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A review of palaeoclimates and palaeoenvironments in the Levant and Eastern Mediterranean from 25,000 to 5000 years BP: setting the environmental background for the evolution of human civilisation

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Abstract

The southern Levant has a long history of human habitation and it has been previously suggested that climatic changes during the Late Pleistocene–Holocene stimulated changes in human behaviour and society. In order to evaluate such linkages, it is necessary to have a detailed understanding of the climate record. We have conducted an extensive and up-to-date review of terrestrial and marine climatic conditions in the Levant and Eastern Mediterranean during the last 25,000 years. We firstly present data from general circulation models (GCMs) simulating the climate for the last glacial maximum (LGM), and evaluate the output of the model by reference to geological climate proxy data. We consider the types of climate data available from different environments and proxies and then present the spatial climatic "picture" for key climatic events. This exercise suggests that the major Northern Hemisphere climatic fluctuations of the last 25,000 years are recorded in the Eastern Mediterranean and Levantine region. However, this review also highlights problems and inadequacies with the existing data.

1. Introduction

The effects of global climate change on human society, the environments we inhabit and the sustainable development of our planet's poorest people are of prime concern to all. Predicting changes in water availability, local environments and climates will be the key to determining which areas of the world will require greatest assistance in dealing with increased global warmth and climate change in the coming century, and beyond. However, the detailed linkages between the development of human civilisation, anthropogenic activities, climate and environmental change still remain poorly understood. Through a more detailed study of how climate and environmental changes influenced the history of human civilisation, it is hoped that a better understanding of the relationship between these key elements can be deduced. The Levantine region (encompassing the modern countries of Israel, Jordan, Lebanon, Syria and parts of Egypt and Saudi Arabia; Fig. 1) has a long history of human habitation (e.g. Issar and Zohar, 2004), making it an ideal area to study linkages between climatic, environmental and societal change.

Palaeoclimatic records, such as those obtained from ice cores, demonstrate the generalised climatic evolution of the last 25,000 years (Fig. 2). The long-term trend during this interval is from the cold, glacial conditions of the Late Pleistocene to the warm, interglacial conditions of the Holocene. This long-term trend is punctuated by several shorter climatic events (Heinrich events, the Younger Dryas and the "8.2 ka cold event") that seem to have had a significant effect on climatic conditions in much of the Northern Hemisphere. The origin of these events may be linked to the rapid input of cold freshwater into the North Atlantic causing a disturbance in oceanic circulation and local climatic regimes (e.g. Broecker and Denton, 1989; Bond et al., 1992, 1993; Alley, 2000; Alley and Ágústsdóttir, 2005; Rohling and Pälike, 2005). The Levantine region straddles the arid/semi-arid boundary and is thus highly

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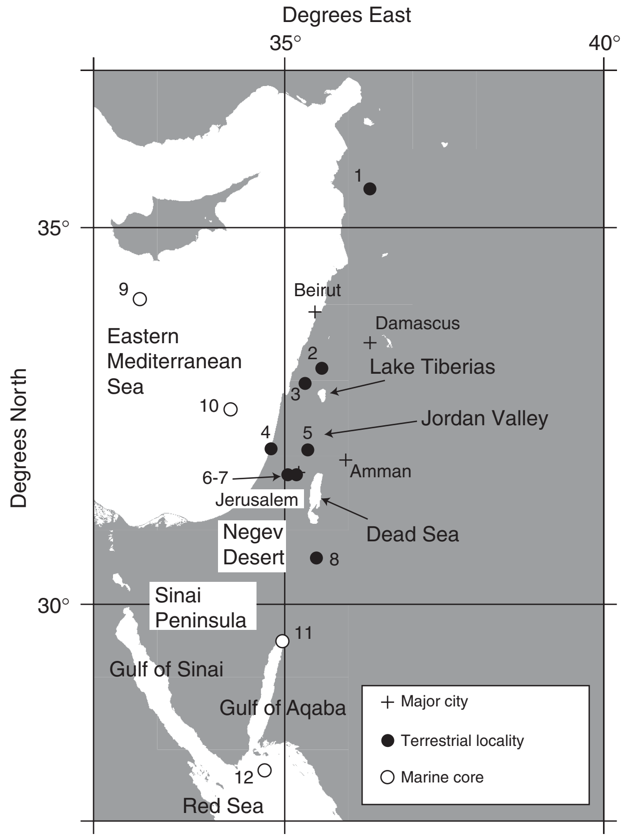

Fig. 1. Map of the Eastern Mediterranean and Levant region showing the locality of major features and sites discussed in the text. 1: Ghab Valley, 2: Hula Basin, 3: Peqiin Cave, 4: Israeli coastal plain, 5: Ma'ale Efrayim Cave, 6: Soreq Cave, 7: Jerusalem West Cave, 8: Wadi Faynan, 9: Ocean Drilling Program (ODP) Site 967, 10: Site of core M44-1-KL83, 11: Site of core GeoB5804-4, 12: Site of core GeoB5844-2. Map produced with GMT (http://gmt.soest.hawaii.edu/).

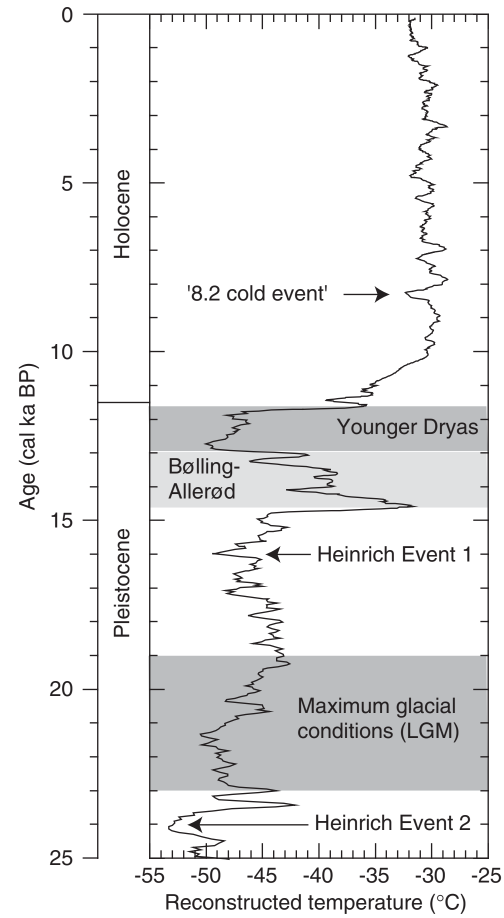

Fig. 2. Reconstructed air temperatures from the GISP 2 Ice core in Greenland (after Alley, 2000). The timing and duration of the last glacial maximum (LGM) is the same as the "LGM Chronozone Level 1" as defined by Mix et al. (2001).

sensitive to climatic changes. Therefore, rapid climatic changes in the past may have had sufficient effect on local environmental conditions to cause adaptation in human behaviour and society.

In the Levant region, where the evidence for human civilisation extends furthest back in time, it has been suggested that there are causal links between major Northern Hemisphere climatic events and the evolution of human civilisation (e.g. as reviewed in: Moore and Hillman, 1992; Harris, 1996; deMenocal, 2001; Mithen, 2003; Issar and Zohar, 2004). The development of hunter-gather sedentism and the earliest semi-permanent settlements in the Levant occurred during the late Glacial Interstadial, known as the Bølling-Allerød warm interval (e.g. Bar-Yosef and Belfer-Cohen, 1992; Moore and Hillman, 1992; Bar-Yosef, 2000). During the Younger Dryas human communities in the Levant generally returned to a hunter-gatherer existence (e.g. Bar-Yosef and Belfer-Cohen, 1992; Bar-Yosef, 2000), possibly as a response to colder, drier climatic conditions. The "Neolithic revolution", including the advent of farming and the first villages, occurred after the Younger Dryas during the warm early Holocene. Unfavourable climatic conditions at about 8.2 and 4.2 cal ka BP have been invoked as the cause of the abandonment of PPNB (Pre-Pottery Neolithic B) farming towns in the Jordan valley and the collapse of the Akkadian civilisation, respectively (e.g. Cullen et al., 2000; deMenocal, 2001; Mithen, 2003).

A number of authors have reviewed various aspects of Late Quaternary palaeoclimates and palaeoenvironments in the Eastern Mediterranean region (e.g. Goodfriend, 1999; Issar, 2003; Issar and Zohar, 2004), but these reviews have often focused on specific time periods (i.e. the Holocene), environments or sets of techniques. The aim of our work is to provide a thorough, yet accessible, review of the current state of our knowledge of palaeoclimates, and palaeoenvironments in the context of palaeoclimatic changes, throughout the Levant and Eastern Mediterranean Sea from 25 to ~5 cal ka BP. We focus on this time period as it represents (i) the interval in which the magnitude of environmental and climatic change was largest (and therefore most easily observed in the geological record); (ii) the period during which human behaviour changed from a mobile to sedentary lifestyle and (iii) is the least subject to anthropogenic environmental change, thereby allowing a clearer examination of the "natural" effects of climate change on human behaviour. After 5 cal ka BP, the magnitude of climate changes is much smaller and it is deemed inappropriate to discuss these here. We note however the possibility of a smaller magnitude event at 4.2 cal ka BP (e.g. Cullen et al., 2000; deMenocal, 2001), but its impact was significant for, by now, large, urbanised communities such as those in Egypt and Akkadia. Because of the unique climatic setting of the Levant, we have restricted our review to the region shown in Fig. 1. Reviews of palaeoclimatic conditions in the adjoining areas (i.e. Turkey and North Africa) during the Late Pleistocene–Holocene have been published elsewhere (e.g. Roberts and Wright, 1983; Gasse and Fontes, 1992; Roberts et al., 2004; Verschuren et al., 2004). We review the different types of palaeoclimatic data available for both terrestrial and marine environments, and synthesise these data for key intervals. In the course of this exercise we pay particular attention to the quality, reliability and accuracy of the various proxies and their age models. This study provides an overview of climatic conditions in the region and allows the identification of key geographic and stratigraphic gaps.

2. Dating

A key aim of this review is to compare and contrast various proxies, which have been dated by a variety of techniques. All ages reported in this paper are in calendar (calibrated) years (or ka) before present (cal years BP or cal ka BP). Uncalibrated radiocarbon ages (^{14}C ages) have been recalibrated using OxCal v. 3.10 (Bronk Ramsey, 1995, using atmospheric data from Reimer et al., 2004) to allow comparison with calendar ages (such as those from ice cores) and uranium-series ages (principally from carbonates). It should be noted that ^{14}C calibration using tree rings extends currently to around 13 cal ka BP; calibrated ages for the preceding Late Pleistocene period are much less secure, so that the possibility of mis-correlation increases.

3. Climate models

In recent years general circulation models (GCMs) have been used to evaluate past climates. These models use the laws of physics and an understanding of past geography to simulate climatic responses. They are objective in character and it is now possible to compare results from different GCMs for a range of times and over a wide range of parameterisations for the past, present and future (e.g. in terms of predictions of surface air temperature, surface moisture, precipitation, etc.)

We present here some outputs on climate for the last glacial maximum (LGM) from model simulations generated using the HadAM3 version of the UK Meteorological Office (UKMO) coupled atmosphere–ocean GCM (Pope et al., 2000). The model was developed at the Hadley Centre for Climate Prediction and Research, which is part of the UKMO. The LGM simulation follows the protocol of the Palaeoclimate Model Intercomparison Project (PMIP, see Joussaume and Taylor, 1995). The boundary conditions on the model include the CLIMAP sea surface temperature (SST) reconstructions (CLIMAP, 1981) and Peltier (1994) ice sheet reconstructions. The climate model provides new insights for the formulation of working hypotheses, providing a range of parameterisations, some of which can be expected to leave geological proxy signatures (e.g. temperature, precipitation, salinity) and others which, though significant, may not (e.g. snowfall, cloud cover, wind velocities, surface pressure). Lakes (as closed systems) respond through time to changes in the balance between precipitation and evaporation (P–E). Similarly rainfall, combined with the overall temperature regime will influence palaeosol generation. The geological record of such sediments can be compared directly with climate model outputs.

We have run simulations for the LGM and it is instructive to compare the outputs from the model with modern observations (Figs. 3 and 4). Most precipitation today occurs in the Eastern Mediterranean during the winter months (December, January and February, DJF; Fig. 3). In the far north of the area, precipitation, mostly from the Atlantic, falls as rain (and some snow) at between 2 and 4 mm/day (Anatolian Uplands northwards of 36°N). There is a rapid decline in the amount of precipitation southwards (see Fig. 3), falling to between 0.5 and 1.0 mm/day around 30°N. This marked contrast in rainfall, between N and S, has been noted by previous authors and, by reference to palaeoclimatic geological proxy data, is believed to have operated in past times too (e.g. Enzel et al., 2003 and references therein). Average winter (DJF) temperatures range between 8 and 12 °C. During the summer months (June, July and August, JJA; Fig. 3) virtually the whole region experiences virtually no precipitation and high temperatures (Fig. 4). In the south, and over the Dead Sea, there is a marked excess of evaporation over precipitation which, coupled with the major abstraction of water from the Jordan system, has resulted in Dead Sea levels dropping by around 0.5 m per annum in recent years. The annual average rainfall between about 31°N and 36°N is around 180 mm/yr, rising along the coastal strip as far east as Jerusalem, to around 400–450 mm per annum (Fig. 3).

Model results for the LGM at first sight seem very similar to the observations for the present, with most precipitation modelled to occur during the winter (DJF; Fig. 3), but with 1–2 mm/day precipitation extending between 37°N and 31°N, particularly adjacent to the coastal strip. Summers, like today, are modelled to be very dry throughout most of the area (Fig. 3). However, during the LGM, much of the winter precipitation over the Anatolian uplands falls as snow, released in a major spring thaw (Fig. 3G and 3H). The resolution of the model (2.5° in latitude and 3.75° in longitude) precludes depiction of the Mt. Hermon Jordan catchment as a discrete area, but it too was likely to have received very heavy snowfalls.

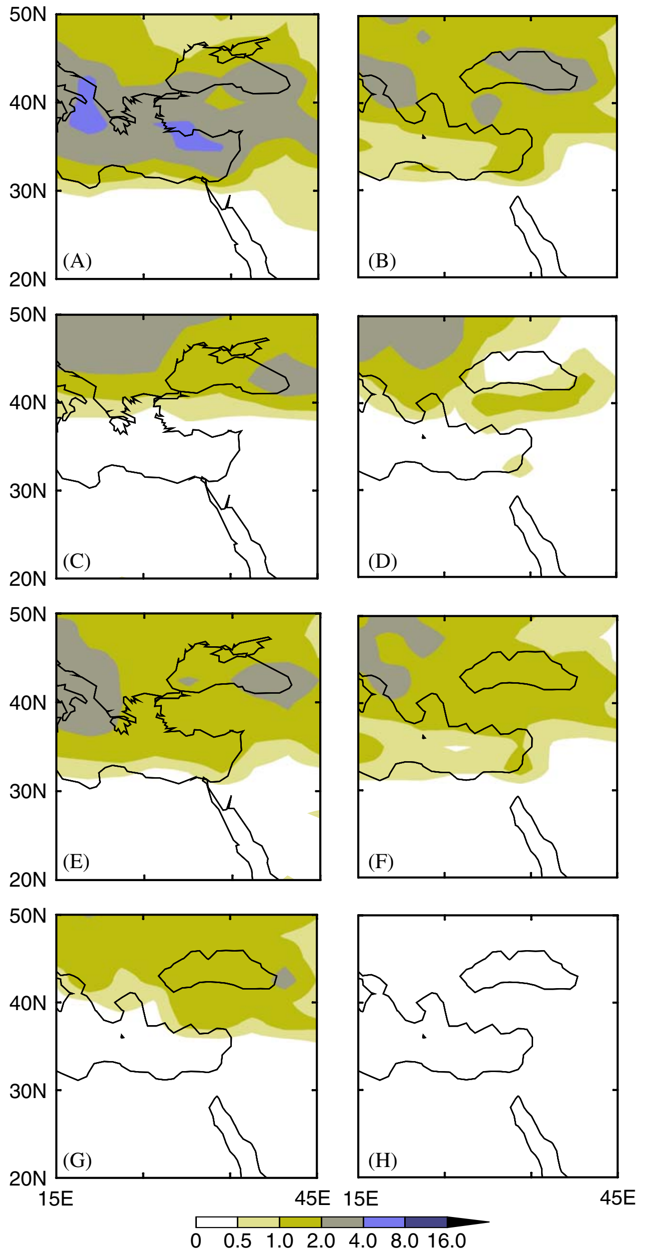

Fig. 3. Climate model outputs for the LGM and the present day. A: present day winter (DJF) precipitation (precipitation in mm/day); B: LGM winter (DJF) precipitation; C: present day summer (JJA) precipitation; D: LGM summer (JJA) precipitation; E: present day annual precipitation; F: LGM annual precipitation; G: LGM winter (DJF) snowfall (snowfall in mm/day); H: LGM summer (JJA) snowfall. Note that panel H is blank because there is no snow fall in summer.

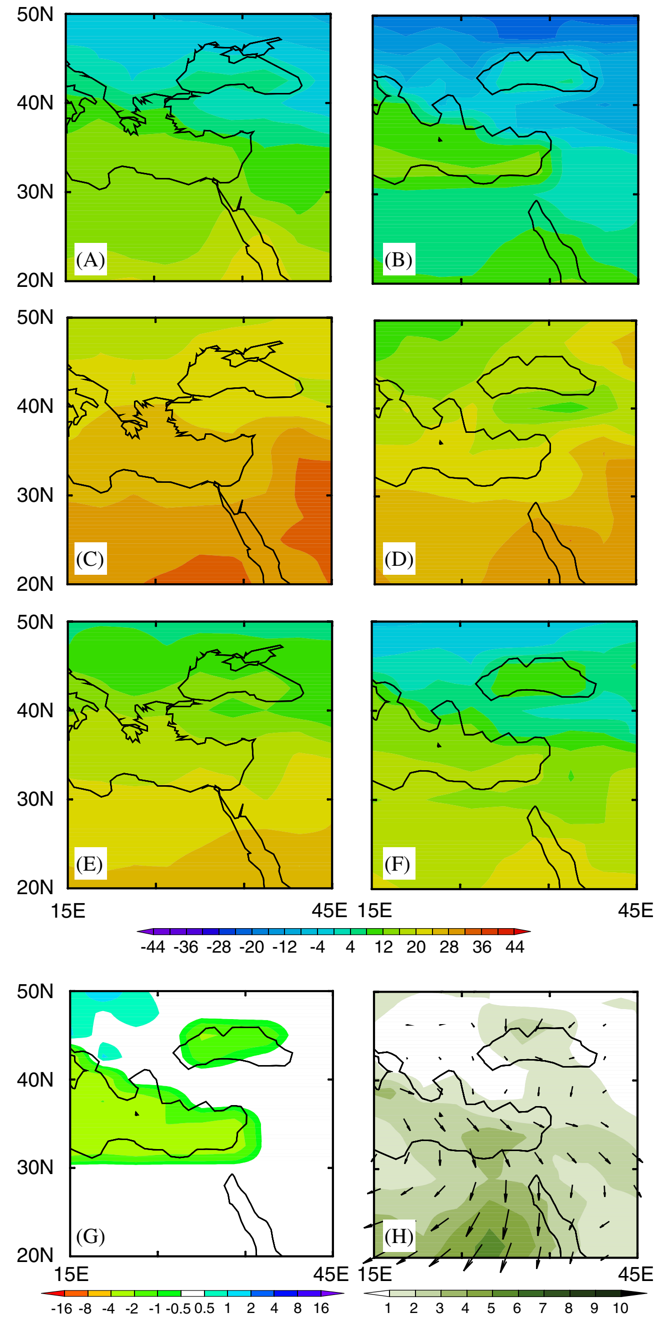

Fig. 4. Climate model outputs for the LGM and present day. A: present day winter (DJF) temperature (= FBC); B: LGM winter (DJF) temperature; C: present day summer (JJA) temperature; D: LGM summer (JJA) temperature; E: present day average annual temperature; F: LGM average annual temperature; G: LGM annual average precipitation minus evaporation in mm/day; H: LGM annual average wind strength (in m/s) and vectors.

Anatolia is modelled to receive some rain even during the summer months (Fig. 3). The overall pattern is for a small but significant increase in precipitation across the northern parts of the area. Modelled temperatures are much lower than present (Fig. 4), and account for less evaporation, but still evaporation exceeds precipitation (Fig. 4G). The lower temperatures help to explain how lake levels (e.g. Lake Lisan) could have grown during the glacial phase, even though the Earth climate system was generally colder and drier. Winds, during the LGM are modelled to be out of the north and northwest, generally strengthening southwards (Fig. 4H). We will now examine the climate record from 25 to 5 cal ka BP, as recorded by a range of geological data, and consider model performance for the LGM in light of these, so that its possible application to intervening times might be evaluated.

4. Description of the climate records

4.1. Non-marine sedimentary records

4.1.1. Lacustrine sediments and lake levels

The long-term rise and fall of lake levels is the result of the interplay between climate (controlling precipitation, run-off, evaporation and global sea-level) and tectonics (controlling basin shape, catchment area and sill depths). Although the Jordan Rift Basin is tectonically active, and has been throughout the Pleistocene and Holocene (e.g. Marco et al., 1996; Enzel et al., 2000; Migowski et al., 2004), most authors have suggested that climate was the dominant control on lake levels in the rift valley (e.g. Begin et al., 1985; Frumkin et al., 1991, 1994, 2001; Yechieli et al., 1993; Neev and Emery, 1995; Stein, 2001; Bartov et al., 2002, 2003; Frumkin and Elitzur, 2002; Landmann et al., 2002; Enzel et al., 2003; Hazan et al., 2005). During the last glacial Lake Lisan covered much of the Jordan Valley (e.g. Neev and Emery, 1967) encompassing the present-day Dead Sea and Lake Tiberias (also known as Sea of Galilee or Lake Kinneret). Sometime in the late glacial (probably after the Younger Dryas), Lake Lisan shrank to a configuration similar to the present situation. Lake levels have been reconstructed by combined chronostratigraphy and facies analysis (e.g. Begin et al., 1985; Yechieli et al., 1993; Neev and Emery, 1995; Bartov et al., 2002, 2003; Landmann et al., 2002; Hazan et al., 2005) and dating of flood sediments in elevated saltcaves and karstic topography (e.g. Frumkin et al., 1991, 1994, 2001; Frumkin and Elitzur, 2002). Fig. 5 shows a compilation of different lake level curves for the interval from 25 cal ka BP to the present for both Lake Lisan and Lake Tiberias. All the studies agree that lake levels in the Lake Lisan/Dead Sea system were relatively high during the last glacial and relatively low during much of the Holocene. However, on a shorter time-scale (<5000 yr) there are clearly a number of inconsistencies between the different studies, which may be critical for understanding the palaeoclimatic significance of lake level curves in the region. A major problem is the difficulty in constructing high-resolution age models for sequences that do not contain sufficient datable materials and do contain a significant number of hiatuses by virtue of their marginal-lake setting. Additionally, the deposition of thick evaporite deposits has meant that some parts of the sections are not dated (e.g. Yechieli et al., 1993). Neev and Emery (1995) studied more distal lake-facies present in boreholes and constructed their lake-level curve by using downhole gamma-ray logs to identify evaporitic (low-stand) and siliciclastic (high-stand) sediments, but their age model is based upon a combination of ^{14}C dates that may be compromised by old carbon and climatic assumptions inferred from the archaeological record. By comparing and carefully considering the merits of the different datasets, we have produced an integrated reconstruction (thick line in Fig. 6) of Lake Lisan/Dead Sea levels to ease comparison between lake level data and other climatic proxies.

Maximum lake levels for the last glacial were achieved at about 25 cal ka BP (Bartov et al., 2002, 2003; Landmann et al., 2002). Both Bartov et al. (2003) and Landmann et al. (2002) suggest that a major lowering in Lake Lisan occurred at about 24 cal ka BP, approximately synchronous with Heinrich Event 2 (H2; 23.8 cal ka BP) although the magnitude of lake level fall is unclear (compare Landmann et al., 2002; Bartov et al., 2003). At the LGM, lake levels were high (compared to H2) but may have been falling during the peak of LGM conditions, synchronous with increased local aridity proposed at this time (see discussion below), and as would be predicted from GCM results (Fig. 4G). All the data sets seem to suggest the existence of a low stand centred about 16–17 cal ka BP, synchronous with Heinrich Event 1 (H1) in the North Atlantic (Bartov et al., 2003). Bartov et al. (2002, 2003) suggest that lake levels peaked after H1 at 15 cal ka BP. Both Neev and Emery (1995) and Landmann et al. (2002) suggest that a highstand was reached about 13 cal ka BP (during the Bølling-Allerød warm interval), although the magnitude and duration are vastly different in both studies. Between 13.2 and 11.4 cal ka BP a substantial halite layer was deposited in the current Dead Sea Basin (Neev and Emery, 1967; Yechieli et al., 1993), which Neev and Emery (1967) interpreted as the result of a lake level drop to ~−700 m. below present day sea level. Yechieli et al. (1993) correlate this extremely arid event with the Younger Dryas of Northern Europe. In contrast, Neev and Emery (1995) found no evidence for salt deposition at this time in the southern basin and suggested that lake levels were high during this period of time. However, the bracketing of ages around the salt layer in the northern basin and other evidence from speleothems and palaeosols (discussed in more detail below) suggests that the Younger Dryas in the Levant was more likely to have been an arid period marked by extremely low lake levels.

Frumkin et al. (1991, 1994) dated Holocene wood fragments found in flood sediments in elevated salt caves to determine the timing of high stands. A first high stand is indicated as occurring ~8.5 cal ka BP, roughly synchronous with a small highstand in the Neev and Emery (1995) curve. By 6.5 cal ka BP, the level of the Dead Sea had fallen sufficiently to result in the desiccation of the Southern Basin (Frumkin et al., 1991, 1994, 2001; Neev and Emery, 1995). Frumkin et al. (1991, 1994) and Neev and Emery (1995) appear to be broadly in agreement over the timing of relatively higher lake levels at about 5 cal ka BP.

Absolute lake levels in Lake Tiberias and Lake Lisan were similar in the last glacial (see Fig. 8 in Hazan et al., 2005) but started to diverge at ~19 cal ka BP for reasons that are currently unclear but are probably a combination of tectonics, basin morphology and climatic change. However, both lakes generally display similar variations in lake-level throughout the last 25,000 years (Fig. 5; Hazan et al., 2005). Maximum levels were attained in Lake Tiberias during the last glacial between 26 and 24 cal ka BP.

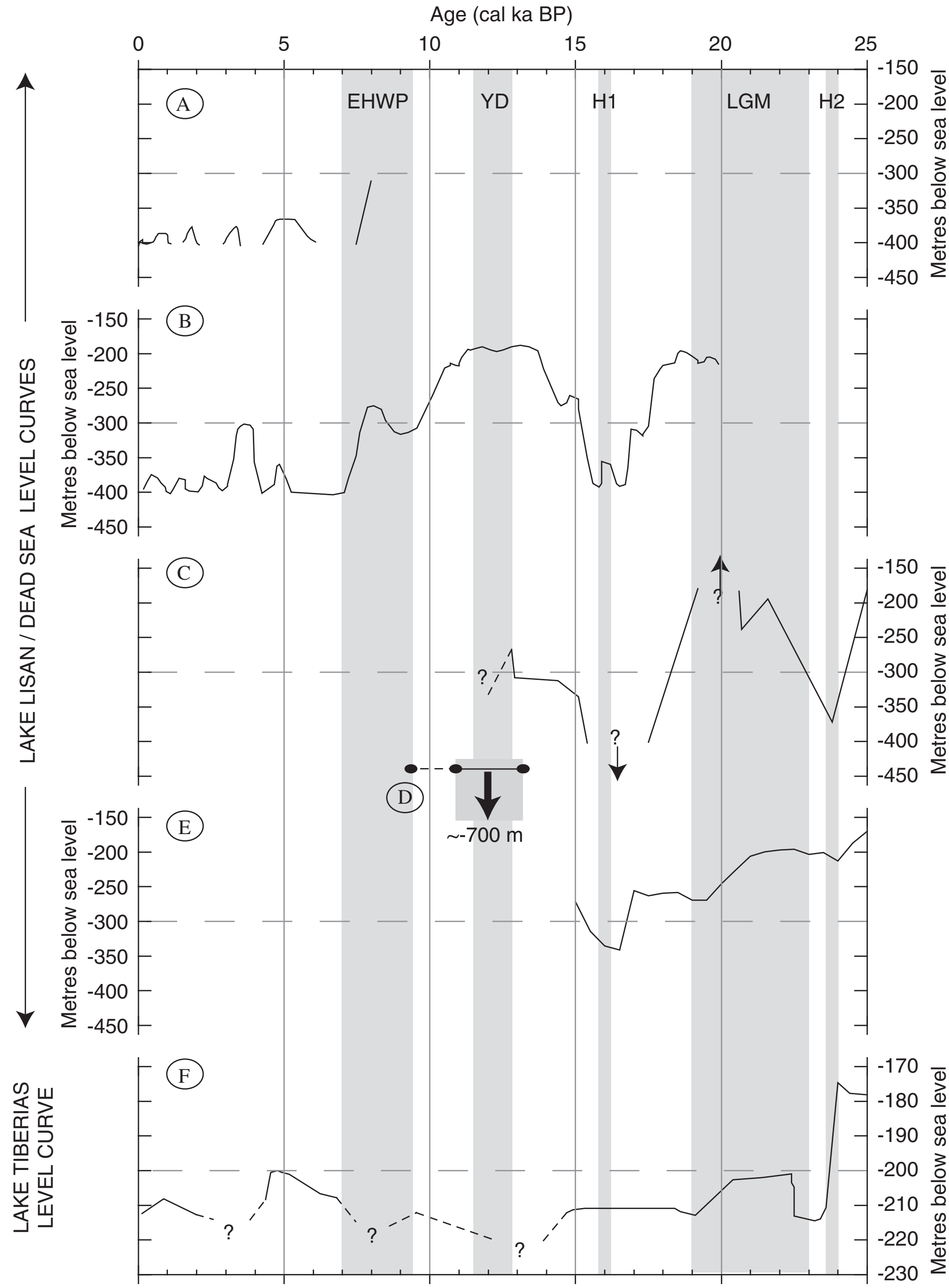

Fig. 5. Compilation of lake level curves for Lake Lisan/the Dead Sea and Lake Tiberias. A: Frumkin et al. (1994); B: Neev and Emery (1995); C: Landman et al. (2000); D: timing of massive salt deposition (Yechieli et al., 1993; Neev and Emery, 1967); E: Bartov et al. (2002, 2003); F: Hazan et al. (2004). EHWP = Early Holocene Wet Phase; YD = Younger Dryas; H1 = Heinrich Event 1; LGM = Last Glacial Maximum; H2 = Heinrich Event 2.

A major lowering of lake levels at 24 cal ka BP may have been a consequence of Heinrich Event H2. The level of Lake Tiberias rose slightly after this event, reaching a highstand at about 22.5 cal ka BP after which there was a gradual decline until about 15 cal ka BP. At that point in time, lake levels fell, although the timing and magnitude is uncertain. During the Younger Dryas Lake Tiberias was probably very low before rising in the early Holocene. Hazan et al. (2005) report a high stand at ~5 cal ka BP, synchronous with high lake levels in the Dead Sea.

4.1.2. Palaeosols

Palaeosol sequences on the coastal plain of Israel have been constrained by luminescence and radiocarbon dating (Gvirtzman and Wieder, 2001). Through examination of a number of soil characteristics (including magnetic susceptibility, particle-size distribution, clay mineralogy and soil micromorphology), Gvirtzman and Wieder (2001) classified sequences of palaeosols, aeolianites and dune sands for the last 53,000 years. The different soil-parent materials and soil types were rated on a semi-quantative ‘‘wet to dry’’

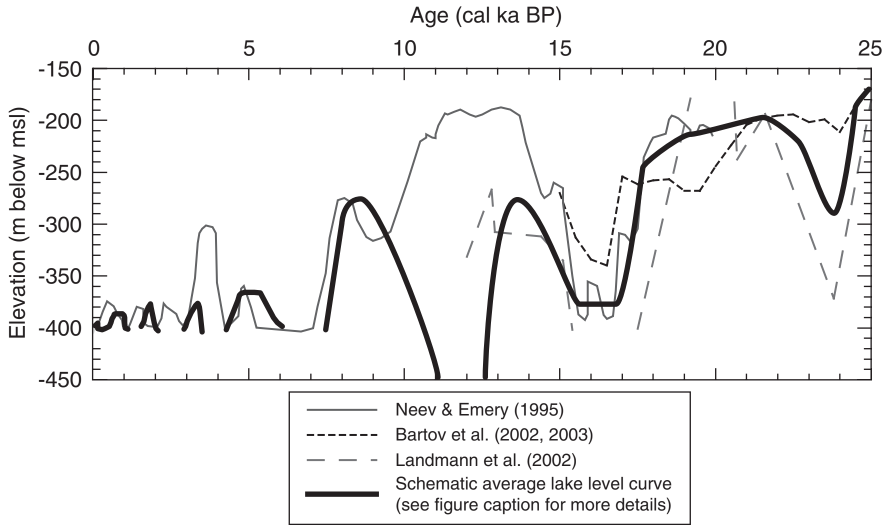

Fig. 6. An integrated, schematic lake level curve (thick bold line) for the Lake Lisan/Dead Sea based upon various studies. This curve is designed primarily to illustrate lake level trends over time for ease of comparison with other proxy data. For the period 25–13 cal ka BP the integrated curve is an approximate average of Neev and Emery (1995), Bartov et al. (2002, 2003) and Landmann et al. (2002). From 13 to 9 cal ka BP we have used the data from Neev and Emery (1967), Begin et al. (1985), Yechieli et al. (1993) and Stein (2001) which suggests a major lake level fall between 13 and 11 cal ka BP. From 9 cal ka BP onwards we have followed the study of Frumkin et al. (1994).

scale. The temporal pattern (Fig. 7) bears some similarities to other proxy data from speleothems and Lake Lisan/Dead Sea levels. However, because of the semi-quantative nature of the palaeosol record (and hence step-like function), some of the nuances of palaeoclimatic changes are lost. Furthermore, the duration of some palaesol events has been estimated on the basis of soil/sediment type, as no dates exist for those horizons (Fig. 7). It is striking to note that the palaeosol record suggests relatively wet (or moist) conditions during the Lateglacial (prior to 12.5 cal ka BP) and from 10 to 7.5 cal ka BP (Fig. 6) that requires an excess of precipitation over evaporation. Gvirtzman and Wieder (2001) suggest that rainfall must have been at least 350 mm/yr but may have been as high as 800 mm/yr (the GCM output for the LGM indicates 360–750 mm/yr with both strong latitudinal zonation and seasonality, Fig. 3F). However, these authors suggest that the cold, dry conditions during the peak of the LGM were not recorded by the palaeosol sequence. From 12.5 to 11.5 cal ka BP there was a rapid accumulation (~700 mm/yr) of atmospheric dust in the palaeosols (compared to 22–83 mm/yr in present day Israel). Gvirtzman and Wieder (2001) consider this aolianite to be the local record of the Younger Dryas interval and representative of increased dust transport from the Saharan and Arabian deserts.

4.1.3. Fluvial sediments

The study of fluvial environments has the potential to yield information about changes in hydrological regime and climate. However, hiatuses and a lack of datable materials can hamper such research. Although Late Pleistocene to Holocene fluvial sedimentation in the Levant has been discussed in terms of environmental change (e.g. Grossman and Gerson, 1987; Oguchi and Oguchi, 2004), the lack of absolute dates makes it difficult to compare these studies with other well-dated climate records. However, in the Wadi Faynan region of southern Jordan, McLaren et al. (2004) were able to identify and date a succession of fluvial terraces, alluvial fans and aeolian sediments, allowing them to place some constraint on the possible palaeoclimatic evolution of this region. Through a limited number of ^{14}C and OSL dates they were able to construct a relative stratigraphy that recorded some of the facies changes in the area during the Quaternary. Aeolian sediments (in places interbedded with fluvial units), dated to about 13.7 cal ka BP, suggest a dry climatic phase. Perennial meandering stream deposits, thought to be early Holocene in age, suggest that the climate was wetter at this time (possibly 9500–8000 cal yr BP; McLaren et al., 2004; Hunt et al., 2004). A wind blown unit indicates that after 7400 cal yr BP climatic conditions were increasingly arid. Unfortunately, the sequence, relative timing and absolute dating of facies changes in Wadi Faynan are extremely complicated and difficult to resolve and thus McLaren et al. (2004) were only able to draw very broad conclusions about the impact of climatic changes on fluvial facies changes.

4.2. Terrestrial palaeobotanical records

Temporal variations in vegetation are important indicators of climatic and environmental change. The best preserved and most informative palynological records come from continuous sedimentary sections with good age control. Lake sediments are often ideal for this purpose, whereas fluvial sediments are less desirable due to possible hiatus's and the increased potential for oxidation of organic matter. Additionally, ''natural''

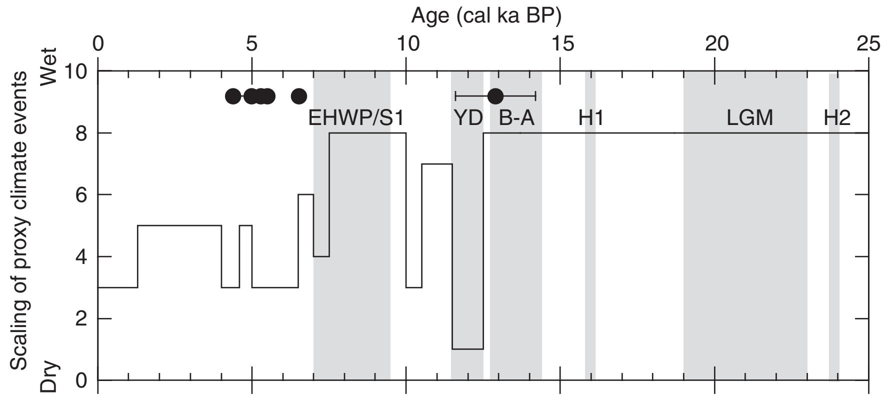

Fig. 7. Palaeoclimate of the Israeli coastal plain, as interpreted from palaeosols (Gvirtzman and Wieder, 2001). Black dots show position of age model tie points. S1 = Sapropel 1, B-A = Bølling-Allerød, other abbreviations as in Fig. 5.

sediment accumulations (i.e. lacustrine or fluvial sediments) are preferable to sediments that accumulate at archaeological sites, where the potential for an anthropogenically induced taphonomic bias is high. Because of this we decided only to include sites without human interference (i.e. natural sites) in this review and so have omitted much of the strictly archaeological record (e.g. Darmon, 1988; Miller, 1998).

High-resolution Late Pleistocene–Holocene continental palynological records from the Levant are available from the Ghab Valley in Syria (e.g. Niklewski and Van Zeist, 1970; van Zeist and Woldring, 1980; van Zeist and Bottema, 1982; Yasuda et al., 2000), the Hula Basin in northern Israel (e.g. Horowitz, 1971, 1989; Baruch and Bottema, 1991, 1999) and the Birkat Ram crater lake also in northern Israel (Weinstein, 1976; Schwab et al., 2004). Other more limited palynological records exist from fluvial (e.g. Hunt et al., 2004) and coastal marsh (e.g. Kadosh et al., 2004) sites, although these will not be discussed in detail in this review.

The Ghab Valley and Hula Basin records (Fig. 8) have formed the basis for most palaeovegetational histories and reconstructions in the Levant region (e.g. van Zeist and Bottema, 1991). However, controversy has existed for some time over the reliability of the age models for these records and the inferred sequences of vegetational changes in the Levant (e.g. Rossignol-Strick, 1995; Hillman, 1996; Meadows, 2005). The original published age models for the Ghab Valley (Niklewski and Van Zeist, 1970; van Zeist and Woldring, 1980; van Zeist and Bottema, 1982) and Hula Basin (Tsukada in van Zeist and Bottema, 1982; Baruch and Bottema, 1991, 1999) suggest that during the Younger Dryas, forest cover and humidity (as defined by arboreal pollen values) increased in Syria, while in northern Israel forest cover contracted and aridity increased (see Baruch and Bottema, 1991; Bottema, 1995; Rossignol-Strick, 1995). Baruch and Bottema (1991) and Bottema (1995) took the view that during the Younger Dryas two climatic subregions existed, unlike the present-day uniform climate regime. In contrast, Rossignol-Strick (1995) suggested that the age models for both the Ghab Valley and the Hula Basin were suspect. Rossignol-Strick (1995) presented palynological data from marine cores in the Mediterranean and the Western Arabian Sea. Through a combination of oxygen-isotope stratigraphy, radiocarbon dating and palynology, Rossignol-Strick (1995) showed that in marine cores the Younger Dryas was marked everywhere by an increase in Chenopodiaceae. This plant is associated with saline soils, arid conditions and areas with less than 100 mm mean annual rainfall. The "Chenopodiaceae Phase" was found to be synchronous in marine cores from the Mediterranean and Arabian Seas, suggesting that this was a regional phenomenon across the entire Levantine and Arabian area. Rossignol-Strick (1995) identified a regional "*Pistacia* Phase" in the early Holocene (10.2–6.7 cal ka BP) suggesting mild winters and mean annual precipitation between 300 and 500 mm. Rossignol-Strick (1995) identified the Chenopodiaceae and *Pistacia* Phases in the Ghab and Hula pollen records and was thus able to provide an alternative chronostratigraphy for these cores (Fig. 8). This revised stratigraphy suggests that, in contrast to the conclusions of Baruch and Bottema (1991), the climatic evolution of the northern (Ghab) and southern (Hula) Levant was actually very similar during the Late Pleistocene to early Holocene. Attempts have been made to refine the radiocarbon age models and chronology of vegetational change of the Hula Basin by applying a correction based upon the δ^{13}C of the sample (e.g. Cappers et al., 1998, 2002). However, Meadows (2005) presented a review of the Cappers et al. (1998, 2002) method from which he concluded that the revised radiocarbon chronologies for the Hula Basin are still contradictory with other marine, terrestrial and archaeological records of palaeoclimate and palaeoenvironment. In common with Meadows (2005), we take the view that the Hula and Ghab pollen sequences record regional changes in pollen assemblages during the Lateglacial transition that can be correlated with the marine record (as in Rossignol-Strick, 1995), suggesting that during the LGM and Younger Dryas, conditions were dry and cold. In contrast the

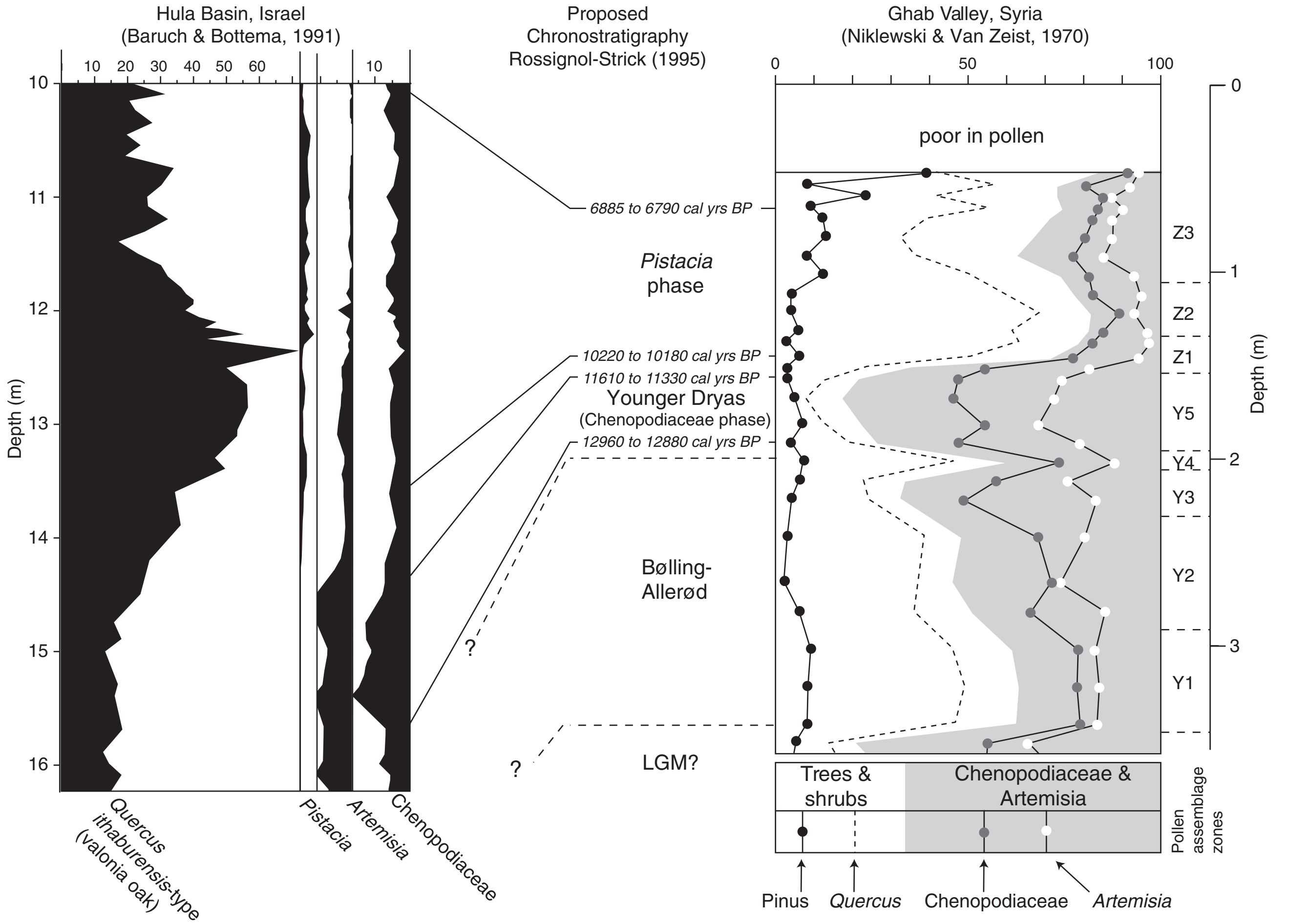

Fig. 8. Palynology of the Hula Basin (Baruch and Bottema, 1991) and the Ghab Valley (Niklewski and Van Zeist, 1970) with the proposed chronostratigraphy of Rossignol-Strick (1995).

Bølling-Allerød and early Holocene were warmer and wetter intervals. In particular, the early Holocene may have been very wet with mild winters, as evidenced by the high abundances of oak and *Pistacia* (Rossignol-Strick, 1999). These inferences are in keeping with the wider regional marine records of palynological change (Rossignol-Strick, 1995).

The Late Pleistocene–Holocene pollen record from the volcanic crater lake of Birket Ram (Golan Heights, Israel) shows clear changes in the abundance of arboreal/non-arboreal pollen (Weinstein, 1976). Unfortunately, there is no independent age model for this record; Weinstein (1976) simply correlated the pollen record with the records from the Hula Basin that were available at the time (Horowitz, 1971). A more recent study by Schwab et al. (2004) described new cores from the lake, but unfortunately these only covered the last 6500 years.

Less continuous palynological records are provided by coastal marsh and fluvial sediments. Kadosh et al. (2004) provided some pollen evidence from the Israeli coastal plain for a drier climate than present during the Younger Dryas and a wetter climate during the early Holocene. Hunt et al. (2004) provided palynological and palaeobotantical evidence from fluvial sediments in Wadi Faynan (southern Jordan) for a diverse forest assemblage in the early Holocene. On the basis of a pollen assemblage rich in tree pollen, Poaceae, *Artemisia, Plantago,* and steppe herbs, Hunt et al. (2004) suggested that before 8.0 cal ka BP Wadi Faynan was on the margin of a Mediterranean forest zone with a mean annual precipitation of about 200 mm. From ~8 to 7.4 cal ka BP, the proportion of steppeland vegetation increased, likely as a response to decreasing rainfall. Hunt et al. (2004) suggested that rainfall during this period was ~150 mm/yr. After this period drought indicators decrease in abundance (suggesting a slight increase in rainfall) but trees fail to recover, which Hunt et al. (2004) attributed to increased grazing pressure and anthropogenic activities rather than climate change.

4.3. Terrestrial geochemical records

4.3.1. Speleothems

Speleothems can provide high-resolution geochemical data ($\delta^{18}O_{spel}$, $\delta^{13}C_{spel}$, and $^{87}Sr/^{86}Sr_{spel}$) that can be dated by uranium-series dating. The oxygen-isotopic composition of speleothem carbonate is a consequence of the $\delta^{18}O$ of local precipitation, itself controlled by several factors including latitude, altitude and air temperature (McDermott, 2004). Evaporative processes in semi-arid/arid regions further modify ground waters in the vadose zone, before reaching the point of speleothem formation. The interpretation of $\delta^{18}O_{spel}$ records also depends on the assumption that the carbonate has been precipitated in isotopic equilibrium with cave drip-water throughout the interval of study (e.g. Hendy, 1971; Schwarz, 1986; McDermott, 2004). The carbon-isotopic composition of speleothem carbonate is controlled by $\delta^{13}C$ of soil CO_2, temperature, and, in closed systems, dissolution of bedrock carbonate. The $\delta^{13}C$ of soil CO_2 in modern soils is primarily controlled by the varying proportions of C_3 ($\delta^{13}C \approx -23$ to -34‰) and C_4 ($\delta^{13}C \approx -8$ to -16‰) plants and, thus, large shifts in $\delta^{13}C_{spel}$ have been attributed to major changes in vegetation. However, in some cases evaporative processes, degassing of CO_2 from ground waters and precipitation of vadose zone carbonate may lead to anomalous $\delta^{13}C_{spel}$ values. As with carbon and oxygen, a number of sources contribute to the strontium-isotopic composition of speleothem carbonate ($^{87}Sr/^{86}Sr_{spel}$). Principal among these are sea-water (directly and through sea-spray), deposition of aeolian dust in the soil cover and dissolution of bedrock (e.g. Bar-Matthews et al., 1999).

Speleothem records covering the entire time interval of interest are rare in the Levant region. However, speleothems from the Soreq Cave (Bar-Matthews et al., 1996, 1997, 1999, 2000, 2003; Ayalon et al., 1999) and the Jerusalem West Cave (Frumkin et al., 1999a, 2000; Frumkin and Stein, 2004) provide continuous, high-resolution records of the palaeoclimate of central Israel during the last ~25 cal ka BP. Other speleothem records in the region provide shorter records of various time slices since the LGM. The peak of the LGM (25–19 cal ka BP) is missing in the Ma'ale Eryaim Cave, possibly due to very low rainfall at this time east of the central mountain ridge in Israel (Vaks et al., 2003). However, the interval from 19 to 16 cal ka BP is present suggesting more precipitation at that time (Vaks et al., 2003). In the very long (250 kyr) low-resolution record from Peqiin Cave (northern Israel) hiatuses exist from 15 to 7 cal ka BP and from 5 to 0 ka cal ka BP (Bar-Matthews et al., 2000). Mid–late Holocene (~0 to 6 cal ka BP) records exist from Nahal Qanah Cave, Israel (Frumkin et al., 1999b) and caves in the Galilee region (Issar et al., 1992), although these will not be discussed here.

The speleothem palaeoclimate record from Soreq Cave consists of a composite of several individual speleothems of varying ages, which overlap in time allowing correlation between them (Bar-Matthews et al., 1999). The chronology of the composite record is provided by $^{230}Th–^{234}U$ (TIMS) ages. Against this time frame, Bar-Matthews et al. (1996, 1997, 1999, 2000, 2003) have generated high-resolution records of $\delta^{13}C_{spel}$, $\delta^{18}O_{spel}$ and $^{87}Sr/^{86}Sr_{spel}$ (Fig. 9). The oxygen-isotope data clearly show peaks and troughs that are likely recording the major climatic events of the last 25,000 years including the LGM, Bølling-Allerød and the Younger Dryas. However, both the timing and duration of these last two events are not quite as would be predicted from ice core data but this could be due to the lack of U–Th ages during this interval (Fig. 9). For the last 250 ka, Bar-Matthews et al. (2003) have demonstrated a close match, and almost constant offset, between speleothem $\delta^{18}O$ and a marine record based upon planktonic foraminiferal $\delta^{18}O$ values in the Eastern Mediterranean Sea. This suggests that climate events in the Eastern Mediterranean Sea and on-land were linked for at least the last 250 kyr. Given this correlation, it is likely that changes in SSTs were also registered in continental air temperatures, although it is not possible to use $\delta^{18}O_{spel}$ to calculate temperature directly (Bar-Matthews et al., 2003). Bar-Matthews et al. (1999, 2000, 2003) highlight the occurrence of very low speleothem and foraminiferal $\delta^{18}O$ events during interglacials, including one in the Holocene at ~8 ka that they suggest is the result of increased rainfall. Bar-Matthews et al. (2003) used alkenone SST reconstructions (see below) as a proxy for land temperatures and were able to calculate from their speleothem record the $\delta^{18}O$ of rain water ($\delta^{18}O_{rain}$) and the amount (in mm) of "palaeo annual rainfall". For the last 15,000 years, the $\delta^{18}O_{rain}$ record is most positive during the Younger Dryas, at about 5.1 ka and after 2.5 ka, indicating that these periods were intervals with low rainfall and increased aridity. Conversely, the $\delta^{18}O_{rain}$ record is more negative prior to the Younger Dryas, during the early Holocene (11–7.5 ka) and at about 4.7 ka, indicating wetter, warmer conditions. According to the calculations of Bar-Matthews et al. (2003), annual rainfall was about 600 mm during the early Holocene (~7.5–8 ka), falling to a low of 220 mm at about 5.1 ka. There was a small peak in rainfall at about 4.7 ka which was followed by increasing aridity. They did not calculate the amount of rainfall for the Younger Dryas.

In order to gain a more quantitative estimate of terrestrial temperatures, McGarry et al. (2004) measured the hydrogen-isotopic composition (δD) of fluid inclusions within speleothems from the Soreq, Peqiin and Ma'ale Efrayim Caves. McGarry et al. (2004) used a modification of a method proposed by Matthews et al. (2000) which considers the relationship between δD and $\delta^{18}O$ of fluid inclusions. The $\delta^{18}O$ of the fluid inclusions are calculated from the $\delta^{18}O$ of the enclosing speleothem carbonate with a consideration of the range of possible temperatures at which the speleothem could have grown at. This approach yielded average estimates of land temperatures that display similar trends to SSTs derived from marine alkenones in the Eastern Mediterranean. The calculated temperatures from McGarry et al. (2004) for the LGM are in the range of ~7–14 °C (Fig. 9). Holocene values are considerably warmer, varying from 14 to 17 °C between 8 and 10 cal ka BP to a maximum estimate of 20–22 °C between 0.8 and 1.2 cal ka BP.

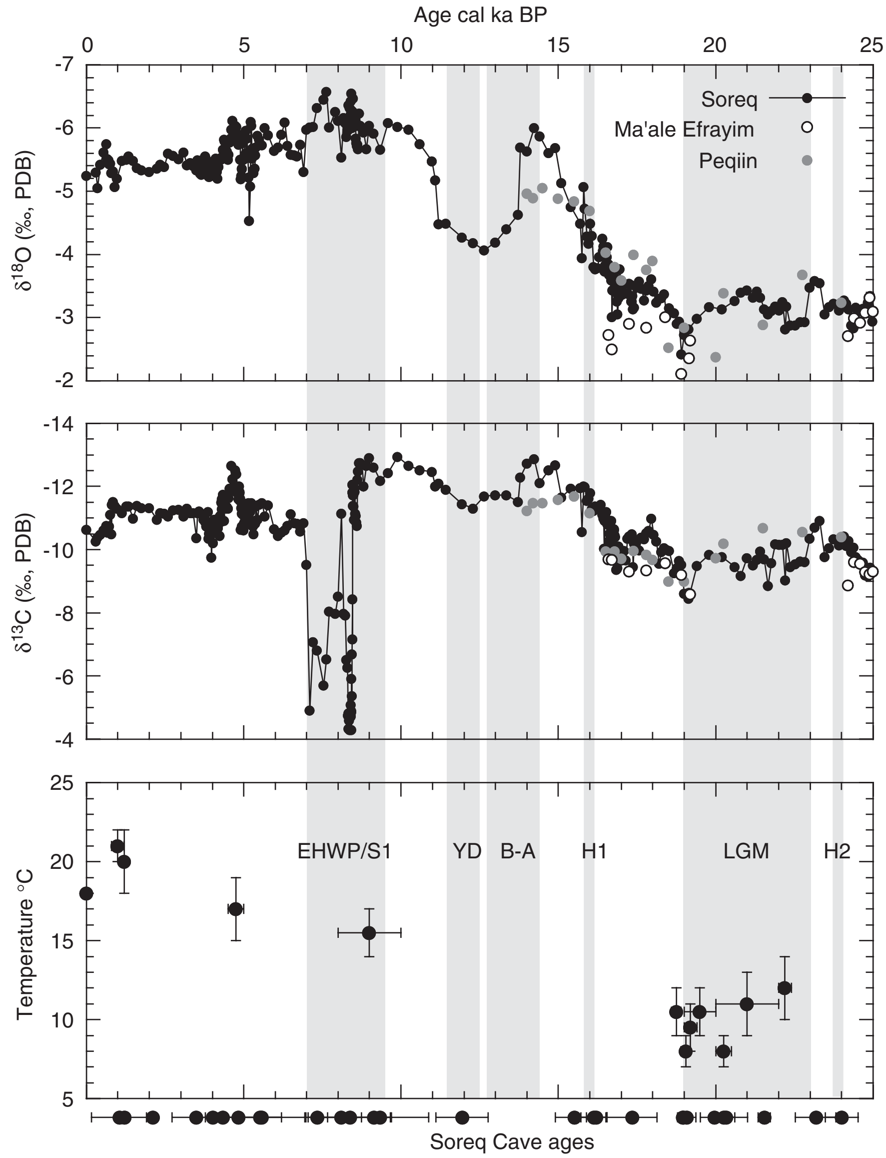

Fig. 9. Speleothem stable-isotope data (Bar-Matthews et al., 2003; Vaks et al., 2003) and reconstructed air temperatures (McGarry et al., 2004).

The different $\delta^{13}C_{spel}$ records all display similar trends from the LGM to the start of the Holocene (e.g. Bar-Matthews et al., 1999, 2000, 2003; Frumkin et al., 2000). A general shift to more negative values is recorded from ~20 to 15 cal ka BP and a plateau from this point to ~10 cal ka BP. These data are consistent with a shift from a C_4-rich flora to a mixed C_3–C_4 vegetation cover. Bar-Matthews et al. (1999, 2000, 2003) reported a large positive carbon-isotope excursion from ~8.5 to 7 cal ka in their record from Soreq Cave in central Israel. They concluded that the positive excursion is related to an early Holocene rainfall "deluge", associated with sapropel S1, which stripped soil cover thereby reducing the influence of soil CO_2 and soil processes on the $\delta^{13}C$ of groundwaters. The Bar-Matthews et al. explanation is supported by other proxy data such as the abundance of detrital material incorporated into the speleothems and Sr-isotope evidence that suggests increased weathering during the positive carbon-isotope excursion. In contrast, Frumkin et al. (2000) found no evidence for this excursion in their record from Jerusalem West Cave. It is possible that the large $\delta^{13}C$ excursion observed in the Soreq Cave is not recognised elsewhere because of either the small number of speleothems available, or because of a unique, highly localised response to increased rainfall in the Soreq Cave region.

Strontium isotope data from the speleothems in Soreq Cave (Ayalon et al., 1999; Bar-Matthews et al., 1999) and the Jerusalem West Cave (Frumkin and Stein, 2004) display similar long-term trends associated with glacial–interglacial transitions. High $^{87}Sr/^{86}Sr$ ratios are observed during the last glacial, with a rapid fall to lower values during deglaciation. Both Bar-Matthews et al. (1999) and Frumkin and Stein (2004) conclude that during the last glacial the principal sources of strontium were sea-spray and dust particles. The wetter and warmer conditions of the Holocene led to increased weathering of bedrock and thus a large decline in $^{87}Sr/^{86}Sr$ ratios towards bedrock values. The lesser influence of aeolian dust during the Holocene is also reflected by the lower Sr-concentrations in speleothem carbonate (Ayalon et al., 1999; Bar-Matthews et al., 1999; Frumkin and Stein, 2004).

4.3.2. Lacustrine sediments

Stein et al. (1997) used Sr-isotope and elemental-ratio (Mg/Ca, Sr/Ca) evidence to demonstrate that during the Late Pleistocene Lake Lisan operated in two modes that resulted in the precipitation of aragonite and gypsum. Aragonite formed when an increased supply of freshwater to the surface of the lake drove a density stratification of the lake. Gypsum was precipitated when this stratification was absent and the lake was able to mix and (possibly) overturn. Precipitation of gypsum is prevalent between 23–22 and 16–15 cal ka BP, suggesting dry climatic conditions during the LGM (e.g. Abed and Yaghan, 2000) and Heinrich Event H1.

Magaritz et al. (1991) measured the $\delta^{13}C$ of bulk organic matter from sediment cores collected from the Dead Sea shore (those used by Yechieli et al., 1993 to reconstruct lake levels, see discussion above). These data show a distinct shift from early Holocene values of −24‰ to −25‰ towards more positive values of between −21‰ to −23‰ at about ~9.4 cal ka BP. At ~5.7 cal ka BP, the data record a return to more negative values before returning to more positive values by <~2 cal ka BP. Magaritz et al. (1991) interpreted the positive shifts as a change from a C_3 flora to a mixed C_3/C_4 flora. In present-day Israel, the presence of C_3 and C_4 plants is strongly controlled by rainfall and C_4 plants are generally restricted to areas with less than 300 mm mean annual rainfall (Goodfriend, 1988). Thus, the more positive carbon-isotope data were interpreted by Magaritz et al. (1991) as representing drier climates and periods of lower mean annual rainfall. The data from Magaritz et al. (1991) would thus seem to indicate a relatively dry and arid Holocene after ~9.4 cal ka BP, with a brief return to wetter conditions at ~5.7 cal ka BP, at odds with other indicators of palaeoclimatic change (e.g. speleothems, marine proxies). It should be noted that the age constraints on this data set are extremely limited and there has been no attempt to identify the origin of the organic matter analysed. There is considerable isotopic variability within individual terrestrial plants (e.g. +3‰ between leaves and wood; Leavitt and Long, 1982) and thus much of the signal described by Magaritz could simply be due to preservational differences. Furthermore, at low total organic carbon (TOC) values (<0.2 wt% in Magaritz et al., 1991) there is potential for unrepresentative values of "terrestrial" $\delta^{13}C$ to be obtained.

4.3.3. Molluscs

In the Levant a number of studies have utilised the stable-isotope geochemistry ($\delta^{13}C$, $\delta^{18}O$) of gastropods to detect changes in rainfall source areas (e.g. Magaritz and Heller, 1980; Goodfriend, 1991), and to investigate temporal vegetational changes related to the amount of precipitation (Goodfriend, 1988, 1990). Low-resolution data from Magaritz and Heller (1980) suggest that between ~12.9 and 12.5 cal ka BP (during the Younger Dryas), the climate of Israel was more arid than at present. Goodfriend (1991) showed that during the early Holocene (~10.95 cal ka BP) $\delta^{18}O$ values of gastropods from southern Israel were similar to modern values (Fig. 10). However, at ~6.8 cal ka BP $\delta^{18}O$ values were depleted by 2‰, which Goodfriend (1991, 1999) suggests is the result of more frequent storms reaching the region from north-east Africa.

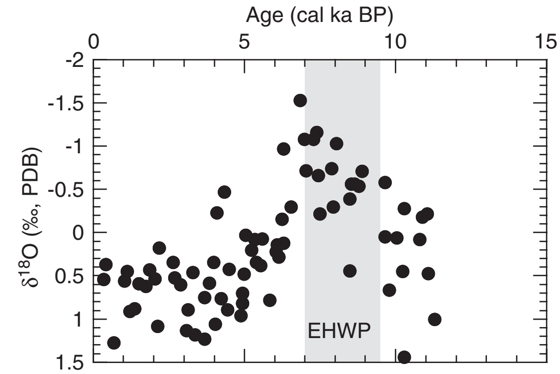

Fig. 10. Gastropod oxygen isotope data from the Negev Desert (Goodfriend, 1991).

The $\delta^{13}C$ of organic matter contained within gastropod shells has been used to track changes in the abundance of C_3/C_4 plants in the Negev Desert during the Holocene (Goodfriend, 1988, 1990, 1999). Based upon the $\delta^{13}C_{org}$ of gastropods, Goodfriend (1988, 1990, 1999) has shown that the C_3/C_4 transition zone was ~20 km further south in the middle Holocene (~7.4–3.2 cal ka BP), compared to its present position, implying that there was considerably more rainfall in the region at this time. In the early Holocene (~11–7.8 cal ka BP) a considerable area of the Negev desert appears to have been covered by C_3 plants, implying even higher rainfall. Goodfriend (1999) suggests that the annual rainfall at this time in the area north of the present-day 100 mm isohyet was more than 290 mm.

4.3.4. Calcretes

In the Negev Desert periods of pedogenesis occurred during wetter climates, resulting in the development of palaeosols containing calcrete nodules (Goodfriend and Magaritz, 1988). Between the phases of pedogenesis drier conditions existed, particularly during the LGM, that resulted in extensive erosion. Stable isotope ($\delta^{13}C_{cal}$, $\delta^{18}O_{cal}$) data from Late Pleistocene (~16.7–12.8 cal ka BP) soil carbonate nodules in the Negev Desert are enriched in ^{13}C and ^{18}O compared to those from older horizons (Magaritz, 1986). This has been interpreted as the result of a southward shift in the desert boundary, widespread pedogenesis and a possible change in the source of the precipitation in the region (Magaritz, 1986; Goodfriend and Magaritz, 1988). Additionally within the isotope data, there is a strong north–south carbon-isotope gradient, which suggests a decrease in plant productivity and an increase in C_4 plants towards the south (Goodfriend and Magaritz, 1988). These factors are consistent with a strong north–south rainfall gradient, suggesting that the major source area of precipitation was to the north or north-west, thereby excluding the possibility that monsoonal rains contributed to rainfall during the Late Pleistocene wet periods. Holocene oxygen-isotope values from the most southerly site (Ramat Hovav) are extremely positive ($\delta^{18}O = +2.5‰$), which Magaritz (1986) suggests is the result of extreme aridity and vigorous evaporative processes. Unfortunately, the lack of nodules in some horizons and changes in facies did not allow Magaritz (1986) to produce a high-resolution time series of calcrete stable-isotope data, making it very difficult to fully understand palaeoclimatic changes from the calcrete data alone.

4.4. Marine records

4.4.1. Deep-sea cores

A large number of deep sea cores have been collected from the Eastern Mediterranean and northern Red Sea, providing a good spatial record of Late Pleistocene–Holocene sedimentation in pelagic and hemipelagic settings. These records are generally continuous, yet are limited in their temporal resolution by typical pelagic sedimentation rates of 2–5 cm/kyr. In a few areas where turbidites are a significant feature of the sedimentary regime, extremely high sedimentation rates are attained (up to 2 m/kyr; Reeder et al., 2002). A number of palaeoceanographic proxies have been applied to these cores yielding information about SSTs, sea-surface salinity (SSS), and continental run-off.

Modern planktonic foraminiferal assemblages in core-top sediments vary with the temperature and salinity of the water in which they lived (see CLIMAP Project Members, 1976; Hayes et al., 2005). Since the 1970s transfer functions have been used to relate quantitatively foraminiferal assemblages with temperature and salinity, allowing the reconstruction of these parameters in the geological past. For the Eastern Mediterranean various reconstructions exist (e.g. CLIMAP Project Members, 1976; Thiede, 1978; Thunell, 1979; Kallel et al., 1997a; Hayes et al., 2005) that utilise a number of different transfer functions. The recent study of Hayes et al. (2005) calculated LGM palaeotemperatures at 37 Mediterranean sites using a core-top calibration data set from 274 sites in the Mediterranean and North Atlantic. Hayes et al. (2005) used a revised analogue technique and artificial neural networks to calculate SSTs. Their reconstructions suggest that mean annual temperatures in the Eastern Mediterranean (Levantine Basin) during the LGM were approximately 2 °C colder than present. Much of this difference seems to have been manifested as colder summer temperatures, whilst winter temperatures show much less difference compared to present values (Fig. 11). Model results (Fig. 4) are broadly in agreement with such an interpretation. Kallel et al. (1997a) calculated SSTs for the Holocene using modern analogues of fossil foraminiferal assemblages. Their results suggest that during the early Holocene (~8–9 cal ka BP) SSTs in the Eastern Mediterranean were similar to present, although in the central and western areas they were somewhat cooler.

An alternative SST proxy is based upon the Uk'_{37} index, a measure of alkenone unsaturation ratios that is dependent on SST (Prahl et al., 1988). This independent measure of SST, in conjunction with estimates of global ice volume, makes it possible to calculate SSS from the oxygen-isotopic composition of planktonic foraminifera ($\delta^{18}O_{foram}$). Using the alkenone method, Emeis et al. (1998, 2000, 2003) have derived estimates of SSTs and SSSs at three Mediterranean sites, including one in the Levantine Basin at ODP Site 967 (Fig. 12). Their reconstruction suggests that during the LGM the Eastern Mediterranean was at least 5–6 °C cooler than present (SST ≈ 12 °C). Between 18 and 9.5 cal ka BP (tie points in the age model) there appears to have been an interval of much warmer temperatures, followed by much cooler temperatures. These intervals could correspond to a warm Bølling-Allerød and a cooler Younger Dryas, although the interpolated dates of the age model are not in agreement with the accepted ages of these events. This highlights the difficulty of sampling and dating short (centennial and millennial) events at low-to-moderate sedimentation rates, which may have been variable. The earliest Holocene is characterised by rising temperatures and lowering salinities. During deposition of sapropel S1 (an organic rich unit dated between ~9.5 and 7.5 cal ka BP) SSS was lower than present (due to increased freshwater input) and SSTs were similar to present. A brief saline event during S1 is interpreted by Emeis et al. (2000) as a reduction in freshwater input during this time of otherwise high freshwater input.

Arz et al. (2003a, b) also used the alkenone method to produce a record of SSTs and SSSs in the Northern Red Sea spanning the Late Pleistocene to Holocene (Fig. 13). This record suggests that during the Late Pleistocene the northern Red Sea was cooler by ~4 °C and more saline by ~10‰ than at present. From about 18 to ~14 cal ka BP, SSTs were even colder, ranging from ~16 to 21 °C. At about 16.5 cal ka BP, coincident with Heinrich Event H1, a minor freshening and a major cooling to ~16 °C is observed in the northern Red Sea. Arz et al. (2003a) suggest that the cooling was either a result of enhanced westerlies or an enhanced NE monsoon. A rapid 4.5 °C increase in palaeotemperatures occurred at ~14.5 cal ka BP marking the onset of the Bølling-Allerød warm period. SSTs during the Younger Dryas were cooler. The transition out of the Younger Dryas was not sharp in the Red Sea, suggesting a gradual improvement in climatic

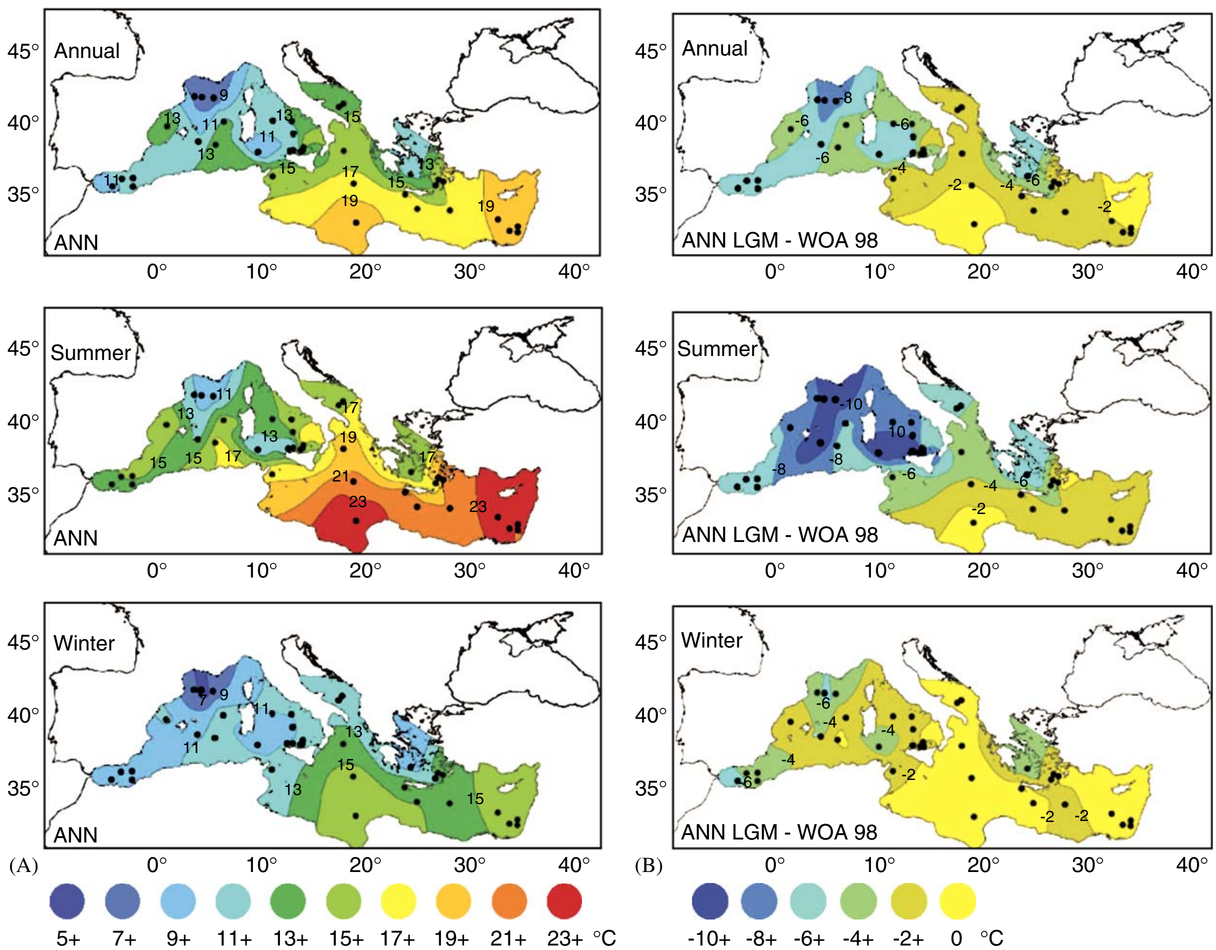

Fig. 11. (A) Foraminiferal LGM annual, summer and winter SST reconstructions (Hayes et al., 2005), calculated using an artificial neural network (ANN). (B) Temperature anomalies for annual, summer and winter SSTs during the last glacial maximum, compared to modern day values (Hayes et al., 2005). Anomaly values were calculated by subtracting modern day SSTs from the glacial values. The black dots represent the sites of the cores from which the LGM data were obtained. This figure is a reproduction of part of Fig. 9 in Hayes et al. (2005).

conditions into the early Holocene. Salinity during the Younger Dryas is slightly higher than during the Bølling-Allerød, suggesting increased aridity. After the Younger Dryas, SSTs gradually increased during the Holocene (Arz et al., 2003a, b). SSSs decreased after the Younger Dryas reaching a minimum during the early Holocene ''humid period'' (Arz et al., 2003a, b), an interval with low aridity index values (Fig. 13). After ~6.5 cal ka BP, the Red Sea became warmer, more arid and more saline, consistent with Mediterranean climatic records (Arz et al., 2003b).

Other attempts at calculating Mediterranean salinities from $\delta^{18}O_{foram}$ have been attempted using various different corrections for temperature. Thunell and Williams, (1989) assumed that the difference in oxygen-isotope values between 8 cal ka BP and the present is entirely due to salinity changes and not due to temperature or ice-volume (considered by Thunell and Williams, 1989, to be identical to present at 8 cal ka BP). Using this assumption and estimates of ice-volume at the LGM, Thunell and Williams (1989) estimated that the Eastern Mediterranean was more saline at the LGM by 2.7‰. They estimated that at 8 cal ka BP, SSSs were fresher by 2.9‰, co-incident with the early Holocene humid period. Whilst this method may yield sensible indications of changes in salinity, it does not represent an overly rigorous estimate of SSSs, as the study was based largely on averages of many sites and does not account for local hydrographic and climatic variations. Kallel et al. (1997a) used their foraminiferal assemblage temperatures and $\delta^{18}O_{foram}$ records to produce a detailed picture of salinity variation during the early Holocene, suggesting fresher surface waters during this interval and that surface salinity was almost homogenous across the entire Mediterranean during the deposition of S1. However, Rohling and De Rijk (1999) delivered a cautionary warning, suggesting that salinities estimated from corrected $\delta^{18}O_{foram}$ values may be overestimated as these studies do not consider temporal changes in the oxygen-isotopic composition of rainfall and river waters.

In addition to reconstructing the physical (T, S) marine environment, geochemical analysis of deep-sea cores can be used to reconstruct changes and sources of terrigenous input and, thus, be used to estimate changes in run-off.

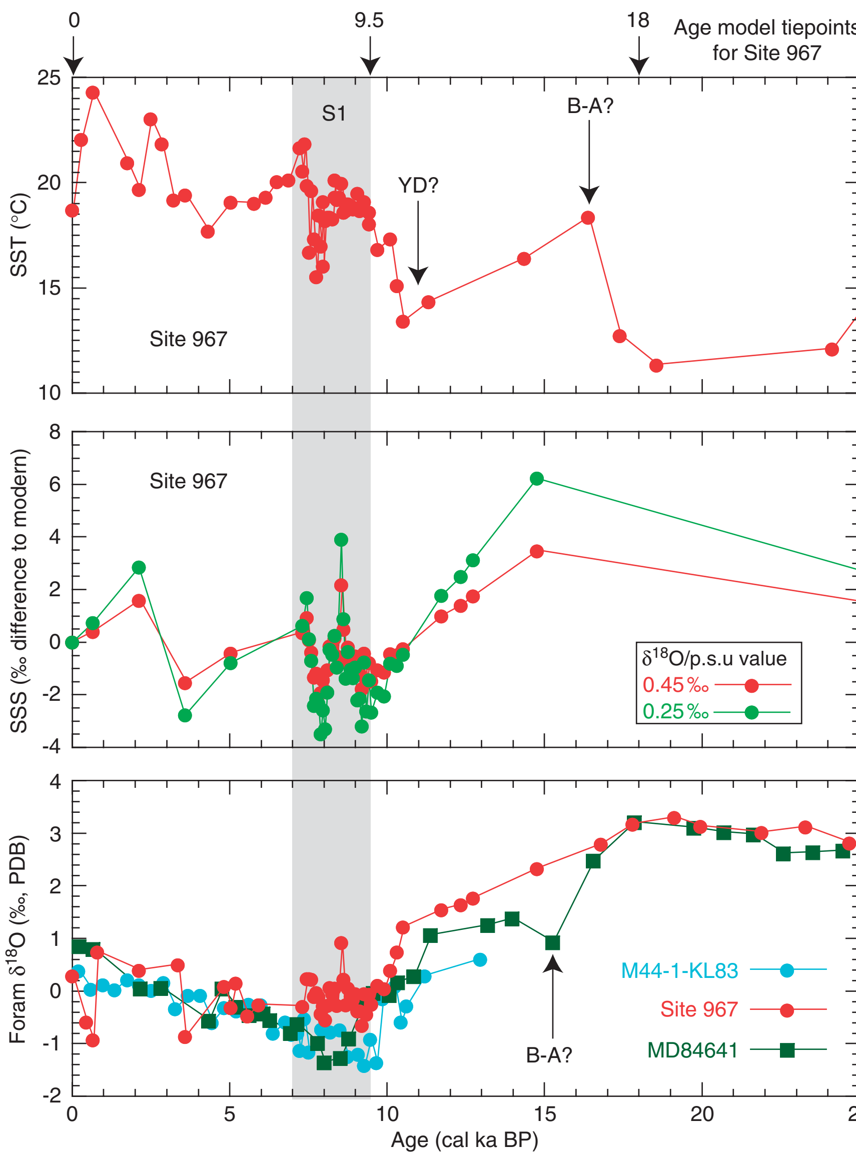

Fig. 12. Compilation of Eastern Mediterranean Sea palaeoclimatic records from Site 967 (Emeis et al., 1998, 2000, 2003) and MD84-461 (Fontugne and Calvert, 1992). The "δ^{18}O/p.s.u. value" is a coefficient used in the calculation of sea-surface salinity that relates salinity and $\delta^{18}O_{seawater}$ (see Emeis et al., 2000 for more details).

Calvert and Fontugne (2001) showed that quartz/clay, Si/Al and Zr/Al ratios were higher during the LGM than the Holocene, suggesting increased wind speeds and dust transport in glacial times. Such an interpretation is compatible with GCM output of increased windspeeds during the LGM (Fig. 4H). Many studies have focused primarily on the early Holocene sapropel S1, which is thought to have resulted from increased freshwater supply into the Mediterranean that inhibited circulation and oxygenation of deep waters (e.g. Rossignol-Strick et al., 1982; Rossignol-Strick, 1985). Krom et al. (1999) attempted to determine the provenance of sediment in the Eastern Mediterranean during S1 by measuring major elements and strontium isotopes ($^{87}Sr/^{86}Sr$) in deep sea cores, a core in the Nile delta and on Saharan dust. They showed that there was a major change in sediment source concomitant with changes in surface salinity as determined by oxygen isotopes. Krom et al. (1999) suggested that a decrease in Saharan dust input coupled with an increase in Ionian and Levantine riverine input best explained their geochemical data in the western Levantine and Ionian basins. This conclusion suggests that the increased outflow from the River Nile did not influence sedimentation over the entire Eastern Mediterranean (as proposed by Rossignol-Strick et al., 1982; Rossignol-Strick, 1985) and that local increases in riverine input (particularly in the west) were also important for the development of S1. Freydier et al. (2001) and Scrivner et al. (2004) used Nd isotopic variations to demonstrate the importance of the Nile outflow for sites east of Crete (approximately 025°E) during S1. Scrivner et al. (2004) also suggest that the peak of Nile outflow may have been in the central part of the

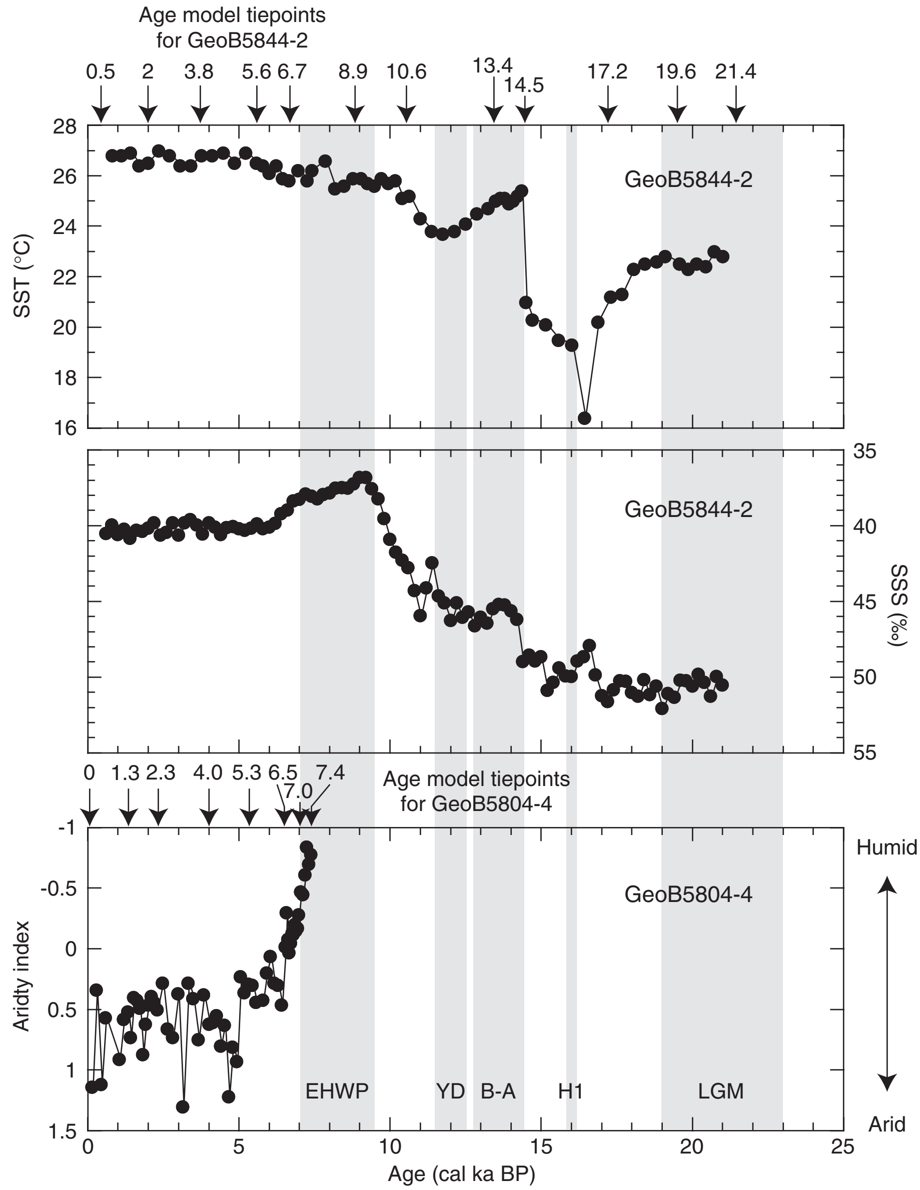

Fig. 13. Compilation of northern Red Sea palaeoclimatic records (Arz et al., 2003a, b).

sapropel event, and that other freshwater discharges were more important than the Nile in the early and late stages of the sapropel.

An alternative method of investigating past changes in runoff is provided by deep water turbidites. Reeder et al. (2002) have shown that Late Pleistocene–Holocene turbidites in the Eastern Mediterranean (Herodotus Basin) were partially controlled by changes in sea-level and climate. These turbidites were dated by radiocarbon and characterised as to their source (Nile Cone, Libyan/Eygptian shelf or Anatolian rise). During the LGM (up to ~17 cal ka BP) no turbidites were derived from the Nile, attributed by Reeder et al. (2002) to the cooler and drier climate in the Nile hinterland at this time. In contrast after the LGM, turbidites deposited in the Herodotus Basin were exclusively sourced from the Nile and contain an abundance of terrestrial organic matter. Reeder et al. (2002) suggest that these units correspond to increased Nile outflow associated with the early Holocene humid phase and sapropel S1. After ~6 cal ka BP, only thin turbidites originating from the North African shelf are present, which is consistent with a decreased Nile outflow and an increase in aridity over the Nile region in the late Holocene.

4.4.2. *Coral records*

Presently, the only coral records for the Levantine region come from the Red Sea. The coral record here is highly discontinuous, with reefs forming predominantly during interglacial times (e.g. Reyss et al., 1993; Gvirtzman et al., 1992). Accordingly, during the last 25,000 years the only corals that exist are Holocene in age (e.g. Gvirtzman et al., 1992; Gvirtzman, 1994; Moustafa et al., 2000; Felis et al., 2004), and even these only date from ~6.5 cal ka BP onwards. The stable-isotope and elemental geochemistry of corals have been used extensively to reconstruct palaeoceanographic conditions at high resolutions, yet the Holocene Red Sea corals have received little attention. Moustafa et al. (2000) presented stable carbon and oxygen isotope data from *Porites* spp. corals from the northern Gulf of Aqaba that range in age from ~6.5 to 5.1 cal ka BP. Some of their samples displayed seasonal variations in oxygen-isotope values. The amplitude of these changes was greater than present and annual growth rates were lower, suggesting to Moustafa et al. (2000) that the mid-Holocene was characterised by a larger seasonal SST contrast, reduced SSS variations and increased terrigenous input. However, Arz et al. (2003b) suggest that in the Red Sea the major humid interval of the Holocene had terminated by ~7.25 cal ka BP. Unfortunately, given the brevity of the Moustafa et al. (2000) coral record it is difficult to know whether the coral data record a transitional phase (between extremes of aridity/humidity) or a minor humid event unrecognised in the Arz et al. (2003b) aridity records.

5. Discussion

The key marine and terrestrial palaeoclimatic records from the Eastern Mediterranean and Levant are summarised in Fig. 14. This figure provides a broad view of the climatic evolution of this region over the last 25,000 years at different localities and from different proxies. The following discussion focuses on the major climatic events during this interval and attempts to summarise the spatial picture of climatic conditions during these events. By viewing "time slices" of climate (Fig. 15) it is hoped that the reader can gain a more informed appreciation for where the data come from and, importantly, where gaps exist in our knowledge.

5.1. Heinrich event 2, H2 (~23.8 cal ka BP)

Heinrich events were times of intense cooling resulting from the addition of freshwater to the North Atlantic inhibiting heat transport from the tropics to northern latitudes (e.g. Bond et al., 1992, 1993). In the Levant and Eastern Mediterranean, H2 appears to have been marked by a sharp lowering in lake levels (Bartov et al., 2002, 2003; Landmann et al., 2002; Hazan et al., 2005) and a small positive excursion in speleothem oxygen isotope values (in the Soreq Cave), interpreted as a cooling event (dated at ~25 cal ka BP by Bar-Matthews et al., 1999). Bartov et al. (2003) suggests that cold-water input into the Mediterranean during H2 caused a reduction of evaporation and precipitation in the Eastern Mediterranean region, leading to transient cooling and an evaporation excess over the Levant region. Unfortunately the abundance of data for H2 in the Eastern Mediterranean and Levantine Region is low.

5.2. Last Glacial Maximum (~23–19 cal ka BP)

The palaeoclimatic records of the LGM in the Eastern Mediterranean and Levant suggest that the region was generally cooler and more arid than present, generally in good agreement with the predictions made by the climate models (Figs. 3 and 4). Mean Annual SSTs predicted by foraminiferal assemblages (Hayes et al., 2005) vary from 15 °C north of Crete to about 19 °C in the far Eastern Mediterranean (Fig. 15A). However, alkenone temperature estimates from ODP Site 967 suggest lower SSTs of ~12 °C (Emeis et al., 2000, 2003), some 1° to 3° less than even the winter temperatures predicted by Hayes et al. (2005). This discrepancy could be due to the absence of foraminiferal assemblage temperature estimates in the region around Site 967 in Hayes et al. (2005). The Hayes et al. (2005) reconstruction is in good agreement with other alkenone temperature estimates further west (e.g. Cacho et al., 2000), suggesting that the techniques should give comparable results in the Eastern Mediterranean. Salinity estimates from sites in the central and western Mediterranean are elevated at the LGM compared to present values, suggesting increased evaporation and/or decreased freshwater input (e.g. Kallel et al., 1997b; Emeis et al., 2000). Unfortunately, reliable SSS estimates for the Eastern Mediterranean are currently unavailable, although Thunell and Williams (1989) suggest an average salinity 2.7‰ greater than present. The northern Red Sea also appears to have been marked by cooler temperatures and increased salinity at the LGM (Arz et al., 2003a), although at this time the reduced connection between the global ocean and the Red Sea would also have effected palaeoceanographic conditions (Siddall et al., 2003, 2004). The speleothem record from Soreq Cave displays a gradual increase in $\delta^{18}O$ values from ~21 to 19 cal ka BP, which Bar-Matthews et al. (1997, 1999) interpret as a decrease in temperature. McGarry et al. (2005) calculate an air temperature range of ~8–12 °C above the Soreq Cave during the LGM. Bar-Matthews et al. (1997) estimated that mean annual rainfall above the Soreq Cave during the period spanning the LGM was between 250 and 400 mm, considerably less than the 500 mm that presently accumulates there. The absence of speleothem precipitation during the interval from 25 to 19 cal ka BP in the Ma'ale Efrayim Cave (in the modern rain shadow of the Judean Mountains) is probably due to the unavailability of soil zone water at this time (Vaks et al., 2003) further suggesting low rainfall at this time in the Levant region. Dry, cool conditions during the LGM are also evidenced by erosion in the Negev Desert (Goodfriend and Magaritz, 1988), the precipitation of gypsum in the Dead Sea (Abed and Yaghan, 2000) and elevated non-arboreal pollen values in Syria (according to the revised Ghab stratigraphy of Rossignol-Strick, 1995). Speleothem carbon-isotope values also become more positive at this time, possibly suggesting an increase in the proportion of C_4 plants consistent with cooler and drier conditions (Bar-Matthews et al., 1997, 1999).

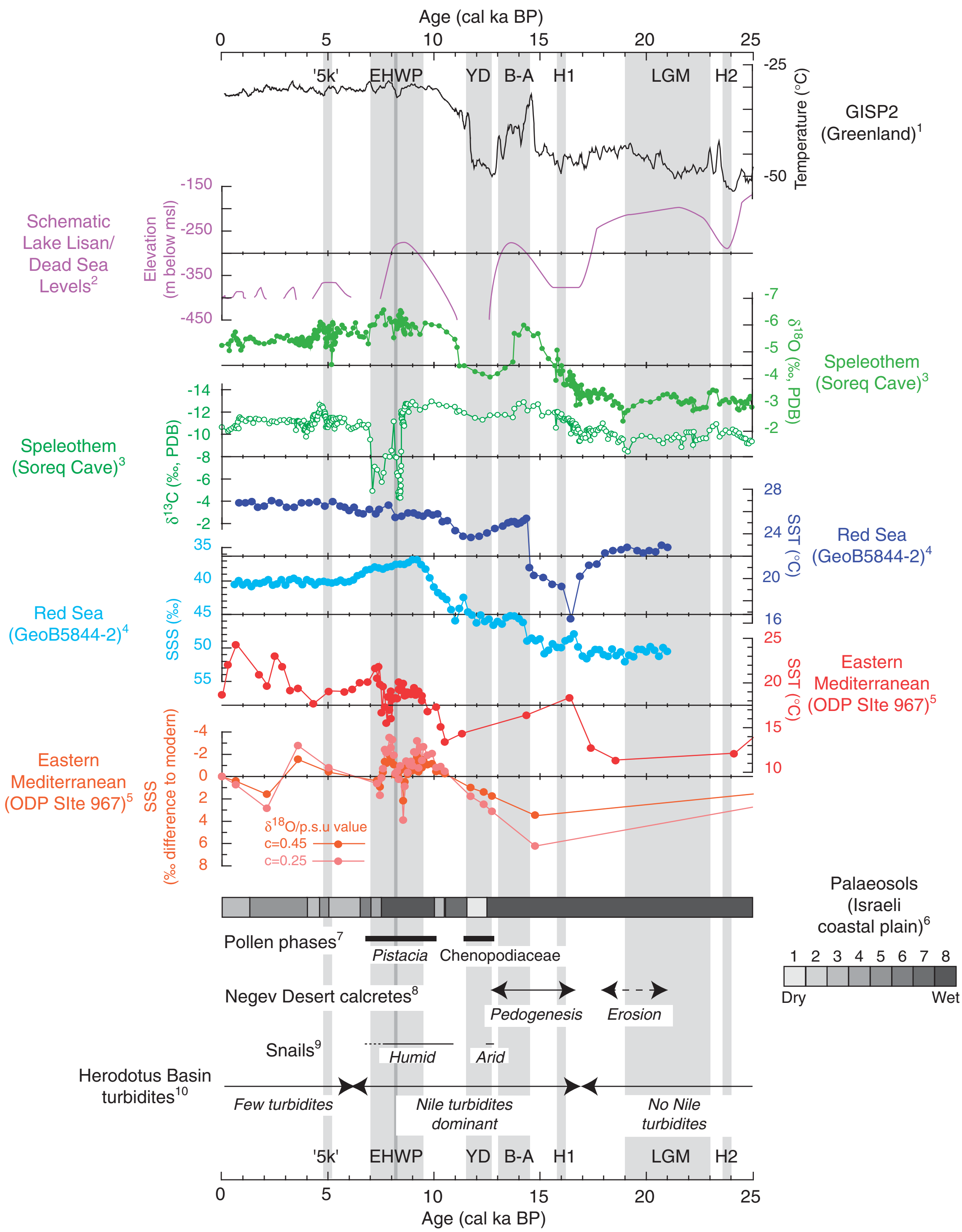

Fig. 14. Compilation of terrestrial and marine palaeoclimatic proxy data for the Levant and Eastern Mediterranean. Also shown is the ice core record from GISP 2 (Greenland). References: 1: Alley (2000); 2: see Fig. 6; 3: Bar-Matthews et al. (2003); 4: Arz et al. (2003a); 5: Emeis et al. (2000, 2003); 6: Gvirtzman and Wieder (2001); 7: Rossignol-Strick (1995); 8: Magaritz (1986), Goodfriend and Magaritz (1988); 9: Magaritz and Heller (1980); Goodfriend (1991, 1990, 1999); 10: Reeder et al. (2002).

The strontium isotopic composition of speleothem carbonate indicates that dust transport and sea spray were elevated compared to bedrock dissolution during the LGM (Bar-Matthews et al., 1999; Frumkin and Stein, 2004), implying lower amounts of precipitation. Elemental ratios in Eastern Mediterranean sediments also suggest increased wind speeds and dust transport during the LGM (Calvert and Fontugne, 2001). Lake Lisan levels remained high during the LGM although they may have been falling. Clearly, despite the dry climate, evaporation in the Dead

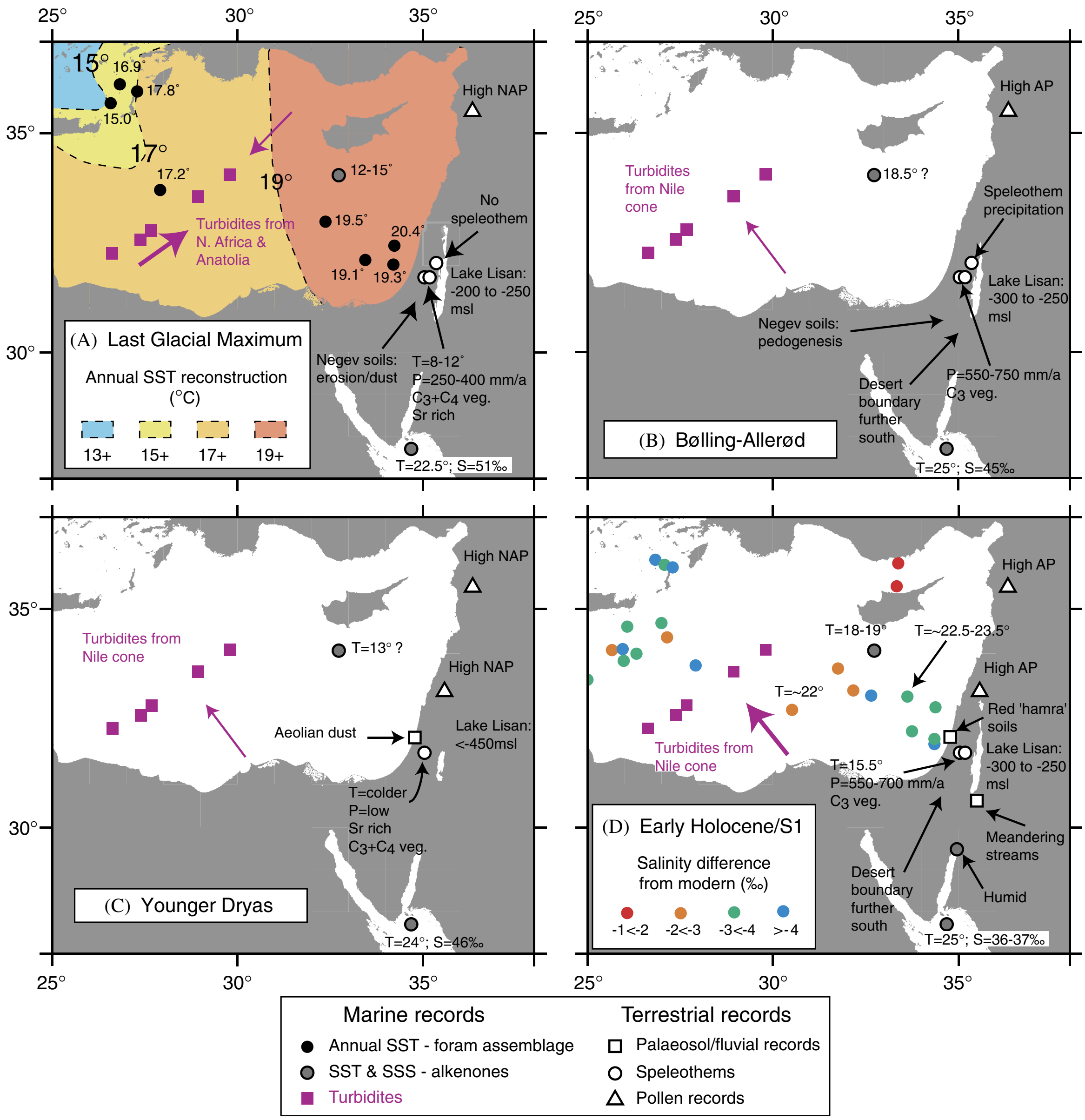

Fig. 15. Summary of climatic conditions at the LGM (A), peak of the Bølling-Allerød warm phase (B), the Younger Dryas (C) and during the early Holocene/S1 (D). Turbidite data from Reeder et al. (2002); Alkenone SSTs and SSS data from Emeis et al. (2000, 2003) and Arz et al. (2003a, b); speleothem data from Bar-Matthews et al. (1997, 1999, 2000, 2003), Frumkin et al. (1999, 2000), Vaks et al. (2003) and McGarry et al. (2004); pollen data from Niklewski and Van Zeist (1970), Baruch and Bottema (1991) and Rossignol-Strick (1995); lake levels from Fig. 6 (this study); LGM annual SSTs calculated from foraminiferal assemblages taken from Hayes et al. (2005); Early Holocene sea-surface salinity values are from Kallel et al. (1997a); Israeli coastal plain palaeosol data from Gvirtzman and Wieder (2001); Negev data from Magaritz (1986), Goodfriend and Magaritz (1988); Goodfriend (1999); Southern Jordan fluvial data from McLaren et al. (2004). T = temperature (all in °C), S = salinity, P = precipitation, NAP = non-arboreal pollen, AP = arboreal pollen, maps drawn with GMT (http://gmt.soest.hawaii.edu/). Coastlines do not account for any changes in sea-level or sedimentation.

Sea Rift was sufficiently compensated by water supply to maintain lake levels or, at least, prevent a extremely large regression. In the absence of rainfall, riverine input from melting snow (as suggested by the GCM) may have been significant.

5.3. Heinrich event 1 H1 (16.0 cal ka BP)

As with H2, H1 in the Levant is marked by a lowering of lake levels (Bartov et al., 2002, 2003) and a small positive oxygen-isotope excursion in the Soreq Cave speleothem record (Bar-Matthews et al., 1999). Unfortunately, the Eastern Mediterranean marine records of SST and SSS are at an insufficient resolution to detect whether H1 had a significant effect. Red Sea SSTs are generally low from ~17 to 15 cal ka BP compared to the LGM. A sharp drop in SST and small decrease in salinity appear to have occurred at ~16 cal ka BP which Arz et al. (2003a) postulate is a response to atmospheric cooling during H1, although strangely there is no oxygen-isotope response during this time period. This cooling is in keeping with Bartov et al.'s (2003) hypothesis that cold-water input into the Mediterranean during H2 caused a reduction of evaporation and precipitation in the Eastern Mediterranean region, leading to transient atmospheric cooling and an evaporation excess over the Levant region. However, how the Red Sea became fresher during an interval of increased evaporation is unclear, although long-term sea-level rise and changes in Red Sea circulation may explain some of the salinity decrease.

5.4. Bølling-Allerød warm interval (~15–13 ka)

Lake levels during the Bølling-Allerød (Fig. 15B) appear to have been reasonably high, suggesting an increase in the precipitation/evaporation ratio from the conditions during H1. The revised pollen diagram from the Ghab Valley seems to suggest that oak forests were prevalent during the Bølling-Allerød, consistent with warmer and wetter conditions (see Rossignol-Strick, 1995). According to Rossignol-Strick (1995) deciduous oak requires at least 500 mm/yr rainfall, consistent with estimates from speleothems of 550–750 mm/yr (Bar-Matthews et al., 1997). The speleothem stable isotope data are also suggestive of an increase in C_3 plants and increased temperatures during this interval (Bar-Matthews et al., 1997, 1999, 2003). The Eastern Mediterranean marine record of the Bølling-Allerød is unclear, probably as a result of age model problems and the low temporal resolution of the samples (a result of low sedimentation rates). However, warm temperatures (~18 °C) occur at Site 967 at ~16.5 cal ka BP (Emeis et al., 2000, 2003), which could feasibly be part of a mis-dated Bølling-Allerød event, although this is impossible to determine at this time. A single data point at ~15 cal ka BP in the MD84641 oxygen-isotope record (data from Fontugne and Clavert, 1992) may record a peak in warmth and/or a low in salinity during the glacial–interglacial transition. In the Red Sea, where sample temporal resolution is higher, SSTs rose dramatically at ~14.5 cal ka BP concomitant with a small decrease in SSSs.

5.5. Younger Dryas (~12.7–11.5 cal ka BP)

Although there has been some debate about climatic conditions during the Younger Dryas in the Levant, based mainly upon the poorly dated palynological records (e.g. Baruch and Bottema, 1991, 1999), it is clear from our review of multiple datasets that the period was extremely arid and, most likely, cold compared to the Bølling-Allerød and the Holocene (Fig. 15C). Extreme aridity is evidenced by the sedimentary record in the form of massive salt deposition and a lowering of lake levels in Lake Lisan (e.g. Yechieli et al., 1993) and deposition of wind blown sediments on the Israel coastal plain (Gvirtzman and Wieder, 2001). The high abundances of Chenopodiaceae and *Artemisia* suggest rainfall of <150 mm/yr. Rossignol-Strick (1995) suggests that in the Levant the Younger Dryas was the most arid period of the Late Pleistocene. Speleothems record a return towards glacial oxygen and carbon-isotope values suggesting cooling, aridity and an increase in C_4 vegetation. Bar-Matthews et al. (2003) calculated that $\delta^{18}O_{rain}$ was more positive during the Younger Dryas compared to any interval of the last 15 cal ka BP consistent with lower rainfall during this period.

The marine record of the Younger Dryas is less conclusive. Cool SSTs (~13 °C) are recorded at Site 967 at about 11 cal ka BP (Emeis et al., 2000, 2003), which could be part of the Younger Dryas. Surprisingly, there does not appear to be any change in salinity at this time or a major shift in foraminiferal $\delta^{18}O$ values. However, further west several records show a return towards positive glacial oxygen isotope values, lower alkenone SSTs, and higher salinities during the Younger Dryas (e.g. Kallel et al., 1997b; Cacho et al., 2001). In the Red Sea, there was a cooling spanning the Younger Dryas, although salinity does not appear to have changed greatly during this interval. Given the dramatic climatic changes recorded by terrestrial proxies it seems surprising that the Eastern Mediterranean marine record does not show a clear climatic response during the Younger Dryas. Higher resolution records from sites with high sedimentation rates are required to determine whether the impact of the Younger Dryas on marine environments in the Eastern Mediterranean Sea was significant or not.

5.6. The early Holocene and Sapropel S1 (~9.5–7 cal ka BP)

The early Holocene appears to have been the wettest phase of the last 25,000 years across much of the Levant and Eastern Mediterranean (Fig. 15D). Elevated rainfall is evidenced on land by increases in *Pistacia* and oak in the Ghab and Hula pollen records (Rossignol-Strick, 1995, 1999), southward migration of the Negev Desert boundary (Goodfriend, 1999), meandering streams in southern Jordan (McLaren et al., 2004) and the deposition of "red hamra" type palaeosols (indicating an excess of precipitation versus evaporation) on the Israeli coastal plain (Gvirtzman and Wieder, 2001). Lake levels also returned to relatively high levels following the massive drawdown during the Younger Dryas (e.g. Frumkin et al., 1994). The Soreq Cave speleothem oxygen-isotope record displays a shift towards more negative values, consistent with warmer, wetter conditions; Bar-Matthews et al. (2003) estimates rainfall amounts of between 550 and 700 mm/yr. The carbon-isotope record from Soreq is remarkable, displaying a large (~8‰) excursion towards more positive values. This is interpreted by Bar-Matthews as resulting from greatly enhanced rainfall removing soil cover and increasing bed-rock dissolution. The absence of this excursion in other speleothem records (e.g. Frumkin et al., 2000) suggests that soil removal may have been a highly localised phenomenon. The speleothem $\delta^{13}C$ record from the Jerusalem West Cave is relatively negative during the early Holocene, compared to the late Pleistocene or the late Holocene, suggesting relatively more C_3 vegetation at this time, consistent with elevated rainfall (Frumkin et al., 2000). McGarry et al. (2004) calculated a terrestrial palaeotemperature of ~16 °C during the early Holocene.

Sediments rich in organic carbon (Sapropel 1 or S1) were deposited in the Eastern Mediterranean under anoxic conditions between ~9.5 and 7 cal ka BP (e.g. Emeis et al., 2003) At Site 967, SSTs are estimated to have been ~20–22 °C whilst salinity was some 2–4‰ less than modern values at the beginning and end of sapropel deposition (Emeis et al., 2000, 2003). Foraminiferal oxygen-isotope values also suggest a freshening of surface waters during S1. Mediterranean sapropels have been linked to minima in the precession cycle and it has been postulated that anoxic conditions were induced through stratification of the water column driven by a low salinity surface layer (e.g. Rossignol-Strick et al., 1982, Rossignol-Strick, 1985; Rohling and Hilgen, 1991; Rohling, 1994; Kallel et al., 1997a). The source of this low salinity water may have been from an enhanced Nile River outflow, driven by increased rainfall in the Sahara (e.g. Rossignol-Strick et al., 1982; Rossignol-Strick, 1985). Evidence for this in the Eastern Mediterranean comes from increased Nile-cone-sourced turbidites in the Herodotus Basin (Reeder et al., 2002) and more radiogenic Nd-isotopic values in planktonic foraminifera at ODP Site 967 (Scrivner et al., 2004). However, further west trace element data and Nd-isotopic compositions suggest that influence of the Nile water there was minimal (Krom et al., 1999; Freydier et al., 2001). This is supported by the work of Kallel et al. (1997a) who argued that the uniform oxygen-isotopic values of surface waters in the eastern and central Mediterranean resulted from a uniformly distributed freshwater source across the entire region that most likely originated from enhanced precipitation over the entire basin. Increased rainfall in the Levant led to an increase in run-off into the Red Sea, as recorded by elevated amounts of terrigenous sediment and a minimum in salinity estimates.

Some records of S1 show a brief interruption of anoxic conditions, lasting approximately 200 years (e.g. Rohling et al., 1997; Myers and Rohling, 2000). Myers and Rohling (2000) link this to a brief cooling event (evident in the Adriatic Sea) that induced some circulation of deep and intermediate waters in the Mediterranean. This event is dated by Rohling et al. (1997) to between 7.9 and 7.7 cal ka BP, which places it very close in age to Northern Hemisphere cooling that started at ~8.6 cal ka BP that may have lasted until 7.9 cal ka BP (Rohling and Pälike, 2005). Rohling and Pälike (2005) have shown that the so-called 8.2 cal ka BP cold event (recognised in the Greenland ice core) is superimposed upon a general background of climatic deterioration. Whether the interruption of S1 is related to the 8.2 cal ka BP cold event or the more general climatic decline at this time is unclear. Bar-Matthews et al. (1999) claim that an excursion in speleothem $\delta^{13}C$ at about 8.2 cal ka BP represents a sudden cooling and decrease in precipitation during the terrestrial "deluge". The SST data from Site 967 and the Red Sea are inconclusive (Fig. 15) but it is interesting to note that at Site 967 there is a peak in salinity that is dated to ~8.2 cal ka BP. Unfortunately, given the uncertainty in the age models for both speleothems and marine cores, it may prove very difficult to determine conclusively whether cooling and increased salinity in the Eastern Mediterranean about 8000 years ago was related to general cooling or the "8.2 ka cold event".

5.7. *Mid-Holocene wet event (~5000 yr BP)*

There is some suggestion in the terrestrial data that there may have been a slightly wetter phase in the mid-Holocene (~5000 yr BP). Lake levels were higher at this time (Frumkin et al., 1994), precipitation calculated from speleothems rose to >500 mm/yr at ~4.8 cal ka BP, and the palaeosol record suggests a slightly wetter climate at about 4.8 cal ka BP. There is a suggestion of a brief wetter phase in the snail record at ~4.5 cal ka BP. Marine records from the Mediterranean and Red Sea do not show any significant variation during the mid-Holocene, although coral records that date from 6.5 to 5.1 cal ka BP suggest increased humidity (relative to present).

6. Conclusions

1. The major northern hemisphere climatic shifts of the last 25,000 years are present in the geological record of the Eastern Mediterranean and Levant.
2. The marine and terrestrial palaeoclimatic records from the region are in general agreement over the timing of changes in climate. However, the records do not always agree as to the severity of some events (such as the Younger Dryas). This may be a problem with the sample resolution of marine sites with low–moderate sedimentation rates or may be a real phenomena caused by differing rates of responses in different proxies and environments.
3. The LGM and early Holocene have received considerable attention, and this is reflected in the availability of data for these intervals. However, data coverage for the transition between these two extremes of climate is poorer. Further studies focused on the Bølling-Allerød and Younger Dryas are required in order to understand the climatic regime of the Levant during these intervals.
4. High-resolution age models are required to understand whether climate events in different regions and environments are synchronous or of the same duration. Where age models are good, there seems to be good agreement between the timing and duration of events, irrespective of environment or geographic position.
5. Output from GCMs for times such as the LGM are generally compatible with geological proxy data, giving confidence in the models themselves.

6. Model output throws light on processes operating within the climate system, and to possible responses within the environment. In particular, models alert us to seasonally dependent factors, which are often difficult to discern from the imperfectly preserved sedimentary record.

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