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Mediterranean climate variability during the Holocene

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Abstract

We present a study on four high sedimentation-rate marine cores with suppressed bioturbation effects, recovered along the northern margin of the eastern Mediterranean. We demonstrate that this region, central to the development of modern civilisation, was substantially affected throughout the Holocene by a distinct cycle of cooling events in the order of 2^{o} C. In the best-preserved cases the onset of these events appears particularly abrupt, within less than a century. The cooling events typically lasted several centuries, and there are compelling indications that they were associated with increased aridity in the Levantine/NE African sector (Rossignol-Strick, 1995; 1998; Alley et al., *1997; Hassan, 1986; 1996; 1997a,b; McKim Malville* et al., *1998). Several of these episodes appear to have been coincident with cultural reorganisations, with indigenous developments (eg. cattle domestication, new technologies) and population migrations and fusion of peoples and ideas (Hassan, 1986; 1996; 1997a,b; McKim Malville, 1998). We infer that climatic events of a likely high-latitude origin (O'Brien* et al., *1995; Bond* et al., *1997; Mayewski* et al., *1997; Alley* et al., *1997) caused cooling and aridity in and around the eastern Mediterranean via a direct atmospheric link, and therefore played an important role in the development of modern civilisation.*

Keywords: Climatic variability, Palaeoceanography, Aegean, Holocene, Foraminifera, Stable isotopes.

Introduction

The Holocene is customarily regarded as a period of climate stability, but recently a distinct millenium-scale repetition of Holocene cold events has been recognised (O'Brien *et al.,* 1995; Bond *et al.,* 1997; Campbell *et al.,* 1998; Bianchi & McCave, 1999). This climatic variability may have driven migratory patterns and settlement behaviour in early societies (HASSAN, 1986; 1996; 1997a, b), but it remains to be established how high-latitude Holocene climate variability affected the near/middle East region. One measure of such climatic variability it is fluctuations in past sea surface temperature (SST) estimated from marine planktonic microfossil records (CITA, 1977; THUNELL, 1977; ROHLING *et al.,* 1993; PFLAUMANN *et al.,* 1996; KALLEL *et al.,* 1997; ORTIZ & MIX, 1997; ROHLING *et al.,* 1997; TARGARONA, 1997; HAYES *et al.,* 1999). We assess the Late Glacial-Holocene SST changes from planktonic foraminiferal abundance variations in four cores (Table 1). IN68-9 from the S Adriatic Sea; C40 from the SW Aegean Sea; LC21 from the SE Aegean Sea; and LC31 from W of Cyprus (Fig. 1), using a simple a-priori grouping of warm versus cool species (expressed as % warm species/(%warm + %cool species), Table 2; Fig. 2 and 3). Similar groupings are found from multivariate statistical analyses of Mediterranean records (CITA, 1977; THUNELL, 1977; ROHLING *et al.*, 1993; HAYES *et al.,* 1999). In subtropical waters, the species of the "warm" group predominate within the shallow, warm, and nutrient-depleted summer mixed layer. Many are spinose and contain photosynthetic symbionts. Species of the "cool" group are mostly present in winter, when SST is reduced and the mixed layer has considerably deepened (> 100 m) and become enriched in nutrients. Despite an obvious overprint of the degree of eutrophication related to the development of the seasonal thermocline, recent modern analogue studies show that such groupings provide useful first-order information on SST fluctuations (PFLAUMANN *et al.,* 1996; KALLEL *et al*., 1997; ORTIZ & MIX, 1997). In corroboration of the SST fluctuations based on planktonic foraminifera (zooplankton), we include a similar warm and cool water grouping for core IN68-9 of selected organic walled cyst species derived from auto-trophic dinoflagellates (phytoplankton) that generally preserve well (Table 2). Both qualitative records differ in detail due to the effects of productivity and preservation constraints, but the basic patterns in intervals with comparable sampling resolution are quite similar (Fig.3).

Table 1
Core characteristics.

Core	Location	Depth
LC-21	35o 39.71' N 26o 34.96' E	1522 m
LC-31	34o 59.76' N 31o 09.81' E	2298 m
IN68-9	41o 47.5' N 17o 54.5' E	1234 m
C-40	36o 56.12' N 24o 04.69' E	852 m

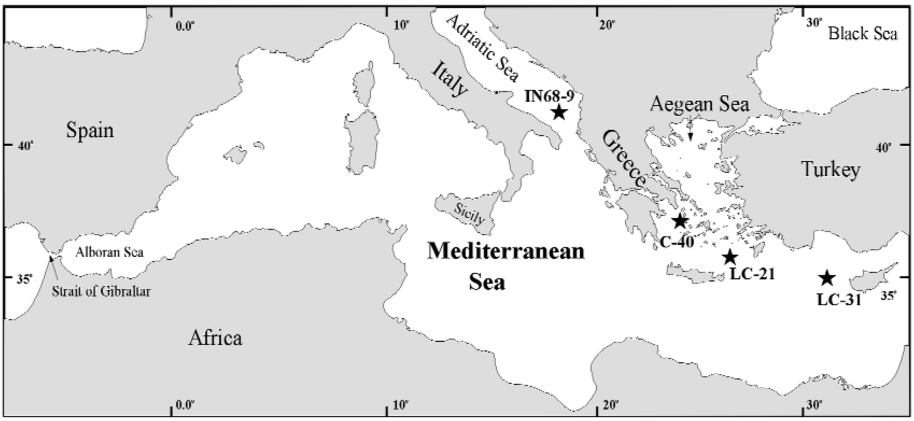

Fig. 1: Map of Eastern Mediterranean, showing locations of cores used in this study.

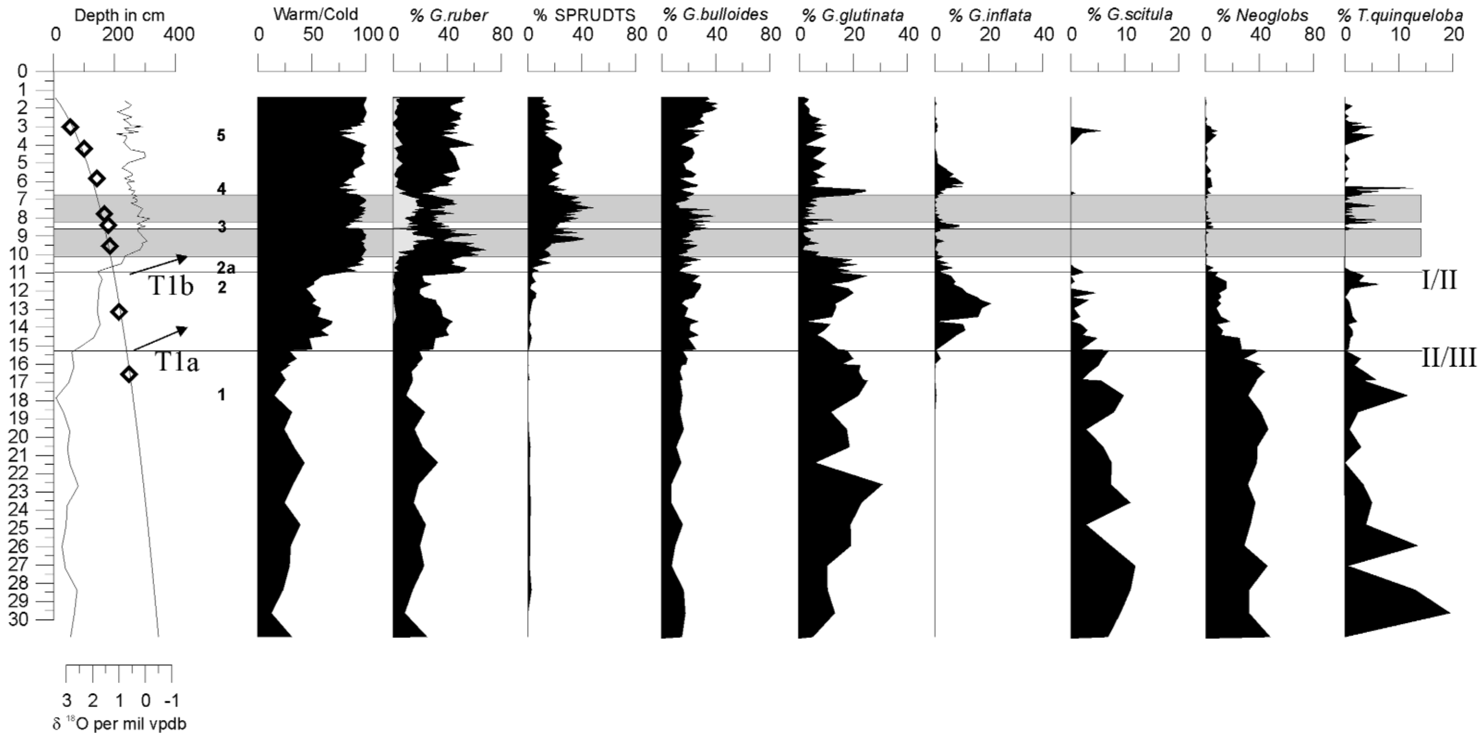

Fig. 2: Time-stratigraphic framework of the core LC-21, using a second order polynomial fit through age-control points given in Table 3. The $\delta^{18}O$ plot is based on measurements of the foraminiferal species *G.ruber* and is expressed in per mil (VPDB). Bold arrows indicate the extent of the isotopic depletions associated with terminations T1a and T1b. The bold numbers indicate the position of cooling events identified in LC-21 primarily from warm/cold faunal plots. Relative abundances of key planktonic foraminifera species are also shown together with the biozonal boundaries I/II and II/III (JORISSEN *et al.*, 1993).

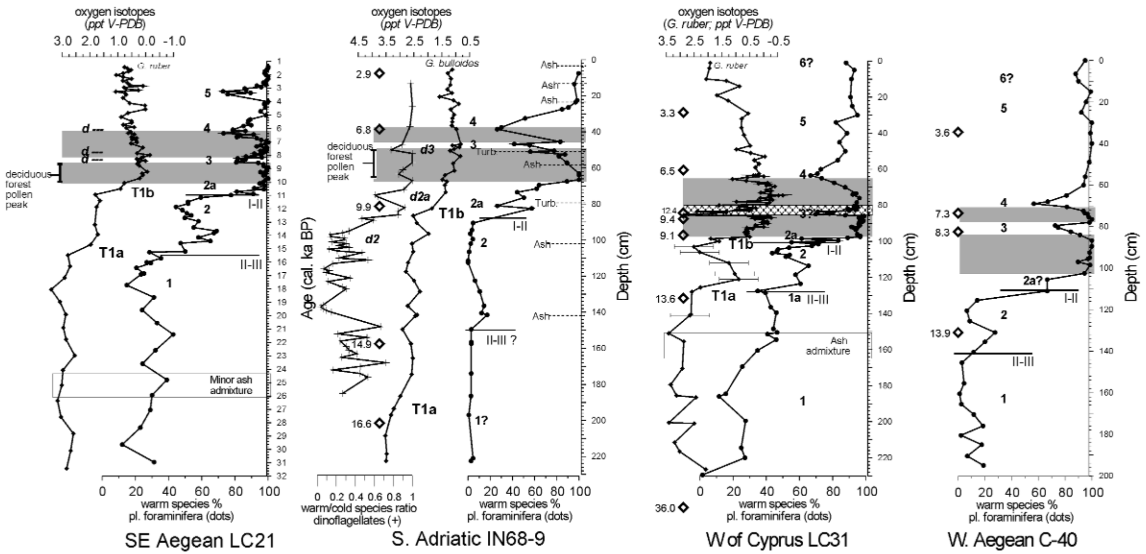

Fig. 3: Records of “warm” versus “cold” planktonic foraminiferal species as discussed in the text. LC-21 is plotted versus time, plots of IN68-9 for clarity show depth subtracting thicknesses of turbidites and ash-layers, and for LC31 and C40 versus true depth. Planktonic foraminiferal biozone boundaries after JORISSEN *et al.* (1993). Cooling intervals are numbered sequentially in bold. Main cooling events in the dinoflagellate-based warm-cold record (see text) for IN68-9 are numbered sequentially in bold italics, with the prefix “d”. In LC21, the dinoflagellate-based temperature record is near 100% throughout the Holocene, and is compromised by low cyst abundances outside the sapropelic intervals; levels where minor cooling events were found are marked “d---”. Grey bands represent sapropelic intervals. The cross-hatched interval within the sapropel in LC31 represents a slump (Table 3). Oxygen isotope stratigraphies are based on *G. bulloides* for IN68-9 and *G. ruber* for LC21 and LC31, from the 250-350 mm size-window. Error-bars in LC31 represent range of variability observed in replicate analyses, for analyses added at the S.O.C. isotope facility to an originally less detailed record analysed at Bremen. The two facilities were intercalibrated by exchange of laboratory standards and replicate sample analyses. Bold lines indicate the two main deglaciation steps: Terminations T1a and T1b. Within the basal sapropel unit, relatively small effects of temperature change on $\delta^{18}O$ are overwhelmed by those of the dramatic summer rainfall increase witnessed by a peak in deciduous forest pollen in both IN68-9 (TARGARONA, 1997) and LC21 (preliminary record; not shown). Warm-cold record for C40 excludes *Turborotalita quinqueloba*, a small-size species that is over represented in this >120 mm study, relative to the other three >150 μm studies.

Table 2
Composition of species-groups used in planktonic foraminiferal and dinoflagellate warm-cold plots.

Planktonic foraminifera (zooplankton)	
Warm group	Cool group
Globigerinoides ruber	*Globorotalia scitula*
Globigerinoides sacculifer	*Turborotalita quinqueloba*
Hastigerina pelagica	*Globorotalia inflata*
Globoturborotalita rubescens	*Neogloboquadrina pachyderma*
Orbulina universa	
Globigerinella digitata	
Globoturborotalita tenella	
Globigerinella siphonifera	
Selected autotrophic dinoflagellate cyst species (phytoplankton)	
Warm group	Cool group
Impagidinium aculeatum	*Bitectatodinium tepikiense*
Impagidinium strialatum	*Spiniferites elongatus*
Impagidinium paradoxum	*Impagidinium pallidum*
Spiniferites mirabilis	

Time-stratigraphic framework

The chronostratigraphy is determined by several AMS^{14}C analyses in each of cores IN68-9, LC21, LC31 and C40, all performed on hand-picked clean planktonic foraminiferal tests (Table 3). A calendar age of 3.578 ka BP (KUNIHOLM *et al.,* 1996) is used for the Santorini ash layer at 81-91 cm in LC21, and 33-36 cm in C40. We calibrate all radiocarbon convention ages using the marine mode of the Calib3.03 programme (STUIVER & REIMER, 1993). It is immediately obvious that some of the AMS^{14}C datings tend towards older ages than expected from the well-established ages of the biozonal boundaries and the Santorini ash layer (Table 3, Fig. 3). This tendency is mostly minor (e.g. the ages around the Santorini ash in LC21), but in some cases a substantial offset is seen. For example, the age immediately below the 5 cm thick slump within the sapropel in LC31 appears about 1000 years too old. This is probably due to the immediate proximity of the mass-transport layer, which itself is about 3000 years older than surrounding sediments. Offsets towards older radiocarbon ages are potentially caused by addition of an uncertain component of older tests to the material analysed, which seems possible in view of the high accumulation rates of the cores and their proximity to the continental slopes. Calibrated ages for LC21 are then plotted versus depth subtracting the thicknesses of the interbedded Santorini ash layer and turbidites. As LC21 has the highest resolution of the four cores in the study and appears to be the most complete record, it is used as a master chronology in the following discussion (Fig. 2). All ages are reported as calibrated AMS^{14}C years BP unless otherwise stated.

Bio-stratigraphic control also provides two well-dated correlation horizons, the biozone I-II and II-III boundaries, as defined on the basis of 11 central Mediterranean marine records with a total of 50 AMS^{14}C ages (JORISSEN *et al.,* 1993). Subsequent studies emphasised Mediterranean-wide applicability and synchroneity of both boundaries (CAPOTONDI *et al.*, 1999; HAYES *et al.,* 1999). This includes the Aegean Sea

Table 3

Age control-points used in the present study, with true depths in the cores, and corrected depths (subtracting thicknesses of turbidites and ash-layers). I-II and II-III are planktonic foraminiferal biozone boundaries (JORISSEN *et al.*, 1993). Radiocarbon convention ages calibrated using marine mode of programme Calib3.03 (STUIVER & REIMER, 1993). Santorini age after KUNIHOLM *et al.* (KUNIHOLM *et al.*, 1996) . Core LC-31 dates include a local ΔR correction of 149±39 years after FACORELLIS et al., (FACORELLIS *et al.*, 1998). Dates in LC-21 are after MERCONE *et al.*, (MERCONE *et al.*, 1999) and in C-40 after GERAGA *et al.*, (GERAGA *et al.*, 2000).

True depth (cm)	Corrected depth (cm)	AMS lab.code	Uncorrected AMS^{14}C age from direct dating or dated horizon (kaBP)	cal.age (ka BP)	cal.age range (1σ) (kaBP)
IN68-9					
11.5	7.5	UTC-500	3.16±0.12	2.93	2.78-3.10
54.5	38.5	UTC-1607	6.39±0.06	6.85	6.77-6.90
157.25	81.25	UTC-501	9.28±0.18	9.93	9.80-10.08
241.5	157.5	UTC-502	13.10±0.20	14.95	14.61-15.30
322.5	201.5	UTC-503	14.20±0.30	16.55	16.15-16.92
510.5	247.5	UTC-504	17.20±0.30	19.82	19.40-20.30
LC-21					
50	50	CAMS-41314	3.35±0.06	3.20	3.11-3.27
95.5	85.5	CAMS-41313	4.29±0.06	4.40	4.32-4.45
137.75	127.75	CAMS-41311	5.59±0.06	5.95	5.90-6.03
161.5	151.5	CAMS-41315	7.48±0.06	7.90	7.94-7.84
174.25	164.25	CAMS-41312	8.12±0.06	8.52	8.59-8.44
179.5	169.5	AA-30364	9.01±0.07	9.54	9.22-9.86
209	199	AA-30365	11.77±0.08	13.14	12.77-13.51
252.5	242.5	CAMS-41316	14.45±0.06	16.85	16.96-16.74
81-91	81	Santorini	3.578		
LC-31					
28.5	28.5	CAMS-45864	3.45±0.05	3.33	3.26-3.36
60.5	60.5	CAMS-45863	6.12±0.05	6.54	6.47-6.62
84.25	81 (slump)	AA-30367	10.90±0.10	12.42	12.27-12.55
87.5	82.5	CAMS-45861	8.74±0.05	9.38	9.34-9.43
96.5	91.5	CAMS-45862	8.50±0.05	9.06	8.98-9.19
131.5	126.5	CAMS-45860	12.04±0.05	13.57	13.47-13.69
247.5	242.5	CAMS-45859	32.96±0.05	35.96 (simple 3kyr addition)	
C-40					
73.5	73.5	Beta-110420	6.83±0.11	7.30	7.20-7.39
82.5	82.5	Beta-110419	7.83±0.14	8.26	8.10-8.38
131	131	Beta-110418	12.35±0.16	13.93	13.72-14.16
33-36	33-36	Santorini	3.578		

(ZACHARIASSE *et al.*, 1997), as is shown also by the results of core C40 presented here. The II/III biozonal boundary in the Adriatic however, does not appear to be entirely synchronous with the corresponding faunal break in our other cores. The chronostratigraphy of Adriatic core IN68-9 gives ages for the zone I-II and II-III boundaries of 10.5 cal ka BP and 14.5 cal ka BP, respectively (JORISSEN *et al.*, 1993; ROHLING *et al.*, 1993, 1997). We note that IN68-9 presents a date almost 1000 years younger for the II/III boundary than that observed in their other cores (JORISSEN *et al.,* 1993) or in the cores presented here. As yet, there is no satisfactory explanation but it is possible that a small hiatus is present in IN68-9 over the II/III boundary.

Cooling events

We identify several cooling events, these are sequentially numbered (in bold), noting their positions relative to the biozonal boundaries, ash layers, and sapropelic units (Fig. 2 and 3). Events 1 and 6 are less accurately timed because of their occurrence near the limits of the investigated records where resolution is lowest. The 2a event (9.6-9.8 ka BP) is only identified positively from core IN68-9 with limited support from the other cores outside the Adriatic. The 1a event (13.4-13.6 ka BP) is only identified in the base of core LC-31, and needs to be considered with caution. The age-ranges determined for the NE Mediterranean cooling events are 1.5-2.0, 3.6-4.0, 5.8-6.2, 8.2-8.6, 9.6-9.8, 11.4-12.4, 13.4-13.6, and 17.0-18.5 ka BP, similar to ages reported for high-latitude N. Atlantic cold events in the Holocene (Table 4) (O'BRIEN *et al.,* 1995; BOND *et al.,* 1997).

The magnitude of the 8 ka BP event (our event 3) in IN68-9 has been estimated in the order of 2 C-degrees (ROHLING *et al.,* 1997). For a full glacial-interglacial temperature difference in the Aegean Sea of about 6 C-degrees (ROHLING & DE RIJK, 1999), comparison of the magnitudes of the cooling events with the full glacial-interglacial amplitude of change in the warm/cold plot of LC21 (Fig. 3) also suggests effects of the order of 2 C-degrees.

The rapidity of the onset of the cooling events is best illustrated using the 8 ka BP event (event 3), since it follows the deposition of the unbioturbated (abenthic) lower sapropelic unit. In LC21, this event starts from one sample to the next. Due to the high accumulation rate in this core we infer that the period of onset for this event was less than a century (Fig. 2). The cooling events typically last several centuries, and appear to be repeated at regular intervals.

Discussion

Together, the timing of the events and the apparent rapidity of changes in the eastern Mediterranean suggest a direct atmospheric link between high latitudes and the NE Mediterranean. Regionally - expressions of cold Atlantic Heinrich events in the glacial Gulf of Lions (NW Mediterranean) also suggest an efficient atmospheric teleconnection between that area and the high latitudes (ROHLING *et al.,* 1998). Today, high latitude Arctic/N.Atlantic perturbations are rapidly transmitted to the northern edge of the Mediterranean via orographically channelled winter outbreaks of cold and dry air from high latitudes over the NW Mediterranean, the Adriatic and the Aegean Sea (LEAMAN & SCHOTT, 1991; MARIOPOULOS, 1961; POULOS *et al.,* 1997). An increased duration and/or intensity of these atmospheric outbreaks would cause not only cooling, but also aridity. Palaeobotanical records around the Aegean Sea (ROSSIGNOL-STRICK, 1995), and the $\delta^{18}O$ and $\delta^{13}C$ profiles of Soreq Cave (Israel) (BAR MATTHEWS *et al.*, 1999) both suggest notable cooling and aridity during the Younger Dryas and around 8 ka BP.

Table 4

Comparison of ages of abrupt climate events. This paper: planktonic foraminifera-based cooling events in the Adriatic and Aegean/NE Levantine Seas. BOND *et al.,* 1997: high latitude N. Atlantic cold events from marine records compared with GISP2 ice record. HASSAN *et al.,* 1997a,b and MCKIM MALVILLE *et al.* 1998: NE African arid/cold spells and impact on early societies. GASSE *et al.,* 1996: aridity in the Lake Bangong area (W. Tibet). O'BRIEN *et al.,* 1995: cold events with increased meridional atmospheric circulation affecting the Greenland Ice Sheet. Last column: mean ages based on boundaries of discrete events listed in the table, with standard error of the mean (cal. ka BP).

This paper	Bond *et al.,*1997	Hassan 1997a,b	McKim. *et al,* 1998	Gasse *et al.,* 1996	O'Brien *et al.*, 1995	Event
--	--	--	--	0.7	0.0-0.6	--
1.5	1.2-1.7	1.5-2.0	--	1.3-2.1	--	6?
3.2-4.0	2.5-3.1	--	--	--	2.4-3.1	5
--	4.1-4.8	4.0-4.4	--	<3.9	--	
5.8-6.2	5.6-6.3	5.5-6.7*	<5.8 arid	5.7-6.3	5.0-6.1	4
8.2-8.6	7.6-8.5	7.7-8.3	7.4-8.1***	7.7-8.2	7.8-8.8	3
9.6-9.8	9.4-9.7	9.2-9.5	--	9.6-9.9	--	2a‡
11.4-12.4	10.7-12.6**	--	--	--	>11.3	2
13.4-13.6	13.0-13.4	13.4-14.0	--	--	--	1a‡‡
--	14.6-15.0	--	--	--	--	--
17.0-18.5						1

[*] Onset of major drought and cold conditions around 6.7 ka BP, culminating around 5.5 ka BP with shift of societies into active Nile channels (HASSAN, 1997b).
[**] Their event 8 and Y.D. combined (BOND *et al.,* 1997).
[***] Combination of two separately reported events (7.4-7.5 and 7.9-8.1) (MCKIM MALVILLE *et al.*, 1998).

‡ Only weakly supported outside of Adriatic.
‡‡ Identified in LC-31 only, to be treated with caution.

Moreover, overviews of continental palaeoclimate and archaeological records from the Levant and NE Africa show arid spells of similar age to our Adriatic/Aegean cooling events (Table 4), coinciding with important reorganisations in settlement patterns and agricultural behaviour of early societies (HASSAN, 1997a,b; MCKIM MALVILLE *et al.,* 1998). For example, "*droughts ... between 8.3 and 7.7 ka BP encouraged movements of nomadic cattle keepers from the Egyptian Sahara and bands possessing sheep/goats and wheat/barley from Southwest Asia to the banks of the Nile Valley and Nile Delta. This event was crucial for the agricultural developments in the Nile Valley*" (HASSAN, 1998) (our event 3). In addition, "*It was during the 7th and 6th millennium bp episodes of severe droughts that dwellers of the desert drifted towards the Nile Valley, and elsewhere, in search of water, food, and fodder*" (HASSAN, 1997), (our event 4, see Table 4). We infer that the development of modern civilisation was punctuated by abrupt cooling/aridity events due to an atmospheric teleconnection between the "cradle of modern civilisation" and the high latitudes. We propose that this connection operates through significant increases in winter northerly air flow over the eastern Mediterranean.

The observed temporal spacing of these cooling events appears (Table 4) similar to that of planet-induced sun-tide cycles (LAMB, 1972). Ice-core 14C residual series indicate this periodicity to be around 1450 years, suggesting that fluctuations in solar forcing may indeed have affected glacial and Holocene climate on these time-scales (MAYEWSKI *et al.,* 1997). However, any external forcing would be likely to drive internal oscillations within the earth's ocean/climate system, such as the recently invoked "oceanic oscillator", a fluctuation in the global thermohaline circulation, the nature of which remains to be determined (BIANCHI & MCCAVE, 1999). We tentatively propose that the Mediterranean may form part of the "oceanic oscillator". Although the Mediterranean appears to behave initially as a passive response-basin to high-latitude cold events, resultant increases in buoyancy loss would increase its saline outflow flux into the Atlantic. Mediterranean salt today contributes to the water-column destabilisation needed for North Atlantic Deep Water (NADW) formation (eg. REID, 1979; HILL & MITCHELSON-JACOB, 1993). If the Mediterranean salt flux into the Atlantic were sufficiently increased to trigger enhanced NADW formation (for assessment of current hypotheses, see BRYDEN & WEBB, 1998), then consequent increases in heat-advection to high latitudes could have contributed to termination of high-latitude cold events. This possibility merits consideration, particularly in view of the anthropogenic impact on Mediterranean deep water properties and the recently-observed and rapid major salt redistribution within the eastern Mediterranean (BETHOUX *et al.*, 1990; LEAMAN & SCHOTT, 1991; ROHLING & BRYDEN, 1992; BETHOUX & GENTILLI, 1996; ROETHER, 1996).

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