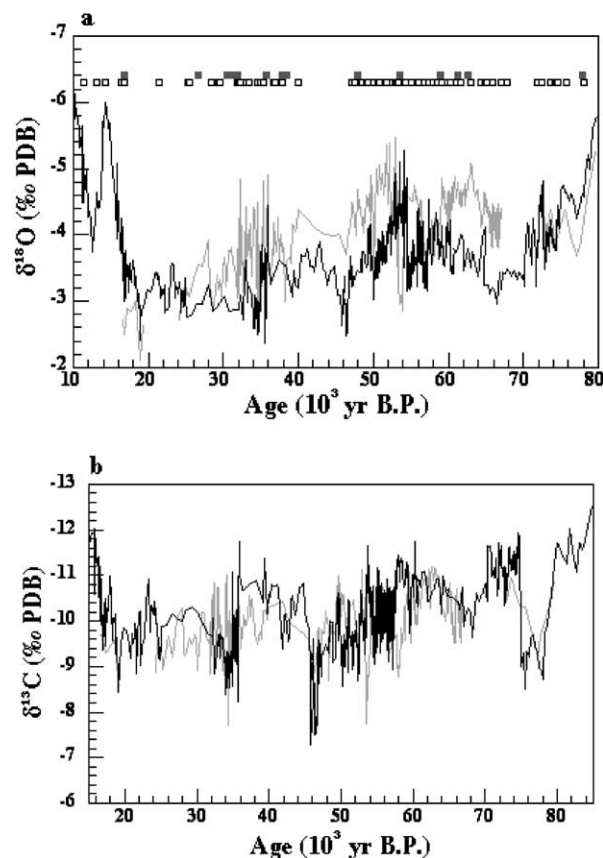


Fig. 5. Superposition of the $\delta^{18}\text{O}$ (a) and $\delta^{13}\text{C}$ (b) records of Ma'ale Efrayim Cave (this study; black) and Soreq Cave (gray) speleothems (after Bar-Matthews et al., 1999), showing the striking similarity between the two records. The squares at the top of (a) indicate the measured TIMS ages (open squares—Soreq Cave, solid squares—Ma'ale Efrayim Cave).

lations of approximately one to several hundred years duration (except for the periods between 46,000 and 40,000 yr B.P.) characterized by amplitude swings as large as 3‰. Low- $\delta^{18}\text{O}$ events occur at 80,000 yr B.P. (−5.3‰), 63,000 yr B.P. (−5.0‰), 53,000 and 52,000 yr B.P. (−5.4‰), 36,000 yr B.P. (−4.8‰), and 32,000 yr B.P. (−4.8‰). High- $\delta^{18}\text{O}$ events occur at 77,000 yr B.P. (−3.6‰), 72,000 yr B.P. (−3.6‰), 66,000 yr B.P. (−3.8‰), 58,000 yr B.P. (−3.6‰), 54,000 yr B.P. (−2.9‰), 50,000 yr B.P. (−3.4‰), 46,000 yr B.P. (−3.7‰), 38,000 yr B.P. (−3.0‰), 35,000 yr B.P. (−2.9‰), 32,000 yr B.P. (−3.1‰), 28,000 yr B.P. (−2.8‰), 24,000 yr B.P. (−2.7‰), and 19,000 yr B.P. (−2.1‰). These events closely match the isotopic events determined for Soreq Cave speleothems (Fig. 5a).

The $\delta^{13}\text{C}$ profile of Ma'ale Efrayim speleothems (Fig. 5b) shows a trend similar to that of the $\delta^{18}\text{O}$ profile, with a slight trend toward higher $\delta^{13}\text{C}$ values over time. Generally, $\delta^{13}\text{C}$ values range from about −11 to −9‰. However, many of the higher $\delta^{13}\text{C}$ values of about −8‰ occur at 54,000 and 35,000 yr B.P., similar to the trend for Soreq Cave (Fig. 5b). Superimposed on the general trend are frequent oscillations lasting one to a few hundred years. A 3‰ fluctuation occurs at around 54,000 yr B.P., from the lowest recorded values of −11 to −8.0‰, the highest values observed throughout the period of deposition.

Discussion

Present-day conditions

The isotopic values of present-day rain and cave waters in Ma'ale Efrayim are generally higher than those of Soreq Cave (Table 2, Fig. 3). This trend is more evident in the drier year 1998–1999 than in the relatively rainy year 1999–2000 (Fig. 3). During the winters of 1998–1999 and 1999–2000 most of the storm tracks bringing rain to both cave sites originated from the eastern Mediterranean Sea. If a Rayleigh-type rainout effect was operating solely (as between the coast and the ridge: Bar-Matthews and Ayalon, 2001), less rain but lower isotopic composition would be expected on the eastern side. The lower amount of rainfall and the higher $\delta^{18}\text{O}$ and δD values of the rain on the eastern side compared with the higher amount of rainfall and the lower $\delta^{18}\text{O}$ and δD values of the rain on the western side thus suggest that a major part of the rainfall evaporated before reaching the Ma'ale Efrayim area, due to the rain shadow effect of the ridge. This view is supported by the $\delta^{18}\text{O}$ – δD relationships of the rainfall, which fall between the MMWL and the MWL and lie (Fig. 3) on a lower slope defining evaporation trends. These trends indicate evaporation below the clouds (e.g., Gat, 1982, 1996). A similar trend was observed by Rindsberger et al. (1990) for rainfall at several stations along a transect from the western to the eastern flank of the central mountain ridge in Israel.

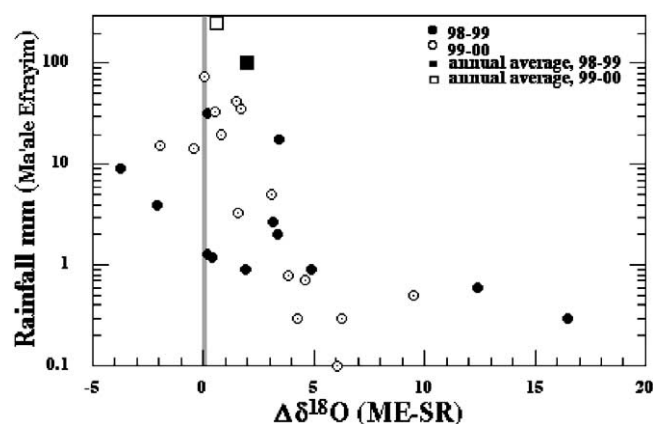


Fig. 6. Difference between Ma'ale Efrayim (ME) rainfall $\delta^{18}\text{O}$ values and Soreq (SR) rainfall values ($\Delta\delta^{18}\text{O}_{\text{ME-SR}}$) vs rainfall amount (mm) in Ma'ale Efrayim, showing that the $\Delta\delta^{18}\text{O}_{\text{ME-SR}}$ increases with decreasing amount of rainfall. The circles represent individual rain events, and the squares represent the average of the 12 rain events in 1998–1999 (solid square) and the 15 rain events during 1999–2000 (open square).

$\delta^{18}\text{O}$ and δD values of Ma'ale Efrayim Cave water have similar composition and reflect a trend similar to that of the local rain water (Fig. 3). This indicates that the source of Ma'ale Efrayim Cave waters is from local precipitation. The difference between the $\delta^{18}\text{O}$ values of the rainfall at the Ma'ale Efrayim Cave site and those of the Soreq rainfall during the same events ($\Delta\delta^{18}\text{O}_{\text{ME-SR}}$) is usually positive. The $\Delta\delta^{18}\text{O}_{\text{ME-SR}}$ difference increases with decreasing amount of rainfall, both during a single event and during the whole rainy season (12 and 15 rain events for 1998–1999 and 1999–2000, respectively; Fig. 6). Ayalon et al. (1998) showed that rainfall events of less than 20 mm at the Soreq Cave site occur mainly when air temperatures exceed 10°C, resulting in evaporation beneath the cloud. Thus, we suggest that the 2–3°C higher surface temperatures at the Ma'ale Efrayim site during rainfall events resulted in more evaporation than that at the Soreq site and increased $\Delta\delta^{18}\text{O}_{\text{ME-SR}}$. Only during relatively massive rainfall events (i.e., above 10 mm at the Ma'ale Efrayim site and above 20 mm at the Soreq Cave site; Fig. 6), when air temperatures in Soreq are between 5° and 10°C and evaporation decreases, does the $\Delta\delta^{18}\text{O}_{\text{ME-SR}}$ value fall to between 2 and −4‰. Under present-day conditions most of the rainfall at the Ma'ale Efrayim site undergoes evaporation. In addition, water loss through evaporation in the unsaturated zone, evapotranspiration, and runoff results in only small amounts of drip water in the cave and consequently the absence of present-day speleothem deposition.

Paleoclimate record

Growth periods

Speleothem deposition in Ma'ale Efrayim Cave occurred during most glacial intervals, marine oxygen isotopic stages 2, 3, 4, and 6 (Fig. 2), with no deposition occurring during

interglacial stage 5 and the Holocene. In addition, no deposition occurred during the height of the Last Glacial Maximum, from about 25,000 to 19,000 yr B.P. (Fig. 2). Thus, water was not available in the unsaturated zone during these periods, unlike the situation on the western side of the same ridge, where speleothem deposition was continuous during the last 240,000 yr (Bar-Matthews et al., 2003).

Previous studies of speleothems showed that in central and northern Israel, on the western side of the central mountain ridge, speleothem deposition was continuous through both glacial and interglacial intervals. The isotopic composition of speleothems recorded both global and regional paleoclimate changes, because the rainfall originated in both the Atlantic Ocean and the eastern Mediterranean Sea (Ayalon et al., 2002; Bar-Matthews et al., 1997, 1999, 2003; Frumkin et al., 1999).

Maximum rainfall on this western flank of the central mountain ridge was suggested to occur in periods when sapropels were formed in the eastern Mediterranean region, mainly between 128,000 and 120,000 yr B.P. and during the early Holocene between 8500 and 7000 yr B.P. Kallel et al. (1997) showed that the salinity of the entire Mediterranean basin was homogeneous during sapropel events. They claimed that this reflected mainly enhanced rainfall throughout the Mediterranean basin, because other processes occurring during sapropel events, such as meltwater discharge from the Eurasian continental ice sheet or enhanced Nile discharge, would have resulted in strong salinity gradients. During the middle Holocene the boundary of the Negev Desert in Israel shifted 20 km south of its present position (Goodfriend and Magaritz, 1988; Goodfriend, 1991, 1999). Goodfriend (1999, Fig. 3) supported this claim on the basis of the isotopic composition of land snails.

The association of extremely high speleothem $\delta^{13}\text{C}$ with minimum $\delta^{18}\text{O}$ values between 128,000 and 120,000 yr B.P. (time equivalent to sapropel S5) lead Frumkin et al. (2000) to suggest that this time interval was hot and dry. They interpreted the very high $\delta^{13}\text{C}$ values of a stalagmite from a cave in Jerusalem as reflecting carbonate rock outcrops that were stripped of vegetation and soil cover due to hot and dry conditions, possibly with the aid of forest fires. An alternative explanation for the strongly defined coupling of low $\delta^{18}\text{O}$ with high $\delta^{13}\text{C}$ that was also found for this time interval and during the beginning of the Holocene in speleothems from Soreq Cave (Bar-Matthews et al., 2000, 2003) is that it is associated with enhanced rainfall during sapropel intervals. The high $\delta^{13}\text{C}$ values are caused by stripping of the soil cover due to deluge events, which resulted in water reaching the cave after little interaction with soil CO_2 . Evidence for this process is provided by the higher pool levels within Soreq Cave (Ayalon et al., 2002; Bar-Matthews et al., 2000, 2003), the large amount of detrital material incorporated in Soreq Cave speleothems during these time intervals (Ayalon et al., 1999), and the sharp drop in the concentrations of Sr, Ba, and U and in the ratios of $(^{234}\text{U}/^{238}\text{U})_0$ and $^{87}\text{Sr}/^{86}\text{Sr}$, which reflect enhanced

weathering of the host rock (Ayalon et al., 1999; Bar-Matthews et al., 2000; Kauffman et al., 1998). A present-day analog is the high $\delta^{13}\text{C}$ value that occurs during massive rainstorms, whereby the $\delta^{13}\text{C}$ values of dissolved inorganic carbon in water infiltrating the cave through thin rock cover have much higher $\delta^{13}\text{C}$ values of ca. -5‰ , compared to values of -14 to -10‰ observed during normal rainfall events (Bar-Matthews et al., 1996).

The occurrence of “wet” eastern Mediterranean conditions in which rainfall occurs on the western flanks of the central mountain ridge of Israel, but is absent on the eastern flanks, is consistent with the rain shadow effect of the ridge, as seen at the present day, leading to the questions: under what conditions does this come about and why were most of the glacial periods “wet?”

The present day is clearly a useful analog since there is now a sharp contrast between the conditions on the western and those on the eastern flanks of the central mountain ridge. Under such conditions very little water reaches the unsaturated zone in the Ma’ale Efrayim region, causing the cave to remain dry. This negative water balance is due to the high evaporation to precipitation ratio, whereby effective precipitation is extremely low. We infer that interglacial conditions were similar and that high evaporation to precipitation ratios led to arid conditions. During the glacial periods, when temperatures were colder, evaporation was lower, and the effective precipitation increased, allowing water to reside in the unsaturated zone. Consequently, speleothem deposition occurred in Ma’ale Efrayim Cave. A similar scenario was suggested to explain speleothem deposition in southeast Australia during glacial periods (Ayliffe et al., 1998). The existence of Lake Lisan (a much larger and deeper late Pleistocene precursor of the Dead Sea) during the last glacial interval (e.g., Bartov et al. 2002; Begin et al., 1974; Neev and Emery, 1967) is a strong indicator that relatively humid conditions prevailed on both sides of the central ridge during glacial intervals.

The short hiatus during the last glaciation, between about 24,000 and 19,000 yr B.P., represents the aridity that developed in the area probably as a result of a sharp drop in the rainfall then (Bar-Matthews et al., 1997), which led also to a sudden drop in the level of Lake Lisan (Bartov et al., 2002) and is evident in the Red Sea (Almogi-Labin et al., 1986).

Isotopic record

The isotopic profiles of Ma’ale Efrayim Cave and Soreq Cave speleothems for this time interval are compared in Fig. 5. There is a very good correspondence in the general trend of both records. Based on the good matching between the well-dated parts of the records, we “wiggly-matched” the Ma’ale Efrayim Cave record with the Soreq Cave record for those time intervals for which only a few age determinations were made. These periods are 80,000–72,000, 67,000–63,000, 30,000–25,000, and 19,000–16,000 yr B.P. The gap between 72,000 and 67,000 yr B.P. is probably

due to lack of sampling. Such similarity of speleothem records from two caves located ~ 60 km away from each other on different sides of the central mountain ridge and in different present-day climatic conditions confirms our inference that the source for the rainfall from which the speleothems acquired their isotopic signal was the same and, as shown by Bar-Matthews et al. (2003) for the Soreq and Peq'iin caves, this source was the eastern Mediterranean Sea. Moreover, the similarity between their oxygen isotopic record trends suggests that, unlike the present day, the climate in both sites during this time interval was generally similar. The matching also supports this suggestion observed between the $\delta^{13}\text{C}$ isotopic profiles of the speleothems from both caves (Fig. 5b), which indicates that both were covered by similar vegetation. However, from 46,000 to 36,000 yr B.P. and from 30,000 to 16,000 yr B.P., the $\delta^{13}\text{C}$ values of Ma'ale Efrayim Cave speleothems were higher by $\sim 0.4\%$ than those of Soreq Cave speleothems, possibly because during these periods the vegetation in Ma'ale Efrayim contained a slightly higher proportion of C4 type vegetation than that in Soreq and/or because of differences in the soil productivity at the two sites (Frumkin et al., 2000).

The $\delta^{18}\text{O}$ values of Ma'ale Efrayim Cave speleothems are lower on average by $\sim 0.5\%$ than those of Soreq Cave speleothems (Fig. 5a; integrated averages are -4.06 ± 0.54 and $-3.53 \pm 0.45\%$, respectively). Two reasons can explain this 0.5% difference; lower $\delta^{18}\text{O}$ of rainfall at the Ma'ale Efrayim site and/or slightly higher temperatures in which speleothems were deposited at the Ma'ale Efrayim site. Under present-day conditions, the rain at the Ma'ale Efrayim site is ^{18}O -enriched relative to that at the Soreq site, due to the evaporation processes brought about by the sharp increase in temperatures on the eastern side of the mountain ridge and the corresponding rainfall/temperature effect of $\sim 0.4\text{‰}/^\circ\text{C}$ (Rozanski et al., 1993). Thus, it is possible that during glacial intervals, when temperatures were colder, these effects were negligible and the rain water $\delta^{18}\text{O}$ values became lower as the rain passed over the central mountain ridge and became isotopically depleted through Rayleigh fractionation processes (rainout effect; Gat, 1996). This alternative is suggested by the present-day analogue, showing that when relatively massive rainfall events occur, and air temperature is below 10°C , the $\Delta\delta^{18}\text{O}_{\text{ME-SR}}$ is low and even negative (Fig. 6). The smaller $\Delta\delta^{18}\text{O}_{\text{ME-SR}}$ of the rainier year (Fig. 6) also supports this possibility. An alternative possibility is that the $\delta^{18}\text{O}$ of rain waters did not change over the ridge and that the lower $\delta^{18}\text{O}$ values in the Ma'ale Efrayim speleothems indicate temperatures $2\text{--}3^\circ\text{C}$ higher at the Ma'ale Efrayim site than at the Soreq site. In both cases, low evaporation is a fundamental factor controlling the availability of water in the epikarst system.

During the short time intervals between 19,000 and 16,000 yr B.P., at the transition from glacial to interglacial conditions, the general pattern changes and $\delta^{18}\text{O}$ values of Ma'ale Efrayim Cave speleothems are on average $\sim 0.5\%$

higher than those of Soreq Cave. This positive $\sim 0.5\%$ difference is that which would be calculated if deposition were hypothetically occurring in the Ma'ale Efrayim Cave under present-day conditions. The $\delta^{18}\text{O}$ values calculated for speleothems forming in equilibrium from the Ma'ale Efrayim Cave waters are -5.0% (calculated by substituting the average cave water $\delta^{18}\text{O}$ value of -3.5% and an average cave temperature of 21°C into the O'Neil et al. (1969) calcite–water fractionation equation). This value is higher by $\sim 0.4\%$ than the average $\delta^{18}\text{O}$ value of -5.4% of modern speleothems from Soreq Cave, which precipitate at average cave temperature of 18°C from cave water with an average $\delta^{18}\text{O}$ value of -4.7% (Bar-Matthews and Ayalon, 2003). The higher hypothetical $\delta^{18}\text{O}$ value of present-day speleothems in Ma'ale Efrayim Cave reflects waters that are isotopically enriched through evaporation; higher temperatures would cause a relative $\delta^{18}\text{O}$ lowering in the speleothems. Similarly, their $\delta^{13}\text{C}$ value also would have been higher than that of Soreq Cave speleothems, because of the dominance of C4 vegetation in the Ma'ale Efrayim area. Thus, the transition from colder glacial to warmer interglacial intervals in the Ma'ale Efrayim area involved both a temperature increase and the development of aridity on the eastern side of the central mountain ridge. A similar trend and enrichment of 0.5% in $\delta^{18}\text{O}$ is also observed for another speleothem (ME-5) deposited between 80,000 and 72,000 yr B.P. However, sample ME-5 was collected from a different site within the cave, and it may reflect local variation in the temperature and the drip isotopic composition.

The comparison of the isotopic records of Ma'ale Efrayim Cave and Soreq Cave shows that during cool glacial conditions, when temperatures were $\sim 6\text{--}10^\circ\text{C}$ lower in the entire eastern Mediterranean region (Emeis et al., 2000), the evaporation was much less significant, the amount of rainfall at both sites was more similar, and the differences in the isotopic compositions of rain, cave water, and speleothems reflect either the $2\text{--}3^\circ\text{C}$ gradient between the sites or the lower $\delta^{18}\text{O}$ values of the rain. The similar $\delta^{13}\text{C}$ values (Fig. 5b) show that, unlike present day, the vegetation at both sites was almost the same, suggesting also that the climate during cool glacial intervals was similar on both sides of the ridge and the fluctuations in $\delta^{13}\text{C}$ values were caused by the same reasons (changes in the C3/C4 proportions) as was suggested for Soreq Cave speleothems (Bar-Matthews et al., 1997). Increasing temperatures during the transition from cool glacial intervals to warm interglacial conditions were associated with a decrease in the rainfall amount and an increase in its isotopic composition. As a result the speleothem $\delta^{18}\text{O}$ values became higher at Ma'ale Efrayim Cave than at Soreq Cave (Fig. 5a). As temperatures increased, the effective precipitation decreased, causing very little water to reside in the unsaturated zone, and eventually speleothem deposition ceased in Ma'ale Efrayim Cave.

The nature of the oxygen and carbon isotopic fluctua-

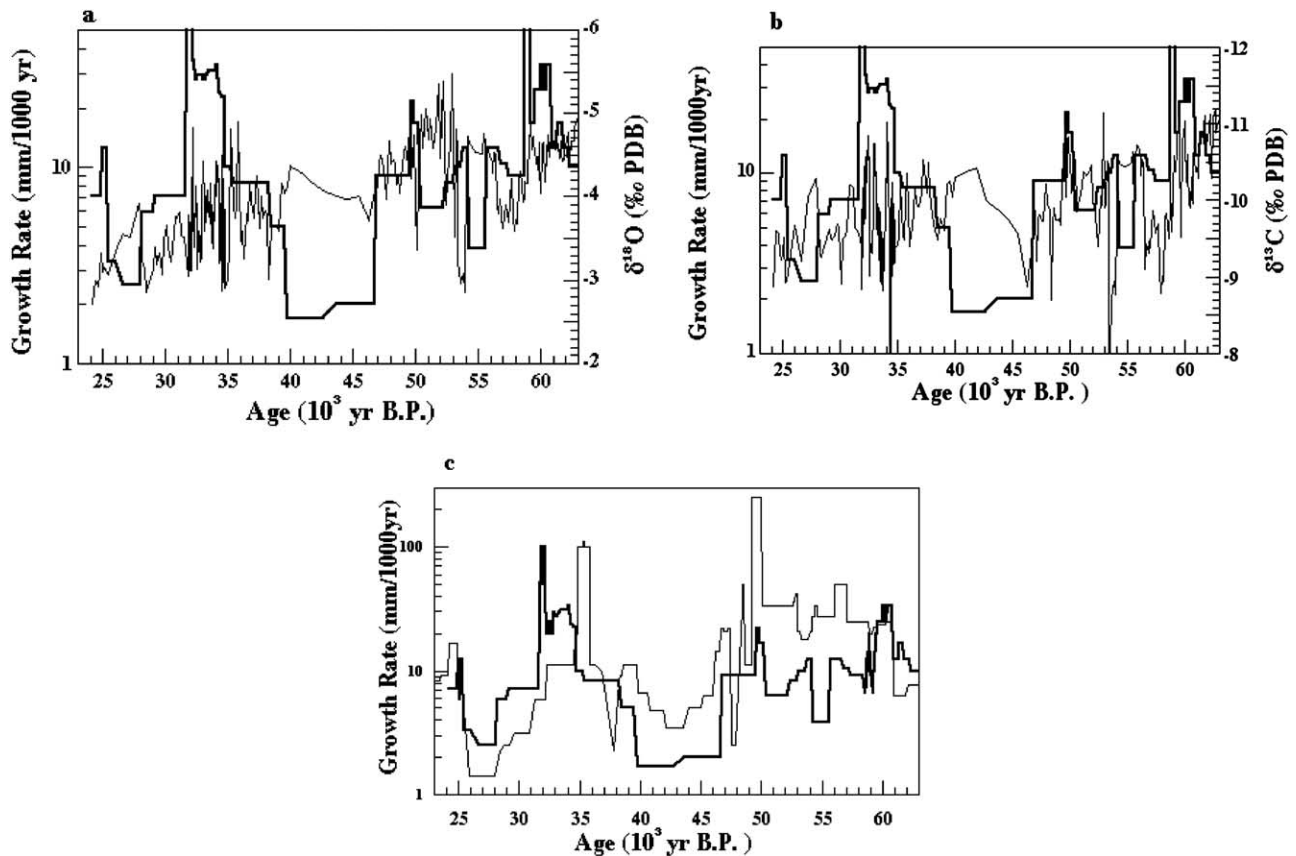


Fig. 7. The calculated growth-rate (black) of a Ma'ale Efrayim Cave speleothem (stalagmite ME-12) superimposed on the $\delta^{18}\text{O}$ (a) and $\delta^{13}\text{C}$ (b) profiles (gray). Fastest growth rates coincide with minimum $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values. In (c) the growth rate of the Soreq Cave speleothems (thin line) is superimposed on that of the Ma'ale Efrayim Cave speleothem (thick line).

tions (the “spikes” and “lows”) in Ma'ale Efrayim Cave can be explained in the light of the Soreq Cave speleothem oscillations during the last glaciation (Bar-Matthews et al., 1999). They usually correlate with global and regional climatic events, such as Heinrich events and the Last Glacial Maximum. Because of the similarity in time and amplitudes of the isotopic events in Ma'ale Efrayim Cave and Soreq Cave speleothems we suggest that they record the same events.

Growth rates

To determine the growth rate, we chose sample ME-12, a large in situ stalagmite that grew during most of the last glacial and was dated with very high resolution (Table 1). Growth rate variations from the Ma'ale Efrayim and Soreq caves closely match the fluctuations in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values (Fig. 7), in which periods with relatively high growth rate are characterized by relatively low $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values. This is most pronounced at ~60,000, 50,000 and 34,000 yr B.P. The trend toward slower growth rates usually follows periods showing increase in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values. We suggest therefore, that higher isotopic values with lower growth rates reflect periods with decreasing rainfall and consequently decreases in the effective precipitation and the

amount of water reaching the unsaturated zone. The similarity in growth rate at the two sites (Fig. 7c) further supports the argument that the rain shadow effect did not exist during glacial conditions.

Conclusions

- Periods of speleothem deposition in Ma'ale Efrayim Cave, located in the rain shadow on the eastern flank of the central mountain ridge, occurred during marine isotopic glacial conditions during periods of increased effective precipitation. Speleothem deposition did not occur in interglacial periods. Rainfall during early stage 2 at 25,000–19,000 yr B.P., the height of the Last Glacial Maximum, appears to have been sufficiently low that speleothem deposition ceased then.
- Water loss through evaporation, evapotranspiration, and runoff results in negligible amounts of drip water in Ma'ale Efrayim Cave and consequently the absence of present-day speleothem deposition. The absence of speleothem growth during interglacial periods is interpreted to reflect hydrologic conditions similar to those of the present day, with the desert boundary in the same position.

- Rainfall in the Ma'ale Efrayim area originates in the eastern Mediterranean Sea. However, the $\delta^{18}\text{O}$ – δD relationships of both rainfall and cave waters fall between the MMWL and the MWL and define evaporation trends. Under present-day conditions, the rain at the Ma'ale Efrayim Cave site is more enriched in ^{18}O than that at the Soreq Cave site on the western flank, due to the evaporation processes brought about by the sharp increase in temperatures on the eastern side of the mountain ridge.
- The general $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ trends of Ma'ale Efrayim Cave speleothems deposited during the last glacial interval is similar to those of the Soreq Cave speleothem located on the western flank of the central mountain ridge, ~60 km away. This indicates that, unlike the present day, the climate on both sides of the ridge during the last glacial interval was similar, and the source for the rainfall from which the speleothems acquired their isotopic signal in both sites was the eastern Mediterranean Sea. The $\delta^{18}\text{O}$ values of Ma'ale Efrayim Cave speleothems deposited during the last glaciation are on average ~0.5‰ lower than those of Soreq Cave speleothems, because the temperatures in Ma'ale Efrayim were slightly higher and/or the $\delta^{18}\text{O}$ of rainfall, associated with higher regional precipitation, was lower.
- Increasing temperatures during the transition from cool glacial intervals to warm interglacial conditions resulted in the relative increase of both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of Ma'ale Efrayim Cave speleothems compared with those of Soreq Cave speleothems, due to decrease in the rainfall amount and increase in the rainfall isotopic composition. Thus, the effect of the rain shadow becomes dominant only during warm interglacial climate conditions, similar to the present day.
- The fluctuations in Ma'ale Efrayim Cave speleothem growth rates closely match the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ fluctuations, with relatively higher growth rates being associated with relatively lower $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values, thus pointing to the correlation between lower isotopic values and increasing amounts of rainfall and effective precipitation. The similarity in contemporary speleothem growth rates at the Ma'ale Efrayim and Soreq cave sites are consistent with the absence of a significant rain shadow during glacial periods and a consequent migration of the desert boundary.

Acknowledgments

This research was supported by The Israel Science Foundation (Grant No. 151/98-13.0 and No. 20/01-13.0). We express our gratitude to the Israel Nature and National Parks Protection Authority for their cooperation. To Irena Segal for help with U-Th measurements and to Yehuda Peled,

Shlomo Ashkenazi, Eli Ram, Rami Solomon, Mark Tz-epelovich, and Meirav Dinor for their help in the field work.

References

- Almogi-Labin, A., Luz, B., Duplessy, J., 1986. Quaternary paleo-oceanography, pteropod preservation and stable isotope record of the Red Sea. *Palaeogeography, Palaeoclimatology, Palaeoecology* 57, 195–211.
- Ayalon, A., Bar-Matthews, M., Sass, E., 1998. Rainfall-recharge relationships within a karstic terrain in the Eastern Mediterranean semi-arid region, Israel, $\delta^{18}\text{O}$ and δD characteristics. *Journal of Hydrology* 207, 18–31.
- Ayalon, A., Bar-Matthews, M., Kaufman, A., 2002. Climatic conditions during marine oxygen isotope stage 6 in the eastern Mediterranean region from the isotopic composition of speleothems of Soreq Cave, Israel. *Geology* 30, 303–306.
- Ayliffe, L.K., Marianelli, P.C., Moriarty, K.C., Wells, R.T., McCulloch, M.T., Mortimer, G.E., Hellstrom, J.C., 1998. 500 ka precipitation record from southeastern Australia: evidence for interglacial relative aridity. *Geology* 26, 147–150.
- Bar-Matthews, M., Ayalon, A., 2001. Eastern Mediterranean paleoclimate during the last 250,000 years as derived from the petrography, mineralogy, trace element and isotopic composition of cave deposits (speleothems), Israel. Geological Survey of Israel Report GSI/41/01, 44 pp.
- Bar-Matthews, M., Ayalon, A., 2003. Speleothems as paleoclimate indicators, a case study from Soreq cave located in the Eastern Mediterranean region, Israel, in: Battarbee, R.W., Gasse, F., Stickly, C.E. (Eds.), *Past Climate Variability through Europe and Africa*, Kluwer Academic Publisher (in press).
- Bar-Matthews, M., Ayalon, A., Matthews, A., Sass, E., Halicz, L., 1996. Carbon and oxygen isotope study of the active water-carbonate system in the karstic Mediterranean cave: implications for paleoclimate research in semiarid regions. *Geochimica et Cosmochimica Acta* 60, 337–347.
- Bar-Matthews, M., Ayalon, A., Kaufman, A., 1997. Late Quaternary paleoclimate in the Eastern Mediterranean Region from stable isotope analysis of speleothems in Soreq cave, Israel. *Quaternary Research* 47, 155–168.
- Bar-Matthews, M., Ayalon, A., Kaufman, A., Wasserburg, G.J., 1999. The Eastern Mediterranean paleoclimate as a reflection of regional events: Soreq cave, Israel. *Earth Planetary Science Letters* 166, 85–95.
- Bar-Matthews, M., Ayalon, A., Kaufman, A., 2000. Timing and hydrological conditions of Sapropel events in the Eastern Mediterranean, as evident from speleothems, Soreq cave, Israel. *Chemical Geology* 169, 145–156.
- Bar-Matthews, M., Ayalon, A., Gilmour, M., Matthews, A., Hawkesworth, C.J., 2003. Sea-land oxygen isotopic relationships from planktonic foraminifera and speleothems in the Eastern Mediterranean region and their implication for paleorainfall during interglacial intervals. *Geochimica et Cosmochimica Acta* (in press).
- Bartov, Y., Stein, M., Enzel, Y., Agnon, A., Reches, Z., 2002. Lake levels and sequence stratigraphy of Lake Lisan, the late Pleistocene precursor of the Dead Sea. *Quaternary Research* 57, 9–21.
- Begin, Z.B., Ehrlich, A., Nathan, Y., 1974. Lake Lisan, the Pleistocene precursor of the Dead Sea. *Geological Survey of Israel Bulletin* 63, 30.
- Coleman, M.L., Shepherd, T.J., Durham, J.J., Rouse, J.E., Moore, G.R., 1982. Reduction of water with zinc for hydrogen isotope analysis. *Analytical Chemistry* 54, 993–995.
- Craig, G., 1961. Isotopic variations in meteoric waters. *Science* 133, 1702–1703.
- Dansgaard, W., 1964. Stable isotopes in precipitation. *Tellus* 16, 438–468.
- Emeis, K.C., Struck, U., Schulz, H.M., Rosenberg, R., Bernasconi, S., Erlenkeuser, H., Sakamoto, T., Martinez-Ruiz, F., 2000. Temperature and salinity variations of Mediterranean Sea surface waters over the last 16,000 years from records of planktonic stable oxygen isotopes and

- alkenone unsaturation ratios. *Palaeogeography Palaeoclimatology Palaeoecology* 158, 259–280.
- Epstein, S., Mayeda, T.K., 1953. Variations of ^{18}O of waters from natural sources. *Geochimica et Cosmochimica Acta* 4, 213–224.
- Frumkin, A., Ford, D.C., Schwarcz, H.P., 1999. Continental oxygen isotopic record of the last 170,000 years in Jerusalem. *Quaternary Research* 51, 317–327.
- Frumkin, A., Ford, D.C., Schwarcz, H.P., 2000. Paleoclimate and vegetation of the last glacial cycles in Jerusalem from a speleothem record. *Global Biogeochemical Cycles* 14, 863–870.
- Gascoyne, M., Schwarcz, H.P., Ford, D.C., 1982. Uranium series ages of speleothem from Northwest England: correlation with Quaternary climate. *Philosophical Transactions Royal Society of London B* 301, 143–164.
- Gat, J.R., 1982. Precipitation, groundwater and surface waters, in: *Paleoclimates and Paleowaters*. International Atomic Energy Agency, Vienna, pp. 3–12.
- Gat, J.R., 1996. Oxygen and hydrogen isotopes in the hydrologic cycle. *Annual Reviews Earth Planetary Science* 24, 225–262.
- Gat, J.R., Carmi, I., 1970. Evolution in the isotopic composition of atmospheric waters in the Mediterranean Sea area. *Journal Geophysical Research* 75, 3039–3048.
- Gat, J.R., Carmi, I., 1987. Effect of climate changes on the precipitation patterns and isotopic composition of water in climatic transition zone: case of Eastern Mediterranean sea area, in: *The influence of Climatic Change Climatic Variability on the Hydrologic Regime and Water Resources*, IANS 168 Symp. Proc. pp. 513–523.
- Goodfriend, G.A., 1991. Holocene trends in ^{18}O in land snail shells from the Negev Desert and their implications for changes in rainfall source areas. *Quaternary Research* 35, 417–426.
- Goodfriend, G.A., 1999. Terrestrial stable isotope records of Late Quaternary paleoclimates in the eastern Mediterranean region. *Quaternary Science Reviews* 18, 501–513.
- Goodfriend, G.A., Magaritz, M., 1988. Palaeosols and late Pleistocene rainfall fluctuations in the Negev Desert. *Nature* 332, 144–146.
- Gordon, D., Smart, P.I., Ford, D.C., Andrew, J.N., Atkinson, T.C., Rowe, P.J., Christopher, N.S.J., 1989. Dating of Late Pleistocene interglacial and interstadial periods in the United Kingdom from speleothem growth frequency. *Quaternary Research* 31, 14–26.
- Hendy, C.H., 1971. The isotopic composition of the speleothems I. The calculation of the effects of different modes of formation on the isotopic composition of speleothems and their applicability as paleoclimatic indicators. *Geochimica et Cosmochimica Acta* 35, 801–824.
- Kallel, N., Paterne, M., Duplessy, J.-C., Vergnaud-Grazzini, C., Pujol, C., Labeyrie, L., Arnold, M., Fontugne, M., Pierre, C., 1997. Enhanced rainfall in the Mediterranean region during the last sapropel event. *Oceanologica Acta* 20, 697–712.
- Kaufman, A., Wasserburg, G.J., Porcelli, D., Bar-Matthews, M., Ayalon, A., Halicz, L., 1998. U-Th isotope systematics from the Soreq Cave Israel and climatic correlations. *Earth Planetary Science Letters* 156, 141–155.
- Lauritzen, S.E., 1995. High-resolution paleotemperature proxy record for the last interglaciation based on the Norwegian speleothems. *Quaternary Research* 43, 133–146.
- McDermott, F., Frisia, S., Huang, Y., Longinelli, A., Spiro, B., Heaton, T.H.E., Hawkesworth, C.J., Borsato, A., Keppens, E., Fairchild, I.J., Borg, K., Verheyden, S., Selmo, E., 1999. Holocene climate variability in Europe: evidence from $\delta^{18}\text{O}$, textural and extension rate variations in three speleothems. *Quaternary Science Reviews* 18, 1021–1038.
- Neev, D., Emery, K.O., 1967. The Dead-Sea; depositional processes and environments of evaporites. *Geological Survey of Israel Bulletin* 41, 147 pp.
- O'Neil, J.R., Clayton, R.N., Mayeda, T.K., 1969. Oxygen isotope fractionation of divalent metal carbonates. *Journal of Chemical Physics* 30, 5547–5558.
- Rindsberger, M., Jaffe, Sh., Rahamin, Sh., Gat, J.R., 1990. Patterns of the isotopic composition of precipitation in time and space: data from the Israeli storm water collection program. *Tellus* 42B, 263–271.
- Rozanski, K., Araguás-Araguás, L., Gonfiantini, R. (1993). Isotopic patterns in modern global precipitation. *Climate Change in Continental Isotopic Records*, Geophysical Monograph 78, American Geophysical Union, 1–36.
- Schwarcz, H.P., 1986. Geochronology and isotopic geochemistry of speleothems, in: Fritz, P., Fontes, J.C. (Eds.), *Handbook of Environmental Isotope Geochemistry*, Vol. 2. Elsevier, Amsterdam, pp. 271–300.
- Tanweer, A., Hut, G., Burgman, J.O., 1988. Optimal conditions for the reduction of water to hydrogen by zinc for mass spectrometric analysis of the deuterium content. *Chemical Geology (Isotope Geoscience Section)* 73, 199–203.

Paleoclimate and location of the border between Mediterranean climate region and the Saharo–Arabian Desert as revealed by speleothems from the northern Negev Desert, Israel

A. Vaks^{a,b,*}, M. Bar-Matthews^b, A. Ayalon^b, A. Matthews^a, A. Frumkin^c, U. Dayan^c,
L. Halicz^b, A. Almogi-Labin^b, B. Schilman^b

^a *Institute of Earth Sciences, Hebrew University of Jerusalem, Jerusalem 91904, Israel*

^b *Geological Survey of Israel, 30 Malchei Israel St, Jerusalem 95501, Israel*

^c *Department of Geography, Hebrew University of Jerusalem, Jerusalem 91905, Israel*

Received 9 January 2006; received in revised form 23 May 2006; accepted 10 July 2006

Available online 30 August 2006

Editor: S. King

Abstract

Speleothem bearing karstic caves of the northern Negev Desert, southern Israel, provides an ideal site for reconstructing the paleoclimate and paleo-location of the border between Mediterranean climate region and the Saharo–Arabian Desert. Major periods of speleothem deposition (representing humid periods) were determined by high resolution ²³⁰Th–U dating and corresponding studies of stable isotope composition were used to identify the source of rainfall during humid periods and the vegetation type. Major humid intervals occurred during glacial at 190–150 ka, 76–25 ka, 23–13 ka and interglacials at 200–190 ka, 137–123 ka and 84–77 ka. The dominant rainfall source was the Eastern Mediterranean Sea, with a possible small contribution from southern tropical sources during the interglacial periods. When the interglacial interval rainfall was of Eastern Mediterranean origin, the minimum annual rainfall was ~300–350 mm; approximately twice than of the present-day. Lower minimum amounts of precipitation could have occurred during glacial periods, due to the cooler temperatures and reduced evaporation. Although during most of the humid periods the vegetation remained steppe with mixed C3+C4 vegetation, Mediterranean C3 type steppe-forest vegetation invaded southward for short periods, and the climate in the northern Negev became closer to Mediterranean type than at present. The climate was similar to present, or even more arid, during intervals when speleothem deposition did not occur: 150–144 ka, 141–140 ka, 117–96 ka, 92–85 ka, 25–23 ka, and 13 ka–present-day.

Precipitation increase occurred in the northern Negev during the interglacial monsoonal intensity maxima at 198 ka, 127 ka, 83 ka and glacial monsoonal maxima at 176 ka, 151 ka, 61 ka and 33 ka. However, during interglacial monsoonal maxima at 105 ka and 11 ka, the northern Negev was arid whereas during glacial monsoonal minima it was usually humid. This implies that there is not always synchronicity between monsoonal activity and humidity in the region.

Oxygen isotopic values of the northern Negev speleothems are systematically lower than contemporaneous speleothems of central and northern Israel. This part is attributed to the increased rainout of the heavy isotopes by Rayleigh fractionation processes, possibly due to the farther distance from the Mediterranean coast.

© 2006 Elsevier B.V. All rights reserved.

Keywords: Speleothems; Paleoclimate; Negev Desert; Northern Saharo–Arabian Desert; ²³⁰Th–U ages

* Corresponding author. Tel.: +972 2 5314343; fax: +972 2 5380688.

E-mail address: antonv@gsi.gov.il (A. Vaks).

1. Introduction

A key subject of paleoclimate studies is the reconstruction of climate changes of the Saharo–Arabian desert belt (Fig. 1A), and how this climate was affected by global events. At present, the Saharo–Arabian Desert is influenced by two main climate systems. One associated with polar fronts originating in the northeast Atlantic Ocean and producing series of cyclones that pass over Western Europe before reaching the Mediterranean Sea. The absence of topographic barriers allows these Atlantic low-pressure cells to progress eastwards above the warm Mediterranean Sea, which provides a secondary moisture source for winter rainfall [1], that mainly affects the northern Saharo–Arabian Desert (e.g., [2–4]). The second system originates in the tropical Atlantic and Indian Oceans, and is manifested as the low latitude monsoon that mainly affects the southern parts of Saharo–Arabian Desert (e.g., [5]).

During the Middle–Late Quaternary, the southern boundary of the Saharo–Arabian Desert shifted several hundred kilometers north of its present location, as

demonstrated by the formation of lacustrine sediments [6,7], travertines, uranium ores and speleothems [8,9]. Their occurrence indicates that wetter climate conditions prevailed during part of previous interglacial periods and during the early Holocene [8]. Wet periods also occurred in the south-eastern Sahara during several glacial interstadials. All these humid episodes resulted from the northward migration of the African and Indian monsoon systems [8,10,11,12]. During the last 200 ka the periodicity of the wet phases was usually driven by the 23 ka precession cycle [9].

Paleoclimate data on the northern border of Saharo–Arabian Desert is significantly more limited, particularly in the Eastern Mediterranean (EM) region. Data from Egypt, Tunisia and Morocco [13–15] show that wetter conditions occurred during the early Holocene. Lake sediments in southern Jordan indicate that humid conditions prevailed during previous interglacial marine isotopic stages (MIS) 5.5 and 7.1, and probably also during glacial interstadials in MIS-6 [16]. The major source of paleoclimatic data on the northern Saharo–Arabian desert in the EM area comes

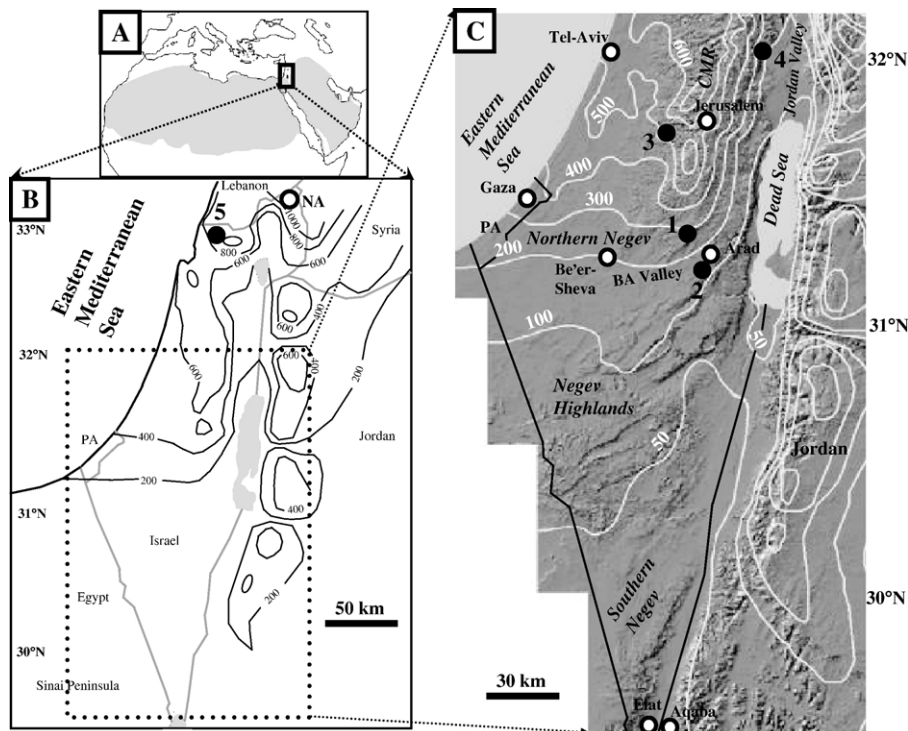


Fig. 1. Geographical, isohyet and location maps of the study area. (A) Map indicating the extent of the Saharo–Arabian desert (in gray). The rectangle marks the research area. (B) Precipitation map of Israel and adjacent lands: Palestinian Authority in Gaza (PA), north-eastern Egypt (eastern Sinai Peninsula), western Jordan, south-western Syria and southern Lebanon. Isohyets are indicated by black lines and present borders by grey lines, the village of Neve–Ativ (NA) is shown by a white circle and Peqi'in Cave by a black circle labelled 5. (C) Map showing the relief [63] and precipitation in the research area in more detail. The isohyets are marked by white lines and the caves by black circles with numbers as following: 1) Ma'ale–Dragot cave system, 2) Tzavoa Cave, 3) Soreq Cave, and 4) Ma'ale–Efrayim Cave. CMR is the Central Mountain Ridge running from north to south, and ending north of Be'er–Sheva–Arad (BA) Valley. Cities are shown by white circles.

from the Negev Desert, southern Israel (Fig. 1—B, C), where ^{230}Th – U ages of travertines [17,18] show that wetter periods occurred at >500 ka, ~ 200 ka, 140–120 ka, 70–25 ka and ~ 15 ka and similarly ^{14}C dates of carbonate nodules from loess soils, at >37 ka, ~ 28 ka, and ~ 13 ka [19]. However, these proxies are only sporadically available and are subject to secondary processes such as erosion and weathering.

The location of the northern boundary of Saharo–Arabian Desert is one of the critical issues for paleoclimate reconstruction of the EM area and determining if climate changes at its northern margins were synchronous with those at the southern boundary and with the monsoonal activity. In this work we determine the timing of shifts in the northern boundary of the desert, and the associated climatic changes, through a study of cave deposits (speleothems) from karstic caves in the northern Negev Desert (Fig. 1C). Today, this region encompasses the transition between the Mediterranean climate region in the north (with annual rainfall of 350–1200 mm, cool rainy winters and hot dry summers) and the arid desert climate region to the south, with less than 200 mm. This study focuses on the region between present-day 300 mm and 150 mm isohyets. The existence of numerous karstic caves rich with speleothems make the region ideal for reconstructing the paleo-position of the desert boundary and understanding its paleoclimate. Speleothems grow in caves when water reaches the unsaturated zone and vegetation is present on the surface to supply the CO_2 necessary for limestone dissolution. The specific aims of the study are: 1) to reconstruct the paleo-climate of the northern Negev region by dating periods of speleothem growth using ^{230}Th – U dating; 2) to determine the rainfall moisture source and the paleovegetation using $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of calcite speleothems. The regional significance of the new data will be explored by a comparison with speleothem records from the Mediterranean climate region in central and northern Israel (mainly with Soreq cave speleothems) [20–28] and the Jordan Valley “rain shadow” desert on the eastern flank of the Central Mountain Ridge (CMR) of Israel [29] (Fig. 1B, C).

2. Climatological background

2.1. Climatic zones of Negev Desert

The present-day Negev Desert can be divided into three main climatic zones (Fig. 1C):

- 1) The *northern Negev Desert*, located at south-eastern corner of the EM Sea, forms a ~ 40 km wide belt that includes the coastal plain near Gaza, the southern edge

of the CMR, and few smaller ridges to the east near the town of Arad. The northern border of the area is approximately defined by the 350 mm isohyet, which marks the southern boundary of Mediterranean climate zone. The southern border is Be’er–Sheva–Arad Valley, dissecting the area from west to east (Fig. 1C). Annual average rainfall varies from ~ 300 to 350 mm in the northern part to ~ 150 mm in the south. Vegetation changes southward from C3 Mediterranean steppe-forest to a mix of C3 and C4 semi-desert Irano–Turanian vegetation [30–34].

- 2) The *Negev Highlands*, south of Be’er–Sheva–Arad Valley consist of several small NE–SW trending ridges with elevations of 500–1033 m asl. The rainfall varies from 150 mm in the north to 50 mm in the south. Vegetation changes southward from semi-desert Irano–Turanian vegetation to Saharo–Arabian desert type, both comprising mixed C3 and C4 vegetation [30,32–34].
- 3) The *southern Negev Desert*, located south of the Negev Highlands, receives ~ 30 –50 mm average annual rainfall and is characterized by Saharo–Arabian desert flora. The region receives its rainfall mostly at the beginning (October–November) and end of the rainy season (March–May) in sporadic short storms, usually accompanied by local floods. Some of this rainfall is associated with synoptic systems that originate in the tropical Atlantic Ocean, pass over Africa and approach the region from the south-southwest. These synoptic conditions infrequently occur in the northern Negev.

2.2. Rainfall gradient in the northern Negev Desert

At present the northern Negev Desert receives most of its rainfall during the winter months (December to February) from mid-latitude cyclones (Cyprus cyclones) moving eastwards above the EM Sea. Dayan [35] found that the typical Cyprus cyclones (the major contributor of rainfall in Israel) correspond to the majority of long fetch of maritime air masses crossing the Mediterranean. Summers, between May and September, are hot and dry and the result from the sinking air of subtropical highs, which develop over the Mediterranean Sea as strong high-pressure ridges push eastwards from the Azores subtropical high.

Rainfall isohyets run from north to south in northern and central Israel parallel to the coastline, but then abruptly change their orientation and run to the west parallel to the northern coastline of the Sinai Peninsula and the N–S rainfall decrease become sharpest even though there isn’t any significant topographic barrier in the 50 km belt inland of the coastline (Fig. 1B, C). This sharp precipitation

gradient change occurs where the Cyprus cyclones cross the Levant region from the west to the east: the northern Sinai coastline forming the southern limit at which rain clouds can form [36]. Consequently, the rainfall amount north of Be'er–Sheva–Arad Valley is much higher than to the south (Fig. 1C). Since the position of the northern Sinai coast changed in the past due to Quaternary sea level fluctuations [37,38], it could affect the amount of rainfall in the Negev Desert.

3. The studied caves

Speleothems were sampled from two cave systems located in limestone and dolomite host rock of Cenomanian–Turonian Judea Group (100–88 Ma):

Ma'ale–Deragot Caves (#1 in Fig. 1C) located between present-day 280–300 mm isohyets, at an elevation of 630–720 m asl, 10–90 m below the surface, and 65 km from the Mediterranean Sea. This is the closest Negev cave system to the region with Mediterranean climate. Until the caves were discovered during quarrying, they had no natural openings.

Tzavoa cave (#2 in Fig. 1C) located within present-day 150–160 mm isohyets, at an elevation of 550 m asl, 20–50 m below the surface, and 80 km from the Mediterranean Sea. Most speleothems used in the present study come from this cave because of its critical location at the southern and most arid part of the research area. The cave has several natural openings.

4. Analytical methods and background data

Fifteen speleothems (3 stalagmites and 12 stalactites) were collected from various locations within the Tzavoa cave. Two stalactites were in-situ, all others were broken. The stalagmites range in size from 5 to 15 cm in length and ~3 to 15 cm in width. The stalactites are up to 30 cm long and up to 8 cm wide. Six randomly located speleothems were collected from the Ma'ale–Deragot cave system: two stalagmites about 0.5 m long and 20 cm wide, and four stalactites varying in size from a few cm to ~0.5 m long and 20 cm wide.

The speleothems were sectioned using a diamond saw to expose their internal structure and to eliminate diagenetically altered samples [21]. The mineralogy and petrography was determined using petrographic microscope, Jeol 840 scanning electron microscope equipped with Oxford ISIS EDS system, and Philips PW 3020 X-ray diffractometer.

For dating purposes, up to 1 g material was drilled using 0.8–4 mm diameter drill bits along the growth axis (for stalagmites) (Fig. 2A) and across the growth axis (for stalactites). Six laminae were sampled twice by

drilling two different spots on the same lamina (these samples with duplicate dating are designated by the roman numerals I and II in Supplemental Table 1). Depending on the uranium concentration (Supplemental Table 1), 80–900 mg calcite powder was dissolved in 7 N HNO₃. The sample was loaded onto mini-columns that contained 2 ml Bio-Rad AG 1X8 200–400 mesh resin. U was eluted by 1 N HBr and Th with 6 N HCl [39]. Afterwards the U and Th solutions were evaporated to dryness and dissolved in 2 ml and 5 ml of 0.1 N HNO₃ respectively.

²³⁰Th–U dating was performed on 125 samples of speleothems from Tzavoa cave and 17 samples from Ma'ale–Deragot caves, using multicollector inductively coupled plasma mass spectrometer (MC-ICP-MS) Nu Instruments Ltd (UK) equipped with 12 Faraday cups and 3 ion counters. The sample was introduced to the MC-ICP-MS through an Aridus® micro-concentric desolvating nebuliser sample introducing system. The instrumental mass bias was corrected (using exponential equation) by measuring the ²³⁵U/²³⁸U ratio and correcting with the natural ²³⁵U/²³⁸U ratio. The calibration of ion-counters relatively to Faraday cups was performed using several cycles of measurement with different collector configurations in each particular analysis. The age determination was possible due to the accurate determination of ²³⁴U and ²³⁰Th concentrations by isotope dilution analysis using the ²³⁶U–²²⁹Th spike. ²³⁰Th/U ages were corrected for detrital ²³⁰Th [40], assuming a ²³²Th/²³⁸U isotope atomic ratio of 3.8 (the mean crustal value) in the detrital components. However, less than 20% of the samples had a ²³⁰Th/²³²Th activity ratio less than 100 and needed this correction (Supplemental Table 1). The reproducibility of ²³⁴U/²³⁸U ratio was 0.11% (2σ).

For δ¹⁸O and δ¹³C analyses, samples of 1–2 mg material were drilled using an 0.8 mm diameter drill, either along or across the growth axis (Fig. 2B). 504 measurements of δ¹⁸O and δ¹³C were made in high resolution on 3 stalagmites and 5 stalactites using VG SIRA-II Mass Spectrometer with ISOCARB system for carbonate analysis [21]. Hendy tests [41] were performed to ensure that the speleothems were deposited in isotopic equilibrium. This test was critical for Tzavoa speleothems since the cave has several present-day natural openings. Calcite δ¹⁸O and δ¹³C values of calcite are reported relative to PDB standard.

For laminae thinner of <5 mm only single dating was performed. Thicker laminae were dated at their top and base, and in several cases where the age difference between the two was greater than 3 ka we increased the dating points. This enabled to precisely calculate the growth rate along the cross section, with the assumption

that the age represents the center of the drilled area, and that there is a constant growth rate between two dated points. The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profiles were measured along the same line, allowing us to define precisely the age of each point.

In order to provide additional data on the present-day conditions, three samples of dripping water from Tzavoa cave were collected in March 2002, at the end of the 2001–2 rainy season. Rain samples were collected during 2004–5 winter from the city of Be'er-Sheva, 380 m asl and ~ 40 km from the EM Sea coast, where each rain event was sampled, and in the town of Arad, 600 m asl, and ~ 80 km from the EM Sea coast. At Arad the rain accumulated in a plastic bucket and to prevent evaporation, a 1 cm thick oil layer was added. Rainwater that accumulated below the oil was sampled at 40–50 day intervals and transferred into sealed plastic bottles. $\delta^{18}\text{O}$ and δD measurements of the cave and

rainwater were performed using methods described in [42] and are reported in the SMOW scale.

Rainfall amounts in Be'er-Sheva and Arad during the 2004–5 winter were 390 mm and 160 mm and their $\delta^{18}\text{O}$ and δD values range from -11.8‰ to 1.3‰ , and -68‰ to 6‰ , respectively (Fig. 3A). At Be'er-Sheva $\delta^{18}\text{O}$ – δD relationships of rain events above 10 mm usually follow the Mediterranean Meteoric Water Line (MMWL) [43], consistent with their EM origin, with the exception of two events on 29.10.2004 and 8–10.3.2005 that originated in the tropical Atlantic Ocean (Israel Meteorological Service data). The $\delta^{18}\text{O}$ – δD data for these events fit the global Meteoric Water Line (MWL). Rain events below 10 mm usually follow the MMWL, with some falling on evaporation trends. $\delta^{18}\text{O}$ – δD relationships of rainfall in Arad follow the MMWL. Present-day $\delta^{18}\text{O}$ values of Tzavoa Cave water are from -4.8‰ to -5.6‰ and δD values are from -14‰ to

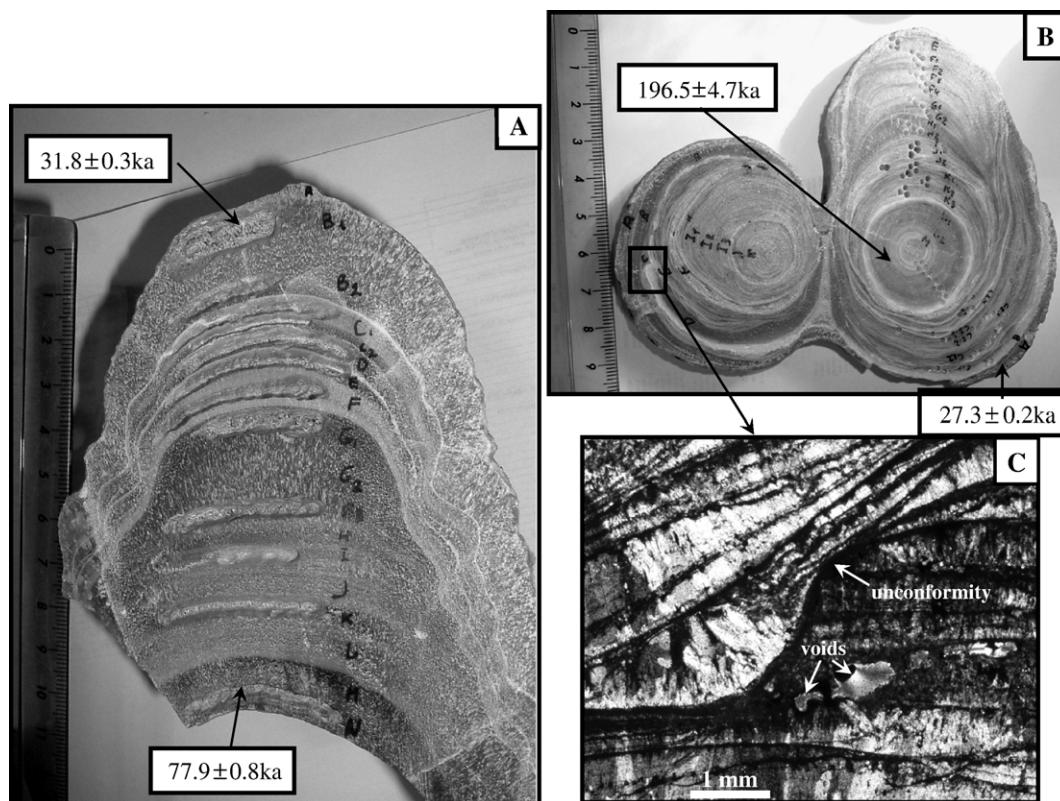


Fig. 2. (A). Typical laminar cross section along the growth axis of stalagmite (TZ-15) from Tzavoa Cave; individual laminae are marked A–N. Two ^{230}Th –U ages are indicated in laminae B1 and N. Three white laminae seen between the laminae B2 and C1, C2 and D, and D and E, mark hiatuses in deposition. Drilling grooves can be seen where samples were taken for ^{230}Th –U dating. (B) Perpendicular cross section of two joined stalactites: TZ-22(1) on the right, and TZ-22(2) on the left. The holes mark the drilling points for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ analyses. (C). Optical microscope image of the lamina between C and D in TZ-22(2) in crossed polarized light. This lamina that defines an unconformity is composed of dark, opaque, micrometer size (micritic) calcite crystals together with detrital material (mainly silica, iron oxide and apatite). This micritization of calcite, as the voids in the lamina D below the unconformity are possibly caused by corrosion during the cease of the stalactite growth between ~ 135 ka and ~ 80 ka.

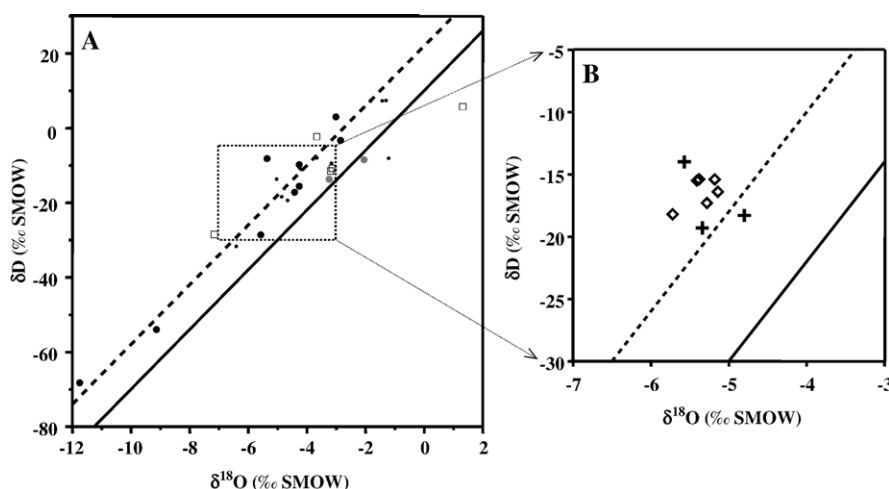


Fig. 3. (A) The $\delta^{18}\text{O}$ and δD values of rainwater from Be'er-Sheva and Arad during the winter 2004–5. Samples from Be'er-Sheva with <10 mm rainfall are shown as small dots; events with >10 mm rainfall are shown as bold black circles. Rain events with >10 mm rainfall but from tropical origin are marked as bold grey circles. Isotope compositions of samples from Arad (shown as small open rectangles) are from five sampling periods: 29.10.2004 event (the first rain of the season); 1.11 to 14.12.2004; 14.12.2004 to 2.2.2005; 2.2 to 13.3.2005; and 13.3 to 22.5.2005. The MWL is marked by the continuous line and MMWL by broken line. (B) $\delta^{18}\text{O}$ and δD values of the dripping water in Tzavoa Cave (crosses) and Soreq Cave (open diamonds) during March 2002, relative to MMWL and MWL.

–19.3‰ (Fig. 3B), and also follow the MMWL, consistent with an EM Sea rain source [43].

No speleothem deposition occurs today because the amounts of rainfall and dripping water in the cave are extremely low, and therefore the present-day relationship between the isotopic compositions of the rainfall and of the speleothems cannot be determined. In the Mediterranean climate region, the speleothem isotopic composition (oxygen of calcite and hydrogen of fluid inclusions) reflects the isotopic composition of rainfall [42,24,43].

5. Results

5.1. Growth periods and petrography

Stalagmites and stalactites from Tzavoa cave are all composed of low magnesium calcite (up to 1% Mg) with laminated texture. The ^{230}Th –U ages show that they grew from ~200 ka to 13 ka (Supplemental Table 1, Fig. 4). None of the individual speleothems grew throughout the entire time span. No speleothems younger than 13 ka were found.

5.1.1. Stalagmites

The stalagmites are composed of 0.1 to 4 cm thick laminae. Columnar honey brown calcite crystals are up to 1 cm in size, devoid of detritus. In each sample, it is possible to identify thin white laminae cutting through the brownish crystals (Fig. 2A). Microscopically, these

white laminae are composed of micrometer (μm) size, dark, opaque, slightly corroded and porous calcite crystals, frequently rich with detritus. The laminae usually mark growth breaks (hiatuses) (Fig. 2B, C).

The stalagmite ages are between 78 ka and 13 ka, (i.e. during the glacial intervals and at the beginning of the deglaciation). Stalagmite TZ-15 grew between 78 ka and 31 ka (Figs. 2A and 4) with hiatus between 46 ka and 40 ka; Stalagmite TZ-6 grew between 46 ka and 30 ka, with an hiatus between 43 ka and 40 ka and stalagmite TZ-14 grew from ~23 ka to 13 ka with hiatuses between ~23 ka and ~20 ka and ~16 ka to ~14 ka.

5.1.2. Stalactites

The laminae of most of the stalactites are thinner than those of the stalagmites (Fig. 2B). In contrast to the stalagmites, stalactites grew from ~200 ka to ~13 ka, during both interglacial and glacial periods (Fig. 4). Stalactites that grew during cool glacial intervals at MIS-6 and during the last glacial (MIS-4 to 2), are characterized by brown honey colored laminae alternating with thin white laminae (Fig. 2B).

The large stalactites: TZ-4, TZ-21 and TZ-22 (up to 8 cm thick and up to 30 cm long) are composed of honey colored, columnar calcite crystals, up to ~1 cm in size, preferably orientated perpendicular to the growth axis. Their growth periods are: TZ-4 between 196 and 47 ka, TZ-21 from 189 to 38 ka and TZ-22 from 200 to 27 ka (Fig. 4). The growth during the last glacial period in

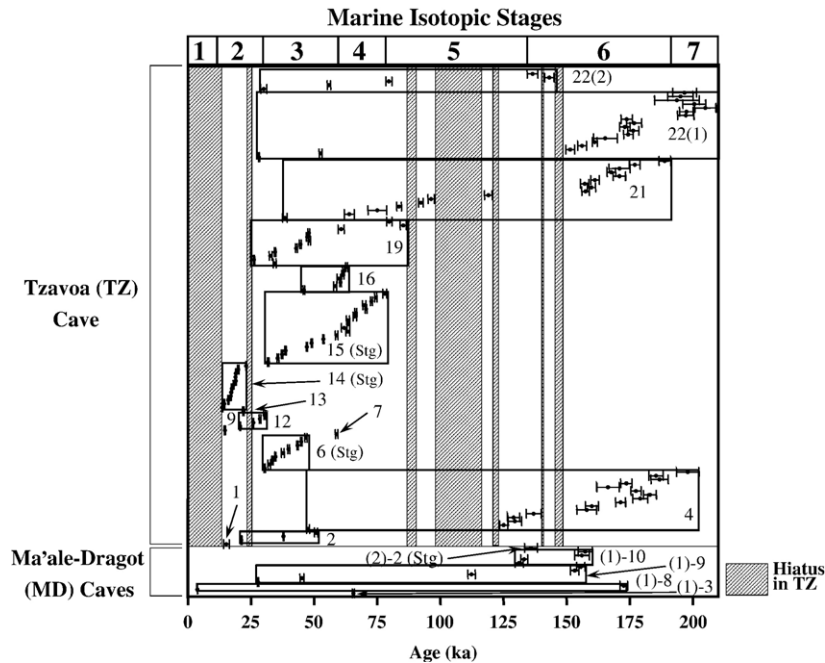


Fig. 4. Periods of speleothem deposition in Ma'ale-Dragot (MD) (bottom) and Tzavoa Cave (TZ) (center and top) as indicated by the age determinations. The horizontal axis marks the age (ka). Age data for different speleothem samples are enclosed in rectangles, together with speleothem sample number. Stalagmites samples are identified by the suffix "Stg". Black dots with error bars define individual measurements. Hiatuses are marked by vertical patterned rectangles. Marine isotopic stages (1–7) are marked at the top of the figure. All data is taken from Supplemental Table 1.

these stalactites was limited to the uppermost laminae with thickness of less than 1 cm, with major hiatuses between 151 ka and 144 ka, 141 ka and 140 ka, 117 ka and ~96 ka, and from ~92 ka to 85 ka (these hiatuses are based on the ages including errors). All these hiatuses are defined by similar petrography as in stalagmites (Fig. 2B, C). Smaller stalactites grew only

during the last glacial period, at similar time intervals to the stalagmites. The stalactites ceased growing at about 13–14 ka (samples TZ-1 and TZ-9), similar to the cessation found for stalagmite TZ-14 (Fig. 4).

The large number of age determinations from the Tzavoa Cave enabled the calculation of the age relative frequency diagram, using Isoplot 3 software [44]. This

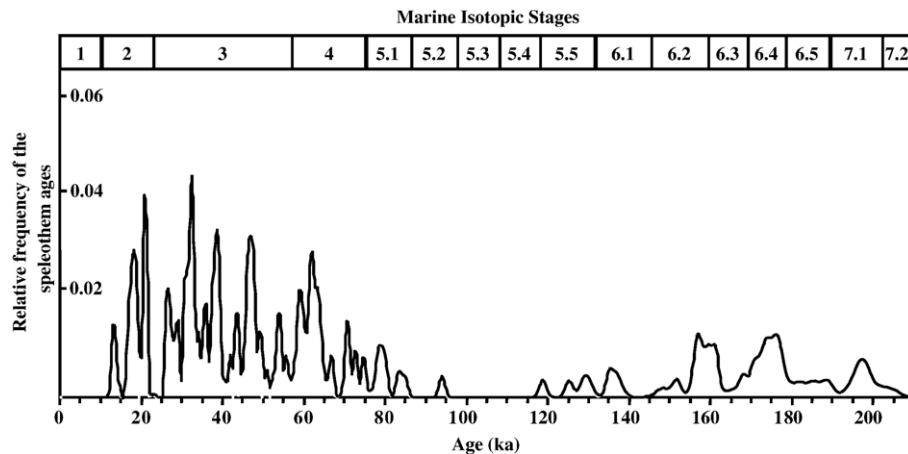


Fig. 5. Running average of speleothem age relative frequency diagram for the Tzavoa Cave during the last 200 ka (see text for explanation). It shows which fraction of the samples formed in certain age with 95% confidence. Marine isotopic stages and substages are marked at the top of the graph.

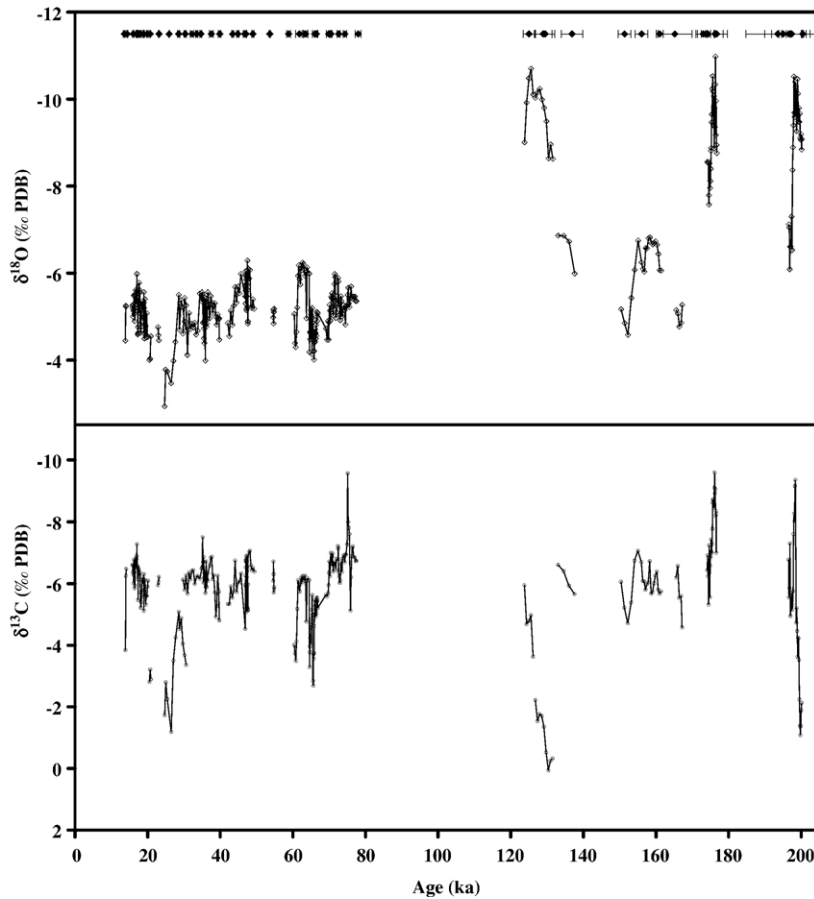


Fig. 6. $\delta^{13}\text{C}$ (lower plot) and $\delta^{18}\text{O}$ profiles (upper plot) of the sampled Tzavoa Cave speleothems for the last 200 ka. The speleothem ages are shown at the top of the plots as black diamonds with horizontal 2 sigma error bars.

analysis (Fig. 5) shows that the numbers of measured ages decrease as the ages become older. This is consistent with a sampling bias towards younger speleothems, which are more accessible in the cave and probably cover the older speleothems. The peaks of higher frequencies also indicate more intensive speleothem deposition and higher humidity. Periods of highest speleothem age frequencies occurred during the last glacial period: at 48–46 ka, ~38 ka, 33–31 ka, 21–20 ka, and 19–17 ka. Longer intervals with relatively high age frequency occurred between 64 and 58 ka, and between 40 and 26 ka. Less significant peaks of age frequencies cluster at 200–196 ka, 178–167 ka, 162–154 ka, 152–150 ka, 137–132 ka, 130–123 ka, ~118 ka, 96–92 ka, 85–81 ka, 80–78 ka, 75–72 ka, ~70 ka, ~54 ka, and ~14 ka.

Growth intervals of speleothems from Ma'ale Dragot Caves are similar to those of the Tzavoa cave (Supplemental Table 1, Fig. 4). However, speleothems older than 200 ka were also found, as well as one sample

with an age of 113 ka that occurs within a long hiatus determined for Tzavoa cave, and one lamina of late Holocene age ~3.8 ka. These two additional ages are consistent with the location of Ma'ale Dragot caves north of Tzavoa, in more humid conditions.

5.2. Oxygen and carbon isotopic composition of speleothems

Eight speleothems, three conical stalagmites and five conical stalactites from Tzavoa Cave were analysed. Where possible, isotopic compositions of speleothems that grew in the same time interval were measured to check if they show similar values and trends in order to verify further that the speleothems were formed in isotopic equilibrium. Similar $\delta^{18}\text{O}$ trends were observed in most samples: stalagmite TZ-6 and stalactite TZ-19 from the 49–43 ka time interval; stalagmite TZ-15 and stalactite TZ-16 between 64 and 58 ka, but with an offset of ~0.5‰. Only in one case, partial matching is evident

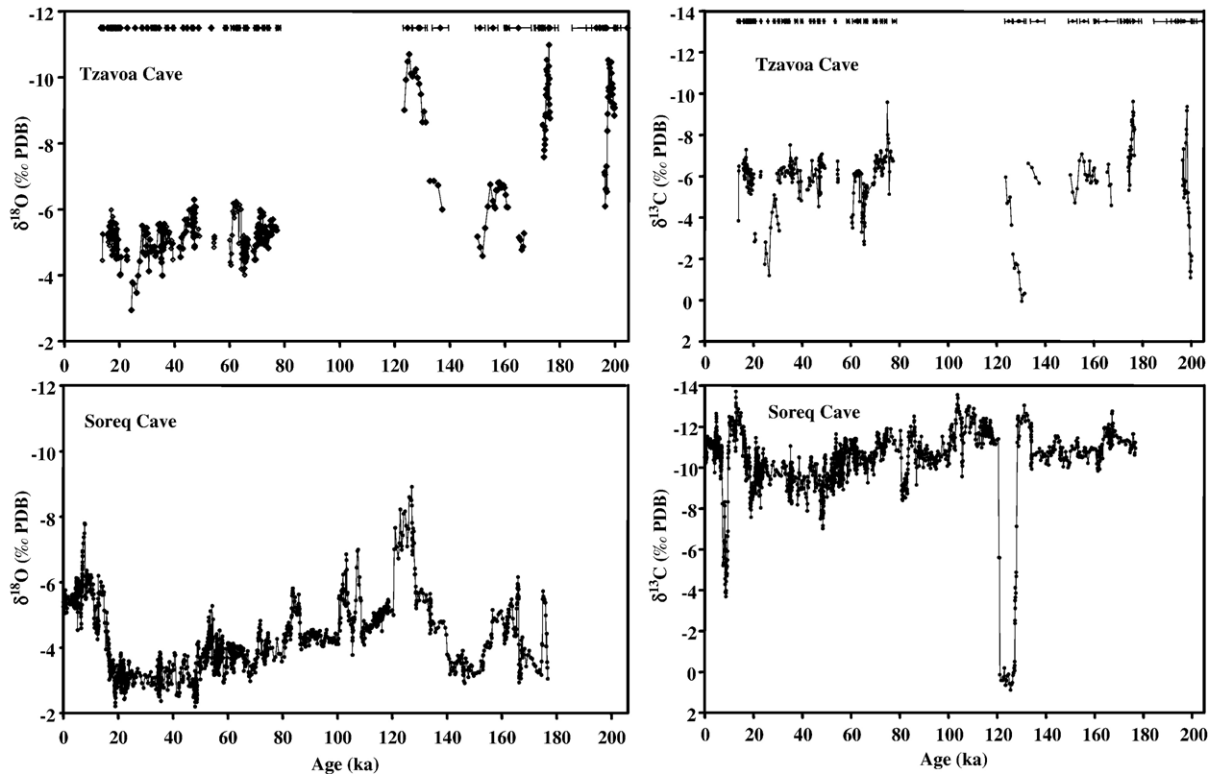


Fig. 7. Comparison of the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profiles of the Tzavoa cave with those of the Soreq cave. $\delta^{18}\text{O}$ profiles of Tzavoa speleothems (upper left plot) compared with Soreq Cave speleothems (lower left plot). $\delta^{13}\text{C}$ profiles of Tzavoa speleothems (upper right plot) compared with Soreq Cave speleothems (lower right plot). The ages of Tzavoa Cave speleothems are shown at the tops of the upper plots by black diamonds with horizontal 2 sigma error bars. The ages of Soreq speleothems are modified from Bar-Matthews et al. [24].

between stalagmite TZ-6, stalactite TZ-19 and stalagmite TZ-15 for the time interval between 39 ka to 30 ka. Since Hendy's test suggests that all samples were deposited in isotopic equilibrium, it is unlikely that these differences are a result of kinetic isotopic fractionation. We therefore suggest that calculation of the ages assuming constant growth rate on both sides of the dating point is not always accurate, causing a slight offset when a very detailed matching is attempted.

The overall range of $\delta^{18}\text{O}$ values is from -11‰ to -3‰ (Fig. 6A). Three $\delta^{18}\text{O}$ minima notably occur in interglacial and glacial intervals when sapropel formation occurred in the EM Sea [45]: 200–196 ka (MIS-7.1) with a $\delta^{18}\text{O}$ variation of -6.0‰ to -10.4‰ ; 177–173 ka (MIS-6.4) with a $\delta^{18}\text{O}$ variation of -7.5‰ to -11‰ and between 131 and 123 ka (MIS-5.5) with a $\delta^{18}\text{O}$ variation of -8.5‰ to -10.7‰ .

During the previous glacial MIS-6.2 and MIS-6.3, at ~ 166 – 150 ka, and the transition between MIS-6.1 and MIS-5.5, at 137–132 ka, the $\delta^{18}\text{O}$ values range between $\sim -7.0\text{‰}$ and -4.5‰ . $\delta^{18}\text{O}$ values of speleothems deposited during the last glacial period (MIS-4–MIS-2,

between ~ 76 ka and ~ 13 ka) vary between $\sim -6.5\text{‰}$ and -3.0‰ . The isotopic compositions of small samples and samples with poor age resolution were not measured; thus the isotopic profiles do not cover all periods of speleothem deposition in the cave and not all gaps in the isotopic profiles reflect lack of deposition (Figs. 4–6).

The oxygen isotopic profile performed for Tzavoa Cave speleothems was compared with Soreq Cave speleothems [24], showing a linear correlations between the $\delta^{18}\text{O}$ profiles ($R^2=0.7$ to 0.8) for the time intervals between 176 ka and 173 ka, 161 and 150 ka, and 137 and 123 ka, but with rough correlation during the last glacial period. Because of the general correspondence between the trends of the two profiles, we considered using wiggle matching (only if it was possible within the 2σ error bars of the ages) in cases when the records did not exactly match each other.

$\delta^{13}\text{C}$ values are shown in Fig. 6B. The overall range of values is from $\sim 0.0\text{‰}$ to $\sim -9.5\text{‰}$, averaging at about -6‰ . $\delta^{13}\text{C}$ values during the last glacial period vary between $\sim -9.2\text{‰}$ and $\sim -1.45\text{‰}$. A sharp drop

from 0.0‰ to −6.0‰ in speleothems that grew between 131 and 123 ka. An extreme change in $\delta^{13}\text{C}$ values occurs between 200 and 196 ka with a drop from −1.0‰ to −9.4‰, which was immediately followed by a rise to −5.5‰. $\delta^{13}\text{C}$ values also dropped to $\sim -9.5\text{‰}$ during the interval between 177 and 173 ka.

6. Discussion

6.1. Sources of rainfall in northern Negev during the last 200 ka

Tzavoa Cave is located today in a desert environment at the southernmost limit of Mediterranean cyclone tracks. Only a small fraction of the rainfall is of tropical origin from the south [46,47]. Although no speleothem deposition occurs today in the Tzavoa Cave, thick speleothems (with sections of tens of cm) were deposited at intervals during the last 200 ka. Further to the south, speleothem deposition during the last 200 ka was very minor, and when it occurred it is represented by very thin laminae (1–3 cm thick). Tzavoa Cave thus represents the southern limit of thick speleothem deposition in the past 200 ka. This suggests that most of the rainfall that reached Tzavoa Cave in the past originated from the north (i.e. from EM Sea moisture) (Fig. 3). This suggestion is supported by the similar $\delta^{18}\text{O}$ trends (but larger amplitude variations) of the Tzavoa speleothems compared with Soreq Cave speleothems (Fig. 7) for most depositional intervals.

6.2. The minimum amount of annual precipitation required for speleothem deposition

Deposition of speleothems depends on the availability of water in the unsaturated zone. In arid environments, periods of deposition are indicative of water availability in this zone [29,48]. We use the term “humid (or wet) period” for time intervals with a high ratio of precipitation (P) to evaporation (E), allowing more water to enter the unsaturated zone for speleothem deposition. Correspondingly, the term “arid (or dry) period” refers to a low P/E ratio, which reduces the amount of water entering the unsaturated zone, thus causing the reduction or cessation of speleothem growth.

In contrast to the Soreq and Jerusalem caves in central Israel and Peqi'in Cave in the northern Israel where speleothem deposition was continuous during the last 250 ka [24], speleothem deposition in Tzavoa Cave was episodic and mainly occurred during the glacial periods: MIS-6 and MIS-4, 3 and 2, with very short hiatuses (Figs. 4 and 5). During interglacials, speleothem deposition only

occurred for relatively short intervals at ~ 200 –190 ka (MIS-7.1), ~ 137 –123 ka and ~ 118 ka (MIS-6.1–MIS-5.5 transition and MIS-5.5), ~ 96 –92 ka (MIS-5.2), and ~ 84 –77 ka (MIS-5.1).

At present, in the Mediterranean climate zone in northern and central Israel only $\sim 1/3$ of the annual rainfall reaches the unsaturated zone (i.e., about 200–300 mm) and the remaining 2/3 is lost either by evaporation and/or by runoff [43,49]. Despite this loss, the amount of water reaching the unsaturated zone is sufficient to allow speleothem deposition today and throughout the Holocene. The area above Tzavoa cave receives only 150–160 mm rainfall, mostly from the same cyclones, but no speleothem deposition occurs today or occurred in the Holocene. We infer that this lack of speleothem deposition in Tzavoa Cave was due to a negative water budget in the unsaturated zone caused by lower rainfall amounts, evaporation and runoff.

Holocene speleothem deposition was also not found in the Ma'ale-Efrayim cave located in the Jordan Valley “rain shadow” desert [29], although at present this region receives a higher amount of annual rainfall (~ 250 –300 mm) than Tzavoa Cave (Fig. 1B). However, in caves from the Ma'ale Dragot quarry, located within slightly more humid conditions (~ 300 mm of annual rainfall), there is minor deposition during the Holocene as evident from a thin 3.8 ka old outer layer covering older stalactites.

These considerations imply that in present-day and Holocene climate conditions, speleothem deposition occurs only where annual rainfall exceeds ~ 300 –350 mm. Thus, the episodic deposition of speleothems in the northern Negev, compared to the continuous deposition in the Mediterranean climate region in central and northern Israel, suggests that significant influx of water into the unsaturated zone only occurs where rainfall exceeds ~ 300 mm.

This discussion is strictly valid only for climatic conditions similar to present-day and Holocene. However, it is possible that during cool glacial periods the minimum amount of rainfall required for speleothem deposition was lower than 300 mm, because temperatures and evaporation rates were lower, and the frequency of the snow events was higher. The minimum amount of rainfall required for speleothem deposition could be also different than ~ 300 mm if the tropical rainfall arrived from the south during the interglacial episodes.

6.3. Migration of the desert boundary during the humid interglacial intervals

Speleothem deposition in the Tzavoa cave during the interglacial intervals between 200–190 ka, 137–123 ka,

and 84–77 ka, must have been associated with significantly more rainfall than the Holocene. Assuming that temperatures were similar to these of present-day, the source of precipitation was the EM Sea and that a minimum of 300–350 mm of annual rainfall was required for speleothem deposition, the annual rainfall during these short periods was $\sim 100\%$ higher than at present. Higher rainfall is in agreement with the contemporaneous presence of sapropels in the EM Sea [23,24,50].

Supportive evidence for the increase in rainfall comes from the $\delta^{13}\text{C}$ values (Figs. 6 and 7). Vegetation type is one of the most important factors affecting the $\delta^{13}\text{C}$ values of speleothems [17,21,26,42,41,51,52]. Speleothems from the Mediterranean climate region, northern and central Israel, usually have $\delta^{13}\text{C}$ values between $\sim -12\text{‰}$ and $\sim -8\text{‰}$ (average $\sim -10.5\text{‰}$). These ranges reflect dominant C3-Mediterranean type vegetation. In the northern Negev the values range between $\sim 0\text{‰}$ and $\sim -9.0\text{‰}$ (average $\sim -6.0\text{‰}$), and are consistent with C3–C4 mixed vegetation of the steppe and semi-desert. $\delta^{13}\text{C}$ values are even more positive further to the south. This increase in $\delta^{13}\text{C}$ values of large numbers of speleothems from several caves along N–S transect from Mediterranean to arid climate zone demonstrates that vegetation plays a major role in this region on the speleothem $\delta^{13}\text{C}$ values.

At the beginning of humid interglacial interval between 200 and 196 ka (MIS-7.1) the speleothem $\delta^{13}\text{C}$ values in Tzavoa Cave decreased from -1‰ to -9.4‰ . Such a shift is indicative of a rapid change from desert vegetation to Mediterranean steppe-forest with significant fraction of C3 plant species. During other interglacial humid intervals (137–123 ka (MIS-6.1 – MIS-5.5 transition and MIS-5.5), and 78–76 ka (MIS-5.1)) the $\delta^{13}\text{C}$ values were -6‰ to -7‰ , consistent with mixed C3–C4 vegetation of steppe proximate to the border of Mediterranean climate zone.

It is concluded that the desert boundary migrated south of Tzavoa Cave during these short interglacial time intervals, (Fig. 8). The increase in rainfall in the northern Negev most probably occurred due to southward shift of the Mediterranean cyclone tracks. Such an analogue we see in the present-day during extremely rainy winters [53]. It is also possible that there was a greater contribution of moisture from tropical Atlantic and Indian Oceans, because the simultaneous increase in monsoon activity has been demonstrated in Oman speleothems [8], by sharp decreases in the salinity of the Red Sea [54], the existence of high lake stands in southern Sahara Desert [6,7,10,12], and the formation of iron ores, travertines and speleothems in the Western desert, Egypt [9].

Northern Negev humid interglacial intervals coincide with monsoon maxima at 198 ka, 127 ka, and 83 ka [55].

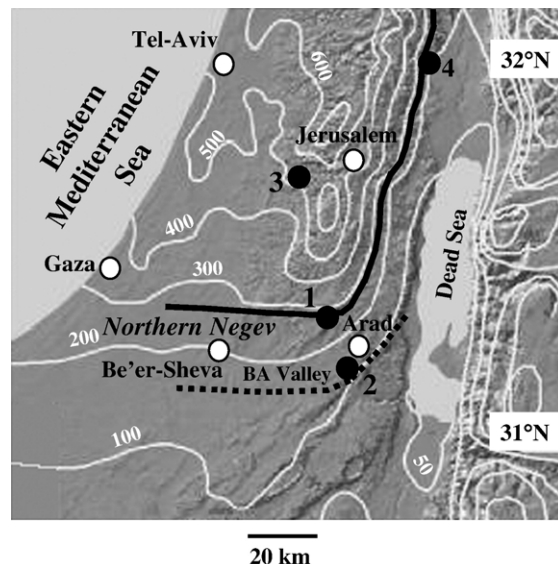


Fig. 8. Map showing the two positions of the southern boundary of speleothem deposition identified in this study: Solid line: the dry episodes between 150–144 ka, 141–140 ka, 117–96 ka, 92–85 ka, 25–23 ka, and 13–0 ka. Dotted line: humid (wet) periods at 200–150 ka, 137–123 ka, 96–92 ka, 85–25 ka, 23–13 ka. Other symbols are as in Fig. 1C.

Humid conditions also occurred in southern and central parts of Saharo–Arabian Desert during these periods [8]. However, during the monsoon maxima at 105 ka and the Holocene (11 ka), humid conditions prevailed in southern and central Saharo–Arabian Desert [8,56], whereas northern Negev was arid. This suggests that during interglacials there is not always a linkage between monsoon activity and humidity in the northern boundary of the Saharo–Arabian Desert.

6.4. Migration of the desert boundary during the glacial periods

Most speleothem formation in the Tzavoa Cave occurred during the two last glacial periods: speleothem deposition was continuous between 190 ka and 150 ka, and between 76 ka and 13 ka except for short hiatus between 25 ka and 23 ka. A similar pattern of increased humidity occurred during the two last glacial periods in Ma'ale–Efrayim cave [29]. Speleothem deposition in both caves probably became possible because of increased precipitation and lowering of the temperatures by 5–10 °C during glacial periods [28,45], resulting in a higher P/E ratio. This indicates that during cool glacial intervals the desert boundary migrated southward compared with its present-day position and caused Mediterranean climate conditions to dominate in the Jordan Valley and northern Negev (Fig. 8).

The average $\delta^{13}\text{C}$ values of Tzavoa Cave speleothems during these periods are of -6‰ indicative of mixed C3–C4 (mostly C4) steppe vegetation proximate to the border Mediterranean climate zone (Fig. 7). Lower values reaching $\sim -9.5\text{‰}$ at ~ 176 ka, and ~ 76 ka suggest that for short intervals Mediterranean C3 type vegetation invaded from the north. Very high $\delta^{13}\text{C}$ values reaching $\sim -1\text{‰}$ to -3‰ at ~ 58 ka, 30–25 ka, and ~ 21 ka are reflecting scarce vegetation due to arid conditions.

Periods of enhanced monsoonal activity in the southern Saharo–Arabian Desert [9] during glacials at 176 ka, 151 ka, 61 ka and 33 ka [55] also coincide with humid periods in the northern Negev. However, even when the monsoonal activity was minimal, the northern Negev region was usually humid. Thus, during glacials monsoon activity and humidity at the northern boundary of the Saharo–Arabian Desert are not always linked, consistent with the Mediterranean Sea being the main humidity source in this region.

6.5. Comparison of the speleothem age frequencies with the levels of the paleo-lake Lisan

Lake Lisan (the precursor of the present-day Dead-Sea) existed during most of the last glacial from ~ 70 ka to ~ 15 ka, and received most of its water from northern Israel via the Jordan river. During most of this period the lake level stand was ~ -265 m, but dropped to ~ -325 and -320 m at ~ 46 and 40 ka, respectively and reached its maximum height of ~ -160 m at 26 ka. From ~ 25 ka the lake level gradually decreased to ~ -400 m for most of the Holocene [57,58]. The period of speleothem deposition in Tzavoa Cave during the last glacial period coincides with the existence of Lake Lisan. The most striking correlation is the sharp drop in the lake level after 13 ka, coinciding with the cessation of speleothem deposition in Tzavoa Cave. Although there is no simple positive co-variance between the highest relative frequencies of speleothem ages and the highest lake levels, the simultaneous speleothem deposition and occurrence of high levels of Lake Lisan indicate that the Mediterranean rain systems shifted southward during the last glacial period. This shift brought about increased precipitation associated with cool temperatures, more frequent snowfall during the winter and low evaporation, which increased infiltration of water to the caves, aquifers and lakes.

6.6. Oxygen isotope evolution of the rainfall in northern Negev during the last 200 ka

Comparison of the $\delta^{18}\text{O}$ values of the Tzavoa and Soreq Cave speleothems (Fig. 7) reveals similar general

trends, but that Tzavoa Cave speleothems are usually depleted by approximately $\sim 1\text{--}2\text{‰}$ relative to Soreq Cave samples, with an even larger depletion of $\sim 4\text{‰}$ between 173 and 177 ka. Assuming that the origin of rainfall to both desert regions is the EM Sea, the large depletion in $\delta^{18}\text{O}$ of Tzavoa Cave speleothems requires explanation.

Two possible explanations arise. One is that Tzavoa speleothems were deposited at higher temperatures than at Soreq cave. During the last glacial period, temperatures in the northern Negev could have been higher than in central and northern Israel owing to much steeper north–south temperature gradients. These could have been the result of extremely cold temperatures in Europe and Turkey, which were under Polar high pressure. Contemporaneous high temperatures could have developed in Saharo–Arabian Desert affected by subtropical high pressure. Two reasons partially argue against this scenario: a) it does not explain why the $\delta^{18}\text{O}$ values of Tzavoa Cave speleothems were depleted by similar values during the humid interglacial intervals; b) based on the offset in the $\delta^{18}\text{O}$ values, the north–south temperature gradient was $6\text{--}8\text{ °C}$ over the ~ 100 km distance between Soreq and Tzavoa caves, which is very unlikely.

A second explanation is that the rain above the Tzavoa Cave was more depleted in heavy isotopes due to Rayleigh fractionation effects. Dansgaard [59] showed that Rayleigh isotopic fractionation during rainfall precipitation results in progressive lowering of δD and $\delta^{18}\text{O}$ values of vapor as the result of increased rainout as a function of distance from the source and cooling due to higher elevation. The rainwater $\delta^{18}\text{O}$ depletion due to elevation in Mediterranean basin varies between -0.1‰ and $-0.6\text{‰}/100$ m, while the most common values are $\sim -0.2\text{--}0.25\text{‰}/100$ m [60].

A simple calculation assessing the viability of a Rayleigh process as an explanation of the $\delta^{18}\text{O}$ offset is made using the present-day rainfall values. Present-day $\delta^{18}\text{O}$ of atmospheric vapor above the EM during the winter varies from -11‰ to -18.6‰ , with an average $\delta^{18}\text{O} \sim -13.5\text{‰}$ [61]. Assuming condensation temperatures of 0 °C , 5 °C and 10 °C , the calculated initial rainfall $\delta^{18}\text{O}$ values above the EM Sea are -1.9‰ , -2.3‰ and -2.7‰ , respectively. Using these initial values and the Rayleigh fractionation equations given by Criss [62], p.108, we then calculate the $\delta^{18}\text{O}$ values of water and vapor as a function of F , the fraction of vapor remaining in the clouds. These are shown in Fig. 9. Estimates of the fraction of rainfall removed from the clouds as they advance inland are obtained by inserting the rainfall $\delta^{18}\text{O}$ values at different sites in Israel [20]. Adjacent to the coastline in central Israel, the rainfall $\delta^{18}\text{O}$ values range between $\sim -4.0\text{‰}$ and -4.5‰ , requiring $\sim 15\%$ rainout.

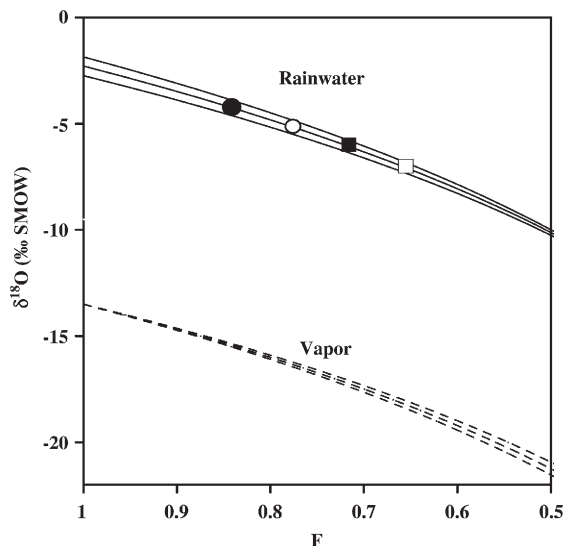


Fig. 9. Possible changes in $\delta^{18}\text{O}$ (SMOW) ‰ of meteoric water due to Rayleigh fractionation of rainfall (continuous lines) from vapor (broken lines). The changes in the isotopic composition are shown for condensation temperatures of 0 °C, 5 °C and 10 °C. F is the fraction of the water remaining in the clouds as they advance further from the source. The $\delta^{18}\text{O}$ data of the present-day rainfall is shown as follows: Black circle — EM Sea coast; open circle — Soreq Cave area; black rectangle — Jerusalem; open rectangle — Neve-Ativ (NA).

40 km away from the coastline, ~400 m asl (Soreq cave area) the average rainfall $\delta^{18}\text{O}$ is ~−5.0‰ to ~−5.5‰, corresponding to 20–22% rainout, 55–60 km from the coast and 700–800 m asl (Jerusalem), the rainfall $\delta^{18}\text{O}$ is ~−6‰ to ~−6.5‰ giving ~25–30% rainout, and in Neve-Ativ, northern Israel (Fig. 1B), 45 km from the coast but at an elevation of 1000 m asl, $\delta^{18}\text{O}$ values of rainfall are ~−7.0‰ to ~−7.5‰, requiring ~35% rainout. Thus, the variability of present-day rainfall $\delta^{18}\text{O}$ values from various sites in Israel can be attributed to moderate Rayleigh fractionation, and mainly to changes in elevation with a less marked distance affect.

Tzavoa cave is located ~80 km from the shore-line and 550 m asl and is slightly more inland and higher than at the Soreq Cave site (45 km and 400 m elevation). Thus, from the point of view of the Rayleigh model, a slight depletion with respect to the Soreq cave is to be anticipated. It is clear, however, that the Rayleigh effect cannot account for all the 1–2‰ $\delta^{18}\text{O}$ offset. Thus, at present it seems that a combination of both enhanced temperature gradient and Rayleigh fractionation provides the best explanation, particularly for glacial intervals, where the offset is most marked. During the brief interglacial intervals, an enhanced component of low $\delta^{18}\text{O}$ rainfall of tropical origin cannot be ruled out.

7. Conclusions

The northern Negev is located in a transition zone between Mediterranean climate in the north and arid climate in the south, with average annual precipitation of 350 to 150 mm (i.e., semi-arid to arid) with typical vegetation changing from C3 Mediterranean steppe-forest to C3+C4 Irano-Turanian semi-desert vegetation. During the last 200 ka this region experienced wetter conditions both during interglacials and glacials at 200–150 ka, 137–123 ka, 85–25 ka, and 23–13 ka.

The dominant source of precipitation throughout these wet intervals was the EM Sea, with a possible small contribution of the southern tropical sources during the short humid interglacial intervals. When the rainfall during the interglacial intervals was of the first origin, the minimum annual rainfall must have been ~300–350 mm, i.e., about a twice than at present. During most of the last glacial and parts of the previous glacial, the region was wetter; however, the minimum amount of precipitation required for speleothem deposition could have been less than 300 mm due to lower temperatures resulting in more frequent snowfall and lower evaporation. Coinciding with the increase in precipitation, Mediterranean C3 type vegetation invaded southward for short periods, and the climate in the northern Negev became more Mediterranean, but still remained mixed C3+C4 steppe most of the time. The climate was arid in the northern Negev during interglacial and glacial intervals when no speleothem deposition occurred: 150–144 ka, 141–140 ka, 117–96 ka, 92–85 ka, 25–23 ka, and 13 ka to present-day. The relatively depleted $\delta^{18}\text{O}$ values of Tzavoa Cave speleothems relative to Soreq cave could in part reflect Rayleigh distillation effects, especially if they were enhanced by a more marked north–south temperature gradient during glacial periods. There is a general correlation between the period of intensive speleothem deposition in caves located in the northern Negev desert and in the “rain shadow” desert, with the period of existence of paleo-lake Lisan during the last glacial.

Humid periods found during interglacials in the northern Negev coincide with monsoon maxima and with humid conditions in southern and central parts of Saharo-Arabian Desert at 198 ka, 176 ka, 151 ka, 127 ka, and 83 ka, 61 ka and 33 ka. However, during monsoon maxima at 105 ka and 11 ka, humid conditions occurred in the southern and central Saharo-Arabian Desert, but the northern Negev was dry. During the glacial monsoonal minima the northern Negev was usually humid, and the southern parts of Saharo-Arabian Desert were dry. This implies that there is not always synchronicity between

monsoon activity and humidity at the northern boundary of the Saharo Arabian Desert, which is consistent with the major source of humidity in the region being the Mediterranean Sea.

Acknowledgements

This research was supported by the Israel Science Foundation (grant No. 910/05). We would like to thank: N. Tepliakov and I. Segal for the help and advice with chemical analyses and with the MC-ICP-MS; E. Vaks and E. Reznik-Vaks, Z. Wiener, and the Bloch family for the help with rainwater sampling; A. Sandler for the XRD analyses; M. Dvorchek for guidance in SEM analyses; S. Ashkenazi, E. Ram, and members of the Cave Research Unit at the Hebrew University for help with field work; the Israeli Nature and Parks Authority for permission to sample speleothems. Special thanks to Y. Enzel and B. Ziv for fruitful discussions and L. Laor, E. Eliani and U. Simchai for their help with laboratory work.

Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at [10.1016/j.epsl.2006.07.009](https://doi.org/10.1016/j.epsl.2006.07.009).

References

- [1] N. Roberts, H.E. Wright, Vegetational, lake level, and climatic history of the Near East and southwest Asia, *Global Climate Scenes at Last Glacial Maximum*, 1993, pp. 194–220.
- [2] T.M.L. Wigley, G. Farmer, Climate of Eastern Mediterranean and Near East, *Paleoclimates, Paleoenvironments and Human Communities in the Eastern Mediterranean Region in Later Prehistory*, 1982, pp. 3–37.
- [3] M. Rindsberger, M. Magaritz, I. Carmi, D. Gilad, The relation between air mass trajectories and the water isotope composition of rain in the Mediterranean Sea area, *Geophys. Res. Lett.* 10 (1983) 43–46.
- [4] G. Eshel, Mediterranean climates, *Isr. J. Earth-Sci.* 51 (2002).
- [5] M. Rossignol-Strick, Mediterranean Quaternary sapropels, an immediate response of the African monsoon to variation of insolation, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 49 (1985) 237–263.
- [6] G.H. Miller, F. Wendorf, R. Ernst, R. Schild, A.E. Close, I. Friedman, H.P. Schwarcz, Dating lacustrine episodes in the eastern Sahara by the epimerization of isoleucine in ostrich eggshells, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 84 (1991) 175–189.
- [7] B.J. Szabo, J.C.V. Haynes, T.A. Maxwell, Ages of Quaternary pluvial episodes determined by uranium-series and radiocarbon dating of lacustrine deposits of Eastern Sahara, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 113 (1995) 227–242.
- [8] D. Fleitmann, S.J. Burns, U. Neff, A. Mangini, A. Matter, Changing moisture sources over the last 330,000 years in Northern Oman from fluid-inclusion evidence in speleothems, *Quat. Res.* 60 (2003) 223–232.
- [9] J.K. Osmond, A.A. Dabous, Timing and intensity of groundwater movement during Egyptian Sahara pluvial periods by U-series analysis of secondary U in ores and carbonates, *Quat. Res.* 61 (2004) 85–94.
- [10] N. Petit-Maire, The Sahara in the Holocene, Map, UNESCO and COMW, IGN (eds.), Aux-an-Provence, France (1993).
- [11] M.K. Crombie, R.E. Arvidson, N.C. Sturchio, Z. El Alf, K.A. Zeid, Age and isotopic constraints on Pleistocene pluvial episodes in the Western Desert, Egypt, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 130 (1997) 337–355.
- [12] P. deMenocal, J. Ortiz, T. Guilderson, J. Adkins, M. Sarnthein, L. Baker, M. Yarusinsky, Abrupt onset and termination of the African humid period: rapid climate responses to gradual insolation forcing, *Quat. Sci. Rev.* 19 (2000) 347–361.
- [13] J.C. Fontes, F. Gasse, On the ages of humid Holocene and late Pleistocene phases in North Africa; remarks on “Late Quaternary climatic reconstruction for the Maghreb (North Africa)” by P. Rognon, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 70 (1989) 393–398.
- [14] H.M. El-Asmar, Quaternary isotope stratigraphy and paleoclimate of coral reef terraces, Gulf of Aqaba, South Sinai, Egypt, *Quat. Sci. Rev.* 16 (1997) 911–924.
- [15] C. Causse, B. Ghaleb, N. Chkir, K. Zouari, H. Ben Oueddou, A. Mamou, Humidity changes in southern Tunisia during the Late Pleistocene inferred from U–Th dating of mollusc shells, *Appl. Geochem.* 18 (2003) 1691–1703.
- [16] N. Petit-Maire, P. Sanlaville, A. Abed, S. Yasin, R. Bourrouilh, P. Carbonel, M. Fontugne, J.L. Reyss, New data for an Eemian lacustrine phase in southern Jordan, *Episodes* 25 (2002) 279–280.
- [17] H.P. Schwarcz, B. Blackwell, P. Goldberg, A.E. Marks, Uranium series dating of travertine from archaeological sites, Nahal Zin, Israel, *Nature* 277 (1979) 558–560.
- [18] A. Livnat, J. Kronfeld, Paleoclimatic implications of U-series dates for lake sediments and travertines in the Arava Rift Valley, Israel, *Quat. Res.* 24 (1985) 164–172.
- [19] G.A. Goodfriend, M. Magaritz, Palaeosols and late Pleistocene rainfall fluctuations in the Negev Desert, *Nature* 332 (1988) 144–146.
- [20] A. Ayalon, M. Bar-Matthews, A. Kaufman, Climatic conditions during marine isotope stage 6 in the eastern Mediterranean region from the isotopic composition of the Soreq Cave, Israel, *Geology* 30 (2002) 303–306.
- [21] M. Bar-Matthews, A. Ayalon, A. Kaufman, Late Quaternary paleoclimate in the Eastern Mediterranean region from stable isotope analysis of speleothems at Soreq Cave, Israel, *Quat. Res.* 47 (1997) 155–168.
- [22] M. Bar-Matthews, A. Ayalon, A. Kaufman, Paleoclimate evolution in the Eastern Mediterranean region during the last 58,000 years as derived from the stable isotopes of speleothems (Soreq Cave, Israel), *Isotope Techniques in the Study of the Environmental Change*, 1998, pp. 673–682.
- [23] M. Bar-Matthews, A. Ayalon, A. Kaufman, Timing and hydrological conditions of Sapropel events in the Eastern Mediterranean, as evident from speleothems, Soreq cave, Israel, *Chem. Geol.* 169 (2000) 145–156.
- [24] M. Bar-Matthews, A. Ayalon, M. Gilmour, A. Matthews, C.J. Hawkesworth, Sea-land oxygen isotopic relationships from planktonic foraminifera and speleothems in the Eastern Mediterranean region and their implication for paleorainfall during interglacial intervals, *Geochim. Cosmochim. Acta* 67 (2003) 3181–3199.

- [25] A. Frumkin, D.C. Ford, H.P. Schwarcz, Continental oxygen isotopic record of the last 170,000 years in Jerusalem, *Quat. Res.* 51 (1999) 317–327.
- [26] A. Frumkin, D. Ford, H. Schwarcz, Paleoclimate and vegetation of the last glacial cycles in Jerusalem from a speleothem record, *Global Biochem. Cycles* 14 (2000) 863–870.
- [27] A. Matthews, A. Ayalon, M. Bar-Matthews, D/H ratios of fluid inclusions of Soreq cave (Israel) speleothems as a guide to the Eastern Mediterranean Meteoric Line relationships in the last 120 ky, *Chem. Geol.* 166 (2000) 183–191.
- [28] S. McGarry, M. Bar-Matthews, A. Matthews, A. Vaks, B. Schilman, A. Ayalon, Constraints on hydrological and paleo-temperature variations in the Eastern Mediterranean region in the last 140 ka given by the δD values of speleothem fluid inclusions, *Quat. Sci. Rev.* 23 (2004) 919–934.
- [29] A. Vaks, M. Bar-Matthews, A. Ayalon, B. Schilman, M. Gilmour, C.J. Hawkesworth, A. Frumkin, A. Kaufman, A. Matthews, Paleoclimate reconstruction based on the timing of speleothem growth and oxygen and carbon isotope composition in a cave located in the rain shadow in Israel, *Quat. Res.* 59 (2003) 182–193.
- [30] J.C. Vogel, A. Fuls, A. Danin, Geographical and environmental distribution of C3 and C4 grasses in Sinai, Negev, and Judean deserts, *Oecologia* 70 (1986) 258–265.
- [31] G.A. Goodfriend, Rainfall in the Negev Desert during the middle Holocene, based on (super 13) C of organic matter in land snail shells, *Quat. Res.* 34 (1990) 186–197.
- [32] G.A. Goodfriend, Terrestrial stable isotope records of Late Quaternary paleoclimates in the eastern Mediterranean region, *Quat. Sci. Rev.* 18 (1999) 501–513.
- [33] T.E. Cerling, J. Quade, Stable carbon and oxygen isotopes in soil carbonates, climate change in continental isotopic records, *Geophys. Monogr.* 78 (1993) 217–231.
- [34] N. Feinbrun-Dothan, A. Danin, Analytical flora of Eretz-Israel, CANA Publishing House, Jerusalem, Israel, 1998, pp. 1–1008.
- [35] U. Dayan, Climatology of back trajectories from Israel based on synoptic analysis, *J. Clim. Appl. Meteorol.* 25 (1986) 591–595.
- [36] A. Zangvil, P. Druian, Upper air trough axis orientation and the spatial distribution of rainfall over Israel, *Int. J. Climatol.* 10 (1990) 57–62.
- [37] Y. Itzhaki, Pleistocene shore-lines in the coastal plane of Israel, *Geol. Surv. Isr., Bull.* 32 (1961) 1–9.
- [38] A. Avital, Geological history of the Plio-Pleistocene–Holocene offshore based on cores off Ashqelon, southern Israel., M. Sc. thesis, Ben Gurion University of the Negev, 2003.
- [39] F. McDermott, D.P. Mathey, C. Hawkesworth, Centennial-scale Holocene climate variability revealed by a high-resolution speleothem δ (super 18) O record from SW Ireland, *Science* 294 (2001) 1328–1331.
- [40] A. Kaufman, G.J. Wasserburg, D. Porcelli, M. Bar-Matthews, A. Ayalon, L. Halicz, U–Th isotope systematics from the Soreq cave, Israel and climatic correlations, *Earth Planet. Sci. Lett.* 156 (1998) 141–155.
- [41] C.H. Hendy, The isotopic geochemistry of speleothems-I. The calculation of the effects of different modes of formation on the isotopic composition of speleothems and their applicability as palaeoclimatic indicators, *Geochim. Cosmochim. Acta* 35 (1971) 801–824.
- [42] M. Bar-Matthews, A. Ayalon, A. Matthews, E. Sass, L. Halicz, Carbon and oxygen isotope study of the active water-carbonate system in a karstic Mediterranean cave: implications for paleoclimate research in semiarid regions, *Geochim. Cosmochim. Acta* 60 (1996) 337–347.
- [43] A. Ayalon, M. Bar-Matthews, E. Sass, Rainfall–recharge relationships within a karstic terrain in the Eastern Mediterranean semi-arid region, Israel: δ 18O and δ D characteristics, *J. Hydrol.* 207 (1998) 18–31.
- [44] K.R. Ludwig, Isoplot 3.00, A Geochronological Toolkit for Microsoft Excel, Berkeley Geochronology Center, Special Publication, vol. 4, 2003.
- [45] K.-C. Emeis, T. Sakamoto, R. Wehausen, H.-J. Brumsack, The sapropel record of the eastern Mediterranean Sea — results of Ocean Drilling Program Leg 160, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 158 (2000) 371–395.
- [46] U. Dayan, R. Abramski, Heavy rain in the Middle East related to unusual jet stream properties, *Bull. Am. Meteorol. Soc.* 64 (1983) 1138–1140.
- [47] B. Ziv, A subtropical rainstorm associated with tropical plume over Africa and the Middle-East, *Theor. Appl. Climatol.* 69 (2001) 91–102.
- [48] L.K. Ayliffe, P.C. Marianelli, K.C. Moriarty, R.T. Wells, M.T. McCulloch, G.E. Mortimer, J.C. Hellstrum, 500 ka precipitation record from southeastern Australia: evidence for interglacial relative aridity, *Geology* 26 (1998) 147–150.
- [49] A. Ayalon, M. Bar-Matthews, B. Schilman, Rainfall isotopic characteristics at various sites in Israel and the relationships with unsaturated zone water, Israel Geological Survey Report GSI/16/04, 2004, 37 pp.
- [50] N. Kallel, J.C. Duplessy, L. Labeyrie, M. Fontugne, M. Paterne, M. Montacer, Mediterranean pluvial periods and sapropel formation over the last 200,000 years, *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 157 (2000) 45–58.
- [51] D. Genty, M. Massault, Carbon transfer dynamics from bomb-14C and δ 13C time series of a laminated stalagmite from SW France—modelling and comparison with other stalagmite records, *Geochim. Cosmochim. Acta* 63 (1999) 1537–1548.
- [52] D. Genty, A. Baker, M. Massault, C. Proctor, M. Gilmour, E. Pons-Branchu, B. Hamelin, Dead carbon in stalagmites: carbonate bedrock paleodissolution vs. ageing of soil organic matter. Implications for 13C variations in speleothems, *Geochim. Cosmochim. Acta* 65 (2001) 3443–3457.
- [53] Y. Enzel, R. Bookman, D. Sharon, H. Gvirtzman, U. Dayan, B. Ziv, M. Stein, Late Holocene climates of the Near East deduced from Dead Sea level variations and modern regional winter rainfall, *Quat. Res.* 60 (2003) 263–273.
- [54] A. Almogi-Labin, C. Hemleben, D. Meischner, Carbonate preservation and climatic changes in the central Red Sea during the last 380 kyr as recorded by pteropods, *Mar. Micropaleontol.* 33 (1998) 87–107.
- [55] M. Rossignol-Strick, M. Paterne, A synthetic pollen record of the eastern Mediterranean sapropels of the last 1 Ma: implications for the time-scale and formation of sapropels, *Mar. Geol.* 153 (1999) 221–237.
- [56] K. Nicoll, Recent environmental change and prehistoric human activity in Egypt and Northern Sudan, *Quat. Sci. Rev.* 23 (2004) 561–580.
- [57] Y. Bartov, M. Stein, Y. Enzel, A. Agnon, Z. Reches, Lake levels and sequence stratigraphy of Lake Lisan, the Late Pleistocene precursor of the Dead Sea, *Quat. Res.* 57 (2002) 9–21.
- [58] Y. Bartov, S.L. Goldstein, M. Stein, Y. Enzel, Catastrophic arid episodes in the Eastern Mediterranean linked with the North Atlantic Heinrich events, *Geology* 31 (2003) 439–442.
- [59] W. Dansgaard, Stable isotopes in precipitation, *Tellus* 16 (1964) 438–468.

- [60] L. Gourcy, A. Aouad-Rizk, L. Araguas, A. Argiriou, P. Bono, M.F. Diaz-Teijeiro, A. Dirican, A.M. El-Asraq, J. Gat, R. Gonfiantini, N. Horvatincic, B. Ouda, P. Carreira-Paquete, D. Rank, O. Saighi, Y. Travi, P. Vreca, Isotopic composition of precipitation in relation to air circulation patterns in the Mediterranean basin: preliminary results, International Workshop on the Application of Isotope Techniques in Hydrological and Environmental Studies, 2004.
- [61] J. Gat, B. Klein, Y. Kushnir, W. Roether, H. Wernli, R. Yam, A. Shemesh, Isotope composition of air moisture over the Mediterranean Sea: an index of the air–sea interaction pattern, *Tellus* 55B (2003) 953–965.
- [62] R.E. Criss, Principles of Stable Isotope Distribution, Oxford University Press, NY, 1999.
- [63] J.K. Hall, V.A. Krashennikov, F. Hirsh, C. Benjamini, A. Flexer, Geological Framework of the Levant, Volume II: The Levantine Basin and Israel, 2005, pp. 1–826.

Desert speleothems reveal climatic window for African exodus of early modern humans

Anton Vaks Institute of Earth Sciences, Hebrew University of Jerusalem, Jerusalem 91904, Israel, and
Geological Survey of Israel, 30 Malkhei Israel St., Jerusalem 95501, Israel

Miriam Bar-Matthews }
Avner Ayalon } Geological Survey of Israel, 30 Malkhei Israel St., Jerusalem 95501, Israel

Alan Matthews Institute of Earth Sciences, Hebrew University of Jerusalem, Jerusalem 91904, Israel

Ludwik Halicz Geological Survey of Israel, 30 Malkhei Israel St., Jerusalem 95501, Israel

Amos Frumkin Department of Geography, Hebrew University of Jerusalem, Jerusalem 91905, Israel

ABSTRACT

One of the first movements of early modern humans out of Africa occurred 130–100 thousand years ago (ka), when they migrated northward to the Levant region. The climatic conditions that accompanied this migration are still under debate. Using high-precision multicollector-inductively coupled plasma-mass spectrometry (MC-ICP-MS) U-Th methods, we dated carbonate cave deposits (speleothems) from the central and southern Negev Desert of Israel, located at the northeastern margin of the Saharan-Arabian Desert. Speleothems grow only when rainwater enters the unsaturated zone, and this study reveals that a major cluster of wet episodes (the last recorded in the area) occurred between 140 and 110 ka. This episodic wet period coincided with increased monsoonal precipitation in the southern parts of the Saharan-Arabian Desert. The disappearance at this time of the desert barrier between central Africa and the Levant, and particularly in the Sinai-Negev land bridge between Africa and Asia, would have created a climatic “window” for early modern human dispersion to the Levant.

Keywords: Negev Desert, speleothems, U-Th dating, paleoclimate, out of Africa.

INTRODUCTION

The African origin of the early modern humans ca. 200–150 ka is now well documented (McBrearty and Brooks, 2000; McDougall et al., 2005), but the routes of their expansion out of Africa are still debated (Vermeersch, 2001; Petraglia and Alsharekh, 2003). Fossil remains of early modern humans and sites of African Middle Stone Age industry in the eastern Sahara (McBrearty and Brooks, 2000; Smith et al., 2004a) and Arabian Peninsula (Petraglia and Alsharekh, 2003; Rose, 2004) suggest a migration route from tropical east Africa to the north and northeast. (The African terminology, the Middle Stone Age, is used in this paper to cover the same time period described in Europe as the Middle Paleolithic; Marean and Assefa, 2005.) Archaeological data from northern Israel indicate that one of the major waves of early modern human expansion out of the African continent occurred between 130 and 100 ka (Schwarcz et al., 1988; Valladas et al., 1988; Mercier et al., 1993; Bar-Yosef, 1998; Grun et al., 2005). All possible migration routes leading from Africa to the Levantine early modern human sites (Fig. 1A) converge in the arid to hyper-arid Negev, Sinai, and southern Jordan Deserts (Fig. 1B), making it a key region for understanding climatic constraints on early modern human dispersal. Derricourt (2005) argued that the Sinai-Negev Desert route was the major (and possibly the only) way out of Africa, and that the passage through this “bottleneck” region

was dependent on the development of suitable climate conditions. The southern and central Saharan-Arabian Desert experienced increased monsoon precipitation during this period of early modern human emergence (Szabo et al., 1995; Rohling et al., 2002; Fleitmann et al., 2003; Osmond and Dabous, 2004), but it is not known if an increase in rainfall also occurred in the northern part of the migration corridor. Here, we present evidence for a period of enhanced rainfall activity between 140 and 110 ka in the central and southern Negev Desert, Israel, based on absolute U-Th dating of speleothems. The climate during this period consisted of episodic wet events that enabled the deserts of the northeastern Sahara, Sinai, and Negev to become more suitable for the movement of early modern humans. This period was preceded and followed by essentially unbroken arid conditions, which may indicate that climate change had a major limiting role in the timing of early modern human dispersal out of Africa.

GEOLOGICAL CONTEXT

The caves of the Negev Desert are dry today and speleothems do not form, but their presence in a number of caves clearly indicates that water reached the unsaturated zone in the past and that surface vegetation was present at times. Today, the mildly arid northern (300–150 mm/yr) and arid central parts of the Negev Desert (150–50 mm/yr) mainly receive their rainfall from Mediterranean cyclones (Dayan, 1986), as in the

Mediterranean climate region of northern and central Israel. In contrast, the southern Negev is hyperarid (<50 mm/yr) (Amit et al., 2006) (Fig. 1B), and its rainfall mostly consists of rare thunderstorms bringing moisture of tropical origin (Kahana et al., 2002). The transition from the Mediterranean conditions in central Israel to the hyperarid desert in the southern Negev occurs over a distance of less than 150 km. The arid and hyperarid Sinai-Negev Deserts form the land bridge that links Africa to Asia.

Speleothem deposition has been continuous during the last 200 k.y. in the Mediterranean climate region of northern and central Israel (Frumkin et al., 1999; Bar-Matthews et al., 2003), whereas in the mildly arid steppe zone in the Jordan Valley and the northern Negev, speleothem deposition has been more sporadic, with hiatuses during dry interglacial episodes (marine oxygen isotope stage [MIS] 5.3–5.2, and Holocene) and the two last glacial maxima (Vaks et al., 2006). Further south, where present-day precipitation is below 150 mm/yr (central and southern Negev), the cross sections of speleothems younger than 200 ka are reduced to 1–3 cm, compared with tens of centimeters in the northern Negev (Vaks et al., 2006).

METHODS

Eleven speleothems (stalactites, stalagmites, and flowstone, see GSA Data Repository Table DR1¹) were collected from five caves located along a north-south transect of the central and southern Negev Desert, from the present-day 150 mm isohyet to the 30 mm isohyet (Fig. 1B). The caves occur in Middle Cretaceous limestone and dolomite host rock. The speleothems were sectioned, and up to 500 mg of powder was drilled out from different laminae using 0.8–1.2-mm-diameter drill bits. Stable isotope tests (Hendy, 1971) performed on several of the speleothems indicate equilibrium deposition. The procedures for accurate U-Th dating using

¹GSA Data Repository item 2007202, Table DR1 and Figure DR1, is available online at www.geosociety.org/pubs/ft2007.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

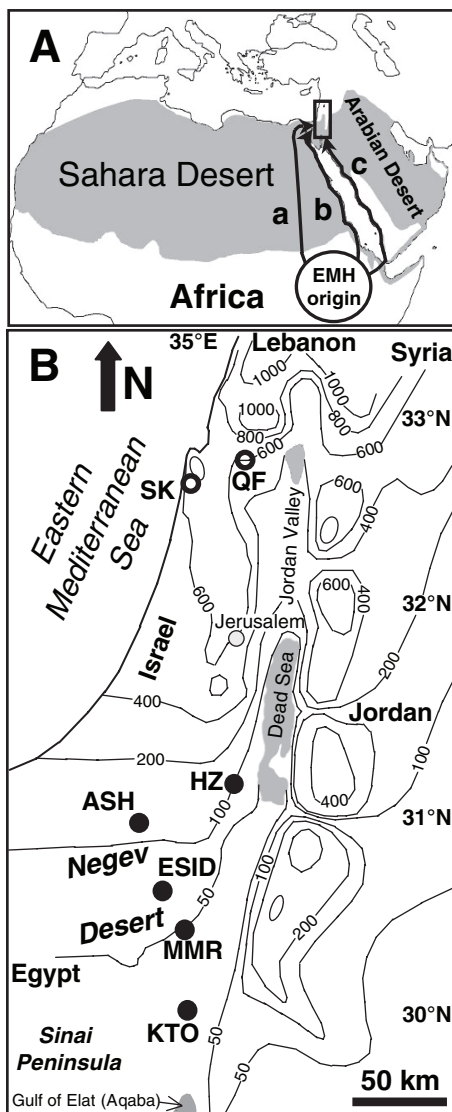


Figure 1. Location maps. **A:** Saharan-Arabian Desert (gray shading indicates rainfall below 100 mm) and three possible routes for migration of early modern humans (EMH) toward Levant early modern human sites, passing through desert areas of Sinai, Negev, and southern Jordan: (a) along Nile River, (b) along African Red Sea coast, (c) along Arabian Red Sea coast after crossing Bab-el-Mandeb Straits (after Petraglia and Alsharekh, 2003; Rose, 2004). Rectangle indicates southern Levant area shown in Figure 1B. **B:** Southern Levant area showing location of five studied cave sites in Negev Desert. Caves are denoted by black circles and labeled as follows: HZ—Hol-Zakh; ASH—Ashalim; ESID—Even-Sid; MMR—Ma'ale-ha-Meyshar; KTO—Ktora Cracks. Open circles indicate caves with early modern human remains in northern Israel: Skhul (SK) and Qafzeh (QF). Lines represent rainfall isohyets (in mm/yr). Present-day Negev Desert can be divided into three subzones: northern—300 mm to 150 mm isohyets; central—150 mm to 50 mm isohyets; southern—50 mm isohyet to Gulf of Elat (Aqaba).

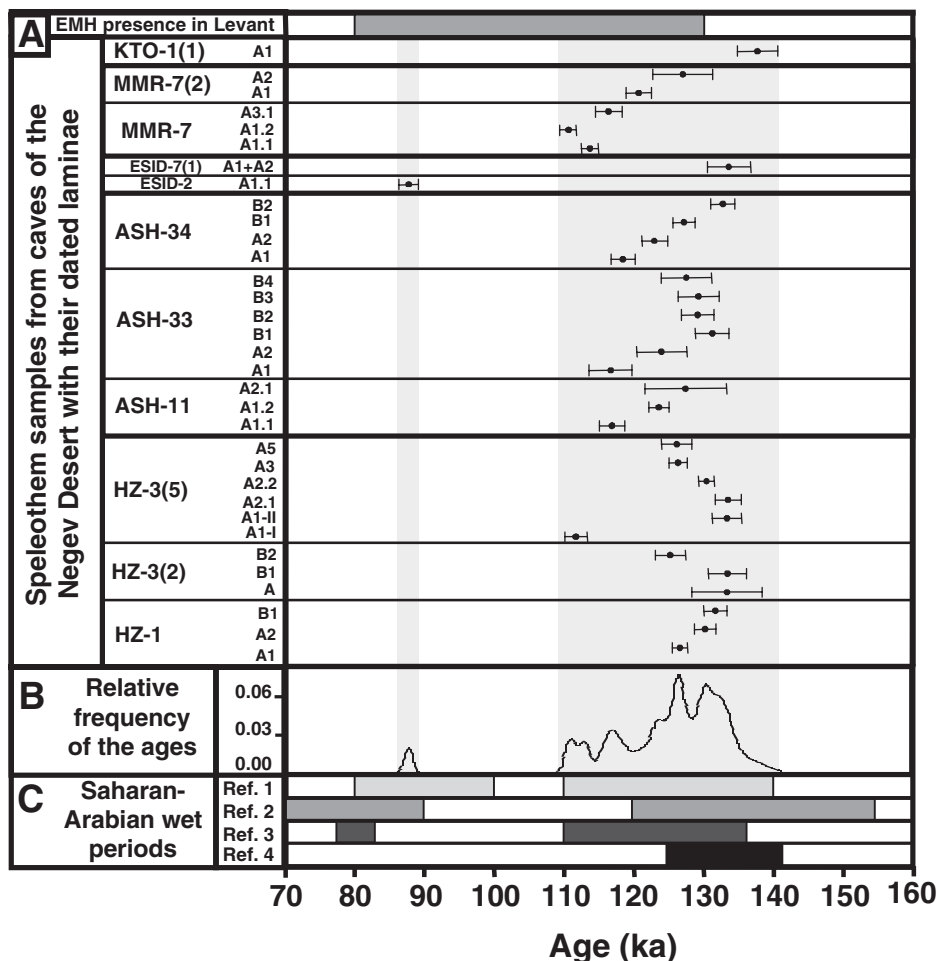


Figure 2. Speleothem ages and enhanced rainfall periods in Negev Desert determined in this study compared to data on occupation of caves in northern Israel by early modern humans (EMH) and periods of enhanced monsoonal rainfall in southern Saharan-Arabian Desert. **A:** U-Th ages (with 95% confidence [2 σ] error bars) of speleothem samples of Negev Desert. Speleothem samples are identified along vertical axis, and data for each speleothem are separated from one another by thin horizontal lines. Bold horizontal lines separate data for different caves (cave locations appear in Fig. 1B). Stratigraphic order of laminae in each individual speleothem is from bottom to top of row. Dark-gray rectangle shows range of ages of wet period indicated by speleothems. Bold gray bar below upper horizontal axis indicates archaeological period associated with early modern human remains in Skhul and Qafzeh Caves in northern Israel (Schwarcz et al., 1988; Valladas et al., 1988; Mercier et al., 1993; Bar-Yosef, 1998; Grun et al., 2005). **B:** Relative age-frequency curve calculated from age data, showing fraction of samples that formed in same age interval at 95% confidence level. **C:** Periods of enhanced monsoonal activity in southern and central parts of Saharan-Arabian Desert indicated by horizontal bars: Ref. 1—Osmond and Dabous (2004); Ref. 2—Szabo et al. (1995); Ref. 3—Fleitmann et al. (2003), and Ref. 4—Smith et al. (2004b). Figure shows that early modern human presence in Levant started at peak wet phase of the desert.

multicollector-inductively coupled plasma-mass spectrometry (MC-ICP-MS) at the Geological Survey of Israel are described in detail in Vaks et al. (2006). Thirty-three samples were dated (Fig. 2A based on data in Table DR1). Although some speleothems were 20–30 cm thick, only the outermost laminae of overall thickness of 5–30 mm were dateable (Fig. DR1, see footnote 1). Some laminae were less than 2 mm thick, which limited the sample size for dating to 10–20 mg. However, the increase in analytical uncertainty in these samples was minor due to the relatively

high U concentrations (see Table DR1). A relative age-frequency diagram was calculated using Isoplot/Ex 3 software (Fig. 2B) (Ludwig, 2003). The higher age-frequency peaks indicate more intensive speleothem deposition.

RESULTS AND DISCUSSION

The ages of speleothem deposition in the five caves mostly cluster between ca. 140 ka and 110 ka (interglacial MIS 5.5 and 5.4), with an additional minor deposition interval at ca. 88 ka (beginning of MIS 5.1) defined by one speleothem

age (Fig. 2A). The speleothem age-frequency distributions (Fig. 2B) show that the major depositional period was between 133 and 122 ka.

No speleothem deposition was found between 185 and ca. 140 ka (MIS 6) (Fig. DR1, between ca. 110 and ca. 90 ka (MIS 5.4–5.2), and after ca. 85 ka (i.e., during the most of the interglacial MIS 5.1, the last glacial period, and the Holocene). We therefore conclude that the central and southern Negev Desert was arid to hyperarid during these periods. The lack of deposition during the glacial periods contrasts with the continuous deposition of speleothems in the Mediterranean climate zone (in northern and central Israel; Bar-Matthews et al., 2003), and with the speleothem deposition during glacial periods in northern Negev (Vaks et al., 2006). Hol-Zakh, the northernmost cave studied in this work (Fig. 1B), and the northern Negev Tzavoa Cave studied by Vaks et al. (2006) are located only 10 km apart, yet the two caves are characterized by a sharp present-day precipitation gradient (150–160 mm/yr at Tzavoa to 90–100 mm/yr at Hol-Zakh), which is marked by thinning of the speleothem laminae at Hol-Zakh and a different chronology of speleothem deposition events.

Speleothem deposition in an otherwise arid to hyperarid region indicates an increase of rainfall above the caves, leading to recharge of water in the unsaturated zone. The increase in precipitation in the Negev Desert could have occurred because of a southern shift of the Mediterranean cyclones and/or an increase in their intensity. However, the periods of enhanced rainfall most probably would have been short in duration and alternated with long droughts. The data supporting this hypothesis of episodic rainfall are: (1) only thin, 5–10 mm speleothem laminae were deposited in the caves, and (2) the absence of calcite horizons in middle-late Quaternary gypsic-salic soils of the southern Negev indicates that precipitation levels greater than 80 mm/yr could not have persisted for periods longer than tens to hundreds of years (Amit et al., 2006). Thus, the wet episodes were too short to cause the development of calcite horizons in the soils but long enough to allow the formation of thin speleothem laminae in the caves.

Although the wet episodes were short, they would have had a significant environmental impact. The caves are located at shallow depths (2–20 m) below the tops of the hills, and thus their catchments are small compared to those of wadi channels, which would have seen significantly greater amounts of water and increased vegetation cover. Pollen records show a significant contemporaneous development of vegetation cover over the entire region (Weinstein-Evron, 1987), and increased spring discharge is marked by travertine deposition in the hyperarid southeastern parts of the Negev Desert (Livnat and Kronfeld, 1985). Higher rainfall is also indi-

cated by the Eemian record of perennial lakes in basins occupied today by playas in southern Jordan (Petit-Maire et al., 2002).

Figure 2 shows that the cluster of humid episodes in the Negev Desert between 140 and 110 ka was synchronous with increased monsoonal intensity in the southern Arabian Peninsula (Fleitmann et al., 2003) and the Saharan Desert (Szabo et al., 1995; Rohling et al., 2002; Osmond and Dabous, 2004; Smith et al., 2004b). This would have sustained habitable conditions for early modern humans over the entire desert belt, and not just in oases and the Nile Valley (Smith et al., 2004b). The northward movement of early modern humans through the Sinai-Negev land bridge would have been facilitated under these conditions, allowing them to reach the more hospitable climates of Levant ca. 130–100 ka, as documented by dated early modern human remains from the Skhul and Qafzeh Caves in northern Israel (Figs. 1B and 2) (Schwarcz et al., 1988; Valladas et al., 1988; Mercier et al., 1993; Bar-Yosef, 1998; Grun et al., 2005). Earlier prehistoric sites outside of Africa have not yielded unequivocal early modern human skeletal remains (Stringer, 2002; Barkai et al., 2003; Grun et al., 2005). The faunal assemblages of the archaeological layer in Qafzeh Cave (QF in Fig. 1B) show that Palearctic fauna, which dominated in the caves of southern Levant during MIS 6, disappeared at the onset of interglacial MIS 5, and African and Saharan-Arabian elements became the highest since 780 ka. These latter elements include species of micromammals, hartebeest, equids, and ostrich egg shells, all of which indicate an abundance of savannah species in the area (Rabinovich and Tchernov, 1995; Tchernov, 1992, 1996). The faunal assemblage of Qafzeh Cave thus provides supportive evidence for the northward expansion of African and Saharan-Arabian biotic zones during the period when early modern humans are first recorded in the area.

Middle Stone Age sites are found in the Negev and southern Jordan. Rink et al. (2003) suggested an age range of 90–220 ka or older for these Middle Stone Age sites. However, the absence of the skeletal remains makes it impossible to determine the hominid type.

The Nile River corridor provides a potential path to the Mediterranean Sea through the Saharan-Arabian Desert (route a in Fig. 1A). Wetter conditions in Sinai and Negev Deserts at this time would have made migration from the Nile Delta to the Levant easier. The increased monsoonal precipitation over the tropics and southern and central Saharan-Arabian Desert at 140–110 ka also would have enhanced the Nile River flow, made this route more usable, and opened new migration routes across the desert. Stringer (2000) also argued that prehistoric human expansions mainly occurred along the coasts,

which provided the first and fastest migration routes. The possible coastal migration routes from Ethiopia to the Levant follow: (1) the African Red Sea coast (route b in Fig. 1A), or (2) the Arabian Red Sea coast (route c in Fig. 1A), if early modern humans crossed the Bab-el-Mandeb Straits (Rose, 2004). The Red Sea coastal routes were extremely arid before 140 ka and after 115 ka (Almogi-Labin et al., 2004). Archaeological evidence supports the proposition of Red Sea coastal migration routes. Walter et al. (2000) found that early modern humans occupied the arid Red Sea coast of Eritrea at 125 ka; Van Peer et al. (1996) reported evidence of early modern human occupation of Red Sea coastal mountains of southeastern Egypt during approximately the same period. Petraglia and Alsharekh (2003) found that Middle Stone Age sites of early modern humans (although not radiometrically dated) are concentrated in the western parts of the Arabian Peninsula and some of them are close to the Red Sea coast.

Whereas the significant increase in wet episodes over northern Africa during the 140–110 ka period could have removed the climatic barriers to human and animal migration to the north, the aridization of the northern parts of Saharan-Arabian Desert at ca. 110 ka (expressed by the cessation of speleothem deposition in the Negev caves) would have suppressed the return route from the Levant to Africa for at least 20,000 yr. The highest frequency of speleothem ages (indicating highest speleothem deposition rate) occurs in the earlier half of the 140–110 ka period, followed by a gradual decrease in deposition as conditions moved to aridity (Fig. 2B). Thus, this period of expanding aridity in the Saharan-Arabian Desert may have encouraged early modern humans to move further into the Mediterranean climate zone of the Levant and possibly to the other parts of Eurasia.

Wet interglacial events have been suggested to be crucial to the occurrence of Pleistocene migrations “out of Africa” (Derricourt, 2005). We therefore suggest that the episodic humid phase between 140 and 110 ka in the northern Saharan-Arabian Desert opened an important interglacial climatic window that allowed early modern humans to disperse out of Africa to other parts of the world.

ACKNOWLEDGMENTS

This research was supported by the Israel Science Foundation (grant 910/05) and by the Ring Foundation. We would like to thank N. Tepliakov, I. Segal, O. Yoffe, S. Ashkenazi, B. Schilman, and E. Eliani from Geological Survey of Israel, and the team of the Cave Research Unit at Hebrew University, Jerusalem, for their assistance in field and laboratory work. We would also like to thank L. Grosman, A. Almogi-Labin, R. Amit, Y. Enzel, E. Hovers, N. Goren-Inbar, and R. Rabinovich for advice, and T. Esat, C. Marean, and the two anonymous reviewers for the fruitful comments, which led to considerable improvement of the manuscript.

REFERENCES CITED

- Almogi-Labin, A., Bar-Matthews, M., and Ayalon, A., 2004, Climate variability in the Levant and northeast Africa during the late Quaternary based on marine and land records, in Goren-Inbar, N., and Speth, D., eds., *Human Paleoeology in the Levantine Corridor*: Oxford, UK, Oxbow Books, p. 117–134.
- Amit, R., Enzel, Y., and Sharon, D., 2006, Permanent Quaternary hyperaridity in the Negev, Israel, resulting from regional tectonics blocking Mediterranean frontal systems: *Geology*, v. 34, p. 509–512, doi: 10.1130/G2254.1.
- Barkai, R., Gopher, A., Lauritzen, S.E., and Frumkin, A., 2003, Uranium series dating from Qesem Cave, Israel, and the end of the Lower Paleolithic: *Nature*, v. 423, p. 977–979, doi: 10.1038/nature01718.
- Bar-Matthews, M., Ayalon, A., Gilmour, M., Matthews, A., and Hawkesworth, C.J., 2003, Sea-land oxygen isotopic relationships from planktonic foraminifera and speleothems in the Eastern Mediterranean region and their implication for paleorainfall during interglacial intervals: *Geochimica et Cosmochimica Acta*, v. 67, p. 3181–3199, doi: 10.1016/S0016-7037(02)01031-1.
- Bar-Yosef, O., 1998, On the nature of transitions: The Middle to Upper Palaeolithic and the Neolithic revolution: *Cambridge Archaeological Journal*, v. 8, p. 141–163.
- Dayan, U., 1986, Climatology of back trajectories from Israel based on synoptic analysis: *Journal of Climate and Applied Meteorology*, v. 25, p. 591–595, doi: 10.1175/1520-0450(1986)025<0591:COBTFI>2.0.CO;2.
- Derricourt, R., 2005, Getting “out of Africa”: Sea crossings, land crossings: *Journal of World Prehistory*, v. 19, p. 119–132, doi: 10.1007/s10963-006-9002-z.
- Fleitsmann, D., Burns, S.J., Neff, U., Mangini, A., and Matter, A., 2003, Changing moisture sources over the last 330,000 years in northern Oman from fluid-inclusion evidence in speleothems: *Quaternary Research*, v. 60, p. 223–232, doi: 10.1016/S0033-5894(03)00086-3.
- Frumkin, A., Ford, D.C., and Schwarcz, H.P., 1999, Continental oxygen isotopic record of the last 170,000 years in Jerusalem: *Quaternary Research*, v. 51, p. 317–327, doi: 10.1029/1999GB001245.
- Grun, R., Stringer, C., McDermott, F., Nathan, R., Porat, N., Robertson, S., Taylor, L., Mortimer, G., Eggins, S., and McCulloch, M., 2005, U-series and ESR analyses of bones and teeth relating to the human burials from Skhul: *Journal of Human Evolution*, v. 49, p. 316–334, doi: 10.1016/j.jhevol.2005.04.006.
- Hendy, C.H., 1971, The isotopic geochemistry of speleothems: I. The calculation of the effects of different modes of formation on the isotopic composition of speleothems and their applicability as palaeoclimatic indicators: *Geochimica et Cosmochimica Acta*, v. 35, p. 801–824.
- Kahana, R., Ziv, B., Enzel, Y., and Dayan, U., 2002, Synoptic climatology of major floods in the Negev Desert, Israel: *International Journal of Climatology*, v. 22, p. 867–882, doi: 10.1002/joc.766.
- Livnat, A., and Kronfeld, J., 1985, Paleoclimatic implications of U-series dates for lake sediments and travertines in the Arava Rift Valley, Israel: *Quaternary Research*, v. 24, p. 164–172, doi: 10.1016/0033-5894(85)90003-1.
- Ludwig, K.R., 2003, *Isoplot 3.00: A Geochronological Toolkit for Microsoft Excel*: Berkeley, California, University of California, Special Publication No. 4, 70 p.
- Marean, C.W., and Assefa, Z., 2005, The Middle and Upper Pleistocene African Record for the Biological and Behavioral Origins of Modern Humans, in Stahl A., ed., *African Archaeology: A Critical Introduction*: Blackwell Studies in Global Archaeology: Malden, Massachusetts, Blackwell Publishing, p. 93–129.
- McBrearty, S., and Brooks, A.S., 2000, The revolution that wasn't: A new interpretation of the origin of modern human behavior: *Journal of Human Evolution*, v. 39, p. 453–563, doi: 10.1006/jhev.2000.0435.
- McDougall, I., Brown, F.H., and Fleagle, J.G., 2005, Stratigraphic placement and age of modern humans from Kibish, Ethiopia: *Nature*, v. 433, p. 733–736, doi: 10.1038/nature03258.
- Mercier, N., Valladas, H., Bar-Yosef, O., Vandermeersch, B., Stringer, C., and Joron, J.L., 1993, Thermoluminescence date for the Mousterian burial site of Es-Skhul, Mt. Carmel: *Journal of Archaeological Science*, v. 20, p. 169–174, doi: 10.1006/jasc.1993.1012.
- Osmond, J.K., and Dabous, A.A., 2004, Timing and intensity of groundwater movement during Egyptian Sahara pluvial periods by U-series analysis of secondary U in ores and carbonates: *Quaternary Research*, v. 61, p. 85–94, doi: 10.1016/j.yqres.2003.09.004.
- Petit-Maire, N., Sanlaville, P., Abed, A., Yasin, S., Bourrouilh, R., Carbonel, P., Fontugne, M., and Reyss, J.L., 2002, New data for an Eemian lacustrine phase in southern Jordan: *Episodes*, v. 25, p. 279–280.
- Petruglia, M.D., and Alsharekh, A., 2003, The Middle Palaeolithic of Arabia: Implications for modern human origins, behaviour and dispersals: *Antiquity*, v. 77, p. 671–684.
- Rabinovich, R., and Tchernov, E., 1995, Chronological, paleoecological and taphonomical aspects of the Middle Paleolithic site of Qafzeh, Israel, in Buitenhuis, H., and Uerpmann, H.P., eds., *Archaeozoology of the Near East*: Leiden, Netherlands, Backhuys Publishers, p. 4–44.
- Rink, W.J., Richter, D., Schwarcz, H.P., Marks, A.E., Monigal, K., and Kaufman, D., 2003, Age of the Middle Palaeolithic site of Rosh Mor, central Negev, Israel: Implications for the age range of the early Levantine Mousterian of the Levantine corridor: *Journal of Archaeological Science*, v. 30, p. 195–204, doi: 10.1006/jasc.2002.0831.
- Rohling, E.J., Cane, T.R., Cooke, S., Sprovieri, M., Bouloubassi, I., Emeis, K.C., Schiebel, R., Kroon, D., Jorissen, F.J., Lorre, A., and Kemp, A.E.S., 2002, African monsoon variability during the previous interglacial maximum: *Earth and Planetary Science Letters*, v. 202, p. 61–75, doi: 10.1016/S0012-821X(02)00775-6.
- Rose, J.I., 2004, The question of Upper Pleistocene connections between east Africa and south Arabia: *Current Anthropology*, v. 45, p. 551–555, doi: 10.1086/423500.
- Schwarcz, H.P., Grun, R.V.B., Bar-Yosef, O., Valladas, H., and Tchernov, E., 1988, ESR dates from the hominid burial site of Kafzeh in Israel: *Journal of Human Evolution*, v. 17, p. 733–737, doi: 10.1016/0047-2484(88)90063-2.
- Smith, J.R., Giegengack, R., and Schwarcz, H.P., 2004a, Constraints on Pleistocene pluvial climates through stable-isotope analysis of fossil-spring tufas and associated gastropods, Kharga Oasis, Egypt: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 206, p. 157–175, doi: 10.1016/j.palaeo.2004.01.021.
- Smith, J.R., Giegengack, R., Schwarcz, H.P., McDonald, M.M.A., Kleindienst, M.R., Hawkins, A.L., and Churcher, C.S., 2004b, A reconstruction of Quaternary pluvial environments and human occupations using stratigraphy and geochronology of fossil-spring tufas, Kharga Oasis, Egypt: *Geochronology: International Journal* (Toronto, Ontario), v. 19, p. 407–439, doi: 10.1002/gea.20004.
- Stringer, C., 2000, Coasting out of Africa: *Nature*, v. 405, p. 24–27, doi: 10.1038/35011166.
- Stringer, C., 2002, Modern human origins: Progress and prospects: *Philosophical Transactions of the Royal Society of London*, v. 357, p. 569–579, doi: 10.1098/rstb.2001.1057.
- Szabo, B.J., Haynes, J.C.V., and Maxwell, T.A., 1995, Ages of Quaternary pluvial episodes determined by uranium-series and radiocarbon dating of lacustrine deposits of eastern Sahara: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 113, p. 227–242, doi: 10.1016/0031-0182(95)00052-N.
- Tchernov, E., 1992, The Afro-Arabian component in the Levantine mammalian fauna—A short biogeographical review: *Israel Journal of Zoology*, v. 38, p. 155–192.
- Tchernov, E., 1996, Rodent faunas, chronostratigraphy and paleobiogeography of the southern Levant during the Quaternary: *Acta Zoologica Cracoviensis*, v. 39, p. 513–530.
- Vaks, A., Bar-Matthews, M., Ayalon, A., Matthews, A., Frumkin, A., Dayan, U., Halicz, L., Almogi-Labin, A., and Schilman, B., 2006, Paleoclimate and location of the border between Mediterranean climate region and the Saharo-Arabian Desert as revealed by speleothems from the northern Negev Desert, Israel: *Earth and Planetary Science Letters*, v. 249, p. 384–399, doi: 10.1016/j.epsl.2006.07.009.
- Valladas, H., Reyss, J.L., Joron, J.L., Valladas, G., Bar-Yosef, O., and Vandermeersch, B., 1988, Thermoluminescence dating of Mousterian ‘proto-Cro-Magnon’ remains from Israel and the origin of the modern man: *Nature*, v. 331, p. 614–616, doi: 10.1038/331614a0.
- Van Peer, P., Vermeersch, P.M., Moeyersons, J., and Van Neer, W., 1996, Paleolithic sequences of Sodmein Cave, Red Sea Mountains, Egypt, in Piviti, G., and Soper, R., eds., *Aspects of African Archaeology*: Harare, Zimbabwe, University of Zimbabwe, p. 149–156.
- Vermeersch, P.M., 2001, ‘Out of Africa’ from an Egyptian point of view: *Quaternary International*, v. 75, p. 103–112, doi: 10.1016/S1040-6182(00)00082-3.
- Walter, C., Buffler, R.T., Bruggemann, J.H., Guillaume, M.M.M., Berhe, S.M., Negassi, B., Libsekal, Y., Cheng, H., Edwards, R.L., von Cosel, R., Neraudeau, D., and Gagnon, M., 2000, Early human occupation of the Red Sea coast of Eritrea during the last interglacial: *Nature*, v. 405, p. 65–69, doi: 10.1038/35011048.
- Weinstein-Evron, M., 1987, Palynology of Pleistocene travertines from the Arava Valley, Israel: *Quaternary Research*, v. 27, p. 82–88, doi: 10.1016/0033-5894(87)90051-2.

Manuscript received 12 February 2007

Revised manuscript received 30 April 2007

Manuscript accepted 2 May 2007

Printed in USA

8. Paleoclimate of northern Saharan-Arabian Desert from speleothem record in central and southern Negev and Judea Desert, Israel

8.1 Introduction

This chapter reconstructs paleoclimate of the arid and hyper-arid central and southern Negev Desert and "rain-shadow" Judea desert, addressing issues that are not discussed in chapters 1-3. The subjects discussed are: speleothems as recorders of humid periods during the Pliocene and Quaternary, the minimum precipitation (rain and snow) amounts necessary for the speleothem deposition during glacial and interglacial periods, sources of rainfall, the vegetation cover, and correlation of the data to global climate changes. This chapter compares the paleoclimate records of the central and southern Negev with those of rain shadow Judea Desert and mildly arid zones of northern Negev and Jordan Valley, and attempts to synthesize these records to provide an overall picture of the paleoclimate evolution. Additional issues discussed are the role of humid periods older than 180 ka in the migration of hominids and animals out of Africa, speleothem deposition as a record of relative erosion rates and Sr isotopes in desert speleothems as tracers of dust origin and supply. The background and methods for this chapter are described in Introduction (1) and Methodology (4) chapters of this thesis.

8.2 Results

8.2.1 Field and petrographic characterization of the speleothems and the cave sediments

8.2.1.1 Vadose speleothems and preliminary U-Pb dating results

Most speleothems from the central and southern Negev Caves (#5-#14 in Fig.1C and Appendix 2) are vadose, consisting of stalagmites, stalactites and flowstones, composed

of low Mg calcite whose morphology shows the direction of the gravitational water flow. Their cross sections range from a few millimeters to ~40 cm thick. The speleothems have a similar stratigraphy in all caves located south of 150 mm present-day isohyet, and can be divided into three main stratigraphic members, based on their field appearance, color, laminae width, and crystal size: the Basal member, the Intermediate member, and the thin Young member (Figs. 3, 4).

The massive Basal Member is composed of large (up to few cm in length) columnar low Mg calcite crystals, mostly light transmitting (Figs. 3(1-2), 4A). The entire thickness of this member is 5-40 cm and it is the thickest of the three stratigraphic members. In Izzim, Ashalim, Mitzpe-Ramon, Wadi Lotz and Ktora Cracks Caves, the Basal member usually comprises more than 80-95% of the speleothem volume. The columnar, light transmitting, low-Mg calcite crystals usually show continuous growth across this stratigraphic member (Figs. 3(1), 4A), suggesting that Basal Member was deposited from continuously slowly dripping water. Rarely, mainly in stalactites, the growth of columnar calcite crystals is interrupted by the thin laminae of milky-white or beige, microcrystalline calcite, or by brown or beige laminae containing detrital material (Fig. 3(5)). The Basal member terminates at its top by gray, white, beige or red colored lamina, 1-8 mm thick, mainly containing microcrystalline calcite, evaporite minerals and detritus (Figs. 3 (1-4); 4A, B). This possibly suggests a hiatus (growth break) between the Basal member and the Intermediate member.

The Basal member is older than ~550 ka, which is the limit of the U-Th dating method. Preliminary U-Pb ages of two laminae, ASH-15-D and ASH-15-E, from the top of the Basal Member in a flowstone from Ashalim Cave (Fig. 3(1)) (performed in University of Melbourne, Australia) are 3.045 ± 0.021 Ma and 3.048 ± 0.014 Ma respectively (Jon Woodhead, personal communication, Table 1). Sample ASH-15-D+E was also dated by U-Pb method at the University of Leeds, UK, giving a preliminary age of 3.015 ± 0.205 Ma (Robert Cliff, personal communication). Laminae KTO(1)-1-J and

KTO(1)-1-K from the bottom of the Basal Member in a flowstone in Ktora Cracks (Fig. 4A), gave a preliminary U-Pb ages of 2.713 ± 0.060 Ma and 3.317 ± 0.063 Ma, respectively (Jon Woodhead, personal communication, Table 1). Thus, deposition of the speleothems of the Basal member occurred probably 3.3-2.7 Ma ago in the Middle Pliocene.

The Intermediate member is 2-15 cm thick, and comprises ~ 4-20% of the speleothem volume in most caves. The member is layered and composed of alternating black, brown, beige, gray or colorless laminae of columnar, light transmitting, low-Mg calcite crystals (Fig. 3, 4). The width of each lamina ranges from a few mm to 5 cm. Several fine grained milky-white or beige, laminae of micro-crystalline calcite cut through the columnar crystalline structure suggesting hiatuses (Fig. 3(1, 5), 4B). This member is also older than 550 ka but was deposited after the Basal member. The lamina ASH-15-C from the base of the Intermediate member of a flowstone ASH-15 from Ashalim Cave (Fig. 3(1)) gave a preliminary U-Pb age of 1.269 ± 0.013 Ma, which is Early Pleistocene (Jon Woodhead, personal communication, Table 1).

The Young member is 0.5-2 cm thick, composed of thinly laminated, columnar, low-Mg crystalline calcite, and comprises less than 1% of the speleothem volume. At some localities it was possible to accurately date the uppermost 2 to 6 calcite laminae by the U-Th method (chapter 8.2.2). The light transmitting crystalline calcite laminae are several mm thick and separated from one another by fine (<1 mm), opaque, milk-white, red, brown grey and beige fine laminae (Fig. 3(6), 4B). The latter are composed of microcrystalline calcite, gypsum, halite, clay minerals, quartz, and minor amounts of celestine, barite, apatite, and minerals of Fe, Mn, Ti, and Zn (Fig. 5; 6; 7). This composition is similar to present day “patina” accumulating on speleothem surfaces in arid and hyper-arid environments, suggesting that these laminae formed due to condensation corrosion and/or accumulation of evaporite minerals or detrital material where conditions were dry. Condensation corrosion occurs when tiny water droplets

undersaturated with respect to calcium carbonate condensate on dry speleothem surface, and corrode the crystalline calcite forming fine, milky-white, opaque, and sometimes porous lamina of micro-crystalline calcite on the speleothem surface (Auler and Smart, (2004); Dreybrodt et al., (2005) and references within). The detrital material accumulated after it was washed into the cave by rare rain events; drying out left a residue of clays and evaporite minerals on the speleothem surface.

In Kanaim Cave, Judea Desert, speleothems show laminated structure similar to speleothems in mildly-arid zones of Jordan Valley (Vaks et al., 2003) and northern Negev (Vaks et al., 2006), with the difference that they are composed of pure calcite with very small amounts of detrital material. Stalactite sections dateable by the U-Th method in this cave reach more than 10 cm in length. These stalactites are composed of beige or gray, 2-15 mm thick laminae of columnar, light transmitting, low-Mg calcite crystals, alternating with fine (<1 mm), opaque, milk-white laminae of micro-crystalline calcite with some halite and gypsum (Fig. 8). The latter represent hiatuses in calcite growth, during which condensation corrosion and deposition of evaporite minerals took place. The laminae older than the dating limit showed no change in petrography relative to younger ones. The three-member speleothem stratigraphy of the central and southern Negev was not observed in Kanaim Cave.

8.2.1.2 Phreatic speleothems

Phreatic speleothems are found in Ma'ale Dragot Cave in northern Negev, Makhtesh-ha-Qatan Cave, the lowermost passages of the Ashalim Cave, Ma'ale-ha-Meyshar Cave and some small caves nearby. These speleothems were deposited below the groundwater table, when the caves were submerged. In all caves, with the exception of Makhtesh-ha-Qatan, the phreatic speleothems are older than the 550 ka. Phreatic speleothems, as evident of any apparent gravitational control on their morphology, form 1-20 cm calcite overgrowth layers on the caves floor, walls and ceilings (Figs. 9A, B). In Ma'ale-ha-

Meyshar Cave and the nearby small caves, calcite "rafts" formed on ancient pool water surface and some of them sank later to the floor, forming a cover of up to 1 m thick on the caves floor (Fig. 9C). The formation of the cave "rafts" probably postdates the calcite overgrowth layer on the walls and ceiling of the caves. Cave "rafts" are the precise marker for the location of the ancient groundwater level (Polyak and Asmerom, 2005).

8.2.1.3 Cave sediments and paleosols washed into the caves

Three major types of soil are found above the caves: 1) In areas of the northern Negev and Jordan Valley where annual precipitation exceeds 250-300 mm, brown Terra-Rosa soil is found; this type of soil becomes more reddish when climate becomes more humid Mediterranean; 2) beige Loess soils are found in northern and central Negev and the Judea Desert, where present-day precipitation is between 250 and 80 mm; 3) beige and gray Hamada soils, typical of hyper-arid desert where present-day annual precipitation is 80 mm or lower, are found in the central and southern Negev, in Dead-Sea and Arava valleys (Dan et al., 1975; Rabikovitz, 1992).

The cave sediments found in most caves resemble the Loess or Hamada soils above the caves. In caves with the natural openings, some animal related organic matter and bat guano were found. In Even-Sid-Ramon and adjacent caves, and the Ma'ale-ha-Meyshar Caves, cave sediments differ from present-day soils due to their reddish color and high contents of iron oxide. They are possibly paleosols that were washed to the caves. In Ma'ale-ha-Meyshar Caves, these sediments clearly postdate the phreatic calcite overgrowths (Fig. 9A), and were probably washed into the caves at the same time as the formation of the cave "rafts" and vadose speleothems older than 550 ka. In the vadose speleothems the laminae of the columnar calcite are separated by fine detrital laminae that have a reddish color similar to the paleosol color (Fig. 6A).

8.2.2 Uranium concentrations of central and southern Negev speleothems

Uranium concentrations in central and southern Negev speleothems (Table 1) range between 0.04 ppm and ~20 ppm. In speleothems located within Turonian limestone host rock in the southern Negev (Shizafon mini-caves and Ktora Cracks), the range is 0.18 - 1 ppm, whereas in speleothems of the central Negev in Turonian host rock the range is 0.7 - 20 ppm. In speleothems located in Cenomanian dolomite and limestone host rocks uranium concentrations are slightly lower, 0.04 - 2 ppm. In a speleothem from Wadi-Lotz Cave located in Eocene limestone, uranium concentration is ~0.4 ppm. Corresponding uranium concentrations in the host rocks are as follows: 1 - 3.5 ppm in Turonian limestone from the southern Negev, 0.1 - 0.6 ppm in Turonian limestones from the central Negev, 0.35 ppm in Cenomanian limestones, 2 - 3.2 ppm in Cenomanian dolomites, and ~0.8 ppm in Eocene limestone. Uranium concentrations in the soils above the caves vary from 1.7 to 3.8 ppm. No direct link was found between the uranium concentrations in speleothems and the host rock and/or the soil, suggesting that uranium concentration in speleothems depends largely on migration routes of the water reaching a cave. Some speleothems of from three caves in Turonian host rock, Kanaim, Hol-Zakh and Ashalim, contain uranium concentrations ranging from ~4 ppm to ~20 ppm, which are much higher than the soil and the host rock. It is suggested that the source water migrated through Senonian phosphorites that contain up to 120 ppm uranium (Soudry et al., 2002), and are located in vicinity of Kanaim and Hol-Zakh Caves. Senonian phosphorites were not found above Ashalim Cave, but they are located above the Turonian host rocks 2.5 km to the north. It is thus possible that phosphorites were located in vicinity to Ashalim Cave in the past, but have been eroded.

8.2.3 U-Th ages and depositional periods

8.2.3.1 Kanaïm Cave, Judea Desert and the modified record of Ma'ale-Efrayim Cave, Jordan Valley

U-Th ages were determined on 14 speleothem samples from Kanaïm Cave (Table 1). The relative age frequency curve, calculated using Isoplot/Ex3 software (Ludwig, 2003) (Fig. 10), shows that speleothem deposition clusters at the following time intervals: 370-340 ka, 335-315 ka, 270-240 ka, 211-204 ka, 157-154 ka, 149-144 ka, 143-140 ka, 133-130 ka, ~38 ka, and 32-31 ka. These results indicate that speleothem deposition occurred only in short episodes during the 370 ka time span.

U-Th ages were determined on 15 speleothem samples from Ma'ale-Efrayim Cave using MC-ICP-MS (Table 1) in order to add more ages to the existing record from TIMS dating (Vaks et al., 2003). The relative age frequency curve, which includes both new and previous age results, calculated using Isoplot/Ex3 software (Fig. 10), shows that speleothem deposition in Jordan Valley clusters between 320-255 ka, 230-205 ka, 185-170 ka, 162-145 ka, ~135 ka, 75-24 ka, 17-16 ka, and 13.5-11 ka. This record expands the earlier work of Vaks et al. (2003) to 320 ka, eliminates an erroneous 78 ka deposition episode, but includes a short depositional episode at 12.1 ± 1.3 ka (sample ME-12-out at the top of the stalagmite ME-12).

8.2.3.2 Caves from the central and southern Negev Desert

The dating study was performed on the outermost laminae of the Young member, 0.2-2 cm thick, which were found to be dateable by the U-Th method. One hundred twenty three U-Th age determinations using MC-ICP-MS were carried out (Table 1, Fig. 11A), and the relative age frequency curve was calculated using Isoplot/Ex3 software (Figs. 11B, 12A, B, C). The results show that speleothem deposition was highly discontinuous, with clusters at the following time intervals: 520-440 ka, 380-285 ka, 255-240 ka, 225-185 ka, 160-155 ka, 142-109 ka, and 89-86 ka.

The age uncertainties for the time interval between 550 ka and 350 ka are large and overlap one another. Thus, it is difficult to precisely determine the periods of speleothem deposition. The relative age frequencies were higher at 520-440 ka and 380-350 ka, and very low between 440 ka and 380 ka (Fig. 12C). Speleothem deposition during the last 350 ka is classified here into two groups: 1) periods with relatively *intensive* deposition, defined by at least two ages from two different caves; 2) episodes with *limited* deposition defined by single ages from single caves. Periods of intensive speleothem deposition occurred during the three major time intervals: 350-290 ka, 220-190 ka, and 142-109 ka. Limited speleothem deposition occurred at 290-285 ka, 255-240 ka, 225-220 ka, 190-185 ka, 160-155 ka and 89-86 ka. No speleothem deposition is observed at 285-255 ka, 240-225 ka, 185-160 ka, 155-142 ka, 109-89 ka, 86-0 ka.

There are more speleothem deposition periods in the central Negev compared to the southern Negev (to the south of 50 mm isohyet). Intensive speleothem deposition in southern Negev only occurred during the period between 350 ka and 310 ka. Limited speleothem deposition occurred at ~137 ka, whereas other periods of speleothem deposition are missing, including the periods at 310-290 ka and 220-190 ka during which many speleothems formed in central Negev (Fig. 11). Even-Sid-Ramon mini-caves are the southern most caves where speleothem deposition aged 310-290 ka period is found. Ma'ale-ha-Meyshar Cave is the southern most cave in which speleothems aged 220-190 ka are found.

During the three periods of intensive speleothem formation, peaks occur at 343 ka, 320 ka, 295 (Fig. 12C); 218 ka, 212 ka, 205 ka, 200 ka, 193 ka (Fig. 12B) and 138 ka, 133-131 ka, 126 ka, 117 ka, and 112-111 ka (Fig. 12A). During the periods between 220-190 ka and 142-109 ka, the average time span between adjacent peaks is 6.4 ± 1.2 ka.

Most ages follow the stratigraphic order with the outermost laminae being the youngest. However, in some cases laminae adjacent to hiatuses or next to the present day

surface show an age reversals, e.g. stalactite KN-8 from Kanaim Cave and flowstone ASH-33 from the Ashalim Cave (Table 1; Fig. 8, Fig. 13).

In KN-8, the measured age of outer lamina A is older (182.6 ± 2.3 ka) than that of lamina B (146.5 ± 2.8 ka). This may be due to condensation corrosion processes that affected lamina A and resulted in radionuclide remobilization (Auler and Smart, 2004). Lamina B most probably represents the true age because of its clear columnar calcite crystals which are typical for a closed system (Fig. 8A, B).

In ASH-33, the ages of upper 13 sub-laminae (A1 to E2.1) follow the stratigraphic order from 218.0 ± 2.0 ka to 116.7 ± 3.1 ka, but the 14-th sub-lamina E2.2 above the hiatus (between laminae E and F (>500 ka)), has an age of 199.5 ± 2.5 (Table 1; Fig. 11A). The hiatus between laminae E and F represents a >300 ka interval with no speleothem deposition, during which detrital material (clay minerals, and host rock fragments) accumulated on speleothem surface (Fig. 13 (1), (2), (3), (4)). On renewal of speleothem deposition at about 220 ka, calcite laminae E2.1-D1 overgrew the detrital fragments, but porosity remained in between the new laminae and the old surface, causing open system conditions, that lead to deposition of the younger secondary calcite, and the age reversal in lamina E2.2 (Figs. 13(4), (5)). The fabric of lamina E2.2 differs from the columnar crystal fabric of laminae A1 to E2.1, consistent with this hypothesis (Fig. 13(3)).

8.2.4 Oxygen and carbon isotopic compositions of the central and southern Negev speleothems

$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (‰ PDB) profiles were measured on 12 speleothems from six caves. Eleven speleothems have sections of 350 ka and younger, and four speleothems include Basal and Intermediate members, several laminae of the Young member older than 550 ka and one lamina dated by U-Th to 516 ± 51 ka.

8.2.4.1 Oxygen isotopic compositions

$\delta^{18}\text{O}$ values of speleothems older than 550 ka were measured in Ashalim Cave across the entire cross section of flowstone ASH-15, which comprises the Basal and Intermediate members, the bottom laminae of the Young member from stalagmite ASH-34 (Ashalim Cave), stalactite ESID-2 (Even-Sid-Ramon mini-caves) and flowstone MMR-7(2) (Ma'ale-ha-Meyshar Cave). $\delta^{18}\text{O}$ values of the Basal member vary slightly from -9.4‰ and -9.7‰ on the bottom ~ 3 cm, than they then decrease to -10.1‰ - -11.8‰ across the remaining major part of the Basal member. $\delta^{18}\text{O}$ values of the Intermediate member range between -9.3‰ to -10‰. At the bottom of the Young member, the $\delta^{18}\text{O}$ values vary between -7.1 ‰ to -10.4 ‰. In Even-Sid-Ramon mini-caves (ESID-2, lamina B) an episode of speleothem deposition was dated to 516 ± 51 ka gave $\delta^{18}\text{O}$ values decreasing from -0.3‰ to -3.5‰ (Fig. 14)

$\delta^{18}\text{O}$ profiles for the period from 350 ka to 315 ka are represented by flowstone SHACH-1 from the Shizafon mini-caves; and for the 305-285 ka period – by phreatic speleothem MKTC-5 from Makhtesh-ha-Qatan Cave and by stalactite ESID-2 from Even-Sid-Ramon mini-caves. During the former interval, the $\delta^{18}\text{O}$ values range between -11.3‰ and -10.2‰ and are the lowest values found in the last 350 ka. During the latter period $\delta^{18}\text{O}$ values decrease from -1.6‰ to -3.7‰ in the Makhtesh-ha-Qatan Cave and then vary between -4‰ and -2.8‰ in Even-Sid-Ramon mini-caves (Fig. 15A, D, E).

One of the most important periods of speleothem deposition in the central Negev Desert occurred between ~220 ka and 190 ka. $\delta^{18}\text{O}$ profiles were measured on five speleothems from four caves: stalactite HZ-3(2) from Hol-Zakh Cave; flowstone ASH-33 and stalagmite ASH-34 from Ashalim Cave; stalactite ESID-2 from Even-Sid-Ramon mini-caves; and flowstone MMR-8 from Ma'ale-ha-Meyshar Cave, covering the periods between 221 ka and 219 ka, and between 213 ka and 197 ka. $\delta^{18}\text{O}$ values range between -6.9‰ and -10.5‰. During the 221-219 ka period $\delta^{18}\text{O}$ values measured on one flowstone from the Ashalim Cave show an increase from -7.7‰ to -7.2‰ and a decrease

from -6.9‰ to -7.7‰ between 213 ka and 208 ka. In contrast, five speleothems formed between 208 ka and 197 ka give $\delta^{18}\text{O}$ values between -7.2‰ and -10.5‰. Although these differences in $\delta^{18}\text{O}$ values exist between the caves, the overall $\delta^{18}\text{O}$ trends are similar and the calculated running average curves indicate three peaks of negative values -9.6‰, -9.65‰ and -10.1‰ at 206.5 ka, 204 ka and 199 ka, respectively. Two peaks of high values of -7.0‰ and -7.2‰ occur at 214 ka and 203 ka respectively (Figs. 15A, 16 C, D). $\delta^{18}\text{O}$ profile of speleothem deposited during the short episode at ~157 ka is represented by lamina C of a phreatic speleothem MKTC-5 from Makhtesh-ha-Qatan Cave, having $\delta^{18}\text{O}$ values varying between -3‰ and -4.7‰ (Fig. 15C).

The major speleothem deposition period in the central Negev Desert during the last 550 ka occurred between 142 ka and 109 ka, and is represented by fourteen speleothems from six caves. $\delta^{18}\text{O}$ profiles were measured on seven speleothems from four caves covering time span from 133 ka to 115 ka. The studied speleothems are flowstone HZ-1 and stalactite HZ-3(2) from Hol-Zakh Cave; stalactite ASH-11, flowstone ASH-33 and stalagmite ASH-34 from Ashalim Cave; stalactite ESID-7 from Even-Sid-Ramon mini-caves and flowstone MMR-7(2) from Ma'ale-ha-Meyshar Cave.

Based on $\delta^{18}\text{O}$ values, it is possible to divide the 133-115 ka period into two parts. The first sub-period occurred between 133 ka and 127 ka, represented by seven speleothems from the four caves. During this period the $\delta^{18}\text{O}$ values range between -7.9‰ and -10.2‰. Although the $\delta^{18}\text{O}$ values differ between the caves, the trends are similar. The running average of $\delta^{18}\text{O}$ shows four peaks of negative values of -9.5‰, -9.4‰, -9.8‰ and -9.6‰ at 131.5 ka, 128.5 ka, 127 ka, and 126.5 ka respectively. Three peaks of positive values of -8.6‰, -8.2‰ and -8.3‰ occur at 132 ka, 130 ka and 127.5 ka respectively. The second sub-period between 127 ka and 117 ka is represented by four speleothems from two caves. Three speleothems from the Ashalim Cave have $\delta^{18}\text{O}$ values between -7.4‰ to -8.5‰ between 127 ka and 118 ka with a decrease to -9.8‰ at 117 ka. The speleothem from Ma'ale-ha-Meyshar Cave which grew between 127 ka and

119 ka, have $\delta^{18}\text{O}$ values ranging from -10.4‰ to -10.7‰. Due to the large difference in $\delta^{18}\text{O}$ values and trends a running average was not calculated for this period (Fig. 15 A, 16A, B). The youngest period of speleothem deposition at 88-87 ka is characterized by $\delta^{18}\text{O}$ values vary between -7.7‰ and -8.1‰ (Fig. 15A, B).

8.2.4.2 Carbon isotopic compositions

Speleothem $\delta^{13}\text{C}$ values vary in the Basal member vary between -2‰ and -3.5‰; in the Intermediate member the values decrease continuously from +2.1‰ at the base to -6.1‰ at the top. In the bottom laminae of the Young member $\delta^{13}\text{C}$ values largely vary from +1.35‰ to -8.5‰. $\delta^{13}\text{C}$ values decrease from -2.0‰ to -3.1‰ during speleothem deposition episode at 516 ± 51 ka (Fig. 17).

Oscillations between high and more negative $\delta^{13}\text{C}$ values are characteristic of all the ensuing periods. Between 350 ka and 315 ka, $\delta^{13}\text{C}$ values range between +1‰ and -1.8‰, compared to the 305-285 ka period where they decrease from -3.2‰ to -7.5‰ (Fig. 18A, D, E). Between 221 ka and 219 ka, $\delta^{13}\text{C}$ values range between +1.5‰ and +0.7‰, which are the highest values during the last 350 ka. Between 213 ka and 208 ka, $\delta^{13}\text{C}$ values vary from -0.2‰ to -0.6‰, and from -0.5‰ to -6.8‰ between 208 ka and 197 ka, with the highest values being found for the Ashalim Cave and the lowest for Even-Sid-Ramon mini-caves. The running average $\delta^{13}\text{C}$ values emphasize two major peaks of negative values of -5.0‰ and -4.3‰, at 205 ka 198 ka, respectively. Three major peaks of positive values of +1‰, -0.3‰, and -0.9‰ are identified at 220 ka, 208 ka and 202 ka respectively (Fig. 18A, 19C, D). $\delta^{13}\text{C}$ values typical to the ~157 ka deposition episode in Makhtesh-ha-Qatan Cave range from -7‰ to -8.5‰, increasing to -6‰ in the end (Fig. 18A, C).

During the 133 ka 115 ka period the $\delta^{13}\text{C}$ values range between -0.5‰ and -7.7‰ (Fig. 18A; 19A, B). Between 133 ka and 127 ka, the $\delta^{13}\text{C}$ values are between -0.5‰ and -7.6‰. Although the $\delta^{13}\text{C}$ values differ between the caves, generally decreasing trends

mark this period with the most significant decreases being found in the northernmost caves. The running average shows four negative peaks of -3.5‰, -5.0‰, -5.6‰ and -5.8‰, at 131 ka, 130 ka, 128.5 ka and 126.5 ka, respectively, and three positive peaks of -1.6‰, -2.9‰, and -3.2‰ occur at 132 ka, 130.5 ka, 128 ka respectively (Fig. 19A, B). Following this, $\delta^{13}\text{C}$ values sharply increase and during the 126-118 ka period the $\delta^{13}\text{C}$ values range between -1.4‰ and -2.8‰ (running average ranges between -1.8‰ and -2.3‰). After 118 ka the $\delta^{13}\text{C}$ values decrease sharply reaching -7.7‰ at 116 ka. Speleothems of Ashalim and Ma'ale-ha-Meyshar caves have similar trends during this period (Fig. 19A, B). During the youngest speleothem deposition event at 88-87 ka (Even-Sid-Ramon Cave), the $\delta^{13}\text{C}$ values vary from -5.6‰ to -4.1‰ (Fig. 18A, B).

8.2.5 Present-day rainfall in the Negev Desert: amounts and hydrogen and oxygen isotopic compositions

Rainfall data: amounts, δD and $\delta^{18}\text{O}$ values for the northern, central and southern Negev Desert measured during the winters of 2004-2005 (including data of Vaks et al., (2006)) and 2005-2006 are given in Table 2. Rainfall amounts vary as follows: in Be'er-Sheva from 207 mm to 389 mm; in Arad from 53 mm to 161 mm; in Makhtesh-ha-Qatan from 25 mm to 122 mm; in Mitzpe-Ramon from 48 mm to 50 mm and in Neot-Smadar from 9 mm to 19 mm.

δD and $\delta^{18}\text{O}$ values of rainfall in the northern Negev (Be'er-Sheva, Arad) are between +37.4‰ and -68.3‰ and between +6.4‰ and -11.8‰, respectively with d-excess values between +35‰ and -13‰ (Table 2, Fig. 20). Average $\delta^{18}\text{O}$ and δD compositions during the years 2004-2006 are respectively -4.7‰ and -17.6‰ (d-excess = +20.2‰) in Be'er-Sheva, and -4.1‰ and -13.8‰ (d-excess = +19.1‰) in Arad (Table 2; Fig. 20A, B; Fig. 21C).

δD and $\delta^{18}\text{O}$ values of rainfall in central and southern Negev are between +39.0‰ and -44.0‰ and between +5.1‰ and -8.0‰, respectively with d-excess values ranging

between +41‰ and -12‰ (Table 2; Fig. 21). All events with d-excess values below +1‰ are smaller than 2 mm. Average rainfall $\delta^{18}\text{O}$ and δD values in Makhtesh-ha-Qatan for 2002-2006 are -2.4‰ and -6.1‰ respectively (d-excess = +13.4‰) (including data of Kuperman (2005) for 2002-2004); -3.3‰ and -11.1‰ (d-excess = +15.1‰) in Mitzpe-Ramon, and -2.0‰ and -7.5‰ (d-excess = +8.8‰) in Neot-Smadar (Table 2; Fig. 21A, B, C). The average $\delta^{18}\text{O}$ and δD values of the Ein-Netafim spring water are -3.9‰ and -18.4‰ respectively, with d-excess of +13.2‰ (Table 2).

8.2.6 δD values of speleothem fluid inclusions from northern, central and southern Negev

Corrected δD values of speleothem fluid inclusions (FI) from Tzavoa, Hol-Zakh, Ashalim and Ma'ale-ha-Meyshar Caves, range between -23‰ and -58‰ (Table 3, and Fig. 22A, B). These values are in the range of the present day rainfall, but significantly lower than the average (Table 2, Fig. 20, 21).

δD values of the FI water of Ashalim Cave speleothems older than the 550 ka, are between -26‰ and -53‰. The speleothems of the Basal member in Ashalim Cave have very low FI water content (<0.1 μl for 200 mg calcite) and only one sample was successfully analyzed, giving a δD of -47.3‰. Speleothems of the Intermediate member have higher water contents and 7 samples from this member gave δD values between -26‰ and -53‰ (Fig. 22B).

δD values of speleothem FI deposited during the last 220 ka range between -23.0‰ and -58‰. Between 220 ka and 190 ka, they vary from -43.8‰ to -56.9‰, with no significant differences between the caves. Between 177 ka and 172 ka, δD values range between -43.6‰ and -48.7‰, and between 161 ka and 157 ka they increase to -34.0‰. Values for the 132-117 ka period vary from -55.9‰ to -31.7‰, with the lowest values in Ma'ale-ha-Meyshar Cave and the highest values in Tzavoa Cave. Between 67 ka and 65

ka, FI δD values rise to -23.0‰ and are the highest in the Negev speleothems and high values of -31.1‰ of was also found between 35 ka and 32 ka.

8.2.7 Sr concentrations and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in speleothems, cave host rocks and soils for the Jordan Valley and the Negev Desert

Sr concentrations in vadose speleothems vary largely between 41 ppm and 1270 ppm (Table 4; Fig. 23A-D).

Speleothem Sr concentrations in central and southern Negev range between 100 ppm and 326 ppm in the Basal member, and between 41 ppm and 110 ppm in the Intermediate member. Sr concentrations in speleothems of the Young member vary from 43 ppm to 557 ppm (Fig. 23A). High Sr concentrations (252-557 ppm) are found in Ashalim and Hol-Zakh Caves, whereas low Sr concentrations (43-203 ppm) occur in Even-Sid-Ramon and Ma'ale-ha-Meyshar caves, Ktora Cracks and Shizafon mini-caves. In younger speleothems the following ranges are observed: at 350-290 ka between 43 and 200 ppm; at 225-187 ka between 111 and 557 ppm; at 135-110 ka from 58 to 510 ppm, and lamina deposited at 89-86 ka episode, 203 ppm (Table 4; Fig. 23D).

Sr concentrations in northern Negev range from 57 ppm to 527 ppm with 1270 ppm being observed in one speleothem lamina. A detailed record of Sr concentrations during the last 200 ka derived from Tzavoa Cave speleothems shows significant changes. Between 199 ka and 197 ka Sr concentrations are low (66-78 ppm), increase to 166-173 ppm between 176 ka and 173 ka, and 354 ppm at ~157 ka. Between 140 ka and 123 ka, Sr concentrations decrease from 151 ppm to 57 ppm. Between 79 ka and 14 ka the Sr concentrations are high 257-527 ppm, with the highest values at 78 ka and 70 ka (335-402 ppm), 38 ka (433 ppm), 21 ka (527 ppm), 18-17 ka (450-462 ppm) and 14 ka (473 ppm). Sr concentrations of Ma'ale-Dragot Cave speleothems vary between 152 ppm (lamina dated to ~3.8 ka) to 1270 ppm (lamina dated to ~65 ka) (Table 4; Fig. 23C).

Sr concentrations in Jordan Valley speleothems (Ma'ale-Efrayim) range between 100 ppm and 350 ppm, the lower values (115-154 ppm) are typical to the period from 310 ka to 205 ka, and higher values (generally >150 ppm) from 160 ka to the 16 ka (Table 4; Fig. 23B).

Sr concentrations in the phreatic speleothems range between 160 ppm and 2140 ppm. The value of 2140 ppm for speleothem in Makhtesh-ha-Qatan (MKTC-5) is the highest in the Negev speleothems.

Sr concentrations in host rocks vary between 78 ppm and 950 ppm. In Terra-Rosa soils above Ma'ale-Dragot and in Ma'ale-Efrayim, Sr concentrations range between 121 ppm and 186 ppm, whereas in the Loess and Hamada soils Sr concentrations are higher, between 314 ppm and 408 ppm (Table 4).

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios in Ma'ale-Efrayim Cave speleothems vary from 0.7082 to 0.7085. Between 310 ka and 205 ka $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vary from 0.7082 to 0.7084. This is followed by generally steady line increase from 0.7083 to 0.7085 from 200 ka to 72 ka. Between 72 ka and 16 ka the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios oscillate between 0.7083 and 0.7085, with highest values of 0.7085 at 72 ka, 63 ka, 38 ka, and 25 ka respectively (Fig. 24A).

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the vadose speleothems from Negev caves vary from 0.7077 to 0.7085. In Basal member $^{87}\text{Sr}/^{86}\text{Sr}$ ratios increase from 0.7077 in the base to 0.7079 at the top, in Intermediate member the values vary between 0.7081 and 0.7085, and in the Young member – between 0.7079 and 0.7085. Low values of 0.7079-0.7080 at 350-310 ka rise to 0.7082-0.7083 between 310 ka and 170 ka, increase to 0.7085 at 160-155 ka and decrease to 0.7082 at 132 ka. Between 132 ka and 110 ka $^{87}\text{Sr}/^{86}\text{Sr}$ ratios oscillate from 0.7082 to 0.7084 and between 110 ka and 13 ka show graduate but irregular increase from 0.7082 to 0.7085, with high peaks of 0.7085 at 63 ka, and 16 ka. In ~3.8 ka stalactite lamina MD(1)-8-A1, Ma'ale-Dragot caves, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio drops to 0.7081 (Table 4, Fig. 25A, B).

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios in phreatic speleothems from Ma'ale-ha-Meyshar Cave range between 0.7077 and 0.7078, and in Makhtesh-ha-Qatan Cave (MKTC-5-C) the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is 0.7079 (Table 4).

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the cave host rocks range between 0.7074 and 0.7078 (Table 4 and Figs. 24B, 26) and typical of Cretaceous and Eocene carbonate rocks (Koepnick et al., 1985).

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the bulk soils above the caves vary from 0.7084 to 0.7110. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the bulk Loess and Hamada desert soils range between 0.7084 and 0.7091. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in bulk Terra-Rosa soils are higher, 0.7098 and 0.7110, with 0.7098-0.7100 above Ma'ale-Dragot Caves in northern Negev, and 0.7110 above the Ma'ale-Efrayim Cave in Jordan Valley.

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios in silicate fraction of soils range between 0.7101 and 0.7123. In silicate fraction of the Negev soils: Loess, Hamada soils and Terra-Rosa soil above Ma'ale-Dragot Cave $^{87}\text{Sr}/^{86}\text{Sr}$ ratios range between 0.7101 and 0.7115, whereas higher value of 0.7123 is found in silicate fraction of the Terra-Rosa soil in Jordan Valley (Ma'ale-Efrayim Cave) (Table 4; Figs. 24B, 26). The silicate fraction of paleosols washed to the caves ranges between 0.7098 and 0.7140 (Table 4).

8.3 Discussion

8.3.1 Humid episodes in central /southern Negev, and Judea deserts

8.3.1.1 Speleothem deposition as recorder of effective precipitation

Periods of speleothem deposition in arid environments are indicative of water availability in unsaturated zone (Holmgren et al., 1995; Fleitmann et al., 2003; Vaks et al., 2003, 2006 and 2007). In this study the term "humid/wet" period is used to define time intervals with a positive effective precipitation, i.e. in which sufficient water enters the unsaturated zone for speleothem deposition to occur. Correspondingly, the term "arid/dry" period refers to a negative effective precipitation where insufficient water

enters the unsaturated zone, for of speleothem growth to occur. During the "humid/wet" period the infiltration coefficients (the precipitation fraction entering the groundwater) are greater than zero, whereas during the "arid/dry" period they are close to zero.

Humid and arid episodes have been correlated with the glacial/interglacial global climate change framework of Marine Isotopic Stages (MIS) as defined by Shackleton and Opdyke (1973), Martinson et al. (1987), and Fontugne and Calvert (1992).

At present, in the Mediterranean climate zone in northern and central Israel $\sim 1/3$ of the annual rainfall reaches the unsaturated zone and the remaining $2/3$, i.e., about 300-350 mm in Soreq Cave area is lost either by evaporation and/or by runoff (Ayalon et al., 1998, 2004). Despite this loss, the amount of water reaching the unsaturated zone is sufficient to allow speleothem deposition today and throughout the Holocene (Bar-Matthews et al., 2003-a). The southernmost Holocene (3.8 ka) speleothem deposition occurred in Ma'ale-Dragot Caves located at the present-day 280-300 mm isohyet (Fig. 1C) (Vaks et al., 2006). No present-day speleothem deposition occurs in semi-desert and desert regions of Israel that receive less than ~ 300 mm of annual rainfall. Thus, 300-350 mm isohyets can be defined as present-day limit of speleothem deposition in Israel, i.e. the zone with positive effective precipitation is located to the north of this boundary, and negative effective precipitation (infiltration coefficient of zero) to the south.

During past interglacial periods, regional temperatures were similar to those of the Holocene, as recorded by alkenone and *G. ruber* paleo-temperatures (Emeis et al., 1998, 2003; Kallel et al., 2000) and speleothem fluid inclusions (McGarry et al., 2004). Based on this evidence is possible to conclude that speleothem deposition during interglacial periods required at least 300-350 mm of annual rainfall (Vaks et al., 2006).

During the glacial periods, regional temperatures were 5-10°C lower than at present (Emeis et al., 2003; McGarry et al., 2004). Thus, a main question is the minimum precipitation amount that was necessary for speleothem deposition during the glacial periods compared to the interglacial periods. At present, most infiltration to the

unsaturated zone aquifers occurs during mid-winter heavy rainstorms when air temperature is below 10°C (Ayalon et al., 1998; 2004). A study made by Bonacci (2001) in Dinaric karst area, Istria Peninsula, Croatia, showed that during the winter months when average temperature is ~3.5°C, infiltration coefficients are 0.9-0.95, i.e. almost all the rain and snow infiltrate to the aquifer. During the April and October when the temperatures are 10°C and 12°C respectively (similar to present day average winter temperatures in Israel), the infiltration coefficients are ~0.6, and during the July and August when average temperature is ~21°C, the infiltration coefficients drop to ~0.2. In the above region the annual precipitation is ~1000 mm and falls throughout the year. During years when most of the rainfall occurred during the hot season, the mean annual infiltration coefficient was 0.36, whereas in years where most of the precipitation fell in the cold season the mean coefficient was 0.76. Similarly, the study of Shuman and Donnelly (2006) in Massachusetts, US, showed that during years when most rainfall and snow falls in winter, the lake levels increase significantly relative to years when most of rainfall falls in summer. In the Oslo region, Norway most precipitation during winter months occurs as snowfall (~200mm) and rainfall of ~600 mm occurs mainly during the warmer season (Norwegian Meteorological institute data <http://www.worldweather.org/008/c00021.htm>). The infiltration is ~1/2 (i.e., 400 mm) of the total precipitation, and French and Binley (2004) (and references within) argue that 50% of the total recharge is from snowmelt, implying that additional 200 mm of recharge is from rainfall, thus emphasizing the importance of infiltration from snowmelt.

The data of Bonacci (2001) suggest that the infiltration coefficients can increase by ~50-60% when the temperatures drop from 10-12°C to 3-4°C. A similar ~50-60% increase in infiltration coefficients in the Mediterranean climate of Israel would reduce the amounts of water lost due to evaporation/runoff by 75-100 mm and imply that 200-275 mm compared to 300-350 mm in interglacial periods, is the minimum amount of annual rainfall and snow required for speleothems deposition in glacial periods.

In desert locations vast bedrock surfaces can be found, and during the rain events bedrock surfaces collect runoff from larger areas. The runoff infiltrates to small soil pockets or cracks in the rock, leading to existence of limited local areas receiving an equivalent of 4-5 times higher rainfall amount (Yair, 1990). In caves like Ktora Cracks, Ma'ale-ha-Meyshar, Even-Sid-Ramon or Makhtesh-ha-Qatan Cave this effect can cause increased water infiltration. However, all these caves are dry today and their major speleothem deposition periods (and major hiatuses) do not significantly differ from those of Ashalim Cave, above which bedrock surfaces contributing to runoff are not found. Thus, although it is possible that runoff infiltration from the bedrock surfaces is able to cause formation of speleothems even for minor increases in precipitation, it is likely that such events will be recorded in single caves, rather than in multiple caves.

8.3.1.2 Humid periods in central and southern Negev Desert

Based on the above rough estimates of the minimum precipitation necessary for speleothem deposition during glacial and interglacials, it becomes possible to semi quantitatively assess variations in precipitation amounts in central and southern Negev Desert and the "rain shadow" and Judea Desert during the glacial-interglacial cycles that occurred in the last 350 ka.

As noted previously, it is possible to describe speleothem deposition events in the central and southern Negev Desert as either periods of *intensive* speleothem deposition, characterized by two or more speleothem ages from two or more caves in the region; or periods of *limited* speleothem deposition, characterized by a single speleothem age. Periods of intensive speleothem deposition are assumed to have occurred when regional precipitation amounts were higher than the minimum precipitation value, leading to positive effective precipitation in the entire region and the formation of numerous speleothems. Limited speleothem deposition occurred when the precipitation amounts were lower in the wider region, but high enough to enable local speleothem deposition.

Such a scenario could occur when runoff from vast bedrock surface contributes water to the small area (Yair, 1990), as described in section 8.3.1.1.

Based on the numbers of speleothem ages for various caves, the timing of deposition, and the correlation with global climate changes, five groups of rainfall/speleothem formation regimes during the last 350 ka in the central and southern Negev Desert can be defined:

1) Intensive speleothem deposition periods during previous interglacial periods, defining the most humid events with annual precipitation exceeding 300 mm. These events occurred between 350 ka and 310 ka (interglacial MIS-9), between 220 and 190 ka (interglacial MIS-7.3, 7.2 and 7.1), and between 142 and 109 ka (including glacial MIS-6.1 – interglacial MIS-5.5 transition (Termination II) (Usami et al., 1998), interglacial MIS-5.5, 5.4 and the beginning of MIS-5.3) (Figs. 11A, B; 12A, B, C).

2) Intensive speleothem deposition during previous glacial periods, most probably reflecting at least 200-275 mm of annual precipitation. This is represented by speleothem growth periods between 310 ka and 290 ka (glacial MIS-8.5) (Figs. 11A, B; 12A, B, C).

3) Episodes of limited speleothem deposition during interglacials, possibly representing events with precipitation higher compared to the present, but not as high as in the case 1. These episodes occurred at 247.4 ± 10.8 ka (probably at the transition from glacial MIS-8.1 to the interglacial MIS-7.5), 225-220 ka (beginning of the interglacial MIS-7.3), 190-185 ka (end of the interglacial MIS-7.1), and 87.7 ± 1.4 ka (transition from interglacial MIS-5.2 to MIS-5.1) (Figs. 11A, B).

4) Limited speleothem deposition during glacial periods with less precipitation than in cases 1, 2 and 3, but possibly more than at present. These humid events occurred between 290 ka and 285 ka (transition from glacial MIS-8.5 to MIS-8.4), and at 157.2 ± 3.8 ka (glacial MIS-6.2) (Figs. 11A, B).

5) Periods with no speleothem deposition (hiatuses) due to dry conditions. These hiatuses occurred during the following periods: 285-255 ka, 240-225 ka, 185-160 ka, 155-142 ka, 109-89 ka, 86-0 ka.

Types 1 and 2 are defined in this study as “Negev Humid Periods” (NHP), as follows: NHP-4 from 350 ka to 310 ka; NHP-3 from 310 ka to 290 ka; NHP-2 from 220 ka to 190 ka, and NHP-1 – from 142 ka to 109 ka. During two older humid periods NHP-4 and NHP-3, the speleothem ages have large analytical errors. Their sub-division is based on the fact that NHP-3 speleothem age frequency is the highest at ~295 ka, i.e. glacial MIS-8.5, whereas the NHP-4 age frequency is the highest at ~320 ka and ~342 ka, i.e. interglacial MIS-9 (Fig. 12C). During periods older than 350 ka large analytical uncertainties of speleothem ages do not enable to determine whether the speleothem deposition was intensive or limited.

Based on the age frequencies, number of laminae, their thickness, and distribution in several caves it is possible to determine which of these four humid periods were the more humid. Speleothems from NHP-1 (Vaks et al., 2007) are found in six caves and the number of laminae and the age frequencies are the highest, their laminae are relatively thicker, suggesting they represent the wettest period. However, their cross-section is not continuous and alternate with short hiatuses suggesting clusters of short wet episodes with long droughts in between them. Speleothems from NHP-2 are found in four caves, their age frequencies are lower, and their cross section is <1.5 cm, possibly suggesting conditions were slightly drier than during NHP-1. Speleothems from NHP-3 and NHP-4 are found in two and four caves respectively, but their laminae are thin, and age frequencies are low, suggesting drier conditions than during NHP-2 and NHP-1 (Fig. 11A, B; 12A, B, C). NHP-4 was at its most intense in the southern Negev and more limited in the central Negev. This contrasts with NHP-3, 2 and 1, which were most intense in the central Negev, but limited or absent in the southern Negev. The short durations of the NHPs are in accord with the absence of calcite horizons in the regolith

soils of the southern Negev of Middle-Late Quaternary (Amit et al., 2006; Enzel et al., 2007).

8.3.1.3 Humid periods in Judea Desert

13 speleothem samples were dated in Kanaim Cave, Judea Desert. Their ages cluster at: 354 ± 19 ka (MIS-9-10), 324 ± 10 ka (MIS-9), four ages cluster between 270 ka and 240 ka (MIS-8), 207.5 ± 3.8 ka (MIS-7), three ages cluster between 160 ka and 140 ka (MIS-6), 131.4 ± 2.1 ka (Termination II and MIS-5.5), and two ages are 37.90 ± 0.25 ka and 31.46 ± 0.22 ka (late MIS-3) (Fig. 10A). Thus, most of the speleothem deposition in Judea Desert occurred during glacial periods, similar to the speleothem deposition in the northern Negev (Vaks et al., 2006) and Jordan Valley (Vaks et al., 2003). However, their cross sections are thinner, suggesting that the paleo-climate in the Judea Desert was slightly drier compared with the northern Negev and the Jordan Valley. Speleothems deposited during interglacial periods in Kanaim Cave (Fig. 10A) overlap with NHP-4, NHP-2 and NHP-1 from the central and southern Negev, with speleothem deposition periods of 200-196 ka and 137-123 ka from the northern Negev, and with speleothem deposition periods of 225-205 ka and ~ 135 ka in Jordan Valley, as found by (Vaks et al., 2003, 2006) and in this study (Table 1, Figs. 10B, 11).

8.3.2 Origin of precipitation in the Negev Desert under present-day conditions and during the past 550 ka

8.3.2.1 Present-day rainfall

The origin of present-day rainfall in Israel is determined by ($\delta D - \delta^{18}O$) relationship (Gat and Carmi, 1987; Ayalon et al., 1998). Precipitation from rain clouds forming above the EM Sea follows the local Mediterranean Meteoric Water Line (MMWL) with average d-excess values of +22‰ (Gat and Carmi, 1987; Ayalon et al., 1998), and sometimes even $>+40$ ‰ (Bar-Matthews et al., 2003-a; Kuperman, 2005) but with the same slope of

the global Meteoric Water Line (MWL) ($\delta D = 8 \times \delta^{18}O + 10$) (Craig, 1961) with d-excess values of +10‰. D-excess in rainfall from tropical origin (African monsoon) follows the MWL (Njitchoua et al., 1999), and the $\delta^{18}O$ values in rainfall south of Israel (Eastern Africa and Arabia) vary from -2‰ to +2‰ (GNIP database, Aggarwal et al., (2007)), thus isotopical signature of the rainfall from tropical origin is different from precipitation originated in EM Sea. Evaporation below the clouds causes the d-excess to be lower and the slope differs from the MMWL and the MWL (these are referred as evaporation lines). This is typical to present-day rainfall in the Jordan Valley (Vaks et al., 2003).

Most rain events in the northern Negev (Be'er-Sheva and Arad), including the majority of rain events of >10 mm, have d-excess between +16‰ and +34‰, and follow the MMWL. Only a small fraction of the rain events follow the MWL, or fall on evaporation lines with d-excess values as low as -13‰ (Table 2 and Fig. 20). This includes rain events with <10 mm, but also some rain events of >10 mm (in this study, 5 events). The ranges of $\delta^{18}O$ and δD values are such that it is impossible to differentiate between rainfall from tropical origin and events that underwent evaporation. Accordingly, data on the origin of each synoptic system was obtained from Israel Meteorological Service (ref: <http://www.ims.gov.il/>), Israel Weather (ref: www.israelweather.co.il) and the Skiron Forecasts of University of Athens (ref: <http://forecast.uoa.gr/forecastamg.html>). Three events with >10 mm were of tropical origin (Fig. 20), and two events associated with Atlantic-Mediterranean fronts occurred during the warmer months of October and April and their d-excess thus reflects evaporation. Particular events at Be'er-Sheva station gave the most negative d-excess values (-8‰ to -13‰), with the data defining an evaporation line with an equation: $\delta D = 4.9 \times \delta^{18}O + 4.0$ (Figs. 20, 21C).

The average (weighted mean (Ayalon et al., 1998)) isotopic compositions of Be'er-Sheva and Arad rains give d-excess values of +20.2‰ and +19.4‰ respectively (Figs. 20, 21C). These values are close to MMWL but slightly lower than average d-excess of

rains from central and northern Israel, possibly because of the combination of evaporation effects and the inclusion of several events of tropical origin.

In the central and southern Negev Desert, rain events have d-excess between +41‰ to -12.6‰ (this study and Kuperman (2005)) (Table 2, Fig. 21-A, B), with averages of +15.1‰ in Mitzpe-Ramon, +13.4‰ in Makhtesh-ha-Qatan (this study and Kuperman (2005)), +8.8‰ in Neot-Smadar, and +13.2‰ in Ein-Netafim spring water near Elat. The average isotopic compositions of the rainfall in Mitzpe-Ramon, Makhtesh-ha-Qatan and Neot-Smadar follow the evaporation line defined for Be'er-Sheva rainfall (Fig. 21C), which makes it difficult to differentiate between rainfall of tropical origin and evaporated rain. Based on data of Israel Meteorological Service, Israel Weather and Skiron Forecasts of University of Athens in Mitzpe-Ramon more than half of the rainfall originated in EM Sea, whereas in Makhtesh-ha-Qatan and in Neot-Smadar ~60% of the rainfall was from tropical origin. However, Kuperman (2005) reported that during the winter 2002-2003 at Makhtesh-ha-Qatan the average d-excess of rainfall was +27‰, indicating that significant amounts of EM rainfall can reach this region. The d-excess value of Ein-Netafim spring water near Elat is +13.2‰, probably suggesting mixing between EM and tropical origin rainfall. This spring, like the speleothems, represents relatively large rain events that infiltrate into the ground and not undergo significant evaporation.

8.3.2.2 *Origin of precipitation in the past*

Speleothem FI $\delta^{18}\text{O}$ - δD relationships can be used as a major proxy to define the precipitation source (Genty et al., 2002; Matthews et al., 2000; McGarry et al., 2004). Since only the δD of FI was measured in this study, the $\delta^{18}\text{O}$ of the FI water were calculated from the measured $\delta^{18}\text{O}$ of the speleothem calcite, using the paleo-temperature equation of (O'Neil et al., 1969):

$$(2) \delta^{18}\text{O}_{\text{calc}} - \delta^{18}\text{O}_{\text{w}} = 2.78 \times 10^6 \times T^{-2} - 2.89$$

Where $\delta^{18}\text{O}_{\text{calc}}$ is $\delta^{18}\text{O}$ of the speleothem, $\delta^{18}\text{O}_w$ is $\delta^{18}\text{O}$ of the paleo-cave water, and the T is cave temperature in Kelvin. Cave temperatures approximate average annual surface temperatures (Poulson and White, 1969; Gascoyne, 1992). The paleo-temperatures used in the $\delta^{18}\text{O}_w$ calculation were taken from sources such as EM Sea surface temperatures (SST) (Emeis et al., 1998, 2003; Kallel et al., 2000; Kolosovsky, 2003), which approximate land temperatures, and terrestrial paleo-temperatures estimated from FI in caves from central and northern Israel and the Jordan Valley (McGarry et al., 2004). Correction for different topographic elevations of cave sites are made by subtracting 1°C for each 150 m change asl. For the time period between 161 ka and 157 ka there are no studies of the paleo-temperature of the EM Sea or other caves in Israel, and the calculated temperature from Western Mediterranean were used (Martrat et al., 2004) assuming that SST in western Mediterranean were $\sim 4\text{-}5^\circ\text{C}$ lower than in EM. With this assumption, paleo-temperatures from this period were $3\text{-}4^\circ\text{C}$ lower than present. The calculated paleo-temperatures for each cave are summarized in Table 3. Because of inherent inaccuracy in the temperature calculations a $\pm 2^\circ\text{C}$ range of temperature was taken for each time period (McGarry et al., 2004). The results are summarized in Fig 27A-D.

$\delta^{18}\text{O}$ and δD of the paleo-cave waters show a wide range, with d-excess values above the MMWL and values below the MWL (Figs. 27A, 28). However, it is possible to characterize these relationships for each time period and for individual caves. For glacial periods only Tzavoa Cave was studied. All samples from Tzavoa Cave either fall above the MMWL (between 177ka and 172 ka), or follow the MMWL during the last glacial (67-65 ka, 64-61 ka and 35-32 ka) (Figs 27B, 28). NHP-1 was studied in Tzavoa, Hol-Zakh, Ashalim and Ma'ale-ha-Meyshar showing that in Tzavoa also during interglacial NHP-1, similar to the glacial periods, the $\delta^{18}\text{O}$ - δD relationships follow the MMWL, indicating an EM source. NHP-1 samples in Ashalim and Ma'ale-ha-Meyshar caves either fall above or close to the MMWL, also implying an EM source. One sample in

Hol-Zakh aged 132-131 ka falls between the two lines (Figs. 27C, 28). During NHP-2, Tzavoa and Ma'ale-ha-Meyshar caves fall near the MMWL whereas Ashalim cave shows completely different pattern with all samples falling on and below the MWL (Figs. 27D, 28). This could imply a tropical source or evaporation. The main difference between the FI from Ashalim Cave during NHP-2 and those from this period at Tzavoa and Ma'ale-ha-Meyshar caves is that $\delta^{18}\text{O}$ values of the first are ~ 1.5 to 3‰ higher (-7.3‰ - -8.1‰), (Fig. 27D). It is not clear why the Ashalim speleothems are ^{18}O enriched, since evaporation will also lead to significant increase in their δD values. Also, Ashalim Cave's location between Tzavoa Cave and Ma'ale-ha-Meyshar Cave (Fig. 1C) make it difficult to suggest different origin for rainfall. A possible explanation is cooler temperatures. The entrance to the Ashalim Cave has a funnel shape up to ~ 6 m depth and the cave chambers are located below the funnel. This cave morphology is known to trap cold air during winter (Forbes, 1998) and probably caused the $\delta^{18}\text{O}$ values of Ashalim speleothems to be 1-2 ‰ more positive than in other caves during NHP-1.

It was impossible to measure δD of fluid inclusions of speleothems between 220 ka and 550 ka owing to their very thin laminae.

Another important evidence for a dominant EM Sea precipitation source comes from the petrography of NHP-1 and NHP-2 speleothems, which show a decrease in their volume along a north-south transect (Fig. 29A-I). To the south of Ma'ale-ha-Meyshar Cave, laminae of NHP-2 and NHP-3 disappear from the caves and the cross sections of speleothems from NHP-1 decrease from >1.5 cm to a few mm. Only the NHP-4 speleothems are more common in southern Negev than in central Negev, possibly indicating an increased supply of moisture from the tropical source (Fig. 29-J). Thinning of speleothem cross sections from the north to the south is a part of a regional scale phenomena in which the length of the depositional periods (Fig. 29-J) and the thickness of entire cross sections of speleothems from Israeli caves decrease drastically along the north-south transect from central Israel to the central and southern Negev. In Soreq Cave

the height of largest stalagmites (from base to the top) exceeds 8 m, whereas in the northern Negev (Ma'ale-Dragot, Tzavoa and Izzim Caves), the largest stalagmites are ~2-2.5 m in height. Further to the south large stalagmites are absent from the caves and the widest cross sections of stalagmites, flowstone and stalactites from the bedrock to the youngest laminae do not exceed 0.5 m.

Other criteria for the origin of rainfall in the Negev speleothems come from the comparison of their $\delta^{18}\text{O}$ profiles with those of speleothems from central and northern Israel (Bar-Matthews et al., 2003-a) and the northern Negev (Vaks et al., 2006). In central and southern Negev there is a large difference between the glacial (-1.5‰ to -5‰) and interglacial (-6.8‰ and -11.4‰) $\delta^{18}\text{O}$ values. The variations between glacial and interglacial $\delta^{18}\text{O}$ values are similar or even higher than in speleothems from central and northern Israel (Bar-Matthews et al., 2003-a) and from the northern Negev (Vaks et al., 2006) (Figs. 15A, 30A). This implies that the cave water $\delta^{18}\text{O}$ values reflect the glacial-interglacial variability originating in the same rain source: the EM Sea. On the other hand, during NHP-2 and NHP-1 $\delta^{18}\text{O}$ profiles of speleothems from the central Negev have lower variability compared with speleothems from central and northern Israel and the northern Negev (Fig. 30). In the latter, there is a sharp change during the Termination-II at 135 ka, and at 127 ka, with minimum values between 128-121 ka whereas in the central and southern Negev, there is very little change throughout the entire period from 133-115 ka (Fig. 30). Moreover, the $\delta^{18}\text{O}$ values of speleothems from central Negev for this period are lower by 1-3‰ compared to those from central and northern Israel, but similar to those of Tzavoa Cave in the northern Negev (Figs 30 B, C). The lower variability of $\delta^{18}\text{O}$ values of the speleothems from the central and southern Negev could may imply that the rainfall source (at least at the beginning and the end of NHP-1 and NHP-2) was of tropical origin, and during the peak of these humid periods the EM source became dominant. This pattern is not evident from the FI data, most probably because of its lower resolution, but also contradicts the speleothem thickness

trend from north to south. A more likely possibility is that Rayleigh fractionation isotopic processes (Vaks et al., 2006) and amount effect (Dansgaard, 1964; Bar-Matthews et al., 1997, 2003-a) mask the fluctuation in the $\delta^{18}\text{O}$ of rainfall as source indicator. A correlation of the lowest $\delta^{18}\text{O}$ values with peak speleothem age frequencies (Fig. 30 B, C) at 208-197 ka and at 132 ka, 126 ka and 117 ka, supports increased rainfall amounts during these time intervals.

The decrease in $\delta^{18}\text{O}$ of speleothems from central Israel to the Negev during NHP-1 and NHP-2 contrasts with the north-south increasing $\delta^{18}\text{O}$ trend observed for present-day rainfall. This contrast can be explained by the very small present-day rainfall amounts, which enable evaporation processes to obscure possible Rayleigh fractionation isotopic effects. Rayleigh fractionation isotopic processes with increasing distance from the EM Sea coast / elevation occur in central and northern Israel, where precipitation amounts are large (Ayalon et al., 2002; Vaks et al., 2006).

To conclude the speleothem record indicates that EM Sea was the major source for the precipitation in the central Negev, with a possible minor contribution of rainfall from tropical sources in the southern Negev.

8.3.3 Vegetation in the central and southern Negev Desert determined from the $\delta^{13}\text{C}$ of the speleothems and its relation to regional climatic change

Speleothem $\delta^{13}\text{C}$ values in central and southern Negev Desert vary between 1.5‰ and -8.4‰ and in most cases are much higher than those of the northern Negev (Vaks et al., 2006), Jordan Valley (Vaks et al., 2003), and central and northern Israel (Bar-Matthews et al., 2003-a) (Figs. 18, 19, 31). $\delta^{13}\text{C}$ values of speleothems from central and northern Israel were interpreted to reflect both the vegetation type and soil-water rock interaction (Bar-Matthews et al., 1996, 1997, 2003-a; Frumkin et al., 1999; 2000). $\delta^{13}\text{C}$ values between -9‰ to -13‰ were interpreted to usually reflect C3 type vegetation and higher values were attributed to an increase in the proportion of C4 type vegetation. However,

(Bar-Matthews et al., 2000, 2003-a) also argue that during certain time intervals that correlate with periods of sapropel formation in the EM Sea (sapropel 1 between 9-7 ka, and sapropel 5, during MIS-5.5 between 127-120 ka), the combination of lowest $\delta^{18}\text{O}$ values with highest $\delta^{13}\text{C}$ values (-7‰ to -9‰ and 1‰ to -6‰ respectively), is indicative of deluge periods with increased ^{13}C contribution from host rock and atmospheric CO_2 . Frumkin et al. (2000) on the other hand relate this combination during sapropel 5 to very dry conditions. In the northern Negev, $\delta^{13}\text{C}$ values range from ~0‰ to -9.5‰ with an average of -6‰ (Vaks et al., 2006). The higher $\delta^{13}\text{C}$ values from 1.5‰ to -8.5‰ in central and southern Negev (Figs. 18, 19, 31) are considered to reflect either of the dominance of scarce C4 vegetation when the values are in the range of -2.5‰ to -8.5 ‰, and/or higher contribution of host rock and atmospheric CO_2 when $\delta^{13}\text{C}$ values are isotopically heavier.

The highest speleothem $\delta^{13}\text{C}$ values usually occur at the beginning of each NHP and decrease during the period to the lowest values at its end (Figs 18, 19, 31). This change correlates well with the highest speleothem age frequencies occurring at 208-197 ka, 133-125 ka, 118-116 ka (Fig 31B,C), suggesting that the lowering in $\delta^{13}\text{C}$ is indicative of increased precipitation and the development of more vegetation above the cave. As the frequency of speleothem deposition decreases between 126 ka and 118 ka, $\delta^{13}\text{C}$ shows a corresponding increase. This period is one of the warmest in the region during the last 220 ka with EM average SST of 22-24°C and annual mean of 17-22°C in Soreq Cave area (Emeis et al., 2003; McGarry et al., 2004); this is suggestive of increased evaporation, and hence lower effective precipitation, relative to the periods at 133-126 ka and 118-116 ka. A decrease in speleothem $\delta^{13}\text{C}$ between 205 ka and 197 ka and between 128 ka and 126 ka in northern and central Negev is associated with the sharp increases in speleothem $\delta^{13}\text{C}$ (from -11.5‰ to -6.5‰ and from ~-13‰ to 1‰, respectively) in Soreq and Peqi'in Caves from central and northern Israel (Fig. 31B, C), and in Jerusalem Cave (Frumkin et al., 2000). As noted earlier, Frumkin et al. (2000) interpreted the latter

increase in speleothem $\delta^{13}\text{C}$ in the Jerusalem Cave to be a result of aridification, reduced vegetation, forest fires and the stripping of the soil cover. However, the fact that speleothems grew in the northern and central Negev, indicate that during this period the Negev Desert was relatively wet. Moreover, speleothem fluid inclusions from Tzavoa Cave during this period have d-excess ranging between 53‰ to 34‰, and in the more southern caves d-excess values range between 37‰ and 14‰ (Fig. 28), indicating that the EM Sea was the major moisture source for the Negev. The fact that $\delta^{13}\text{C}$ values in Tzavoa and Hol-Zakh decreased from -2‰ to -7.5‰ and further to the south the $\delta^{13}\text{C}$ values increased from $\sim -3.8\text{‰}$ to $\sim -2\text{‰}$ may suggest a relative decrease in rainfall from the northern Negev, which is closer to the EM Sea compared to the central Negev. Thus the speleothem record from the Negev desert probably indicates an increase in intensity of the Atlantic-Mediterranean (Cyprus) cyclones resulted in extremely wet conditions above the EM Sea as evident also from the formation of EM sapropels which are suggested to reflect increase precipitation over the entire Mediterranean Sea (Kallel et al., 1997, 2000). Bar-Matthews et al., (2000, 2003-a) interpreted the increase in $\delta^{13}\text{C}$ values to reflect deluge events in central and northern Israel that resulted in fast water infiltration to the cave with little interaction with soil CO_2 . In northern Negev the amounts of rainfall also increased but were lower, with no deluge effects, resulting in increase in vegetation cover and decrease in speleothem $\delta^{13}\text{C}$ values. The increase in rainfall in the region is also supported by Rossignol-Strick and Paterne (1999) who recorded an increase in deciduous oak pollen in sapropel sediments indicating precipitation amounts of 650 mm or higher.

Only short episodes of speleothems formation occur in the central Negev in Makhtesh-ha-Qatan Cave at ~ 157 ka (glacial MIS-6.2), with $\delta^{13}\text{C}$ values of -8.4‰ (Fig. 18, 31A). These are the lowest values known for the central Negev speleothems. This speleothem is phreatic and the cave was probably located below the stream channel of Wadi Hazera. Therefore it is possible that the low speleothem $\delta^{13}\text{C}$ indicates vegetation that locally

developed in the wadi channel and does not reflect the entire region vegetation. Another short event of speleothem deposition in the central Negev occurred between 89-86 ka; the speleothems $\delta^{13}\text{C}$ values are $\sim -4.2\text{‰}$ to -5.6‰ , suggesting the dominance of C4 semi-desert type vegetation.

8.3.4 Correlation between the Negev Humid Periods and the regional and global climate changes

Good correlation was found during the NHP-1 between the timing of $\delta^{18}\text{O}$ decrease between MIS-6.1 and MIS-5.5 at 132 ka in the Soreq Cave speleothems and the timing of maximum speleothem deposition in the Negev. The second high peak of speleothem deposition in the Negev at 127-126 ka correlates with the $\delta^{18}\text{O}$ decrease to the minimal values at the climax of MIS-5.5 in Soreq Cave. There is also a good correlation between one of the negative peaks of the $\delta^{18}\text{O}$ in Peqi'in Cave at 200 ka and the highest peak of the speleothem deposition in the Negev during the NHP-2 (Fig. 30). Bar-Matthews et al. (1997, 2000, 2003-a) interpreted these decreases in $\delta^{18}\text{O}$ values as associated with increased rainfall in central and northern Israel. As was shown in chapter 8.3.2.2, the precipitation in the Negev Desert during the NHPs originated at EM Sea, indicating that strongest humid episodes above the EM Sea brought the precipitation to the Negev Desert. Not all humid periods described by Bar-Matthews et al., (1997, 2000, 2003-a) reached the Negev Desert because of its southern position.

Episodes of speleothem deposition NHP-s 4, 3, 2 and 1 in the central and southern Negev Desert are very short and coincide with: periods of high amplitude oscillations in the solar energy in the northern and southern hemispheres (Berger, 1978) (Fig. 32), peaks of northern Hemisphere insolation, an African monsoon index of $\geq 51 \text{ cal/cm}^2 \times \text{day}$ (Rossignol-Strick (1983), for explanation see caption of Fig. 33) and the formation of the Mediterranean sapropels (Rossignol-Strick and Paterne, 1999) (Fig. 33).

NHP-4, 3, 2 and 1 occurred contemporaneously with speleothem deposition in Hoti Cave, northern Oman, at 330-300 ka, 200-180 ka, and 135-120 ka (Fig. 11C – Ref. 1). Fleitmann et al. (2003) suggested that speleothem deposition coincided with the northward migration of the ITCZ and intensification of the Indian Monsoon. Comparison between the timing of NHPs and humid periods from other parts of the southern Saharan-Arabian Desert such as north-western Sudan and central and southern Egypt, shows that wet periods in the southern part of the Sahara desert coincide with wetter conditions in the central and southern Negev (Fig. 11, Refs. 2-4) (Szabo et al., 1995; Crombie et al., 1997; Osmond and Dabous, 2004); however the large age uncertainties of the latter does not enable a precise correlation with NHP-s.

Since this study demonstrates that the EM was the major precipitation source to the Negev Desert, this link between maxima of the African monsoon and increased precipitation from Mediterranean sources suggests that the intensity of Atlantic-Mediterranean Cyclones must have increased contemporaneously with the African Monsoon. This was also suggested by Almogi-Labin et al. (2004), Bar-Matthews et al. (2000, 2003-a), based on the timing of the anomalously low $\delta^{18}\text{O}$ and anomalously high $\delta^{13}\text{C}$ values of speleothems from the central and northern Israel (Fig. 34). However, all the NHP commenced several thousand years before the peaks of maximum insolation and the beginning of the sapropel formation, and they ended several thousands year afterward (Figs. 34, 35). This suggests that the NHP were triggered before the monsoon maxima.

A change in the isotopic composition of speleothems from central and northern Israel during Termination II and the transition from MIS 7-2 to MIS 7-1, and the sharp increase in the age frequencies during NHP-1 and 2, at 204 ka and 135 ka (Fig. 34), coincides with rapid global warming recorded by speleothem deposition in Alps (Spannagel Cave, Austria) (Spotl et al., 2002), the rise of the global sea level (Henderson and Slowey, 2000; Gallup et al., 2002; Antonioli et al., 2004), the start of speleothem deposition in

Oman (Fleitmann et al., 2003), and the timing of speleothem deposition in northern England (Baker et al., 1993).

The contemporaneous tropical and mid-latitude humid periods can have a common cause. During higher northern hemisphere summer insolation periods, the SST in subtropical Atlantic Ocean were higher causing the Azorean high pressure cell to weaken, enabling the intensive monsoon rainfall in the southern Sahara during the summer (deMenocal, 1995; 2004). The weaker Azorean high pressure cell potentially can lead to a more negative NAO index during winter, resulting in increasing rainfall above the Mediterranean Sea (Hurrell, 1995; Krichak and Alpert, 2005) and consequently above the Negev Desert. During the Holocene and MIS-11 (430-400 ka) no speleothems formed in the central and southern Negev Desert. It could be explained by lower northern Hemisphere insolation, lower African Monsoon Index (Fig. 32, 33) and stronger Azorean high pressure cell, which prevented from Atlantic-Mediterranean cyclones to reach as far to the south, as during the NHPs.

8.3.5 "North – south paradox" of the Negev paleoclimate

The speleothem record of northern Negev, Judea Desert and Jordan Valley shows that speleothem deposition was intensive during most of the two last glacial periods MIS-6, and MIS-4-2. These glacial speleothems are thick, indicating high effective precipitation. They were deposited at the time of high levels of Lake Lisan between 70-17 ka (Neev and Emery, 1967; Begin et al., 1974; Bartov et al. 2002, 2003; Bookman et al., 2006) and Lake Amora, the precursor of modern Dead Sea, which existed during MIS-6 (Waldman et al., 2007). Speleothems were also deposited during interglacials, mainly at 200-186 ka, 137-123 ka, ~118 ka and 85-76 ka, but their deposition was episodic compared to glacial speleothems, and their cross sections are thinner, indicating lower effective precipitation. During the MIS-5, the levels of the Lake Samra (the precursor of the modern Dead Sea

during MIS-5) were also lower, apart from a minor increase between 130-120 ka (Waldman et al., 2007), compared to the levels of glacial Lake Lisan.

In contrast to the northern Negev, Judea Desert and Jordan Valley, most deposition in the central and southern Negev Desert occurred during interglacials, with almost no deposition during the two last glacial periods. The cross sections of these interglacial speleothems are even thinner than those of the interglacial speleothems from northern Negev, indicating that the central and southern Negev speleothems were deposited under lower effective precipitation; nevertheless their deposition occurred throughout the entire Negev Desert. The absence of glacial speleothems from the central and southern Negev, when the climate was apparently wetter in the northern Negev and "rain shadow" desert, and the corresponding formation of thin interglacial speleothems in the central and southern Negev Desert, when conditions in the north were apparently drier, is considered as the "north-south paradox". This apparent paradox is emphasized by the fact that the major source of precipitation of the entire region was the EM Sea.

One of the factors limiting the penetration of rainfall and snow from the EM Sea to the central and southern Negev during the glacial periods could be the northern shift of the Sinai Mediterranean coast due to glacial sea level decrease (Fig. 36, Enzel et al. (2007)). This however does not explain the penetration of rainfall to the entire Negev during interglacial NHP-s. An increase in precipitation to at least 300-350 mm associated with Cyprus cyclones could explain speleothem deposition in the entire Negev Desert during the interglacials. The intensity of these cyclones during the glacial periods could have been lower, bringing less than 200 mm of annual precipitation to the central and southern Negev, not enough for speleothem deposition. This in turn implies that the amount of precipitation during glacials in the northern Negev, Judea Desert and Jordan Valley was at least higher than 200 mm (as was discussed in chapter 8.3.1), enabling water infiltration to the caves.

Low temperatures would result in higher infiltration coefficients during glacial periods, giving a higher influx of water to caves and aquifers, and allowing deposition of thick speleothems in the northern Negev, Judea Desert and the Jordan Valley. The high lake levels of the two last glacial precursors of the modern Dead Sea, Lake Lisan and Lake Amora (Waldman et al., 2007), were probably the result of the high infiltration coefficients and possible snowmelt runoff into the Dead Sea Rift. Low infiltration coefficients during interglacial periods could also explain the low levels of Lake Samra during MIS-5 (Waldman et al., 2007).

Kolodny et al. (2005) calculated the relative humidity above the Lake Lisan during the last glacial period to be ~ 0.88 compared with values from 0.46 to 0.625 calculated for the present day Dead Sea (Gat, 1984; Krumgalz et al., 2000). The higher relative humidity during glacial most probably led to less evaporation of rainfall and from soil surface, resulting in increased amounts of water infiltrating to the caves and aquifers. This most probably led to a rise of the regional water table and higher lake levels.

An increase in infiltration of the rainwater occurred also in the Negev highlands during the last glacial period, as evident by travertine deposition in the central Negev springs (Schwarcz et al., 1979). However, this increase in infiltration evidently was not enough to form speleothems, possibly because of the small recharge areas of the caves, compared to the large catchments of the springs.

Thus it is suggested that Negev "north-south paradox" can be explained as follows. During the last glacial periods Cyprus cyclones brought more than 200 mm of annual precipitation to the area of the present day 150-160 mm isohyet (Tzavoa Cave). This amount was enough for the development of steppe vegetation above Tzavoa Cave. The fact that $\delta^{13}\text{C}$ values of the northern Negev speleothems are always higher (Vaks et al., 2006) than those of contemporaneous speleothems from central and northern Israel, indicates that the amount of precipitation reaching the northern Negev during glacial periods must have been lower. In contrast, the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of the speleothems in the

Jordan Valley, are similar to those of central and northern Israel (Vaks et al., 2003) implying higher precipitation than in the northern Negev. In both regions low temperatures and high infiltration coefficients enabled water to enter the caves and form thick speleothems. In the drier Judea desert (Kanaïm Cave) precipitation reached more than 200 mm only for short time episodes during the two last glacial periods, as evident from the relatively thin laminae. In the central and southern Negev precipitation must have been below ~200 mm as evident from the lack of speleothem deposition apart from NHP-3.

During the interglacials NHP-1, 2 and 4, the intensity of Cyprus cyclones must have been higher compared with glacial periods, bringing precipitation of at least 300-350 mm as far south as present day 50 mm isohyet (Ma'ale-ha-Meyshar Cave). However, evidently temperatures were high and infiltration coefficients were low, thus bringing only small amounts of water to the caves and resulting in the formation of only thin speleothem laminae.

8.3.6 Pliocene-Pleistocene humid periods in the central and southern Negev Desert as evident from speleothems older than 550 ka

The Negev speleothems of the Basal member with four preliminary U-Pb ages of 3.3-2.7 Ma (see chapter 8.2.1.1) must have been deposited under humid conditions, they are massive (up to 40 cm in width), contain large crystals with very little signs of hiatuses in between laminae. In the Intermediate and Young members the number of hiatuses increases and the laminae become thinner. The change in the appearance and the petrography towards the younger part are indicative of drying since middle Pliocene (Fig 3, 4). The drying-out of the Negev Desert during the last ~ 2.7 Ma occurred together with the desertification of the Sahara (deMenocal, 1995; 2004).

The Basal Member was probably deposited contemporaneously with the Pliocene Arava Formation lakes and (Avni, 1997; Avni et al., 2001-b) suggested that these lakes

formed at 4 - 2 Ma. The lake sediments contain lacustrine marls including abundant freshwater fauna, and they covered area larger than 1300 km² in the southern Negev and Arava Valley. These lakes were formed under humid conditions except for one hypersaline event (Avni, 1997). This is very similar to the massive appearance of the Basal member showing very few signs of hiatuses.

The lamina ASH-15-C from the base of the Intermediate member of Ashalim Cave has a preliminary U-Pb age of 1.269 ± 0.013 Ma (see chapter 8.2.1.1) which suggests that these speleothems formed contemporaneously with lakes of Early Pleistocene Zhiha formation (Ginat et al., 2003). The major lake (Lake Zihor) formed in the southern Negev at ~1.4 Ma and covered area of 18 km². The lake sediments show 3 periods of high stand with abundant freshwater fauna and periods of desiccation in between. The latter were marked by the formation of reddish soils containing carbonate horizons. Annual precipitation varied from 150 mm to 500 mm during the deposition of the lake sediments (Ginat et al., 2003). This is similar to the discontinuous formation of the Intermediate member, which shows more evidence of hiatuses than the Basal member.

Middle-Late Pleistocene sediments in the southern Negev include series of alluvial terraces up to 20 m above the present day wadi channels. Gypsic-salic soils typical to present day hyper arid climate formed on this terraces. (Avni et al., 2001-a) suggests that their formation was controlled mainly by climatic variations. Thus it is possible that the Young member of the speleothems formed during these phases of terraces formation.

The fact that several growth / hiatus cycles can be observed in single speleothem over >3 Ma shows that in the Negev Desert environment the water conduit system from the surface to the caves (fractures) seem to remain open over long times. It is the reflection of absence of prior precipitation of calcite in the cracks due to the arid conditions.

$\delta^{18}\text{O}$ values of the Pliocene Basal member speleothems in Ashalim Cave range from -9.4‰ to -11.7‰ and $\delta^{18}\text{O}$ values of the Intermediate member in Ashalim Cave and Even-Sid-Ramon mini-caves are -9.3‰ - -10‰. All fluid inclusion δD values range

between -32‰ and -55‰. These values are similar or slightly lower than those of the speleothems deposited during NHP-4, 2 and 1. The lack of independent temperature estimates for these periods make it impossible to reconstruct the source of precipitation.

8.3.7 Phreatic speleothems in Makhtesh-ha-Qatan as markers of paleo-groundwater levels

Phreatic speleothems of Makhtesh-ha-Qatan Cave were dated at ~290 ka and older, with one lamina deposited at ~157 ka. At present the cave is located 4 m above the present day channel of Wadi Hazera (an ephemeral stream). In the past, before the incision of the Wadi Hazera channel, the cave was probably located below the ancient channel. Speleothems deposition suggests higher ground water level, or periods when Wadi Hazera was a perennial stream and the cave was flooded.

Kronfeld and Livnat (1987) found that tufa was deposited at Makhtesh-ha-Qatan area, about 1.5 km to the west, during the last glacial period (between 73 ka and 30 ka) and during older episodes. However, no speleothems were deposited in Makhtesh-ha-Qatan Cave during the last glacial, probably because the precipitation amounts were too low to form vadose speleothems (like in central Negev, section 8.3.5), and groundwater levels in Judea Group aquifer were too low to form phreatic speleothems. Because of large analytical uncertainties of Kronfeld and Livnat (1987) ages it is impossible to compare their older tufa ages to the speleothem ages determined in Makhtesh-ha-Qatan Cave.

8.3.8 – Humid periods in the northern Saharan-Arabian Desert as climatic “windows” for Out of Africa dispersals of hominids and animals

Humid episodes in the Saharan-Arabian Desert were an important control of hominid migration from the African continent to the other parts of the world. This was discussed in detail as part of this thesis in the article published in *Geology* (Vaks et al., 2007). In this paper, the emphasis was on the critical importance of NHP-1. Similarly, all other

NHP-s and older humid periods could have played very important role as a “climatic windows” controlling the dispersion of hominids and animals out of the African continent. Derricourt, (2005) argues that Sinai-Negev land bridge was the major and probably the only route out of Africa before the early modern humans developed the ability of seafaring during the last glacial period. The study of Fernandes et al. (2006) argues for the absence of post-Miocene land bridges across the Red Sea, with the exception of the Sinai-Negev Desert. Thus, wet climate conditions in this "bottleneck" region were of great importance during the Late Pliocene and Pleistocene.

Today the Nile River provides a way across the Sahara, but the usability of this route by hominids and animals in the past was highly dependant on the intensity of the African monsoon. Interglacial phases of high monsoonal intensity increased the Nile flow, and other Sahara-crossing rivers possibly formed (Rohling et al., 2002). During the phases of low monsoon intensity, mainly during glacial periods, the flow of the Nile decreased, becoming discontinuous in extremely arid conditions similar to Last Glacial Maximum (Almogi-Labin et al., 2004), when southward expansion of Sahara Desert reached 5-11°N (Roche et al., 2007). Said (1993) argues that the sources of River Nile did not reach as far as tropical Africa until about 800 ka. Thus, the Saharan-Arabian Desert that began to form at ~ 2.7 Ma (deMenocal, 2004) could have been an even more significant barrier for migration out of Africa for the ~2 millions years, prior to the establishment of the connection between the Nile and the tropical Africa. This emphasizes the importance of wet phases in the Sahara and Sinai-Negev land bridge for the out of Africa dispersal during the Early Pleistocene.

Hand axes, similar to those found in the Ubeidiya site in northern Jordan Valley and dated at ~1.4 Ma, were also found by Ginat et al. (2003) (and references within) in sediments of the Early Pleistocene Lake Zihor in the southern Negev, identifying a probable route between the southern Negev Desert and the Jordan Valley.

NHP-3, 4 and earlier humid periods in the Negev also could be associated with migrations of hominids and animals. During the NHP-2 there is data showing that between 200 ka and 195 ka, humid conditions prevailed both in the northern and southern parts of Saharan-Arabian Desert (Szabo et al., 1995; Crombie et al., 1997; Fleitmann et al., 2003; Osmond and Dabous, 2004; and this study). This potentially opened the passage through Sahara to the Levant. Recently the early Mousterian site in Rosh-Mor, central Negev was dated to ~200 ka, supporting this idea (Rink et al., 2003).

8.3.9 Sr isotope ratios from Israeli desert speleothems as tracers of the origin of dust and host rock weathering

Sr substitutes for Ca in the speleothem calcite lattice (Fairchild et al., 2001). Three major sources potentially supply Sr to the groundwater reaching the cave system: host rock, the soil above the cave and sea spray (Ayalon et al., 1999; Bar-Matthews et al., 1999; Frumkin and Stein, 2004). Sr contribution from sea spray ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7092$) is most probably the highest near the coast and decreases inland. Sr isotopic ratios in Negev rainfall vary between 0.7079 and 0.7084, compared to 0.7078-0.7092 in coastal and northern parts of Israel (Herut et al., 1993), suggesting a very low contribution of the sea spray to the Negev rainwater relative to the northern and coastal region. Soils in Israel are dust-born, originating from distal Saharan-Arabian sources and from the proximal source of northern and central Sinai Peninsula (Yaalon, 1971; Yaalon and Ganor, 1973; Yaalon and Dan, 1974; Amit and Gerson, 1986; Tsoar and Pye, 1987). (Crouvi et al., 2007; Enzel et al., 2007) showed that dust that forms the Negev loess soils mainly originates from wind blown sediments of the marine shelf of the Nile Delta and the northern and central Sinai Peninsula.

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the Negev desert speleothems range from 0.7083 to 0.7085 for two last glacial periods at 160-155 ka and 75-13 ka; is lower and ranges between 0.7081 and 0.7084 during the glacial intervals between 310-285 ka and 176-172 ka and during the

interglacial periods 350-310 ka, 220-185 ka, 142-76 ka. Speleothem $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the Jordan Valley during two last glacial periods and interglacial intervals are similar to the Negev values. (Table 4, Fig. 24-25). In Soreq and Jerusalem Caves, central Israel the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in glacial and interglacial speleothems are very similar to each other (Ayalon et al., 1999; Bar-Matthews et al., 1999; Frumkin and Stein, 2004); during the two last glacial periods they range between 0.7083-0.7085 (similar to the desert speleothems), whereas during the penultimate and the Holocene interglacials the speleothem $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are 0.7078-0.7080 (Fig. 37). The glacial-interglacial fluctuations of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in speleothems from central Israel are matched by similar trends in Sr concentrations. During glacial intervals the Sr concentrations range between 55 and 200 ppm in the Jerusalem Cave (Frumkin and Stein, 2004) and 45 and 220 ppm in the Soreq Cave (Bar-Matthews and Ayalon, 2001), but during the interglacials, the Sr concentrations are lower between 7 and 55 ppm in Jerusalem Cave (Frumkin and Stein, 2004), and 50 and 170 ppm in Soreq Cave (Bar-Matthews and Ayalon, 2001) (Fig. 38). Similar trends, but with higher concentrations, are observed in the Negev speleothems. During the two last glacial periods concentrations range from 120 to 340 ppm in Ma'ale-Efrayim Cave, and 255 to 525 ppm in Tzavoa cave, whereas during the two last interglacials in Ma'ale-Efrayim Cave values are 100 to 260 ppm and Tzavoa Cave values are 60 to 150 ppm (Table 4, Fig. 23B,C, Fig. 38).

Speleothem $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Intermediate and Young members (0.7079-0.7085) lie between the host rock ratios (0.7074-0.7078) and the bulk soils values (0.7084-0.7110) (Fig. 39A). $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of Ma'ale-Efrayim and Tzavoa speleothems slightly approach the host rock composition during the interglacial periods but the soil compositions during glacials (Fig. 39C).

Frumkin and Stein (2004) interpreted the high glacial Sr concentrations and $^{87}\text{Sr}/^{86}\text{Sr}$ values to intensive dust fluxes from Sahara to the EM region due to the hyper arid conditions in the Saharan areas. Reduced speleothem Sr concentrations and $^{87}\text{Sr}/^{86}\text{Sr}$

ratios during the penultimate interglacial and the early-middle Holocene, were interpreted to reflect reduced dust fluxes from Sahara due to the intensive monsoon precipitation. In contrast, Ayalon et al. (1999) and Bar-Matthews et al. (1999) interpreted the same trends to reflect increase in chemical erosion of the host rock during the interglacial humid periods, whereas during glacials weathering was less intensive and the contribution from exogenic sources increased.

High interglacial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and Sr concentrations of the Negev speleothems relative to speleothems from caves in central Israel can be attributed to the fact that precipitation amounts in the Negev were always significantly lower. Because of the lower precipitation, chemical erosion of host rock was less significant, and the high Sr concentration and Sr isotopic values mainly originated from dust-born soil. Supportive evidence for low relative rates of chemical erosion in the central and southern Negev comes from the fact that speleothem cross-section become thinner southward by ~ 2 orders of the magnitude (see section 8.3.2.2), which is indicative of reduced rates of host-rock dissolution above the caves. It further supported by the study of Haviv et al. (2006) showing that the erosion rates in the arid and mildly arid zones in Israel (100-400 mm rainfall) are 1-4 m/Ma, whereas in Mediterranean climate zone, central Israel (500-600 mm), the erosion rates increase to 15-40 m/Ma.

The similar speleothem $^{87}\text{Sr}/^{86}\text{Sr}$ in caves from the Negev Desert, Jordan Valley and central Israel during the glacial periods (Fig. 37) suggests common sources. Frumkin and Stein (2004) suggest the Sahara Desert. However, other more proximal sources such as the Nile delta, and northern and central Sinai are possible (Crouvi et al., 2007; Enzel et al., 2007). The contribution of these sources is evident from the dust fall map of Ganor and Foner (1996; 2001) which shows an increase in dust fall above the north-eastern Sinai and northern and central Negev relative to north-western Sinai (Fig. 40). During glacial periods vast parts of continental shelf and the Nile Delta were exposed due to lower glacial sea level, and probably increased the dust contribution to the region during

the winter storms (Enzel et al., 2007) (Fig. 36). The similarity between $^{87}\text{Sr}/^{86}\text{Sr}$ of present-day rainfall in the Negev Desert (0.7079-0.7084) (Herut et al., 1993) and the $^{87}\text{Sr}/^{86}\text{Sr}$ of glacial speleothems from all Israeli caves (0.7083-0.7085) suggest that the main source remained the same.

During the interglacial periods there is a marked difference in Sr isotopic composition between Negev speleothems and central Israel speleothems. Whereas in central Israel the $^{87}\text{Sr}/^{86}\text{Sr}$ values are closer to that of the host rock, in the Negev they only slightly decrease relative to the glacial values (Fig. 37, 39). This supports a higher dust contribution of dust during interglacials to the Negev. Even during periods when Saharan dust source was possibly shut down (Frumkin and Stein, 2004), the Negev Desert continued to receive dust, most probably from the proximal Sinai source (Fig. 36, 40). The reason why such Sinai dust was not transported to the north is because it is composed of coarse silt particles ($>20\mu\text{m}$) traveling in low altitude, whereas dust from the Saharan-Arabian source is composed of fine particles traveling in high altitude in atmosphere (Frumkin and Stein, 2004; Enzel et al., 2007). During interglacial periods the coarse dust was transported from Sinai to the Negev by strong south-westerly and westerly winds of Cyprus cyclones, but washed down by rain before reaching the central Israel. This is very similar to present day situation, where dust fall is maximal in the Negev, but sharply decreases to the north (Ganor and Foner, 2001; Enzel et al., 2007) (Fig. 40). Supportive evidence for this process is the present-day $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Jordan Valley bulk soils and of their silicate components, which are 0.7110 and 0.7123 respectively, and nearer to the Saharan dust end member (0.7165-0.7200) (Krom et al., 1999) than the Negev soils, with bulk ratios of 0.7084-0.7100 and 0.7101-0.7115 in silicate components (Figs. 24, 26, 39A, Table 4). This difference may indicate northward enrichment in the Saharan dust source, whereas in the south the Sinai proximal dust source is more dominant. The $^{87}\text{Sr}/^{86}\text{Sr}$ and $1/\text{Sr}$ of the bulk soils define a mixing line between the two dust sources, starting at ~ 0.7084 , which is also the average composition of most of the speleothems

(Fig. 39A). Consequently, this is probably an average $^{87}\text{Sr}/^{86}\text{Sr}$ of the Sinai proximal source.

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Base Member speleothems formed at 3.3-2.7 Ma are 0.7077-0.7079, similar to the host-rock values (Fig. 39B), whereas in the Intermediate (~ 1.3 Ma) and Young member the ratios are 0.7079-0.7085 (Fig. 39A-C). (deMenocal, 1995; 2004) suggested that the formation of Saharan-Arabian desert belt started at ~ 2.7 Ma. Since Intermediate and Young members were deposited later, it is possible that there was higher contribution of dust from Sahara-Sinai deserts during their deposition.

9. Summary and conclusions

Speleothem deposition in present-day arid regions of Israel indicates that humid climatic conditions (i.e. periods with positive effective precipitation/infiltration index) occurred in the past. Effective precipitation tends to decrease with increasing temperature and evaporation. The minimum precipitation amounts necessary to deposit speleothems are estimated in this study to be: 200-275 mm/year during glacial periods and 300-350 mm/year during interglacial periods.

In the Jordan Valley Ma'ale-Efrayim Cave, speleothem deposition mainly occurred during the three last glacial periods, with minor deposition during Termination II (~135 ka) and MIS-7 interglacial (225-205 ka), and no speleothem deposition during the Holocene, the peak of last glacial maximum (~19 ka) and the previous interglacial MIS-5(1-4). This contrasts with the continuous speleothem growth that occurred in the Mediterranean climate zone on the western side of the Central Mountain Ridge (CMR) during the last 240 ka. Comparison of the present-day $\delta^{18}\text{O}$ and δD values of the cave and rain water on the western and the eastern sides of the CMR shows that at present, higher temperatures and greater evaporation on the eastern flank and in the Jordan Valley are the major influences on the isotopic compositions and lack of rainfall. The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ profiles of speleothems deposited between 67 ka and 25 ka in Ma'ale-Efrayim Cave

match the general trends of the isotopic profiles of Soreq Cave speleothems, suggesting the similar precipitation source (EM Sea) and similar climatic conditions. Thus, during the last glacial the desert boundary in the Jordan Valley effectively migrated further south from the present day location, whereas during the Holocene and the last interglacial the desert boundary was in its present-day position. The decrease in temperature and consequent increase in effective precipitation in the Jordan Valley are probably the major factors controlling the decay of the "rain shadow" effect during glacial periods.

Most of the speleothem deposition in the Tzavoa Cave, northern Negev, occurred during the glacial periods (as in Jordan Valley), whereas during the interglacial periods and glacial maxima speleothem deposition was limited and episodic. No speleothem deposition occurred at MIS-5(3-2), Younger Dryas and Holocene, indicating that climate in these periods was similar to present or even more arid. Holocene speleothems were only found in Ma'ale-Dragot caves located on the present-day 280-300 mm isohyets. These observations are consistent with the proposal that the precipitation that caused speleothem deposition during the interglacials in Tzavoa Cave was 300-350 mm (i. e. twice than the present day), whereas during the glacial periods, precipitation amounts were 200-275 mm, owing to cooler temperatures and reduced evaporation.

$\delta^{13}\text{C}$ values of the speleothems in the northern Negev during glacial periods (-3‰ - -9‰) show that vegetation was usually C3+C4 steppe, whereas the $\delta^{13}\text{C}$ values in Jordan Valley speleothems (-8‰ - -11‰) formed at the same time indicate a higher percentage of C3 Mediterranean vegetation, and wetter conditions. Both in the Jordan Valley and northern Negev the $\delta^{18}\text{O}$ values of the speleothems generally show the same overall trends as the Soreq Cave in the Mediterranean climate, pointing to the EM Sea as the precipitation source. δD - $\delta^{18}\text{O}$ relations of speleothem fluid inclusions also indicate the EM precipitation source in northern Negev and Jordan Valley. The $\delta^{18}\text{O}$ values of speleothems in the northern Negev are, however, systematically lower than in contemporaneous speleothems from the Jordan Valley and central and northern Israel.

This in part is attributed to the increased rainout of the heavy isotopes by the Rayleigh fractionation processes, possibly due to further distance of these caves from the Mediterranean Sea. The study shows that the maximum southward shift of the border between the Mediterranean climate and the semi-desert zone during the humid periods in the northern Negev was 20-25 km.

Major humid periods in the central and southern Negev Desert (Negev Humid Periods – NHP) during the last 350 ka occurred at: 350-310 ka (NHP-4), 310-290 ka (NHP-3), 220-190 ka (NHP-2), and 142-109 ka (NHP-1). NHP-4, NHP-2 and NHP-1 occurred during interglacial conditions (MIS-9, MIS-7(3-1) and Termination II + MIS-5(5-4) respectively), whereas NHP-3 is associated with glacial period (MIS-8). δD - $\delta^{18}O$ relations in the paleo-cave waters trapped in fluid inclusions indicate that the EM Sea was the major source of the rainfall during the last 220 ka in Jordan Valley, northern and central Negev. Another important evidence for major EM Sea rain source during the NHP-1, 2 and 3 is the decrease in speleothem amounts from north to the south. Intensive speleothem deposition occurred in the central Negev during the NHP-1, 2 and 3, whereas speleothem deposition was limited during NHP-4. In the southern Negev intensive speleothem deposition occurred only during the NHP-4, with limited speleothem deposition occurring during the NHP-1 and no speleothems were found from NHP-2 and 3. This comparison indicates that the major source of precipitation during NHP-1, 2 and 3 was the EM Sea, whereas tropical rain source probably became a prominent source during NHP-4. However, the record of the central and southern Negev is in contrast with northern Negev, Jordan Valley and Judea Desert, where the most of the speleothems were deposited during the glacial periods. Glacial speleothem deposition in the latter occurred due to precipitation amounts of 200 mm or higher associated with high infiltration coefficients during the cold glacial periods, whereas central and southern Negev remained dry. During the humid interglacial episodes the intensities of the Cyprus cyclones were higher than during the glacials, bringing precipitation high as 300-350 mm as far south as

the present-day ~50 mm isohyet. However, the higher temperatures and low effective precipitation caused only limited water infiltration to the caves leading only to deposition of thin speleothem laminae.

The major decreases of speleothem $\delta^{18}\text{O}$ in Soreq Cave at 132 ka and 127-126 ka and in Peqi'in Cave at 200 ka occurred simultaneously with the highest peaks of speleothem deposition in the Negev Desert during the NHP-1 and NHP-2. These events were probably associated with most intensive humid episodes above the EM Sea.

NHP-s were contemporaneous with periods of monsoon index of ≥ 51 ($\text{cal}/\text{cm}^2 \times \text{day}$) and formation of the sapropels in the Mediterranean Sea. Such simultaneous intensification of the monsoon and Atlantic-Mediterranean cyclones is probably related to the weakening of the high pressure cell above sub-tropical Atlantic Ocean. Absence of speleothem deposition in the Negev during the Holocene and MIS-11 (430-400 ka) interglacials could be explained by lower Northern Hemisphere insolation and lower African Monsoon index than during the NHPs. Thus, the high pressure cell above subtropical Atlantic Ocean was probably stronger, leading to more northern trajectories of Atlantic-Mediterranean cyclones and preventing from Mediterranean precipitation to reach central and southern Negev.

Speleothem $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the speleothems in the “rain shadow” and the Negev deserts usually vary between 0.7082 and 0.7085, and only slightly change from the glacial to interglacial conditions, compared to central Israel. This could be due to lower host rock dissolution rates, or higher dust supply. The dust supply to the Negev Desert was higher than to central Israel because it is located near the proximal Sinai-Nile Delta dust source. Speleothem and soil $^{87}\text{Sr}/^{86}\text{Sr}$ ratios probably indicate ~ 0.7084 as the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio for this source.

Most of the speleothems older than the uranium series dating limit deposited in the Negev belong to the Basal member and have preliminary U-Pb ages ranging between 3.3 Ma and 2.7 Ma, simultaneous with formation of the lakes of the Arava formation. A few

speleothems (Intermediate member) were deposited at ~1.3 Ma, probably simultaneously with the occurrence of the lakes of Zhiha Formation. A reduction in the rate of speleothem deposition, together with an increase in speleothem $^{87}\text{Sr}/^{86}\text{Sr}$ ratios between Basal and Intermediate-Young members, probably indicates increased desertification of this region and of entire Saharan-Arabian Desert after ~2.7 Ma, which caused increased dust supply from the Sinai and Saharan-Arabian sources.

Multiple growth / hiatus cycles in a single speleothem over >3 Ma show that in arid environment the fractures conducting water from the surface to the caves remain open over long times.

The Sinai-Negev land bridge was the major, and possibly the only land-bridge, connecting the Africa with Asia since the Miocene, and the Saharan-Arabian Desert was significant obstacle for hominid and animal migrations from Africa to other parts of the world. This study suggests that humid phases in the Sinai-Negev land bridge that occurred contemporaneously with humid phases in southern and central Saharan-Arabian Desert probably opened the climatic “windows” for migration of hominids and animals out of the African continent. Migration of the early modern humans from Africa to the Levant was probably associated with NHP-1 at 142-109 ka.

Bibliography:

- Aggarwal, P. K., Alduchov, O., Araguas, L., Dogramaci, S., Katzlberger, G., Kriz, K., Kulkarni, K. M., Kurtas, T., Newman, B. D., and Pucher, A. (2007). New capabilities for studies using isotopes in the water cycle. *EOS, Transactions, American Geophysical Union* **88**, 537-538.
- Almogi-Labin, A., Bar-Matthews, M., and Ayalon, A. (2004). Climate variability in the Levant and northeast Africa during the Late Quaternary based on marine and land records. In: (Goren-Inbar, N. and Speth, D. Eds) *Human paleoecology in the Levantine corridor*, p. 117-134.
- Amit, R., Enzel, Y., and Sharon, D. (2006). Permanent Quaternary hyperaridity in the Negev, Israel, resulting from regional tectonics blocking Mediterranean frontal systems. *Geology* **34**, 509-512.
- Amit, R., and Gerson, R. (1986). The evolution of Holocene Reg (gravelly) soils in the deserts; an example from Dead Sea region *Catena* **13**, 59-79.
- Antonioli, F., Bard, E., Potter, E.-K., Silenzi, S., and Imbrota, S. (2004). 215-ka history of sea level oscillations from marine and continental layers in Argentarola Cave speleothems (Italy). *Global and Planetary Change* **43**, 57-78.
- Auler, A. S., and Smart, P. L. (2004). Rates of condensation corrosion in speleothems. *Speleogenesis and Evolution of Karst Aquifers* www.speleogenesis.info **2**, p.2 (4 pages)
- Avni, Y. (1991). "The geology, paleogeography and landscape evolution in Central Negev Highlands and the western Ramon structure." M.Sc. (in Hebrew, English Abstract), Hebrew University.
- Avni, Y. (1997). Geological evolution of the central and southern Negev as an indicator of the evolution of the Dead Sea Transform western margin in the late Neogene and Quaternary. *Ph.D. Thesis (in Hebrew, English Abstract)*, 240 pp.
- Avni, Y., Bartov, Y., Garfunkel, Z., and Ginat, H. (2001-a). Evolution of the Paran drainage basin and its relation to the Plio-Pleistocene history of the Arava Rift western margin, Israel. *Israel Journal of Earth Sciences* **49**, 215-238.
- Avni, Y., Bartov, Y., Garfunkel, Z., and Ginat, H. (2001-b). The Arava formation - A Pliocene sequence in the Arava Valley and its western margin, southern Israel. *Israel Journal of Earth Sciences* **50**, 101-120.
- Avni, Y., Porat, N., Plakht, J., and Avni, G. (2006). Geomorphic changes leading to natural desertification versus anthropogenic land conservation in an arid environment, the Negev Highlands, Israel. *Geomorphology* **82**, 177-200.
- Ayalon, A., Bar-Matthews, M., and Sass, E. (1998). Rainfall-recharge relationships within a karstic terrain in the Eastern Mediterranean semi-arid region, Israel: $\delta^{18}\text{O}$ and δD characteristics. *Journal of Hydrology* **207**, 18-31.
- Ayalon, A., Bar-Matthews, M., and Kaufman, A. (1999). Petrography, strontium, barium and uranium isotope ratios in speleothems as paleoclimatic proxies: Soreq Cave, Israel. *The Holocene* **9**, 715-722.
- Ayalon, A., Bar-Matthews, M., and Kaufman, A. (2002). Climatic conditions during marine isotope stage 6 in the eastern Mediterranean region from the isotopic composition of the Soreq Cave, Israel. *Geology* **30**, 303-306.
- Ayalon, A., Bar-Matthews, M., Schilman, B. (2004). Rainfall isotopic characteristics at various sites in Israel and the relationships with unsaturated zone water. *Israel Geological Survey Report GSI/16/04*, 37pp.

- Baker, A., Smart, P. L., and Ford, D. C. (1993). North-west European paleoclimate as indicated by growth frequency variations of secondary calcite deposits. *Palaeogeography, Palaeoclimatology, Palaeoecology* **100**, 291-301.
- Baker, A., and Smart, P. L. (1995). Recent flowstone growth rates: field measurements in comparison with theoretical predictions. *Chemical Geology* **122**, 121-128.
- Banner, J. L. (1995). Application of the trace element and isotope geochemistry of strontium to studies of carbonate diagenesis. *Sedimentology* **42**, 805-824.
- Bar-Matthews, M., Matthews, A., and Ayalon, A. (1991). Environmental controls of speleothems mineralogy in a karstic dolomitic terrain (Soreq Cave, Israel). *Journal of Geology* **99**, 189-207.
- Bar-Matthews, M., Ayalon, A., Matthews, A., Sass, E., and Halicz, L. (1996). Carbon and oxygen isotope study of the active water-carbonate system in a karstic Mediterranean cave: Implications for paleoclimate research in semiarid regions. *Geochimica et Cosmochimica Acta* **60**, 337-347.
- Bar-Matthews, M., Ayalon, A., and Kaufman, A. (1997). Late Quaternary Paleoclimate in the Eastern Mediterranean Region from Stable Isotope Analysis of Speleothems at Soreq Cave, Israel. *Quaternary Research* **47**, 155-168.
- Bar-Matthews, M., Ayalon, A., Kaufman, A., and Wasserburg, G. J. (1999). The Eastern Mediterranean paleoclimate as a reflection of regional events: Soreq cave, Israel. *Earth and Planetary Science Letters* **166**, 85.
- Bar-Matthews, M., Ayalon, A., and Kaufman, A. (2000). Timing and hydrological conditions of Sapropel events in the Eastern Mediterranean, as evident from speleothems, Soreq cave, Israel. *Chemical Geology* **169**, 145-156.
- Bar-Matthews, M., Ayalon, A., Matthews, A., and Frumkin, A. (2000-b). Eastern Mediterranean Paleoclimate during the last 250,000 years as derived from the petrography, mineralogy, trace element and isotopic composition of cave deposits (speleothems), Israel. *Israel Geological Survey Report TR-GSI/10/2000*, 17 pp.
- Bar-Matthews, M., and Ayalon, A. (2001). Eastern Mediterranean paleoclimate during the last 250,000 years as derived from the petrography, mineralogy, trace element and isotopic composition of cave deposits (speleothems), Israel, pp. 73. Report GSI/41/01, Geological Survey of Israel, The Ministry of National Infrastructures, Jerusalem, Israel.
- Bar-Matthews, M., Ayalon, A., Gilmour, M., Matthews, A., and Hawkesworth, C. J. (2003-a). Sea-land oxygen isotopic relationships from planktonic foraminifera and speleothems in the Eastern Mediterranean region and their implication for paleorainfall during interglacial intervals. *Geochimica et Cosmochimica Acta* **67**, 3181-3199.
- Bar-Matthews, M., Almogi-Labin, A., Ayalon, A., Schilman, B., Kolosovsky, E., Aizenshtat, Z., McGarry, S., and Matthews, A. (2003-b). The relationship between the marine and terrestrial (speleothems) climatic record during the Quaternary in the Eastern Mediterranean region. *Israel Geological Survey Report GSI/10/2003*, 31 pp.
- Bartov, Y., Stein, M., Enzel, Y., Agnon, A., and Reches, Z. (2002). Lake Levels and Sequence Stratigraphy of Lake Lisan, the Late Pleistocene Precursor of the Dead Sea. *Quaternary Research* **57**, 9-21.
- Bartov, Y., Goldstein, S. L., Stein, M., Enzel, Y. (2003). Catastrophic arid episodes in the Eastern Mediterranean linked with the North Atlantic Heinrich events. *Geology* **31**, 439-442.
- Begin, Z. B., Ehrlich, A., and Nathan, Y. (1974). Lake Lisan, the Pleistocene precursor of the Dead Sea. *Bulletin of the Geological Survey of Israel* **63**, 30 pp.

- Berger, A. (1978). Long-term variations of caloric insolation resulting from the Earth's orbital elements. *Quaternary Research* **9**, 139-167.
- Bonacci, O. (2001). Monthly and annual effective infiltration coefficients in Dinaric karst: example of the Gradole karst spring catchment. *Hydrological Sciences-Journal-des Sciences Hydrologiques* **46**.
- Bookman, R., Bartov, Y., Enzel, Y., and Stein, M. (2006). The levels of the Quaternary lakes in the Dead Sea basin: two centuries of research. In: Enzel, Y., Agnon, A., Stein, M. (Eds.), *New frontiers in Dead Sea paleoenvironmental research*, Geological Society of America Special Paper **401**, 155-170.
- Brook, G. A., Embabi, N. S., Ashour, M. M., Edwards, R. L., Cheng, H., Cowart, J. B., and Dabous, A. A. (2002). Djara Cave in the Western Desert of Egypt: morphology and evidence of Quaternary climatic change. *Cave and Karst Science: The transactions of the British Cave Research Association* **29**, 57-66.
- Buchmann, D., and Dreybrodt, W. (1985). The kinetics of calcite dissolution and precipitation in geologically relevant situations of karst areas (2. Closed system). *Chemical Geology* **53**, 109-124.
- Burns, S. J., Fleitmann, D., Matter, A., Kramers, J., and Al-Subbary, A. (2003). Indian ocean climate and an absolute chronology over Dansgaard/Oeschger events 9 and 13. *Science* **301**, 1365-1367.
- Calvo, R. (2002). "Stratigraphy and petrology of the Hazeva Formation in the Arava and the Negev: Implications for the developments of the sedimentary basins and morphotectonics of the Dead Sea Rift Valley." Ph.D. thesis (in Hebrew, English Abstract) Hebrew University.
- Causse, C., Ghaleb, B., Chkir, N., Zouari, K., Ben Oueddou, H., and Mamou, A. (2003). Humidity changes in southern Tunisia during the Late Pleistocene inferred from U-Th dating of mollusc shells. *Applied Geochemistry* **18**, 1691-1703.
- Cerling, T. E., Quade, J. (1993). Stable carbon and oxygen isotopes in soil carbonates. *Climate Change in continental Isotopic Records, Geophysical Monograph* **78**, 217-231.
- Craig, H. (1961). Isotopic variations in Meteoric Waters. *Science* **133**, 1702-1703.
- Crombie, M. K., Arvidson, R. E., Sturchio, N. C., El Alfy, Z., and Zeid, K. A. (1997). Age and isotopic constraints on Pleistocene pluvial episodes in the Western Desert, Egypt. *Palaeogeography, Palaeoclimatology, Palaeoecology* **130**, 337-355.
- Crouvi, O., Amit, R., Enzel, Y., Porat, N., and McDonald, E. V. (2007). Regional pedosedimentary stratigraphic units in the primary eolian loess of the Negev Desert, Israel. In: Crouvi, O., Malik, U., Oren, O., Shalev, E., Friedman, V., and Yechieli, Y., eds., *Israel Geological Society, Annual Meeting: Neve Zohar, The Dead Sea*, p.23.
- Dan, Y., Yaalon, D., Koyumdjisky, H., and Raz, Z. (1975). The Soils of Israel (in Hebrew, with 1:500,000 soil map of Israel). In "Bulletin 168 of The Volcanic Institute of Agricultural Research, ." pp. 28, Beit Dagan, Israel.
- Dansgaard, W. (1964). Stable isotopes in precipitation. *Tellus* **16**, 438-468.
- Dayan, U. (1986). Climatology of Back Trajectories from Israel Based on Synoptic Analysis. *Journal of climate and applied meteorology* **25**, 591-595.
- deMenocal, P. (1995). Plio-Pleistocene African Climate. *Science* **270**, 53-59.
- deMenocal, P. B. (2004). African climate change and faunal evolution during the Pliocene-Pleistocene. *Earth and Planetary Science Letters* **220**, 3.

- Dennis, P. F., Rowe, P. J., and Atkinson, T. C. (2001). The recovery and isotopic measurement of water from fluid inclusions in speleothems. *Geochimica et Cosmochimica Acta* **65**, 871-884.
- Derricourt, R. (2005). Getting "Out of Africa": sea crossings, land crossings. *Journal of World Prehistory* **19**, 119-132.
- Drake, N., and Bristow, C. (2006). Shorelines in Sahara: geomorphological evidence for an enhanced monsoon from paleolake Megachad. *The Holocene* **16**, 901-911.
- Dreybrodt, W., Lauckner, J., Liu, Z., Svensson, U., and Buchmann, D. (1996). The kinetics of the reaction $\text{CO}_2 + \text{H}_2\text{O} \gg \text{H}^+ + \text{HCO}_3^-$ as one of the rate limiting steps of the dissolution of calcite in the system $\text{H}_2\text{O} - \text{CO}_2 - \text{CaCO}_3$. *Geochimica et Cosmochimica Acta* **60**, 3375-3381.
- Dreybrodt, W., Gabrovšek, F., and Perne, M. (2005). Condensation corrosion: a theoretical approach. *Acta Carsologica* **34**, 317-348.
- Ehrlich, S., Gavrieli, I., Ben-Dor, L., and Halicz, L. (2001). Direct high-precision measurements of the $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratio in natural water, carbonates and related materials by multiple collector inductively coupled plasma mass spectrometry (MC-ICP-MS). *Journal of Analytic Atomic Spectrometry* **16**, 1389-1392.
- Emeis, K. C., Schulz, H. M., Struck, U., Sakamoto, T., Dooze, H., Erlenkeuser, H., Howell, M., Kroon, D., and Paterne, M. (1998). Stable isotope and alkenone temperature records of sapropels from sites 964 and 967: constraining the physical environment of sapropel formation in the eastern Mediterranean Sea. *Proceedings of the Ocean Drilling Program, Scientific Results* **160**, 309-331.
- Emeis, K.-C., Struck, U., Schulz, H.-M., Rosenberg, R., Bernasconi, S., Erlenkeuser, H., Sakamoto, T., and Martinez-Ruiz, F. (2000). Temperature and salinity variations of Mediterranean Sea surface waters over the last 16,000 years from records of planktonic stable oxygen isotopes and alkenone unsaturation ratios. *Palaeogeography, Palaeoclimatology, Palaeoecology* **158**, 259.
- Emeis, K. C., Schulz, H. M., Struck, U., Rossignol-Strick, M., Sakamoto, T., Erlenkeuser, H., Howell, M., Kroon, D., Mackensen, A., Ishizuka, S., Oba, T., Sakamoto, T., and Koizumi, I. (2003). Eastern Mediterranean surface water temperatures and $\delta^{18}\text{O}$ composition during deposition of sapropels in the late Quaternary. *Palaeogeography* **18** (1), doi:10.1029/2000PA000617.
- Enmar, L. (1999). "The travertines in the northern and central Arava, petrography and geochemistry." M.Sc. (in hebrew, english abstract), Hebrew University.
- Enzel, Y., Amit, R., Dayan, U., Crouvi, O., Kahana, R., Ziv, B., and Sharon, D. (2007). The climatic and physiographic controls of the eastern Mediterranean over the late Pleistocene climates in the southern Levant and its neighboring deserts. *Global and Planetary Change* **in press**.
- Fairchild, I. J., Baker, A., Borsato, A., Frisia, S., Hinton, R., W., McDermott, F., and Tooth, A. (2001). Annual to sub-annual resolution of multiple trace-element trends in speleothems. *Journal of the Geological Society* **158**, 831-841.
- Feinbrun-Dothan, N., Danin, A. (1998). "Analytical flora of Eretz-Israel." CANA Publishing House Ltd., Jerusalem, Israel.
- Fernandes, C. A., J., R. E., and Siddall, M. (2006). Absence of post-Miocene Red Sea land bridges: biogeographic implications. *Journal of Biogeography* **33**, 961-966.
- Fleitmann, D., Burns, S. J., Neff, U., Mangini, A., and Matter, A. (2003). Changing moisture sources over the last 330,000 years in Northern Oman from fluid-inclusion evidence in speleothems. *Quaternary Research* **60**, 223-232.

- Fontugne, M., and Calvert, S. E. (1992). Late Pleistocene variability of the carbon isotopic composition of organic matter in the eastern Mediterranean: monitor of changes in carbon sources and atmospheric CO₂ concentrations. *Palaeoceanography* **7**, 1-20.
- Forbes, J. (1998). Air temperature and relative humidity study: Torgac Cave, New Mexico. *Cave and Karst Studies* **60**, 27-32.
- Ford, D., and Williams, P. W. (1989). "Karst Geomorphology and Hydrology." Unwin Hyman, London.
- French, H., and Binley, A. (2004). Snowmelt infiltration: monitoring temporal and spatial variability using time-lapse electrical resistivity. *Journal of Hydrology* **297**, 174-186.
- Frisia, S., Borsato, A., Fairchild, I. J., and McDermott, F. (2000). Calcite fabric, growth mechanisms, and environments of formation in speleothems from Italian Alps and southwestern Ireland. *Journal of Sedimentary Research* **70**, 1183-1196.
- Frisia, S., Borsato, A., Fairchild, I. J., McDermott, F., and Selmo, E. M. (2002). Aragonite-calcite relations in speleothems (Grotte de Clamouse, France): Environment, fabrics and carbonate geochemistry. *Journal of Sedimentary Research* **72**, 687-699.
- Frumkin, A., Ford, D. C., and Schwarcz, H. P. (1999). Continental Oxygen Isotopic Record of the Last 170,000 Years in Jerusalem. *Quaternary Research* **51**, 317-327.
- Frumkin, A., Ford, D., and Schwarcz, H. (2000). Paleoclimate and vegetation of the last glacial cycles in Jerusalem from a speleothem record. *Global Biogeochemical Cycles* **14**, 863-870.
- Frumkin, A., and Stein, M. (2004). The Sahara-East Mediterranean dust and climate connection revealed by strontium and uranium isotopes in a Jerusalem speleothem. *Earth and Planetary Science Letters* **217**, 451-464.
- Frumkin, A., and Fischhendler, I. (2005). Morphometry and distribution of isolated caves as a guide for phreatic and confined paleohydrological conditions. *Geomorphology* **67**, 457-471.
- Gallup, C. D., Cheng, H., Taylor, F. W., and Edwards, L. R. (2002). Direct determination of the timing of sea level change during Termination II. *Science* **295**, 310-313.
- Ganor, E., and Foner, H. A. (1996). The mineralogical and chemical properties and the behavior of the eolian Saharan dust over Israel. In: *Guerzoni, S. and Chester, R. (Eds.) The impact of desert dust across the Mediterranean* pp. 163-172.
- Ganor, E., and Foner, H. A. (2001). Mineral dust concentrations, deposition fluxes and deposition velocities in dust episodes over Israel. *Journal of Geophysical Research* **106**, 18431-18437.
- Garfunkel, Z. (1970). The tectonics of the western margin of the southern Arava. *Ph.D. Thesis (in Hebrew, English Abstract)*, 204 pp.
- Garfunkel, Z. (1978). The Negev - regional synthesis of sedimentary basin. *International Association of Sedimentology (IAS). Tenth International Congress of Sedimentology. Part 1: Precongress, Israel.*, pp. 35-110.
- Garfunkel, Z. (1982). The structures of the Syrian Arc in Israel. *Israel Geological Society Annual Meeting*, Elat, p. 23.
- Garfunkel, Z. (1988-a). The history of the Dead Sea Basin. *Israel Geological Society Annual Meeting*, En-Boqeq, p. 32.
- Garfunkel, Z. (1988-b). Relations between continental rifting and uplifting: evidence from the Suez rift and northern Red Sea. *Tectonophysics* **150**, 33-49.

- Garfunkel, Z., and Horowitz, A. (1966). The Upper Tertiary and Quaternary morphology of the Negev. *Israel Journal of Earth Sciences* **15**, 101-117.
- Gascoyne, M. (1992). Paleoclimate determination from the cave calcite deposits. *Quaternary Science Reviews* **11**, 609-632.
- Gat, J. R. (1984). The stable isotope composition of the Dead Sea waters. *Earth and Planetary Science Letters* **71**, 361-376.
- Gat, J. R., and Carmi, I. (1987). Effect of climate changes on the precipitation patterns and isotopic composition of water in a climatic transition zone: case of the Eastern Mediterranean Sea area. In "The influence of Climatic Change and Climatic Variability on the Hydrologic Regime and Water Resources." pp. 513-523. IANS 168 Symp. Proc.
- Genty, D., Plagnes, V., Causse, C., Cattani, O., Stievenard, M., Falourd, S., Blamart, D., Ouahdi, R., and Van-Exter, S. (2002). Fossil water in large stalagmite voids as a tool for paleoprecipitation stable isotope composition reconstruction and paleotemperature calculation. *Chemical Geology* **184**, 83-95.
- Ginat, H. (1997). The geology and landscape evolution in Hayyon and Zihor valleys. *Ph.D. Thesis (in Hebrew, English Abstract)*, 115 pp.
- Ginat, H., Zilberman, E., and Saragusti, I. (2003). Early Pleistocene lake deposits and Lower Paleolithic finds in Nahal (wadi) Zihor, Southern Negev desert, Israel. *Quaternary Research* **59**, 445-458.
- Goede, A., McCulloch, M., McDermott, F., and Hawkesworth, C. (1998). Aeolian contribution to strontium and strontium isotope variations in a Tasmanian speleothem. *Chemical Geology* **149**, 37-50.
- Goodfriend, G. A. (1990). Rainfall in the Negev Desert during the middle Holocene, based on (super 13) C of organic matter in land snail shells. *Quaternary Research* **34**, 186-197.
- Goodfriend, G. A. (1991). Holocene trends in ¹⁸O in land snail shells from the Negev Desert and their implications for changes in rainfall source areas. *Quaternary Research* **35**, 417-426.
- Goodfriend, G. A. (1999). Terrestrial stable isotope records of Late Quaternary paleoclimates in the eastern Mediterranean region. *Quaternary Science Reviews* **18**, 501-513.
- Goodfriend, G. A., and Magaritz, M. (1988). Palaeosols and late Pleistocene rainfall fluctuations in the Negev Desert. *Nature* **332**, 144-146.
- Greenbaum, N., Porat, N., Rhodes, E., and Enzel, Y. (2006). Large floods during late Oxygen Isotope Stage 3, southern Negev desert, Israel. *Quaternary Science Reviews* **25**, 704-719.
- Harmon, R. S., Ford, D. C., and Schwarcz, H. P. (1977). Interglacial chronology of the Rocky and Mackenzie Mountains based on ²³⁰Th/²³⁴U dating of calcite speleothems. *Canadian Journal of Earth Sciences* **14**, 1543-1552.
- Haviv, I., Stone, J., Enzel, Y., Zilberman, E., Whipple, K., Matmon, A., and Fifield, L. K. (2006). Climatic control on erosion rates of dolo-limestone hilltops. In: (add editors) *Israel Geological Society Annual Meeting, Bet-Shean*, add page.
- Henderson, G. M., and Slowey, N. C. (2000). Evidence from U-Th dating against Northern Hemisphere forcing of the penultimate deglaciation. *Nature*, 61-66.
- Hendy, C. H. (1971). The isotopic geochemistry of speleothems-I. The calculation of the effects of different modes of formation on the isotopic composition of speleothems and their applicability as palaeoclimatic indicators. *Geochimica et Cosmochimica Acta* **35**, 801-824.

- Hendy, C. H., and Wilson, A. T. (1968). Palaeoclimatic data from speleothems. *Nature* **219**, 48-51.
- Herut, B., Starinsky, A., and Katz, A. (1993). Strontium in rainwater from Israel: Sources, isotopes and chemistry. *Earth and Planetary Science Letters* **120**, 77-84.
- Holmgren, K., Karlen, W., and Shaw, P. A. (1995). Paleoclimatic Significance of the Stable Isotopic Composition and Petrology of a Late Pleistocene Stalagmite from Botswana. *Quaternary Research* **43**, 320-328.
- Hurrell, J. W. (1995). Decadal trends in the north atlantic oscillation: regional temperatures and precipitation. *Science* **269**, 676-679.
- Kahana, R., Ziv, B., Enzel, Y., and Dayan, U. (2002). Synoptic climatology of major floods in the Negev Desert, Israel. *International Journal of Climatology* **22**, 867-882.
- Kallel, N., Paterne, M., Duplessy, J. C., Vergnaud-Grazzini, C., Pujol, C., Labeyrie, L., Arnold, M., Fontugne, M., and Pierre, C. (1997). Enhanced rainfall on Mediterranean region during the last sapropel event. *Oceanologica Acta* **20**, 697-712.
- Kallel, N., Duplessy, J. C., Labeyrie, L., Fontugne, M., Paterne, M., and Montacer, M. (2000). Mediterranean pluvial periods and sapropel formation over the last 200 000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* **157**, 45-58.
- Kaufman, A., Wasserburg, G. J., Porcelli, D., Bar-Matthews, M., Ayalon, A., and Halicz, L. (1998). U-Th isotope systematics from the Soreq cave, Israel and climatic correlations. *Earth and Planetary Science Letters* **156**, 141-155.
- Koepnick, R. B., Burke, W. H., Denison, R. E., Hetherington, E. A., Nelson, H. F., Otto, J. B., and Waite, L. E. (1985). Construction of the seawater $^{87}\text{Sr}/^{86}\text{Sr}$ curve for the Cenozoic and Cretaceous supporting data. *Chemical Geology (Isotope Geoscience Section)* **58**, 55-81.
- Kolodny, Y., Stein, M., and Machlus, M. (2005). Sea-rain-lake relation in the Last Glacial East Mediterranean revealed by $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ in Lake Lisan aragonites. *Geochimica et Cosmochimica Acta* **69**, 4045-4060.
- Kolosovsky, A. (2003). "Reconstruction of paleotemperatures and determination of origin of organic matter in sediments from the eastern Mediterranean Sea (MSc Thesis, in Hebrew, English abstract)." Hebrew University, Jerusalem.
- Krichak, S. O., and Alpert, P. (2005). Signatures of the NAO in the atmospheric circulation during wet winter months over the Mediterranean region. *Theoretical and Applied Climatology* **82**, 27-39.
- Krom, M. D., Cliff, R. A., Eijssink, L. M., Herut, B., and Chester, R. (1999). The characterisation of Saharan dusts and Nile particulate matter in surface sediments from the Levantine basin using Sr isotopes. *Marine Geology* **155**, 319-330.
- Kronfeld, J., and Livnat, A. (1987). Uranium-series dating of Rift Valley travertines and lacustrine deposits: paleoclimatic implications and stratigraphic application. Israel Ministry of Energy and Infrastructure.
- Krumgalz, B. S., Hecht, A., Starinsky, A., and Katz, A. (2000). Thermodynamic constraints of the Dead Sea evaporation: Can the Dead Sea dry up? *Chemical Geology* **165**, 1-11.
- Kuperman, G. (2005). "Reconstruction of flood periods by dating speleothems in Nahal Hazera, northern Negev." Hebrew University of Jerusalem, M.Sc. Thesis (in Hebrew, English Abstract).
- Lauritzen, S. E. (1995). High resolution paleotemperature proxy record for the last interglaciation based on Norwegian speleothems. *Quaternary Research* **43**, 133-146.

- Li, H.-C., Ku, T.-L., You, C.-F., Cheng, H., Edwards, R. L., Ma, Z.-B., Tsai, W.-s., and Li, M.-D. (2005). $^{87}\text{Sr}/^{86}\text{Sr}$ and Sr/Ca in speleothems for paleoclimate reconstruction in Central China between 70 and 280 kyr ago. *Geochimica et Cosmochimica Acta* **69**, 3933-3947.
- Livnat, A., and Kronfeld, J. (1985). Paleoclimatic implications of U-series dates for lake sediments and travertines in the Arava Rift Valley, Israel. *Quaternary Research* **24**, 164-172.
- Livnat, A., and Kronfeld, J. (1990). Pleistocene lakes in the Arava Rift of Israel (Sayif Formation): facies and paleoenvironmental setting. *Journal of African Earth Sciences* **10**, 409.
- Ludwig, K. R. (2003). Isoplot 3.00, Special Publication No. 4. In "A Geochronological Toolkit for Microsoft Excel." (K. R. Ludwig, Ed.). Berkeley Geochronology Center, 2455 Ridge Road, Berkeley CA 94709, USA.
- Magaritz, M., and Goodfriend, G. A. (1987). "Abrupt Climatic Change." Reidel D., Publishing Company.
- Martinson, D. G., Pisias, N., Hays, J. D., Imbrie, J., Moore, T. C. J., and Shackleton, N. J. (1987). Age dating and the orbital theory of the ice ages: development of the high-resolution 0-300,000 year chronostratigraphy. *Quaternary Research* **27**, 1-29.
- Martrat, B., Grimalt, J. O., Lopez-Martinez, C., Cacho, I., Sierro, F. J., Flores, J. A., Zahn, R., Canals, M., Curtis, J. H., and A., H. D. (2004). Abrupt temperature changes in the Western Mediterranean over the past 250,000 years. *Science* **306**, 1762-1765.
- Matthews, A., Ayalon, A., and Bar-Matthews, M. (2000). D/H ratios of fluid inclusions of Soreq cave (Israel) speleothems as a guide to the Eastern Mediterranean Meteoric Line relationships in the last 120 ky. *Chemical Geology* **166**, 183-191.
- McDermott, F. (2004). Palaeo-climate reconstruction from stable isotope variations in speleothems: a review. *Quaternary Science Reviews* **23**, 901-918.
- McDermott, F., Frisia, S., Huang, Y., Longinelli, A., Spiro, B., Heaton, T. H. E., Hawkesworth, C., Borsato, A., Keppens, E., Fairchild, I. J., Borg, K., Verheyden, S., and Selmo, E. M. (1999). Holocene climate variability in Europe: Evidence from $\delta^{18}\text{O}$, textural and extension rate variations in three speleothems. *Quaternary Science Reviews* **18**, 1021-1038.
- McDermott, F., Mathey, D. P.; Hawkesworth, C. (2001). Centennial-scale Holocene climate variability revealed by a high-resolution speleothem delta (super 18) O record from SW Ireland. *Science* **294**, 1328-1331.
- McGarry, S., Bar-Matthews, M., Matthews, A., Vaks, A., Schilman, B., and Ayalon, A. (2004). Constraints on hydrological and paleotemperature variations in the Eastern Mediterranean region in the last 140 ka given by the [delta]D values of speleothem fluid inclusions. *Quaternary Science Reviews* **23**, 919-934.
- Miller, G. H., Wendorf, F., Ernst, R., Schild, R., Close, A. E., Friedman, I., and Schwarcz, H. P. (1991). Dating lacustrine episodes in the eastern Sahara by the epimerization of isoleucine in ostrich eggshells. *Palaeogeography, Palaeoclimatology, Palaeoecology* **84**, 175-189.
- Neev, D., and Emery, K. O. (1967). The Dead Sea. *Bulletin of the Geological Survey of Israel* **41**, 147pp.
- Njitchoua, R., Sigha-Nkamdjou, L., Dever, L., Marlin, C., Sighomnou, D., and Nia, P. (1999). Variations of the stable isotopic compositions of rainfall events from the Cameroon rain forest, Central Africa. *Journal of Hydrology* **223**, 17.

- O'Neil, J. R., Clayton, R. N., and Mayeda, T. K. (1969). Oxygen isotope fractionation of divalent metal carbonates. *Journal of Chemical Physics* **30**, 5547-5558.
- Olausson, E. (1961). Studies of deep sea cores. Sediment cores from the Mediterranean Sea and the Red Sea. *Rep. Sweden Deep Sea Expedition 1947-48 VIII*, 337-391.
- Osmond, J. K., and Dabous, A. A. (2004). Timing and intensity of groundwater movement during Egyptian Sahara pluvial periods by U-series analysis of secondary U in ores and carbonates. *Quaternary Research* **61**, 85-94.
- Petit-Maire, N., Sanlaville, P., Abed, A., Yasin, S., Bourrouilh, R., Carbonel, P., Fontugne, M., and Reyss, J. L. (2002). New data for an Eemian lacustrine phase in southern Jordan. *Episodes* **25**, 279-280.
- Picard, L. (1943). Structure and evolution of Palestine (with comparative notes on neighbouring countries). *Hebrew University, Jerusalem, Department Geological Bulletin* **4**, 187 pp.
- Plakht, J. (1995). Climatic conditions during the last stages of development of Makhtesh Ramon. *Israel Journal of Earth Sciences* **44**, 149-157.
- Plakht, J. (2000). Quaternary units in makhteshim, Negev Desert: Similarities and peculiarities. *Israel Journal of Earth Sciences* **49**, 179-187.
- Polyak, V., and Asmerom, Y. (2005). Orbital control of long term moisture in southwestern USA. *Geophysical Research Letters* **32**, 1-4.
- Poulson, T. L., and White, W. B. (1969). The cave environment. *Science* **165**, 971-981.
- Rabikovit, S. (1992). The Soils of Israel: Formation, Nature, and Properties (In Hebrew, with 1:250,000 soil map of Israel in English and Hebrew). , pp. 489. Hakibbutz Hamehuchad Publishing House, Israel. .
- Rink, W. J., Richter, D., Schwarcz, H. P., Marks, A. E., Monigal, K., and Kaufman, D. (2003). Age of the Middle Palaeolithic site of Rosh Mor, central Negev, Israel: Implications for the age range of the Early Levantine Mousterian of the Levantine corridor. *Journal of Archaeological Science* **30**, 195-204.
- Robinson, L. F., Henderson, G. M., and Slowey, N. C. (2002). U-Th dating of marine isotope stage 7 in Bahamas slope sediments. *Earth and Planetary Science Letters* **196**, 175-187.
- Roche, D. M., Dokken, T. M., Goosse, H., Renssen, H., and Weber, S. L. (2007). Climate of the Last Glacial Maximum: sensitivity studies and model-data comparison with the LOVECLIM coupled model. *Climate of the past* **3**, 205-224.
- Rohling, E. J., Cane, T. R., Cooke, S., Sprovieri, M., Bouloubassi, I., Emeis, K. C., Schiebel, R., Kroon, D., Jorissen, F. J., Lorre, A., and Kemp, A. E. S. (2002). African monsoon variability during the previous interglacial maximum. *Earth and Planetary Science Letters* **202**, 61-75.
- Rosignol-Strick, M. (1983). African monsoons, an immediate climate response to orbital insolation. *Nature* **304**, 46-49.
- Rosignol-Strick, M., and Paterne, M. (1999). A synthetic pollen record of the eastern Mediterranean sapropels of the last 1 Ma: implications for the time-scale and formation of sapropels. *Marine Geology* **153**, 221-237.
- Ryan, W. B. F. (1972). Stratigraphy of the late Quaternary sediments in the Eastern Mediterranean. In "The Mediterranean Sea: A Natural Sedimentation Laboratory " (D. J. Stanley, Ed.), pp. 149-169. Dowden, Hutchinson and Ross, Stroudsburg, Pennsylvania, US.
- Said, R. (1993). "The river Nile: Geology, Hydrology and Utilization." Pergamon Press Ltd, Oxford.

- Schwarcz, H. P. (1986). Geochronology and isotopic geochemistry of speleothems. In "Handbook of Environmental Isotope Geochemistry." (P. Fritz, and J. C. Fontes, Eds.), pp. 271-300. Elsevier, Amsterdam.
- Schwarcz, H. P., Blackwell, B., Goldberg, P., and Marks, A. E. (1979). Uranium series dating of travertine from archaeological sites, Nahal Zin, Israel. *Nature* **277**, 558-560.
- Shackleton, N. J., and Opdyke, N. D. (1973). Oxygen isotopes and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale. *Quaternary Research* **3**, 39-55.
- Shay-El, Y., and Alpert, P. (1991). A diagnostic study of winter diabatic heating in the Mediterranean in relation with cyclones *Quarterly Journal of Royal Meteorological Society* **117**, 715-747.
- Shuman, B., and Donnelly, J. P. (2006). The influence of seasonal precipitation and temperature regimes on lake levels in the northeastern United States during the Holocene. *Quaternary Research* **65**, 44-56.
- Solomon, A., and Garfunkel, Z. (1989). The evolution of the northern Negev monoclines. *Israel Geological Society Annual Meeting*, Ramot, p.131.
- Soudry, D., Ehrlich, S., Yoffe, O., and Nathan, Y. (2002). Uranium oxidation state and related variations in geochemistry of phosphorites from the Negev (southern Israel). *Chemical Geology* **189**, 213-230.
- Soudry, D., Nathan, Y., and Roded, R. (1985). The Ashosh-Haroz facies and their significance for the Mishash paleogeography phosphorite accumulation in the northern and central Negev, southern Israel *Israel Journal of Earth Sciences* **34**, 211-220.
- Spotl, C., Mangini, A., Frank, N., Eichstadter, R., and Burns, S. J. (2002). Start of the last interglacial period at 135 ka: evidence from a high Alpine speleothem. *Geology* **30**, 815-818.
- Steinitz, G., and Yosef, B. (1991). The Miocene-Pleistocene history of the Dead Sea segment of the Rift in light of K-Ar ages of basalts. *Israel Journal of Earth Sciences* **40**, 199-208.
- Street, A. F., and Grove, A. T. (1979). Global Maps of Lake-Level Fluctuations since 30,000 yr B. P. *Quaternary Research* **12**, 83-118.
- Szabo, B. J., Haynes, J. C. V., and Maxwell, T. A. (1995). Ages of Quaternary pluvial episodes determined by uranium-series and radiocarbon dating of lacustrine deposits of Eastern Sahara. *Palaeogeography, Palaeoclimatology, Palaeoecology* **113**, 227-242.
- Tsoar, H., and Pye, K. (1987). Dust transport and the question of desert loess formation. *Sedimentology* **34**, 139-153.
- Usami, M., Suga, S. I., Lototskaya, A., Ziveri, P., Ganssen, G. M., and van Hinte, J. E. (1998). Calcareous nannofloral response to Termination II at 45°N, 25°W (northeast Atlantic). *Marine Micropaleontology* **34**, 47-70.
- Vaks, A. (2001). "Climatic changes during the last 160 kyr from the speleothem record of karstic cave located on desert boundary, Ma'ale-Efrayim, Eastern Shomeron, Israel." M.Sc. Thesis (in Hebrew, English Abstract), Hebrew University.
- Vaks, A., Bar-Matthews, M., Ayalon, A., Schilman, B., Gilmour, M., Hawkesworth, C. J., Frumkin, A., Kaufman, A., and Matthews, A. (2003). Paleoclimate reconstruction based on the timing of speleothem growth and oxygen and carbon isotope composition in a cave located in the rain shadow in Israel. *Quaternary Research* **59**, 182-193.

- Vaks, A., Bar-Matthews, M., Ayalon, A., Matthews, A., Frumkin, A., Dayan, U., Halicz, L., Almogi-Labin, A., and Schilman, B. (2006). Paleoclimate and location of the border between Mediterranean climate region and the Saharo-Arabian Desert as revealed by speleothems from the northern Negev Desert, Israel *Earth and Planetary Science Letters* **249**, 384-399.
- Vaks, A., Bar-Matthews, M., Ayalon, A., Matthews, A., Halicz, L., and Frumkin, A. (2007). Desert speleothems reveal climatic window for African exodus of early modern humans. *Geology* **35**, 831-834.
- Vogel, J. C., Fuls, A., Danin, A. (1986). Geographical and environmental distribution of C3 and C4 grasses in Sinai, Negev, and Judean deserts. *Oecologia* **70**, 258-265.
- Waldman, N., Starinsky, A., and Stein, M. (2007). Primary carbonates and Ca-chloride brines as monitors of a paleo-hydrological regime in the Dead Sea basin. *Quaternary Science Reviews* **in-press**.
- Walker, J., Cliff, R. A., and Latham, A., G. (2007). U-Pb isotopic age of the StW 573 Hominid from Sterkfontein, South Africa *Science* **314**, 1592-1594.
- Woodhead, J., Hellstrom, J., Maas, R., Drysdale, R., Zanchetta, G., Devine, P., and Taylor, E. (2006). U-Pb geochronology of speleothems by MC-ICPMS. *Quaternary Geochronology* **1**, 208-221.
- Yaalon, D. H. (1971). Soil forming processes in time and space. In: Yaalon, D. H. (Ed.) *Paleopedology: Origin, nature and dating of paleosols. International Society of soil Science and Israel Universities Press*, pp. 29-39.
- Yaalon, D. H., and Dan, J. (1974). Accumulation and distribution of loess-derived deposits in the semi-desert fringe areas of Israel. *Zeitschrift fur Geomorphologie Supplementband* **20**, 91-105.
- Yaalon, D. H., and Ganor, E. (1973). The influence of dust on soils during the Quaternary. *Soil Science* **116**, 146-155.
- Yair, A. (1990). The role of topography and surface cover upon soil formation along hillslope in arid climate. *Geomorphology* **3**, 287-299.
- Yonge, C. J. (1982). Stable isotope studies from the water extracted from speleothems. *Ph.D. Thesis, McMaster University*, 298 pp.
- Zangvil, A., and Druian, P. (1990). Upper Air Trough Axis Orientation and the Spatial Distribution of Rainfall over Israel. *International Journal of Climatology IJCLEU* **10**, 57-62.
- Zilberman, E. (1991). "Landscape evolution in the central, northern, and northwestern Negev during the Neogene and the Quaternary (Ph.D Thesis, in Hebrew, English abstract)." Hebrew University, Jerusalem.

Appendix 1: Figures

Figure 1: Geographical location of the study area, annual precipitation and sample sites.

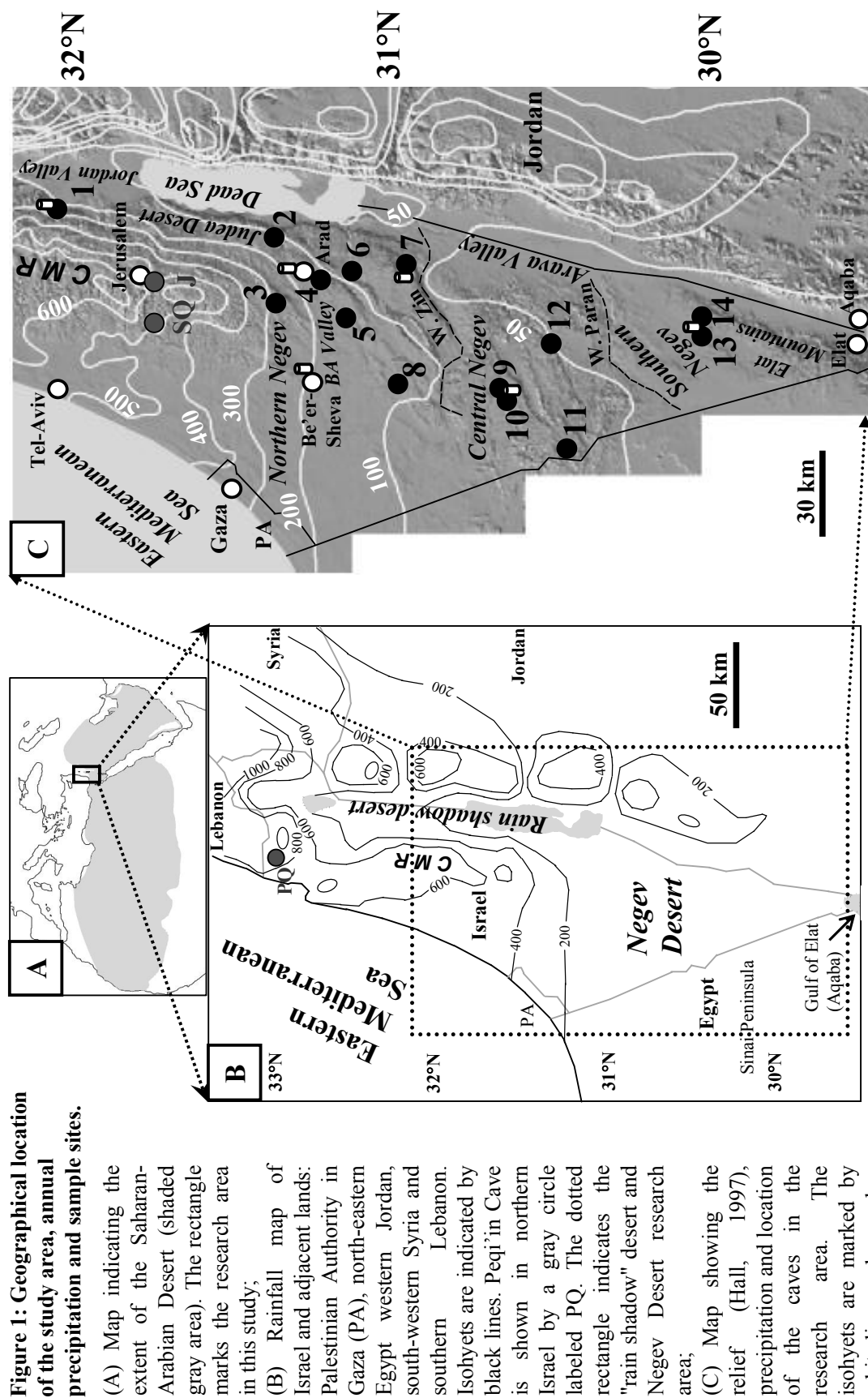


Figure 2: Geology of the research area. Most caves (#1-10, and #12-14, Fig. 1C) are located in Cenomanian and Turonian limestone and dolomite of Judea Group, but one cave (Wadi-Lotz Cave, # 11, Fig. 1C) is located in Eocene limestone of Avdat Group.

2A: Geological stratigraphic section of the research area.

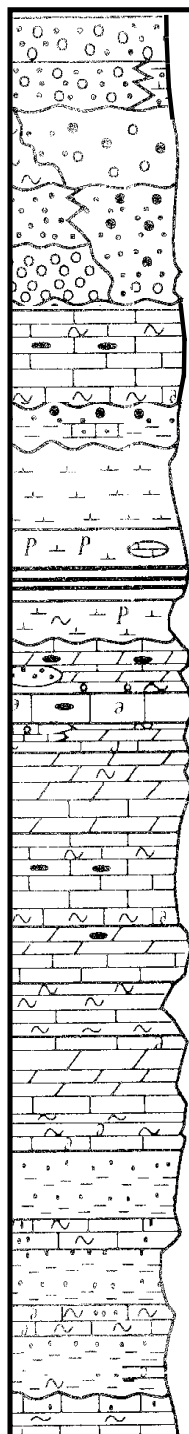
	Loess / Alluvial terraces	Pleistocene	Quaternary
	Zhiha Formation Dead Sea Group Arava Formation	Pliocene	
	Hazeva Group	Miocene	Tertiary
	Avdat Group	Eocene	
	Mount Scopus Group	Paleocene	
		Maastricht	Cretaceous
		Senonian	
	Judea Group	Turonian	
		Cenomanian	
		Albian	
	Curnub Group	Aptian	
	Arad Group	Callovian	Jurassic

Figure 2B: Geologic map of the research area. Caves marked by numbers and black circles as in Fig. 1C.

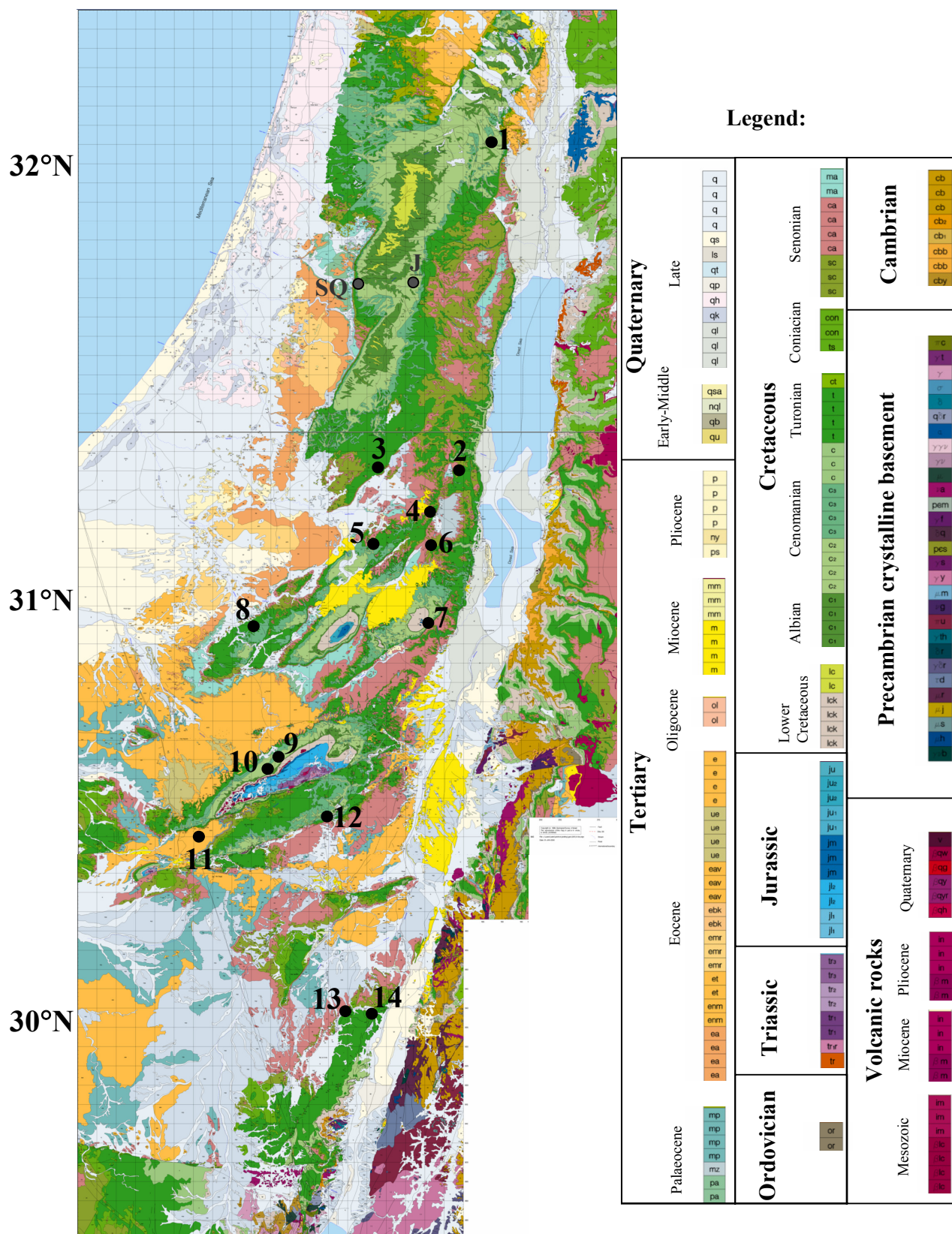


Figure 3: Typical cross sections of the Ashalim Cave speleothems as observed in the field, hand specimens and under SEM:

1 – Flowstone ASH-15 in Ashalim Cave (#8, Fig. 1): the Basal massive member is yellow (top centre – laminae D-I) and composed of large (> 2 cm) calcite crystals. The laminae C, D and E (marked by rectangles), are dated by U-Pb method to 1.269 ± 0.013 Ma, 3.045 ± 0.021 Ma and 3.048 ± 0.014 Ma respectively. The white and brownish thin lamina between the two stratigraphic members (enlarged and marked by arrow) represents a growth break (hiatus). (2) Enlargements of boundary between the Basal member (lamina D) and Intermediate member (laminae A-C): 2.1 – in plain light, 2.2 – under SEM. (3, 4) SEM analysis shows that it contains silica, alumina, microcrystalline calcite, Fe-oxides and halite.

5 – Stalactite ASH-11 in Ashalim Cave: the Basal member is yellow (top centre), the Intermediate member is brown, and the Young member composed of 4 thin laminae is shown enlarged in (6). Numerous fine white laminae (white arrows in top) representing hiatuses divide between the wider calcite laminae.

6 – Enlargement of the Young member in stalactite ASH-11 showing the 5 calcite laminae with overall thickness of less than 2 cm. Four white thin laminae dividing between the calcite laminae represent hiatuses (marked by black arrows in top). The U-Th ages of the laminae are marked by arrows in bottom.

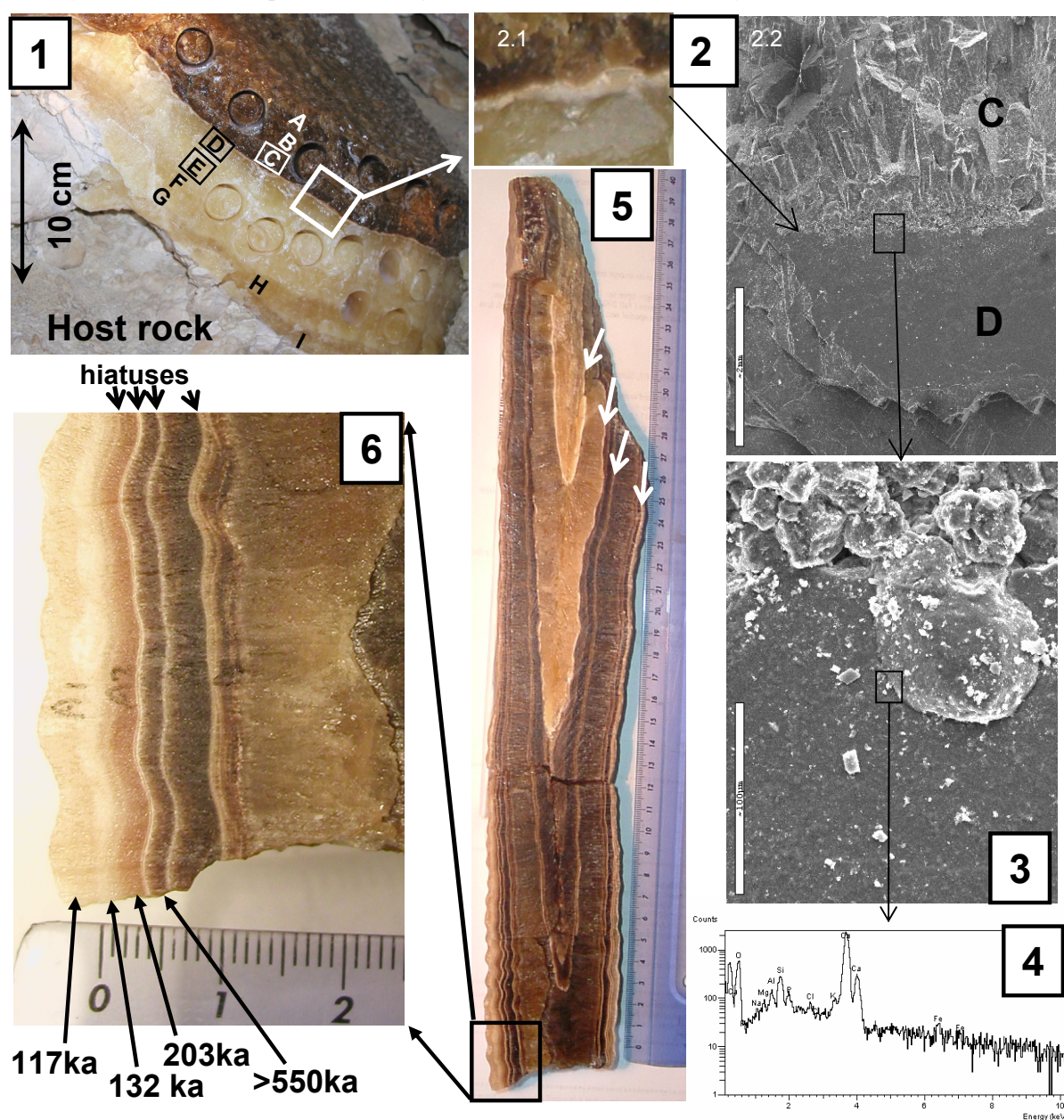


Figure 4: Cross section of the flowstone KTO(1)-1 from Ktora Cracks (#14, Fig. 1C)

A - The Basal member (laminae I, J, and K) makes up the most of the speleothem volume and contains only 2 hiatuses at the bottom. This is indicative of generally continuous growth from slowly dripping water. The oldest laminae J and K that are marked by rectangles were dated by U-Pb method to 2.71 ± 0.06 Ma and 3.32 ± 0.06 Ma, respectively.

B - The cross section of flowstone KTO(1)-1 from the Ktora Cracks (#14 in Fig. 1). Black arrows on the right mark the hiatuses in calcite deposition between the Basal and the Intermediate members, and inside the Intermediate member. The oldest major hiatus is lamina H, composed of two gypsum laminae (beige) and thin calcite lamina in between (dark). The two gypsum laminae represent periods of hyper-arid conditions that occurred between the deposition of the Basal and the Intermediate members. The laminated Intermediate member (laminae E, F and G) is of lesser width, and shows frequent alternations between calcite laminae and hiatuses (arrows) and therefore represents generally drier conditions than the Basal member.

The Young member is thinly laminated (laminae A1, A2, B, C and D at the top, <1 cm width). The laminae A1 and A2 are dateable by the U-Th method and their ages are 137 ka and 319 ka respectively, whereas laminae C and D are older than 550 ka. The gypsum lamina B in the Young member formed during the long hyper-arid period during Middle-Pleistocene.

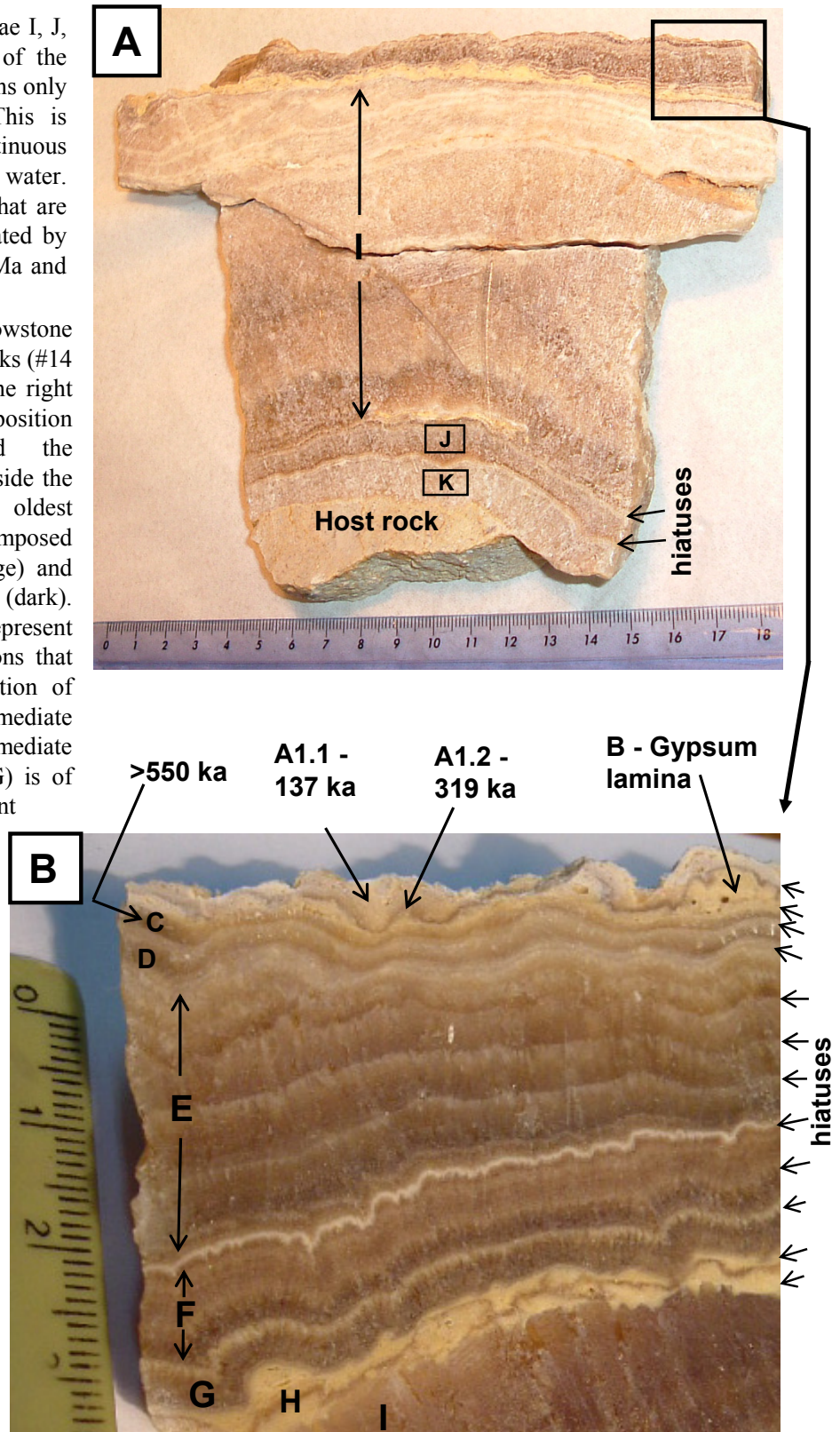


Figure 5: The mineralogy and petrography of the calcite and fine laminae representing hiatuses within the Young member of the stalactite ASH-11. A - Abrasion of the calcite crystals on the present day surface of the stalactite; B – etched calcite crystals of the ancient surface (hiatus plane) between laminae A2 (203 ka) and A3 (>550 ka); C – petrographic microscope picture (crossed polar light) showing the gypsum on the stalactite surface, calcite crystals of the lamina A1, detrital matter in hiatus between the laminae A1 and A2, and columnar fabric of calcite crystals in lamina A2; D – SEM picture of the present day surface of the stalactite with gypsum plates over etched calcite crystals; E – SEM picture of a halite sphere and the flakes of the clay minerals within the fine lamina representing the hiatus between the laminae A1 and A2; F - SEM picture of the gypsum plates over calcite crystals formed during the growth break between the laminae A2 and A3.

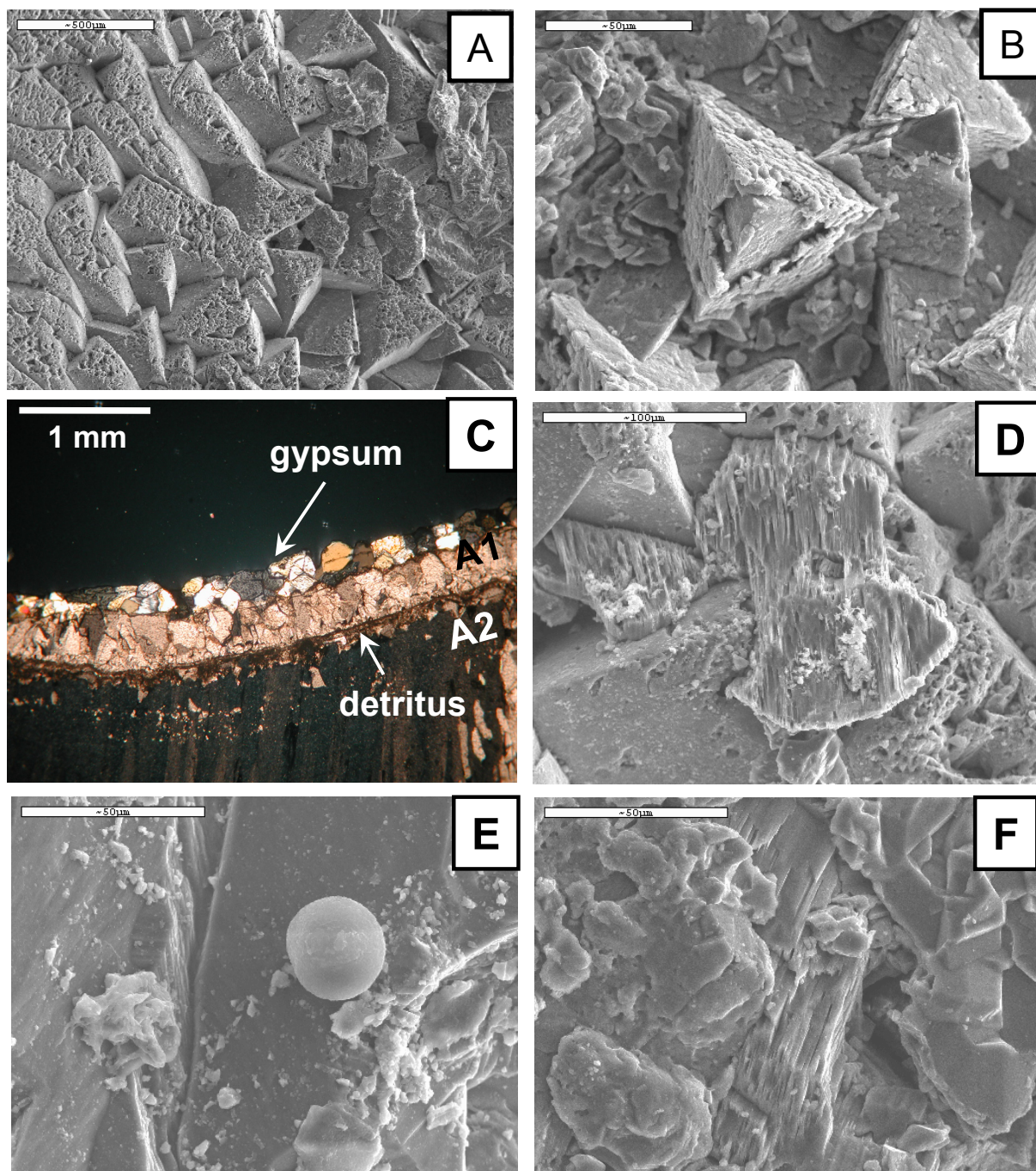
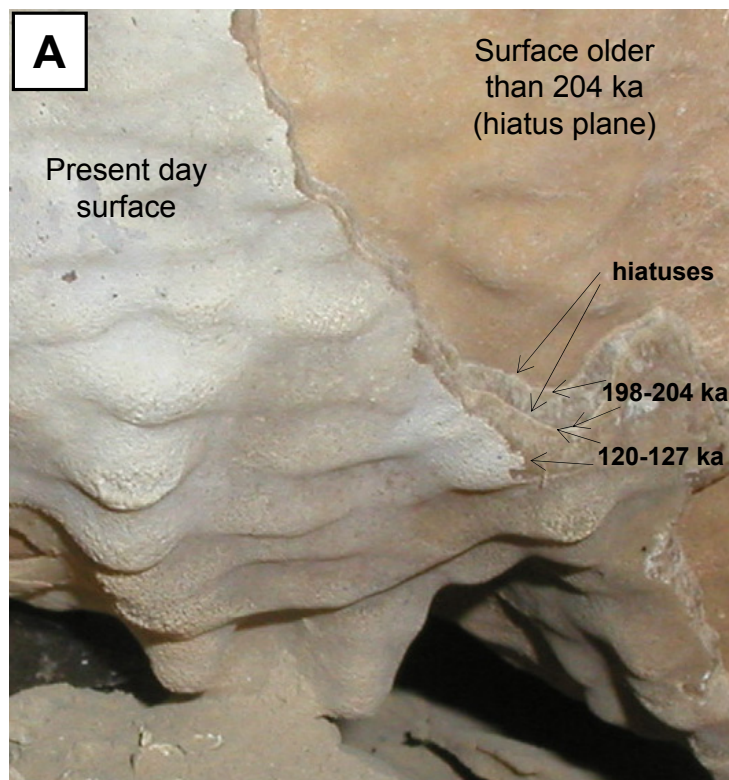


Figure 6: Flowstone MMR-7(2) from Ma’ale-ha-Meyshar Cave. A – Photograph of the present-day surface of the flowstone covered by a “patina” of grey-whitish microcrystalline calcite and detrital materials. The two youngest calcite laminae were dated at 204-198 ka and 127-120 ka. These laminae are separated one from other by fine laminae rich with detrital materials, microcrystalline calcite and evaporite minerals, representing hiatuses. It is easy to separate between the calcite laminae along these hiatus planes exposing the older speleothem surfaces.



B – SEM analysis of the material accumulated on present-day speleothem surface shows microcrystalline calcite, clay minerals (silica, alumina), gypsum, halite, quartz grains, phosphate and minerals of Fe, Ti, and Zn.

C – SEM analysis of the material accumulated on speleothem surface older than 204 ka (reddish colored hiatus plane in A) shows relatively similar mineralogy as in B (microcrystalline calcite, clay minerals (silica, alumina), quartz grains, and minerals of Fe, Mg and Zn), with larger amounts of Fe, Cr and Zn, smaller amounts of gypsum and halite, and an absence of Ti and phosphate.

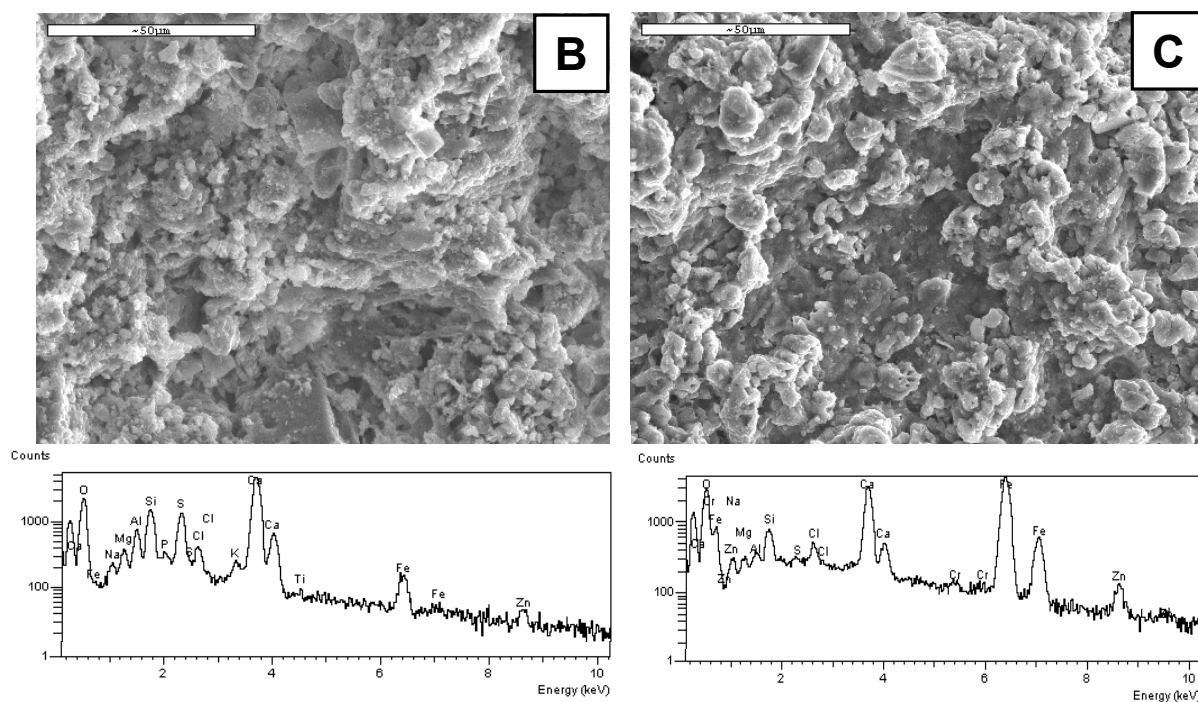


Figure 7: SEM analysis of the minerals. (A) Celestine (SrSO_4), (B) Minerals of Ti, Fe, Mn, Mg, with clay minerals from present day surface of flowstone MMR-8; (C) Barite (BaSO_4) and celestine from the hiatus between laminae E and F in flowstone ASH-33.

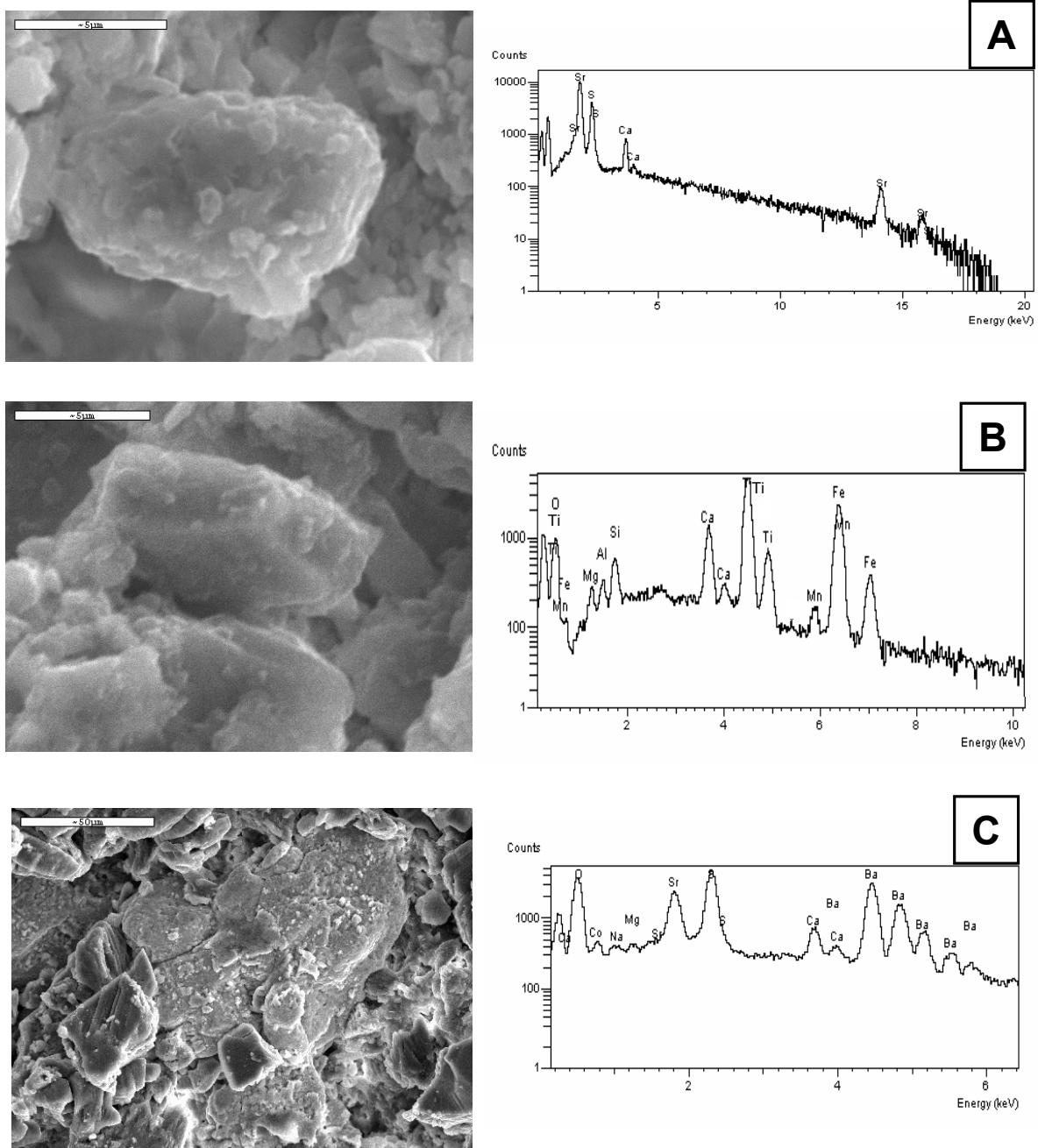
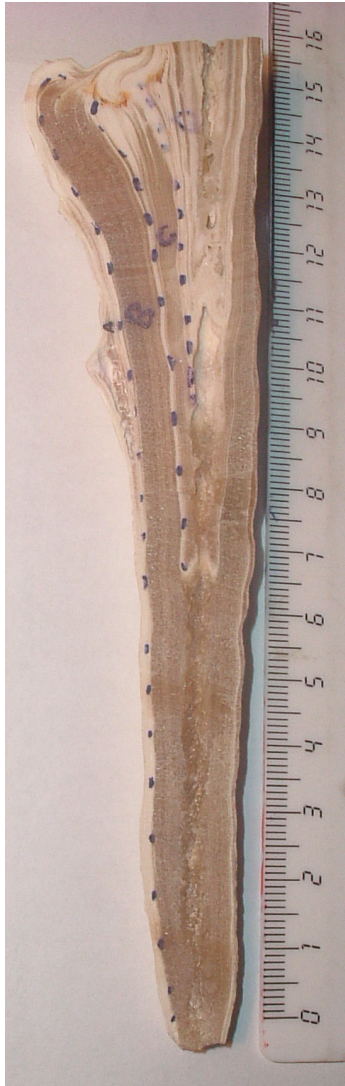


Figure 8: Petrography of stalactite KN-8 from the Kanaim Cave showing laminae of beige columnar calcite alternating with white laminae of microcrystalline calcite, probably formed by condensation corrosion and representing hiatuses in speleothem deposition. The corrosion of calcite probably induced radionuclide remobilization causing the age reversal between laminae A and B.

A



B

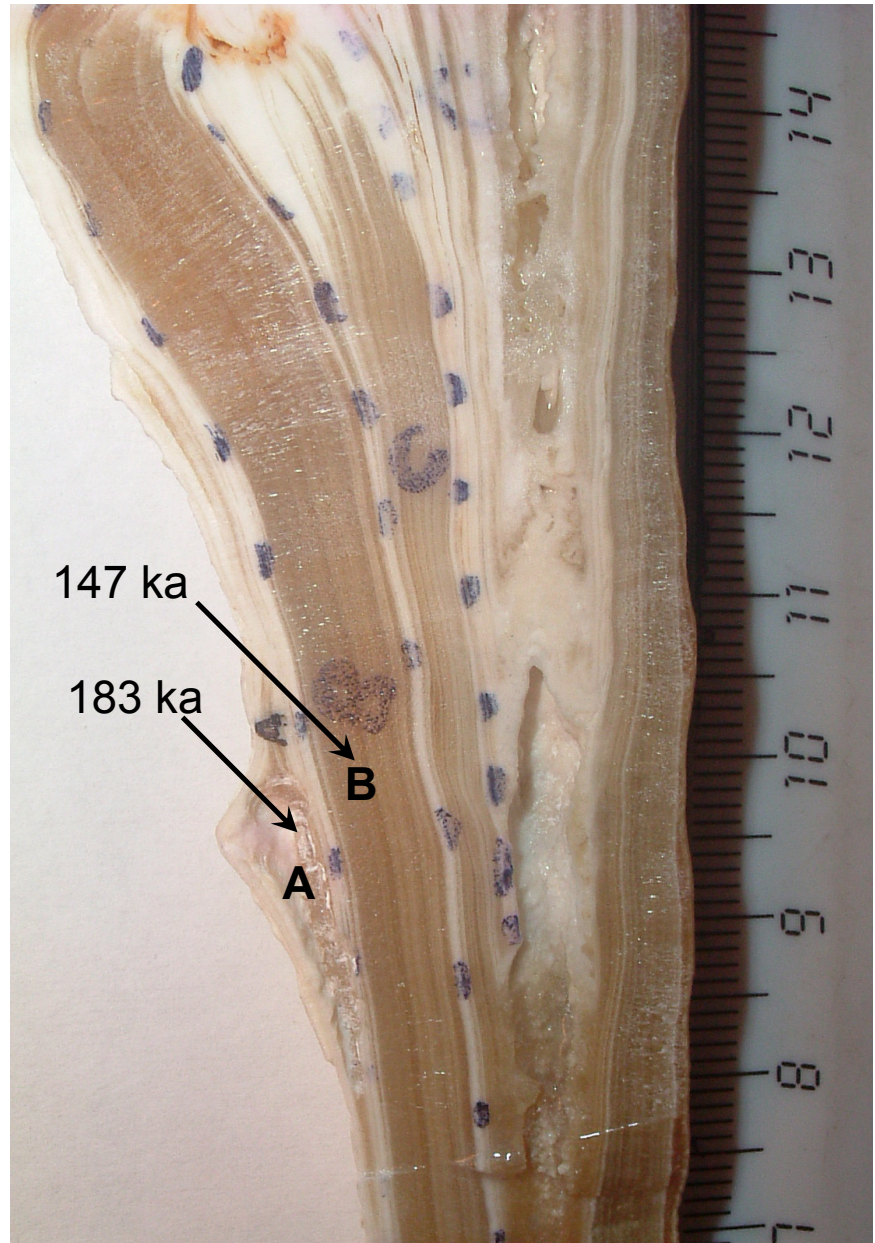
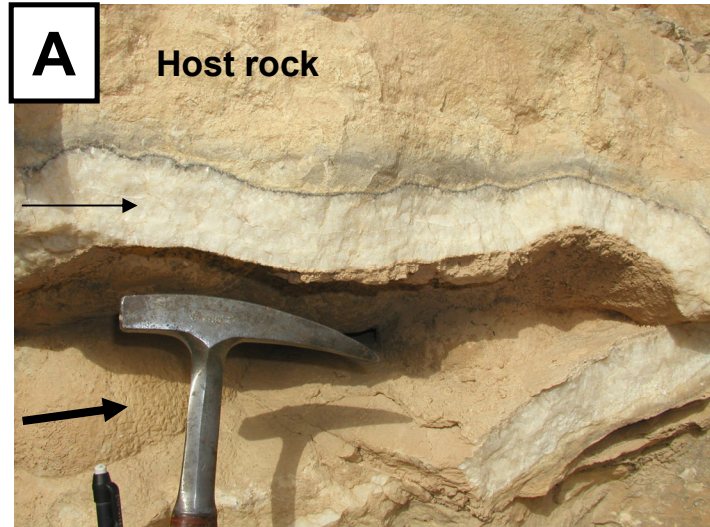
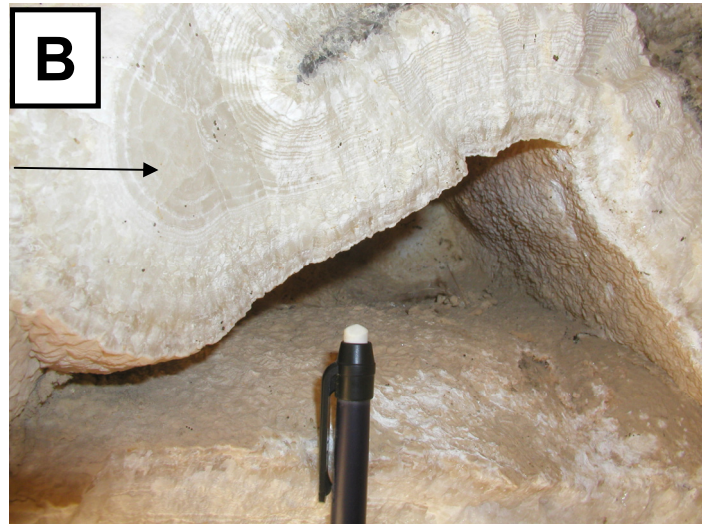


Figure 9: Phreatic speleothems in Ma’ale-ha-Meyshar Cave.

A : Phreatic calcite overgrowth layer (thin arrows) covering the cave walls and ceiling. The reddish paleosols (thick arrow) postdate the phreatic speleothems and possibly were washed into the cave after it was uplifted above the groundwater level.



B: Laminated calcite overgrowth on the cave walls



C: the cave “rafts” covering the floor of a small cave near Ma’ale-ha-Meyshar major cave.



Figure 10: The relative frequencies of the speleothem ages in “rain shadow” desert.

A - Kanaim Cave, Judea Desert; B – Ma’ale-Efrayim Cave, including the modified data of Vaks et al., (2003) and the new 7 ages.

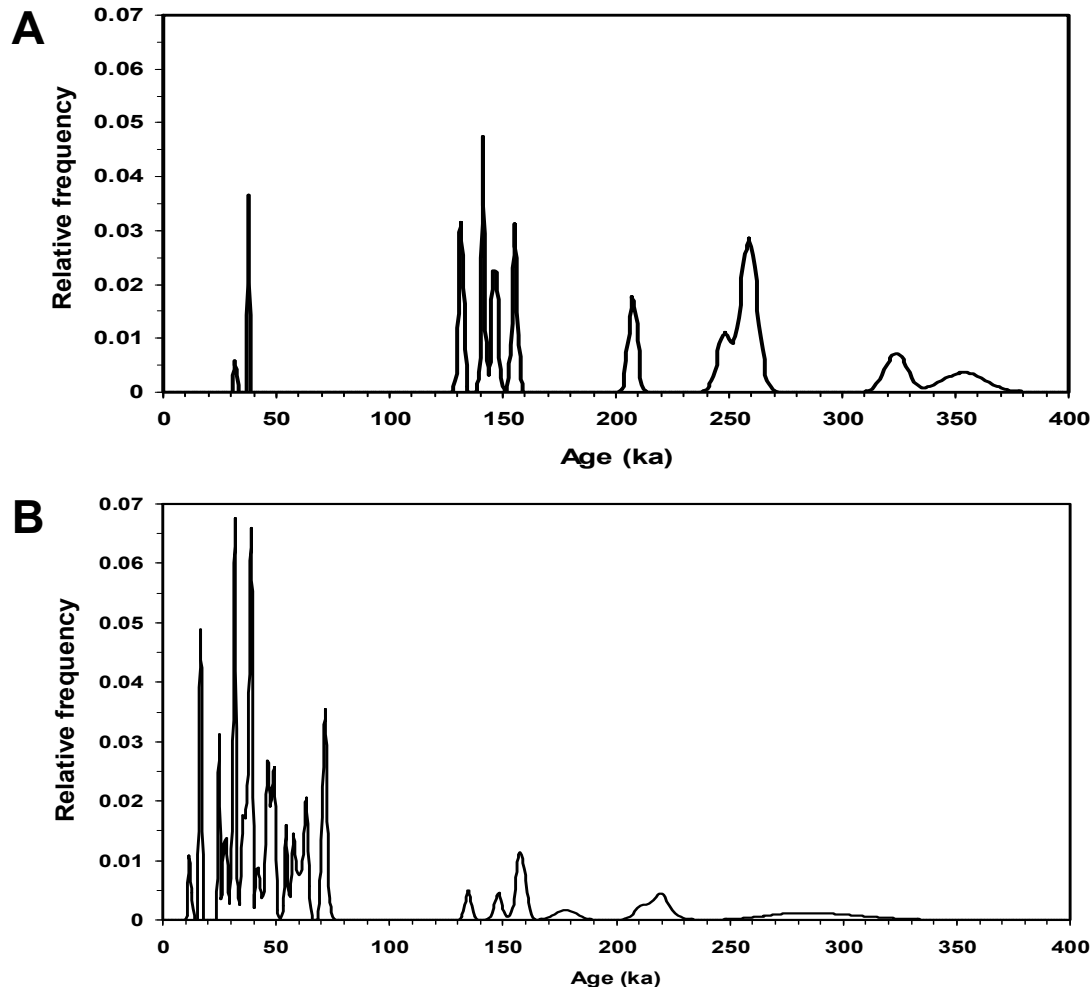


Figure 11 (next page): Dating of central and southern Negev speleothems.

A – Age distribution and periods of speleothem deposition of central and southern Negev Desert as determined by U-Th dating. The horizontal axis marks the age (ka). The ages are arranged vertically according to their speleothem number (bold font) and individual sample number (smaller font). Black circles with error bars define individual age measurements. Vertical rectangles mark speleothem depositional periods: green rectangles indicate periods of intensive speleothem deposition (two or more contemporaneous ages), and yellow rectangles indicate periods of limited speleothem deposition (one age only). The grey rectangles on the right define the periods between 350 ka to 600 ka in which the age $\pm 2\sigma$ uncertainties are large, and precise timing of the speleothem deposition periods is impossible. Dark grey rectangles show periods of higher relative frequency of the speleothem deposition, light grey rectangles show periods when the frequency is low.

B – The relative age frequencies indicating the fraction of speleothems deposited at a certain age with 95% confidence.

C – Timing of humid periods in southern Saharan-Arabian Desert (rectangles mark the humid periods) compared with periods of speleothem deposition in the Negev: Ref. 1 – (Fleitmann et al., 2003); Ref. 2 – (Osmond and Dabous, 2004) (dark blue – most humid periods, light blue – less humid periods, white – dry periods); Ref. 3 – (Crombie et al., 1997); Ref. 4 – (Szabo et al., 1995)).

Figure 11:

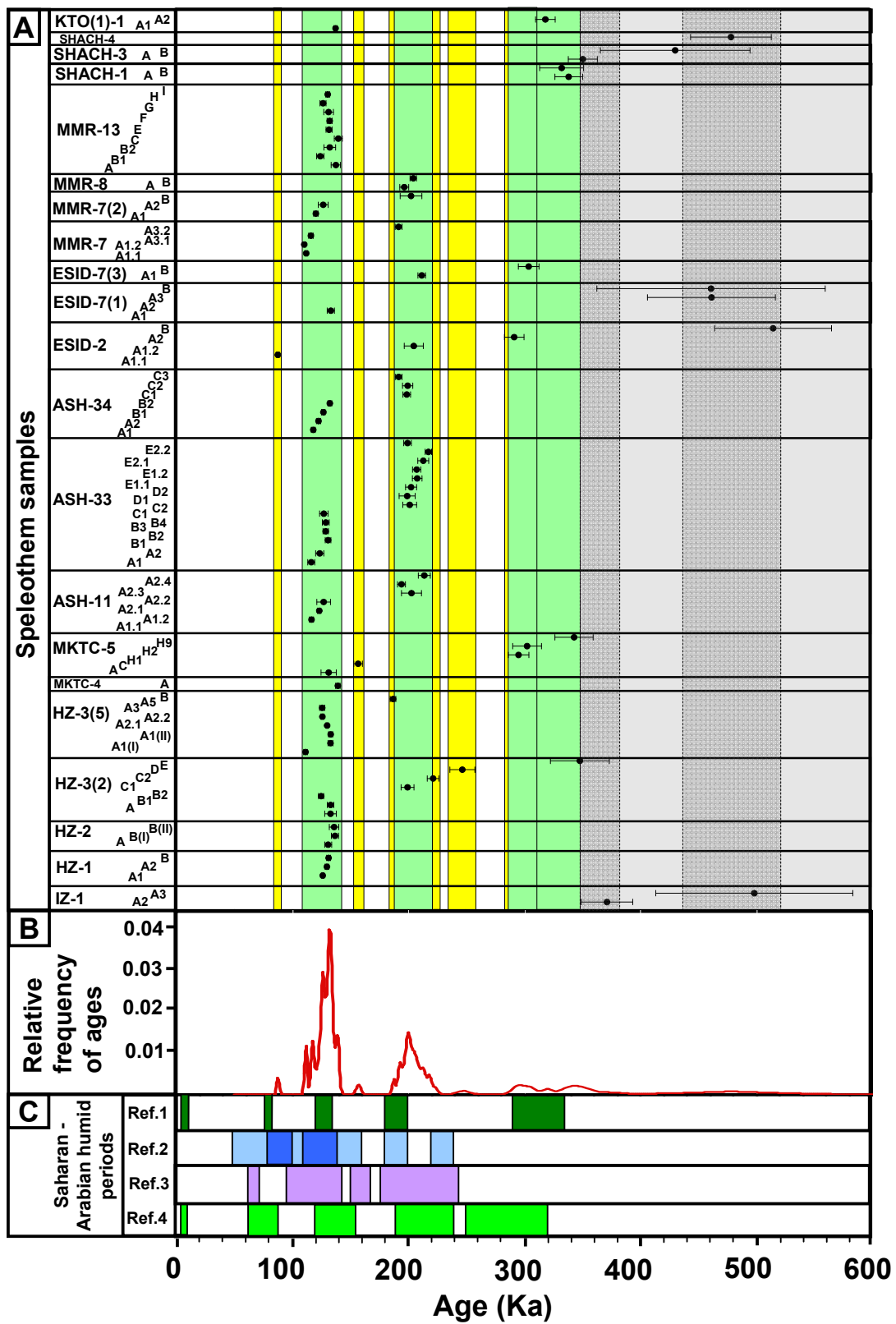


Figure 12: Plots of the relative age frequencies versus age for 150-100 ka (A); 230-180 ka (B) and 550-250 ka (C).

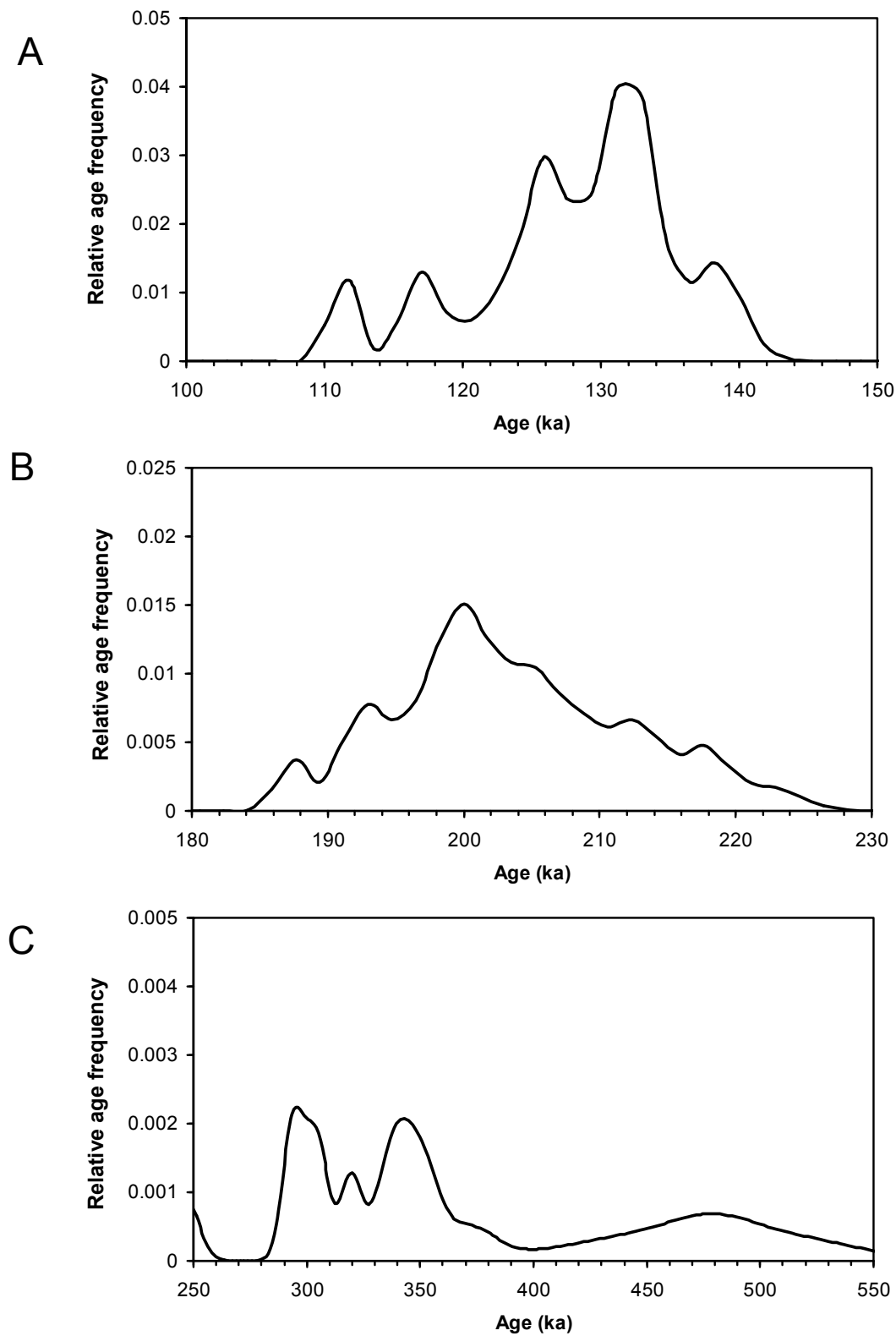


Figure 13: Sample ASH-33 from the Ashalim Cave. 1 – 2 pieces of flowstone ASH-33 with a drill holes for U-Th dating in the left piece. The dotted lines correlate between the top of the lamina C and the base of lamina E in the two pieces; 2 – the detailed stratigraphy of the sample. The hiatus E-F appears as orange clay lamina, two rock fragments (RF) overlay the hiatus. The areas photographed under microscope (3, 4, and 5) are marked by rectangles in 2; 3 - columnar fabric typical for speleothems grown in a closed system; 4 - the lamina E2.2 grown in a open system when the secondary calcite has grown in the pores between the rock fragments and the clay lamina (plane polar light); 5 - the same in cross polar light.

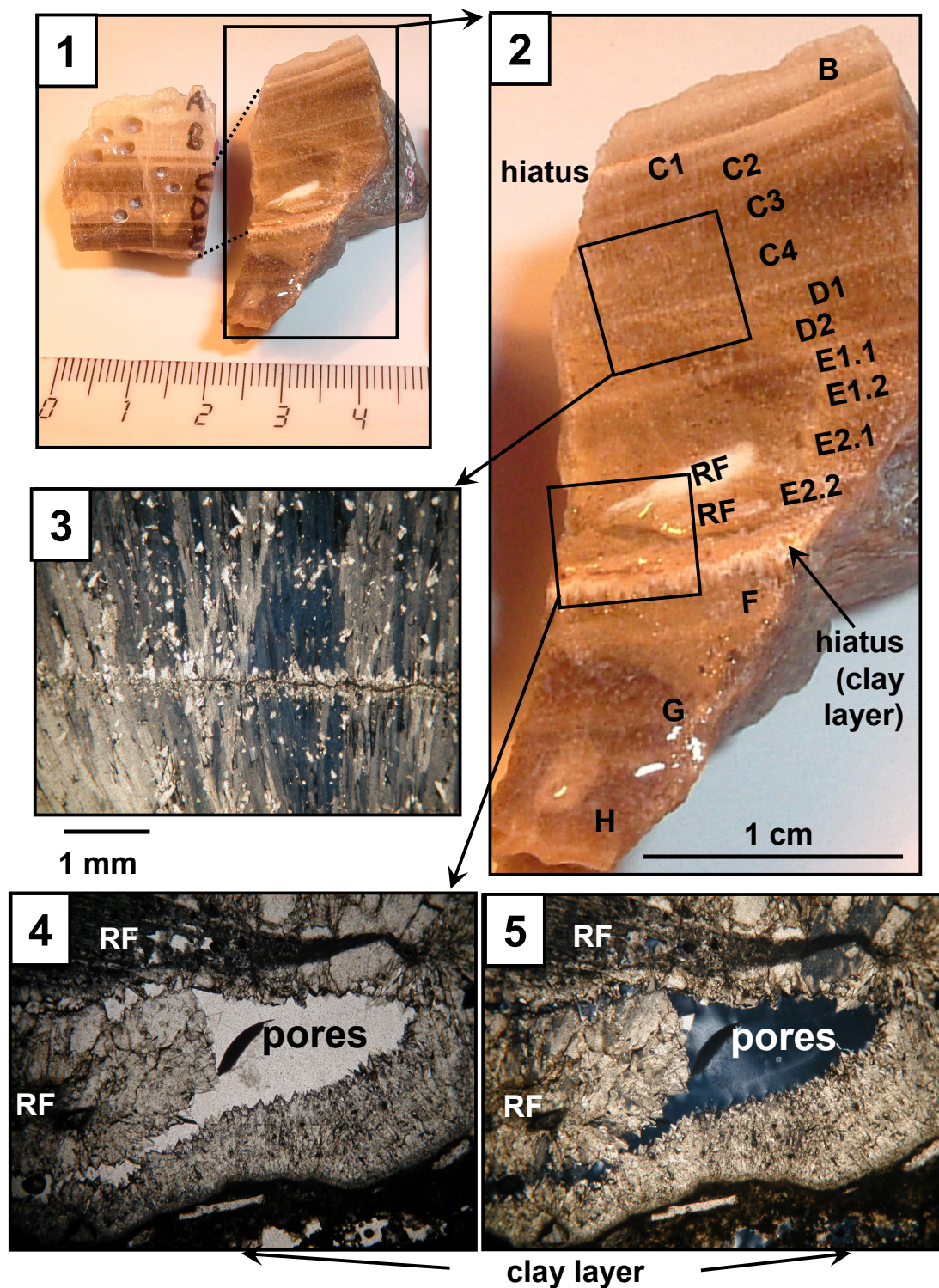


Figure 14: $\delta^{18}\text{O}$ values of central and southern Negev speleothems during the last ~3Ma. Samples younger than 350 ka are enclosed by the large rectangle. The small rectangles enclose samples that grew at 516 ± 51 ka, 1.269 ± 0.013 Ma (preliminary U-Pb age) and between 3.02 Ma and 3.07 Ma (preliminary U-Pb ages). Samples older than 350 ka are marked by green circles. The samples in the light blue oval fields are between 550 ka to 1.3 Ma, and older than 3.07 Ma, they were not dated and are arranged according to their stratigraphic order.

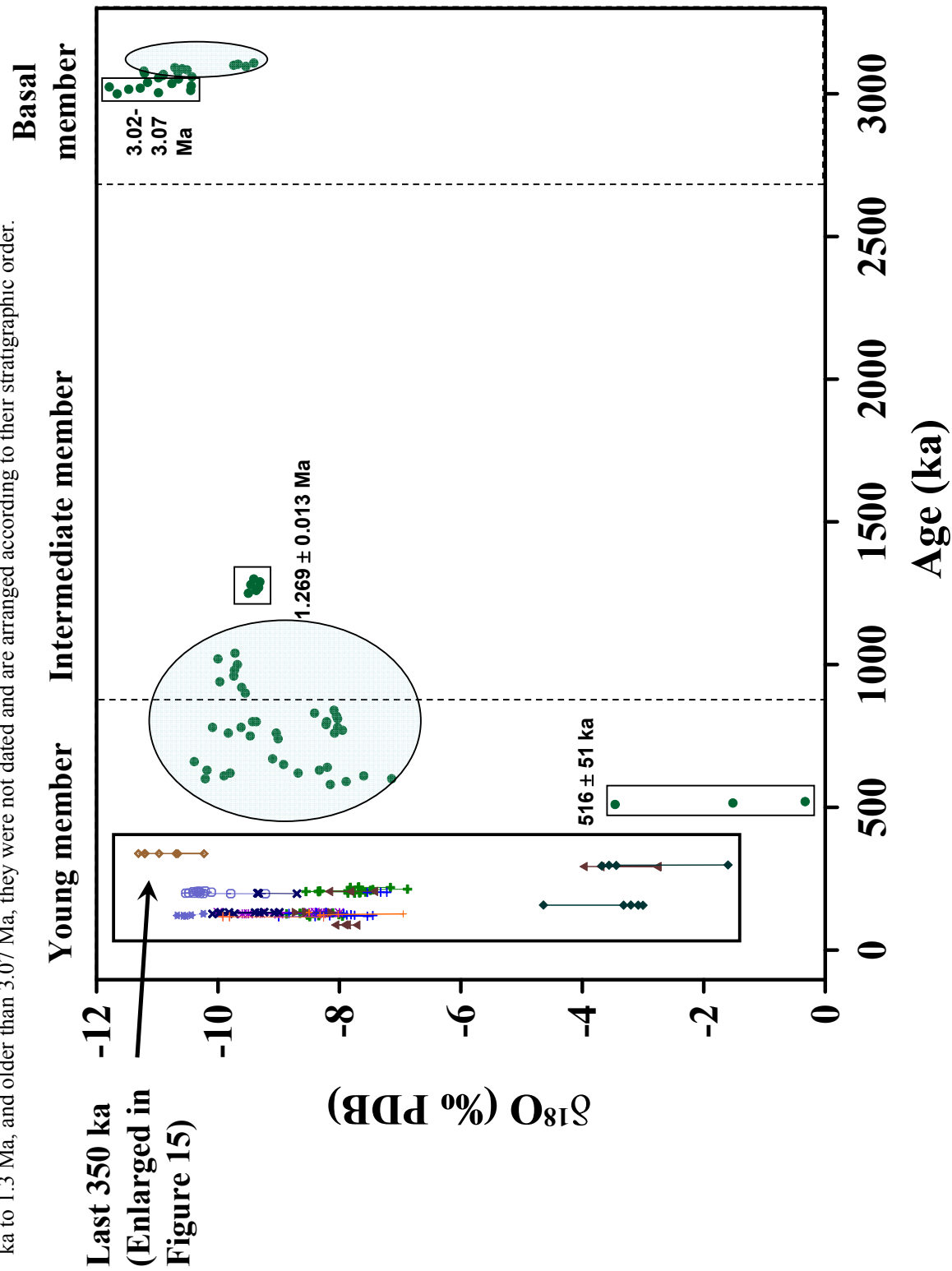


Figure 15: Oxygen isotope profiles of central and southern Negev speleothems during the last 350 ka. (A) speleothems deposited during the last 350 ka; (B) between 89 and 86 ka; (C) between 161 and 154 ka; (D) between 305 ka and 285 ka; (E) between 350 ka and 315 ka. U-Th ages with $\pm 2\sigma$ error bars for samples deposited during short episodes are shown in top of plots B-E.

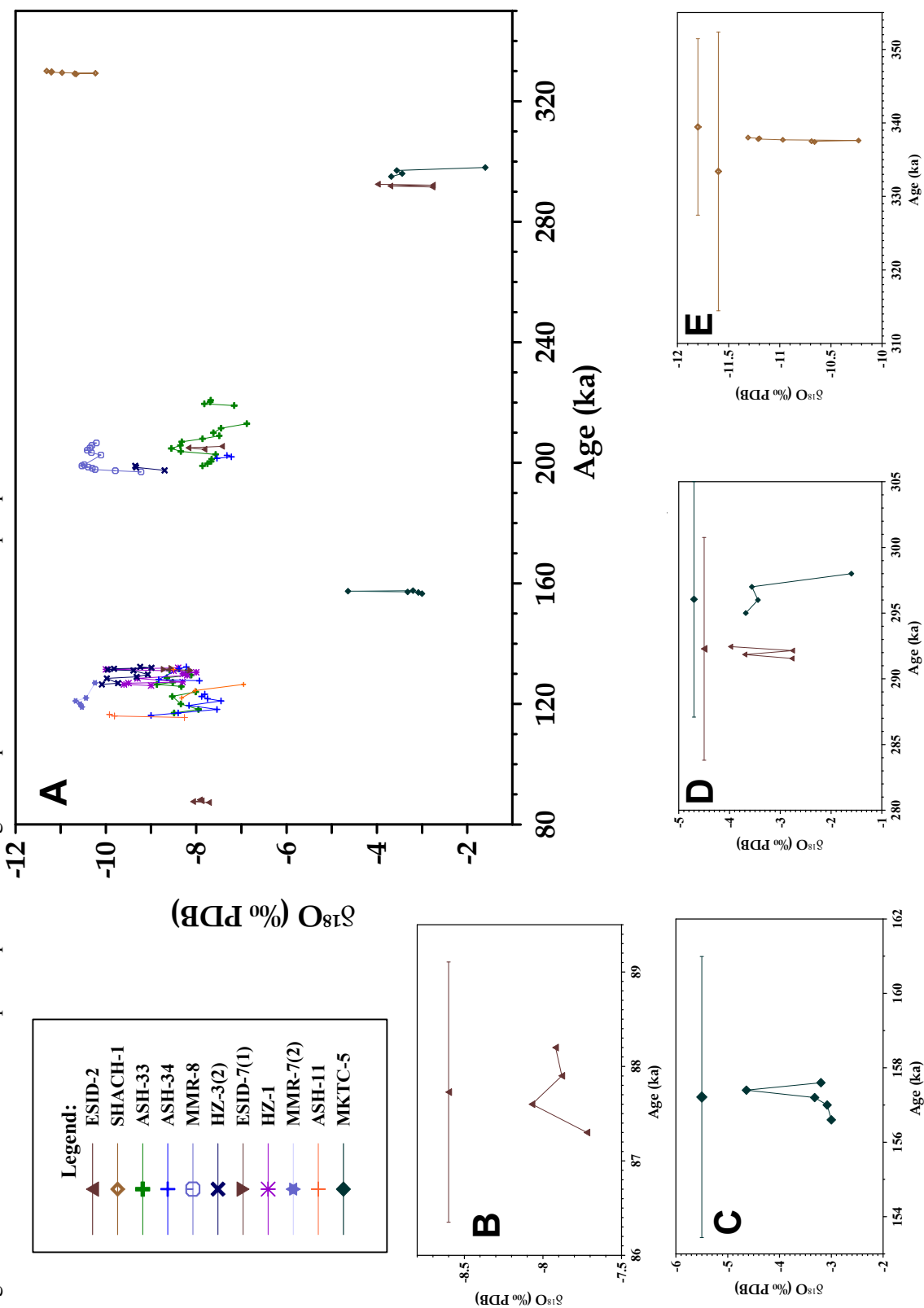


Figure 16: Oxygen isotope profiles in central Negev during the two major periods of speleothem deposition. (A) Oxygen isotope profiles of central and southern Negev speleothems sampled between 133 ka and 115 ka. U-Th ages with 2 σ errors are given in the top of the diagrams; (B) data as in A with running average of $\delta^{18}\text{O}$ between 133 ka and 126 ka (marked by black line); (C) Oxygen isotopic profiles between 221 ka and 197 ka. U-Th ages with 2 σ errors are given in the tops of the diagram; (D) data as in C with running average of $\delta^{18}\text{O}$ (marked by black line).

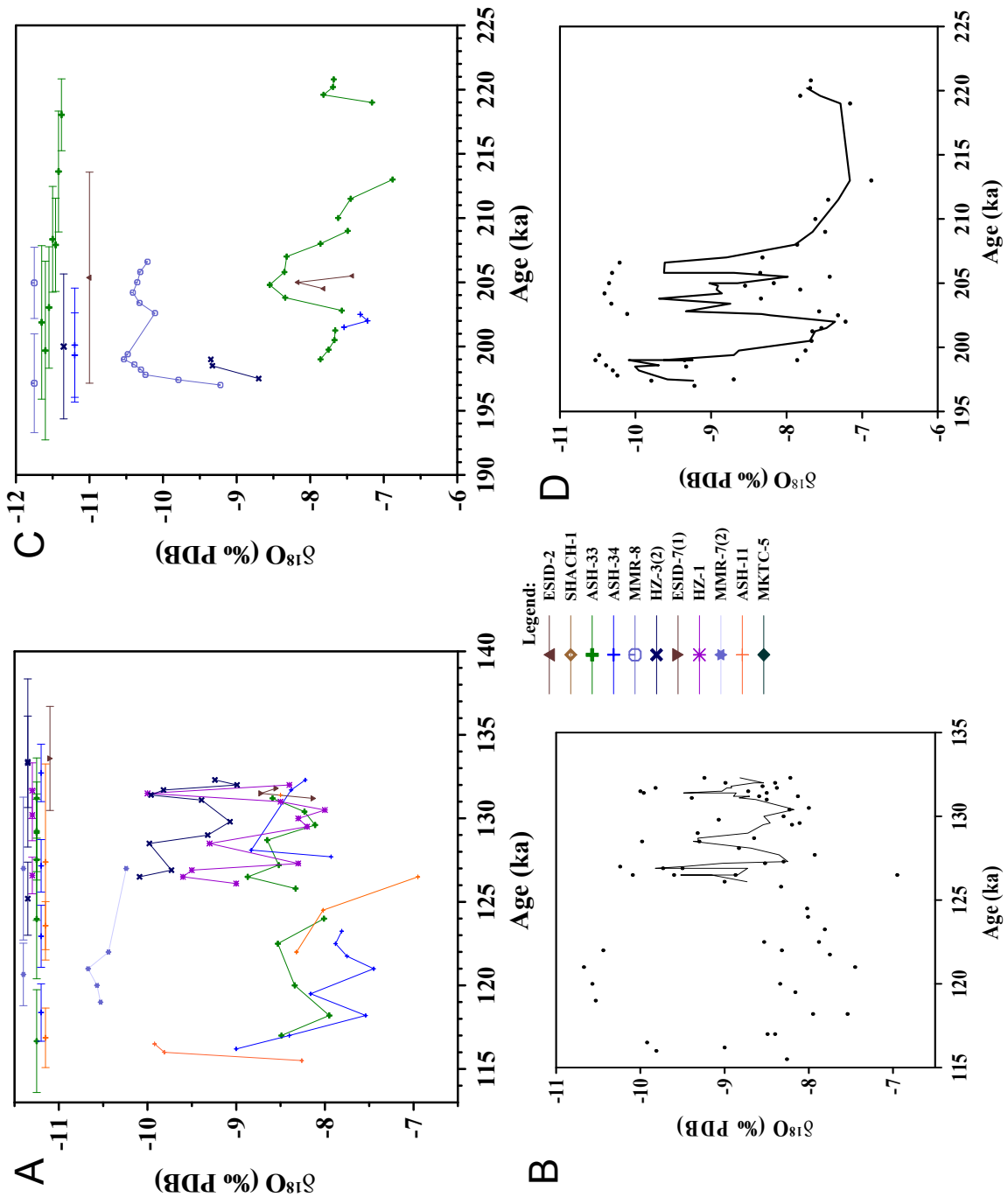


Figure 17: $\delta^{13}\text{C}$ values of central and southern Negev speleothems during the last ~3 Ma. Samples younger than 350 ka are enclosed by a large rectangle. The small rectangles enclose samples that grew at 516 ± 51 ka, 1.269 ± 0.013 Ma (preliminary U-Pb age) and between 3.02 Ma and 3.07 Ma (preliminary U-Pb ages). Samples older than 350 ka are marked by green circles. The samples in the light blue fields are between 550 ka to 1.3 Ma, and older than 3.07 Ma, were not dated and are arranged according to their stratigraphic order.

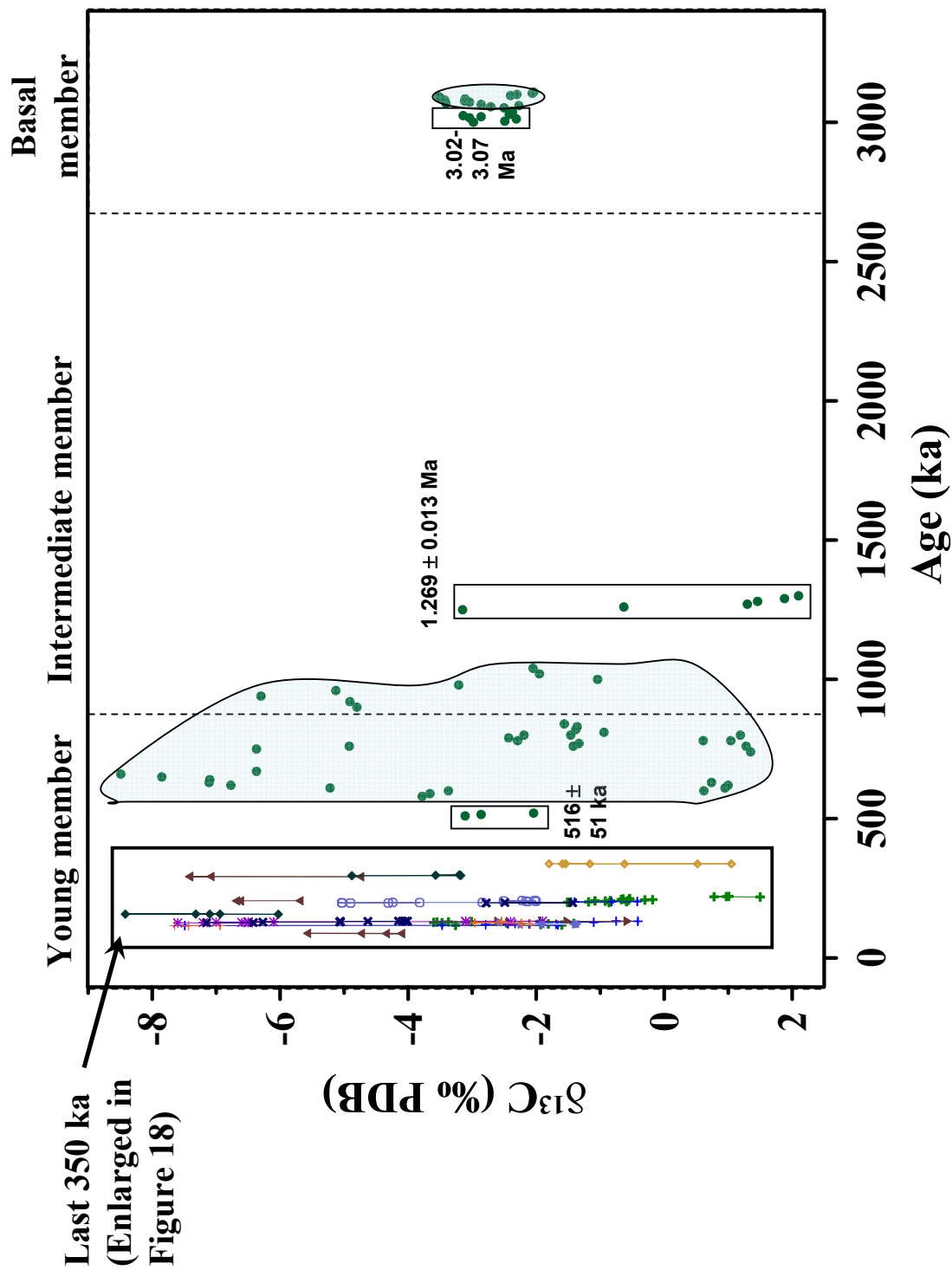


Figure 18: Carbon isotope profiles of central and southern Negev speleothems during the last 350 ka. (A) all speleothems formed during the last 350 ka; (B) between 89 and 86 ka; (C) between 161 and 154 ka; (D) between 305 ka and 285 ka; (E) between 350 ka and 315 ka. U-Th ages with $\pm 2\sigma$ error bars for samples deposited during short episodes are shown in the top of plots B-E.

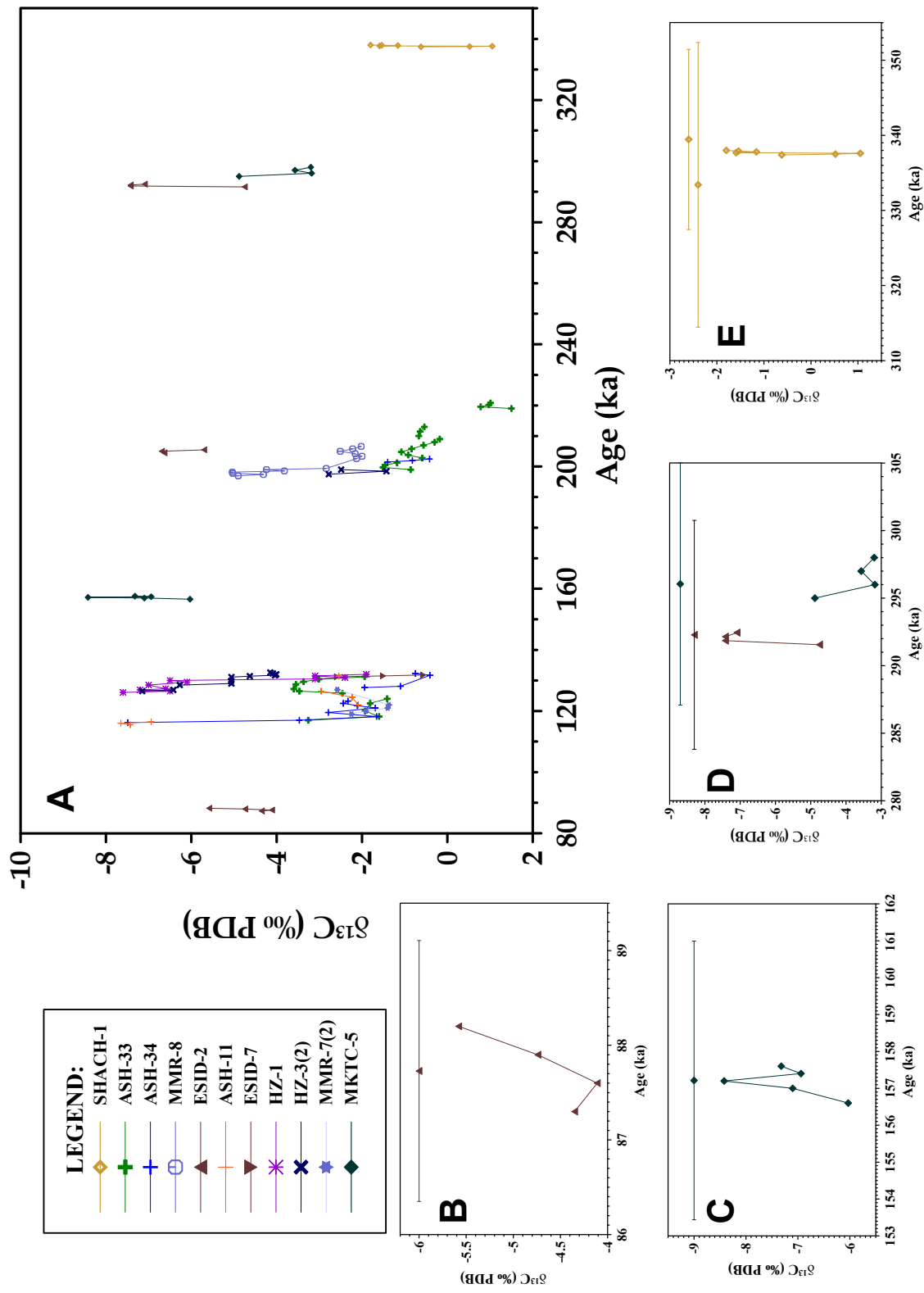


Figure 19: Carbon isotope profiles in central Negev during the two major periods of speleothem deposition (A) Carbon isotope profiles of central and southern Negev speleothems between 133 ka and 115 ka. U-Th ages with 2σ errors are given at the top of the diagram; (B) Data as in A, with running average of $\delta^{13}\text{C}$ marked by the black line; (C) Carbon isotopic profiles of central and southern Negev speleothems between 221 ka and 197 ka. U-Th ages with 2σ errors are given on the top of the diagram; (D) Data as in C, with running average of $\delta^{13}\text{C}$ marked by the black line.

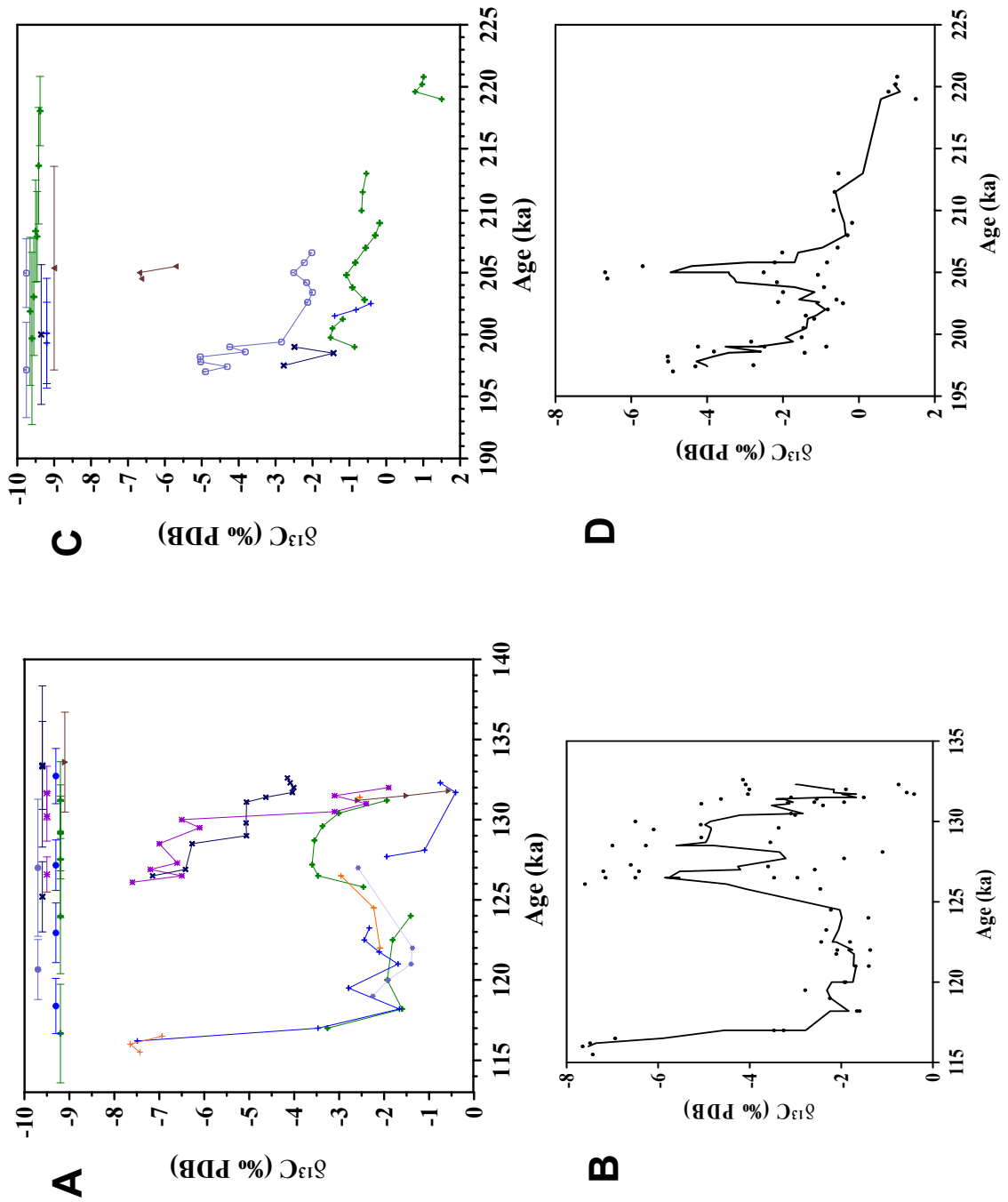


Figure 20: Northern Negev rainfall isotopic composition. A - Oxygen and hydrogen isotopic composition of northern Negev rainfall during the 2004-2006 rainy seasons (September-May) including results of Vaks et al. (2006). (B) enlarged central part of A. Events that have undergone most significant evaporation marked by oval in A and form the Be'er-Sheva evaporation line (black dashed line in A and B).

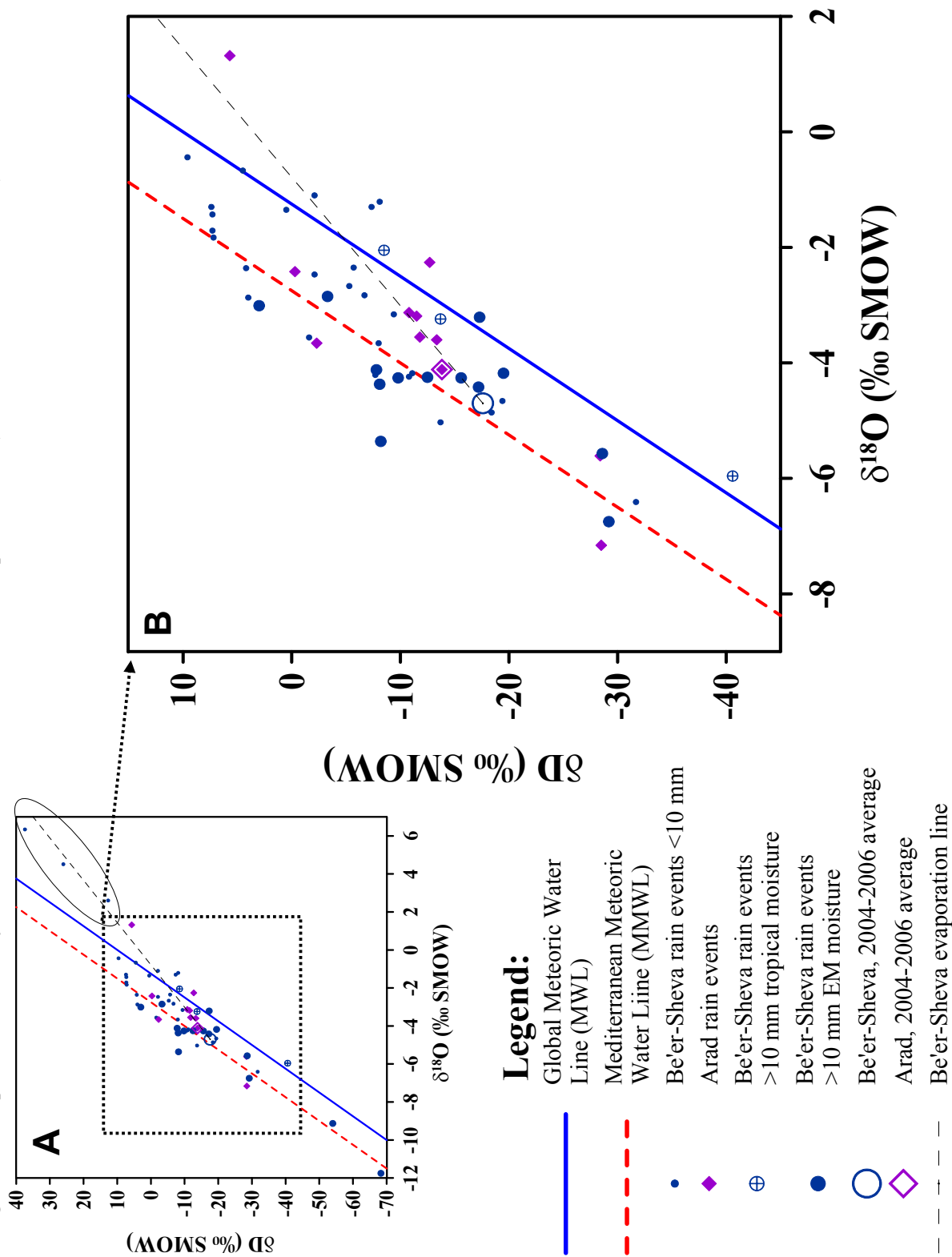


Figure 21: Central and southern Negev rainfall isotopic composition. A - Oxygen and hydrogen isotopic composition of the rainfall in central and southern Negev during 2004-2006 (September to May) in Mitzpe-Ramon and Neot-Smadar, and during 2002-2006 in Makhtesh-ha-Qatan (including the data of Kuperman (2005)); B - enlarged central part of A; C - the average isotopic compositions of the five rainfall collecting stations in the Negev plotted relative to the evaporation line of Be'er-Sheva rainfall shown in Fig. 20 (A, B).

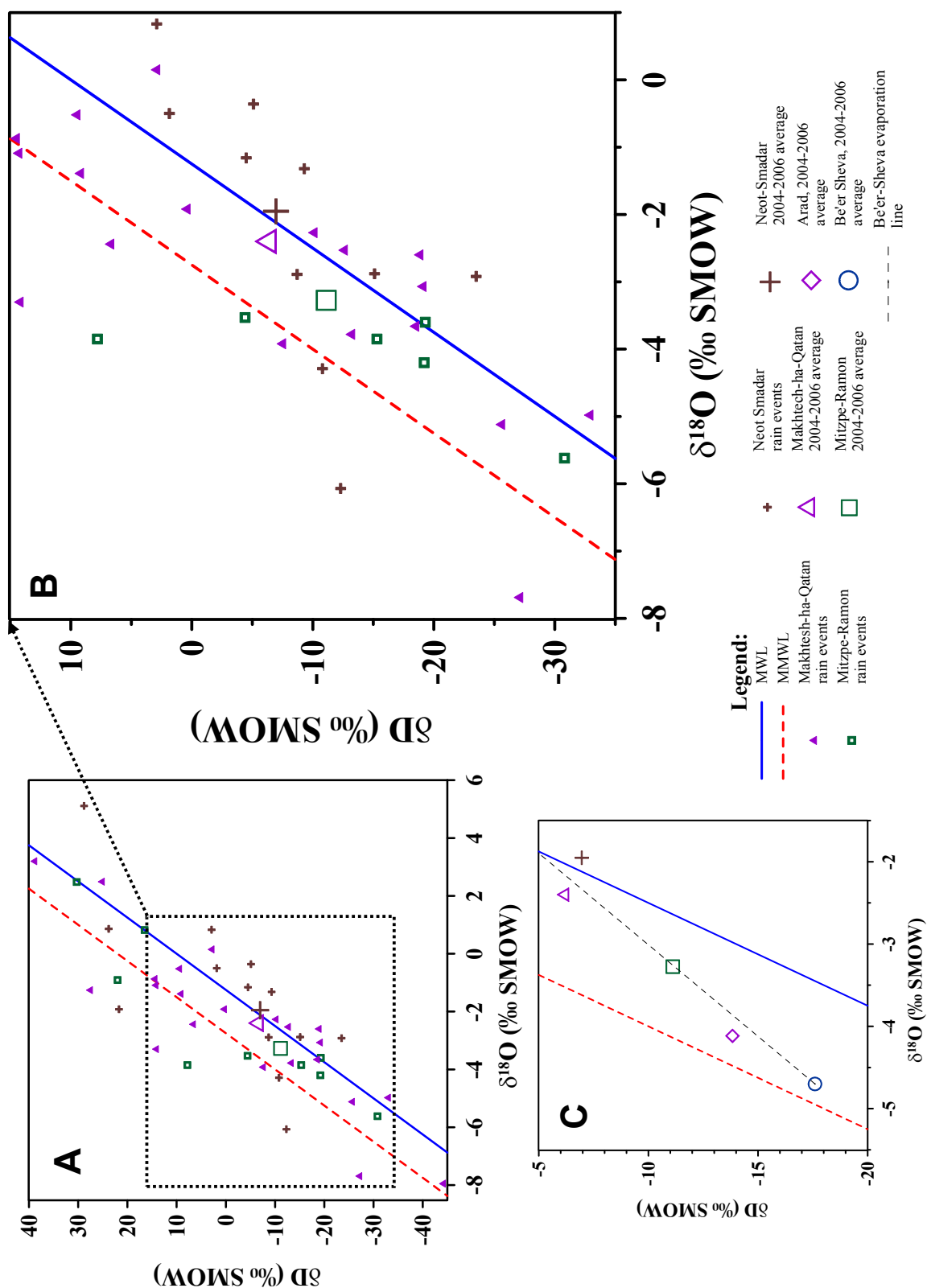


Figure 22: δD values of Negev Desert speleothem fluid inclusions (FI) during the last 220 ka with the trend line in grey (A); during the last 3.1 Ma, samples with preliminary U-Pb ages are enclosed in two small rectangles (B). FI of Tzavoa, Ashalim, Hol-Zakh and Ma'ale-ha-Meyshar Caves are shown by blue, green, purple and orange triangles, respectively. The samples in the light blue oval filed (B) are between 550 ka to 1.3 Ma, they were not dated and are arranged according to their stratigraphic order.

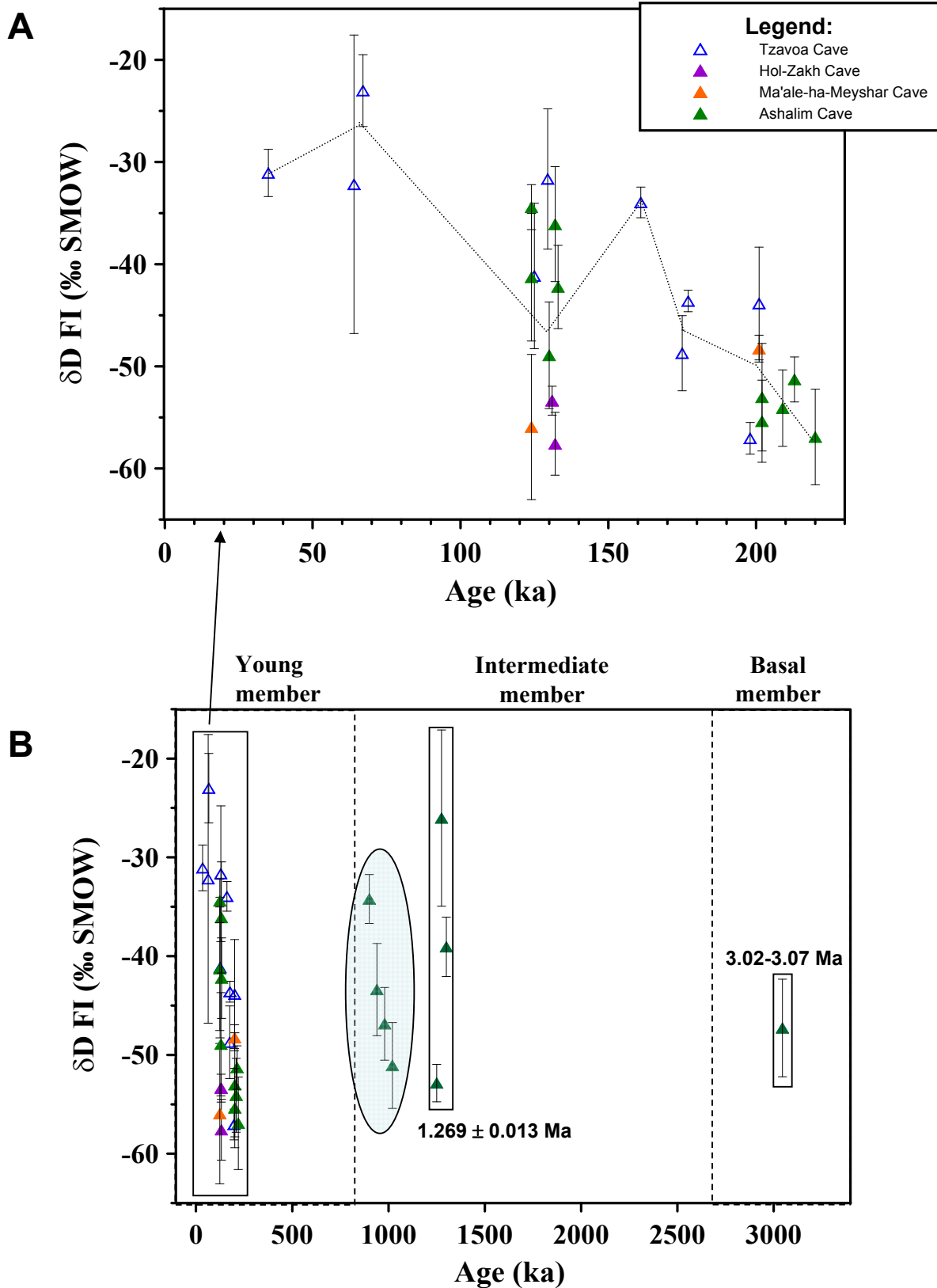


Figure 23: Sr concentrations in the Negev and Jordan Valley speleothems; during the last 3.3 Ma, samples with preliminary U-Pb ages enclosed in rectangles (A); during the last 300 ka in Jordan Valley (B); during the last 200 ka in the northern Negev: Tzavoa cave speleothems are shown by olive green data points linked by lines, and Ma'ale-Dragot by green data points (C); during the last 350 ka in the central and southern Negev (D). The samples in the light blue oval fields are between 550 ka to 1.3 Ma, and older than 3.07 Ma, they were not dated and are arranged according to their stratigraphic order (A).

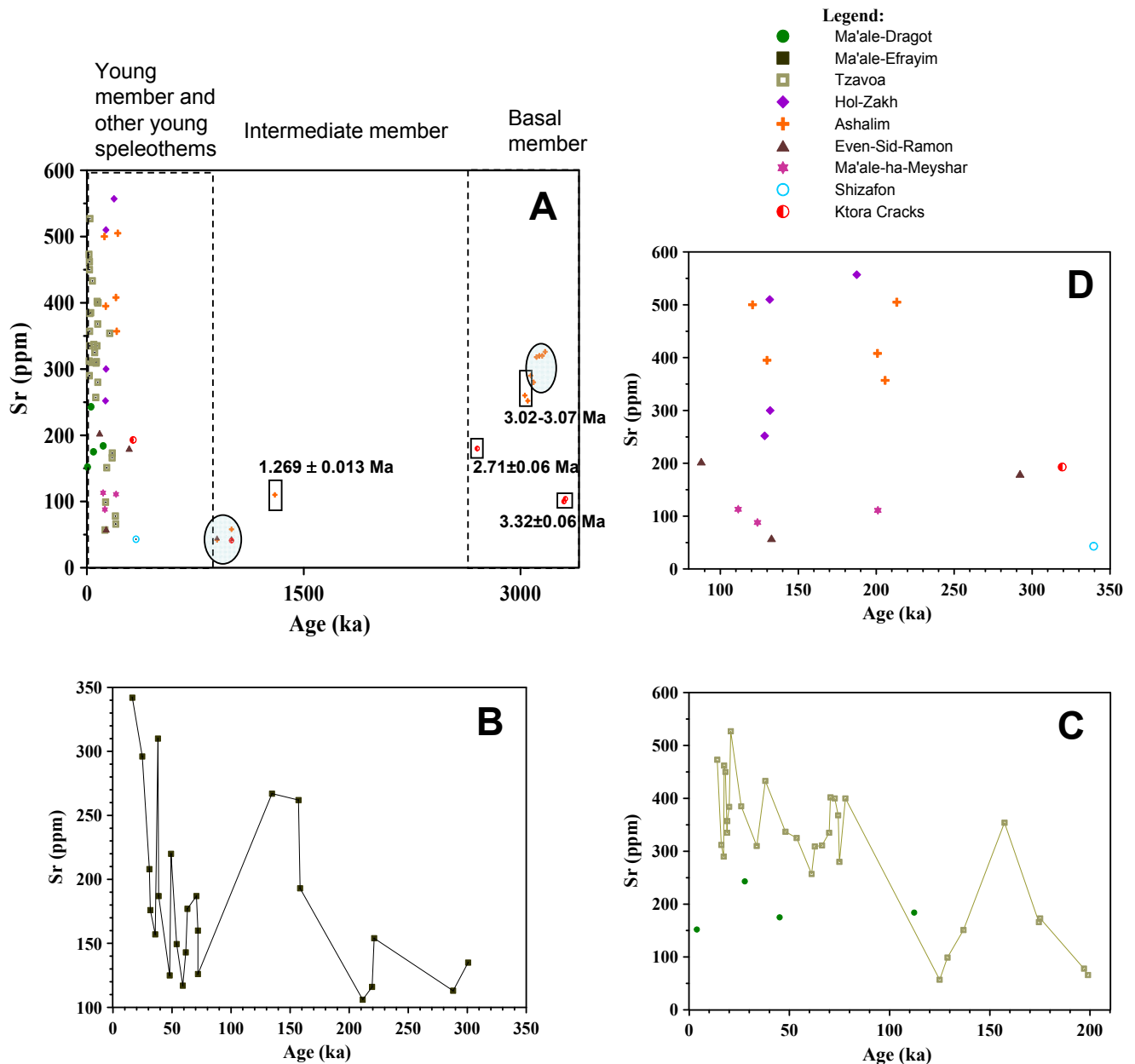


Figure 24: $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios of Ma'ale-Efrayim Cave speleothems; during the last 300 ka (A); $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios of Ma'ale-Efrayim Cave speleothems compared with $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic composition of dolomitic host rock, bulk soil above the cave and silicate fraction of the soil (B).

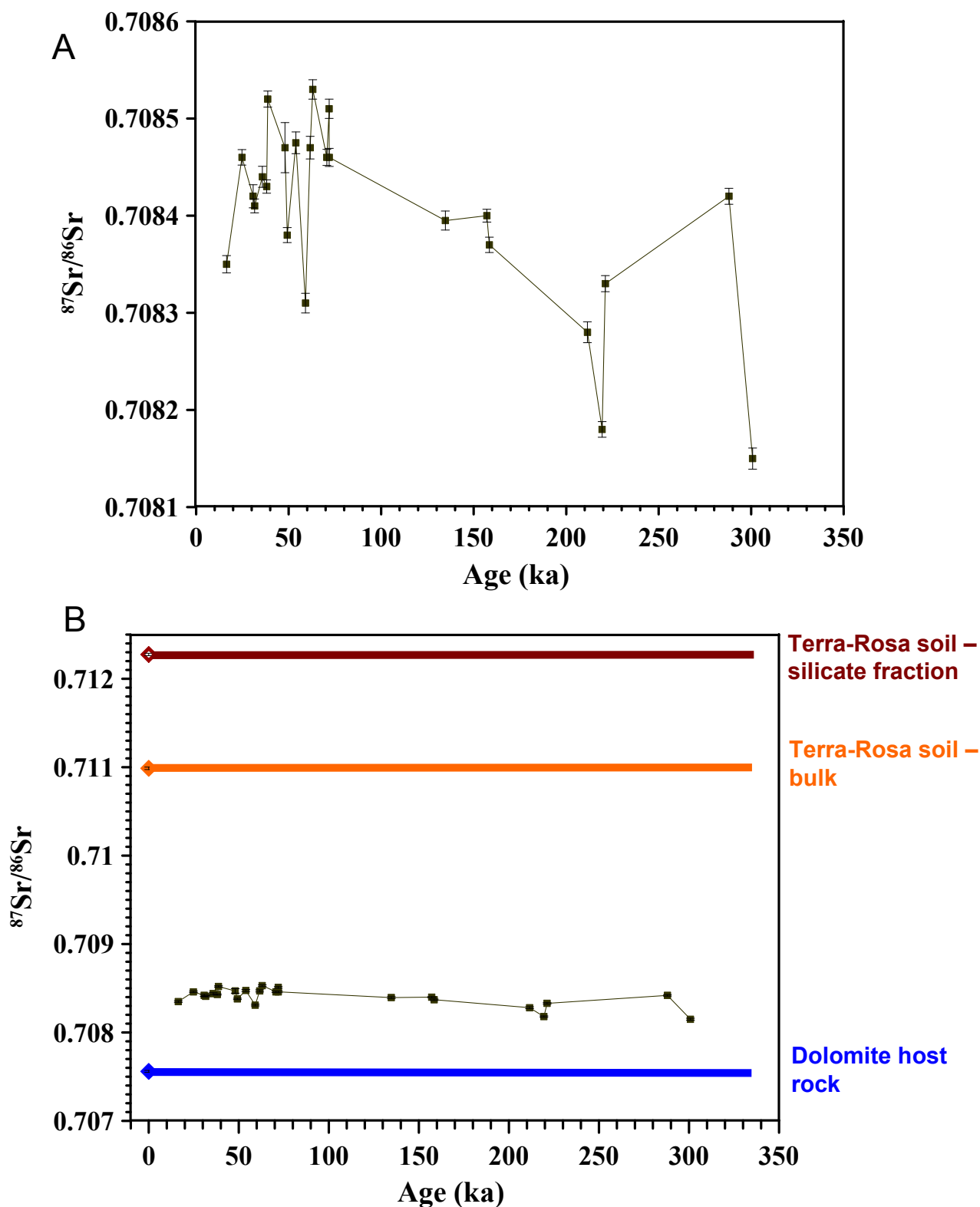


Figure 25: $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Negev Desert vadose speleothems; formed during the last 3.3 Ma, preliminary U-Pb ages shown by small rectangles (A); and during the last 350 ka (B). The samples in the light blue oval fields (A) are between 550 ka to 1.3 Ma, and older than 3.07 Ma, they were not dated and are arranged according to their stratigraphic order.

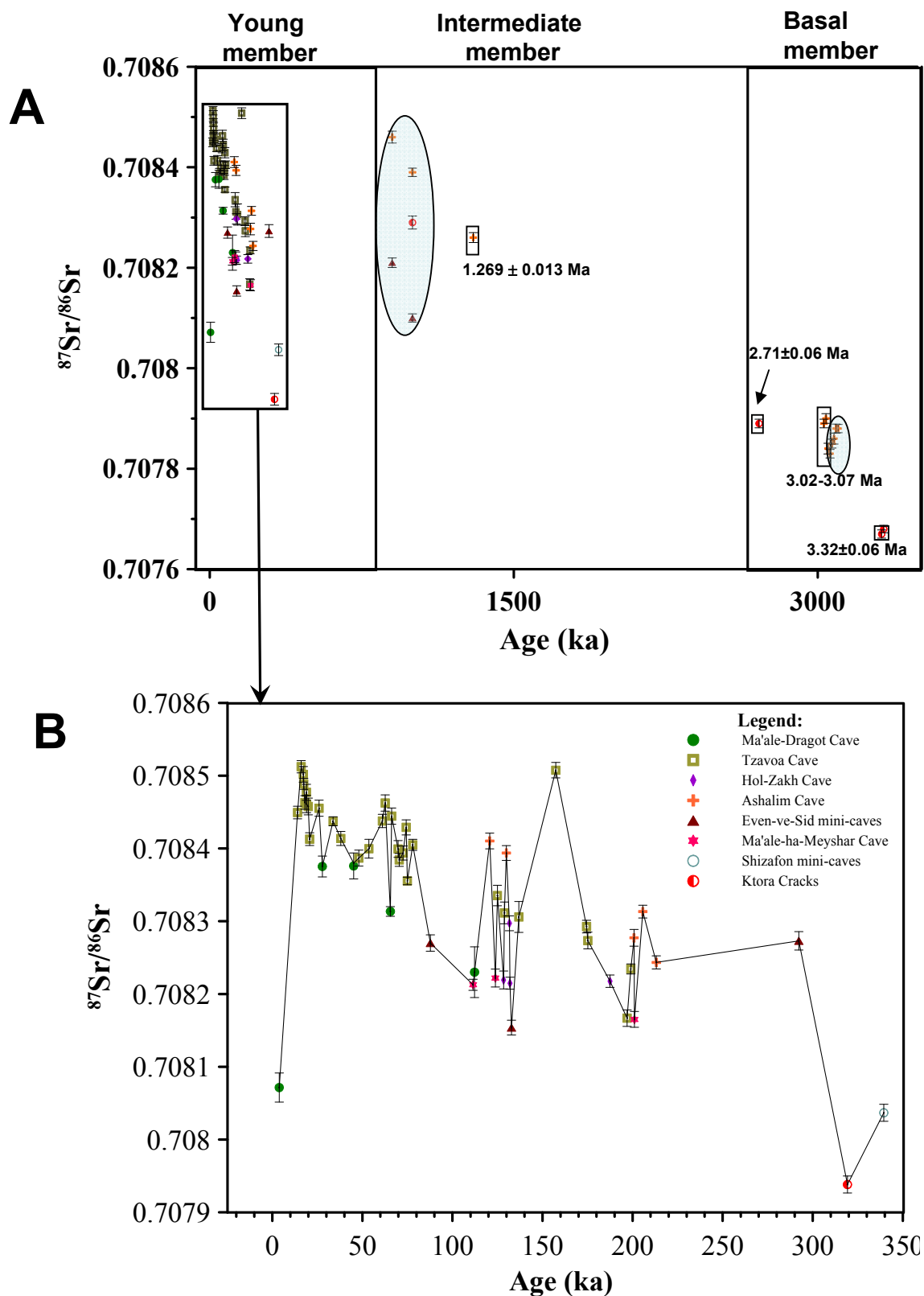


Figure 26: $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in Negev Desert vadose speleothems compared with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of dolomitic host rock, of bulk soil above the caves and of its silicate fraction; during the last 3.3 Ma (A); $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in Negev Desert vadose speleothems during the last 350 ka compared with the same parameters as in A. Preliminary U-Pb ages are enclosed in small rectangles. The samples in the light blue oval fields are between 550 ka to 1.3 Ma, and older than 3.07 Ma, they were not dated and are arranged according to their stratigraphic order.

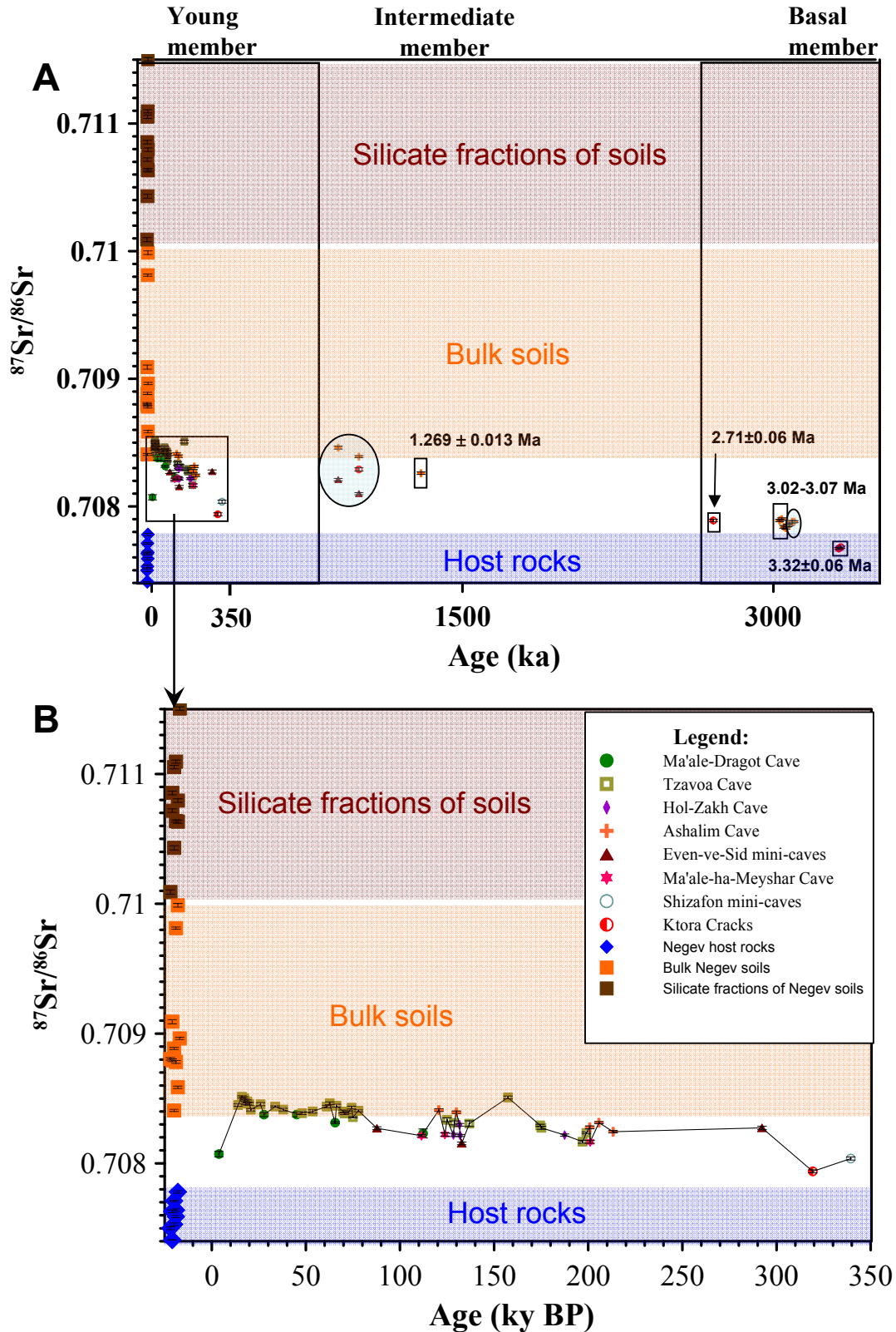
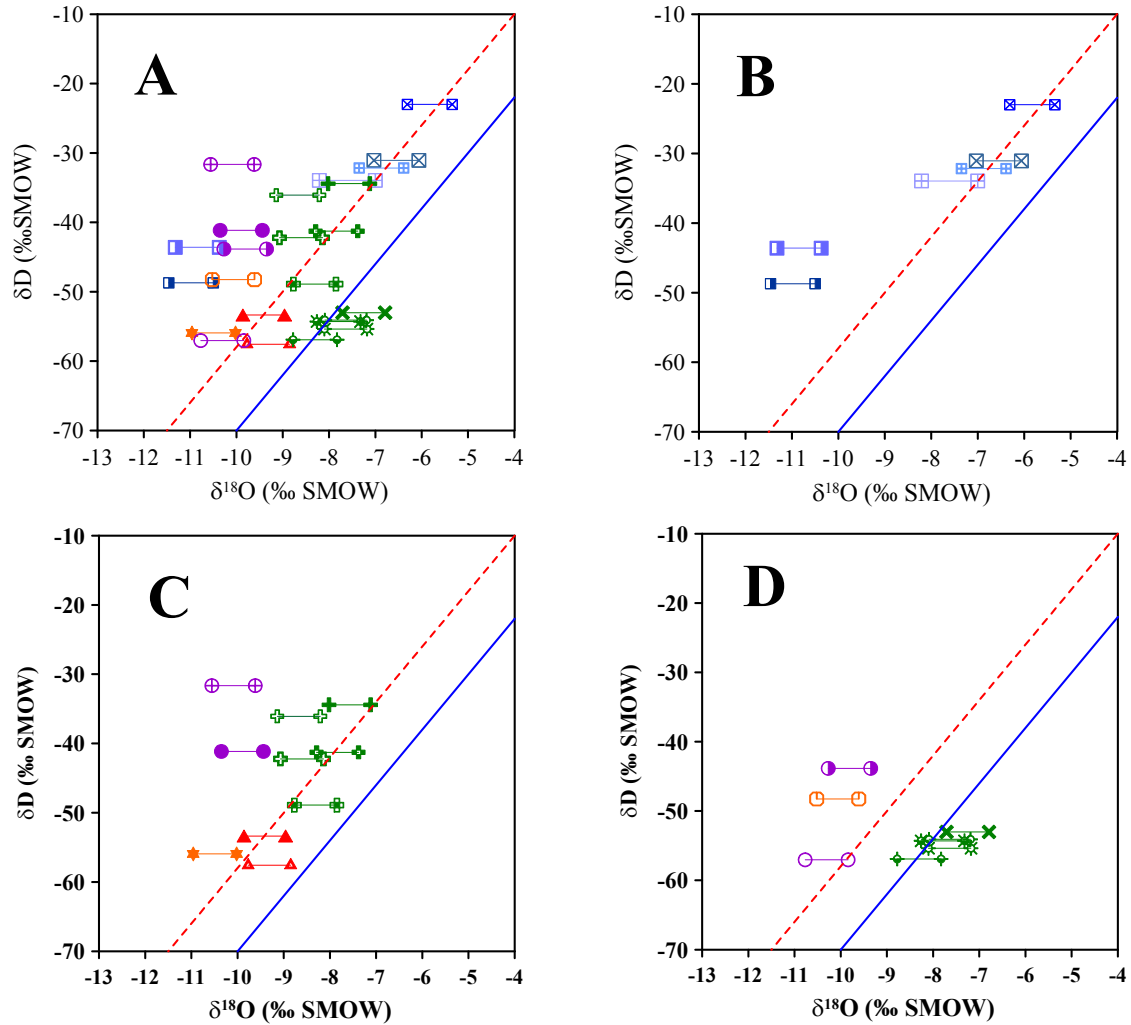


Figure 27: Hydrogen and oxygen isotopic composition of paleo-cave waters based of speleothem fluid inclusion analysis: (A) Last 220 ka in all caves; (B) 2 last glacial periods (Tzavoa Cave); (C) NHP-1 in central and southern Negev; (D) NHP-2 in central and southern Negev. The MMWL (dashed red line) and MWL (blue line) are shown for reference. For each sample the range in the calculated $\delta^{18}\text{O}$ represents $\pm 2^\circ\text{C}$ variations in cave temperatures (see text for explanation).



Legend:

- | | |
|------------------------|----------------------------------|
| —□— Tzavoa 32-35 ka | —+— Ashalim 126-130 ka |
| —□— Tzavoa 61-64 ka | —+— Ashalim 127-132 ka |
| —□— Tzavoa 65-67 ka | —+— Ashalim 131-132 ka |
| —●— Tzavoa 124-126 ka | —*— Ashalim 199-202 ka |
| —⊕— Tzavoa 128-131 ka | —x— Ashalim 201-203 ka |
| —□— Tzavoa 157-161 ka | —)— Ashalim 202-208 ka |
| —□— Tzavoa 172-175 ka | —◇— Ashalim 210-213 ka |
| —□— Tzavoa 175-177 ka | —+— Ashalim 218-220 ka |
| —○— Tzavoa 195-198 ka | —★— Ma'ale-ha-Meyshar 120-128 ka |
| —⊖— Tzavoa 198-201 ka | —⊕— Ma'ale-ha-Meyshar 195-206 ka |
| —+— Ashalim 117-124 ka | —▲— Hol-Zakh 126-131 ka |
| —+— Ashalim 117-126 ka | —▲— Hol-Zakh 131-132 ka |
| —+— Ashalim 126-130 ka | — — — MWL |
| —+— Ashalim 127-132 ka | --- MMWL |
| —+— Ashalim 131-132 ka | |

Figure 28: d-excess values from speleothem fluid inclusions in central and southern Negev Desert during the last 220 ka plotted relative to the MWL (blue horizontal line), MMWL (red horizontal line) and maximum d-excess in present-day rainfall from EM origin (red dotted horizontal line) (Bar-Matthews et al., 2003; Kuperman, 2005). Tzavoa Cave samples are marked by blue, Ashelim Cave by green, Hol-Zakh by red and Ma'ale-ha-Meyshar by orange two-end arrows. For each sample the range d-excess represents $\pm 2^\circ\text{C}$ variations in cave temperatures (see text for explanation).

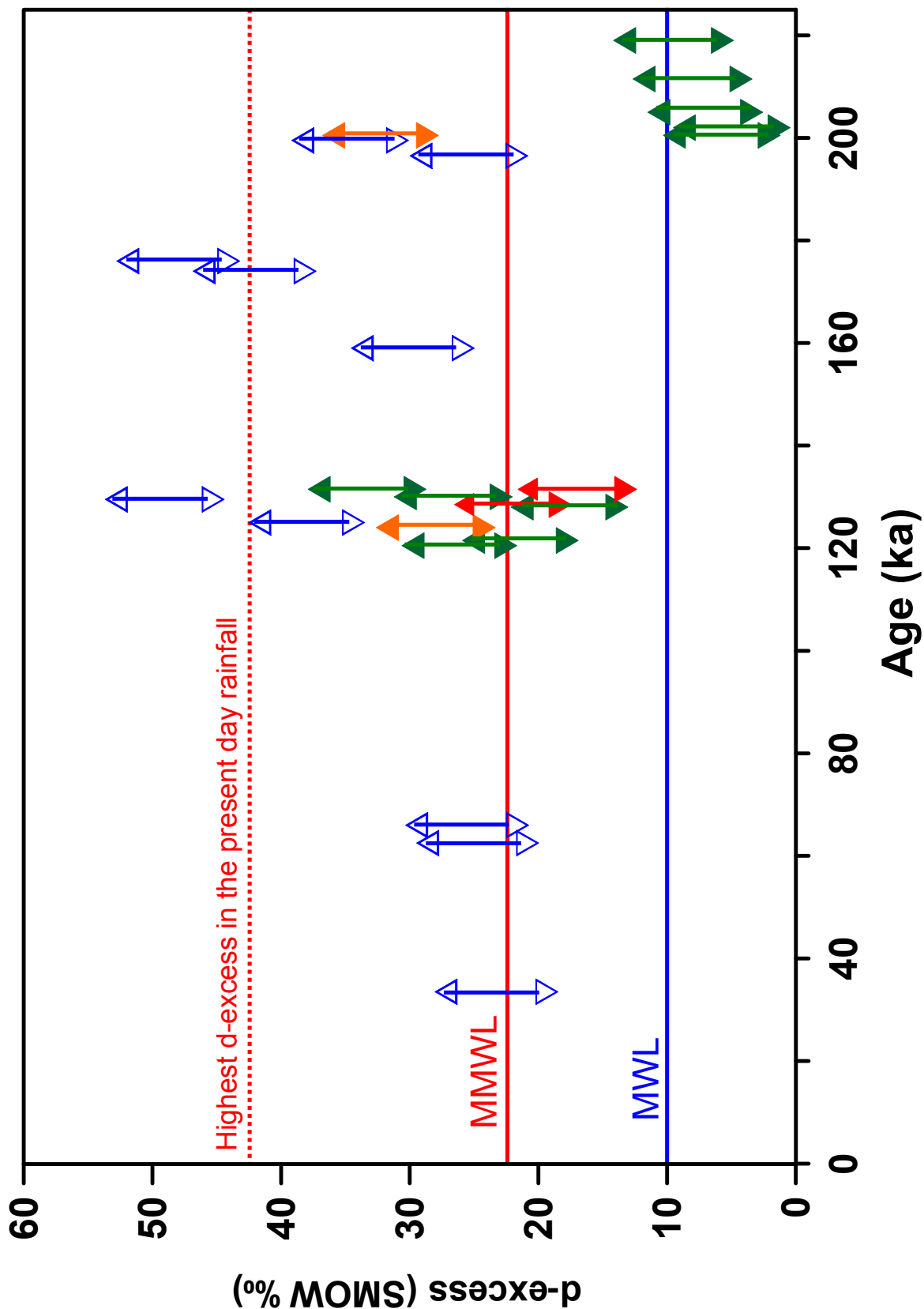


Figure 29: Thinning of the speleothem sequences from the north to the south.

Figure 29 (A-G): Width of the NHP-1 and NHP-2 sequences in the Negev speleothems. In Tzavoa Cave, northern Negev: width of the NHP-1 and NHP-2 sequences is ~ 6 cm (A, B). Ashalim Cave central Negev (C, D), and Ma'ale-ha-Meyshar Cave, the boundary between central and southern Negev (E, F): width of NHP-1 and NHP-2 sequences is up to 2.5 cm. Ktora Cracks, southern Negev: the width of NHP-1 sequence is reduced to a few millimeters, while the NHP-2 sequence disappears (G).

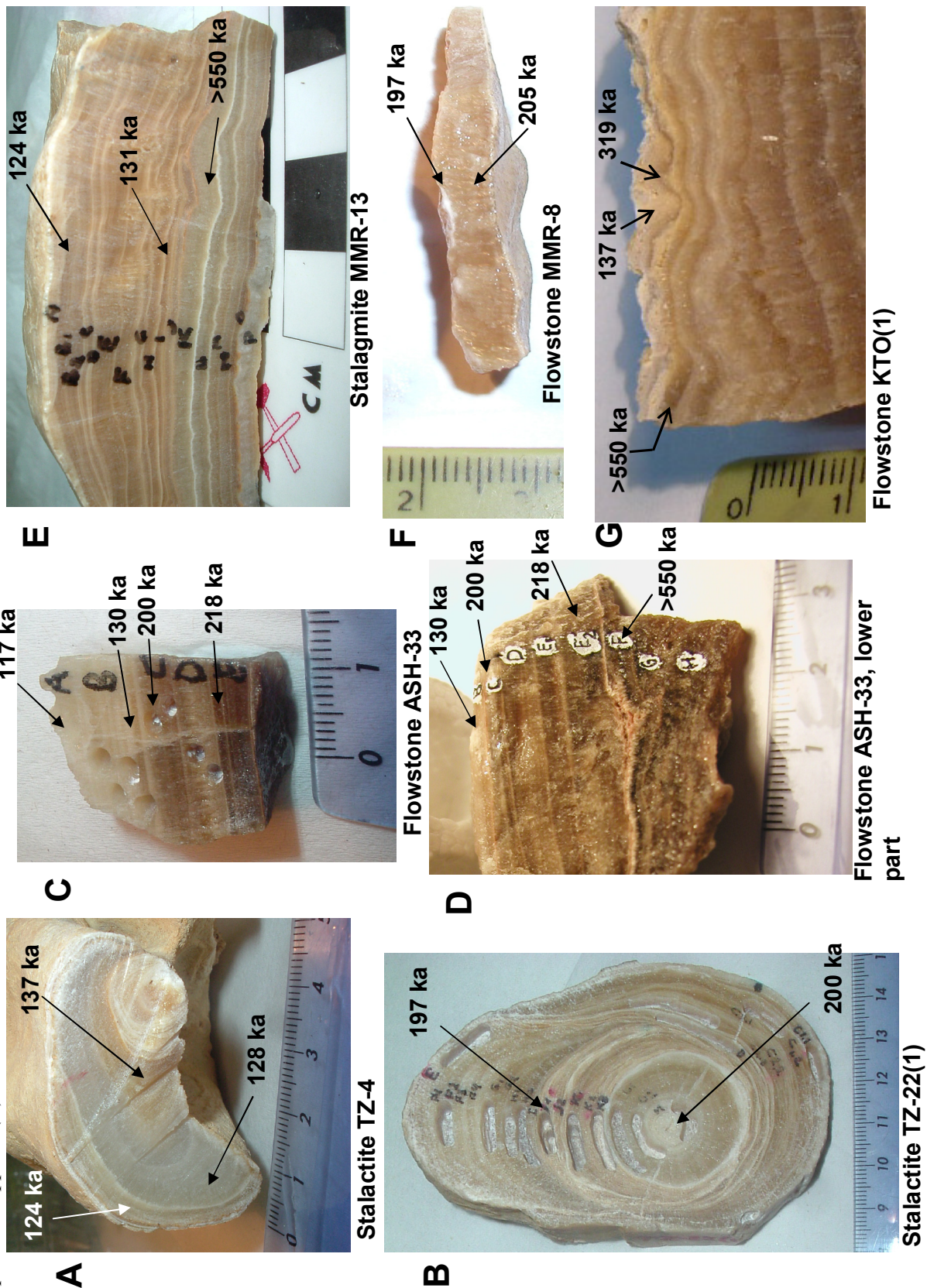


Figure 29 (H, I): Thinning of the speleothem laminae from the north to the south in the Negev Desert during the NHP-1 (H) and during the NHP-2 (I). The maximal thickness of the laminae decrease from the north to the south. Laminae with thickness of 0 cm represent depositional hiatuses.

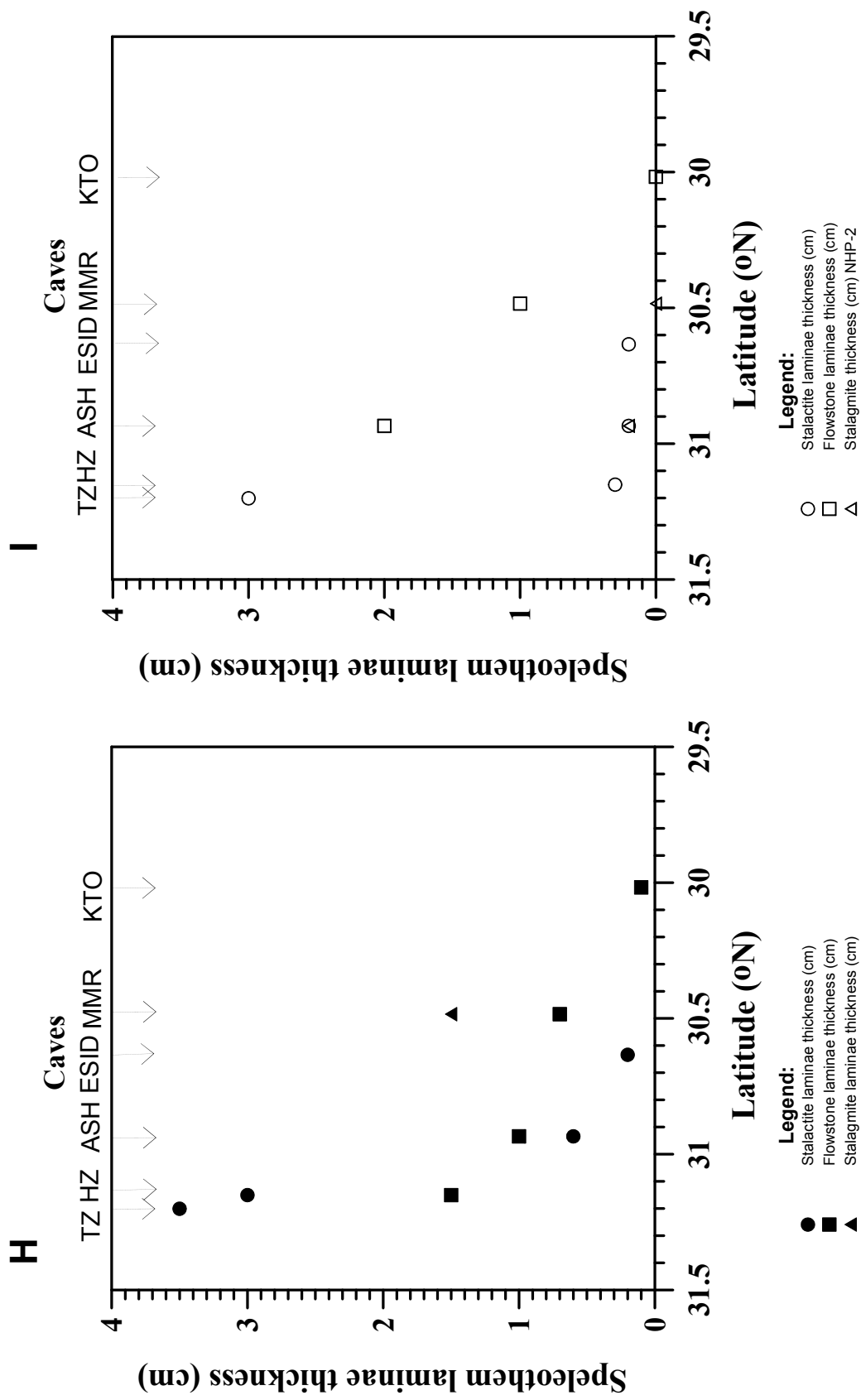


Fig. 29J: Shortening of the speleothem deposition periods from the north to the south.

Speleothem deposition during the glacial periods
 Speleothem deposition during the interglacial periods
 (incomplete data shown by striped pattern with question marks)

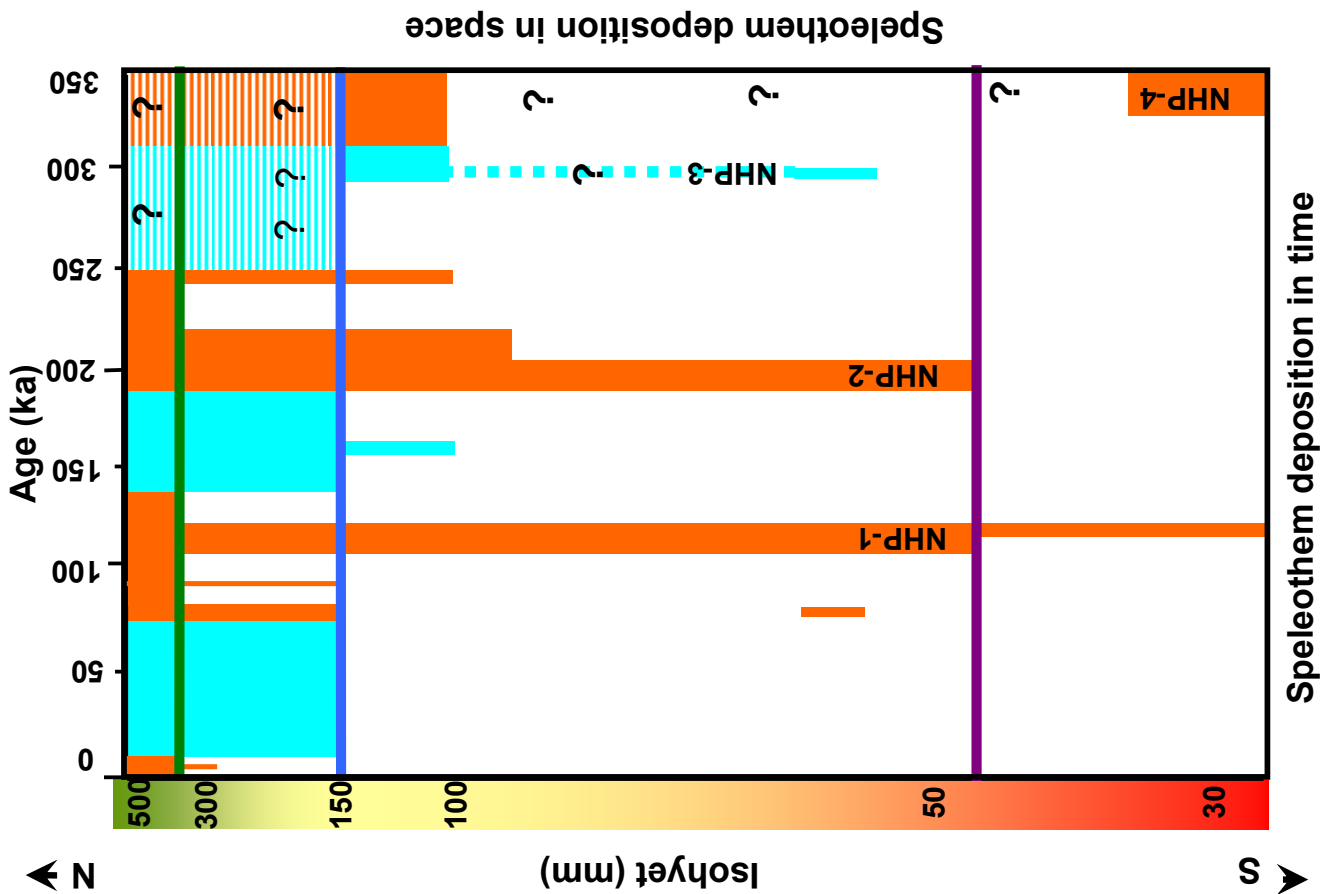
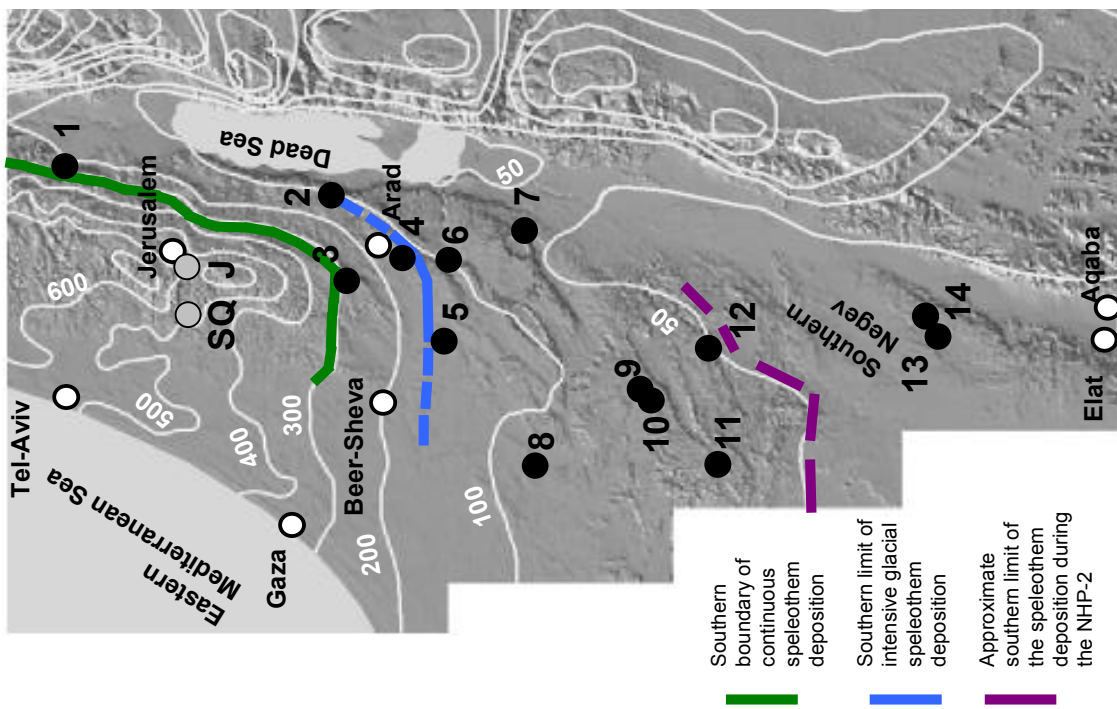


Figure 30: Comparison of the $\delta^{18}\text{O}$ profiles and age frequencies of central Negev speleothems with $\delta^{18}\text{O}$ values of Tzavoa, Soreq and Peqi'in speleothems. (A) $\delta^{18}\text{O}$ profiles of the central Negev speleothems (color) compared with those of Tzavoa Cave, northern Negev (grey) and Soreq and Peqi'in Caves in central and northern Israel (black). The rectangles in (A) indicate NHP-1 (left) and NHP-2 (right). (B-C) $\delta^{18}\text{O}$ profiles of central Negev speleothems (top) compared with $\delta^{18}\text{O}$ profiles of the Tzavoa, Soreq and Peqi'in caves (middle), and with relative frequencies of speleothem deposition in central and southern Negev (bottom) during NHP-1 (B) and NHP-2 (C). The green rectangles show the highest relative frequencies of speleothem ages.

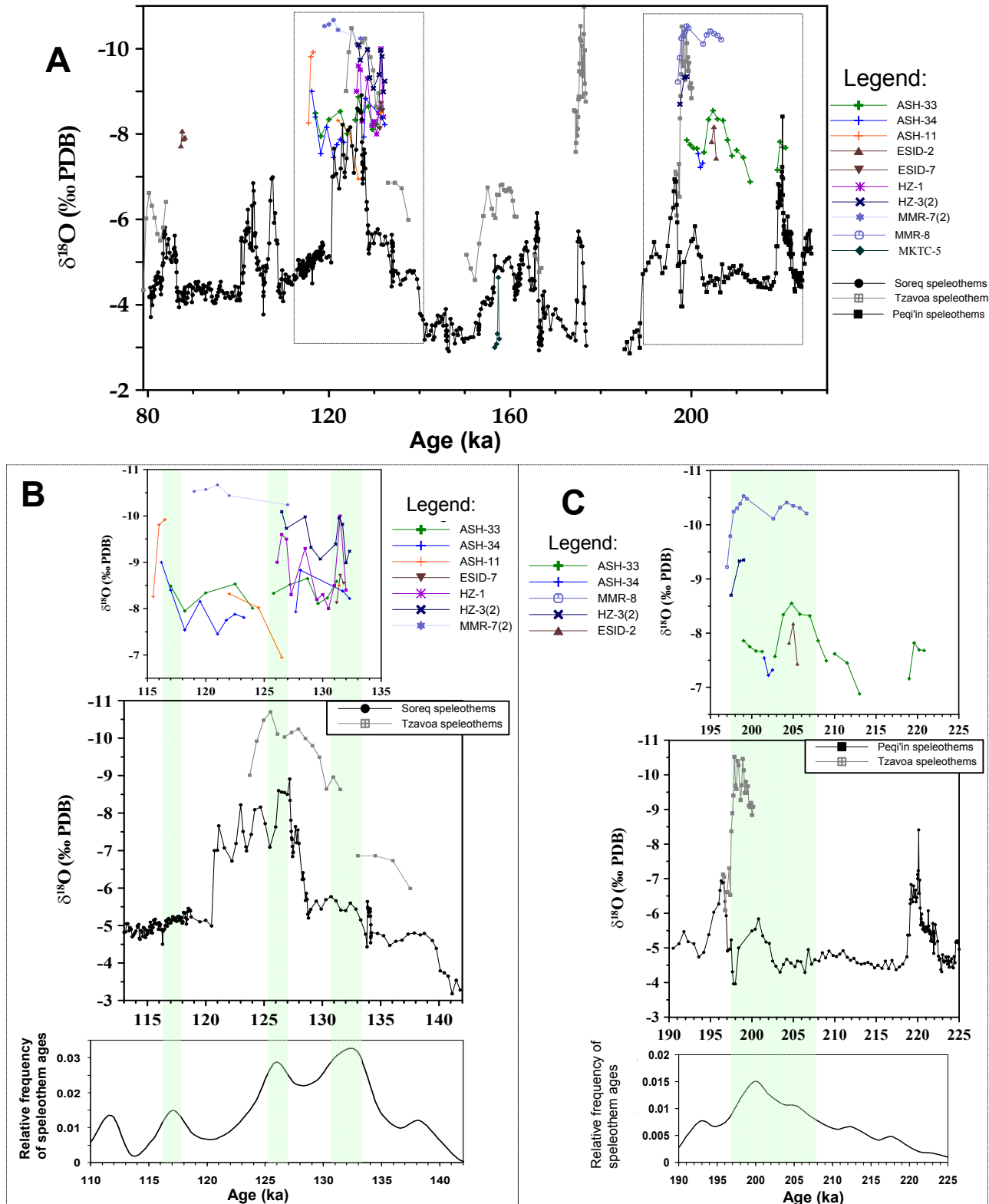


Figure 31: Comparison of the $\delta^{13}\text{C}$ profiles and age frequencies of central Negev speleothems with $\delta^{13}\text{C}$ values of Tzavoa, Soreq and Peqi'in speleothems. (A) $\delta^{13}\text{C}$ profiles of central and southern Negev caves (color) compared with those of Tzavoa Cave, northern Negev (grey) and Soreq and Peqi'in Caves in central and northern Israel (black). The rectangles in (A) show NHP-1 (left) and NHP-2 (right). (B-C) $\delta^{13}\text{C}$ profiles of central and southern Negev speleothems (top) compared with $\delta^{13}\text{C}$ profiles of the Tzavoa, Soreq and Peqi'in caves (middle), and with relative frequencies of speleothem deposition in central and southern Negev (bottom) during NHP-1 (B) and NHP-2 (C). The green rectangles show the highest relative frequencies of speleothem ages.

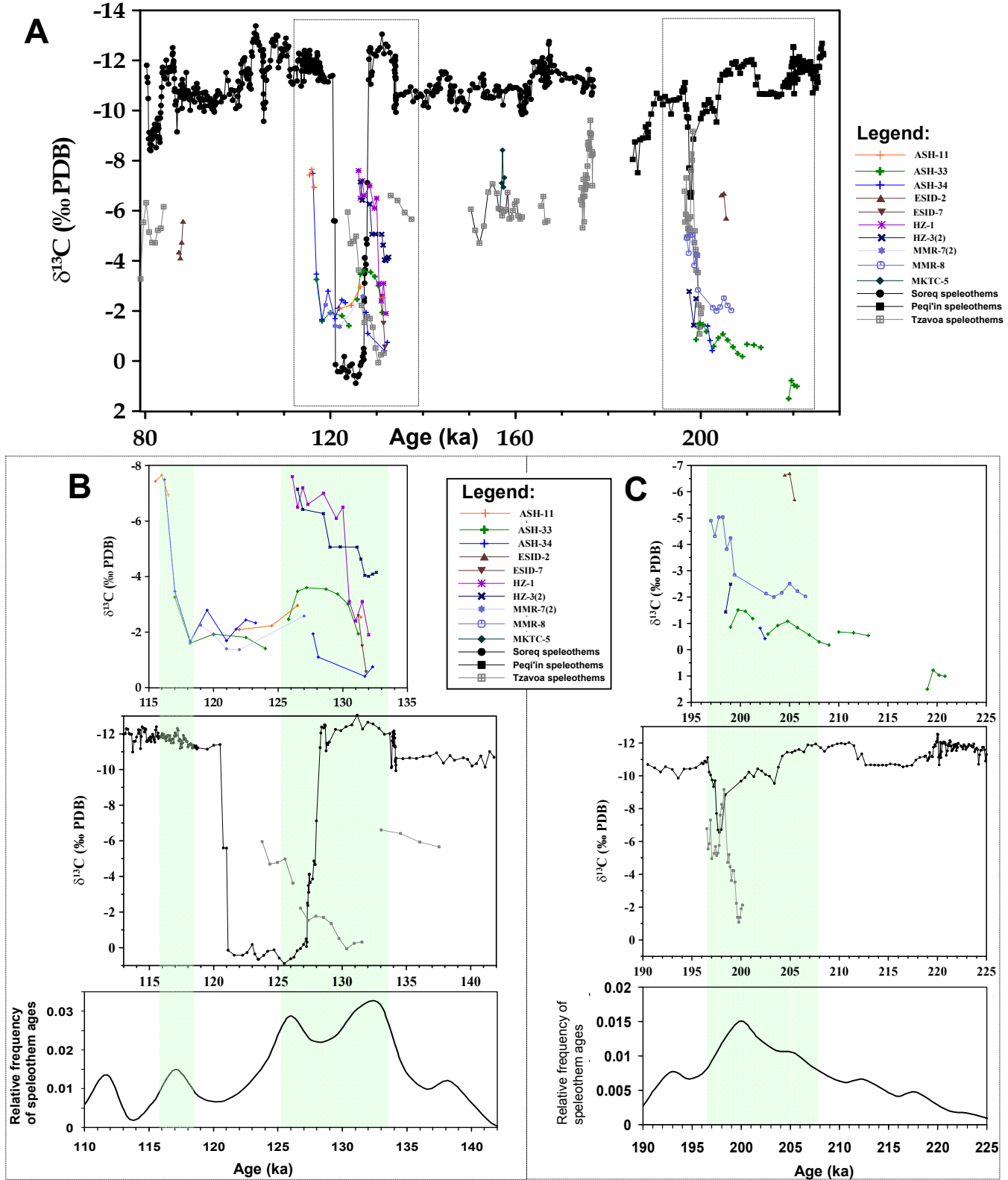


Figure 32: Correlation between the peaks of relative frequencies of the speleothem ages during the last 550 ka (B) and solar radiation energy in July at 65°N (Berger, 1978) (A), and in January at 65°S (Berger, 1978) (C). NHP-s are indicated by green rectangles and periods of lower relative frequency of speleothem ages are marked by the yellow rectangles. Periods of higher relative frequency of speleothem ages between 550 ka and 350 ka are indicated by grey rectangles.

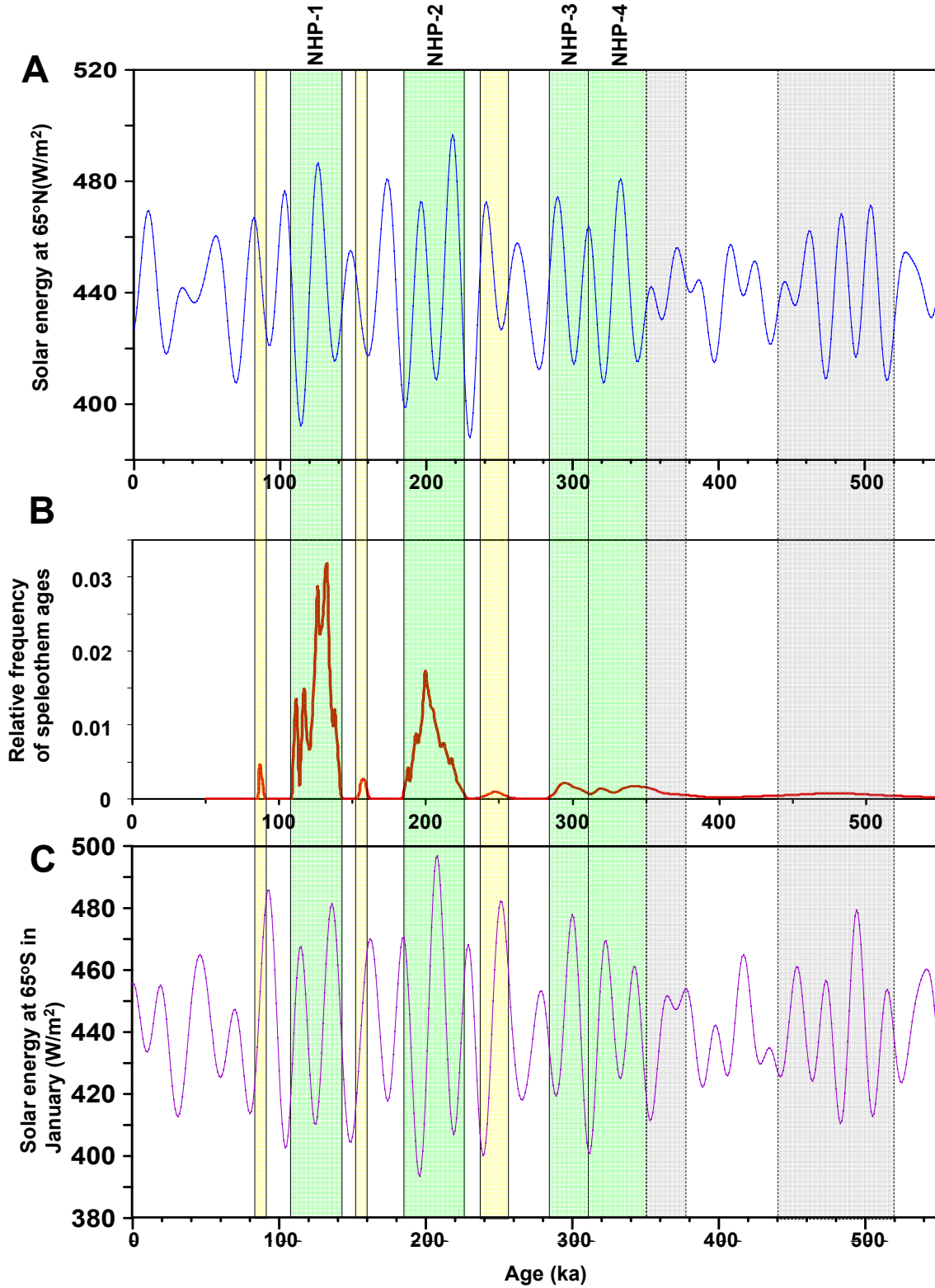


Figure 33: Relative frequencies of the speleothem ages compared with African monsoon index and timing of Mediterranean sapropels (Rossignol-Strick and Paterné, 1999): NHP-s (green rectangles), periods of limited speleothem deposition (yellow rectangles) during the last 350 ka, and periods of higher relative frequency of speleothem ages between 550 ka and 350 ka (grey rectangles). African monsoon index (M) in time (t) defined by Rossignol-Strick (1983) as: $M_t = I_T^t + (I_T^t - I_E^t)$, where I_T and I_E is insolation in units ($\text{cal} \cdot \text{cm}^{-2}$) $\cdot \text{day}^{-1}$ at Tropic of Cancer and Equator, respectively.

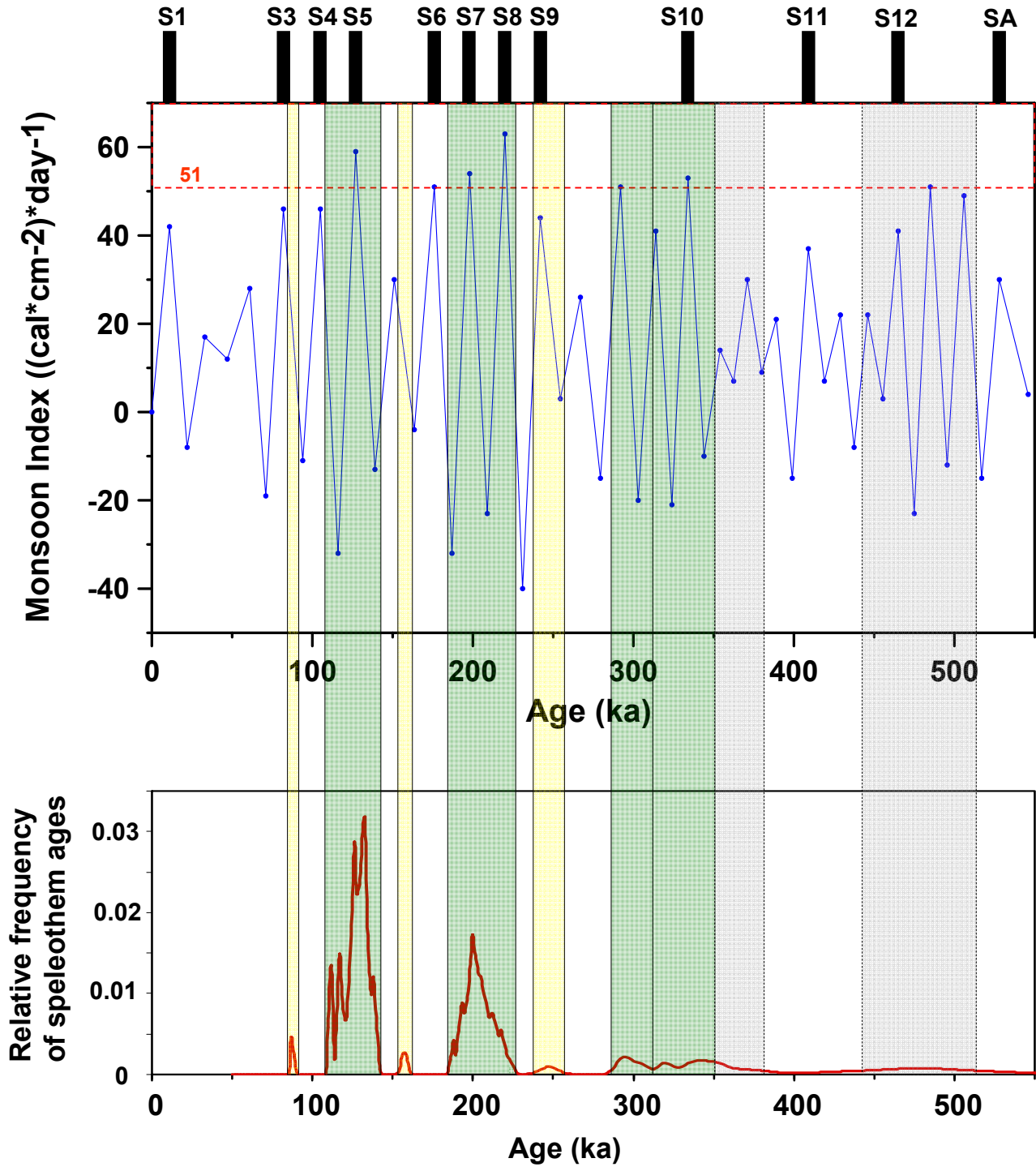


Figure 34: Correlation between the relative frequencies of the speleothem ages during the NHP-1 (A) and NHP-2 (B) and the: $\delta^{18}\text{O}$ profiles of Soreq and Peqi'in Caves (central and northern Israel), monsoon index, and parameters of solar radiation in July at 65°N , in December at 30°N and 60°S (Berger, 1978). Peaks of positive monsoon index are marked by red lines; green lines show peaks of negative monsoon index. Timing of the sapropels S5, S7 and S8 is shown by the black bars (Fontugne and Calvert, 1992) and grey bars (Emeis et al., 2003) at the top.

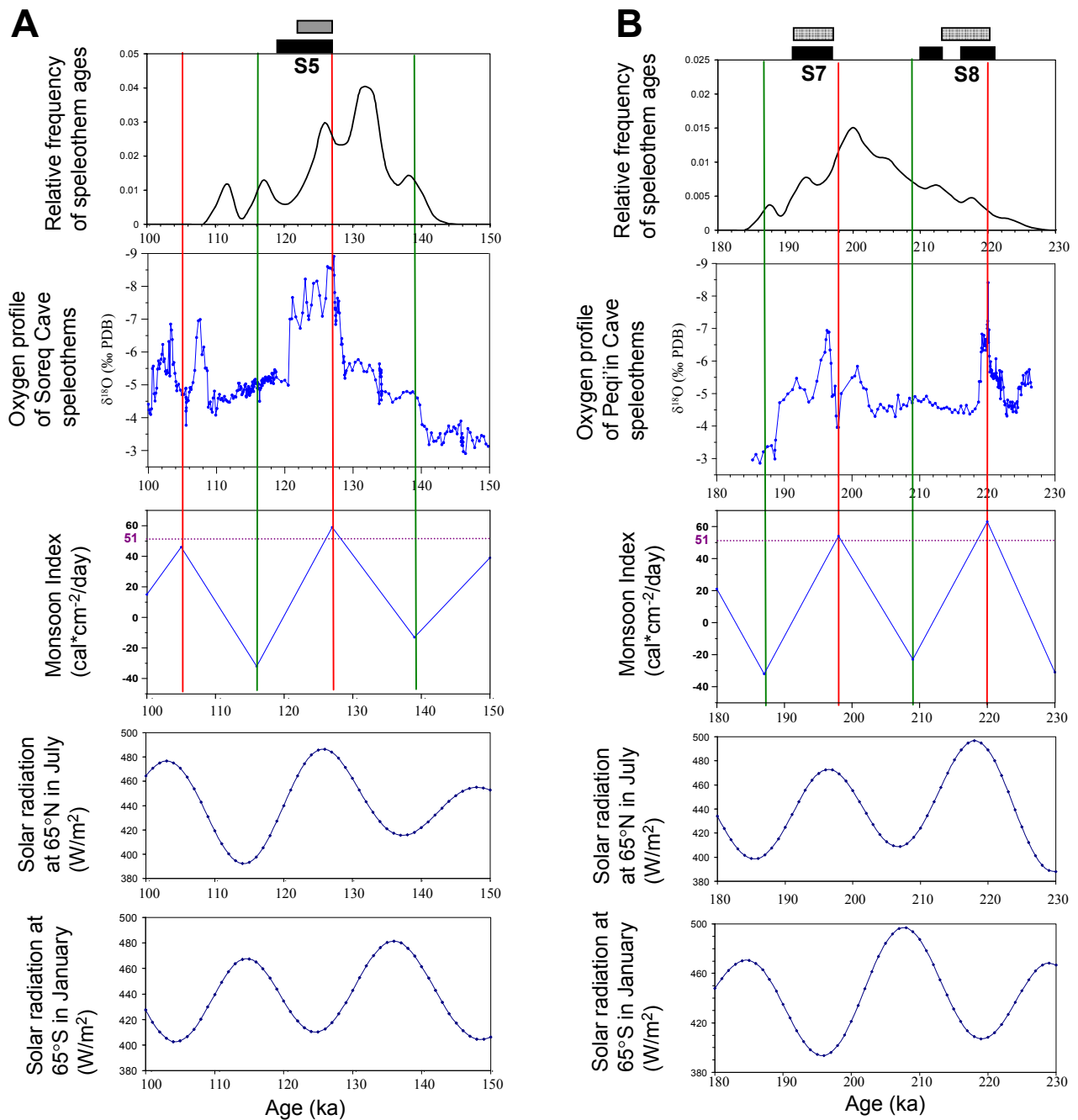


Figure 35: Correlation between the relative frequencies of the speleothem ages and sapropel events (black vertical bars at the top) during the period between 550 ka and 250 ka (A) with: African monsoon index values (B), solar radiation energy in July at 65°N (C), and in December at 60°S (D) (Berger, 1978). NHP-3 and 4 are indicated by green rectangles, periods of less intensive speleothem deposition by yellow rectangles; light yellow rectangles at the right mark higher frequencies of speleothem deposition during the period between 550 ka and 350 ka, where the age uncertainties are large.

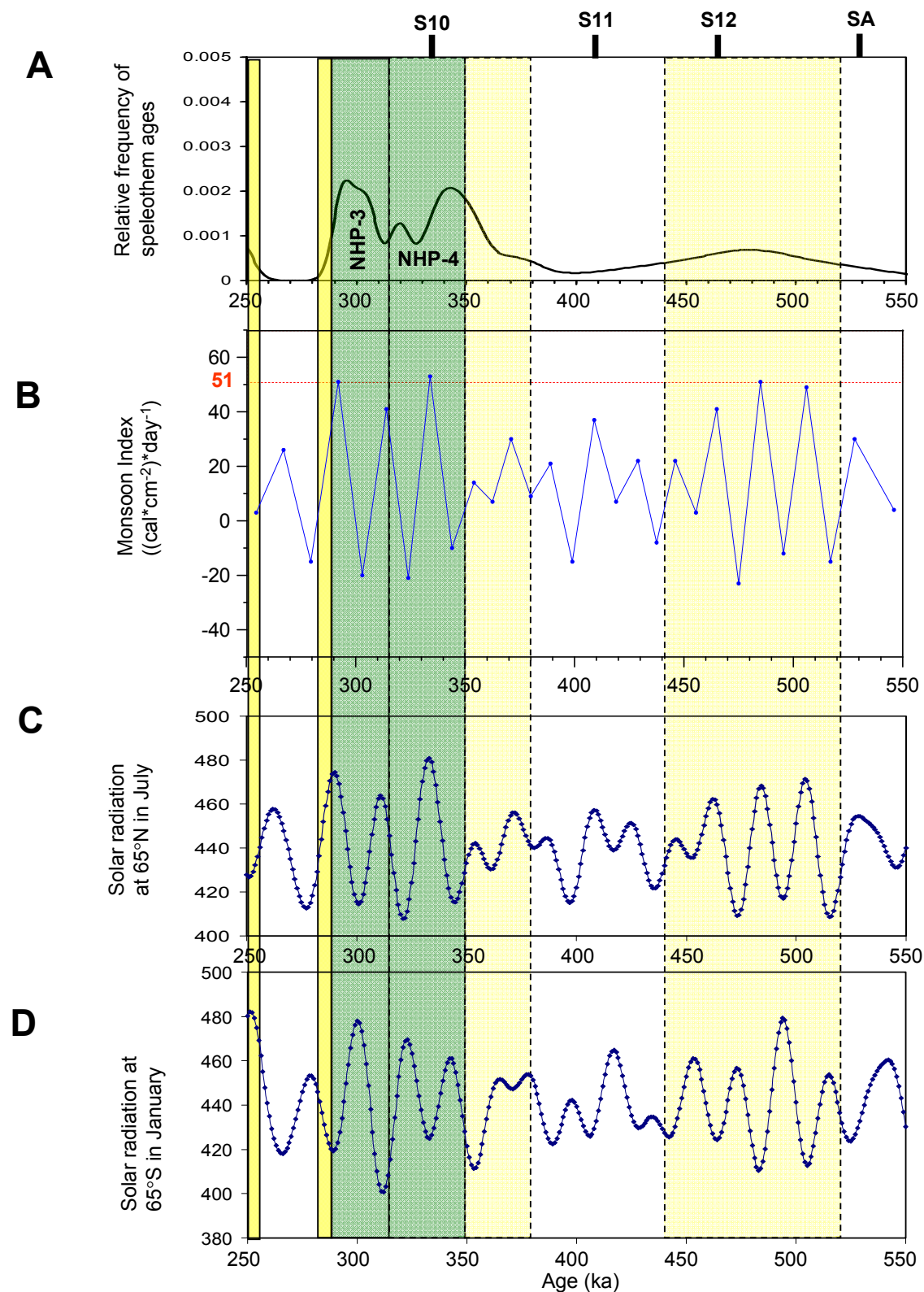


Figure 36: Location of the caves where Sr isotopic composition of speleothems was studied (black filled circles with labels given in Appendix 2 and Fig.1C) Soreq (SQ) and Jerusalem (J) caves, central Israel are indicated with open circles. Glacial (G) and interglacial/present-day (I) locations of the EM coastline are indicated in the figure, together with the glacial Lake Lisan, Sinai Peninsula and Nile Delta. The area exposed during the glacial sea low stands shown in light gray; Wind direction during the passage of the Cyprus cyclones is marked by arrows.

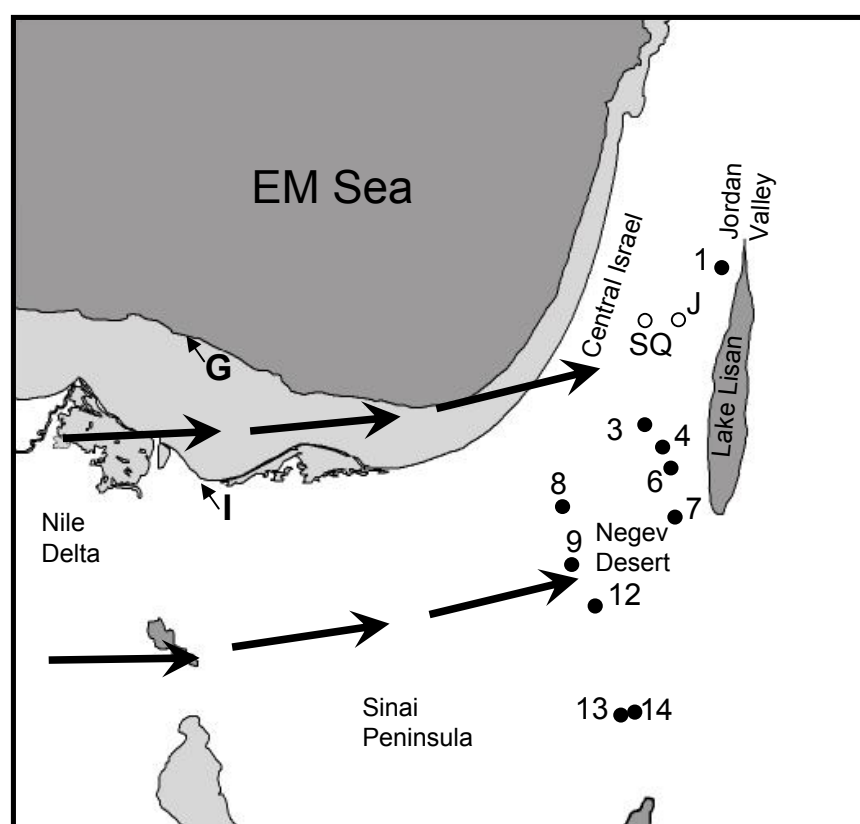


Figure 37: $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of Ma'ale-Efrayim Cave speleothems, Jordan Valley (A) and Negev Desert speleothems (B) compared with those of the speleothems of Soreq (Ayalon et al., 1999) and Jerusalem (Frumkin and Stein, 2004) caves.

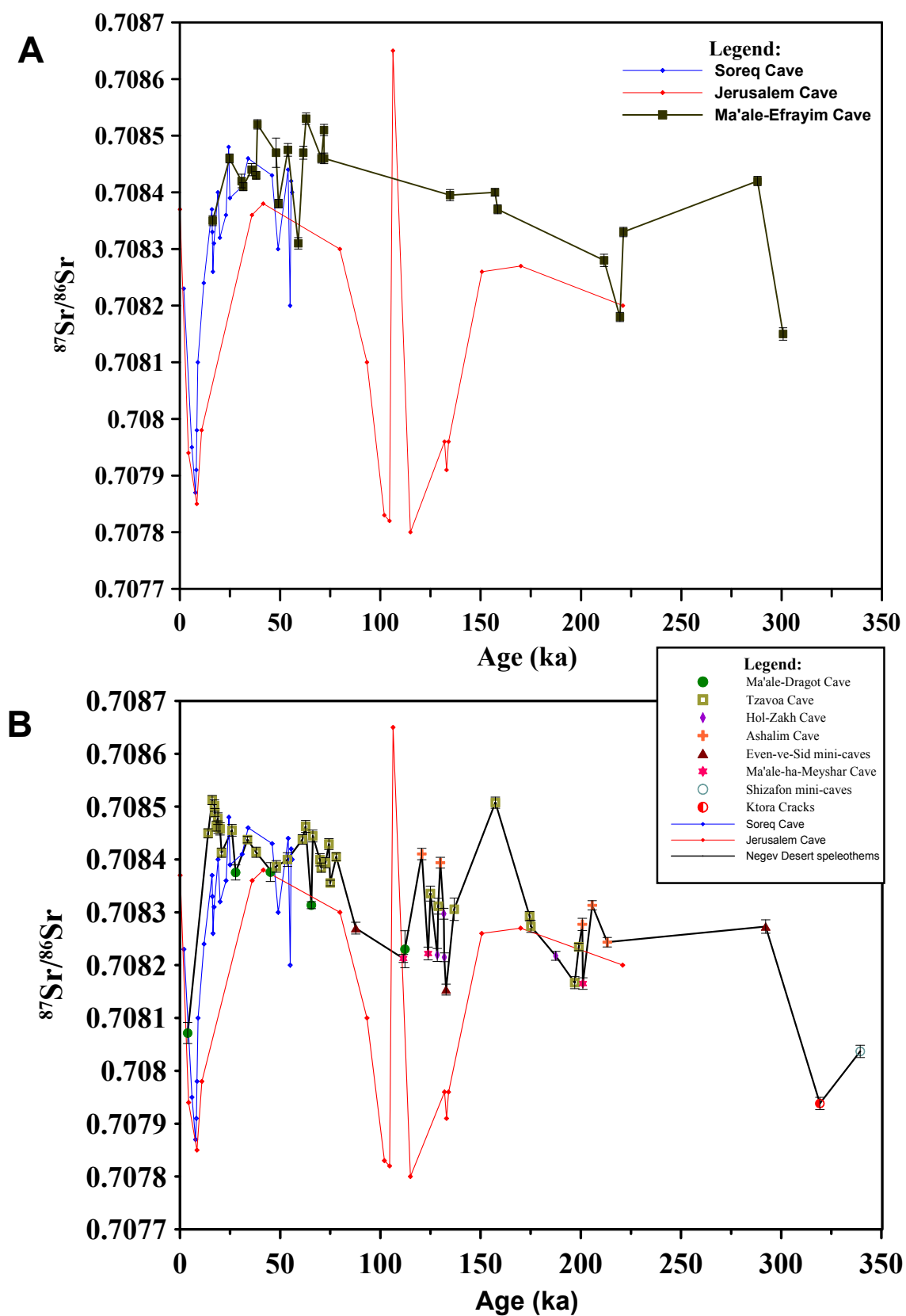


Figure 38: Comparison between Sr concentrations in central Israel caves (Soreq and Jerusalem) (Ayalon et al, 1999; Bar-Matthews et al., 2001; Frumkin and Stein, 2004) and caves of northern Negev (Tzavoa and Ma’ale-Dragot), Jordan Valley (Ma’ale-Efrayim) (A), and central and southern Negev (B).

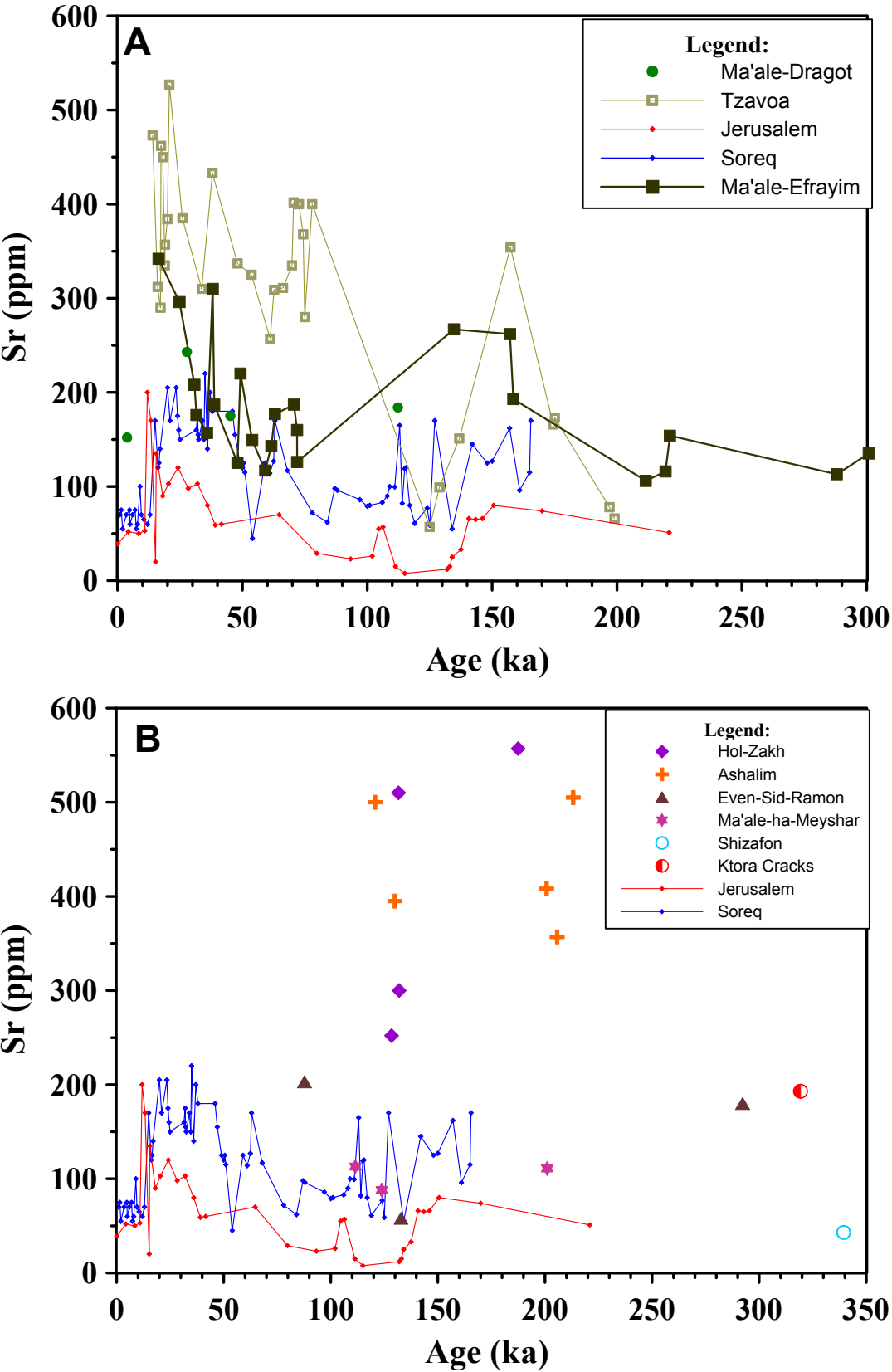


Figure 39: (A) $^{87}\text{Sr}/^{86}\text{Sr}$ vs $1/\text{Sr}$ plot in speleothems, host rocks and bulk soils. The bulk compositions of soils probably define a mixing line (orange) between the proximal source (0.7083–0.7085?) and Saharan dust (0.7165–0.72) (Krom et al., 1999), whereas speleothems define an apparent mixing line, possibly a result of dilution with rainwater (green); (B) $^{87}\text{Sr}/^{86}\text{Sr}$ vs $1/\text{Sr}$ plot for vadose and phreatic speleothems including the Basal member; (C) $^{87}\text{Sr}/^{86}\text{Sr}$ vs $1/\text{Sr}$ plot for vadose speleothems younger than 550 ka, as well as in Intermediate member. Enclosed data fields are for Tzavoa and Ma'ale-Efrayim Cave speleothems. Their Sr isotopic values define the composition of the proximal source during the glacial and interglacial periods, with a slight increase in host rock contribution and dilution evident during the interglacials.

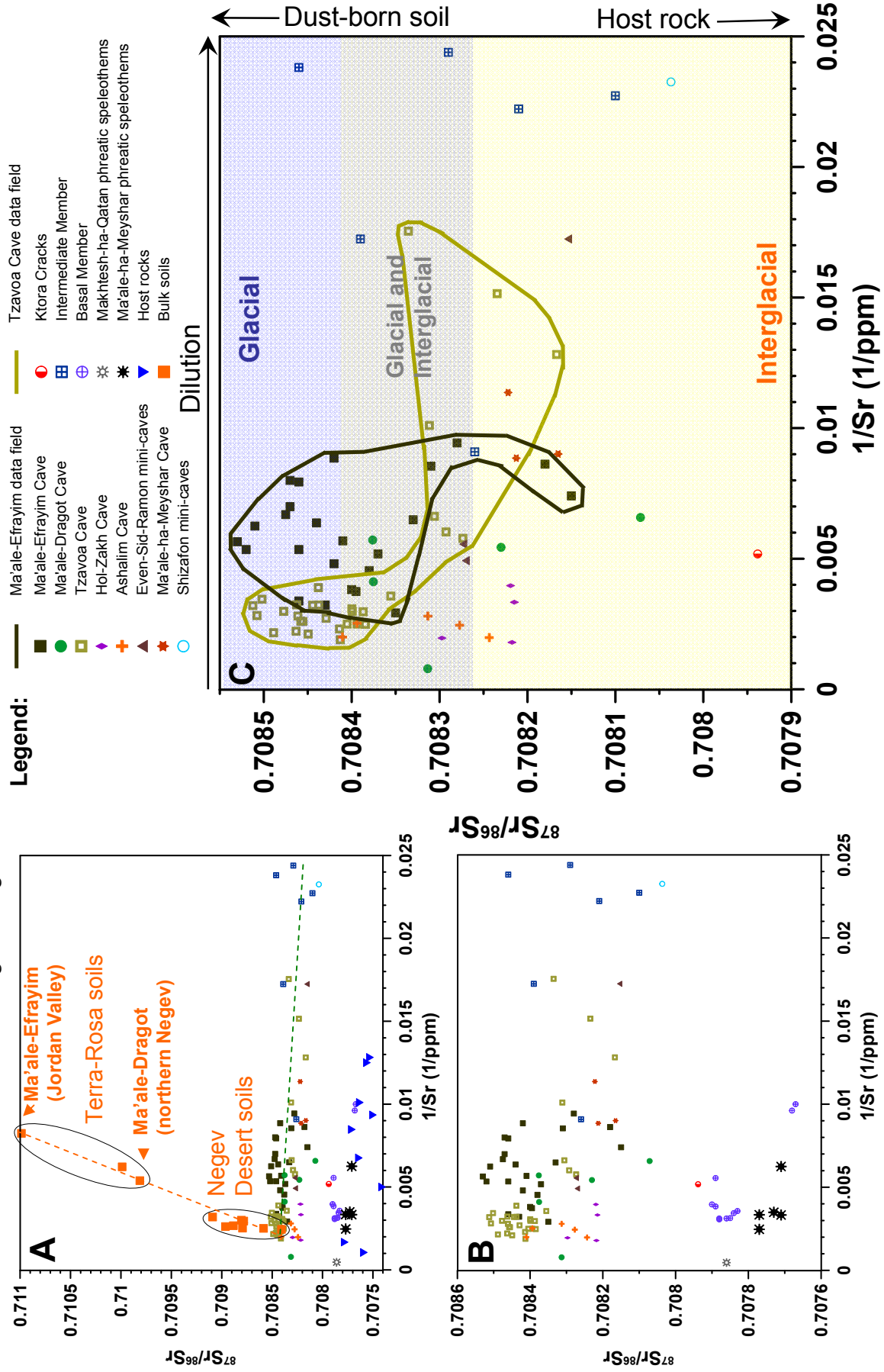
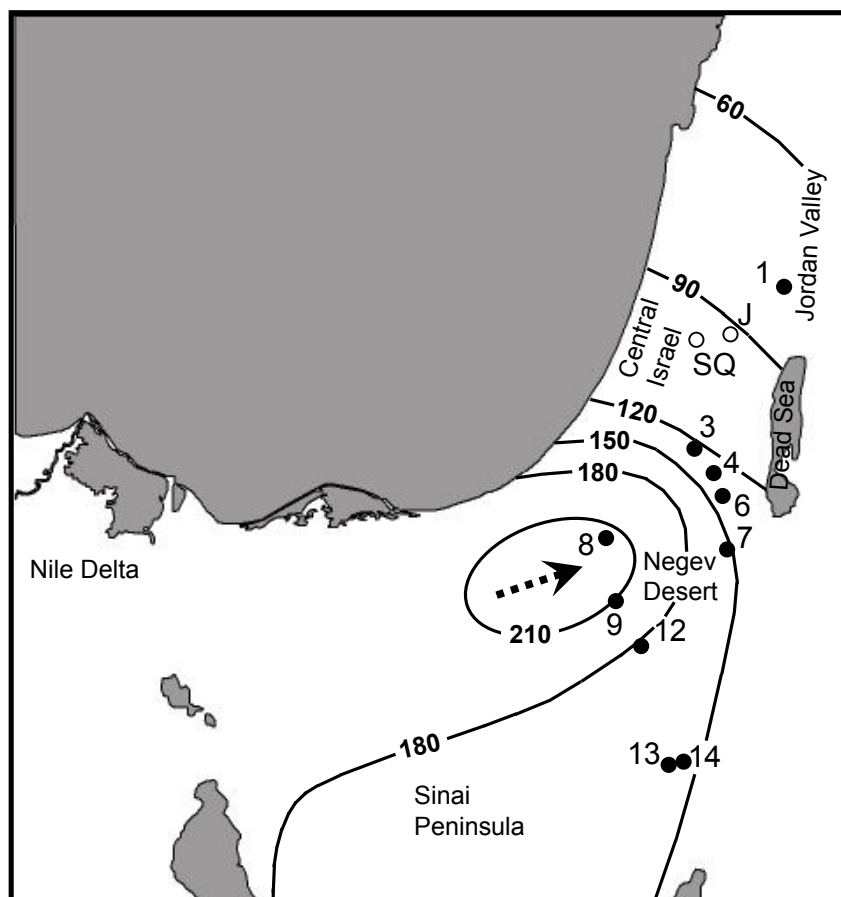


Figure 40: Annual dust fall (gr/m^2) map of Ganor and Foner (1996, 2001) and the location of the studied caves and the caves of central Israel. The marks and labels of the caves are as in Fig. 36. The wind direction during the passage of the Cyprus cyclones is marked by the arrow.



Appendix 2 – The studied caves

Cave positions were determined by global position (GP) system.

Ma'ale-Efrayim Cave (#1 in Fig. 1C) – GP: 32°03'05.81"N, 35°23'03.34"E. At an elevation of 250 m asl, located on the eastern slope of the Central Mountain Ridge (CMR) of Israel near the Jordan Valley, 60 km of the east from the Mediterranean Sea coast. The crest of the CMR is 900-1000 m asl and 10 km to the west of the cave, and receives 650-700 mm annual rainfall. The topographically depressed position of the cave relatively to westerly winds causes a deep "rain-shadow", and present day annual precipitation above the cave is 250-300 mm, with average yearly temperature of 21-22°C. The cave depth is 2-20 m below the surface. The cave was probably formed along NE-SW tectonic fissure in Cenomanian dolomite host rock (Fig. 2). The cave was discovered during road cutting and before it had no natural opening. At present, the lowest part of the cave traps cold air during the winter, whereas the temperature in the upper part is around the annual mean value. The speleothems inside the cave are vadose (formed in unsaturated zone) and include stalagmites (more than 2 m in height), stalactites and flowstone.

Kanaim Cave (#2 in Fig. 1C) – GP: 31°18'07.43"N, 35°18'23.42"E. At an elevation of 280 m asl, located on the eastern slope of the CMR in Judea Desert, 9 km NE of town of Arad and 8 km to the west of Dead Sea. The southern edge of CMR is 600-900 m asl and located 22 km to the west, receives 200-350 mm annual rainfall. The topographically depressed position of the cave relatively to westerly winds causes a "rain shadow" and the present day precipitation above the cave is ~100 mm, with average yearly temperature of ~21-22°C. The cave depth is 5-30 m below the surface. The cave system is hundreds of meters in length and contains gypsum speleothems, which locally reach thicknesses of a few cm on the cave walls. Vadose calcite speleothems are found in a few deep chambers >50 meters from the entrance and mainly include stalagmites and stalactites (up to tens of cm in length). The cave has one natural opening.

Ma'ale-Deragot Cave system (#3 in Fig. 1C) - GP: 31°18'3.09"N, 35°04'4.12"E. At an elevation of 630-720 m asl, located in northern Ma'ale-Dragot (Tel-Arad) Quarry. The caves are located on the southern edge of CMR of Israel a few km to the north from Be'er-Sheva Arad Valley and 65 km to the E-SE from the Mediterranean Sea. The cave

system is within Cenomanian and Turonian dolomite and limestone host rocks (Fig. 2) and the depth of the caves is 10-90 m below the surface. The mean annual present-day precipitation above the caves is 280-300 mm; and the mean annual temperature is ~17-18°C. Until the caves were discovered during quarrying, they had no natural openings. The stratigraphically oldest speleothems in lowest part of the caves are phreatic, composed of alternating calcite and hematite laminae. Most of the speleothems are vadose and includes stalagmites (more than 2 m in height), stalactites and flowstone.

Tzavoa cave (#4 in Fig. 1C) – GP: 31°12'3.24"N, 35°13'4.38"E. At an elevation of 550 m asl on the eastern flank of mountain ridge (~70 m below the ridge crest), about 4 km of the south from town of Arad, and 80 km E-SE from the Mediterranean Sea. The cave system is located in Turonian limestone host rock (Fig. 2), 20-50 m below the surface, within present-day 150-160 mm isohyets, with mean annual temperature of 18-19°C. The cave system is hundreds of meters in length and contains vadose speleothems: stalagmites (more than 1 m in height), stalactites and flowstone. Calcite veins tens of cm in width are found in the host rock. Field relations between the cave and the veins show that these veins are older than the cave space. The cave has several natural openings.

Izzim Cave (#5 in Fig. 1C) – GP: 31°08'2.24"N, 35°03'5.11"E (limestone quarry). At an elevation of ~500 m above the sea level, 72 km SE from the Mediterranean Sea coast. The cave is a few tens of meters in length, and located in Turonian limestone host rock (Fig. 2), 10-15 m below the surface. Before it was discovered by quarrying, the cave had no natural openings. The mean annual rainfall above the cave is of 120-140 mm and the mean annual temperature is ~18-19°C. The cave contains vadose calcite speleothems: stalagmites (of up to 2 m in height), stalactites and flowstone. Numerous gypsum speleothems are also found in the cave, some of them formed on the surface of the calcite speleothems.

Hol-Zakh Cave (#6 in Fig. 1C) – GP: 31°09'2.24"N, 35°12'0.37"E (limestone quarry). At an elevation 150 m asl. This is a small cave located within Turonian limestone host rock (Fig. 2), at depth of 15-20 m below the surface, 84 km south-east from the coast of eastern Mediterranean Sea, between present-day 80-100 mm isohyets, with mean annual temperature of 21-22°C. Although the cave located in the same latitude of the Izzim Cave, the rainfall amount is smaller because of its location in the "rain shadow" of the

mountain ridge to the west. The cave was exposed during quarrying, has no natural openings, and contains vadose speleothems: stalactites and flowstone crusts of maximum length/thickness of ~3 cm.

Ashalim Cave (#7 in Fig. 1C) – GP: 30°56'3.62"N, 34°44'2.25"E. At an elevation of ~400 m asl, located ~2 km SE to Ashalim village and 67 km SE from the coast of eastern Mediterranean Sea. The cave system is hundreds of meters long, located in Turonian limestone host rock (Fig. 2) at depths of 0-25 m below the surface; between present-day 120 mm and 100 mm isohyets, with mean annual temperature of 19-20°C. The cave has one natural opening, a few meters in diameter with a funnel-like shape. Most of the speleothems are vadose: stalagmites, stalactites and flowstone with maximal thickness of 30-40 cm. In lowest parts of the cave some phreatic speleothems cover the cave walls, and "stalactites" of bat guano were found.

Makhtesh-ha-Qatan Cave (#8 in Fig. 1C) – GP: 30°57'0.53"N, 35°13'1.16"E. Elevation of ~20 m bsl. This is small cave located ~4 m above the Wadi Hazera (ephemeral stream) channel, in the outlet of Makhtesh-ha-Qatan (erosional crater), on the eastern flanks of mountain ridge rising ~550 m asl, and 99 km south-east from the coast of eastern Mediterranean Sea. The precipitation on the ridge crest west of the cave is ~100 mm, but the depressed position of the cave relatively to westerly winds creates a "rain shadow" effect, and the annual precipitation above the cave is 50-70 mm with mean annual temperature of ~23°C. The cave located within Cenomanian dolomite host rock (Fig. 2), at depth of 0-4 m below the surface; it has one natural opening and its length is several meters. Today this cave is intermittently flooded by Wadi Hazera, and sandy flood sediments cover the cave floor. In the past the cave was probably below the channel of Wadi-Hazera and was flooded during the high groundwater levels. The cave contains phreatic calcite speleothems ~20 cm thick, alternating with thin (<1 cm) laminae of vadose speleothems. Speleothems are composed of calcite and thin iron oxide laminae. Gypsum crystals are also found on the surface of the calcite speleothems.

Even-Sid-Ramon mini-caves (#9 in Fig. 1C) – GP: 30°38'2.28"N, 34°48'3.57"E. At an elevation of ~800 m asl, located in a quarry 2 km to the north of the town of Mitzpe-Ramon, 96 km south-east from the coast of the eastern Mediterranean Sea. The precipitation is 90-100 mm, with mean annual temperature of ~17°C. Small caverns and

fissures in Cenomanian limestone host rock (Fig. 2), are located 2-6 m below the surface and were exposed during the quarrying. Before it they had no natural openings. They contain mainly vadose speleothems: stalactites, flowstone and pool deposits with maximum thickness of ~8 cm. Some of the caves contain red clay sediments, probably paleosols washed into the caves from the surface.

Mitzpe-Ramon Cave – (#10 in Fig. 1C) – GP: 30°36'3.54"N, 34°47'3.99"E. At an elevation of ~850 m asl. The cave is located in the western limit of the town of Mitzpe-Ramon, 95 km from the coast of the eastern Mediterranean Sea. The cave length is a few tens of meters, and it is located at a depth of 0-20 m below the surface in Cenomanian dolomite host rock (Fig. 2). Present day annual rainfall is 90-100 mm with mean annual temperature of 16-17°C. The cave has one natural opening. It contains vadose speleothems, mainly flowstone and stalactites up to 20 cm thick.

Wadi-Lotz Cave – (#11 in Fig. 1C) – GP: 30°28'1.22"N, 34°34'3.85"E. At an elevation of ~900 m asl. The cave is located ~2 km from Israeli-Egyptian border, ~100 km from the coast of the EM Sea. The rainfall above the cave is ~100 mm, and the mean annual temperature is 16-17°C. The cave is located within Eocene limestone host rock (Fig. 2), inside a steep cliff. The cave has the shape of an overturned funnel 15 m in depth, starting with natural opening in its top (3-4 m in diameter), and widening towards its bottom, forming the large chamber with the major natural entrance of ~10 m in diameter. Most of the speleothems are vadose flowstone up to 15 cm thick, and either covers the walls in the upper part of the shaft, or are exposed on the surface above the upper opening and partly eroded. The cave floor is covered by thick layer of bird guano.

Ma'ale-ha-Meyshar Cave (#12 in Fig. 1C) – GP: 30°29'3.56"N, 34°55'5.47"E (near road #40), at an elevation of 450 m asl. The cave is located 115 km S-SE from Mediterranean Sea coast, on 50 mm isohyet, with mean annual temperature ~19°C. The major cave was partly destroyed by road cutting. The remaining chamber is 10-12 m in length, 5-12 m below the surface, and it is a part of cave system composed of isolated small caves containing phreatic speleothems 1-25 cm thick, covered by thin vadose flowstone, stalactites and stalagmites up to 3 cm thick. Reddish and beige clay sediments are found inside the caves (probably paleosols washed to the caves). The host rock is

Cenomanian dolomite (Fig. 2). The cave had no natural opening before it was exposed by road cutting.

Shizafon Quarry mini-caves (#13 in Fig. 1C) – GP: 30°02'1.83"N, 35°00'1.13"E (limestone quarry), at an elevation of ~400 m asl. The site is located 165 km S-SE off the coast of eastern Mediterranean Sea, between present-day 40 mm to 20 mm isohyets, with the mean annual temperature is 20-21°C. Small vugs, caverns and cracks containing vadose flowstone 0.5-5 cm thick occur in Upper-Turonian limestone host rock (Fig. 2). The mini-caves are located at a depth of ~10 m below the surface and had no natural openings before exposed.

Ktora Cracks (#14 in Fig. 1C) – GP: 30°01'5.68"N, 35°03'4.93"E; at an elevation of ~400 m asl. The site is located ~55 km north of town of Elat, 168 km S-SE from the coast of eastern Mediterranean Sea, between present-day 40 mm to 20 mm isohyets, with the mean annual temperature is 20-21°C. This is an exposed wall of a collapsed cave covered by vadose flowstone 5-20 cm thick. Some of the speleothems are covered by rock fall and talus. This collapsed cave, and other caves in this site were formed within large fissures in Upper Turonian limestone host rock (Fig. 2) which developed as a result of sliding of large blocks of rock down slope (Yacoby, 2004). The walls collapsed in many of the cracks clogging the upper parts of the cracks with breccias formed the cave ceiling. The direction of the crack forming the cave is 230° (SW-NE).

Reference:

Yacoby, Y. (2004). The open fractures in Ma'ale-Grofit area. *MSc Thesis (in Hebrew, English abstract), Ben Gurion University of the Negev, Be'er-Sheva, Israel.*

Appendix 3: Analytical procedures of U-Th and U-Pb dating methods

A.3.1. U-Th dating: chemical separation between U and Th (Kuperman, 2005)

1) Depending on the uranium concentration (Table 1), 10-800 mg calcite powder was dissolved in 7N HNO₃.

2) The sample solutions were centrifuged 10 minutes in 4000 rpm to deposit the insoluble residue (most of it was silicates).

3) The aliquots were separated from the insoluble residues.

4) The insoluble residues were dissolved using a 5 ml mixture of concentrated HF and HNO₃ and evaporated to dryness by heating in Teflon beaker. Silica was removed by reaction with HF yielding gaseous SiF₄. The residues were dissolved in 7M HNO₃ and added to the sample aliquots.

5) ²³⁶U-²²⁹Th spike was added to the solution and mixed.

6) The solution was evaporated to dryness in a Teflon beaker and then dissolved in 5 ml 7M HNO₃.

7) 2ml Bio-Rad AG 1X8 200-400 mesh resin was added into mini-columns and cleaned by passing through 2 aliquots of 5 ml 7M HNO₃ and three aliquots of double-distilled H₂O.

8) The resin on columns was pre-conditioned to the sample solution by passing through 5 ml of 7M HNO₃.

9) The sample solution was loaded onto mini-columns.

10) The matrix was removed by passing through 2 aliquots of 2 ml 7M HNO₃.

11) The resin on columns was pre-conditioned for Th collection by the addition of 1.5 ml HCL 6M.

12) Th was eluted into Teflon beakers by passing 3.5 M HCl 6M through the column.

13) The resin on columns was pre-conditioned for U collection by adding of 1.5 ml of 1M HBr.

- 14) U was eluted to Teflon beakers by addition of 3.5 ml 1N HBr to the columns.
- 15) Evaporation of the U and Th solutions to dryness.
- 16) Dissolution of the residues in 2 ml (U) and 5 ml (Th) of 0.1M HNO₃.
- 17) 50 µl of U solution were dissolved in 2.5 ml 0.1M HNO₃ (diluted by factor of 50) and the original U concentration in solution was measured using Perkin Elmer Sciex ICP-MS Elan 6000.
- 18) Samples with U concentration >50 ppb were diluted to 50 ppb to adjust the concentration to MC-ICP-MS Faraday Cups working conditions.
- 19) Addition of 25 µl of UNBL 112A standard containing 50 ppb U to Th solutions and mixing.

A.3.2. U-Th dating: measurement of U and Th isotopic ratios by MC-ICP-MS (Kuperman, 2005)

High precision U-Th dating was performed using a Nu Instruments MC-ICP-MS, equipped with 12 Faraday cups and 3 ion counters. The sample was introduced to the MC-ICP-MS through an Aridus[®] micro-concentric desolvating nebuliser sample introduction system. The solution injected into an Ar plasma at ~6000K, the various elements are ionized and their beam is accelerated in electric field. The beam shape is smoothed by a system of high voltage lenses. The isotopic species of different masses are separated by magnet and are counted by the collector system of 12 Faraday cups (H6, H5, H4, H3, H2, H1, Ax, L1, L2, L3, L4 and L5) and 3 ion counters (IC) (IC0, IC1 and IC2). The difference between the Faraday cups and the ion counters is in their sensitivity. Ion counters are sensitive to ion beams with low concentration, giving weak electric signal (²³⁴U, ²³⁰Th, ²³⁶U, ²²⁹Th) whereas the Faraday cups are built to measure ion beams with high concentration giving strong electric signal (i.e. ²³⁸U, ²³⁵U).

U measurement were made in 3 cycles in which ion counters were calibrated relatively to Faraday cups and three atomic ratios ²³⁵U/²³⁸U, ²³⁴U/²³⁶U, and ²³⁵U/²³⁶U were determined. During the first cycle the instrumental mass bias was corrected

(using the exponential equation) by measuring the $^{235}\text{U}/^{238}\text{U}$ atomic ratio and correcting with the natural $^{235}\text{U}/^{238}\text{U}$ ratio (0.0072), and the Faraday cups were calibrated to ion counter IC0. During the second cycle, ^{235}U and ^{238}U were measured by the Faraday cups and then were calibrated to ion counter IC1. During the third cycle ^{235}U and ^{238}U were measured by the Faraday cups, and ^{236}U and ^{234}U were measured by ion counters, IC0 and IC1, respectively.

Determination of $^{230}\text{Th}/^{229}\text{Th}$ and $^{232}\text{Th}/^{229}\text{Th}$ was performed by calibration to the isotopic ratios of uranium, because the collector configuration of MC-ICP-MS does not enable the direct measurement of ^{230}Th . The measurement was performed in three cycles. During the first cycle ^{235}U and ^{238}U were measured by the Faraday cups and then they were calibrated to ion counter IC1, by which ^{229}Th was measured. In second cycle ^{235}U was measured by IC1 and ^{238}U was measured by Faraday cups. During the third cycle, ^{235}U and ^{238}U were measured by the Faraday cups, and then they were calibrated to ion counter IC1 by which ^{230}Th was measured.

The age determination was possible due to the accurate determination of ^{234}U and ^{230}Th concentrations by isotope dilution analysis using the ^{236}U - ^{229}Th spike: $^{234}\text{U}/^{236}\text{U}$, $^{235}\text{U}/^{236}\text{U}$, $^{235}\text{U}/^{238}\text{U}$, $^{230}\text{Th}/^{229}\text{Th}$ and $^{232}\text{Th}/^{229}\text{Th}$ ratios were determined and the known ratio of $^{236}\text{U}/^{229}\text{Th}$ in spike (2000) enabled calculation of $^{230}\text{Th}/^{234}\text{U}$ ratio. $^{234}\text{U}/^{238}\text{U}$ ratio was calculated with reproducibility of 0.11% (2σ).

A3.3 U-Pb dating method (Woodhead et al., 2006).

A3.3.1 Chemical purification and extraction of U and Pb

Five speleothem laminae ASH-15-C, ASH-15-D, ASH-15-E, KTO(1)-1-J and KTO(1)-1-K were sawn into slabs ~5-10 mm thick and 10-30 mm in length. From each lamina 4-5 sub-samples of 70-280 mg were cut, providing between 100 pg and 1.2 ng of Pb, depending on Pb content. After extraction from the speleothem, sub surfaces were cleaned by repeated ultrasonification in 0.5 M HCl, followed by washing in ultra-pure water. Samples were then dried and weighted. All material handling subsequent to cutting was conducted in a multiple-HEPA-filtered

environment, and all acids utilized in the dissolution and chemical extraction of Pb were (as a minimum) twice distilled in either quartz or Teflon stills. Following Woodhead and Hergt (1997), samples were placed in new, cleaned 30 ml TPX[®] polymethyl-pentene beakers with ~1 ml of ultrapure water, and dissolved by dropwise addition of 6M HCl. After complete dissolution (no inorganic residues were observed in any of the samples), sample solutions were aliquoted for spiking, using ²³³U-²⁰⁵Pb tracer. Pb and U were purified using conventional anion exchange chemistry with HBr/HCl media (employing a single column pass only to reduce blank contributions), and EICHRON TRU-resin with HNO₃/HCl/HF media respectively (Luo et al., 1997; Manhès et al., 1978). Total processing blanks were 5-15 pg Pb and < 5 pg U; a value of 10±5 pg has been used for the former in the blank subtraction calculation.

A3.3.2 Measurement using MC-ICP-MS

Isotopic analyses were carried on a Nu Plasma MC-ICP-MS coupled to a Cetac Aridus desolvation system with a Glass Expansion OpalMist nebulizer (uptake rate ~40 µl/min). At a typical sensitivity of 100-120V/ppm, Pb samples, dissolved in 0.5 ml of 0.3 M HNO₃, yielded total Pb signals upwards of 4×10⁻¹³A (40mV) for a 200 pg sample size, allowing ²⁰⁵, ²⁰⁶, ²⁰⁷, ²⁰⁸Pb to be measured on Faraday cups. The correction for mass bias was performed by reference to external standards. For radiogenic samples (samples without non radiogenic Pb) the error attributable to the mass bias correction was only a minor component of the overall analytical uncertainty.

U isotope dilution analyses were performed using the natural ²³⁵U/²³⁸U ratio in the sample for internal bias correction, after correction for a minor ²³⁵U contribution from the ²³³U spike. The ²³⁴U/²³⁸U of the samples was measured in Geological Survey of Israel using methods described in Vaks et al. (2006). Initial ²³⁴U/²³⁸U activity ratios were calculated using the formula $[\frac{^{234}\text{U}}{^{238}\text{U}}]_0 = 1 + (([\frac{^{234}\text{U}}{^{238}\text{U}}]_s - 1) \times \exp(\lambda^{234}t))$ (Kaufman et al., 1998), where square brackets indicate the activity ratio, subscripts 0 = initial, s = measured in sample, t = time in years, and λ^{234} is the decay constant for ²³⁴U, for which value of 2.835×10⁻⁶ was used. Blank correction and preliminary

isochron construction (prior to disequilibrium correction) were performed using the data reduction software PBDAT (Ludwig, 1993) and Isoplot Ex. 3.04 (Ludwig, 2001).

A3.3.3 Isochrone formalization

Use of the isochron constructions including ^{204}Pb and ^{208}Pb data was avoided because of the interference of the ^{204}Pb mass by ^{204}Hg , and because the concentrations of ^{208}Pb in highly radiogenic speleothem samples ($^{206}\text{Pb}/^{204}\text{Pb} > 500$) are vanishingly small. Isochron construction instead used the Terra-Wasserburg (T-W) diagram (Terra and Wasserburg, 1972), renamed the “semi-total Pb-U isochron” by Ludwig (1998). The T-W diagram uses parameters ^{238}U , ^{207}Pb and ^{206}Pb , which are well determined on the Faraday cups. In the case of a suite of coeval, cogenetic speleothem sub-samples with differing $^{206}\text{Pb}/^{204}\text{Pb}$, plots of $^{238}\text{U}/^{206}\text{Pb}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ data uncorrected for both blank and common Pb should define a mixing line between common Pb (a combination of blank and intrinsic common Pb) and pure radiogenic Pb located on the T-W Concordia (although in the case of speleothems the latter curve is influenced by initial isotopic disequilibrium effects, see below). Plots of isotope data determined on different sub-samples from the same lamina show little systematic difference in the resulting ages when calculated either with or without blank correction, although usually age errors are reduced for blank corrected data. Therefore the data were blank corrected before plotting and calculating T-W disequilibrium Concordia intercepts with Isoplot Ex (Ludwig, 2001).

A3.3.4 Correction for U series disequilibrium

Speleothems precipitate out of U-series equilibrium and $^{234}\text{U}/^{238}\text{U}$ excess in cave seepage waters frequently leading to wide range of $^{234}\text{U}/^{238}\text{U}$ values in speleothems. In addition, the time required for ^{230}Th and ^{231}Pa ingrowth (clean speleothems precipitate with essentially no Th and Pa) in the ^{238}U - ^{206}Pb and ^{235}U - ^{207}Pb decay chains must also be taken into account. The correction for initial ^{234}U and ^{230}Th was achieved using the Isoplot spreadsheet functions of Ludwig (2001) incorporated into

an 'in house' iterative routine. The present day $^{234}\text{U}/^{238}\text{U}$ activity ratios of all five speleothems are very close to equilibrium. For estimation of $^{234}\text{U}/^{238}\text{U}$ activity ratios in flowstone samples KTO(1)-1-J and K, the initial $^{234}\text{U}/^{238}\text{U}=1.06$ activity ratio of the youngest lamina A1 of this flowstone (dated by U-Th to 137.7 ± 1.5 ka, Table 1) have been used. In ASH-15-C, D and E laminae the average initial $^{234}\text{U}/^{238}\text{U}=1.05$ activity ratio for the young speleothems from this cave have been used.

References of Appendix 3:

- Kaufman, A., Wasserburg, G. J., Porcelli, D., Bar-Matthews, M., Ayalon, A., and Halicz, L. (1998). U-Th isotope systematics from the Soreq cave, Israel and climatic correlations. *Earth and Planetary Science Letters* **156**, 141-155.
- Kuperman, G. (2005). "Reconstruction of flood periods by dating speleothems in Nahal Hazera, northern Negev." Hebrew University of Jerusalem, M.Sc. Thesis (in Hebrew, English Abstract).
- Ludwig, K. R. (1993). PBDAT: a computer program for processing Pb-U-Th isotope data. In "U.S.G.S. Open File Report 88-542." pp. 33.
- Ludwig, K. R. (1998). On the treatment of concordant uranium-lead ages. *Geochimica et Cosmochimica Acta* **62**, 665-676.
- Ludwig, K. R. (2001). Isoplot/Ex rev. 2.49. Special Publication, 1a. In "A geochronological toolkit for Microsoft Excel." (K. R. Ludwig, Ed.). Berkeley Geochronology Center, Berkeley, USA.
- Luo, X., Renkammer, M., Lee, D.-C., and Halliday, A. N. (1997). High precision $^{230}\text{Th}/^{232}\text{Th}$ and $^{234}\text{U}/^{238}\text{U}$ measurements using energy-filtered ICP magnetic sector multiple collector mass spectrometry *International Journal of Mass Spectrometry Ion Proceedings* **171**, 105-117.
- Manhes, G., Minster, J. F., and Allegre, C. J. (1978). Comparative uranium-thorium-lead and rubidium-strontium study of the Saint Severin amphoterite: consequences for early solar system chronology. *Earth and Planetary Science Letters* **39**, 14-24.

- Terra, F., and Wasserburg, G. J. (1972). U-Th-Pb systematics in three Apollo 14 basalts and the problem of initial Pb in lunar rocks. *Earth and Planetary Science Letters* **14**, 281-304.
- Vaks, A., Bar-Matthews, M., Ayalon, A., Matthews, A., Frumkin, A., Dayan, U., Halicz, L., Almogi-Labin, A., and Schilman, B. (2006). Paleoclimate and location of the border between Mediterranean climate region and the Saharo-Arabian Desert as revealed by speleothems from the northern Negev Desert, Israel *Earth and Planetary Science Letters* **249**, 384-399.
- Woodhead, J. D., and Hergt, J. M. (1997). Application to the 'double spike' technique to Pb-isotope geochronology. *Chemical Geology (Isotope Geoscience)* **138**, 311-321.
- Woodhead, J., Hellstrom, J., Maas, R., Drysdale, R., Zanchetta, G., Devine, P., and Taylor, E. (2006). U-Pb geochronology of speleothems by MC-ICPMS. *Quaternary Geochronology* **1**, 208-221.

Appendix 4: Tables of results

Table 1 page 1: U-Pb ages of speleothems

Cave number at Fig. 1C	Cave/Cite	Sample ID	Sample mass (gr)	Pb (ppb)	U (ppb)	$^{206}\text{Pb}/^{238}\text{U}$	$\pm \sigma$ (%)	$^{206}\text{U}/^{238}\text{U}$	$\pm \sigma$ (%)	$^{238}\text{U}/^{206}\text{Pb}$	$\pm \sigma$ (%)	$^{207}\text{Pb}/^{206}\text{Pb}$	$\pm \sigma$ (%)	Isochron slope	Intersept with $^{207}\text{Pb}/^{206}\text{Pb}$ axis	Age (Ma) for equilibrium initial $^{234}\text{U}/^{238}\text{U}$	Age (Ma) for assumed initial $^{234}\text{U}/^{238}\text{U}$	Assumed initial $^{234}\text{U}/^{238}\text{U}$	$\pm 2\sigma$			
8	Ashalim Cave	ASH-15-C Flowstone lamina	0.110	1.53	4298	0.0002420	0.656	4.3470	1.900	4132.7	0.656	0.230044	1.900	-0.000134	0.0000001	1.286	0.008	1.269	0.013	1.05	0.03	
			0.130	6.41	3288	0.0007317	0.265	1.6675	0.156	1366.7	0.265	0.599700	0.156									
			0.070	1.64	4102	0.0002550	0.966	3.8683	2.420	3921.6	0.966	0.258511	2.420									
			0.090	1.73	3870	0.0002692	0.779	3.4878	1.650	3714.2	0.779	0.286714	1.650									
			0.070	2.35	3313	0.0003517	0.869	2.5271	1.120	2843.0	0.869	0.395710	1.120									
		ASH-15-D Flowstone lamina	0.254	2.42	1946	0.0007046	0.241	3.1371	0.388	1419.2	0.241	0.318766	0.388	-0.000358	0.0000007	3.063	0.018	3.045	0.021	1.05	0.03	
			0.129	1.65	1612	0.0006390	0.542	3.7323	1.270	1564.9	0.542	0.267931	1.270									
			0.267	1.92	1859	0.0006398	0.261	3.7043	0.535	1563.1	0.261	0.269957	0.535									
			0.267	1.56	1848	0.0005857	0.275	4.6425	0.791	1707.4	0.275	0.215401	0.791									
			0.278	1.19	1650	0.0005499	0.307	5.6635	1.160	1818.4	0.307	0.176569	1.160									
14	K'tora Cracks	ASH-15-E Flowstone lamina	0.187	1.36	1960	0.0005406	0.376	5.9810	1.580	1850.0	0.376	0.167196	1.580	-0.000360	0.0000008	3.066	0.009	3.048	0.014	1.05	0.03	
			0.225	1.29	2426	0.0004920	0.301	10.0470	2.130	2032.4	0.301	0.099532	2.130									
			0.146	1.51	1846	0.0005781	0.465	4.8113	1.520	1729.8	0.465	0.207844	1.520									
			0.174	1.17	1931	0.0005151	0.425	7.4687	2.360	1941.4	0.425	0.133892	2.360									
			0.204	1.13	1945	0.0005084	0.37	8.0294	2.190	1967.0	0.370	0.124542	2.190									
		KTO(1)-1-J Flowstone lamina	0.194	3.04	616	0.0019118	0.264	1.7436	0.155	523.1	0.264	0.573526	0.155	-0.000272	0.0000008	2.734	0.059	2.713	0.060	1.06	0.02	
			0.153	3.33	626	0.0020394	0.38	1.7215	0.229	490.3	0.380	0.580889	0.229									
			0.270	2.37	532	0.0017548	0.308	1.7882	0.202	569.9	0.308	0.559222	0.202									
			0.280	2.02	586	0.0014228	0.324	1.9085	0.243	702.8	0.324	0.523972	0.243									
			KTO(1)-1-K Flowstone lamina	0.146	1.54	338	0.0018544	0.776	1.8809	0.556	539.3	0.776	0.531660	0.556	-0.000334	0.0000008	3.338	0.062	3.317	0.063	1.06	0.02
				0.186	6.50	552	0.0042229	0.204	1.5755	0.124	236.8	0.204	0.634719	0.124								
				0.224	3.15	376	0.0031084	0.295	1.6498	0.169	321.7	0.295	0.606134	0.169								
				0.230	2.29	347	0.0025233	0.374	1.7171	0.221	396.3	0.374	0.582377	0.221								

Table 1 page 2: U-Th ages of speleothems

Cave number at Fig. 1C	Region	Cave/Cite	Sample description	Sub-sample number	Lamina	Age (ka)	$\pm\sigma$	^{238}U ppm	$^{234}\text{U}/^{238}\text{U}$	$\pm 2\sigma$	$^{230}\text{Th}/^{234}\text{U}$	$\pm 2\sigma$	^{230}Th ppb	^{232}Th ppb	$^{230}\text{Th}/^{232}\text{Th}$
1	Jordan Valley	Ma'ale-Efrayim Cave	Stalagmite ME-12	1	ME-12-out	12.06	1.27	1.045	1.04353	0.00263	0.10508	0.00524	0.00187	10.05	35.72
				2	ME-12-A1	16.40	0.28	0.717	1.08865	0.00092	0.14027	0.00113	0.00178	1.833	185.8
				3	ME-12-A2	24.78	0.46	0.738	1.10869	0.00191	0.20446	0.00170	0.00273	3.326	155.8
				4	ME-12-K7	71.88	2.53	0.172	1.09009	0.00196	0.48749	0.00604	0.00149	5.106	55.98
			Conical stalactite ME-5	5	ME-5-A1	49.214	1.354	0.317	1.07602	0.00277	0.36572	0.00394	0.00203	4.71	82.77
				6	ME-5-A2.1	70.629	1.621	0.273	1.09671	0.00265	0.48168	0.00387	0.00235	10.65	42.11
				7	ME-5-A2.2	71.922	1.041	0.295	1.07887	0.00090	0.48726	0.00248	0.00253	2.76	173.7
				8	ME-5-B2	158.475	3.549	0.300	1.05979	0.00167	0.77563	0.00369	0.00403	9.61	79.29
				9	ME-5-B3	157.122	3.673	0.344	1.07333	0.00151	0.77439	0.00394	0.00466	15.03	58.43
				10	ME-5-C1	221.511	6.644	0.284	1.04162	0.00187	0.86547	0.00415	0.00418	18.09	43.66
				11	ME-5-C2	221.186	9.817	0.327	1.04564	0.00202	0.87887	0.00571	0.00490	2.52	367.6
				12	ME-5-C3	219.437	5.442	0.329	1.04813	0.00126	0.87718	0.00320	0.00494	12.02	77.37
				13	ME-5-D1	288.0	34.3	0.394	1.03881	0.00357	0.93987	0.01064	0.00628	13.82	85.63
				14	ME-5-D3	300.8	34.8	0.413	1.02240	0.00284	0.94353	0.00946	0.00649	1.935	631.6
				15	ME-5-D5	275.1	26.8	0.405	1.03320	0.00261	0.92966	0.00940	0.00635	4.316	277.6
2	Judea Desert near Dead Sea	Kana'im Cave	Small conical stalactite KN-1	16	KN-1-A1	31.46	0.22	4.295	1.03130	0.00110	0.25153	0.00073	0.01816	44.12	77.39
				17	KN-1-A2	131.4	2.1	6.054	0.95736	0.00086	0.69607	0.00270	0.06576	6.142	2015
				18	KN-1-B1	256.9	6.0	3.907	0.99768	0.00112	0.90522	0.00230	0.05753	1.979	5456
			Small conical stalactite KN-6	19	KN-6-A1	155.4	2.2	3.716	0.97870	0.00125	0.75715	0.00220	0.44889	0.681	12387
				20	KN-8-A	182.6 ⁶	2.3	3.565	0.95881	0.00113	0.80574	0.00168	0.04490	1.588	5314
			Small conical stalactite KN-8	21	KN-8-B	146.5	2.8	4.055	1.00082	0.00104	0.74003	0.00321	0.04896	0.844	10876
				22	KN-9-A1	37.90*	0.25	3.577	1.03120	0.00110	0.29468	0.00076	0.01816	145.12	23.52
			Stalactite KN-9	23	KN-9-A2	141.3	1.4	6.007	0.95529	0.00121	0.72112	0.00159	0.06746	1.256	10078
				24	KN-9-B	260.0	6.5	2.666	0.98943	0.00091	0.90569	0.00241	0.03895	0.511	14311
				25	KN-10(1)-A (I); KN-10(1)-A (II); KN-10(1)-B1 KN-10(1)-B2 KN-10(1)-B3	207.5 247.9 259.8 323.7 353.9	3.8 6.3 7.1 9.6 18.9	3.578 3.196 2.368 5.149 4.486	1.01872 1.00671 1.00552 0.99754 0.99752	0.00147 0.00096 0.00116 0.00116 0.00126	0.85547 0.89936 0.90969 0.94824 0.96056	0.00228 0.00273 0.00248 0.00187 0.00292	0.05084 0.04717 0.03531 0.07940 0.07007	12.89 0.322 0.328 0.310 0.367	740.1 27466 20206 48117 35860
5	Central Negev	Izzim Cave	Stalagmite IZ-1	30	IZ-1-A2	372.6	22.3	0.837	1.01512	0.00095	0.97260	0.00310	0.01347	15.90	160
				31	IZ-1-A3	500.1	85.3	0.673	1.00734	0.00142	0.99280	0.00329	0.01097	2.996	696
6	Central Negev	Hol-Zakh Cave	Flowstone HZ-1	32	HZ-1-A	129.2	1.6	7.726	0.95261	0.00168	0.68950	0.00197	0.08273	1.398	11101
				33	HZ-1-A1	126.6	1.1	6.970	0.96041	0.00143	0.68313	0.00137	0.07454	8.784	1596
				34	HZ-1-A2	130.2	1.5	6.982	0.95025	0.0009	0.69188	0.00198	0.07484	5.758	2442
				35	HZ-1-B	131.7	1.7	12.927	0.93024	0.00099	0.69314	0.00208	0.13587	1.899	13432
				36	HZ-2-A	131.3	3.0	9.182	0.93652	0.00139	0.69293	0.00380	0.09714	12.98	1405
			Stalactite HZ-2	37	HZ-2-B(I)	137.2	3.1	13.70	0.95356	0.00126	0.71066	0.00372	0.15131	10.81	2626
				38	HZ-2-B(II)	136.4	4.1	12.13	0.96693	0.00141	0.71051	0.00512	0.13583	11.33	2249
			Stalactite HZ-3(2)	39	HZ-3(2)-A	133.3	5.0	11.40	0.94994	0.00191	0.70017	0.00637	0.12359	3.418	6820
				40	HZ-3(2)-B1	133.4	2.7	17.46	0.92278	0.00134	0.69663	0.00336	0.18292	30.02	1149
				41	HZ-3(2)-B2	125.2	2.2	7.598	0.99703	0.00138	0.68340	0.00302	0.08439	2.313	6897
				42	HZ-3(2)-C1	200.0	5.6	13.00	0.9705	0.00153	0.83496	0.00371	0.17176	3.365	9626
				43	HZ-3(2)-C2	222.2	5.0	19.08	0.96496	0.00132	0.86220	0.00257	0.25884	2.860	17008
				44	HZ-3(2)-D	247.4	10.8	4.859	1.00906	0.00228	0.89942	0.00602	0.07188	2.816	4825
				45	HZ-3(2)-E	349.1	25.4	8.769	0.98493	0.00185	0.95462	0.00395	0.13441	2.429	10420
			Stalactite HZ-3(5)	46	HZ-3(5)-A1(I)	111.7	1.6	8.533	0.9366	0.00107	0.63560	0.00248	0.08281	5.626	1781
				47	HZ-3(5)-A1(II)	133.3	2.1	7.270	0.93991	0.00116	0.69884	0.00258	0.07784	4.513	3283
				48	HZ-3(5)-A2.1	133.5	1.9	7.990	0.93593	0.00131	0.69864	0.00227	0.08517	5.842	2763
				49	HZ-3(5)-A2.2	130.4	1.1	7.251	0.95327	0.00089	0.69279	0.00141	0.07807	5.281	2791
				50	HZ-3(5)-A3	126.3	1.3	18.36	0.93151	0.00076	0.67881	0.00174	0.18929	28.34	1260
				51	HZ-3(5)-A5	126.1	2.2	7.173	1.01206	0.0011	0.68775	0.00305	0.08139	4.964	3110
				52	HZ-3(5)-B	187.5	2.4	15.96	0.96631	0.00090	0.81518	0.00178	0.20496	4.484	8603

Table 1 page 3: U-Th ages of speleothems

Cave number at Fig. 1C	Region	Cave/Cite	Sample description	Sub-sample number	Lamina	Age (ka)	$\pm 2\sigma$	^{23}U ppm	$^{234}\text{U}/^{238}\text{U}$	$\pm 2\sigma$	$^{230}\text{Th}/^{234}\text{U}$	$\pm 2\sigma$	^{230}Th ppb	^{232}Th ppb	$^{230}\text{Th}/^{232}\text{Th}$
7	Central Negev	Makhtesh-ha-Qatan Cave	Stalactite MKTC-4 (intermittently grown submerged under water) Stalactite MKTC-5 (intermittently grown submerged under water)	53	MKTC-4-A	139.7*	2.0	0.1520	1.18990	0.00220	0.74355	0.00238	0.00220	19.69	21
				54	MKTC-4-B	>550		0.2328	1.15158	0.00273	1.07539	0.00497	0.00470	1.132	812
				55	MKTC-5-A	131.7*	6.6	0.7642	1.11020	0.00220	0.71385	0.00869	0.01000	137.40	14
				56	MKTC-5-C	157.2	3.8	1.5729	1.23165	0.00201	0.79250	0.00432	0.02503	1.840	2590
				57	MKTC-5-H1	296.1	9.0	0.9829	1.17163	0.00131	0.97697	0.00295	0.01834	0.819	4205
				58	MKTC-5-H2	303.5	12.5	1.2663	1.17871	0.00181	0.98367	0.00386	0.02394	1.970	2279
				59	MKTC-5-H9	344.2	16.5	2.2479	1.23056	0.00169	1.01884	0.00383	0.04594	3.110	2271
8	Central Negev	Ashalim Cave	Stalactite ASH-2 Stalactite (Column) ASH-11	60	ASH-2-A	>550		8.416	0.99963	0.00151	0.99630	0.00522	0.13664	1.715	14944
				61	ASH-11-A1.1	116.9	1.8	3.891	1.03981	0.00095	0.66237	0.00275	0.0437	1.431	5761
				62	ASH-11-A1.2	123.6	1.4	4.055	1.02781	0.00078	0.68192	0.00208	0.0463	0.791	11039
				63	ASH-11-A2.1	127.4	5.9	2.751	1.0434	0.00157	0.69474	0.00837	0.0325	0.953	6541
				64	ASH-11-A2.2	203.4	8.7	12.56	1.02947	0.00085	0.85177	0.00837	0.1795	1.113	30363
				65	ASH-11-A2.3	194.8	3.4	12.56	1.03011	0.00083	0.83886	0.00254	0.1769	1.293	25755
				66	ASH-11-A2.4	214.5	5.2	12.63	1.0271	0.00060	0.86654	0.00329	0.1833	1.328	25979
				67	ASH-11-A3	>550		11.92	1.01049	0.00122	1.03557	0.00207	0.2033	1.748	21815
				68	ASH-11-A4	>550		16.17	0.99917	0.00203	1.03400	0.00294	0.2723	2.384	21428
			Flowstone ASH-15	69	ASH-15-A+B	>550		1.861	0.99939	0.00108	1.03532	0.00113	0.0314	2.394	2460
				70	ASH-15-C1	>550		3.412	0.99925	0.00150	1.02507	0.00219	0.0570	0.766	13956
				71	ASH-33-A1	116.7	3.1	3.298	1.03892	0.00209	0.66166	0.00471	0.03696	2.371	3013
				72	ASH-33-A2	124.0	3.6	3.371	1.02948	0.00191	0.68327	0.00513	0.03866	2.172	3508
				73	ASH-33-B1	131.2	2.4	3.583	1.02585	0.00103	0.70373	0.00322	0.04217	1.396	5805
				74	ASH-33-B2	129.1	2.3	4.156	1.02745	0.00088	0.69807	0.00320	0.04859	1.035	8969
				75	ASH-33-B3	129.2	2.9	4.161	1.02453	0.00087	0.69803	0.00406	0.04852	1.188	7780
				76	ASH-33-B4	127.5	3.6	3.634	1.02456	0.00138	0.69312	0.00506	0.04207	1.824	4447
				77	ASH-33-C1	201.9	6.0	9.854	1.0167	0.00080	0.84705	0.00419	0.13834	2.576	10180
				78	ASH-33-C2	199.7	7.0	11.73	1.01826	0.00078	0.84413	0.00502	0.16434	2.306	13498
			Flowstone ASH-34	79	ASH-33-D1	203.0	4.7	11.53	1.01834	0.00088	0.84906	0.00323	0.16248	4.444	6934
				80	ASH-33-D2	208.3	4.1	11.00	1.01839	0.00117	0.85654	0.00256	0.15644	6.406	4829
				81	ASH-33-E1.1	207.9	3.6	12.37	1.01819	0.00061	0.85590	0.00238	0.17575	1.254	26447
				82	ASH-33-E1.2	213.6	4.7	14.27	1.01996	0.00060	0.86394	0.00295	0.20495	2.627	14713
				83	ASH-33-E2.1	218.0	2.8	13.29	1.01997	0.00073	0.86961	0.00161	0.19217	3.041	11929
				84	ASH-33-E2.2(I)	199.5	2.5	15.59	1.02737	0.00110	0.84628	0.00226	0.22102	2.258	18499
				85	ASH-33-E22(II)	199.7	4.2	14.63	1.02247	0.00081	0.84495	0.00295	0.20600	21.01	1850
				86	ASH-33-E22(III)	198.8	4.3	14.05	1.02368	0.00247	0.84368	0.00311	0.19785	5.700	6574
				87	ASH-33-F-up	>550		13.18	1.01930	0.00121	1.01496	0.00351	0.22235	2.589	16219
				88	ASH-34-A1	118.4	1.7	3.805	1.04200	0.00078	0.66737	0.00264	0.04314	5.333	1526
				89	ASH-34-A2	122.9	1.9	3.626	1.05737	0.00108	0.68293	0.00272	0.04269	0.463	17381
				90	ASH-34-B1	127.2	1.6	3.093	1.04600	0.00069	0.69434	0.00222	0.03662	1.510	4581
				91	ASH-34-B2	132.7	1.7	4.673	1.05782	0.00106	0.71139	0.00229	0.05732	1.542	7003
				92	ASH-34-C1	199.3	3.3	12.84	1.02963	0.00093	0.84575	0.00229	0.18229	10.75	3196
				93	ASH-34-C2	200.1	4.4	15.25	1.02501	0.00066	0.84605	0.00317	0.21560	14.47	2820
				94	ASH-34-C3(I)	192.3	2.9	19.68	1.02945	0.0007	0.83484	0.00222	0.27572	15.40	3370
				95	ASH-34-C3 (II)	188.4	6.9	13.99	1.02299	0.00095	0.82727	0.00557	0.19304	13.87	2636
				96	ASH-34-C3 (III)	194.6	4.3	15.19	1.02689	0.00231	0.83797	0.00296	0.21301	11.00	3667
				97	ASH-34-D	>550		16.45	1.01215	0.0013	0.99939	0.00431	0.27129	1.740	29461

Table 1 page 4: U-Th ages of speleothems

Cave number at Fig. 1C	Region	Cave/Cite	Sample description	Sub-sample number	Lamina	Age (ka)	$\pm 2\sigma$	^{238}U ppm	$^{234}\text{U}/^{238}\text{U}$	$\pm 2\sigma$	$^{230}\text{Th}/^{234}\text{U}$	$\pm 2\sigma$	^{230}Th ppb	^{232}Th ppb	$^{230}\text{Th}/^{232}\text{Th}$
9	Central Negev	Ever-Sid-Ramon mini-caves	Flowstone ESID-1	98	ESID-1 (single lamina)	>500*		0.05894	1.01328	0.00139	1.11336	0.00153	7.2760	0.001084	29
				99	ESID-2-A1.1	87.7	1.4	0.4282	1.09346	0.00132	0.55917	0.00283	0.0043	1.465	566
				100	ESID-2-A1.2	205.4	8.2	0.2079	1.06844	0.00126	0.86183	0.00578	0.0031	2.690	239
				101	ESID-2-A2	292.3	8.5	0.6746	1.06505	0.00061	0.94957	0.00270	0.0111	0.989	2134
				102	ESID-2-B	516.4	50.6	2.193	1.05601	0.00068	1.01256	0.00209	0.0382	1.362	5282
			Stalactite SHAR-1	103	ESID-2-C	>550		0.1182	1.00633	0.00125	1.42795	0.01955	0.0028	0.336	1553
				104	SHAR-1-A	>550		0.1506	1.01788	0.00105	1.10966	0.00401	8.2529	0.002773	64
				105	ESID-7(1)-A1+A2	133.6	3.1	0.1486	1.05006	0.00104	0.71297	0.00417	0.0018	2.065	177
				106	ESID-7(1)-A3	463.1	55.4	0.4954	1.02994	0.0021	0.99701	0.00304	0.0083	4.245	375
				107	ESID-7(1)-B	462.7	98.8	0.8810	1.02007	0.00402	0.99334	0.00487	0.0146	9.399	298
				108	ESID-7(1)-C1	>550		0.3153	1.01146	0.00078	1.02404	0.00207	0.0053	1.476	680
			Stalactite ESID-7(3)	109	ESID-7(3)-A1	212.3*	3.3	0.2811	1.04159	0.00098	0.87388	0.00192	0.00417	26.66	30
				110	ESID-7(3)-B	304.9	9.0	1.538	1.01369	0.00008	0.94341	0.00253	0.02397	2.864	1582
			Stalactite ESID-11	111	ESID-11-A	555.7	201.3	0.1891	1.05595	0.00196	1.01566	0.04700	0.00331	0.889	701
			Stalactite ESID-13	112	ESID-13-A	>550		2.255	1.02911	0.0013	1.00976	0.00208	0.03821	2.299	3119
				113	ESID-13-B1	>550		0.9365	1.01763	0.00149	1.03555	0.00337	0.01609	1.393	2172
				114	ESID-13-C	>550		0.7605	1.05324	0.00109	1.07713	0.00220	0.01407	1.090	2427
			Stalagmite ESID-18	115	ESID-18-A1	>550		0.7366	1.02939	0.00089	1.03234	0.00258	0.01276	3.023	793
				116	ESID-18-A2	>550		1.140	1.02873	0.00089	1.04006	0.00352	0.01989	2.850	1312
				117	ESID-18-E	>550		1.887	1.01371	0.00106	1.04703	0.00243	0.03265	1.212	5055
				118	ESID-22-A1	>550		0.6074	1.02949	0.00121	1.02173	0.00304	0.01042	4.430	442
			Stalagmite ESID-22	119	ESID-22-A2	>550		1.136	1.02549	0.00107	1.03753	0.00295	0.01971	1.653	2240
				120	ESID-22-D	>550		1.558	1.01595	0.00124	1.03545	0.00308	0.02672	1.613	3111
11	Central Negev	Wadi Lotz Cave	Stalagmite fragment LOTZ-3	121	LOTZ-3-A	>550		0.3794	1.0484	0.00085	1.04894	0.00185	0.00680	2.530	509

Table 1 page 5: U-Th ages of speleothems

Cave number at Fig. 1C	Region	Cave/Cite	Sample description	Sub-sample number	Lamina	Age (ka)	$\pm 2\sigma$	^{238}U ppm	$^{234}\text{U}/^{238}\text{U}$	$\pm 2\sigma$	$^{230}\text{Th}/^{234}\text{U}$	$\pm 2\sigma$	^{230}Th ppb	^{232}Th ppb	$^{230}\text{Th}/^{232}\text{Th}$
12	Southern Negev	Ma'ale-ha-Meishar Cave	Flowstone MMR-7	122	MMR-7-A1.1	112.3	1.2	0.4743	1.07182	0.00116	0.65483	0.00191	0.00543	11.58	89
				123	MMR-7-A1.2	110.6	1.2	0.4888	1.07082	0.00133	0.66441	0.00194	0.00546	3.685	281
				124	MMR-7-A3.1	116.4	1.9	0.5324	1.06428	0.00203	0.66295	0.00297	0.00378	0.881	820
				125	MMR-7-A3.2	194.6	3.0	0.3824	1.07651	0.00106	0.84665	0.00223	0.00568	16.11	67
				126	MMR-7-B	>550		0.7164	1.02994	0.00066	1.01282	0.00286	0.01218	9.154	280
				127	MMR-7(2)-A1	120.7	1.9	0.4587	1.07295	0.00199	0.67735	0.00273	0.00544	3.358	305
			Flowstone MMR-7(2)	128	MMR-7(2)-A2	127.0*	4.3	0.6127	1.0783	0.00204	0.69780	0.00290	0.00757	49.76	29
				129	MMR-7(2)-B	203.0	9.4	0.2986	1.06326	0.00136	0.85747	0.00675	0.00444	3.283	256
			Flowstone MMR-8	130	MMR-8-A	198.4	3.8	0.2771	1.06865	0.00119	0.85137	0.00279	0.00411	6.268	125
				131	MMR-8-B	206.1	2.8	0.2475	1.06427	0.00092	0.86219	0.00184	0.00370	5.193	135
			Stalagmite MMR-13	132	MMR-13-A	138.0	4.1	0.3757	1.06167	0.00303	0.72610	0.00512	0.00472	10.23	87
				133	MMR-13-B1	124.5	3.2	0.3685	1.06027	0.00141	0.68786	0.00464	0.00438	11.69	71
				134	MMR-13-B2	132.7	5.1	0.3704	1.05825	0.00141	0.71131	0.00703	0.00455	2.532	339
				135	MMR-13-C	140.1	3.5	0.3740	1.05825	0.00153	0.73075	0.00436	0.00470	7.731	115
				136	MMR-13-E	132.0	2.6	0.3654	1.06014	0.00126	0.70999	0.00334	0.00448	1.665	507
				137	MMR-13-F	132.6	2.2	0.3671	1.06593	0.00075	0.71259	0.00298	0.00455	4.782	179
				138	MMR-13-G	131.6	4.2	0.4035	1.07034	0.00152	0.70963	0.00577	0.00500	3.591	263
				139	MMR-13-H	126.9	2.6	0.4831	1.07206	0.00139	0.69675	0.00332	0.00588	2.862	387
				140	MMR-13-I	130.9	1.7	1.716	1.07753	0.00093	0.70830	0.00239	0.02135	7.516	533
				141	MMR-13-K	>550		0.3811	1.01435	0.0027	1.01946	0.00565	0.00642	3.767	321
13	Southern Negev	Shizafon mini-caves	Flowstone SHACH-1	142	SHACH-1-A	341.7	12.0	0.6578	1.02498	0.00117	0.96470	0.00212	0.01060	24.99	80
				143	SHACH-1-B	333.4	19.0	0.3081	1.0115	0.00114	0.95700	0.00379	0.00486	3.376	274
			Flowstone SHACH-3	144	SHACH-3-A	351.7	12.6	0.2135	1.03883	0.00105	0.97296	0.00211	0.00352	4.426	150
				145	SHACH-3-B	431.6	64.7	0.2232	1.02417	0.00217	0.98988	0.00500	0.00369	1.732	401
				146	SHACH-3-C	>550		0.2051	1.02391	0.00066	1.01832	0.00349	0.00349	1.108	600
				147	SHACH-3-E	>550		0.1845	1.04516	0.00115	1.03029	0.00290	0.00324	1.389	449
			Pool crust SHACH-4	148	SHACH-4	488.9*	34.9	0.4721	1.02872	0.00141	0.99978	0.00140	0.00792	72.76	21
			Flowstone SHACH-6	149	SHACH-6-A	>550		0.3552	1.03231	0.00081	1.01145	0.00084	0.00605	4.021	284
			Flowstone KTO-1(1)	150	KTO-1(1)-A1	137.7*	1.5	0.9823	1.0433	0.00147	0.72318	0.00160	0.01232	224.52	10
				151	KTO-1(1)-A2	319.6	8.5	0.9264	1.02516	0.00082	0.95465	0.00190	0.01593	73.22	41
14	Southern Negev	Klora Cracks		152	KTO-1(1)-C	>550		0.8018	1.00737	0.00119	1.00108	0.00368	0.01318	4.085	619

*Ages corrected for ^{230}Th from detrital origin, other ages don't need correction because their $^{230}\text{Th}/^{232}\text{Th}$ ratios are above 30 (Kaufman et al., 1998).
 @Age overturn in lamina KN-8-A, relative to KN-8-B. KN-8-A is a surface lamina, affected by condensation corrosion which possibly induced radionuclide remobilisation (Auler and Smart, 2004). Thus the age is not reliable.

References:

Auler, A. S., and Smart, P. L. (2004). Rates of condensation corrosion in speleothems. *Speleogenesis and Evolution of Karst Aquifers* (www.speleogenesis.info) 2, p.2 (4 pages).
 Kaufman, A., Wasserburg, G. J., Porcelli, D., Bar-Matthews, M., Ayalon, A., and Halicz, L. 1998. U-Th isotope systematics from the Soreq Cave, Israel, and climatic correlations. *Earth and Planetary Science Letters*, v. 156, p. 141-155.

Table 2: δD and $\delta^{18}O$ values of present-day rainfall in the Negev Desert - page 1

Station with GP and elevation	sample	date of rain event	amount (mm)	$\delta^{18}O$	δD	d-excess
Beer-Sheva	BSH-1	29.10.04	52.3	-2.1	-8.5	7.9
GP: 31° 16' 39" N; 034° 49' 23" E	BSH-2	17.-18.11.04	18.4	-4.3	-9.8	24.3
Elevation: 380 m asl	BSH-3	18.11.04	9.2	-4.2	-11.1	22.3
	BSH-4	21-22.11.04	68.1	-5.4	-8.2	34.7
	BSH-5	26-28.11.04	19.1	-3.0	3.0	27.1
	BSH-6	07-08.12.04	10.4	-4.3	-15.6	18.5
	BSH-7	12.12.04	drops			
	BSH-8	15-16.12.04	3.6	-5.0	-13.7	26.5
	BSH-9	24-25.12.04	26.6	-11.8	-68.3	25.7
	BSH-10	25-26.12.04	3.7	-4.9	-18.4	20.5
	BSH-11	02-04.01.05	22.8	-5.6	-28.6	16.0
	BSH-12	05-06.01.05	36.1	-9.1	-54.0	19.0
	BSH-13	06-07.01.05	1.2	-4.2	-10.8	23.1
	BSH-14	19-20.01.05	3.7	-4.7	-19.4	17.9
	BSH-15	24.01.05	14.8	-4.4	-17.2	18.2
	BSH-16	06-09.02.05	39.6	-2.9	-3.3	19.5
	BSH-17	11-12.02.05	6.6	-3.7	-8.0	21.3
	BSH-18	12-13.02.05	2.5	-6.4	-31.7	19.6
	BSH-19	21.02.05	1.4	-1.3	7.4	17.8
	BSH-20	04.03.05	1.5	-1.2	-8.1	1.6
	BSH-21	08-10.03.05	40.9	-3.2	-13.7	12.2
	BSH-22	12.03.05	4.3	-1.8	7.2	21.8
	BSH-23	02.04.05	0.6	-3.2	-9.4	15.9
	BSH-24	03.04.05	1.4	-1.4	7.3	18.7
	Summary	Winter 2004-2005	388.7 mm	-4.8	-18.2	20.6
		(Vaks et al., 2006)				
	BSH-101	18-21.10.05	4.1	-0.4	9.6	13.1
	BSH-102	07.11.05	1.9	-2.7	-5.3	16.1
	BSH-103	15-16.11.05	4.6	-2.5	-2.1	17.7
	BSH-104	21.11.05	13.3	-4.3	-12.5	21.5
	BSH-105	22.11.05	5.9	-1.1	-2.1	6.7
	BSH-106	20.12.05	3.3	-2.9	4.0	27.0
	BSH-107	23.12.05	5.5	-3.6	-1.6	26.9
	BSH-108	24-26.12.05	23.1	-4.1	-7.8	25.2
	BSH-109	04.01.06	0.6	6.3	37.4	-13.3
	BSH-110	07-09.01.06	11.5	-4.4	-8.1	26.9
	BSH-111	15.01.06	0.3	-2.4	-5.7	13.1
	BSH-112	16-18.01.06	1.7	-1.7	7.3	21.0
	BSH-113	26-28.01.6	5.0	-4.2	-7.7	26.0
	BSH-114	28.01.06	3.6	-2.4	4.2	23.1
	BSH-115	03.02.06	4.2	-2.8	-6.7	15.9
	BSH-116	04.02.06	0.7	-0.7	4.5	9.9
	BSH-117	13-17.02.06	61.2	-6.8	-29.2	24.8
	BSH-118	26.02.06	0.2	4.5	26.0	-10.1
	BSH-119	09-10.03.06	3.6	-1.4	0.5	11.3
	BSH-120	01-02.04.06	12.6	-6.0	-40.6	7.1
	BSH-121	04-06.04.06	23.7	-4.2	-19.5	13.9
	BSH-122	16.04.05	10.2	-3.2	-17.3	8.4
	BSH-123	24.04.06	5.3	-1.3	-7.4	3.1
	BSH-124	25.04.06	0.5	2.6	12.7	-8.1
	Summary	Winter 2005-2006	206.5 mm	-4.5	-16.5	19.4
	Summary	2004-2006	595.2 mm	-4.7	-17.6	20.2

Table 2 - page 2

Station with GP and elevation	sample	date of rain event	amount (mm)	$\delta^{18}\text{O}$	δD	d-excess
Arad						
GP: 31° 16' 01" N; 35° 13' 11" E	ARAD-1	29.10.04	31.3	-3.1	-10.8	14.2
Elevation: 605 m asl	ARAD-2	01.11-14.12.04	51.8	-3.2	-11.5	14.0
	ARAD-3	14.12-02.02.05	35.9	-7.2	-28.5	28.8
	ARAD-4	03.02-13.03.05	41.4	-3.7	-2.3	27.0
	ARAD-5	13.03-22.05.05	0.7	1.3	5.7	-4.9
	Summary	Winter 2004-2005	161.1 mm	-4.2	-12.7	20.6
	ARAD-101	02.10-15.12.05	5.1	-2.4	-0.3	19.1
	ARAD-102	15.12.05-02.02.06	11.1	-3.6	-11.8	16.6
	ARAD-103	02.02-07.03.06	19.7	-5.6	-28.4	16.5
	ARAD-104	07.03-10.04.06	6.9	-3.6	-13.35	15.5
	ARAD-105	10-30.04.06	10.1	-2.3	-12.7	5.4
	Summary	Winter 2005-2006	52.9 mm	-4.0	-17.2	14.5
	Summary	2004-2006	214 mm	-4.1	-13.8	19.1
Makhtesh-ha-Qatan						
GP: 30° 57' 14" N; 35° 12' 34" E						
Elevation: Sea level	MQ-1	29.10.04	65.0	-2.5	-12.5	7.7
	MQ-2	01.11-14.12.04	14.6	2.5	25.3	5.4
	MQ-3	14.12.04-02.02.05	15.6	-3.9	-7.4	24.0
	MQ-4	03.02-13.03.05	26.3	-1.9	0.5	15.9
	Summary	Winter 2004-2005	121.5 mm	-1.97	-4.49	11.3
	***	02.10-15.12.05	no rain			
	***	15.12.05-02.02.06	no rain			
	MQ-101	02.02-07.03.06	10.4	0.2	3.0	1.8
	MQ-102	07.03-10.04.06	7.3	-7.7	-27.0	34.52
	MQ-103	10-30.04.06	7.0	-2.6	-18.8	2.05
	Summary	Winter 2005-2006	24.7 mm	-2.9	-11.9	11.4
	Summary^s	2002-2006^s	216.9 mm^s	-2.4	-6.1	13.4
Mitzpe-Ramon						
GP: 30° 36' 59" N; 034° 48' 06" E	MR-1	29.10.04	5.6	2.5	30.3	10.5
Elevation: 820 m asl	MR-2	01.11-14.12.04	11.13	0.8	16.5	9.9
	MR-3	14.12.04-02.02.05	18.2	-5.6	-30.8	14.2
	MR-4	03.02-13.03.05	13.3	-3.5	-4.4	23.8
	Summary	Winter 2004-2005	48.2 mm	-2.62	-5.51	15.4
	MR-101	02.10-15.12.05	1	-0.9	22	29.28
	MR-102	15.12.05-02.02.06	4.1	-3.9	7.8	38.6
	MR-103	02.02-07.03.06	25.1	-4.2	-21.1	12.5
	MR-104	07.03-10.04.06	7.9	-3.9	-15.3	15.5
	MR-105	10-30.04.06	11.8	-3.6	-19.3	9.5
	Summary	Winter 2005-2006	49.9 mm	-3.9	-16.5	14.7
	Summary	2004-2006	98.1 mm	-3.3	-11.1	15.1

Table 2 - page 3

Station with GP and elevation	sample	date of rain event	amount (mm)	$\delta^{18}\text{O}$	δD	d-excess
Neot-Smadar						
GP: 30° 02' 55"N; 35° 01' 27" E	not sampled [#]	29.10.04	5.0	lost	lost	
Elevation: 405 m asl	NSM-1	22.11.04	0.2	-1.9	21.7	37.1
	NSM-2A	16.12.04	0.7	-4.3	-10.8	23.5
	NSM-2B	24-25.12.04	0.2	lost	lost	
	NSM-2C	05.01.05	0.8	lost	lost	
	NSM-2D	06.01.05	0.1	lost	lost	
	NSM-2E	09.01.05	1.5	-2.9	-23.5	-0.1
	NSM-2F	16.01.05	0.8	0.8	2.9	-3.7
	NSM-3A	11.02.05	0.5	0.9	23.8	16.9
	NSM-3B	12.02.05	drops			
	NSM-3C	20.02.05	0.1	0.8	2.9	-3.7
	NSM-3D	08.03.05	2.2	-1.3	-9.3	1.3
	NSM-3E	09.03.05	5.0	-2.9	-8.7	14.4
	NSM-3F	10.03.05	1.3	-6.1	-12.3	36.3
	NSM-4	23.04.05	0.5	5.1	28.8	-12.1
	Summary*	Winter 2004-2005	18.9 mm*	-2.3	-7.0	11.3
	no rain	02.10-15.12.05				
	NSM-101	15.12.05-02.02.06	1.4	-1.2	-4.5	4.8
	NSM-102	02.02-07.03.06	4.04	-2.9	-15.1	7.9
	NSM-103	07.03-10.04.06	1.7	-0.4	-5.1	-2.2
	NSM-104	10-30.04.06	1.8	-0.5	1.85	5.9
	Summary	Winter 2004-2005	8.9 mm	-1.7	-8.2	5.1
	Summary*	2004-2006	27.8 mm*	-2.0	-7.5	8.8
Ein-Netafim Spring						
GP: 29° 35' 48.69"; 34° 52' 51.52"	NET-1	23.06.2004	spring water	-4.2	-18.0	15.8
550 m asl	NET-2	23.06.2004	spring water	-3.7	-18.8	10.6
	Average			-3.9	-18.4	13.2

^sThe summary of the rainfall data for Makhtesh-ha-Qatan includes the data from this study (winters 2004-2006) and from Kuperman (2005) (winters 2002-2004).

[#]Neot-Smadar station was constructed on 02.11.2004.

*In Neot-Smadar during the winter 2004-2005 the $\delta^{18}\text{O}$ and δD were measured only on 12.8 mm from the 18.9 mm of total rainfall.

Table 3: δD values and calculated $\delta^{18}O$ values of the Negev Desert speleothem fluid inclusions - page 1

Cave	Sample Number	Age range (ka)	$\delta^{18}O$ calcite (‰ PDB)	Standard deviation	$\delta D_{\text{extracted water}}$ (‰ SMOW)	Standard deviation	δD_{FI} (‰ SMOW)	Number of replicates	Minimum temperature (°C)	Maximum temperature (°C)	Calculated minimum $\delta^{18}O$ (‰ SMOW)	Calculated maximum $\delta^{18}O$ (‰ SMOW)	Average d-excess ($\pm 3.5\%$)
Tzavoa	TZ-4-A2	124-126	-10.0	0.7	-80.0	7.1	-41.2	3	18	22	-10.35	-9.44	38
	TZ-4-B1-2	128-131	-9.6	0.6	-71.0	6.9	-31.7	4	15	19	-10.56	-9.62	49
	TZ-15-B-1	32-35	-5.42	0.09	-67.5	2.3	-31.1	6	12	16	-7.03	-6.06	21
	TZ-15-F	61-64	-6.0	0.3	-64.0	14.6	-32.2	4	13	17	-7.35	-6.39	23
	TZ-15-G	65-67	-4.7	0.4	-57.4	3.5	-23.0	4	12	16	-6.31	-5.34	24
	TZ-22(I)-C2	157-161	-6.6	0.3	-69.4	1.5	-34.0	4	12	17	-8.21	-7.00	27
	TZ-22(I)-F4+G1	172-175	-9.8	0.5	-84.9	3.7	-48.7	3	12.5	16.5	-11.46	-10.50	39
	TZ-22(I)-H	175-177	-9.9	0.7	-78.8	1.1	-43.6	4	13.5	17.5	-11.32	-10.37	43
	TZ-22(I)-K2+K3	195-198	-10.1	0.4	-92.1	1.5	-57.0	3	16.5	20.5	-10.77	-9.84	25
	TZ-22(I)-L	198-201	-9.6	0.4	-83.0	5.5	-43.8	3	16.5	20.5	-10.27	-9.35	35
Hol-Zakh	HZ-1-A	126-131	-9.9	0.3	-87.9	2.8	-53.4	4	19.5	23.5	-9.86	-8.96	22
	HZ-1-B	131-132	-9.5	0.4	-92.0	6.1	-57.6	2	18	22	-9.77	-8.85	17
Ashalim	ASH-33-A	117-124	-8.3	0.3	-80	6.3	-41.3	3	19	23	-8.29	-7.38	21
	ASH-33-B123	126-130	-8.5	0.3	-87.2	5.2	-48.9	3	18	22	-8.77	-7.85	18
	ASH-33-B4	130-132	-8.4	0.2	-73.7	5.6	-36.1	2	16	20	-9.14	-8.21	33
	ASH-33-C	199-202	-7.7	0.1	-94.6	4.0	-55.4	5	17.5	21.5	-8.10	-7.18	6
	ASH-33-D	203-209	-8.1	0.4	-90.3	3.7	-54.1	8	19	23	-8.09	-7.19	7
	ASH-33-E1	210-213	-7.3	0.4	-84.5	2.2	-51.3	4	15	19	-8.26	-7.32	11
	ASH-33-E2.1	218-220	-7.6	0.3	-90	4.7	-56.9	3	14	18	-8.78	-7.83	10

Table 3 - page 2

Cave	Sample Number	Age range (ka)	$\delta^{18}\text{O}$ calcite (‰ PDB)	Standard deviation	$\delta\text{D}_{\text{extracted water}}$ (‰ SMOW)	Standard deviation	$\delta\text{D}_{\text{PI}}$ (‰ SMOW)	Number of replicates	Minimum temperature (°C)	Maximum temperature (°C)	Calculated minimum $\delta^{18}\text{O}$ (‰ SMOW)	Calculated maximum $\delta^{18}\text{O}$ (‰ SMOW)	Average d-excess ($\pm 3.5\%$)
Ashlaim	ASH-34-A	117-124	-8.0	0.5	-72.1	2.2	-34.4	3	19	23	-8.02	-7.12	26
	ASH-34-B	127-133	-8.3	0.4	-80.5	4.1	-42.2	4	16	20	-9.07	-8.14	27
	ASH-34-C	201-203	-7.4	0.1	-87.6	5.3	-53.0	4	17.5	21.5	-7.71	-6.79	5
	ASH-15-A1	Early-Middle Pleistocene	-9.58	0.04	-73.1	4.9	-34.2	3	?	?	?	?	?
	ASH-15-A2	Early-Middle Pleistocene	-9.9	0.1	-82.6	2.5	-43.4	3	?	?	?	?	?
	ASH-15-B1	Early-Middle Pleistocene	-9.71	0.04	-85.4	4.7	-46.9	4	?	?	?	?	?
	ASH-15-B2	Early-Middle Pleistocene	-9.9	0.1	-89.2	3.7	-51.1	5	?	?	?	?	?
	ASH-15-C1	Early-Middle Pleistocene	-9.44	0.09	-90.6	4.3	-52.9	2	?	?	?	?	?
	ASH-15-C2	Early-Middle Pleistocene	-9.40	0.09	-85.1	1.9	-26.0	4	?	?	?	?	?
	ASH-15-C3	Early-Middle Pleistocene	-9.4	0.1	-77.8	8.9	-39.1	3	?	?	?	?	?
	ASH-15-YEE	3000-3200	-11.2	0.5	-87.3	3.0	-47.3	4	?	?	?	?	?
Matale-ha-Meyshar	MMR-7(2)-A	120-128	-10.5	0.2	-92.0	7.1	-55.9	2	17.5	21.5	-10.93	-10.02	28
	MMR-8	195-207	-10.2	0.3	-86.4	1.3	-48.3	3	18	22	-10.52	-9.61	32

Table 4: Sr concentrations and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios: speleothems, host rocks, bulk soils and their silicate fractions in the Jordan Valley and the Negev Desert. Page 1: speleothems younger than 550 ka.

Cave/Site/Group	Sample	Age (ka)	$\pm 2 \sigma$	Sr concentration (ppm)	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm 2 \sigma$
U-Th dateable speleothems:						
Ma'ale-Efrayim Cave	ME-3-E	134.70	2.80	267	0.70840	0.00001
	ME-5-A1	49.21	1.35	220	0.70838	0.00001
	ME-5-A2-1	70.63	1.62	187	0.70846	0.00001
	ME-5-A2-2	71.92	1.04	126	0.70846	0.00001
	ME-5-B3	157.12	3.67	262	0.70840	0.00001
	ME-5-B2	158.48	3.55	193	0.70837	0.00001
	ME-5-C1	211.51	6.64	106	0.70828	0.00001
	ME-5-C3	219.43	5.44	116	0.70818	0.00001
	ME-5-C2	221.19	9.82	154	0.70833	0.00001
	ME-5-D1	288.0	34.3	113	0.70842	0.00001
	ME-5-D3	300.8	34.8	135	0.70815	0.00001
	ME-12-A1	16.40	0.28	342	0.70835	0.00001
	ME-12-A2	24.79	0.46	296	0.70846	0.00001
	ME-12-C	30.80	0.60	208	0.70842	0.00001
	ME-12-F	31.60	0.80	176	0.70841	0.00001
	ME-12-I2	35.80	1.00	157	0.70844	0.00001
	ME-12-I3	38.10	1.00	310	0.70843	0.00001
	ME-12-I4	38.70	0.40	187	0.70852	0.00001
	ME-12-J1	48.00	1.40	125	0.70847	0.00003
	ME-12-J4	53.90	1.00	150	0.70848	0.00001
	ME-12-J5	59.10	2.00	117	0.70831	0.00001
	ME-12-K3	61.60	2.20	143	0.70847	0.00001
	ME-12-K5	63.00	1.80	177	0.70853	0.00001
	ME-12-K7	71.88	2.53	160	0.70851	0.00001
Ma'ale-Dragot Caves	MD(1)-3-A	65.54	0.45	1270	0.70831	0.00001
	MD(1)-8-A1	3.83	0.05	152	0.70807	0.00002
	MD(1)-9-A	27.75	0.28	243	0.70838	0.00001
	MD(1)-9-B	45.14	0.78	175	0.70838	0.00002
	MD(1)-9-C	112.26	1.44	184	0.70823	0.00003
Tzavoa Cave	TZ-4-A2	124.97	1.62	57	0.70834	0.00001
	TZ-4-B1,2	128.88	2.43	99	0.70831	0.00002
	TZ-4-B2	136.83	2.95	151	0.70831	0.00002
	TZ-12-A1	20.73	0.14	527	0.70841	0.00001
	TZ-12-A2	25.95	0.10	385	0.70846	0.00001
	TZ-14-A	14.02	0.21	473	0.70845	0.00001
	TZ-14-B+C	16.04	0.21	312	0.70851	0.00001
	TZ-14-E	17.21	0.16	290	0.70850	0.00001
	TZ-14-F	17.43	0.21	462	0.70849	0.00001
	TZ-14-G	18.11	0.19	450	0.70846	0.00001
	TZ-14-H	18.92	0.23	335	0.70848	0.00001
	TZ-14-J	18.99	0.11	357	0.70846	0.00001
	TZ-14-K	19.95	0.28	384	0.70846	0.00001

Table 4 - Sr data, page 2: speleothems younger than 550 ka.

Cave/Site/Group	Sample	Age (ka)	$\pm 2 \sigma$	Sr concentration (ppm)	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm 2 \sigma$
	TZ-15-B1	33.68	0.32	310	0.70844	0.00001
	TZ-15-B2	37.99	0.29	433	0.70841	0.00001
	TZ-15-C	48.04	0.27	337	0.70839	0.00001
	TZ-15-D	53.64	0.25	325	0.70840	0.00001
	TZ-15-E	61.09	0.64	257	0.70844	0.00001
	TZ-15-F	62.68	0.82	309	0.70846	0.00001
	TZ-15-G	66.26	0.64	311	0.70844	0.00001
	TZ-15-I	70.49	0.32	402	0.70838	0.00001
	TZ-15-K	69.84	0.67	335	0.70840	0.00001
	TZ-15-L	72.63	0.40	400	0.70839	0.00001
	TZ-15-M	74.32	0.48	368	0.70843	0.00001
	TZ-15-N	77.94	0.79	400	0.70841	0.00001
	TZ-21-A3	74.96	3.71	280	0.70836	0.00001
	TZ-22(1)-C2	157.35	1.48	354	0.70851	0.00001
	TZ-22(1)-F2+F3	174.46	2.03	166	0.70829	0.00001
	TZ-22(1)-H	175.11	2.64	173	0.70827	0.00001
	TZ-22(1)-K2+K3	197.00	3.00	78	0.70817	0.00001
	TZ-22(1)-L	199.00	3.00	66	0.70823	0.00001
Hol-Zakh Cave	HZ-1-A	128.39	1.32	252	0.70822	0.00001
	HZ-1-B	131.66	1.67	510	0.70830	0.00001
	HZ-3(5)-A2	131.93	1.50	300	0.70821	0.00001
	HZ-3(5)-B	187.49	2.45	557	0.70822	0.00001
Ashalim Cave	ASH-33-C	200.78	6.47	408	0.70828	0.00001
	ASH-33-D	205.69	4.42	357	0.70831	0.00001
	ASH-33-E	213.19	3.71	505	0.70824	0.00001
	ASH-34-A	120.66	1.79	500	0.70841	0.00001
	ASH-34-B	129.94	1.65	395	0.70839	0.00001
Even-Sid-Ramon mini-caves	ESID-2-A2	292.28	8.47	180	0.70827	0.00001
	ESID-2-A1.1	87.73	1.38	203	0.70827	0.00001
	ESID-7-A1	132.78	1.56	58	0.70815	0.00001
Ma'ale-ha-Meyshar Cave	MMR-7-A1	111.45	1.20	113	0.70821	0.00001
	MMR-7(2)-A+B	123.84	2.00	88	0.70822	0.00001
	MMR-8	201.05	3.31	111	0.70817	0.00001
Shizafon Quarry mini-caves	SHACH-1-A	339.45	12.00	43	0.70804	0.00001
Ma'ale-Ktora Cracks	KTO-1(1)-A2	319.29	8.51	193	0.70794	0.00001

Table 4 - Sr data, page 3: speleothems older than 550 ka and host rocks.

Cave/Site/Group	Sample	Age (ka)	±2 σ	Sr concentration (ppm)	⁸⁷ Sr/ ⁸⁶ Sr	±2 σ
Intermediate Member (IM)						
Ashalim Cave	ASH-15-A	550-1269		42	0.70846	0.00001
	ASH-15-B	550-1269		58	0.70839	0.00001
	ASH-15-C	1269	13	110	0.70826	0.00001
Even-Sid-Ramon mini-caves	ESID-2-C	550-2700?		45	0.70821	0.00001
	ESID-2-J	550-2700?		44	0.70810	0.00001
Ma'ale-Ktora Cracks	KTO(1)-1-E	550-2713		41	0.70829	0.00001
Basal Member (BM)		Average IM			0.70829	0.00012
Ashalim Cave	ASH-15-D1	3045	21	260	0.70789	0.00001
	ASH-15-D2	3045	21	252	0.70790	0.00001
	ASH-15-E	3048	14	290	0.70784	0.00001
	ASH-15-F	3048-3300?		280	0.70783	0.00001
	ASH-15-G1	3048-3300?		318	0.70785	0.00001
	ASH-15-G2	3048-3300?		320	0.70786	0.00001
	ASH-15-I	3048-3300?		320	0.70788	0.00001
	ASH-15-J	3048-3300?		326	0.70788	0.00001
Ma'ale-Ktora Cracks	KTO(1)-1-J	2713	60	180	0.70789	0.00001
	KTO(1)-1-K (I)	3317	63	100	0.70767	0.00001
	KTO(1)-1-K (II)	3317	63	104	0.70768	0.00001
	Average BM				0.70783	0.00008
Phreatic Speleothems						
Ma'ale-ha-Meyshar Cave	MMR-12-B	?		160	0.70771	0.00001
	MMR-12-C	?		286	0.70773	0.00001
	MMR-12-E up	?		300	0.70771	0.00001
	MMR-12-E up (1-8)	?		299	0.70777	0.00001
	MMR-12-E low (9,10)	?		405	0.70777	0.00001
	Average MMR-12				0.70774	0.00003
Makhtesh-ha-Qatan	MKTC-5-C	157.21	3.78	2140	0.70786	0.00001
Carbonate Host Rocks:						
Ma'ale-Efrayim	AMMINADAV-ME	Cenomanian		80	0.70756	0.00001
Hol-Zakh	SHIVTA-HZ	Turonian		99	0.70763	0.00001
overly the cave bearing formations:	MENUCHA-ROAD224	Senonian		950	0.70759	0.00001
Ashalim	SHIVTA-CHALUKKIM	Turonian		107	0.70750	0.00001
Ashalim	NEZER-CHALUKKIM	Turonian		118	0.70771	0.00001
overly the cave bearing formations:	MOR-NAFCHA	Eocenian		596	0.70778	0.00001
Even-Sid-Ramon	TAMAR MR-1	Cenomanian		78	0.70753	0.00001
Ma'ale-ha-Meyshar	TAMAR MMR	Cenomanian		148	0.70764	0.00001
Ma'ale-Ktora and Shizafor	GROFIT-KTO	Turonian		200	0.70741	0.00001
	Average carbonate rock				0.70759	0.00011

Table 4 - Sr data, page 4: bulk soils and their silicate fractions.

Soils above the caves (bulk):	Sample	Age (ka)	$\pm 2 \sigma$	Sr concentration (ppm)	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm 2 \sigma$
Mildly-Arid Zone:	Terra-Rossa soil samples:					
Ma'ale-Dragot	AMASA-TR-1	Present		161	0.70999	0.00001
Ma'ale-Dragot	AMASA-TR-2	Present		186	0.70981	0.00001
Ma'ale-Efrayim	ME-TR	Present		121	0.71099	0.00001
	Average Terra-Rossa soil (bulk)				0.71026	0.00052
Arid Zone:	Loess soil samples:					
Tzavoa, Mishash formation	TZ-MISH-LOESS	Present		374	0.70889	0.00001
Tzavoa, Shivta formatior	TZ-SHIV-LOESS	Present		314	0.70909	0.00002
Ashalim	ASH-LOESS	Present		333	0.70881	0.00001
Even-Sid-Ramon (above)	ESID-LOESS	Present		397	0.70879	0.00001
Even-Sid-Ramon (1.5 km from)	RUCHOT-LOESS	Present		408	0.70841	0.00001
	Average Loess soil (bulk)				0.70880	0.00022
Hyper-Arid Zone:	Hamada soil samples					
Ma'ale-ha-Meyshar	MMR-SOIL-1	Present		339	0.70878	0.00001
Ma'ale-ha-Meyshar	MMR-SOIL-2	Present		399	0.70859	0.00001
Shizafon	SHACH-HAMADA	Present		382	0.70896	0.00001
	Average Hamada soil (bulk)				0.70878	0.00015
Paleosoils washed to the caves (silicates):						
Even-Sid-Ramon mini-caves	ESID-paleosol	?			0.70977	0.00002
Ma'ale-ha-Meyshar	MMR-paleosol	?			0.71397	0.00002
Cave in road cut near Even-Sid-Ramon	STRAM-loess	?			0.71027	0.00002
Soils above the caves (silicates):						
Mildly-Arid Zone:	Terra-Rossa soil samples:					
Jordan Valley						
Ma'ale-Efrayim	ME-TR	Present			0.71228	0.00001
Northern Negev						
Ma'ale-Dragot	AMASA-TR-1	Present			0.71063	0.00001
Ma'ale-Dragot	AMASA-TR-2	Present			0.71064	0.00001
Arid Zone:	Loess soil samples:					
Tzavoa, Mishash formation	TZ-MISH-LOESS	Present			0.71043	0.00002
Tzavoa, Shivta formatior	TZ-SHIV-LOESS	Present			0.71085	0.00001
Ashalim	ASH-LOESS	Present			0.71009	0.00002
Even-Sid-Ramon (1.5 km from)	RUCHOT-LOESS	Present			0.71072	0.00001
Even-Sid-Ramon (above)	ESID-LOESS	Present			0.71105	0.00001
	Average soils northern and central Negev (silicates)				0.71063	0.00028
Hyper-Arid Zone:	Hamada soil samples					
Ma'ale-ha-Meyshar	MMR-SOIL-1	Present			0.71110	0.00001
Ma'ale-ha-Meyshar	MMR-SOIL-2	Present			0.71080	0.00001
Shizafon	SHACH-HAMADA	Present			0.71150	0.00001
	Average Hamada soil (silicates)				0.71113	0.00029
	Average Negev Soils (silicates):				0.71088	0.00025

NHP. לכן תא הלהץ הגבוה הסובטרופי מעל האוקיינוס האטלנטי היה מפותח יותר וגרם לשקעים אטלנטיים – ים-תיכוניים לנוע במסלול צפוני יותר, כאשר הוא מונע מהמשקעים ממקור ים-תיכוני להגיע לנגב.

יחסי $^{87}\text{Sr}/^{86}\text{Sr}$ בספלאוטמים במדבר "צל הגשם" ובנגב בדרך כלל נעים בין 0.7082 לבין 0.7085 ומשתנים רק במעט בין תקופות הקרח לבין תקופות בינו-קרחוניות, בהשוואה לספלאוטמים במרכז ישראל. השינוי הקטן יחסית מצביע על בליה מועטה של סלע האם. ערכים אלה מצביעים על אספקת אבק אינטנסיבית ממקורות קרובים יחסית, סיני ודלתת הנילוס, בהם יחסי $^{87}\text{Sr}/^{86}\text{Sr}$ קרובים ל - 0.7084.

מרבית הספלאוטמים בנגב המרכזי והדרומי הם בעלי חתכים עבים ועתיקים מטווח התיארוך בשיטת U-Th. גילים ראשוניים בשיטת U-Pb של מרבית ספלאוטמים האלה הם 2.7-3.3 מיליון שנה. שקיעתם התרחשה במקביל להיווצרות אגמי תצורת ערבה הפליוקנית. מקצת הספלאוטמים שקעו לפני 1.3~ מיליוני שנים במקביל לאגמי תצורת צחיחה שגילם פליסטוקן מוקדם. ירידה בקצב שקיעת הספלאוטמים ביחד עם עלייה ביחסי $^{87}\text{Sr}/^{86}\text{Sr}$ שלהם לפני כ-2.7 מיליוני שנים מעידים כנראה על תהליך המידבור בנגב ובכל הרצועה הסהרו-ערבית, שלוותה בעליה בכמויות האבק שמקורן היה כנראה בסיני ובמדבר סהרו-ערבי. מחזורים רבים של השקעת קלצית והפסקות גידול שהתרחשו לסירוגין באותו ספלאוטם לאורך יותר משלושה מיליוני שנים, מראים כי באזורים ארידיים הסדקים שמובילים את המים מפני השטח אל תוך המערה נשארים פתוחים לאורך תקופות זמן ארוכות.

סיני והנגב היוו כנראה את הגשר היבשתי העיקרי, ואולי היחיד מאפריקה לאסיה מאז תקופת המיוקן, בעוד שהמדבר הסהרו-ערבי היווה מחסום משמעותי בהתפשטותם של ההומינידים ובעלי החיים ממרכז אפריקה לאסיה ולשאר חלקי העולם. מחקר זה מציע כי תקופות לחות בסיני ובנגב שהתרחשו ביחד עם תקופות לחות בדרומה ובמרכז של רצועת המדבר הסהרו-ערבית פתחו "חלונות אקלימיים" לנדידת ההומינידים ובעלי החיים אל מחוץ ליבשת אפריקה. נדידה של האדם המודרני הקדום מאפריקה ללבנט הייתה קשורה כנראה ל-NHP-1 לפני 109-142 אלף שנה.

למינות הספלאוטמים והערכים הגבוהים יחסית של ה- $\delta^{13}\text{C}$, שירדו אך במעט במהלך האפיזודות הלחות, ומצביעים על התפתחות מעטה בלבד של צמחיה.

מקור המשקעים במהלך NHP-3, NHP-2 ו-NHP-1 היה ים התיכון המזרחי (שקעים אטלנטיים-ים-תיכוניים), כמו שמעידים ערכי δD באינקלוזיות הנוזליות בספלאוטמים במהלך 220 אלף השנים האחרונות וכן הירידה בעובי חתכי הספלאוטמים מצפון לדרום. ערכי $\delta^{18}\text{O}$ בנגב נמוכים בהשוואה לספלאוטמים מצפון, מרכז הארץ ומבקעת הירדן כתוצאה מזיקוק ריילי במהלך נדידת הסופות, כנראה בגלל מרחק גדול יותר של המערות מהים.

קיים הבדל בולט בין זמני השקעת הספלאוטמים במרכז ודרום הנגב בעיקר בתקופות הבינקרחוניות, לבין השקעת הספלאוטמים בספר-המדבר בצפון הנגב, מדבר יהודה ובקעת הירדן בעיקר בתקופות הקרח. ההסבר לכך הוא שהשקעת הספלאוטמים בספר המדבר התרחשה בעקבות שילוב של מקדמי חלחול גבוהים בעקבות הטמפרטורות הנמוכות האופייניות לתקופות הקרח, שלוו בכמויות המשקעים גבוהות מ-200 מ"מ לשנה באזורים אלה. העוצמה הנמוכה יחסית של שקעים אטלנטיים-ים תיכוניים בתקופה זו גרמה לנגב המרכזי והדרומי להישאר יבש יחסית. לעומת זאת, באפיזודות הלחות בתקופות הבינקרחוניות עוצמת השקעים הייתה גבוהה יותר מאשר בתקופות הקרח, הגשמים הים תיכוניים הדרימו אל מרכז הנגב וגרמו לכמות המשקעים השנתית לעלות עד 300-350 מ"מ באזור של איזויטת ה-50 מ"מ של היום. הטמפרטורות הגבוהות ומקדמי החלחול הנמוכים הובילו לחלחול מועט יחסית של מים למערות, ולכן שקעו שכבות דקות בלבד של ספלאוטמים בכל אפיזודה לחה.

הירידות החדות ב- $\delta^{18}\text{O}$ של הספלאוטמים במערת שורק לפני 132 אלף שנה ולפני 126-127 אלף שנים במערת שורק, כמו גם לפני 200 אלף שנה במערת פקיעין, התרחשו בו-זמנית לשיאים בקצב השקעת ספלאוטמים בנגב במהלך NHP-1 ו-NHP-2. ככל הנראה כל האירועים האלו היו תוצאה של האפיזודות האקלימיות הלחות ביותר מעל הים התיכון המזרחי.

NHP התרחשו במקביל לתקופות של מונסון אפריקני חזק, המאופיינות באינדקס מונסון גבוה במיוחד ($\geq 51 \text{ cal/cm}^2 \times \text{day}$), ובמקביל להיווצרות שכבות הספרופל בים התיכון. התחזקות בו-זמנית של המונסון האפריקאי ושקעים אטלנטיים-ים תיכוניים מקורה כנראה בהיחלשותו של תא הלחץ הגבוה הסובטרופי מעל האוקיינוס האטלנטי. העדר השקעת הספלאוטמים בנגב בהולוקן ובתקופה בינקרחונית MIS-11 ניתן להסביר על ידי עוצמה נמוכה יותר של קרינת השמש ועל ידי אינדקס נמוך יותר של המונסון האפריקני בהשוואה ל-

בצפון הנגב במערת צבוע ליד ערד, הנמצאת על איזוויטה של 150-160 מ"מ גשם בשנה, מרבית הספלאוטמים שקעו גם כן בתקופות הקרח, בדומה לבקעת הירדן ומדבר יהודה. במהלך התקופות הבינקרוניות ובשיאי תקופות הקרח השקעת הספלאוטמים הייתה מוגבלת ואפיזודית, אך עדיין מעט יותר משמעותית מאשר במדבר "צל הגשם". לא התרחשה השקעת ספלאוטמים במערת צבוע לפני 117 עד 96 אלף שנה, לפני 92 עד 85 אלף שנה וכמו כן במהלך ההולוקן, לכן תקופות אלה היו יבשות בדומה לתנאים השוררים היום ואולי אף יותר. ספלאוטמים מגיל הולוקן תיכון-מאוחר נמצאו רק במערות מעלה-דרגות, הממוקמות על איזוויטות של 280-300 מ"מ בצפון הנגב.

ערכי $\delta^{18}\text{O}$ בספלאוטמים ממערות בצפון הנגב מראים מגמות דומות לאלו שנמדדו במערות באזור אקלים ים-תיכוני בצפון הארץ ובמרכזה, ומצביעים על ים התיכון המזרחי כמקור עיקרי של משקעים שירדו בצפון הנגב. עדות תומכת למקור ים-תיכוני הם ערכי δD שנמדדו באינקלוזיות הנוזליות של הספלאוטמים. ערכי $\delta^{13}\text{C}$ בנגב הצפוני מראים כי צמחייה מעל המערות הייתה בדרך כלל צמחיית ערבה מסוג C3+C4, כאשר לעיתים רחוקות התרחשו תקופות לחות בהן פלשו לאזור מינים ים תיכוניים מסוג C3, או שלהיפך, תקופות בהן התרחש מדבור. כללית, תנאי האקלים בנגב הצפוני היו יבשים יותר מאשר בבקעת הירדן. בנגב הצפוני התזוזה המקסימאלית דרומה של הגבול בין אזור אקלים ים תיכוני לבין אזור האקלים המדברי לא עלתה על 25 ק"מ אפילו בתקופות הלחות ביותר.

למרות שבספלאוטמים ממערות קנאים, מדבר יהודה, רוב הספלאוטמים שקעו בתקופות הקרח, העובי הדק של החתכים של ספלאוטמים אלו מראה על כך שאזור מדבר יהודה היה ברוב התקופות יבש יותר מאשר בקעת הירדן או הנגב הצפוני.

תקופות לחות עיקריות בנגב המרכזי והדרומי (Negev Humid Periods - NHP) התרחשו במהלך 350 אלף השנים האחרונות לפני 310-350 אלף שנה (NHP-4), 290-310 אלף שנה (NHP-3), 190-220 אלף שנה (NHP-2), ו-109-142 אלף שנה (NHP-1). NHP-4 היא התקופה היחידה שהתאפיינה בהשקעת ספלאוטמים אינטנסיבית בנגב הדרומי בהשוואה לנגב המרכזי. סביר לכן, שמקור הגשמים בנגב הדרומי היה טרופי. שלושת התקופות הלחות הצעירות יותר ששוחזרו בנגב המרכזי והדרומי התאפיינו בהשקעת ספלאוטמים אינטנסיבית בנגב המרכזי בלבד ורק NHP-1 שהייתה הלחה ביותר מבין השלוש הובילה להשקעת ספלאוטמים מעטה גם בנגב הדרומי. תקופה אחת בלבד מבין הארבע (NHP-3) התרחשה בתקופת הקרח (MIS-8). האפיזודות הלחות המאפיינות את ה- NHP היו קצרות, כפי שמעידים העובי הדק של

תקציר

במחקר זה נעשה שיחזור פליאואקלים בגבולה הצפוני של רצועת המדבר הסהרו-ערבית, באמצעות משקעי מערות קרבונטיים (ספלאוטמים) ממערות באזורי המדבר של ישראל. בישראל קיימים שלושה אזורים אקלים עיקריים: אקלים ים תיכוני (עם כמות משקעים מעל 350 מ"מ בשנה, עם חורפים גשומים וקרירים וקייצים חמים ויבשים); אקלים ערבה וספר-המדבר עם כמויות משקעים שנתיות שבין 350 מ"מ לבין 150 מ"מ ואקלים מדברי ומדברי קיצוני בדרום הארץ ובמזרחה עם כמות משקעים שנתית הנעה מ-150 מ"מ ועד 30 מ"מ. את המדבריות בישראל ניתן לחלק לשני סוגים: הנגב, המהווה חלק מרצועת המדבר הסהרו-ערבית ומדבר "צל הגשם" ממזרח לשדרת ההר המרכזית, בבקעת הירדן, מדבר יהודה וים המלח. הימצאות ספלאוטמים במערות רבות באזורי המדבר בישראל מעידה על תנאי אקלים לחים יותר בעבר (כלומר, כמות משקעים אפקטיבית/מקדם חלחול חיוביים).

תקופות השקעת הספלאוטמים תוארכו במחקר זה באופן מדויק וברזולוציה גבוהה בשיטת U-Th ושימוש כאינדיקטורים לתקופות לחות. מקור המשקעים, כמותם וסוגי הצמחייה מעל המערות שוחזרו על ידי מדידת ערכי $\delta^{18}\text{O}$ ו- $\delta^{13}\text{C}$ בניצב לטבעות הגידול של הספלאוטמים ובאמצעות ערכי δD של האינקלוזיות הנוזליות הכלואות בתוכם. שיחזור של מקורות האבק נעשה על סמך ריכוזי Sr ויחסי $^{87}\text{Sr}/^{86}\text{Sr}$ בספלאוטמים.

בעבודה זו נעשה אומדן של כמויות המשקעים המינימאליות הנחוצות להשקעת ספלאוטמים, והן 200-275 מ"מ לשנה בתקופות הקרח ו-300-350 מ"מ לשנה בתקופות הבינקרחוניות. במערת מעלה אפריים בבקעת הירדן ובמערת קנאים במדבר יהודה השקעת הספלאוטמים התרחשה בעיקר במהלך שלושת תקופות הקרח האחרונות, עם השקעה מוגבלת של ספלאוטמים במעבר מתקופת הקרח MIS-6 לתקופה הבינקרחונית MIS-5 לפני כ-135 אלף שנה (המכונה Termination-II) ולפני 205-225 אלף שנה. לא שקעו ספלאוטמים במהלך רובה של תקופה בינקרחונית MIS-5, בשיא תקופת הקרח האחרונה ובתקופת ההולוקן. הפרופילים האיזוטופיים של $\delta^{18}\text{O}$ ו- $\delta^{13}\text{C}$ בספלאוטמים ששקעו במערת מעלה-אפריים בבקעת הירדן לפני 67 עד 25 אלף שנה דומים מאוד לאלו של הספלאוטמים בצפון הארץ ובמרכזה (מערות שורק, ירושלים ופקיעין), ומעידים על מקור משקעים משותף והוא הים התיכון המזרחי. משקעי מערת מעלה-אפריים מראים כי בתקופת הקרח האחרונה נדד גבול המדבר בבקעת הירדן דרומה ממקומו הנוכחי. ירידה בטמפרטורות ובעוצמת האידוי ועלייה בכמות המשקעים האפקטיבית בבקעת הירדן בעקבות כך, הם הגורמים העיקריים לדעיכת אפקט "צל הגשם", בבקעת הירדן בתקופות הקרח.



משרד התשתיות הלאומיות
המכון הגיאולוגי

שיחזור פלאואקלים של תקופת הרביעון בגבולו הצפון-מזרחי של מדבר הסהרה בעזרת הספלאוטמים של הנגב

ווקס אנטון

עבודת זו הוגשה כחיבור לקבלת תואר "דוקטור לפילוסופיה"
לסנאט האוניברסיטה העברית, ירושלים.

העבודה נעשתה בהדרכתם של:

ד"ר מרים בר-מטיוס, המכון הגיאולוגי, ירושלים

פרופ' אלן מטיוס, המכון למדעי כדור הארץ, האוניברסיטה העברית, ירושלים

פרופ' עמוס פרומקין, המחלקה לגיאוגרפיה, האוניברסיטה העברית, ירושלים