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Palaeolimnological evidence for an east–west climate see-saw in the Mediterranean since AD 900

Neil Roberts [a,*], Ana Moreno [b], Blas L. Valero-Garcés [b], Juan Pablo Corella [b], Matthew Jones [c], Samantha Allcock [a], Jessie Woodbridge [a], Mario Morellón [d], Juerg Luterbacher [e], Elena Xoplaki [e], Murat Türkeş [f]

[a] School of Geography, Earth and Environmental Sciences, Plymouth University, PL4 8AA, UK
[b] Instituto Pirenaico de Ecología (CSIC), Zaragoza, Spain
[c] School of Geography, University of Nottingham, UK
[d] Department of Surface Waters—Eawag, Dübendorf, Switzerland
[e] Department of Geography, Climatology, Climate Dynamics and Climate Change, Justus Liebig University, Giessen, Germany
[f] Çanakkale Onsekiz Mart University, Department of Geography, Çanakkale, Turkey

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ABSTRACT

During the period of instrumental records, the North Atlantic Oscillation (NAO) has strongly influenced inter-annual precipitation variations in the western Mediterranean, while some eastern parts of the basin have shown an anti-phase relationship in precipitation and atmospheric pressure. Here we explore how the NAO and other atmospheric circulation modes operated over the longer timescales of the Medieval Climate Anomaly (MCA) and Little Ice Age (LIA). High-resolution palaeolimnological evidence from opposite ends of the Mediterranean basin, supplemented by other palaeoclimate data, is used to track shifts in regional hydro-climatic conditions. Multiple geochemical, sedimentological, isotopic and palaeoecological proxies from Estanya and Montcortés lakes in northeast Spain and Nar lake in central Turkey have been cross-correlated at decadal time intervals since AD 900. These dryland lakes capture sensitively changes in precipitation/evaporation (P/E) balance by adjustments in water level and salinity, and are especially valuable for reconstructing variability over decadal–centennial timescales. Iberian lakes show lower water levels and higher salinities during the 11th to 13th centuries synchronous with the MCA and generally more humid conditions during the 'LIA' (15th–19th centuries). This pattern is also clearly evident in tree-ring records from Morocco and from marine cores in the western Mediterranean Sea. In the eastern Mediterranean, palaeoclimatic records from Turkey, Greece and the Levant show generally drier hydro-climatic conditions during the LIA and a wetter phase during the MCA. This implies that a bipolar climate see-saw has operated in the Mediterranean for the last 1100 years. However, while western Mediterranean aridity appears consistent with persistent positive NAO state during the MCA, the pattern is less clear in the eastern Mediterranean. Here the strongest evidence for higher winter season precipitation during the MCA comes from central Turkey in the northeastern sector of the Mediterranean basin. This in turn implies that the LIA/MCA hydro-climatic pattern in the Mediterranean was determined by a combination of different climate modes along with major physical geographical controls, and not by NAO forcing alone, or that the character of the NAO and its teleconnections have been non-stationary.

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1. Introduction

A recent issue of debate in the reconstruction and understanding of climate dynamics during the Medieval Climate Anomaly (MCA, Stine, 1994) is the spatio-temporal character of temperature anomalies and their synchroneity around the globe (Bradley et al., 2003; Mann et al., 2009; Diaz et al., 2011 and references therein, Xoplaki et al., 2011). The issue of spatial coherency applies even more acutely to reconstructed changes in precipitation and water balance during the last millennium, with some areas becoming wetter at the same time that others experienced drought (Graham et al., 2007, 2011; Seager et al., 2007; Diaz et al., 2011). For example, it is clear from tree-ring studies that drought episodes during medieval times in North America had a geographical as well as a temporal expression (Cook et al., 2010). Well-distributed proxy climate data have the potential to reconstruct spatial patterns of regional water balance during the MCA and Little Ice Age (LIA). When combined with numerical climate modelling experiments and analysis of atmospheric

dynamics, it may be possible to link patterns of precipitation to different forcing mechanisms, including solar variations (Steinhilber et al., 2009; Gray et al., 2010; Hegerl et al., 2011) and volcanic eruptions (Mann et al., 2005; Gao et al., 2008; Hegerl et al., 2011), and to known atmospheric and oceanic circulation modes, such as the Arctic Oscillation (Zhao et al., 2001; Feng et al., 2008; Mann et al., 2009; Graham et al., 2011 and references therein). Changes in the frequency or persistency of climate modes such as ENSO (El Niño-Southern Oscillation) may partly account for thermal features during the MCA (e.g. Mann et al., 2005, 2009; Graham et al., 2007, 2011; Seager et al., 2007; Yan et al., 2011).

The Mediterranean basin is influenced by some of the most important mechanisms acting upon the global climate system (Xoplaki, 2002). It marks a transitional zone between the North African–Arabian arid zone dominated by subtropical high pressure and central–northern Europe affected by westerly circulation. In addition, the Mediterranean climate is exposed to the South Asian Monsoon in summer and the western Russian/Siberian High Pressure System in winter (e.g. Corte-Real et al., 1995; Ribera et al., 2000; Xoplaki, 2002; Lionello et al., 2006 and references therein). As well as the influence of atmospheric circulation, climatic conditions over the Mediterranean are affected by physico-geographical factors such as orography, land–sea interactions and the Mediterranean Sea itself (e.g. Lolis et al., 1999; Xoplaki et al., 2000).The prominence of semi-arid conditions at the present day makes this area very sensitive to climatic variations (Lionello et al., 2006, and references therein; Diffenbaugh et al., 2007; Giorgi and Lionello, 2008; Kuglitsch et al., 2010; Toreti et al., 2010). The overall hydrological deficit has required intensive water management during its long history of human occupation, highlighting the central role of hydrological resources in the Mediterranean region. In consequence, reconstructing the timing, intensity, and spatial patterns of hydrological variability in the Mediterranean during the last millennium is crucial to understanding the climate forcing mechanisms behind these changes. The region offers a broad spectrum of documentary information and natural archives, both in time and space, which in turn allow climate reconstructions for past centuries with high temporal and spatial resolution, as well as the analysis of climatic extreme events and socio-economic impacts prior to the instrumental period (e.g. Luterbacher et al., 2006, in press and references therein).

During the period of instrumental records, there is clear evidence of a spatial signature to climate, with annual–decadal variability in winter precipitation in the western Mediterranean being strongly influenced by the North Atlantic Oscillation (NAO; Dünkeloh and Jacobeit, 2003; Trigo et al., 2004; Xoplaki et al., 2004). The strongest positive correlation with the NAO index is located over Iberia (Fig. 1) linked to anomalous high pressure dominance over Northern Europe and below normal sea level pressure over the Azores region (negative NAO conditions). With this surface NAO pattern, the 500-hPa geopotential level is anomalously high (low) in the area of the Icelandic Low and anomalously low (high) across the regions of the subtropical anticyclone and Europe in general, which forces North Atlantic low pressure systems to follow a more northern route and is associated with drier winters in Spain. It has been suggested that the NAO also affects eastern parts of the Mediterranean basin, with the strongest positive (negative) cold-season precipitation anomalies in western Anatolia and parts of the Balkans related to a negative (positive) phase of the NAO (Cullen and deMenocal, 2000; Türkeş and Erlat, 2003, 2005, 2006). In contrast, parts of the southeastern Mediterranean have shown an anti-phase relationship in precipitation and atmospheric pressure with the western Mediterranean (Cullen and deMenocal, 2000; Oldfield and Thompson, 2004; Xoplaki, 2002; Xoplaki et al., 2004; see Fig. 1). In particular, an upper air trough extending from western Europe to the eastern Mediterranean in combination with the connected strong Cyprus Low leads to increased rainfall in the coastal areas of the southern Levant (Ziv et al., 2006; Saaroni et al., 2010). Inter-annual precipitation trends in these southeastern areas have therefore had an inverse correlation with those in the western Mediterranean during the past ~100 years. The associated see-saw pattern in atmospheric pressure has been labelled the Mediterranean Oscillation (MO; Conte et al., 1989).

It has been proposed that the western Mediterranean experienced more frequent negative NAO index states during parts of the LIA (e.g. Luterbacher et al., 2006) and a persistent positive NAO state during the MCA (Trouet et al., 2009). This therefore raises the question of whether these quasi-persistent NAO states were associated with the same kind of see-saw relationship between western and eastern Mediterranean that has been observed for the last ~100 years. In this paper, we use high-resolution palaeolimnological evidence for shifts in regional water balance, supplemented by other proxy climate
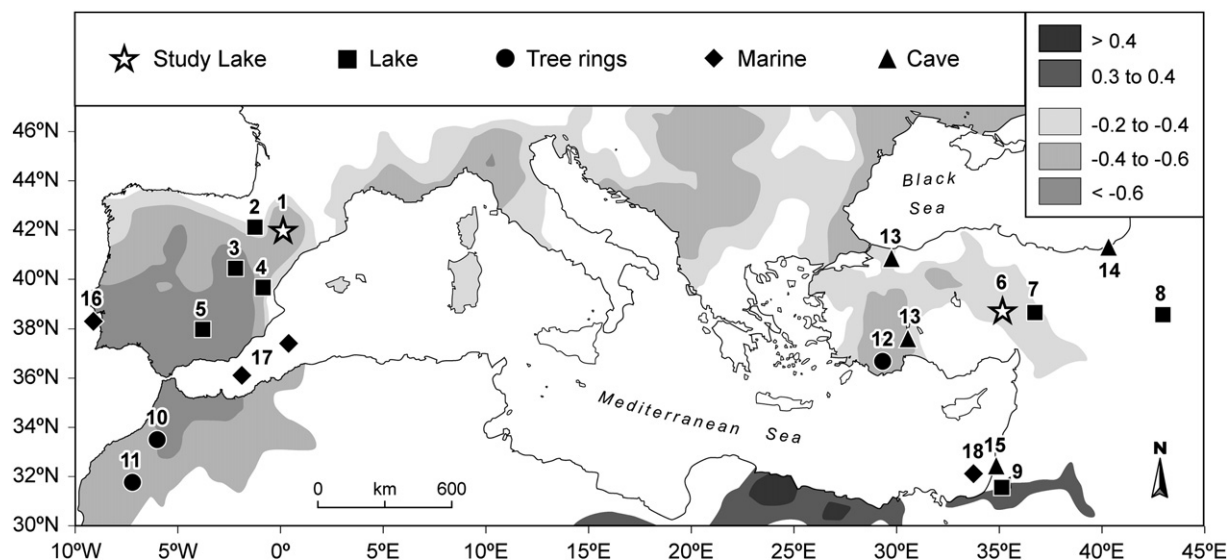

Fig. 1. Location of study sites in relation to the Mediterranean Oscillation shown by spatial Spearman correlation between the NAO atmospheric teleconnection pattern (defined by Barnston and Livezey, 1987) and winter precipitation (1951–2006). The shaded areas denote significant at the 95% level. 1) Montcortès, Estanya and Basa de la Mora, 2) Arreo, 3) Taravilla, 4) La Cruz, 5) Zoñar, 6) Nar, 7) Tecer, 8) Van, 9) Dead Sea, 10) Middle Atlas, 11) High Atlas, 12) SW Taurus, 13) W Turkey, 14) NE Turkey, 15) Soreq, 16) Tagus estuary, 17) Alboran and Balearic basins, 18) Israel coast.

and instrumental data sets, to explore whether the NAO and MO operated over the multi-centennial timescales of the LIA and the MCA. We use statistical analysis to examine correlations between lake records located at the western and easternmost extremes of the Mediterranean basin in order to test the spatial coherence of hydrological changes during the last millennium.

2. Lakes as archives of Late Holocene water balance changes

Dryland lakes can capture sensitively changes in Precipitation/ Evaporation (P/E) balance by adjustments in water level and salinity, and are especially valuable for reconstructing hydro-climatic variability over decadal–centennial timescales (Fritz, 2008). Most lakes in the climatically wetter parts of the Mediterranean Basin have a positive water balance such that their waters remain fresh. By contrast, those in drier Mediterranean regions lose water through evaporation from the lake surface and may be hydrologically closed and contain saline waters. Water levels change in these closed lakes in response to climatic variations, with lake level and/or salinity variations reflecting past changes in effective moisture. Lakes are widely distributed across the Mediterranean Basin, allowing past spatial patterns of climate to be reconstructed (e.g. Roberts et al., 2008). On the other hand, in comparison with some other archives such as tree rings, most lake records are less precisely dated in terms of absolute age and are rarely able to capture inter-annual variability, due to sediment mixing on the lake bed. Lake water balance changes generally integrate winter-season precipitation and summer-season evaporation and evapotranspiration into a single signal, rather than representing climatic conditions during specific seasons. They are also potentially subject to the confounding impact of human disturbance which can overlay any climatic signal. There is widespread evidence that catchment land-use changes have increased the influx of mineral detritus, organic matter and nutrients into many lakes during recent centuries (Roberts and Reed, 2009). Deforestation may have also altered the hydrology of some catchments, notably by increasing runoff rates, while in other cases lake levels have fallen as a consequence of water extraction for irrigation. For these reasons, we focus here on selected Mediterranean lakes with highly-resolved, multi-proxy late Holocene records from well dated core sequences in which human and climatic signals have been differentiated, supplementing them by more ubiquitous lower-resolution sequences.

Past changes in lake water balance are recorded by a number of sedimentary and geomorphological indicators. Water level fluctuations higher than present day can be reconstructed via dated lake marginal deposits (e.g. Bookman et al., 2004), or by changes in the planktonic–benthic ratio of diatoms and other biological indicators (e.g. Barker et al., 1994). Shifts in species assemblage composition may also reflect variations in lake water salinity. In the case of diatoms, past changes in lake salinity have been quantified via transfer functions which model statistically the relationship between modern water chemistry and species assemblages (e.g. Reed, 1998). Oxygen isotope analyses of precipitated carbonate crystals, mollusc and ostracod shells, and diatom frustules provide another important way to reconstruct lake water balance over a hierarchy of timescales (Leng and Marshall, 2004). For a given set of geographical and climatic boundary conditions freshwater lakes with a short residence time have an oxygen isotopic composition similar to that of incoming precipitation, while those with a longer residence time and large evaporative losses have increasingly positive $\delta^{18}O$ values. Even "freshwater" systems in the Mediterranean can be isotopically sensitive to regional water balance changes, so long as significant hydrological losses occur through evaporation (Roberts et al., 2008).

When comparing different lake records, it is necessary to take account of three main sets of controls, namely,

(1) Climate forcing, including inter-annual variations and seasonality of precipitation, temperature, etc., along with the response time of each lake proxy (e.g. linked to lake water residence time)
(2) Non-climatic factors, including human impact and other lake- and parameter-specific controls, e.g. catchment size and type
(3) Chronological precision and accuracy.

Although there are no rigorous ways to partition each lake proxy signal between these different controls, one simple way to do this is via statistical correlations of proxy data across a hierarchy of spatial scales, using results from

(1) The same sediment core, for which chronological correlation can be assured. Close correlation between two proxies from the same core would imply that they have been subjected to a common forcing. If they are not correlated, then this would be consistent with separate forcings; for instance, lake water temperature (inferred from chironomids) vs nutrient loading (inferred from diatoms)
(2) The same lake, for which common forcing can be broadly assured for individual parameters. The same proxy from different cores in the same lake would normally be expected to show a close correlation. If they do not, then the most likely explanation lies with imprecision/inaccuracy in core dating and correlation, or significant spatial variability within the lake
(3) Different lakes in the same region, for which common climate forcing is assured except for localised weather events. Even if subject to common climate forcing, different lake records in the same region may diverge due to (1) non-climatic differences between lakes and (2) errors in chronological correlation
(d) Lakes in different climatic regions, which will be subjected to all three sets of factors and between which, correlations are least likely to be strong. Where correlations are demonstrated, it may be inferred that the lakes have been subjected to common forcing—most likely climatic—and that their chronologies are robust.

We have employed this approach in our research design to include multiple proxies from each of our three principal study lakes, replicated lake and core records within one of our two study regions (northeast Spain), and lakes from geographically distant locations that have experienced contrasting climate histories during the period of instrumental records.

3. Palaeoclimate data from three Mediterranean lakes

In the Iberian Peninsula, there are now many lake-based reconstructions using geochemical and biological proxies with decadal or better resolution spanning the MCA, LIA and modern periods, summarised in Moreno et al. (2011, in review). As well as lying near the centre of action of the 'Mediterranean Oscillation', Northern Spain also possesses highly-resolved sequences which permit replication of records within and between lake basins. Here we focus on the records from two nearby lakes in the Pre-Pyrenean range, at Montcortès (Corella et al., 2011) and Estanya (Morellón et al., 2009; Morellón et al., 2011a).

In karstic Montcortès Lake (Corella et al., 2011) (Table 1) increased carbonate production (high Total Inorganic Carbon, TIC) and lower clastic input (low magnetic susceptibility) occurred during wetter climatic conditions that characterised the main part of the LIA (1330–1840 AD), while higher clastic input occurred during the more arid MCA (1000–1330 AD) (Fig. 2). The presence of a low Mediterranean scrub community—today restricted to elevation below 800 m—suggests warmer temperatures during the MCA (Rull et al., 2011). However, higher erosional input after 1840 AD was almost certainly triggered by increased human occupation (Corella et al.,

Table 1
Modern limnological characteristics of study lakes.

	Montcortès, Spain	Estanya, Spain	Nar, Turkey
Location coordinates	42°19′ N	42°02′ N	38°20′ N
	0° 59′ E	01° 32′ E	34°27′ E
Altitude (masl)	1027	670	1370
Z_{max} (m)	27.7 (2011)	20 (2008)	22.5 (2009)
Lake area, $m^2 \times 10^5$	1.7	1.9	5.6
Electrical conductivity mS cm^{-1}	0.372	3.4	3.37 (2009)
Lake type	Meromictic	Monomictic	Monomictic

2011; Rull et al., 2011). Strong human impact therefore obscures the climate signal during the last part of the LIA and 20th century, and this part of the Montcortès record is excluded from further consideration here. The proxies selected from this core sequence are dated by AMS ^{14}C and varve chronology, and have a mean sampling of <2 yr

for magnetic susceptibility and 6.5 yr for TIC/TOC (Total Inorganic/Organic Carbon).

At nearby brackish karstic Estanya Lake (Fig. 2), shallower water levels and saline conditions predominated during medieval times (870–1300 AD), and generally higher water levels and more diluted waters from 1300 to 1900 AD, although this period shows a complex pattern of wet and arid intervals (Morellón et al., 2011a). Maximum lake levels occurred during the 19th century, and declined during the 20th century. A suite of geochemical elements derived from AVAATECH XRF II core scanning has been selected here as a climate-salinity proxy, notably Ca and S reflecting carbonate and gypsum concentration. These data derive from two separate cores in the same lake dated by AMS ^{14}C, ^{210}Pb and [137]Cs. One of them (core 2) covers the last 860 years with a mean sampling interval of 4 yr (Morellón et al., 2011b), while core 1 covers a much longer timespan but not the last seven centuries (Morellón et al., 2009). There is consequently core overlap and replication for part of the medieval period around AD 1140–1330 (Fig. 2). The northern Spanish lake records from
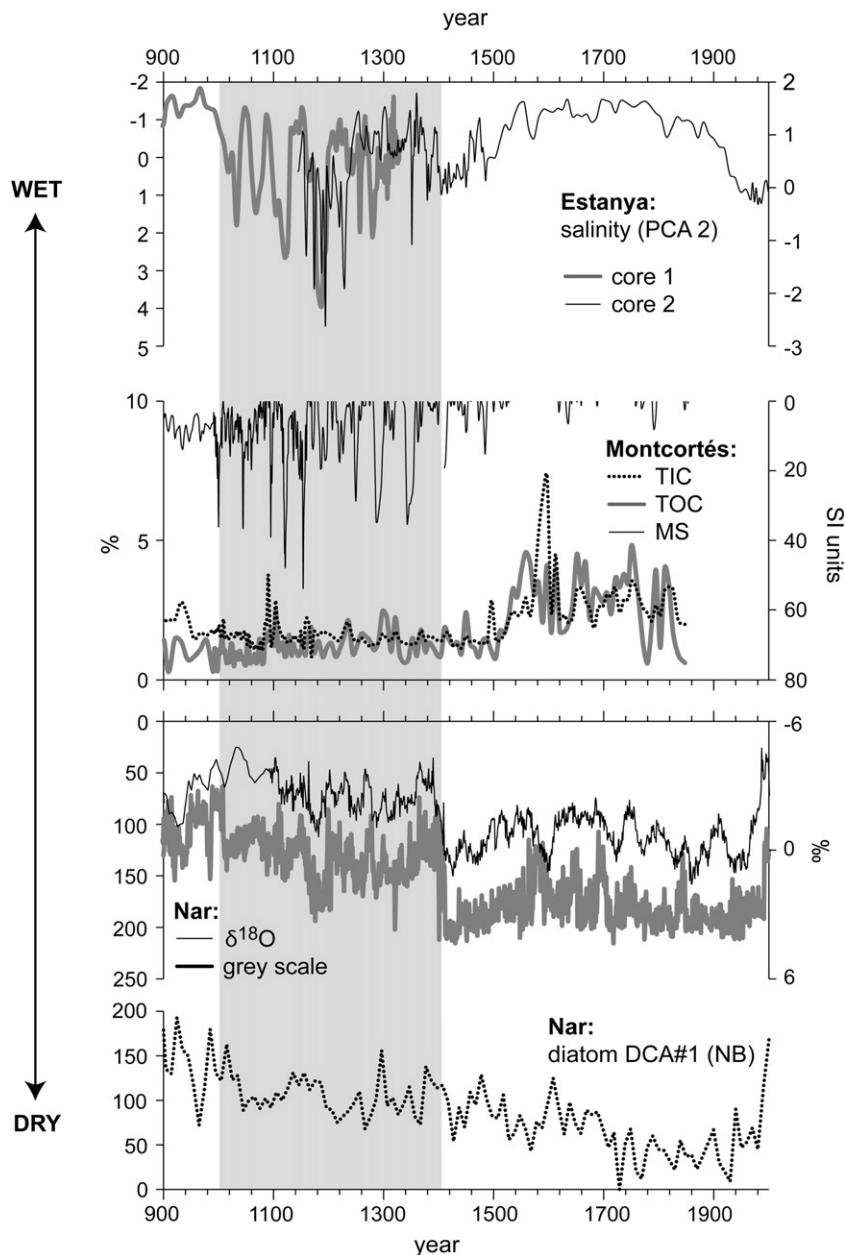

Fig. 2. Selected proxy-climate data for Montcortès, Estanya and Nar lakes, AD900–2000. The Medieval period 1000–1400 AD is shaded.

Table 2
Climate proxy data and decadal Pearson's cross-correlation coefficients for three Mediterranean lake records.

Lake		Nar, Turkey			Montcortès, Spain			Estanya, Spain	
Proxy/core		$\delta^{18}O$ per mil	grey scale	diatom DCA axis1 (nb)	magnetic susceptibility SI units	total organic carbon %	total inorganic carbon %	Itrax XRF PCA 2 Core 1	Itrax XRF PCA 2 Core 2
Original sample interval		1-5 yr [a]	1 yr	10 yr	0.7-3 yr	3-12 yr [b]	3-12 yr [b]	<1-11 yr	1.5-11 yr [b]
Nar	$\delta^{18}O$	-	0.85	-0.56	-0.46	0.38	0.42	-0.18	-0.23
	grey scale		-	-0.64	-0.47	0.45	0.26	-0.55	-0.22
	diatom DCA 1(nb)			-	0.47	-0.42	-0.28	0.19	0.32
Mont-cortès	Mag sus				-	-0.50	-0.48	-0.03	0.44
	TOC					-	0.63	-0.08	-0.52
	TIC						-	0.33	-0.47
Estanya	PCA2 Core1							-	-0.31
Time period		900-2000 AD			900-1850 AD			900-1330	Post-1140[c]

[a] 1 yr interval post-AD1100, 5 yr interval pre-AD1100.
[b] Except 2 intervals of 19 yr.
[c] 1140–2000 (Nar), 1140–1850 (Montcortès), 1140–1330 (Estanya core 1).
Values significant at $p = <0.05$ level are shaded.

Montcortès and Estanya show good overall agreement for the last 1100 years, giving confidence that they reflect common climatic forcing.

At the eastern end of the Mediterranean, Nar crater lake in central Turkey (Table 1) has one of the best-resolved late Holocene climate records. Its continuously-varved sediments provide a well dated proxy-climate sequence for the last 1720 years, with annual to decadal sample resolution. $\delta^{18}O$ measurements on authigenic carbonates show positive values, indicating drier climatic conditions, from 1400 to 1960 AD, with more negative isotopic values, and a wetter climate, between AD 1000–1400 and after 1960 (Jones et al., 2006). These interpretations of the stable isotope variations are confirmed by shifts in carbonate mineralogy between aragonite, indicative of more evaporated lake conditions during the LIA and early 20th century, and calcite, indicative of less evaporated conditions during the other phases. The calcite–aragonite shifts are reflected in grey-scale analysis of the core sediments (Jones, 2004). Diatom analysis of the Nar cores shows significant assemblage changes that are broadly synchronous with the $\delta^{18}O$ record, although there is no evidence for strongly elevated lake salinities during the last 1100 years, partly because of a threshold salinity response in this lake (see Woodbridge and Roberts, 2011). Here we use DCA axis1 of non-blooming diatoms as a palaeohydrological index (see Woodbridge and Roberts, 2010 for further explanation).

The proxy-climate record from Nar lake shows a pattern of change that is almost the mirror image of those from Iberia. In combination, these three high-resolution lake sequences indicate that northern Spain was dry during the MCA and wet during most of the LIA, whereas central Anatolia showed a reverse pattern. This is consistent with contrasting hydro-climatic histories for the east and west Mediterranean over the past millennium.

To test these relationships more fully, we have cross-correlated eight climate proxies from the three study lakes (Table 2). Because the raw data have non-standard time intervals, they have been processed to create decadal averages by being placed in ten-year time "bins". It should be noted that this processing can introduce some additional errors; for example, the Nar diatom data derive from 3 adjacent varves taken every 10 years, whereas other proxy data represent an average of measurements spanning the whole decade. The time intervals used for correlation also vary, with Nar spanning the whole of the last 1110 years, Montcortès covering the period from 900 to 1850 AD, and the two Estanya cores covering 900–1330 AD and 1140–2000 AD respectively. In addition, we recognise that there may be a risk of temporal auto-correlation, which can be a particular problem with long records where N (number of samples) is high, because the r value can be low and still show a significant p-value.
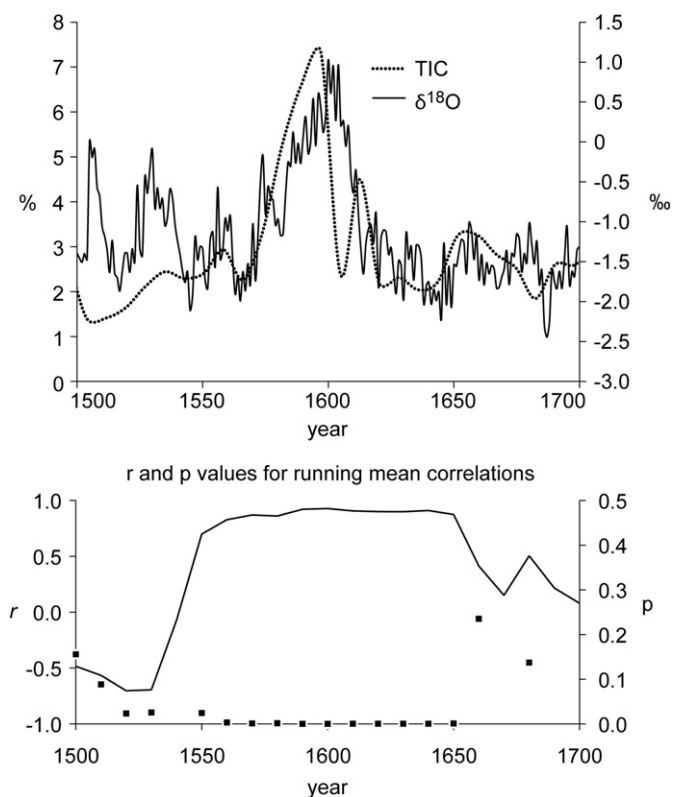

Fig. 3. Nar isotopes vs Montcortès TIC during the late C16th climate anomaly (drought in central Turkey; wetness inferred for northern Iberia).
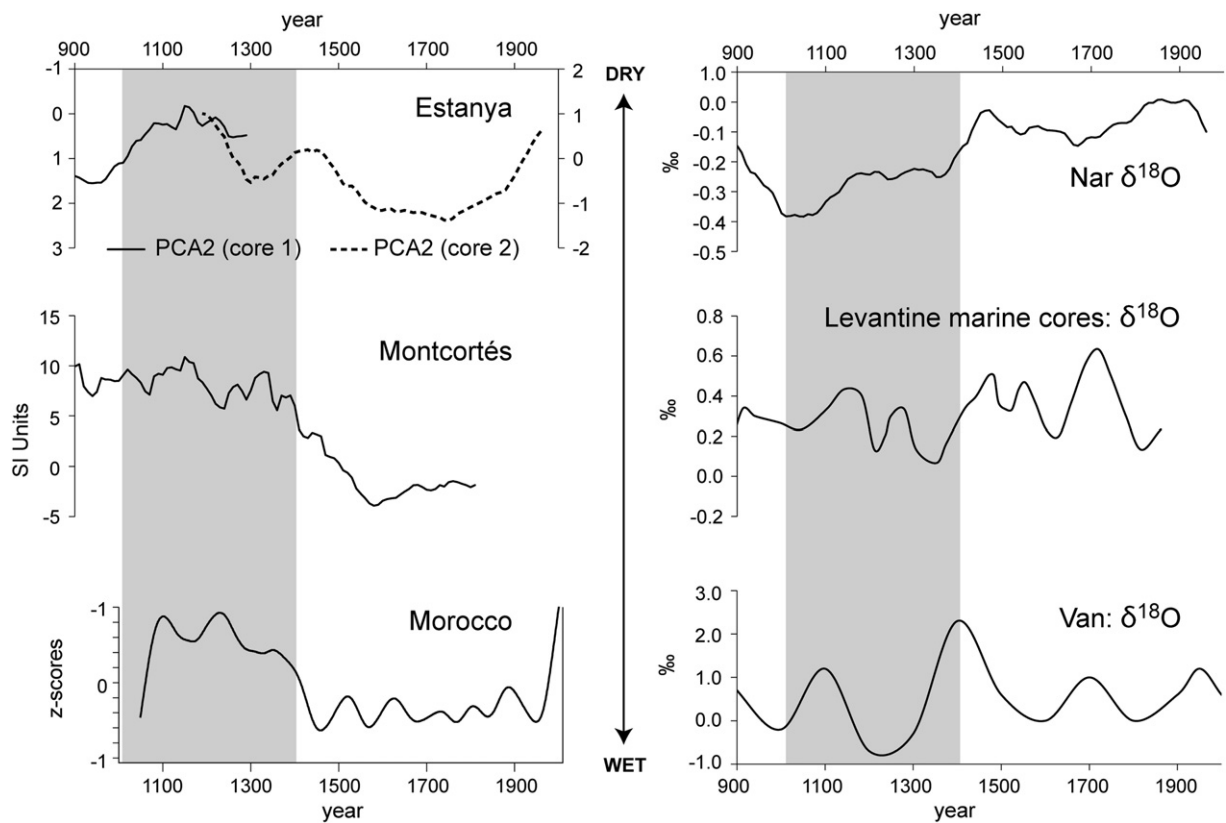
Fig. 4. Centennial trends in proxy climate records: a) western Mediterranean; Montcortès MS, Estanya PCA axis 2, Moroccan tree ring data (Esper et al., 2007); b) eastern Mediter-ranean; $\delta^{18}O$ for Nar, Van (Wick et al., 2003), and Levantine marine cores (Schilman et al., 2001). The Medieval period 1000–1400 AD is shaded.

Following the reasoning outlined in Section 2 above, it would be expected that different hydro-climate proxies from the same core se-quences should exhibit strongest correlation coefficients, and indeed this is the case overall. The highest correlations are shown between $\delta^{18}O$ and grey scale at Nar ($r = 0.85$; $p = 0.000$), while for Nar diatoms and for the three proxies from Montcortès, within-core correlations range from 0.48 to 0.65 and are significant at the 0.05 level or better. For Estanya, for which two separate core records are available, decadal-resolution correlation for the 200-yr overlap period is not significant ($r = 0.31$, $p = 0.217$). Even though visual comparison of the two records shows similar centennial trends, at annual to decadal timescales there is a relatively poor match, most likely due to dating imprecision. Even a 2% age error at AD 1000 would be sufficient to substantially weaken decadal correlations between two cores. Per-haps significantly, r values between the Montcortès proxies and the 800-yr Estanya core 2 record are much higher (0.44 to 0.52) than with Estanya core 1. This might imply that the dating problem lies primarily with the longer Estanya core 1, which is less well-dated for the late Holocene (Morellón et al., 2009).

Correlation coefficients are therefore consistently higher than 0.44 for different proxies within the same core record at both Nar and Montcortès, and between lake records in the same climatic region (Montcortès vs Estanya core 2). In terms of comparing the records from different climate regions, namely northern Spain and central Turkey, six of the nine comparisons between Nar and Montcortès proxies have r values between 0.42 and 0.47. All of the Nar–Mon-tcortès–Estanya core 2 correlations are statistically significant at the 0.05 level (Table 2). Given the existence of non-climatic as well as cli-matic factors, it seems probable that Nar and Montcortès are record-ing common, but inverse, responses to climate forcing at decadal and longer timescales over the period 900–1850 AD. This is given support by the fact that the Nar $\delta^{18}O$ record shows a drought from 1580 to

1610 AD which is independently recorded in tree-ring data and his-torical sources from Anatolia (Kuniholm, 1990; Touchan et al., 2007; White, 2011 and references therein). This dry phase, also present in the Nar diatom record (Woodbridge and Roberts, 2011), is marked by a sharp peak in Total Inorganic Carbon at Montcortès (Fig. 3 upper graph) indicative of positive lake water balance and wet hydro-climatic conditions, and also coincides with geochemistry PCA axis 1 minimum at Estanya (Morellón et al., 2009). At both Nar and Montcortès, this is the single largest inferred climatic anomaly within the LIA, and is consistent with a chronological precision be-tween the two records of ±5 years. 100-year moving window corre-lations between "binned" Montcortès TIC and Nar $\delta^{18}O$ show decadal-mean r values above +0.8 and p values lower than 0.003 for the period from 1560 to 1650 AD (Fig. 3 lower graph), highlighting the strength of the anti-phase relationship during this part of the LIA. This late 16th-century climate episode was matched by notably cold summer temperatures in the Alps (Büntgen et al., 2006) and central Europe (Luterbacher et al., 2004; Dobrovolný et al., 2010).

4. Regional multi-proxy climate evidence

To what extent are the changes observed in these three well-resolved lake sediment records also indicated in other hydro-climatic archives from the Mediterranean? Here we review a range of evidence from the west and then the east Mediterranean, in each case starting with other palaeolimnological data, since these should be most directly commensurable with those analysed above, before examining selected dendro-climatological, marine and other data sets. These are described more fully in Luterbacher et al. (in press) and in Moreno et al. (in review).

A broadly consistent hydrological pattern has been found in other lacustrine sequences of northeast Spain (Moreno et al., 2011, in

review; Morellón et al., 2011a). Decreased carbonate precipitation and higher runoff occurred between the 14th and the 19th centuries in Lake Basa de la Mora (Pyrenees), and variable but lower carbonate precipitation and better preservation of varves, indicative of higher lake levels, occurred in Lake Arreo (NW Ebro Watershed) during the same period. In central Spain, the Taravilla lake sequence reflects changes in the intensity of palaeofloods (Moreno et al., 2008) also reflected in fluvial activity reconstructions from the Tagus River (Benito et al., 2003). The Taravilla lake record shows minimum frequency of extreme flood events during medieval times and an increase since the 14th century. Pollen analyses suggest that these flood events are not related to deforestation or increased human impact. In La Cruz Lake (Cuenca province) (Julià et al., 1998), lower lake levels occurred during the 9th–11th centuries, indicative of drier conditions. The development of meromictic conditions during the LIA was related to the synergetic effects of colder temperatures and higher lake levels suggesting wetter conditions. In southern Spain, sedimentological data from the karstic, 15 m deep Zoñar Lake (Martín-Puertas et al., 2008) indicates arid conditions synchronous with the MCA and two humid periods between 1200 and 1400 AD and around 1600 AD during the LIA. Lower lake levels and higher salinities are therefore reconstructed from all Iberian records for medieval times (9th to 13th centuries, drier), synchronous with the MCA, and generally colder and more humid conditions during the LIA (15th–19th centuries).

These findings are replicated in shallow marine records from the Tagus estuary and from the rias of northwest Spain (Lebreiro et al., 2006) where the input of fluvial-derived sediments clearly decreases during medieval times. Deep marine sediment cores from the Alboran and Balearic basins indicate an increased influx of Saharan dust and a reduction in fluvial particles between 950 and 1250 AD (Moreno et al., 2011, in review). These records therefore point to arid conditions in northwest Africa and/or more persistent winds coming from a southwesterly direction. On land, the cedar tree ring sequence from Morocco's Middle and High Atlas mountains provides a highly-resolved late winter to early summer hydro-climatic record for the last 950 years (Esper et al., 2007) and gives support to the MCA having been climatically drier than the subsequent LIA. Values of the long-term February–June Palmer Drought Severity Index (PDSI) were above average for the period AD 1400–1980, and below average before that time (Fig. 4).

In the eastern Mediterranean there are currently no other lake records with dating or sampling resolution similar to Nar for the last millennium. However, the Nar sequence can be compared with other lower-resolution lake records, such as the centennial-resolution isotope sequence from Lake Van (Wick et al., 2003) which is also varved. Van is the largest soda lake in the world by volume, and is likely to have a somewhat damped response to short-term fluctuations in climate. Fig. 4 compares $\delta^{18}O$ data for Van with the 100-year mean values for Nar. These two lake records show similar overall trends for the last 1100 years, with a shift to drier hydro-climatic conditions at 1350–1400 AD, following a generally wetter phase during the MCA (AD 950 to 1300). Further south, there is a record of Late Holocene lake-level fluctuations from the Dead Sea from a sequence of well-dated palaeo-shorelines (Bookman et al., 2004; Migowski et al., 2006). This provides a semi-continuous record of lake highstands, with low lake-level stages largely based on inference, for example the presence of an unconformity at AD 1400, following higher water levels at 1100–1300 AD. Dead Sea water level fluctuations during recent times have been significantly affected by human impact, including abstraction of Jordan river water which has led to desiccation of the shallow southern basin (Enzel et al., 2006).

Although there is now a significant body of tree ring width data for the eastern Mediterranean, measurement series has been standardised in a manner that eliminates low-frequency variability, so that long-term hydro-climatic changes cannot currently be reconstructed from these archives (Luterbacher et al., in press, and references therein). Most tree ring reconstructions (e.g. Touchan et al., 2005; 2007) also find their strongest relationships with early summer precipitation, which is not a significant rainfall season for most of the Mediterranean.

There are a number of excellent speleothem records from the eastern Mediterranean spanning Holocene or longer timescales, but none so far published which covers the last 1100 years at high temporal resolution. In western Turkey three cave sequences are currently under analysis which cover this time period (Göktürk et al., 2011; Luterbacher et al., in press), but preliminary results suggest that the MCA and LIA may only be weakly expressed, with poor coherence between the stalagmite records from the three caves. There is also a speleothem isotope record spanning the last 500 years from the summer-green climate region of northeast Turkey which shows a period of anomalously low precipitation during the 16th century (Jex et al., 2010, 2011).

Finally, a few marine records have been analysed at relatively good resolution for the Late Holocene, notably two cores taken at intermediate water depths from a high sedimentation site off the coast of Israel (Schilman et al., 2001). This isotope record has a mean sampling interval of ~30 years for the period 900–1900 AD, although it is not well dated or resolved for the last two centuries. As at Nar it indicates a wet phase during the latter part of the MCA, 1200–1400 AD, followed by a dry shift during the 15th century AD (Fig. 4). Following isotope maxima around 1450–1580 AD, a second LIA dry phase is indicated around 1700 AD.

MCA–LIA hydroclimatic conditions across the Mediterranean region were the result not only of annual average precipitation totals, but also of changes in temperature and seasonality. Temperature changes are clearly indicated in marine records from the Mediterranean region (e.g. Taricco et al., 2009) and in altitudinal shifts in montane regions such as the Pyrenees (Morellón et al., 2011a). Nonetheless, for proxies such as lake sediments and tree rings, the clearest integrated expression of climatic changes over the past 1110 years in the Mediterranean has been effective moisture availability. Our synthesis shows only partial agreement with the tree ring-derived PDSI reconstruction for the Mediterranean since 1500 AD by Brewer et al. (2007). In particular, they infer relatively dry (wet) conditions in the western (eastern) Mediterranean prior to ~1670 AD, which are not evident in the lake records presented here, nor in the wider array of proxy-climate data discussed by Luterbacher et al. (in press).

In addition to these hydro-climatic trends during the MCA and LIA, it should be noted that many lake records show climatic changes of significantly larger amplitude in the preceding 1500 years, notably during and after the Roman period. For example, Zoñar Lake in southern Spain contains a varved interval deposited during the Iberian–Roman ages (550 BC–AD 350), although it includes an arid interval during the Roman Imperial Epoch (190 BC–AD 150) (Martín-Puertas et al., 2009). The interval from ~300 BC to 600 AD includes the most humid conditions of the last three millennia not only in southern Spain, but also in parts of the eastern Mediterranean, where Dead Sea lake levels reached a maximum between 200 BC and 100 AD (Bookman et al., 2004). A wet Roman interval (pre-140 BC to AD 100) is also clearly evident in high-resolution isotopic analysis of speleothems from Soreq cave in Israel (Orland et al., 2009). There is evidence of major wet–dry shifts in central Anatolia during the first millennium AD, including a major drought event around 380–540 AD. In contrast to the LIA, this dry phase was of sufficient intensity and duration to make lake water strongly saline at Nar according to diatom data (Woodbridge and Roberts, 2011) and for aragonite precipitation in Tecer lake (Kuzucuoğlu et al., 2011). Because absolute chronologies for Classical times are not always as precise as during the last millennium, correlations between records can in turn be less secure. This makes it uncertain whether hydro-climatic changes between east and west Mediterranean were in or out of phase with each other during the Roman period, although an anti-phase relationship between southern Spain and the southern Levant is suggested
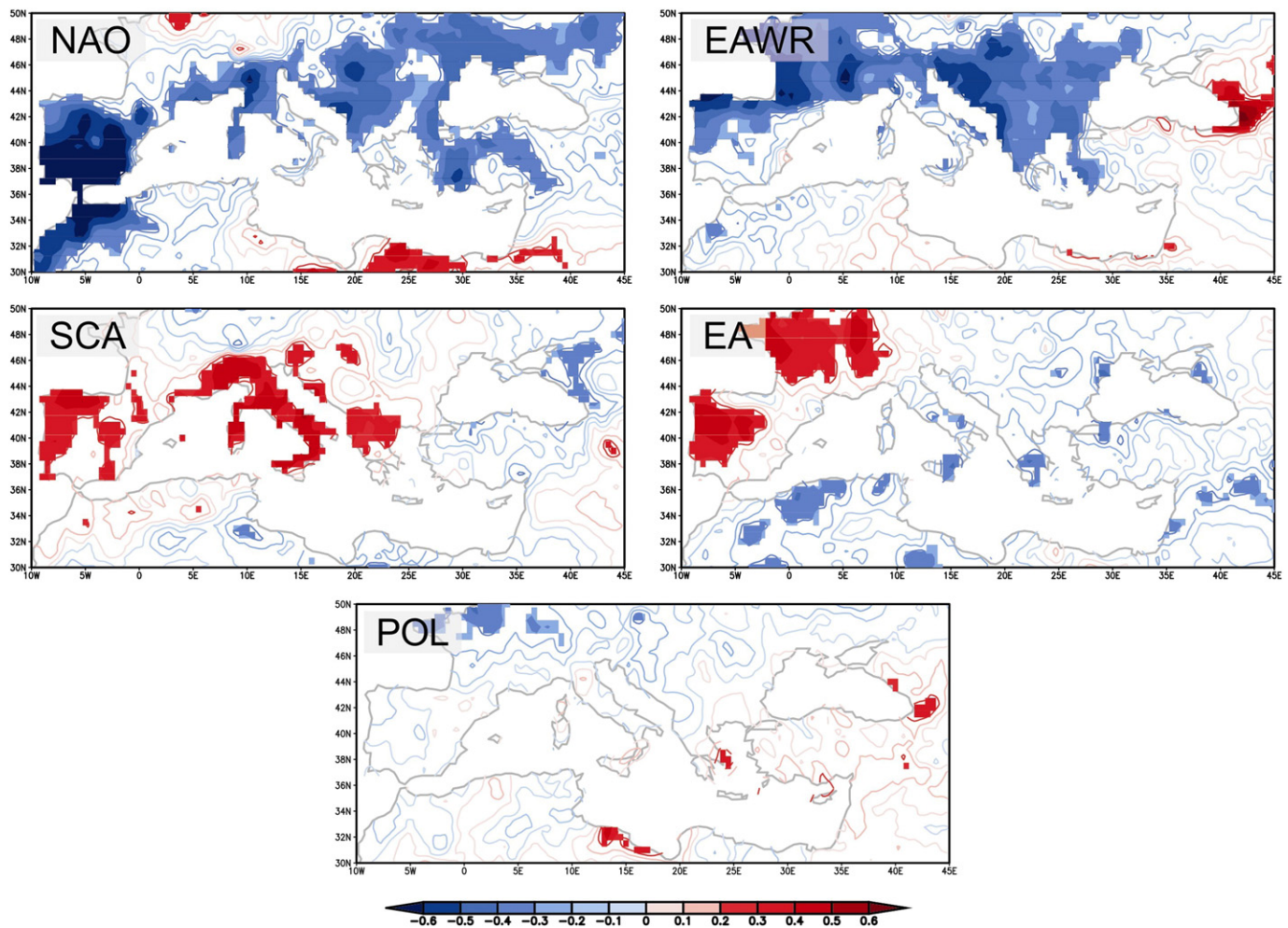
Fig. 5. Spatial Spearman correlations between 5 different atmospheric teleconnection patterns (defined by Barnston and Livezey, 1987) and winter precipitation (1951–2006) in the Mediterranean. The shaded areas denote significant at the 95% level. NAO = North Atlantic Oscillation; EAWR = East Atlantic/Western Russia; SCA = Scandinavian; EA = East Atlantic; POL = Polar/Eurasia.

for the period from ~200 BC to AD ~100. In any case, it is significant that the MCA and LIA do not capture the full amplitude of climatic variability in the Mediterranean under Late Holocene (i.e. modern, pre-Anthropocene) atmospheric boundary conditions.

5. Comparing atmospheric circulation modes and proxy data

The pattern of climatic change across Iberia, Morocco and the adjacent western Mediterranean seas for the last millennium is clear and consistent; the MCA (1000–1400 AD) was overall drier than the

LIA (1400–1850 AD; Esper et al., 2007; Moreno et al., 2011, in review). This appears to be consistent with quasi-permanent NAO forcing, with negative index states being more common during the LIA and positive ones during the MCA as proposed by Trouet et al. (2009). In the eastern Mediterranean, the well-dated evidence from Nar lake points to a pattern since 900 AD opposite to that in the western Mediterranean at both decadal and centennial timescales. This appears to indicate that a precipitation see-saw has operated in the Mediterranean during the last eleven centuries, with the west being wet when the east was dry, and vice-versa.
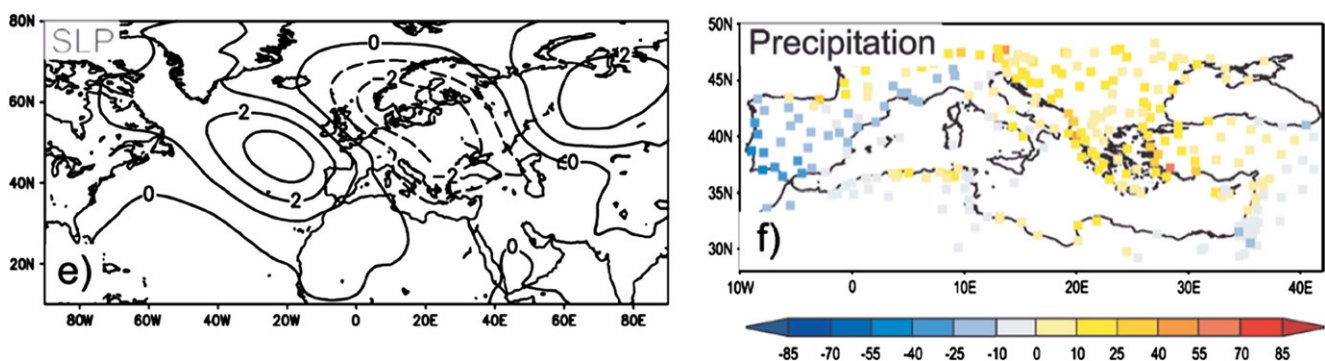

Fig. 6. Canonical spatial patterns of the second CCA of Xoplaki et al. (2004), depicting typical sea level pressure anomalies in hPa (left) and wet season precipitation anomalies in mm (right).

On the other hand, proxy climate records from the southern Levant, such as the Dead Sea levels and the marine isotope sequence shown in Fig. 4, carry a signal that is less clearly opposite to that from the western Mediterranean during the last millennium, even if they bear a partial resemblance to the Nar lake sequence. Similarly, seasonal instrumental precipitation variations in central Anatolia and northern Spain (using gridded CRUTS3 data; Mitchell and Jones, 2005, updated) reveal no correlation within the 20th century. Previous studies (e.g. Cullen and deMenocal, 2000; Türkeş and Erlat, 2003) have shown that the Central Anatolian rainfall region has had an overall positive correlation with the negative NAO index state during the last ~100 years; that is, one broadly in-phase rather than out-of-phase with precipitation changes in Iberia. This could imply that the forcing which generated the east–west precipitation see-saw for the 1000-year period prior to instrumental records was not directly NAO-related, or that the character of the NAO and its teleconnections has been non-stationary over the timescales of the MCA and LIA (c.f. Jones et al., 2003). Mann (2002), for example, concluded that whereas the NAO has been linked to eastern Mediterranean temperature variations over inter-annual to decadal timescales, other patterns have been more important on multi-decadal and longer timescales. Could the MCA/LIA east–west see-saw therefore be explained by other climate modes? For the instrumental period, a number of other atmospheric teleconnections have been proposed to be of relevance for the Mediterranean. For central Turkey, one of the strongest winter-season relationships is the North Sea–Caspian Pattern Index (NCPI; Kutiel and Türkeş, 2005), a local expression of the East Atlantic/Western Russia (EAWR) pattern identified by Barnston and Livezey (1987). On the other hand absolute differences in precipitation amounts between the two NCPI index states are less significant than those in temperature.

Fig. 5 shows the spatial Spearman correlation between the leading five northern hemispheric teleconnection modes (Barnston and Livezey, 1987) and winter land-based precipitation across the Mediterranean for the period 1951–2006. It can clearly be seen that there is no single mode that can account for the west–east differences in precipitation. The EAWR pattern and the NAO show a significant negative correlation with eastern Spain but not a significant positive correlation for the eastern basin. The Scandinavian (SCA) and East Atlantic (EA) patterns show a positive relationship with eastern Spain, but none of those teleconnection patterns on their own can account for a significant amount of winter precipitation variability over the past ~60 years across the whole Mediterranean basin.

The seesaw-like oscillation between the drier conditions in the western and wetter conditions in the eastern Mediterranean during the MCA shows a resemblance with results presented by Dünkeloh and Jacobeit (2003) and Xoplaki et al. (2004). Xoplaki et al. (2004) applied a Canonical Correlation Analysis (CCA) to the extended wintertime wet period (October–March) precipitation anomalies in the Mediterranean and large-scale dynamics at different heights and SSTs. The combination effect of large scale circulation influence represented by the first canonical pair accounts for ~10% of the precipitation variations in the Mediterranean. The spatial characteristics of this pattern are in agreement with Dünkeloh and Jacobeit (2003) for coastal Mediterranean precipitation. The first canonical mode shows significant negative correlation with the NAO and EAWR patterns (−0.66 and −0.50, respectively, Xoplaki et al., 2004) suggesting a synergistic influence of the two patterns with different signature on the western and eastern parts of the Mediterranean. Eshel and Farrell (2000) and Eshel et al. (2000) advanced a simple theory explaining extended winter (October–March) eastern Mediterranean rainfall variability in terms of subsidence anomalies associated with large-scale North Atlantic anomalies. Their concept of anomalous high pressure over Greenland/Iceland and accompanying concurrent depressions over the Mediterranean connected with anomalous warm southerlies and higher precipitation amounts in

the region, is very similar to the structure of the first CCA of extended winter precipitation in Xoplaki et al. (2004). The 700 to 300 hPa anomaly patterns indicate an increasing influence of a positive (negative) anomaly over southwest Asia. This anomaly is absent in the lower troposphere (Eshel and Farrell, 2000; Eshel et al., 2000). The analysis of Xoplaki et al. (2004) indicated that it is important to include the mid- and upper level large-scale atmospheric circulation in order to explain regional differences (west–east) in precipitation variability.

The second CCA pattern of Xoplaki et al. (2004) is linked to a predominantly meridional circulation associated with wet northeast Mediterranean and northern Levant, as well as a dry Iberian Peninsula (Fig. 6) and is significantly correlated with the Polar/Eurasia pattern (POL; Barnston and Livezey, 1987). This cyclonic anomaly pattern causes similar influences with a large-scale cyclonic anomaly circulation pattern over the central and eastern Mediterranean basins and parts of North Africa, producing increased precipitation (Türkeş and Erlat, 2005, 2006) and increased temperature conditions (Türkeş and Erlat, 2008) for most of west, central and southern Turkey during the winter months. The responsible atmospheric pattern is connected to the influence of the subtropical high (positive geopotential height anomalies) that is restricted to the Iberian Peninsula and northwestern Africa connected with subsidence, stable conditions and reduced precipitation. The areas of enhanced precipitation amounts are located in the southeastern part of the anomalous trough stretching from Greenland over Central Europe to the coast of North Africa. In this sector of the trough, the vorticity advection is maximally connected with strong uplift, instability, condensation and a high chance of precipitation (Xoplaki et al., 2004). As in Xoplaki et al. (2004) the second canonical mode of Dünkeloh and Jacobeit (2003) is connected with a precipitation pattern across the Mediterranean that appears closer to that manifest from palaeo-data for the MCA and LIA, namely with opposing centres of variability in central Turkey and northern Spain. This inverse relationship is particularly strong for the period 1560–1650 AD, and suggests that a meridional atmospheric circulation may have predominated during this 100-year period. Perhaps significantly, this east–west hydro-climatic pattern broadly corresponds to that displayed both in model simulations and palaeo-data for longer Holocene timescales (Roberts et al., 2011).

Additionally, seasonal factors may have come into play, notably in spring, which in central Turkey is more important than winter in terms of total precipitation. Spring season CCP2 of Dünkeloh and Jacobeit (2003) resembles that of winter CCP1, but the western centre shifts north towards the Bay of Biscay, creating an inverse pattern of precipitation variability between Iberia–Morocco and Greece–Turkey, similar to that observed over the last millennium. Xoplaki et al. (2004) clearly point to the fact, that despite the importance of the large-scale atmospheric features, smaller scale processes also influence regional rainfall variability in the Mediterranean. Among them, land–sea interactions, the influence of the SSTs connected with latent and sensible heat flux, orographical features and thermodynamical aspects interact with each other on different timescales and are superimposed on the quasi-stationary planetary waves which control large-scale advection.

6. Conclusions

Spatial patterns in hydro-climate revealed by proxy data have the potential to distinguish between alternative modes of past atmospheric circulation. During the period of instrumental records, there is clear evidence of a spatial signature to variations in precipitation in the Mediterranean region which can partly be linked to the NAO. This relationship is clearest in parts of Iberia and northern Morocco, where increases in precipitation are associated with anomalous high pressure dominance over Northern Europe (negative NAO conditions). In parts of the eastern Mediterranean, notably the southern

Levant and northeast Africa, an anti-phase relationship in precipitation and atmospheric pressure with the western Mediterranean has operated over the last two centuries.

Trouet et al. (2009) proposed that the western Mediterranean experienced a persistent positive NAO state during the MCA. If correct, then it might be expected that the northwest–southeast "Mediterranean Oscillation" would also have operated during the LIA and MCA. High-resolution palaeolimnological data from northern Spain show good inter-site coherence, and indicate lower water levels and higher salinities synchronous with the MCA and generally more humid conditions during the LIA. This pattern is confirmed by other lake, marine and tree-ring records from Iberia and Morocco (Moreno et al., 2011, in review). In contrast, Nar lake in central Turkey shows an opposite pattern of wet MCA and a dry LIA. This is supported by statistically-significant cross-correlations of decadal-average proxy-climate data between lake records, and by other, lower-resolution marine and lake data from the eastern Mediterranean. The relationship between different proxies and hydro-climatic indices (e.g. PDSI) is unlikely to have been simple or linear, for example, because of proxy threshold responses. None the less, it seems likely that an east–west bipolar climate see-saw has operated in the Mediterranean for the last 1100 years.

While western Mediterranean aridity appears consistent with NAO forcing during the MCA, the relationship is less clear in the eastern Mediterranean. Currently the strongest evidence for higher winter-season precipitation during the MCA comes primarily from central Anatolia, which—unlike the southeast Mediterranean—has not shown an inverse relationship with NAO phases during the period of instrumental record. Our results therefore do not give support to the Mediterranean Oscillation as a meaningful atmospheric pattern over the pre-instrumental time period. Apparent differences in the strength and positions of the anomaly circulation patterns of the NAO phases may be due to the additional influences of other climate variability modes related to more meridional atmospheric configurations (e.g. anomalous higher pressure in the west, anomalous lower pressure in east). This in turn implies that the LIA/MCA hydro-climatic pattern in the Mediterranean was determined by a combination of climate modes, such as the EAWR in synergy with POL, along with major physical geographical controls, and not solely by NAO forcing. Alternatively or additionally, the character of the NAO and its teleconnections may have varied over the timescales of the MCA and LIA (c.f. Jones et al., 2003). Whatever the precise explanation, it is clear that changes in the climate of the Mediterranean have not been uniform over the past millennium, but instead have shown contrasting spatio-temporal trends and patterns across the basin.

Acknowledgements

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