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Introduction

Holocene climate and cultural evolution in late prehistoric–early historic West Asia

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Abstract

The precipitation climatology and the underlying climate mechanisms of the eastern Mediterranean, West Asia, and the Indian subcontinent are reviewed, with emphasis on upper and middle tropospheric flow in the subtropics and its steering of precipitation. Holocene climate change of the region is summarized from proxy records. The Indian monsoon weakened during the Holocene over its northernmost region, the Ganges and Indus catchments and the western Arabian Sea. Southern regions, the Indian Peninsula, do not show a reduction, but an increase of summer monsoon rain across the Holocene. The long-term trend towards drier conditions in the eastern Mediterranean can be linked to a regionally complex monsoon evolution. Abrupt climate change events, such as the widespread droughts around 8200, 5200 and 4200 cal yr BP, are suggested to be the result of altered subtropical upper-level flow over the eastern Mediterranean and Asia.

The abrupt climate change events of the Holocene radically altered precipitation, fundamental for cereal agriculture, across the expanse of late prehistoric–early historic cultures known from the archaeological record in these regions. Social adaptations to reduced agro-production, in both dry-farming and irrigation agriculture regions, are visible in the archaeological record during each abrupt climate change event in West Asia. Chronological refinement, in both the paleoclimate and archaeological records, and transfer functions for both precipitation and agro-production are needed to understand precisely the evident causal linkages.

Keywords: Climate change; West Asia; Monsoon; Neolithic; Early Bronze Age

Introduction

It is now well established that climate varied significantly since the last glacial maximum. This variation includes not only the deglacial oscillation of the Bolling/Allerod and the Younger Dryas, but also the Holocene (Bond et al., 2001; Mayewski et al., 2004). During the Holocene, the Asian continent South of about 45° latitude witnessed the rise and occasionally abrupt collapse of many agriculture-based societies coincident with abrupt climate changes (Weiss, 2000; Weiss and Bradley 2001). Around the 8.2 ka event, farming communities across West Asia were reduced radically with some habitat-tracking to sustainable environments (Weninger et al., this volume). Climate and societal change also co-occurred prominently during the late mid Holocene and early late Holocene, roughly 5500 to 3500 yr BP. Abrupt climate change during the earlier part is documented in Africa (Gasse, 2000; deMenocal et al., 2000), and Asia (Enzel et al., 1999). In west Asia, at the 5.2 ka event, late Uruk period society in southern and northern Mesopotamia collapsed (Postgate, 1986; Weiss, 2003). Similarly, at the 4.2 ka event, synchronous changes among Early Bronze Age societies suggest a causal link between the event's precipitation diminution and collapse of the politico-economic superstructures dependent upon cereal agriculture (Weiss et al., 1993; Cullen et al., 2000; Staubwasser et al., 2003; Drysdale et al., 2006; Arz et al., this volume).

These kinds of data have forced accelerated discussion of climate events and the link between climate events and cultural change in both the paleoclimate and archaeology communities, where the causal link has been criticized or rejected by some (e.g., Possehl, 1997a,b; Coombes and Barber, 2005; Madella and Fuller, 2006).

From the climatic viewpoint, confusion may arise from conflicting observation of change between seemingly adjacent

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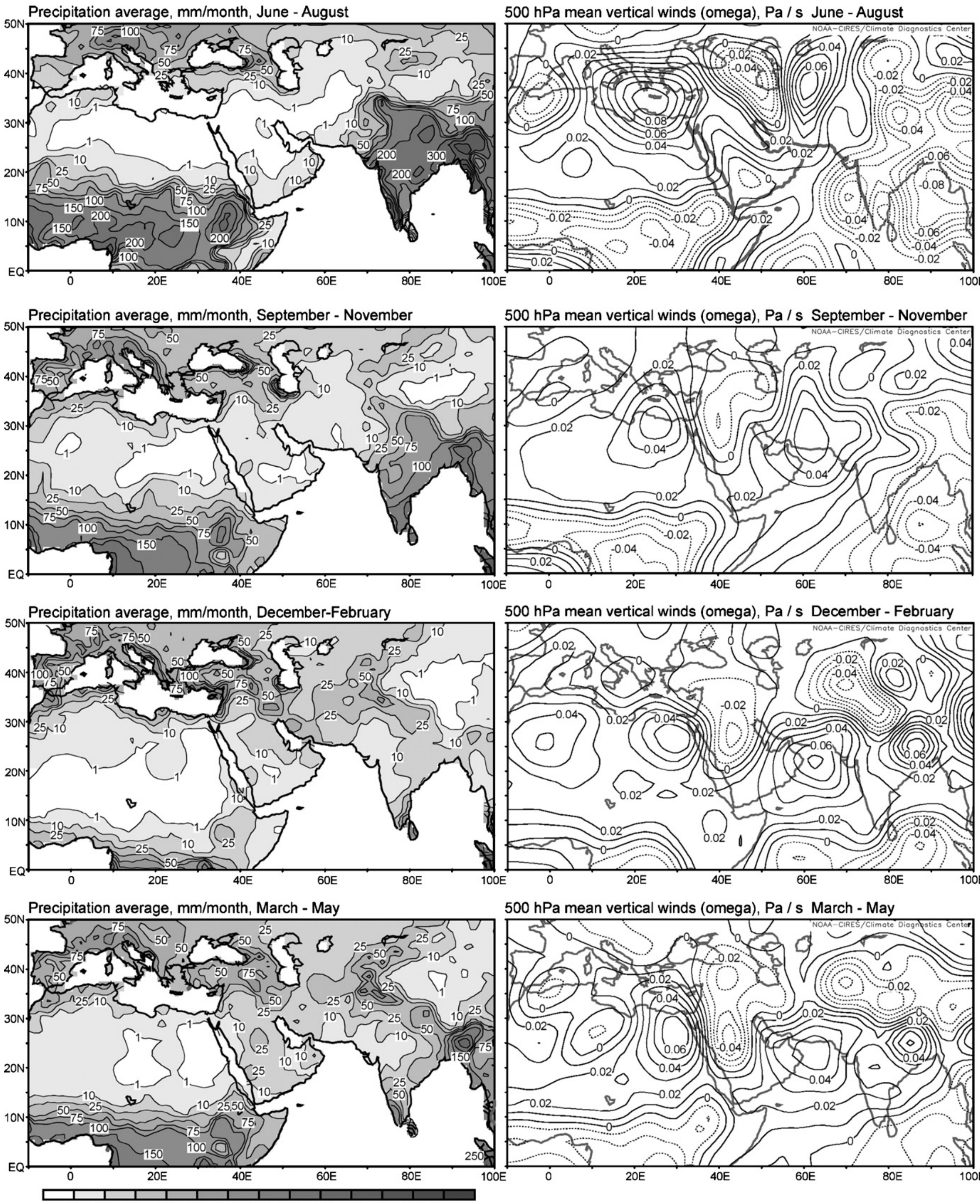

Figure 1. Seasonal averaged of precipitation (mm/month) and vertical winds at 500 hPa (Pa/s). The precipitation maps are the 1960–1991 averages (1° resolution) from the Global Precipitation Climatology Center (Rudolf et al., 1994). The vertical winds (omega) are from the NCEP/NCAR Reanalysis Project (Kalnay et al., 1996). Units of Pa/s are proportional to m/s and reflect the common use of pressure units (Pa) for height levels instead of meters.

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regions. For example, a 4.2 ka drought event is observed in a number of places in the Northern Hemisphere (Booth et al., 2005), but the database remains patchy. From the archaeological viewpoint, uncertainty about the causal linkage between abrupt climate change and social collapse derives from chronological imprecision and the uncertain ability of societies to adapt to the abruptness, magnitude and duration of environmental change (Weiss and Bradley, 2001).

This overview attempts to summarize the climate dynamics for the region roughly between the eastern Mediterranean and South Asia. We discuss present day precipitation patterns and controlling mechanisms. Changes in past precipitation within the region are summarized and potential causes are suggested. The social impact of precipitation change is discussed, particularly for the events around 8200 cal yr BP (8.2 ka event), 5200 cal yr BP (5.2 ka event), and 4200 cal yr BP (4.2 ka event).

Present day precipitation climatology between the eastern Mediterranean and South Asia

There is a strong contrast in the distribution of rainfall throughout the year between the regions of the eastern Mediterranean and West Asia on the one hand, and southern Asia on the other (Fig. 1). Roughly, east of the Arabian Sea the moist season lasts from June to September, i.e. during the south Asian summer monsoon. West of the Arabian Sea and across the eastern Mediterranean the moist season comprises the winter (November–February) and spring months (March–May). The transitional region between the northern Arabian Sea and Tibet receives rain both in winter/spring and during the summer monsoon. Part of the spatial distribution pattern of rainfall can be attributed to the orographic setting (Fisher and Membery, 1998; Evans et al., 2004). The seasonality pattern is largely the result of the circulation regime caused by the South Asian summer monsoon from June to August and early September (Rodwell and Hoskins, 1996; Webster et al., 1998). Deep convection brings plenty of rainfall to India (Fig. 1). The corresponding region of descending air lies over Arabia, where summers are dry. This was termed the “transverse monsoon” circulation by Webster et al. (1998). However, a close inspection of all rainfall charts brings more features into light. An alternating pattern of relatively dry and relatively moist regions emerges in the latitude band roughly between 20° and 45°N from 20° West to 100° East that does not entirely mimic the orographic setting. For example, in the summer season (June–August, Fig. 1) the eastern Mediterranean is dry, the eastern Black Sea, Caucasus and southern Caspian is wet, central Asia east of the Caspian is generally dry despite the presence of mountains, the northwestern margin of the Asian high altitudes is wet, but the northeastern margin is dry again despite the presence of mountains. A similar alternating pattern can be found in the same latitude band for the seasonally averaged vertical wind field. Over the eastern Mediterranean subsidence prevails, whereas over the eastern Black Sea and southern Caspian, uplift occurs. Over central Asia, air tends to descend, but uplift prevails at the northwestern margin of the Asian highlands. Further to the East, subsidence occurs once more. During the other seasons, a similar alternating pattern of comparably dry and moist regions corresponds in the same manner with an alternating pattern in the vertical wind field, only displaced to the south (Fig. 1). Uplift occurs over Arabia without the presence of high mountains. Obviously, the above latitude bands and alternating pattern coincide with the location of the subtropical jet stream, but how does flow in the upper troposphere connect to the observed features on the ground and the lower to middle troposphere?

Precipitation is generally favored when air is allowed to rise. Subsiding air warms on descent, so relative humidity falls and the probability of rainfall is low. As such, the correlation between rainfall and vertical winds observed above is not surprising. The causality between convection over India and broad subsidence over Arabia is intuitive; rising air must be compensated by descent elsewhere. The causality for the alternating pattern in the zone of westerlies further north, however, is more complicated. The vertical wind field is affected by Rossby waves in the upper level flow (e.g. Gordon et al., 1998). Depending on whether flow within these horizontal waves is towards the equator or towards the poles, air underneath is forced to descend or rise respectively (Fig. 2). Poleward flow within a long-wavelength Rossby wave in the subtropics is associated with divergence. This favors cyclone development in the lower troposphere. Under the upper-level ridge region, there is northward advection of warm air. Divergence in the upper level and warm advection below lead to uplift in the lower troposphere. Consequently, surface cyclones may intensify. At the back of a surface cyclone, a cold front may be present which then also brings rain. In contrast, upper-level flow towards the equator is associated with convergence. Below the upper level trough there is southward advection of cool air. Unless a cold front is present, southward advection of cold air into warmer regions behind the preceding surface low and upper-level convergence both result is subsidence and low probability of rain.

Rodwell and Hoskins (1996) demonstrated that the South Asian summer monsoon forces upper-level Rossby waves and causes the observed alternating uplift and descent pattern over West Asia and the eastern Mediterranean (Fig. 1). Rossby waves in the other seasons exist for different reasons, but the fundamental relationship between vertical winds induced by the waves and precipitation is equally valid (Fisher and Membery, 1998). In the following chapters we will discuss the precipitation pattern in relation to vertical winds induced by

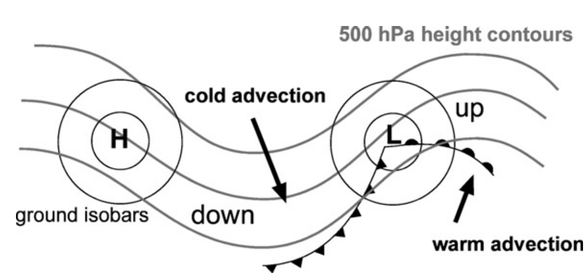

Figure 2. The simplified relationship between Rossby waves in upper-level horizontal flow, vertical flow in the lower level and surface pressure pattern drawn after Gordon et al., 1998).

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the transverse monsoon as well as the Rossby waves and attempt to explain the Holocene climate evolution of West Asia in these terms.

The influence of the South Asian summer monsoon on West Asia, North Africa and the Mediterranean

The fundamental driving forces of the South Asian summer monsoon over South Asia have been extensively studied and summarized (Ramage, 1966; Krishnamurti and Surgi, 1987; Webster et al., 1998; Gadgil, 2003). The summer precipitation maximum over the Indian Peninsula is the direct result of deep convection over the northern Bay of Bengal forced by sensible and latent heating over Tibet and the warm Bay of Bengal. Convective systems may propagate northwestwards along the Ganges plain, dissipating as they progress. This causes the strong precipitation gradient in Northwest India (Fig. 1). In the upper troposphere a corresponding broad center of divergence with outwards flowing air persists over Tibet and the northern Bay of Bengal. An important feature in the South of this region is an upper level easterly jet over southern India and the Arabian Sea, the tropical easterly jet. Slowing of the jet over the Western Arabian Sea causes subsidence over Pakistan, Arabia and Somalia. Any convection from the heated ground is capped by the subsiding air from above and confined to the lower troposphere (Ramage, 1966; Webster and Fasullo, 2003). The resulting shallow heat lows are generally associated with drought. They are a characteristic of the deserts surrounding the Arabian Sea.

Westerly winds in the middle and upper troposphere between the eastern Mediterranean and the Aral Sea meander in horizontal Rossby waves (Rodwell and Hoskins, 1996). These waves are the result of diabatic heating from the region of deep monsoon convection. The wave pattern is such that flow towards the equator occurs over the eastern Mediterranean and large-scale subsidence prevails as a consequence of cold advection from the North. This, in turn, prohibits the development of surface cyclones. Subsidence prevails in the eastern Mediterranean, Northeast Africa and the Red Sea (Rodwell and Hoskins, 1996; Eshel, 2002; Ziv et al., 2004a); see also Figure 1. As the northerly air warms during descent, relative humidity decreases and overall drought prevails. In contrast, warm northward advection and uplift occurs over the eastern Black Sea, Caucasus and southern Caspian region, where summers are comparably moist. This allows the development of cyclones in the eastern Black Sea (Trigo et al., 1999). The same mechanisms of opposite signs operate over the dry Aral Sea region and the comparatively wet mountains at the northwestern fringe of the Asian highlands. It should be noted that this describes a seasonal average of essentially transient features.

To understand the potential impact of past monsoon variability on the climate of West Asia and the eastern Mediterranean, a more detailed look at the dynamic South Asian monsoon is necessary. An important aspect of the South Asian monsoon is its intra-seasonal variability, expressed as intervals of rain over India, termed “active monsoon”, and intervals of reduced rain, termed “weak monsoon” or even drought, termed “break monsoon”. The common perception of the South Asian monsoon is equal to the synoptic situation during the “active monsoon”. Subsidence and dry heat lows West of the Arabian Sea as explained above are also part of the “active monsoon” (Ramage, 1966; Webster et al., 1998; Gadgil, 2003). Spells of weak rainfall interrupt the “active monsoon” with a frequency of about 40 days. In about three out of four summer monsoon seasons, one to three of these “weak monsoon” spells intensify to “break monsoon” spells, lasting up to 2 weeks (Krishnan et al., 2000; Gadgil and Joseph, 2003). Years of poor total monsoon rainfall are associated with long and frequent “break monsoon” spells. Atmospheric circulation and rainfall pattern across the entire Indo-Pacific and South Asian domain are fundamentally different during a “break monsoon” (Krishnan et al., 2000; Gadgil and Joseph, 2003; Joseph and Sijikumar, 2004). Monsoon convection no longer occurs over the Bay of Bengal, but resides over the eastern equatorial Indian Ocean. Surface winds are altered. While most of India experiences drought, rain falls over the Himalayas and their foothills. Tibet and the northwestern Chinese deserts experience rainfall just prior to a ”break” onset (Krishnan et al., 2000). Two features of the “break monsoon” are important for rainfall distribution over West Asia. First, the tropical easterly jet and associated subsidence over Arabia is diminished, no longer confining the heat lows to the lowest troposphere (Ramage, 1966). Second, an anomalous Rossby wave pattern occurs in upper level westerly flow between 30° and 40° N (Gadgil and Joseph, 2003, see their Fig. 13c). The consequences of this anomalous Rossby pattern have not yet been worked out in detail, but they most likely affect vertical winds and rainfall pattern in the entire latitude band. Interestingly, it is such variability of upper tropospheric flow that leads to rare rainfall events in the present day eastern Mediterranean summer (Spanos et al., 2003). In their idealized model, Rodwell and Hoskins (1996) do not observe wide spread subsidence over the eastern Mediterranean when the center of monsoon convection resides over the equatorial Indian Ocean.

The evolution of the South Asian summer monsoon and its effects on Holocene climate of West Asia and North Africa

A considerable number of paleoclimate records exist from the Asian and North African monsoon as well as the eastern Mediterranean. The North African monsoon was stronger and may have reached farther northwards than at the present day (Gasse, 2000; Rohling et al., in press). Northeast Africa was generally wetter (Hoelzmann et al., 2004). Records thought to represent rainfall in the South Asian summer monsoon domain, for example from southern Tibet (Gasse and van Campo, 1994), off the Ganges-Brahmaputra delta (Kudrass et al., 2001), off the Indus delta (Staubwasser et al., 2002), or Oman (Neff et al., 2001; Fleitmann et al., 2003) all show an early and mid Holocene rainfall maximum (Fig. 3). Enhanced upwelling in the western Arabian Sea is interpreted to reflect stronger early and mid Holocene summer monsoon winds (Overpeck et al., 1996; Sirocko et al., 1996; Gupta et al., 2003; Jung et al., 2004).

Enhanced African monsoon rain in the early Holocene led to higher discharge from the river Nile (Scrivner et al., 2004). Stratification of the Mediterranean Sea resulted in the

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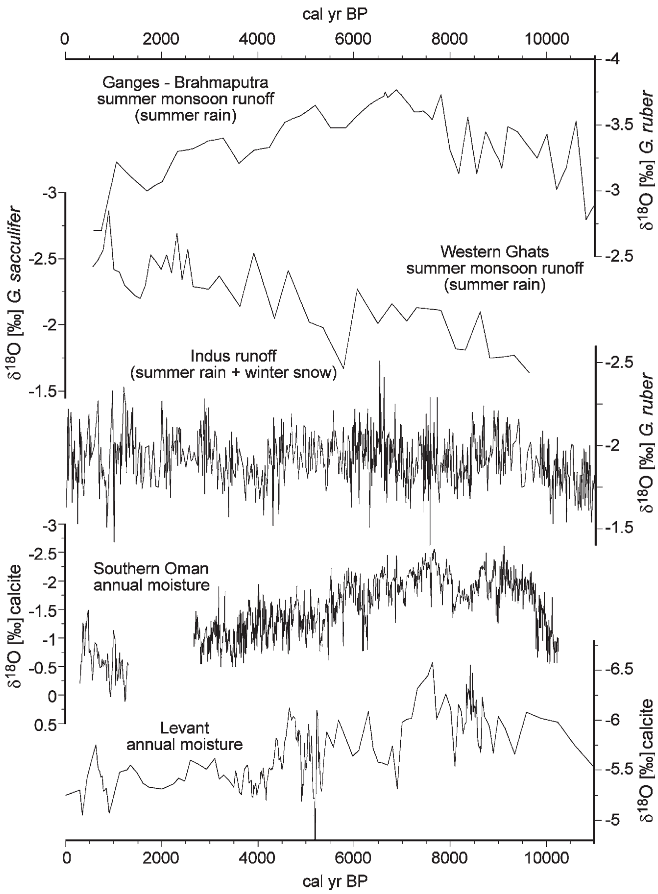

Figure 3. A selection of oxygen isotope proxy records of moisture change during the Holocene for South Asia, West Asia, and the eastern Mediterranean. From Top to bottom: Bay of Bengal off the Ganges-Brahmaputra delta (Kudrass et al., 2001); Eastern Arabian Sea off the Western Ghats (Sarkar et al., 2000); Northeastern Arabian Sea off the Indus delta (Staubwasser et al., 2002, 2003); continental South Oman (Fleitmann et al., 2003); continental Israel (Bar-Matthews et al., 1997).

deposition of a sapropel layer at that time (Calvert and Fontugne, 2001). Similarly, northern Red Sea summers are believed to have been wetter (Arz et al., 2003). The entire eastern Mediterranean region – well outside the direct influence of monsoon rain – also appears to have been considerably moister in summer, as shown by pollen records (Rossignol-Strick, 1999). Evaporation from the basin was lower (Rohling, 1999, Myers, 2002), a result broadly confirmed by marine foraminifer records (Emeis et al., 2000; Rohling et al., 2002). Levantine paleo-rainfall records also indicate a wetter early Holocene (Bar-Matthews et al., 1997, 2003) (Fig. 3), although these are believed to reflect higher winter rainfall.

From the discussion in the previous section it follows that a stronger monsoon over South Asia with a synoptic situation comparable to the present day “active monsoon” is not compatible with enhanced summer monsoon rainfall across the Arabian Peninsula or Northeast Africa. Neither is it compatible with moister summers in the eastern Mediterranean. A stronger South Asian monsoon would have enhanced subsidence and drought in all these regions. How can this apparent contrast be reconciled? It is important to note that proxy records of the South Asian paleo-monsoon rainfall listed so far do not adequately represent the region of present day rainfall during the “active monsoon”. Instead, they represent runoff from the eastern Himalaya (Ganges-Brahmaputra) and western Himalaya and Karakoram (Indus), or rainfall on Tibet. In these regions, rain falls predominantly in association with the “break monsoon” (Krishnan et al., 2000; Gadgil and Joseph, 2003). Only a few records exist from inside the domain of the “active monsoon”, mostly situated off the Western Ghats on the Indian Peninsula. These records suggest that rainfall was reduced during the early Holocene (Sarkar et al., 2000; Thamban et al., 2001) (Fig. 3). South Asian summer monsoon rain was thus not uniformly enhanced during the early and mid-Holocene, but appears to have been reduced over the Indian peninsula. Consequently, the early-mid Holocene monsoon maximum observed in the other regions cannot simply represent a stronger South Asian monsoon of the same kind as at present. Instead, the paleoclimate records suggest that a different type of monsoon prevailed with a different rainfall distribution.

The complex Holocene evolution of the South Asian monsoon may be explained by a more frequent or longer occurrence of a synoptic situation resembling the present day “monsoon breaks” – with convection over the equatorial Indian Ocean and the Himalaya – and a weaker or less frequent occurrence of the “active monsoon” (Staubwasser, 2006). A dominant “break”-type synoptic situation with enhanced South Asian monsoon precipitation along the southern front of Tibet, but not the Indian Peninsula, is consistent with results of most recent global circulation model studies (Bush, 2004, see his Fig. 2). This interpretation is also consistent with enhanced early Holocene upwelling. Upwelling is mostly a function of axis position of the low-level monsoon jet over the Arabian Sea (Anderson and Prell, 1992), which remains favorable during “break monsoon” spells (Joseph and Sijikumar, 2004, see their Fig. 6). The lack of correlation between the abundance of planktonic foraminifers *G. bulloides* (a proxy for upwelling) and rainfall in India during the last one and a half century (Anderson et al., 2002), however, suggests that upwelling records are not very representative of paleo-monsoon rainfall over India.

As a result of a weakened “active monsoon” over the Indian Peninsula during the early- mid Holocene, subsidence over Arabia should also have weakened. Summer rainfall would thus have been more probable over the Arabian Peninsula. This may explain the early Holocene paleo-rainfall maximum observed in Oman stalagmites (Neff et al., 2001; Fleitmann et al., 2003), the presence of lakes in the central Arabian desert (McClure, 1976) and moister conditions in Yemen (Davies, this volume). Considering the dominant proportion of winter and spring rain over Arabia (Fisher and Membery, 1998), however, it is unlikely that different summer conditions were the only reason for a moister early Holocene.

Moister summer conditions over the eastern Mediterranean (Rossignol-Strick, 1999) require significantly reduced subsidence. Applying the observation by Rodwell and Hoskins (1996), longer or more frequent “monsoon breaks” would have altered the average upper level Rossby wave pattern and reduced subsidence over the eastern Mediterranean, Northeast

Africa, and Red Sea (see discussion above). In the light of the anomalous Rossby pattern over the whole of West Asia, as modeled by Rodwell and Hoskins (1996) and observed during the "break monsoon" (Gadgil and Joseph, 2003), a fundamentally different rainfall pattern over West Asia is conceivable for the early Holocene.

Controls of fall, winter and spring rainfall between the eastern Mediterranean and South Asia

Trigo et al. (1999) defined regions in the eastern Mediterranean of preferential cyclone genesis. In winter and spring, these are the Aegean Sea and Cyprus. Similar to the situation in summer, the principal control of eastern Mediterranean precipitation in fall, winter and spring is the vertical motion (instability) within the lower and middle troposphere as a consequence of Rossby waves in the upper level (Eshel and Farrell, 2000, 2001; Eshel, 2002, Ziv et al., 2006). Rainfall occurs when the upper level wave pattern favors cyclone development over the eastern Mediterranean (Kahana et al., 2002, Zangvil et al., 2003; Tsvieli and Zangvil, 2005; Türkes and Erlat, 2006) or moisture transport from tropical sources (Ziv et al., 2004b). Note that the eastern Mediterranean is situated at the boundary between ascent over the Middle East and descent over northeastern Africa in winter and spring (Fig. 1), and therefore should be sensitive to slight shifts in the average Rossby wave and associated vertical wind pattern. Further to the south, in the central Red Sea, it is also the upper level Rossby wave pattern that controls precipitation (Flohn, 1965). Over Arabia, rain falls when upper-level Rossby waves favor cyclone development at the surface (Fisher and Membery, 1998). Similarly, in Iran rainfall is significantly related to the upper level flow (Alijani, 2002). Cyclones constantly move eastwards across West Asia. They may provide moisture during winter and spring as far west as the Karakoram and western Himalaya (Ramage, 1966; Karim and Veizer, 2002). However, only where tropospheric instability leads to ascent such cyclones may be reactivated, intensify and produce rainfall. A regional climate model study of West Asia indicates that the variation of atmospheric instability is perhaps the most important factor that determines interannual rainfall variability (Evans et al., 2004).

Studies that relate surface parameters to rainfall generally observe a correlation between rainfall in the eastern Mediterranean with the North Atlantic oscillation (NAO) index, i.e. the interannual variation of the surface pressure gradient between the Icelandic low and the Azores High (Cullen and deMenocal, 2000; Felis et al., 2000, Rimbu et al., 2001; Lolis et al., 2002). Elevated near-surface pressure over Greenland correlates with anomalous cyclonic activity and enhanced rainfall over the eastern Mediterranean. The physical mechanism responsible for the correlation is upper-level flow (Eshel and Farrell, 2000, 2001; Türkes and Erlat, 2006, Ziv et al., 2006). About a third to a quarter of northeastern Mediterranean precipitation variability can be linked to the NAO (Lolis et al., 2002). In Asia Minor, about a quarter of precipitation variability can be explained by the NAO, but southwards and eastwards into the Levant, Syria and Iraq, the NAO influence weakens substantially (Cullen and deMenocal, 2000; Ziv et al., 2006). Evans et al. (2004) used a regional circulation model embedded in a global model to study the mechanisms controlling precipitation in West Asia and found little influence of the NAO in West Asia beyond Asia Minor. Although upper-level flow remains a fundamental control, it is no longer governed by variability in the North Atlantic.

In summary, it appears that the key factor in the understanding of mechanisms controlling eastern Mediterranean and West Asian rainfall is upper and middle tropospheric flow and the associated Rossby wave pattern. How may the Rossby wave pattern have been influenced during the Holocene? Rossby waves in the troposphere are generally produced by orographic and thermal anomalies (Grotjahn, 1993). The orographic configuration determines Mediterranean and West Asian rainfall distribution in general (Fisher and Membery, 1998; Trigo et al., 1999; Evans et al., 2004) and cannot directly contribute to climate variability on the time scale considered. Thermal anomalies, however, have occurred during the Holocene, for example around 8200 yr ago in the North Atlantic (Grootes and Stuiver, 1997). Considering the impact of modern North Atlantic sea surface temperature changes on the thermal structure of the atmosphere and upper tropospheric flow (Rodwell et al., 1999) and, in turn, its effect on vertical winds and rainfall probability in the eastern Mediterranean (Eshel and Farrell, 2000; 2001), Atlantic sea surface temperature changes are one likely driver of changing rainfall in the eastern Mediterranean. However, other factors are clearly present and appear to become increasingly important eastwards of the Mediterranean. Snow cover of central Asia (and the thermal structure of the atmosphere) is an important factor in modern monsoon variability and West Asian winter climate (Meehl, 1994). Asian snow cover, in turn, is a function of tropical interannual climate variability, of which the El Niño-Southern oscillation (ENSO) is a fundamental aspect. A correlation between rainfall in parts of West Asia and ENSO, or aspects of it, has been observed (Rimbu et al., 2003; Ziv et al., 2006). The physical mechanism behind such observation may again be related to upper tropospheric flow, particularly in subtropical latitudes. One factor that has a profound influence on the Northern Hemispheric upper tropospheric flow in winter is the intensity of Southern Hemisphere tropical convection and the associated Hadley circulation (Yang and Webster, 1990). As a consequence of high tropical convection intensity over Africa, Australia, Indonesia and New Guinea, upper-level flow over the northern subtropics in the same meridional band is strengthened. Notably, the upper-level Rossby wave pattern over the eastern Mediterranean is altered as a result of changes in western Pacific sea surface temperatures and convection intensity (Ziv et al., 2006). Consequently, the paleo-precipitation climatology of the Northern Hemisphere subtropics of Africa, southern Europe, and Asia, cannot be explained without consideration of past tropical convection intensity and ENSO evolution.

Snow cover and glaciation, particularly of the central Asian high altitudes, may provide a positive feedback to forced

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changes in the upper and mid-level Rossby wave pattern. A study by Bush (2000) shows an anomalous Rossby wave pattern in spring as a consequence of a cold anomaly over the western Himalaya and Karakoram because of enhanced glaciation in the mid-Holocene. As a result, a change in the alternating uplift/descent pattern between the eastern Mediterranean and the Himalaya is observed. For example, uplift expands from the western Himalaya over the longitude of the Thar desert between India and Pakistan, a situation that would promote rainfall and is in contrast to the present day. The mid-Holocene is the only time where lakes were permanently present in the Thar desert (Prasad and Enzel, this volume).

Winter and spring rain during the Holocene

As winter/spring is the main rainy season in the Middle East and eastern Mediterranean, several paleoclimate records have been interpreted as reflecting trends in Holocene amounts of winter and spring rain. Pollen records of the region corroborate this interpretation (Rossignol-Strick, 1999). While winter/spring rain over the central and eastern Mediterranean decreased across the mid-Holocene (Bar-Matthews et al., 1997; 2003, Frisia et al., this volume), and probably around the Aegean Sea (Rohling et al., 2002), it increased over Iran (Stevens et al., 2001, Griffiths et al., 2001; Stevens et al., this volume; Wasylikowa et al., this volume). Over northern parts of South Asia, the early and mid-Holocene appears to have been wetter than the late Holocene and the present day (Swain et al., 1983; Enzel et al., 1999; Bush, 2004, Staubwasser, 2006; Prasad and Enzel, this volume). Maximum moisture availability occurred around 6000 yr ago. Continuous records from the Northern Arabian Sea (Fig. 3), show a complex Holocene climate evolution, which is attributed to the combined influence of winter and spring rain as well as summer monsoon rain in this transitional region.

The long-term trends as well as multi-centennial and millennial climate variations in most recent high resolution studies of West Asia have been tentatively explained as the result of changes in the Northern Hemisphere high latitudes by some authors. Variations in the North Atlantic thermohaline circulation (Bond et al., 2001) and the NAO (Lamy et al., 2006) have both been invoked as the driving force of climate change in West Asia. Likewise, the relationship between North Atlantic climate variability and that of the wider eastern Mediterranean region remain elusive for the remaining Holocene (Lamy et al., 2006). As reviewed above, other factors, such as glaciation of the central Asian high altitudes and the onset of modern-type ENSO variability in the mid-Holocene, will have to be considered in future studies. A response of eastern Mediterranean winter climate to the extension of the Siberian High, as proposed byRohling et al. (2002), could be viewed in relation to the teleconnenction between Asian snow cover and the coupled systems of ENSO and the Asian monsoon with the subsequent alternation of westerly flow pattern and temperatures over Eurasia (Meehl, 1994).

An interesting feature of Holocene climate change in the Northern Hemisphere and between the eastern Mediterranean and South Asia in particular is its quasi-cyclic evolution in the multi-centennial band (Bond et al., 2001; Staubwasser et al., 2003; Arz et al., 2003; Wang et al., 2005; Lamy et al., 2006). This have been linked to solar radiation variability, although the mechanism through which the small solar changes propagate into the climate system remains unclear. The apparent solar influence on Holocene climate cycles of West and South Asia, the heating at Asian high altitudes (Cubasch et al., 1997), particularly where glaciated (Bush, 2000), and the associated Rossby wave response, may be worth further examination in numerical simulations.

The 8.2 ka abrupt climate change event

The 8.2 ka abrupt climate change event has recently been reviewed and analyzed by Rohling and Pälike (2005). This review includes several records from the eastern Mediterranean, West Asia and South Asia (Bar-Matthews et al., 1997; Neff et al., 2001; Staubwasser et al., 2002; Rohling et al., 2002; Fleitmann et al., 2003). Additional evidence of the impacts of this event are available for Morocco (Chedadi et al., 1998; Gupta et al., 2003). This volume adds the papers by Frisia et al. for Italy, Migowski et al. for the Levant and Weninger et al. for eastern Europe and Anatolia.

Social adaptations to the abruptness, magnitude and duration of the 8.2 ka climate change appear varied and complex across the regions of West Asia (Table 1). The earliest cultivation of wild cereals began in the Levant at the end of the Pleistocene and was established as field agriculture during the Younger Dryas (13,000–11,500 cal yr BP). Domesticated species of cereals were first cultivated by Pre-Pottery Neolithic A sedentary villagers (11,800–10,500 cal yr BP). By the middle Pre-Pottery Neolithic B period (10,250–9,500 cal yr BP) domesticated cereals included einkorn, emmer and barley, and the first domesticated caprines, sheep and goat, all retrieved at village site excavations across the Levant, southwest Anatolia and southwestern Iran (Harris, 2002). The Final PPNB/PPNC (8,800–8,200 cal yr BP) phase of agricultural village life was marked by the abrupt 8.2 ka event (Weninger et al., this volume). During the Final PPNB/PPNC collapse in the Levant, large villages probably became small hamlets as water necessary for agriculture was reduced. In Mesopotamia, this interval may have witnessed the partial abandonment of northern and central Mesopotamia and the first colonization of southern Mesopotamia by habitat-tracking irrigation agriculture villagers, to judge from their material culture (Weiss, 1978) and two basal period radiocarbon dates from the excavations at Oueilli (Valladas et al., 1996).

Cause of the 8.2 ka climate event

In light of the physical mechanism responsible for the present day correlation between NAO and rainfall in parts of the eastern Mediterranean (see above), enhanced meltwater discharge into the North Atlantic from the collapsing North American ice margin and the associated sea surface temperature drop around 8200 yr ago (Ellison et al., 2006) could directly

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Table 1
Holocene abrupt climate changes and mesopotamian culture

BP	BC	Mesopotamian culture period	
	1900		Synchronus collapse Med. to Indus, and beyond
			HABITAT-TRACKING
			Collapse of Akkadian Empire
4200	2200	3	4.2 kaBP arid/cooling event
	2300		Akkadian Empire
			First palaces in south Mesopotamia
	3000		Early Dynastic Sumerian cities
			Sumerian Isolation
			HABITAT-TRACKING
			Collapse of Late Uruk society, north and south
	3200	2	5.2 kaBP arid/cooling event
	4000		Beginning of Uruk period cities
			Ubaid period villages in south Mesopotamia
7500	5500		Beginning of Ubaid period irrigation agriculture
			First temples in south Mesopotamia
			Samarra period irrigation in south Mesopotamia
			HABITAT-TRACKING
			Collapse of Hassuna villages
8200	6200	1	8.2 kaBP arid/cooling event
	6800		First Hassuna villages in rain-fed north Mesopotamia
11,000–9000	9000–7000		Beginning of agriculture
			HABITAT-TRACKING
			Collapse of hunting and gathering
12,800–11,400	10,800–9400		Younger Dryas arid/cooling event

The sequence of abrupt, arid, cooling event, followed by collapse, and habitat-tracking, marks the Holocene cultural trajectory through the early second millenium BC.

explain reduced winter rainfall inferred at that time for the eastern Mediterranean (Bar-Matthews et al., 1997; Migowski et al., this volume; see also Fig. 3). A similar event is observed in a number of summer monsoon records, but the sequence of events and underlying mechanism is not always clear, partly owing to resolution and chronological problems (Morrill et al., 2003; Rohling and Pälike, 2005). Rohling and Pälike (2005) noted that in many of the Asian records, the 8.2 ka event was part of a longer anomaly that lasted 400 to 600 yr, while in Greenand a sharp event of ~150 yr duration was observed. These authors argued for solar radiation variability as a possible cause. As the following section on the 4.2 ka event will suggest, the origin of abrupt climate events in the Holocene should not be viewed solely in terms of Atlantic thermohaline circulation variability.

A climate event at 6000 cal yr BP?

Although Elmoslimany (1986, 1990) suggests aridification in Iran beginning at ca. 6000 cal yr BP from increases in Chenopod and Artemesia pollen, the coincident increase in oak pollen make this uncertain in her Mirabad and Zeribar cores. Mirabad ostracods suggest that low early Holocene lake levels rose after ca. 4000 cal yr BP (Griffiths et al., 2001), maximum winter cooling at 6 ka BP is recorded for the Adriatic (Sangiorgi et al., 2003), and Soreq Cave (Bar-Matthews et al., 1997) records a gradual precipitation drop from ca. 6200 to 5800 yr ago. Evidence for well-dated abrupt climate changes at 6000 cal yr BP are not evident in this volume's papers (see also Gasse, 2000 and Robinson et al., 2006).

The 5.2 ka abrupt climate change event

The 5.2 ka event is well known in the paleoclimate and archaeological communities from data retrieved at Soreq Cave (Bar-Matthews et al., 1997), Lake Van (Lemcke and Sturm, 1997), the Gulf of Oman (Cullen et al., 2000), the Arabian Sea (Sirocko et al., 1993) and Kilimanjaro (Thompson et al., 2006). Here new data for this event is presented in the papers by Arz et al., Migowski et al., Stevens et al., Wasylikova et al. and Parker et al. Data recently published (Preusser et al., 2005; Magny and Haas, 2004; Parker et al., 2004) reinforce the magnitude, duration and abruptness of the event, as well as its hemispheric and possibly global nature.

The social effects of the event in west Asia and Egypt, visible archaeologically, include the abrupt Late Uruk colony collapse across Greater Mesopotamia, extending from the Zagros to Syria, as precipitation in rainfed cereal agriculture regions was reduced beyond sustainable limits (Weiss, 1986, 2003). The synchronous collapse of Late Uruk society in irrigation agriculture southern Mesopotamia has been identified (Postgate, 1986) but not yet explained.

The 5.2 ka reduction in Anatolian precipitation, the major source of Tigris-Euphrates stream flow (Cullen and deMenocal, 2000), suggests that irrigation agriculture in southern Mesopotamia, in the absence of any agro-technology innovations, was similarly reduced during this period (Nutzel, 2004, p. 122). Severe drought was hypothesized to have been responsible for settlement urbanization in southern Mesopotamia at this time (Nissen, 1988). The social and economic adaptation to reduced Late Uruk irrigation agro-production may also have included the collapse of its ideological and politico-economic frame, the temple.

The popular understanding of third millennium southern Mesopotamian society and economy is "the temple city," derived from archaeological retrieval of an unbroken sequence of temple rebuildings that extended from the earliest Ubaid period settlement of southern Mesopotamia through the Ur III period, summarized by Heinrich (1982), and from the interpretations of Lagash epigraphic data summarized by Falkenstein (1954). The groundbreaking reanalysis of Early Dynastic IIIb records from Lagash (Maekawa, 1973–1974), however, indicated that the House of the Queen, within the palace, owned and administered the third millennium Early Dynastic city's land, labor and cereal production. The third millennium southern Mesopotamian city's temple was only the ideological façade for secular dynastic power.

This situation in southern Mesopotamia obtained still earlier, at the end of the Early Dynastic II and beginning of Early Dynastic IIIa period, when the House of the Queen was the operative office, within the palace, for the entire urban

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administration, controlling all arable land, grain production and cattle breeding. Temples did not own land and were part of the central palace administration (Visicato, 1995, 2000).

Palace control of the southern Mesopotamian urban political economy appears to have emerged immediately after the late Uruk collapse following three thousand years of temple domination. The last temples of the late Uruk IV Eanna precinct were abandoned, replaced by terraces and light post and reed constructions (Eichmann, 2001). At the same time, archaeological excavation has retrieved the first palaces, administrative buildings distinguishable clearly from temples, at Jemdet Nasr, in mudbrick (Moorey, 1976) and then at Early Dynastic I period Uruk, in pisé (Eichmann, 1989). Synchronously, "council" rulership disappears from the proto-Sumerian lexicon and the title "king" is commonly documented (Glassner, 2000, pp. 271–272). During the course, then, of two centuries, at the late Uruk 5.2 ka collapse, the southern Mesopotamian temple lost urban economic and political control to the emergent secular palace. After collapse the palace and its dynasty controlled the city, and the temple and its clergy were reduced to control of the gods to provide the legitimation of kingship. In essence, this was the recrudescence of ancient myth. How land ownership concentration within dynastic lineages was effected during this 5.2 ka abrupt climate change remains a process to be examined within finer chronological comparisons between late Uruk collapse phenomena and the 5.2 ka event and with paleoclimate proxy estimates of ancient Tigris-Euphrates flow (Kumke et al., 2004).

At this point, the database of records showing a climate event around 5200 yr ago is unfortunately not sufficient to allow a meaningful discussion of potential forcing mechanisms. It is not clear if there is any relation between the 5.2 ka event and the well described drying event of the Sahara around 5500 cal yr BP (deMenocal et al., 2000).

The 4.2 ka abrupt climate change event

This globally observed climate event occurred roughly between 4500 and 3500 cal yr BP (Gasse, 2000; Weiss, 2000; Booth et al., 2005). This interval includes some chronological imprecision, usually ±100–200 yr. In the highest resolution records, the event begins at 4200 cal yr BP and lasts about 300 yr. In records from the eastern Mediterranean region and

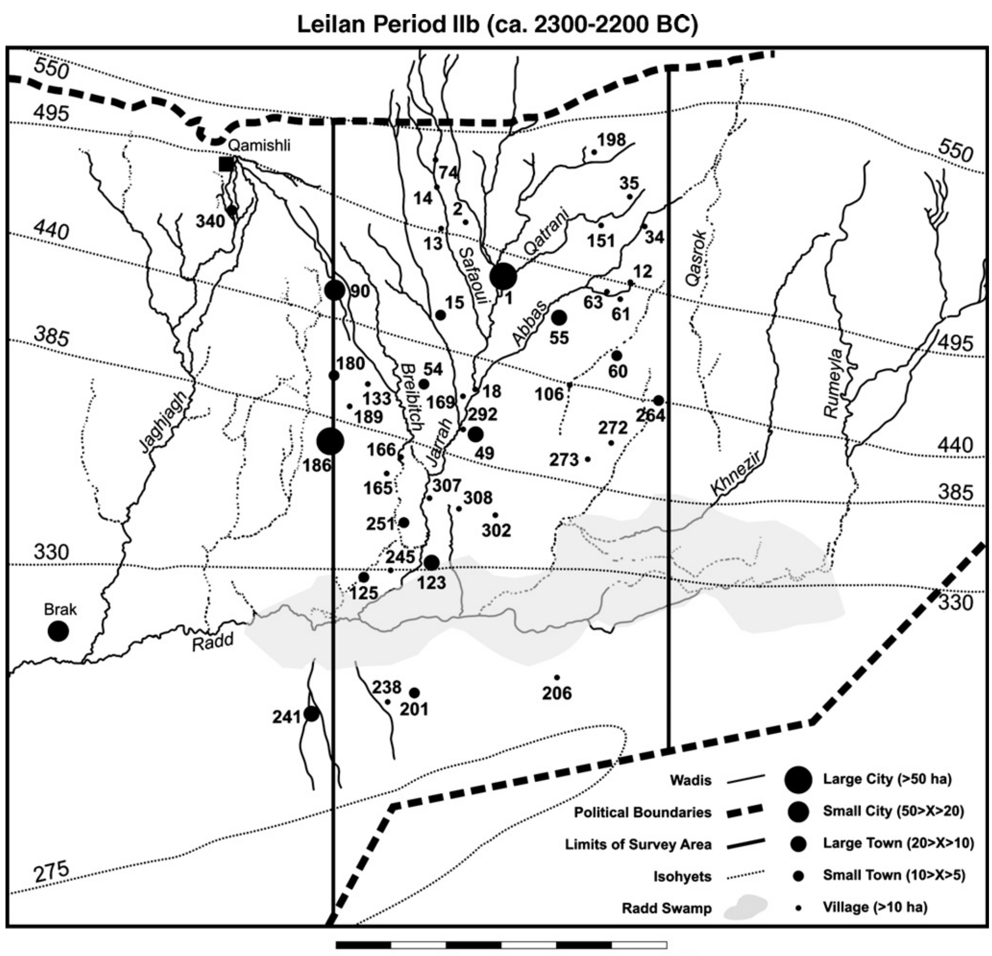

Figure 4. Leilan Region Survey Period IIb (2300–2200 BC). Tell Leilan (N36°57′42″ E41°30′35″), site 1, was the 100 hectare Akkadian imperial production center for the most fertile portion of the Habur Plains, northeast Syria, during this period. Here and in Figures 5 and 6 the precipitation isohyets are estimated from Bar-Matthews et al. (1997) (Ristvet and Weiss, in press).

West Asia, a severe drought is observed almost everywhere. The 4.2 ka event apparently displaced the Mediterranean westerlies and possibly the Indian monsoon (see below), thereby reducing the seasonal precipitation necessary for rain-fed cereal agriculture.

A cultural response to the 4.2 ka climate event may also be seen within the Harappan civilization centered around the Indus valley into the Makran (West Pakistan) and Northwest India. A transition from an urbanized (mature or urban Harappan) to a rural (post-urban) society is well documented beginning at approximately at 3950 cal yr BP (Possehl, 1997a). At the end of the third millennium and the beginning of the second millennium BC, the Great Bath and Granary at Mohenjo-Daro were abandoned, settlement in Sindh, the Indus-Sarasvati valley and the Baluchi highlands collapsed and shifted east to the headwaters of the Sarasvati and south to the Saurashtra Peninsula (Possehl, 1997b; 2000). The climate factor in this relocation has been disputed for lack of clear records. However, such climate records are now available (Staubwasser et al., 2003). Chronological imprecision for the transition from urban Harappan to post-urban remains a problem. Most radiocarbon dates available have an uncertainty of ±100 to 150 yr (Possehl, 1997a); that is, a total uncertainty interval of 300 to 450 yr on the calibrated radiocarbon time scale (Reimer et al., 2004). Some of the more precise dates, e.g. from the city of Mohenjo Daro in the lower Indus valley, or from Rojdi ~700 km South on Saurashtra, India, have a precision of±60 yr, or a calibrated uncertainty interval of 200 yr. This is comparable to the precision of the paleoclimate record off the Indus delta (Staubwasser et al., 2003). The transition from urban to post-urban, as recorded in Mohenjo Daro, began with the abandonment of the Great Bath, which preceeded the abandonment of the whole city by perhaps 200 yr (Possehl, 1997a), apparently just after 4200 cal yr BP. As such, the Harappan cultural transition is within dating uncertainty of the 4.2 ka drought event observed in the marine Indus discharge record (Staubwasser et al., 2003), which began at 4200 cal yr BP and was most severe for the following 200–300 yr (Fig. 3). A return to previous values can be observed at 3600 cal yr BP.

In Palestine, where precipitation and dry farming agro-production are a function of the Mediterranean westerlies, the precipitation reduction has been estimated at 20 to 30% by Bar-Matthews et al. (1997) . Here, cities disappeared and only a few villages and a town near surface water remained (Dever, 1995; Palumbo, 2001). Similar city and town abandonments occurred

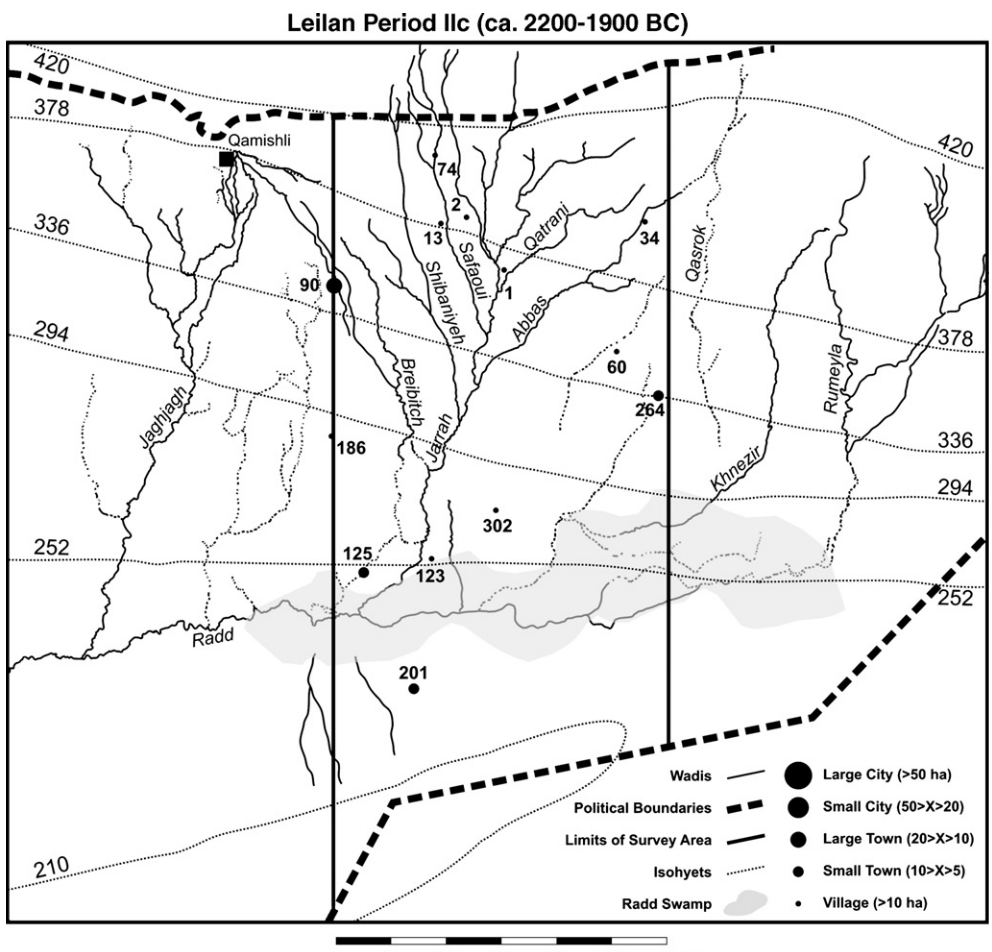

Figure 5. Leilan Region Survey Period IIc (2200–1900 BC). At the Akkadian imperial collapse, during the 4.2 ka abrupt climate change, precipitation in northern Mesopotamia was reduced ca. 20–30% for ca. 300 yr. Settlement here was reduced 73% (Ristvet and Weiss, in press) and occupation at Tell Leilan was reduced to ca. 200 m^2 of reoccupied Akkadian palace rooms. At this time, habitat-tracking to irrigation agriculture south Mesopotamia is documented epigraphically.

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across western Syria, for instance, Ebla (Matthiae, 1995) and Umm al-Marra (Schwartz et al., 2000), and even along the middle Euphrates at Jerablus Tahtani (Peltenburg, 1999), Selenkahiyeh (Van Loon, 2001), Halawa (Orthmann, 1989) and Sweyhat (Zettler, 1997), where reduced settlements, buffered by river access, survived.

At the beginning of the ca. 300 yr term of this event, aridification, dust, and lowered seasonal temperature across the northern Mesopotamian Habur Plains forced massive settlement desertion of the Akkadian imperialized landscape, and left only reduced settlement at strategic loci (Weiss et al., 1993; Cullen et al., 2000). In the Tell Leilan region, across a 30 km wide transect of 1900 km^2, from the Turkish to the Iraqi borders, total population fell dramatically: 74% of sites were abandoned and total area occupied declined by 93% (Figs. 4 and 5; Ristvet and Weiss, in press).

In southern Mesopotamia the social effects of this event are perceptible unevenly as there is no high-resolution archaeological survey data for this period (Weiss, 1975) and because the laconic cuneiform sources provide much information but few details for the Akkadian imperial collapse (Glassner, 1986). Drought, reduced agro-production, political collapse, chaos and foreign invasion befell the capital according to the epic Curse of Akkade (Jacobsen, 1957). Abandonment of Akkade left an archaeological ruin, the target of royal Babylonian excavation hundreds of years later, aside a small town (Weiss, 1997). At Mari, on the middle Euphrates, a post-Akkadian hiatus in the dynastic succession has been defined and regional desertion hypothesized (Durand, 1998), but the complete abandonment of the site during this period has not been observed archaeologically (Margueron, 2004).

Following termination of the 4.2 ka event, according to regional survey data, utilizing relatively fine resolution ceramic chronology, entire regions of northern Mesopotamia, Syria and Palestine were resettled intensively and reorganized fundamentally, as on the Habur Plains (Fig. 6; Ristvet and Weiss, in press). Reoccupied settlements now served different functions within regions resettled on a different scale (Weiss, 1985), with ethnically different populations, sedentarized nomadic pastoralists (Heimpel, 2003; Guichard, 2002; Durand, 1998), and regional economic and spatial organization adapted to the previously abandoned, now open, arable, dry-farming 19th century BC landscapes.

The tephrochronostratigraphic linkage between the Akkadian collapse at imperialized Tell Leilan and the initiation of the 4.2 ka event recorded in the Gulf of Oman (Cullen et al.,

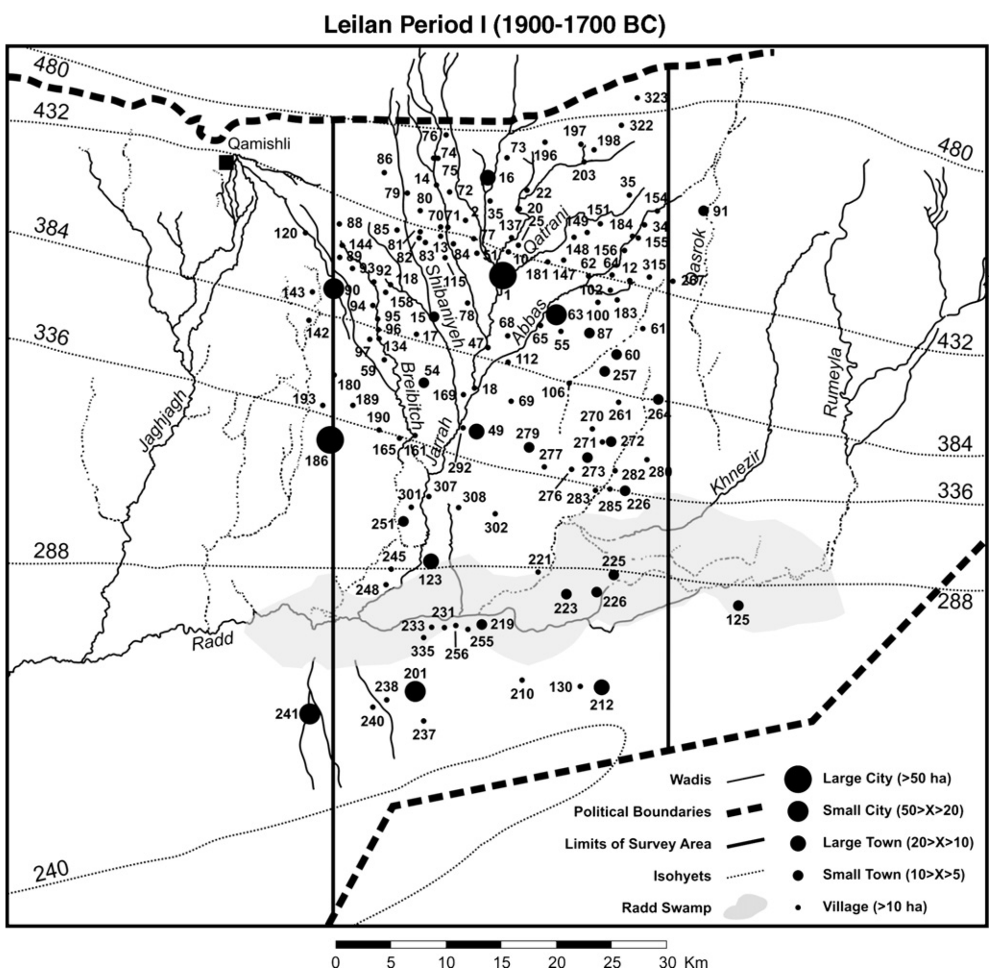

Figure 6. Leilan Region Survey Period I (1900–1700 BC). Favorable conditions for cereal agriculture and settlement returned at ca. 1900 BC. The sedentarization of formerly pastoral nomadic tribes probably accounts for much of the twelve-fold settlement increase during this period (Ristvet and Weiss, in press).

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2000; Weiss et al., 1993; see cover), will now see further radiocarbon chronology refinement. The terminus of this event has already been observed with dendrochronological resolution (Zhang and Hebda, 2005). With abandonment of the instrumental record, precipitation functions can now be deployed to generate realistic rain-fed and irrigation agriculture production functions for the various landscapes of this period and their wide range of political structures. Quantified regional models of the synchronous collapses at 4.2 ka are foreseeable.

Speculative cause and mechanism of the 4.2 ka climate change event in Asia

Climate change at the 4.2 ka event is observed on a global scale, including many places in Asia and Africa, where a severe and lasting drought is indicated (Gasse, 2000; Weiss, 2001; Booth et al., 2005). In contrast, the change in South America is to moister conditions in most cases (Marchant and Hooghiemstra, 2004), although not uniformly. Ice cores from the Andes (Southern Peru) show much enhanced dust concentration and suggest enhanced drought in western South America (Thompson, 2000). Regions of direct interest for this summary, where a clear 4.2 ka drought event is observed, include northern Italy (Drysdale et al., 2006), the Levant (Bar Matthews et al., 1997; Enzel et al., 2003, Migowski et al., this volume), the northern Red Sea (Arz et al., this volume), Arabia, and Mesopotamia (Cullen et al., 2000; Parker et al., 2004; this volume), Caucasus (Alexandrovskiy et al., 2001), Northwest India and Pakistan (Phadtare, 2000; Staubwasser et al., 2003), Tibet (Gasse and van Campo, 1994; Hong et al., 2003, and central and South China (An et al., 2005; 2006; Wang et al., 2005). In eastern Anatolia a clear 4.2 ka drought is recorded in the Lake Van geochemical record (Lemcke and Sturm, 1997). The remainder of Turkey experienced drought as well (Weiss, 2001). Here, the evidence comes largely from pollen records, as at Eski Açigöl where the diminution of oak pollen is interpreted as anthropogenic (Roberts et al., 2001). However, chronological imprecision in these lacustrine records is generally a problem. In a high-resolution stalagmite record from Oman (Fleitmann et al., 2003), the 4.2 ka event is not present. Apart from the Northwest (Phadtare, 2000) the records from the Indian Peninsula unfortunately lack resolution to clearly resolve the 4.2 ka event. Several records from the Thar desert show an event at that time, but see the detailed discussion by Prasad and Enzel (this volume). In western Iran, multiple records exist that indicate the 4.2 ka event, but different proxies conflict regarding the nature of change (Stevens et al., this volume). Several African and central to East Asian records indicate climate change around 4200 cal yr BP, but some apparently of opposite sign. The East African lake record for a 4.2 ka drought event is exceptionally clear (Gasse, 2000; Weldeab et al., 2005; Russell and Johnson, 2005; Krom et al., 2002). In Morocco (Lamb et al., 1995, but see Chedadi et al., 1998) and northeastern China (Hong et al., 2005) as well as in some records from the northwestern margin of the central Asian high altitudes (Ricketts et al., 2001; Wunnemann et al., 2006) conditions at the 4.2 ka event appear to indicate more available moisture. This could be a significant observation and would not be unexpected for a latitude band where rainfall is influenced by atmospheric Rossby waves, but additional records would be helpful for confirmation.

Importantly, records from tropical northern West and East Africa (Gasse and van Campo, 1994) and southern tropical East Africa (Thompson et al., 2002) also record a prominent drought at the 4.2 ka event. In contrast, high latitude records either do not show a significant event (Grootes and Stuiver, 1997), or only record change of a magnitude not different form overall Holocene variability (Bond et al., 2001). Thus, the 4.2 ka event appears to be most prominent between the tropics and the middle latitudes (see also Booth et al., 2005), with implications for the underlying mechanisms involved. Precipitation pattern in the subtropical latitudes in Asia, Africa, and the Mediterranean is significantly affected by the prevailing Rossby wave pattern in upper-level westerly flow (see above). Upper-level flow, in turn, is significantly related to tropical convection intensity. Yang and Webster (1990) show that maximum speed in the northern subtropical jet corresponds to maximum tropical intensity in the Southern Hemisphere tropics. Taken at face value, this may suggest a direct link between southern tropical East Africa and the Middle East. Both regions show a very pronounced 4.2 ka drought event. A change in southern East African tropical convection intensity around 4.2 ka may have altered upper-level flow and wave pattern in the Northern Hemisphere subtropics particularly over the Middle East. This does not reveal the underlying forcing, but it is worth noting that convection intensity in the southern East African tropics responds to ENSO variability as does, in fact, the entire Indian Ocean–West Pacific monsoon system (Meehl, 1994).

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References

Alexandrovskiy, A.L., van der Plicht, J., Belinskiy, A., Khoklova, O., 2001. Chronology of soil evolution and climatic changes in the dry steppe zone of the northern Caucasus, Russia, during the 3rd millennium BC. Radiocarbon 43, 629–635.

Alijani, B., 2002. Variations of 500 hPa flow patterns over Iran and surrounding areas and their relationship with the climate of Iran. Theoretical and Applied Climatology 72, 41–54.

An, C.-B., Tang, L.Y., Barton, L., Chen, F.H., 2005. Climate change and cultural response around 4000 cal yr BP in the western part of Chinese Loess Plateau. Quaternary Research 63, 347–352.

An, C.-B., Feng, Z.-D., Barton, L., 2006. Dry or humid? Mid Holocene humidity changes in arid and semi-arid China. Quaternary Science Reviews 25, 351–361.

Anderson, D.M., Prell, W.L., 1992. The structure of the Southwest monsoon winds over the Arabian Sea during the late quaternary—observations, simulations, and marine geologic evidence. Journal of Geophysical Research-Oceans 97, 15481–15487.

Anderson, D.M., Overpeck, J.T., Gupta, A.K., 2002. Increase in the Asian southwest monsoon during the past four centuries. Science 297, 596–599.

Arz, H.W., Lamy, F., Patzold, J., Muller, P.J., Prins, M., 2003. Mediterranean moisture source for an early-Holocene humid period in the northern Red Sea. Science 300, 118–121.

Arz, H.W., Lamy, F., Pätzold, J., 2006. A pronounced dry event recorded around 4.2 kyr in brine sediments from the Northern Red Sea. Quaternary Research, THIS VOLUME

Bar-Matthews, M., Ayalon, A., Kaufman, A., 1997. Late quaternary paleoclimate in the eastern Mediterranean region from stable isotope analysis of speleothems at Soreq Cave, Israel. Quaternary Research 47, 155–168.

Bar-Matthews, M., Ayalon, A., Gilmour, M., Matthews, A., Hawkesworth, C.J., 2003. Sea–land oxygen isotopic relationships from planktonic foraminifera and speleothems in the Eastern Mediterranean region and their implication for paleorainfall during interglacial intervals. Geochimica Et Cosmochimica Acta 67, 3181–3199.

Bond, G., Kromer, B., Beer, J., Muscheler, R., Evans, M.N., Showers, W., Hoffmann, S., Lotti-Bond, R., Hajdas, I., Bonani, G., 2001. Persistent solar influence on north Atlantic climate during the Holocene. Science 294, 2130–2136.

Booth, R.K., Jackson, S.T., Forman, S.L., Kutzbach, J.E., Bettis, E.A., Kreig, J., Wright, D.K., 2005. A severe centennial-scale drought in mid-continental North America 4200 years ago and apparent global linkages. Holocene 15, 321–328.

Bush, A.B.G., 2000. A positive climatic feedback mechanism for Himalayan glaciation. Quaternary International 65–6, 3–13.

Bush, A.B.G., 2004. Modelling of late Quaternary climate over Asia: a synthesis. Boreas 33, 155–163.

Calvert, S.E., Fontugne, M.R., 2001. On the late Pleistocene–Holocene sapropel record of climatic and oceanographic variability in the eastern Mediterranean. Paleoceanography 16, 78–94.

Chedadi, R., Lamb, H.F., Guiot, J., van der Kaars, 1998. Holocene climate change in Morocco: a quantitative reconstruction from pollen data. Climate Dynamics 14, 883–890.

Coombes, P., Barber, K., 2005. Environmental determinism in Holocene research: causality or coincidence? Area 37, 303–311.

Cubasch, U., Voss, R., Hegerl, G.C., Waszkewitz, J., Crowley, T.J., 1997. Simulation of the influence of solar radiation variations on the global climate with an ocean–atmosphere general circulation model. Climate Dynamics 13, 757–767.

Cullen, H.M., deMenocal, P.B., 2000. North Atlantic influence on Tigris-Euphrates streamflow. International Journal of Climatology 20, 853–863.

Cullen, H.M., deMenocal, P.B., Hemming, S., Hemming, G., Brown, F.H., Guilderson, T., Sirocko, F., 2000. Climate change and the collapse of the Akkadian empire: evidence from the deep sea. Geology 28, 379–382.

Davies, C.P., 2006. Holocene Paleoclimates of Southern Arabia from Lacustrine Deposits of the Dhamar Highlands. Yemen Quaternary Research, THIS VOLUME

deMenocal, P., Ortiz, J., Guilderson, T., Sarnthein, M., 2000. Coherent high- and low-latitude climate variability during the Holocene warm period. Science 288, 2198–2202.

Dever, W., 1995. Social structure in the Early Bronze IV period in Palestine. In: Levy, T. (Ed.), The Archaeology of Society in the Hold Land. Facts on File, New York, pp. 282–296.

Durand, J.M., 1998. Les documents épistolaires du palais de Mari. Littératures anciennes du Proche-Orient, 17. Les Editions du Cerf, Paris, pp. 481–482.

Drysdale, R., Zanchetta, G., Hellstrom, J., Maas, R., Fallick, A., Pickett, M., Cartwright, I., Piccini, L., 2006. Late Holocene drought responsible for the collapse of Old World civilizations is recorded in an Italian cave flowstone. Geology 34, 101–104.

Eichmann, R., 1989. Uruk, die Stratigraphie: Grabungen 1912–1977 in den Bereichen 'Eanna' und 'Anu-Ziqqurrat'. P. von Zabern, Mainz.

Eichmann, R., 2001. Die leichte Bauweise in den archaischen Schichten von Uruk-Warka. In: R. M. Boehmer, J. Maran, (Eds.), Lux Orientis: Archäologie zwischen Asien und Europa, Festschrift für Harald Hauptmann zum 65. Geburtstag, M. Leidorf, Rahden, pp. 121–131.

Ellison, C.R.W., Chapman, M.R., Hall, I.R., 2006. Surface and deep ocean interactions during the cold climate event 8200 yrs ago. Science 312, 1929–1932.

Elmoslimany, A.P., 1986. Ecology and late-quaternary history of the kurdo-zagrosian oak forest near lake zeribar, western Iran. Vegetatio 68, 55–63.

Elmoslimany, A.P., 1990. Ecological significance of common nonarboreal pollen—examples from drylands of the middle-east. Review of Palaeobotany and Palynology 64, 343–350.

Emeis, K.C., Struck, U., Schulz, H.M., Rosenberg, R., Bernasconi, S., Erlenkeuser, H., Sakamoto, T., Martinez-Ruiz, F., 2000. Temperature and salinity variations of Mediterranean Sea surface waters over the last 16,000 yrs from records of planktonic stable oxygen isotopes and alkenone unsaturation ratios. Palaeogeography Palaeoclimatology Palaeoecology 158, 259–280.

Enzel, Y., Ely, L.L., Mishra, S., Ramesh, R., Amit, R., Lazar, B., Rajaguru, S.N., Baker, V.R., Sandler, A., 1999. High-resolution Holocene environmental changes in the Thar Desert, northwestern India. Science 284, 125–128.

Enzel, Y., Bookman, R., Sharon, D., Gvirtzman, H., Dayan, U., Ziv, B., Stein, M., 2003. Late Holocene climates of the Near East deduced from Dead Sea level variations and modem regional winter rainfall. Quaternary Research 60, 263–273.

Eshel, G., 2002. Mediterranean climates. Israel Journal of Earth Science 51, 157–168.

Eshel, G., Farrell, B.F., 2000. Mechanisms of eastern Mediterranean rainfall variability. Journal of the Atmospheric Sciences 57, 3219–3232.

Eshel, G., Farrell, B.F., 2001. Thermodynamics of Eastern Mediterranean rainfall variability. Journal of the Atmospheric Sciences 58, 87–92.

Evans, J.P., Smith, R.B., Oglesby, R.J., 2004. Middle East climate simulation and dominant precipitation processes. International Journal of Climatology 24, 1671–1694.

Falkenstein, A., 1954. La cite-temple sumerienne. Cahiers d'histoire mondiale 1, 784–814.

Felis, T., Patzold, J., Loya, Y., Fine, M., Nawar, A.H., Wefer, G., 2000. A coral oxygen isotope record from the northern Red Sea documenting NAO, ENSO, and North Pacific teleconnections on Middle East climate variability since the year 1750. Paleoceanography 15, 679–694.

Fisher, M., Membery, D.A., 1998. Chapter 2, climate. In: Ghanzafar, A., Fisher, M. (Eds.), Vegetation of the Arabian Peninsula. Kluver, dordrecht, pp. 5–38.

Fleitmann, D., Burns, S.J., Mudelsee, M., Neff, U., Kramers, J., Mangini, A., Matter, A., 2003. Holocene forcing of the Indian monsoon recorded in a stalagmite from Southern Oman. Science 300, 1737–1739.

Flohn, H., 1965. Klimaprobleme am Roten Meer. Erdkunde 19, 179–191.

Frisia, S., Borsato, A., Mangini, A., Spötl, C., Madonia, G., Sauro, U., 2006. Holocene climate variability in Sicily from a discontinuous stalagmite record: implications for Mesolithic to Neolithic transition. Quaternary Research, THIS VOLUME

Gadgil, S., 2003. The Indian monsoon and its variability. Annual Review of Earth and Planetary Sciences 31, 429–467.

Gadgil, S., Joseph, P.V., 2003. On breaks of the Indian monsoon. Proceedings of the Indian Academy of Sciences—Earth and Planetary Sciences 112, 529–558.

Gasse, F., 2000. Hydrological changes in the African tropics since the Last Glacial Maximum. Quaternary Science Reviews 19, 189–211.

Gasse, F., van Campo, E., 1994. Abrupt postglacial climate events in West Asia and North-Africa Monsoon Domains. Earth and Planetary Science Letters 126, 435–456.

Glassner, J.-J., 1986. La chute d'Akkadé: l'evènement et sa mémoire, vol. 5. Beiträge zum Vorderen Orient, Berlin.

Glassner, J.-J., 2000. Ecrire à Sumer: l'invention du cunéiforme. Editions du Seuil, Paris.

Gordon, A., Grace, W., Schwerdtfeger, P., Byron-Scott, R., 1998. Dynamic Meteorology: A Basic Course. Wiley, New York.

Griffiths, H.I., Schwalb, A., Stevens, L.R., 2001. Environmental change in southwestern Iran: the Holocene ostracod fauna of Lake Mirabad. Holocene 11, 757–764.

Grootes, P.M., Stuiver, M., 1997. Oxygen 18/16 variability in Greenland snow and ice with 10^3-to 10^5-year time resolution. Journal of Geophysical Research 102, 26455–26470.

Grotjahn, R., 1993. Atmospheric Circulation: Observation and Theories. Oxford Univ. Press, New York.

Guichard, M., 2002. Le Šubartum occidental a l'avènement de Zimri-Lim. Mémoires de N.A.B.U. 7, 119–168.

Gupta, A.K., Anderson, D.M., Overpeck, J.T., 2003. Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean. Nature 421, 354–357.

Harris, D., 2002. Development of the agro-pastoral economy in the Fertile Crescent during the Pre-Pottery Neolithic period. In: Cappers, R., Bottema, S. (Eds.), The Dawn of Farming in the Near East . Ex Oriente, Berlin, pp. 67–83.Heimpel, W., 2003. Letters to the King of Mari. Eisenbraun's, Winona Lake.

Heinrich, E., 1982. Die Tempel und Heiligtümer im alten Mesopotamien. DeGruyter, Berlin.

Hoelzmann, P., Gasse, F., Dupont, L.M., Salzmann, U., Staubwasser, M., Leuschner, D.C., Sirocko, F., 2004. Paleoenvironmental changes in the arid and subarid belt (Sahara–Sahel–Arabian Peninsula) from 150 kyr to present. In: Battarbee, R.W., et al. (Ed.), Past Climate Variability through Europe and Africa. Springer, Dordrecht.

Hong, Y.T., Hong, B., Lin, Q.H., Zhu, Y.X., Shibata, Y., Hirota, M., Uchida, M., Leng, X.T., Jiang, H.B., Xu, H., Wang, H., Yi, L., 2003. Correlation between Indian Ocean summer monsoon and North Atlantic climate during the Holocene. Earth and Planetary Science Letters 211, 371–380.

Hong, Y.T., Hong, B., Lin, Q.H., Shibata, Y., Hirota, M., Zhu, Y.X., Leng, X.T., Wang, Y., Wang, H., Yi, L., 2005. Inverse phase oscillations between the East Asian and Indian Ocean summer monsoons during the last 12,000 yrs and paleo-El Nino. Earth and Planetary Science Letters 231, 337–346.

Jacobsen, T., 1957. Early political development in Mesopotamia. Zeitschrfit für Assyriologie 52, 91–140.

Joseph, P.V., Sijikumar, S., 2004. Intraseasonal variability of the low-level jet stream of the Asian summer monsoon. Journal of Climate 17, 1449–1458.

Jung, S.J.A., Davies, G.R., Ganssen, G.M., Kroon, D., 2004. Synchronous Holocene sea surface temperature and rainfall variations in the Asian monsoon system. Quaternary Science Reviews 23, 2207–2218.

Kahana, R., Ziv, B., Enzel, Y., Dayan, U., 2002. Synoptic climatology of major floods in the Negev Desert, Israel. International Journal of Climatology 22, 867–882.

Kalnay, E., Kanamitsu, M., Kistler, R., 1996. The NCEP/NCAR 40-year reanalysis project. Bulletin of the American Meteorological Society 77, 437–471.

Karim, A., Veizer, J., 2002. Water balance of the Indus River Basin and moisture source in the Karakoram and western Himalayas: implications from hydrogen and oxygen isotopes in river water. Journal of Geophysical 107.

Krishnamurti, T.N., Surgi, N., 1987. Observational aspects of summer monsoon. In: Chang, C.-P., Krishnamurti, T.N. (Eds.), Monsoon Meteorology. Clarendon Press, Oxford, pp. 3–25.

Krishnan, K., Zhang, C., Sugi, M., 2000. Dynamics of breaks in the Indian summer monsoon. Journal of the Atmospheric Sciences 57, 1354–1372.

Krom, M., Stanley, J., Cliff, R., Woodward, J., 2002. Nile river sediment fluctuations over the past 7000 yr and their key role in sapropel development. Geology 30, 71–74.

Kudrass, H.R., Hofmann, A., Doose, H., Emeis, K., Erlenkeuser, H., 2001. Modulation and amplification of climatic changes in the Northern Hemisphere by the Indian summer monsoon during the past 80 k.y. Geology 29, 63–66.

Kumke, T., Hense, A., Schölzel, C., Andreev, A.A., Brüchmann, C., Gebhardt, C., Helle, G., Kienel, U., Kühl, N., Litt, T., Neumann, F., Schleser, G., 2004. Transfer functions for paleoclimate reconstructions—applications. In: Fischer, H., Floeser, G., Kumke, T., Lohmann, G., Miller, H., Negendank, J.F.W., von Storch, H. (Eds.), The KIHZ Project: Towards a Synthesis of Holocene Proxy Data and Climate Models. Springer, Berlin, pp. 245–262.

Lamb, H.F., Gasse, F., Benkaddour, A., El Hamouti, N., van der Kaars, S., Perkins, W.T., Pearce, N.J., Roberts, C.N., 1995. Relation between century-scale Holocene arid intervals in tropical and temperate zones. Nature 373, 134–137.

Lamy, F., Arz, H.W., Bond, G.C., Bahr, A., Patzold, J., 2006. Multicentennial-scale hydrological changes in the Black Sea and northern Red Sea during the Holocene and the Arctic/North Atlantic oscillation. Paleoceanography 21, A1008 doi:10.1029/2005PA001184.

Lemcke, G., Sturm, M., 1997. δ18O and trace element measurements as proxies for the reconstruction of climate changes at lake Van. In: Dalfes, H.N., Kukla, G., Weiss, H. (Eds.), Third Millenium BC Climate Change and Old World Collapse. NATO ASI Ser. 1, vol. 49. Springer, New York, pp. 653–678.

Lolis, C.J., Bartzokas, A., Katsoulis, B.D., 2002. Spatial and temporal 850 hPA air temperature and sea-surface temperature covariances in the Mediterranean region and their connection to atmospheric circulation. International Journal of Climatology 22, 663–676.

Madella, M., Fuller, D.Q., 2006. Palaeoecology and the Harappan Civilisation of South Asia: a reconsideration. Quaternary Science Reviews 25, 1283–1301.

Maekawa, K., 1973–1974. The development of the É-MÍ in Lagash during Early Dynastic III. Mesopotamia 8–9, 77–144.

Magny, M., Haas, J.N., 2004. A major widespread climatic change around 5300 cal. Yr BP at the time of the Alpine Iceman. Journal of Quaternary Science 19, 423–430.

Marchant, R., Hooghiemstra, H., 2004. Rapid environmental change in African and South American tropics around 4000 yrs before present. Earth-Science Reviews 66, 217–260.

Margueron, J.-Cl., 2004. Mari: métropole de l'Euphrate. Picard, Paris.

Matthiae, P., 1995. Fouilles à Ebla en 1993–1994: le palais de la ville bas-nord. Compte rendus de l'Académie des Inscriptions et Belles Lettres, Avril-Juin 651–681.

Mayewski, P.A., Rohling, E., Stager, E., Karlen, J.C., Maasch, W., Meeker, A., Meyerson, L.D., Gasse, E.A., van Kreveld, F., Holmgren, S., Lee-Thorp, K., Rosqvist, J., Rack, G., Staubwasser, F., Schneider, M., Steig, R.R., 2004. Holocene climate variability. Quaternary Research 62, 243–255.

McClure, H.A., 1976. Radiocarbon chronology of late quaternary lakes in Arabian Desert. Nature 263, 755–756.

Meehl, G.A., 1994. Coupled land–ocean–atmosphere processes and South Asian Monsoon variability. Science 266, 263–267.

Migowski, C., Stein, M., Prasad, S., Negendank, J.F.W., 2006. Dead Sea levels, climate variability and human culture development in the Holocene Near East. Quaternary Research, THIS VOLUME

Moorey, R., 1976. The late pre-historic administrative building at Jamdat Nasr. Iraq 38, 95–106.

Morrill, C., Overpeck, J.T., Cole, J.E., 2003. A synthesis of abrupt changes in the Asian summer monsoon since the last deglaciation. The Holocene 13, 465–476.

Myers, P.G., 2002. Flux-forced simulations of the paleocirculation of the Mediterranean. Paleoceanography 17, 1009 doi:10.1029/2000PA000613.

Neff, U., Burns, S.J., Mangini, A., Mudelsee, M., Fleitmann, D., Matter, A., 2001. Strong coherence between solar variability and the monsoon in Oman between 9 and 6 kyr ago. Nature 411, 290–293.

Nissen, H.-J., 1988. The early history of the ancient Near East. University of Chicago Press, Chicago.

Nutzel, W., 2004. Einführung in die Geo-Archäologie des Vorderen Orients. Reichert, Wiesbaden.

Orthmann, W., 1989. Halawa, 1980–1986. Habelt, Bonn.

Overpeck, J., Anderson, D., Trumbore, S., Prell, W., 1996. The southwest Indian Monsoon over the last 18000 yrs. Climate Dynamics 12, 213–225.

Palumbo, G., 2001. The Early Bronze IV. In: MacDonald, B., Bienkowski, R. (Eds.), The Archaeology of Jordan. Sheffield Academic Press, Sheffield, pp. 233–269.

Parker, A.G., Eckersley, L., Smith, M.M., Goudie, A.S., Stokes, S., White, K., Hodson, M.J., 2004. Holocene vegetation dynamics in the northeastern Rub'al-Khali desert, Arabian Peninsula: a pollen, phytolith and carbon isotope study. Journal of Quaternary Science 19, 665–676.

Parker, A.G., Goudie, A.S., Stokes, S., White, K., Hodson, M.J., Manning, M., Kennet, D., 2006. A record of Holocene Climate Change from lake geochemical analyses in southeastern Arabia. Quternary Research, THIS VOLUME

Peltenburg, E., 1999. Tell Jerablus-Tahtani, 1992–1996, a summary. In: del Olmo Lete, G., Montero Fenellos, J.-L. (Eds.), Archaeology of the Upper Syrian Euphrates. Editorial Ausa, Barcelona, pp. 97–105.

Phadtare, N.R., 2000. Sharp decrease in summer monsoon strength 4000–3500 cal yr BP in the central higher Himalaya of India based on pollen evidence from alpine peat. Quaternary Research 53, 122–129.

Possehl, G., 1997a. Climate and the eclipse of the ancient cities of the Indus. In: Dalfes, H.N., Kukla, G., Weiss, H. (Eds.), Third Millenium BC Climate Change and Old World Collapse. NATO ASI Ser. 1, vol. 49. Springer, New York, pp. 193–244.

Possehl, G.L., 1997b. The transformation of the Indus civilization. Journal of World Prehistory 11, 425–472.

Possehl, G., 2000. The drying up of the Sarasvati. In: Bawden, G., Reycraft, R. (Eds.), Environmental Disaster and the Archaeology of Human Response. University of New Mexico Press, Albuquerque, pp. 63–74.

Postgate, N., 1986. The transition from Uruk to Early Dynastic: continuities and discontinuities in the record of settlement. In: Finkbeiner, U., Rollig, W. (Eds.), Đamdat Nasr: Period or Regional style? Reichert, Wiesbaden, pp. 90–106.

Prasad, S., Enzel, Y., 2006. Holocene paleoclimates of India. Quaternary Research, THIS VOLUME.

Preusser, F., Radies, D., Driehorst, F., Matter, A., 2005. Late quaternary history of the coastal WahibaSands, Sultanate of Oman. Journal of Quaternary Science 20, 395–405.

Ramage, C.S., 1966. Summer atmospheric circulation over Arabian Sea. Journal of the Atmospheric Sciences 23, 144–148.

Reimer, P.J., Baillie, M.G.L., Bard, E., Bayliss, A., Beck, J.W., Bertrand, C.J.H., Blackwell, P.G., Buck, C.E., Burr, G.S., Cutler, K.B., Damon, P.E., Edwards, R.L., Fairbanks, R.G., Friedrich, M., Guilderson, T.P., Hogg, A.G., Hughen, K.A., Kromer, B., McCormac, F.G., Manning, S.W., Ramsey, C.B., Reimer, R.W., Remmele, S., Southon, J.R., Stuiver, M., Talamo, S., Taylor, F.W., van der Plicht, J., Weyhenmeyer, C.E., 2004. IntCal04 Terrestrial radiocarbon age calibration, 26-0 ka BP. Radiocarbon 46, 1029–1058.

Ricketts, R.D., Johnson, T.C., Brown, E.T., Rasmussen, K.A., Romanovski, V.V., 2001. The Holocene limnology of lake Issyk-Kul, Kyrgyzstan: trace elements and stable isotope composition of ostracodes. Palaeogeography, Palaeoclimatology, Palaeoecology 176, 207–227.

Rimbu, N., Lohmann, G., Felis, T., Patzold, J., 2001. Arctic oscillation signature in a Red Sea coral. Geophysical Research Letters 28, 2959–2962.

Rimbu, N., Lohmann, G., Felis, T., Patzold, J., 2003. Shift in ENSO teleconnections recorded by a northern Red Sea coral. Journal of Climate 16, 1414–1422.

Ristvet. L., Weiss, H. in press. The Hâbûr Region in the Late Third and Early Second Millennium B.C., In: Orthmann, W. (Ed.), The History and Archaeology of Syria. Vol. 1. Saarbrucken: Saarbrucken Verlag. http://research.yale.edu/leilan/RistvetWeissHAS_10.pdf.

Roberts, N., Reed, J.M., Leng, M.J., Kuzucuoglu, C., Fontugne, M., Bertaux, J., Woldring, H., Bottema, S., Black, S., Hunt, E., Karabiyikoglu, M., 2001. The tempo of Holocene climatic change in the eastern Mediterranean region: new high-resolution crater-lake sediment data from central Turkey. The Holocene 11, 721–736.

Robinson, S., Black, S., Sellwood, B., Valdes, P., 2006. A review of paleoclimate and paleoenvironments in the Levant and Eastern Mediterranean from 25,000 to 5000 yrs BP: setting the environmental background for the evolution of human civilization. Quaternary Science Reviews 25, 1517–1541.

Rodwell, M.J., Hoskins, B.J., 1996. Monsoons and the dynamics of deserts. Quarterly Journal of the Royal Meteorological Society 122, 1385–1404.

Rodwell, M.J., Rowell, D.P., Folland, C.K., 1999. Oceanic forcing of the wintertime North Atlantic oscillation and European climate. Nature 398, 320–323.

Rohling, E.J., 1999. Environmental control on Mediterranean salinity and delta O-18. Paleoceanography 14, 706–715.

Rohling, E.J., Pälike, H., 2005. Centennial-scale climate cooling with a sudden cold event around 8,200 yrs ago. Nature 434, 975–979.

Rohling, E.J., Mayewski, P.A., Abu-Zied, R.H., Casford, J.S.L., Hayes, A., 2002. Holocene atmosphere–ocean interactions: records from Greenland and the Aegean Sea. Climate Dynamics 18, 587–593.

Rohling, E.J., Abu-Zied, R., Casford, C.S.L., Hayes, A., and Hoogakker, B.A.A, in press. The Mediterranean Sea: Present and Past. In: Woodward, J. Physical Geography of the Mediterranean Basin. Oxford University Press, Oxford.

Rossignol-Strick, M., 1999. The Holocene climatic optimum and pollen records of sapropel 1 in the eastern Mediterranean, 9000–6000 BP. Quaternary Science Reviews 18, 515–530.

Rudolf, B., Hauschild, H., Rueth, W., Schneider, U., 1994. Terrestrial Precipitation Analysis: Operational Method and Required Density of Point Measurements. In: Desbois, M., Desalmond, F. (Eds.), Global Precipitations and Climate Change. NATO ASI Series 1, Vol. 26. Springer-Verlag, pp. 173–186.

Russell, J.M., Johnson, TC., 2005. A high-resolution geochemical record from Lake Edward, Uganda Congo and the timing and causes of tropical African drought during the late Holocene. Quaternary Science Reviews 24, 1375–1389.

Sangiorgi, F., Capotondi, L., Nebout, N., Vigliotti, L., Brinkhaus, H., Giunta, S., Lotter, A., Morigi, A., Negri, A., Reichart, G., 2003. Holocene seasonal sea-surface temperature variations in the southern Adriatic Sea inferred from a multiproxy approach. Journal of Quaternary Science 18, 723–732.

Sarkar, A., Ramesh, R., Somayajulu, B.L.K., Agnihotri, R., Jull, A.J.T., Burr, G.S., 2000. High resolution Holocene monsoon record from the eastern Arabian Sea. Earth and Planetary Science Letters 177, 209–218.

Schwartz, G.., Curvers, H., Geritsen, F.A., MacCormack, J., Miller, N.F., Weber, J.A., 2000. Excavation and survey in the Jabbul plain. American Journal of Archaeology 104, 419–462.

Scrivner, A.E., Vance, D., Rohling, E.J., 2004. New neodymium isotope data quantify Nile involvement in Mediterranean anoxic episodes. Geology 32, 565–568.

Sirocko, F., Sarnthein, M., Erlenkeuser, H., Lange, H., Arnold, M., Duplessy, J.C., 1993. Century-scale events in monsoonal climate over the past 24,000 yrs. Nature 364, 322–324.

Sirocko, F., GarbeSchonberg, D., McIntyre, A., Molfino, B., 1996. Teleconnections between the subtropical monsoons and high-latitude climates during the last deglaciation. Science 272, 526–529.

Spanos, S., Maheras, P., Karacostas, T., Pennas, P., 2003. Objective climatology of 500-hPa cyclones in central and east Mediterranean Region during warm-dry period of the year. Theoretical and Applied Climatology 75, 167–178.

Staubwasser, M., Sirocko, F., Grootes, P.M., Erlenkeuser, H., 2002. South Asian monsoon climate change and radiocarbon in the Arabian Sea during early and middle Holocene. Paleoceanography 17, 1063 doi:10.1029/2000PA000608.

Staubwasser, M., Sirocko, F., Grootes, P., Segl, M., 2003. Climate change at the 4.2 ka BP termination of the Indus valley civilization and Holocene south Asian monsoon variability. Geophysical Research Letters 30, 1425 doi:10.1029/2002GL016822.

Staubwasser, M., 2006. An overview of Holocene South Asian monsoon records—monsoon domains and regional contrasts. Journal of the Geological Society of India 68, 433–446.

Stevens, L.R., Wright, H.E., Ito, E., 2001. Proposed changes in seasonality of climate during the Lateglacial and Holocene at Lake Zeribar, Iran. Holocene 11, 747–755.

Stevens, L.R., Ito, E., Schwalb, A., Wright, H.E., 2006. Timing of atmospheric precipitation in the Zagros Mountains inferred from a multi-proxy record from Lake Mirabad, Iran. Quaternary Research, THIS VOLUME

Swain, A.M., Kutzbach, J.E., Hastenrath, S., 1983. Estimates of Holocene Precipitation for Rajasthan, India, Based on Pollen and Lake-Level Data. Quaternary Research 19, 1–17.

Thamban, M., Rao, V.P., Schneider, R.R., Grootes, P.M., 2001. Glacial to Holocene fluctuations in hydrography and productivity along the southwestern continental margin of India. Palaeogeography Palaeoclimatology Palaeoecology 165, 113–127.

Thompson, L.G., 2000. Ice core evidence for climate change in the tropics: implication for our future. Quaternary Science Reviews 19, 19–35.

Thompson, L.G., Mosley-Thompson, E., Davis, M.E., Henderson, K.A., Brecher, H.H., Zagorodnov, V.S., Mashiotta, T.A., Lin, P.N., Mikhalenko, V.N., Hardy, D.R., Beer, J., 2002. Kilimanjaro ice core records: evidence of Holocene climate change in tropical Africa. Science 298, 589–593.

Thompson, L., Mosley-Thompson, G., Brecher, E., Davis, H., Leon, M., Les, B., Lin, D., Mashiotta, P.-N., Mountain, T., 2006. Abrupt tropical climate change: past and present. Proceedings of the National Academy of Sciences 103, 10536–10543.

Trigo, I.F., Davies, T.D., Bigg, G.R., 1999. Objective climatology of cyclones in the Mediterranean region. Journal of Climate 12, 1685–1696.

Tsvieli, Y., Zangvil, A., 2005. Synoptic climatological analysis of 'wet' and 'dry' Red Sea Troughs over Israel. International Journal of Climatology 25, 1997–2015.

Türkes, M., Erlat, E., 2006. Influences of the North Atlantic oscillation on precipitation variability and changes in Turkey. Il Nuovo Cimento 29C doi:10.1393/ncc/i2005-10228-8.

Valladas, H., Evin, J., Arnold, M., 1996. Datation par la methode du carbone 14 des couches Obeid 0 et 1 de Tell Oueilli (Iraq). In: Huot, J.-J. (Ed.), Oueili: travaux de 1987 et 1989. Editions Recherce sur les Civilizations, Paris, pp. 381–384.

Van Loon, M., 2001. Selenkahiyeh. Nederlands Historisch-Archaeologish Instituut te Istanbul, Leiden.

Visicato, G., 1995. The Bureaucracy of Šuruppak. Ugarit Verlag, Münster.

Visicato, G., 2000. The Power and the Writing: the early scribes of Mesopotamia. CDL, Bethesda.

Wang, Y.J., Cheng, H., Edwards, R.L., He, Y.Q., Kong, X.G., An, Z.S., Wu, J.Y., Kelly, M.J., Dykoski, C.A., Li, X.D., 2005. The Holocene Asian monsoon: links to solar changes and North Atlantic climate. Science 308, 854–857.

Wasylikowa, K., Witkowski, A., Walanus, A., Hutorowicz, A., Alexandrowicz, S.W., Langer, J.J., 2006. Palaeolimnology of Lake Zeribar, Iran, and its climatic implications. Quaternary Research, THIS VOLUME

Webster, P.J., Fasullo, J., 2003. Monsoon: Dynamical Theory. In: Holton, J., Pyle, J., Curry, J. (Eds), Encyclopedia of Atmospheric Sciences, Elsevier, Amsterdam, pp. 1370–1383.

Webster, P.J., Magana, V.O., Palmer, T.N., Shukla, J., Tomas, R.A., Yanai, M., Yasunari, T., 1998. Monsoons: Processes, predictability, and the prospects for prediction. Journal of Geophysical Research—Oceans 103, 14451–14510.

Weiss, H., 1975. Kish, Akkad and Agade. Journal of the American Oriental Society 95, 434–453.

Weiss, H., 1978. Irrigation Agriculture. Science 200, 1377–1378.

Weiss, H., 1985. Tell Leilân and Shubat-Enlil. Mari (Paris) 4, 269–292.

Weiss, H., 1986. The Origins of Tell Leilan and the Conquest of Space in Third Millennium Mesopotamia. In: Weiss, H. (Ed.), The Origins of cities in dry farming Syria and Mesopotamia in Third Millenium B.C. Four Quarters, Guilford, pp. 71–108.

Weiss, H., 1997. Akkade. In: Meyers, M. (Ed.), The Oxford Encyclopedia of Archaeology in the Near East, 1, pp. 41–44. New York.

Weiss, H., 2000. Beyond the Younger Dryas: Collapse as Adaptation to Abrupt Climate Change in Ancient West Asia and the Eastern Mediterranean. In: Bawden, G., Reycraft, R. (Eds.), Confronting Natural Disaster: Engaging the Past to Understand the Future. University of New Mexico Press, Albuquerque, 2000, pp. 75–98.

Weiss, H., 2003. Ninevite Periods and Processes. In: Rova, E., Weiss, H. (Eds.), The Origins of North Mesopotamian Civilization. Subartu IX, Brepols, Turnhout, pp. 593–624.

Weiss, H., Bradley, R.S., 2001. Archaeology—What drives societal collapse? Science 291, 609–610.

Weiss, H., Courty, M.A., Wetterstrom, W., Guichard, F., Senior, L., Meadow, R., Curnow, A., 1993. The genesis and collapse of 3rd millennium north mesopotamian civilization. Science 261, 995–1004.

Weldeab, S., Schneider, R., Kölling, M., Wefer, G., 2005. Holocene African droughts relate to eastern equatorial African cooling. Geology 33, 981–984.

Weninger, B., Alram-Stern, E., Bauer, E., Clare, L., Danzeglocke, U., Jöris, O., Kubatzki, C., Rollefson, G., Todorova, H., van Andel, T., 2006. Climate Forcing due to the 8200 cal BP event observed at Early Neolithic sites in the Eastern Mediterranean. Quaternary Research, THIS VOLUME

Wunnemann, B., Mischke, S., Chen, F.H., 2006. A Holocene sedimentary record from Bosten Lake, China. Palaeogeography Palaeoclimatology Palaeoecology 234, 223–238.

Yang, S., Webster, P.J., 1990. The effect of summer tropical heating on the location and intensity of the extratropical westerly jet streams. Journal of Geophysical Research—Atmospheres 95, 18705–18721.

Zangvil, A., Karas, S., Sasson, A., 2003. Connection between eastern Mediterranean seasonal mean 500 hPa height and sea-level pressure patterns and the spatial rainfall distribution over Israel. International Journal of Climatology 23, 1567–1576.

Zettler, R. (Ed.), 1997. Subsistence and Settlement in a Marginal Environment: Tell es-Sweyhat, 1989–1995 Preliminary Report. MASCA, Philadelphia.

Zhang, Q.-B., Hebda, R., 2005. Abrupt climate change and variability in the past four millennia of the southern Vancouver Island, Canada. Geophysical Research Letters 32, L16708.

Ziv, B., Saaroni, H., Alpert, P., 2004a. The factors governing the summer regime of the eastern Mediterranean. International Journal of Climatology 24, 1859–1871.

Ziv, B., Dayan, U., Sharon, D., 2004b. A mid-winter, tropical extreme flood-producing storm in southern Israel: Synoptic scale analysis. Meteorology and Atmospheric Physics 88, 53–63.

Ziv, B., Dayan, U., Kushnir, Y., Roth, C., Enzel, Y., 2006. Regional and global atmospheric patterns governing rainfall in the southern Levant. International Journal of Climatology 26, 55–73.

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