IMPACT OF THE NORTH ATLANTIC OSCILLATION ON MIDDLE EASTERN CLIMATE AND STREAMFLOW

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Abstract. Interannual to decadal variations in Middle Eastern temperature, precipitation and streamflow reflect the far-field influence of the North Atlantic Oscillation (NAO), a dominant mode of Atlantic sector climate variability. Using a new sea surface temperature (SST) based index of the NAO and available streamflow data from five Middle Eastern rivers, we show that the first principal component of December through March streamflow variability reflects changes in the NAO. However, Middle East rivers have two primary flooding periods. The first is rainfall-driven runoff from December through March, regulated on interannual to decadal timescales by the NAO as reflected in local precipitation and temperature. The second period, from April through June, reflects spring snowmelt and contributes in excess of 50% of annual runoff. This period, known locally as the *khamsin*, displays no significant NAO connections and a less direct relationship with local climatic factors, suggesting that streamflow variability during this period reflects land-cover change, possibly related to agriculture and hydropower generation, and snowmelt.

1. Introduction

Although Turkey has the good fortune of being home to the headwaters of the Tigris and Euphrates rivers, with primary control on the limited freshwater available to the Middle East and geopolitical advantage over Syria and Iraq, a widespread drought throughout the Middle East and Central Asia (Barlow et al., 2002) since 1998 has demonstrated that surface freshwater exhibits temporal as well as spatial variability. The map in Figure 1 depicts the accumulated precipitation deficits (at intervals of 150 mm) for the period November through April, 1998–1999 to 2000–2001, using the CAMS-OPI dataset, a blend of satellite and station data (Janowiak and Xie, 1999).

As energy demands climb in Turkey, Syria, and Iraq, and each country moves to increase withdrawals from Middle Eastern rivers (Figure 1), it is critical that climate variability, both natural and anthropogenically forced, be objectively incorporated into water resource related decision making. Turkey has significant hydroelectric power resources (104 hydropower stations, installed capacity 10.2 gi-



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Figure 1. Inset globe shows the domain $(23^{\circ}-55^{\circ} \text{ E}; 29^{\circ}-45^{\circ} \text{ N})$ (box) used in the statistical analysis. The enlarged map depicts accumulated precipitation deficits (at intervals of 150 mm) for the period November through April, 1998–1999 to 2000–2001, using the CAMS-OPI dataset, a blend of satellite and station data (Janowiak and Xie, 1999). The six streamflow stations used in the analysis are marked with circles and the major rivers are labelled.

gawatts (GW)). Major Turkish dams include the Ataturk (2,400 megawatts (MW)), the Keban (1,330 MW) and the Karakaya Dam (1,800 MW) all on the Euphrates River. The Ilisu Dam (1,200 MW) is the largest hydropower project on the Tigris River. Electricity production in 1999 was 116.5 billion kWh with hydropower making up ~30.5%. Syria's total installed electric generating capacity as of 1999 was 4.5 GW with ~20% coming from hydropower. New hydopower facilities include the 630 MW Tishreen hydropower station built on the Euphrates and completed in 1997. Electricity production in 1998 was 17.5 billion kWh with hydropower making up ~57%. Approximately 90% of Iraq's national power grid was destroyed in the Gulf War. In early 1992, Iraq claimed to have restarted 75% of the national grid, including the Saddam Dam (750 MW) on the Tigris River. Electricity production in 1998 was 28.4 billion kWh with hydropower making up less than 5% (DOE, 2001).

In addition to management factors and land-use change, interannual to decadal changes in December through March precipitation and streamflow in the eastern Mediterranean are tied to the North Atlantic Oscillation (NAO), a dominant mode of Atlantic sector climate variability (Cullen and deMenocal, 2000a; Eshel and Farrell, 2000). Like the El Niño Southern Oscillation (ENSO) (Simpson et al., 1993a; Cane et al., 1994), the NAO has been shown to impact both marine and terrestrial ecosystems (Fromentin and Planque, 1996; Alheit and Hagen, 1997; Beniston, 1997). However, unlike ENSO, there is still considerable doubt as to whether the NAO is a coupled ocean-atmosphere process. Coupled, or two-way interactions between the ocean and atmosphere suggest the potential for a degree

of predictability. Predictability resides primarily within the long-term memory of the ocean and provides the opportunity for long-range climate forecasting (Mason et al., 1999). The only unquestioned example, of a coupled ocean-atmosphere interaction is ENSO (Philander, 1983, 1985; Cane and Zebiak, 1985). Some have argued the case for coupled ocean-atmosphere interactions in the North Atlantic (Rodwell et al., 1999), however this remains to be seen.

This paper attempts to explore the seasonal relationship between streamflow and large-scale climate variations with the hope of generating an awareness that natural climate variability, coupled with global warming, management factors, and land-use change, could significantly alter interannual to decadal streamflow. In Section 2, we introduce the major rivers of the Middle East, and describe the motivation for a statistical reconstruction of streamflow in the region. As data is often limited in this region we create a reconstruction of the streamflow time series using a statistical technique which employs reduced space optimal interpolation to fill in missing records. In Section 3, we describe the NAO and provide a description of the dynamics linking the NAO and the Middle East. In Section 4, we define a new sea surface temperature (SST) based index of the NAO and investigate its relationship with Middle East temperature, precipitation and streamflow. We then look at the first principal component of this optimally interpolated streamflow dataset and quantify its climatic connections to the NAO during both winter and spring.

2. Major Rivers of the Middle East

The Middle East has a Mediterranean macroclimate, characterized by cool, wet winters and hot, dry summers (Figure 1). Most of the Middle East lacks access to surface and groundwater resources due to the subtropical predominance of evaporation over precipitation. One important exception is Turkey, which has abundant precipitation resulting from the orographic capture of winter rainfall from eastward propagating mid-latitude cyclones generated in the Atlantic Ocean and the eastern Mediterranean Sea. Among the major river systems of the Middle East are the Tigris-Euphrates and the Jordan-Yarmouk Rivers. Both river systems cross international borders and transboundary water disputes over rightful entitlement have been ongoing. Not considered a major river, the Ceyhan is contained completely within Turkey, flowing down from the Taurus Mountains with a combined discharge of 14 billion cubic meters (BCM) per year.

2.1. THE TIGRIS AND EUPHRATES RIVERS

Originating in the Taurus Mountains and Anatolian Highlands of Turkey, the Tigris and Euphrates Rivers, annually discharge \sim 74 BCM and exhibit significant interannual variability. After leaving Turkey, the Tigris and Euphrates Rivers flow into Syria and Iraq, ultimately converging to form the Shatt Al-Arab and draining into the Persian Gulf (Soffer, 1999). The Tigris-Euphrates system, with a combined drainage area of $4.1 \, 10^5 \, \text{km}^2$, has two primary flooding periods: the first, December through March, is due mostly to surface runoff from winter rainfall; the second rise, from April to June, reflects spring snowmelt (Hillel, 1994).

The Euphrates is the combination of two tributaries: the Karasu and the Murat which begin on the slopes of Mount Ararat north of Lake Van. The Keban Dam is situated at a point where these two tributaries meet (Figure 1). After Keban, the Euphrates flows through the southeastern Taurus Mountains and crosses into Syria where the major tributaries are the Balikh and the Khabur. At 360 km downstream from the Iraqi-Syrian border, the river reaches its alluvial delta in Iraq. Farther downstream, a portion of the river drains into Lake Hammar while the remainder joins the Tigris. Mean annual Euphrates streamflow is approximately 35 BCM, of which 30 BCM forms from surface runoff and spring melt within Turkey; with large interannual variations (Kolars, 1997). The Euphrates is 2,700 km long with 41% of its flow in Turkey, 24% in Syria and 35% in Iraq.

The Tigris forms in the eastern part of Turkey near Lake Hazar, only 30 km from the headwaters of the Euphrates, and flows southeast, where it forms the border between Turkey and Syria before entering Iraq. The major tributaries of the Tigris are the Greater Zab, the Lesser Zab, the Adhaim, the Diyala, and the Karun. The longest tributary of the Tigris, the Karun, begins in Iran flowing westward into Iraq. Mean annual Tigris streamflow is approximately 50 BCM, of which 23 BCM originates in Turkey, with the rest added in Iraq. The Tigris is 1,900 km long with 21% of its flow in Turkey, 77% in Iraq, and 2% in Syria.

2.2. THE JORDAN-YARMOUK RIVERS

Although small when compared to the Tigris-Euphrates, the Jordan-Yarmouk drainage basin extends over Israel, Jordan, Syria, and Lebanon making it geopolitically important. The Jordan originates at the foot of Mount Hermon, a region which receives 1,300 mm of precipitation per year. The northern sources of the Jordan are the Dan which delivers on average 245 million cubic meters (MCM) per year, the Banias which delivers average 120 MCM per year, and the Hasbani which delivers on average 128 MCM per year. These three tributaries converge in Israel and form the main stream of the Jordan which then flows southward into Lake Kinnerat, eventually draining into the Dead Sea (Shapland, 1997). Shortly after its exit from Lake Kinnerat, the Jordan is joined by its primary tributary, the Yarmouk. Farther south, the Jordan is joined by two additional tributaries, the Harod and the Yabis, then finally the Fariah, the Zarqa, and the Nusayrat. Before its waters were diverted for irrigation purposes the average annual discharge was 1.2 BCM of which 660 MCM came from the upper Jordan, 475 MCM came from the Yarmouk. The Jordan is 330 km long with 54% in Jordan, 30% in Syria, 14% in Israel and 2% in Lebanon (Hillel, 1994; Soffer, 1999).

2.3. STREAMFLOW ANALYSIS

Despite significant steps taken to gather and maintain a global archive of streamflow data during the International Hydrological Decade, coordinated by the United Nations Educational, Scientific, and Cultural Organization (UNESCO) in the 1960s, there has been little effort to maintain this network in the Middle East and most instrumental records cover relatively short, non-overlapping time intervals ending in the mid-70s or early 80s. This lack of data greatly complicates the use of standard statistical techniques like principal component analysis (PCA), canonical correlation analysis (CCA), and singular value decomposition (SVD) (Bretherton et al., 1992). Evans et al. (2000b) developed a method to approach such problems and applied it successfully to isolating modes of global variability in coral δ^{18} O, an important climate proxy. As a result, isolating the dominant climatic contributions to variability in individual records of streamflow can be employed as a means of reconstructing stations with sparse coverage. In fact, the strong connection between streamflow and ENSO allowed seasonal to interannual streamflow prediction of the Australian rivers, Darling and Murray, (Simpson et al., 1993b) as well as the Nile River (Elfatih and Eltahir, 1996). Here we apply, with some modification, this method to five streamflow records representing discharge in the Tigris-Euphrates, Jordan-Yarmouk, and Ceyhan Rivers, as a means of quantifying the control of the NAO on Middle Eastern water resources.

3. The North Atlantic Oscillation (NAO)

The NAO is the dominant source of Atlantic sector climate variability, accounting for one-third of the total variance in sea-level pressure and 20–60% of the December through March (DJFM) temperature and rainfall variability in Scandinavia, Greenland, Europe, and the Mediterranean over the past 150 years (Hurrell, 1995). Because the signature of the NAO is strongly regional, a simple NAO index has been defined as the difference between the normalized mean DJFM sea-level pressure (SLP) anomalies at locations representative of the relative strengths of the Azores High (AH) and Icelandic Low (IL). A lower-than-normal (higher-thannormal) IL and higher-than-normal (lower-than-normal) AH result in an enhanced (reduced) pressure gradient and a positive (negative) NAO index. During this +NAO (–NAO) phase, surface winds and wintertime storms moving from west to east across the North Atlantic are stronger (weaker) than usual. As a result, wetter and warmer (drier and cooler) winter conditions occur in northern Europe, Scandinavia and the east coast of the U.S., while cooler and drier (warmer and wetter) winters occur in Greenland and the Mediterranean extending into the Middle East.

Investigation of the power spectrum of the instrumental NAO sea level pressure (NAO_{SLP}) index for the past 130 winters reveals dominant periods of variability, centered on 6–10 years and 2–3 years (Hurrell and Van Loon, 1997). Also significant is the red portion of the NAO power spectrum, which may exhibit significant

modulation on interdecadal and century timescales (Delworth and Mann, 2000). Thus, the possibility exists that NAO-related interdecadal-centennial scale variability may play a principle role in Middle Eastern Holocene climate variability (Mann, 2002).

Over the past several decades, the NAO index has steadily strengthened, rising from its low index state in the 1960s to a historic maximum in the 1990s. This trend accounts for a significant portion of the DJFM temperature increase over Eurasia (Hurrell and Van Loon, 1997; Hurrell et al., 2001). Consequently, the NAO has been introduced into the global warming debate with a search for mechanisms that could resolve to what extent this trend is a combination of anthropogenic perturbation and natural variability (Hoerling et al., 2001).

3.1. PRECIPITATION VARIABILITY ASSOCIATED WITH THE NAO

Here we describe more fully the influence of the NAO on interannual to decadal precipitation variability in the Middle East. We compute composite precipitation anomalies using the 21-year CMAP monthly data set (Xie and Arkin, 1996) over the region (90° W-60° E; 20° N-70° N) extending from the east coast of the U.S. to the Middle East for months with positive and negative values of the NAO index for the period January 1979 through December 1998 (Figure 2). High and low index values were defined as the highest and lowest quartile in the 20-year record for each month. Composite anomalies averaged over the months DJFM, conventionally the period when the NAO is most active and well defined, were then computed. The accepted signature of the NAO, with enhanced precipitation in western Europe during high index periods, and in the eastern Mediterranean during low index periods, is evident (Cullen and deMenocal, 2000a; Eshel and Farrell, 2000). However, this picture provides additional detail, much of it not visible in earlier studies based solely on station observations (Dai et al., 1997). An axis of enhanced monthly wintertime precipitation is seen during low index periods extending from the southeastern U.S. across the Atlantic to the Iberian Peninsula, with negative anomalies approximately parallel to the north and south. During high index periods, the pattern is nearly reversed, but departures from strict linearity can be seen. The enhanced precipitation axis during low index months extends across the northern Mediterranean nearly to the Caspian Sea with an enhanced secondary maximum along the west coast of Spain and Portugal and over Turkey.

A relationship between the eastern Mediterranean and Atlantic sector is to be expected because the NAO regulates Atlantic heat and moisture fluxes into the Mediterranean region (Turkes, 1996a,b). Since these winter cyclones are the dominant source of Middle Eastern rainfall and river runoff, NAO-related changes in Atlantic westerly heat/moisture transport and Atlantic/Mediterranean SST's can be expected to influence Middle Eastern climate.



Figure 2. (a) Composite of the most negative quartile of NAO_{SLP} index years (–NAO phase) and (b) the most positive quartile of NAO_{SLP} index years (+NAO phase).

4. Data and Methodology

The NAO has a coherent signal in SST involving a tripole pattern of almost zonally oriented anomalies with subtropical and high-latitude SST's varying in-phase and mid-latitude SST's varying out-of-phase (Seager et al., 2000). While the NAO reflects changes in the strength and orientation of mid-Atlantic westerlies, which dictate heat and moisture flux trajectories, the NAO is also expressed in terms of distinct North Atlantic SST, sub-surface temperature (Molinari et al., 1997; Reverdin et al., 1997) and sea-ice extent anomalies (Deser and Blackmon, 1993). During the positive phase of the NAO, SST's in the eastern Labrador Sea, eastern Mediterranean, and subtropical Atlantic are anomalously cold by 0.5–1 °C, whereas SST's in the North Sea and Sargasso Sea are anomalously warm by approximately the same amount (Deser and Blackmon, 1993).

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Table I

The five 'centers of action' selected from the North Atlantic sector and their respective correlation with the NAO_{SLP} index (Hurrell, 1995). All five centers are found to be significantly correlated at the 99% confidence level according to the two-tailed t-test

Region	Lat.	Lon.	<i>r</i> -value
North Sea	50–55° N	0–5° E	0.61
Labrador Sea	55–60° N	$45-40^{\circ} \mathrm{W}$	0.57
Sargasso Sea	30–35° N	65–60° W	0.30
E. Mediterranean	30–35° N	30–35° N	0.55
S. Atlantic	10–15° N	45–40° E	0.40

4.1. AN SST-BASED NAO INDEX

Here we develop an SST anomaly index of the NAO to extract the NAO-related component of Atlantic SST variability. We then use this NAO_{SST} index to explore linkages between changes in North Atlantic surface ocean conditions and Middle Eastern streamflow.

The technique we employ is similar in principle to that used for the NINO3 index which tracks ENSO in the tropical Pacific Ocean (Rasmusson and Carpenter, 1982). The NAO_{SLP} index of Hurrell (1995) (Figure 3a) was spatially correlated against global, gridded ($5^{\circ} \times 5^{\circ}$), and standardized SST anomalies between 1857–2000 (Kaplan et al., 1998). From the resulting spatial correlation field the five highest correlation (r_i) 'centers of action' (C_{SST}) were identified, each corresponding to one $5^{\circ} \times 5^{\circ}$ grid SST time series (Table I). All five centers are found to be significantly correlated at the 99% confidence level according to the two-tailed t-test. The five SST anomaly time series ($C_{SST1}-C_{SST5}$) were then combined using the following relation to develop the 1857–2000 NAO_{SST} index, (I(t), in units of °C):

$$I(t) = \frac{\sum_{i=1}^{5} [r_i \times (C_{\text{SST}})]}{5}.$$
 (1)

Criteria for selecting the centers was based upon similar level of correlation among the five and a physically grounded connection linking each center to the NAO. The North Sea, Labrador Sea, and Sargasso Sea regions were identified by Kushnir (1994) as dominant 'centers of action' and explain 70% of the total variance. The eastern Mediterranean and equatorial region were selected based on significant correlation.



Figure 3. (a) The NAO_{SLP} index (Hurrell, 1995) is shown as a dashed line and the NAO_{SST} in $^{\circ}$ C is shown as a solid line. The correlation between the two indices is 0.67 and is significant at the 99.9% confidence level according to the two-tailed t-test. (b) The spectrum the NAO_{SLP} index (dashed) and NAO_{SST} (solid) indices.

4.2. INDEX INTER-COMPARISON

The resulting NAO_{SST} index (Figure 3a) is highly correlated to the NAO_{SLP} index (r = 0.67, significant at the 99.9% confidence level according to the two-tailed t-test). The North Sea, Labrador Sea, and eastern Mediterranean centers are the three strongest variance contributors to the NAO_{SST} index (Table I). The fourth and fifth correlation centers in the central tropical Atlantic and the Sargasso Sea contribute the remaining NAO_{SST} index variance. In Figure 3b we compute the spectral density of both the NAO_{SLP} and NAO_{SST} index. The NAO_{SST} index reflects NAO-related changes in Atlantic surface ocean temperature variability which, because of its high thermal inertia, is an efficient integrator of atmospheric fluctuations and shows less year to year variability.

4.3. SEASON-LAG DIAGRAMS

Middle East (ME) temperature, precipitation, and streamflow display strong seasonality. As a result it is useful to introduce season lag (SL) diagrams where:

$$f(s, l) = corr[x(s+l, t), y(s, t)].$$
 (2)

Contour plots of these fields display the lagged correlation coefficient between climatic timeseries x(s, t) and y(s, t) computed overall available years t and are presented as a function of season s and seasonal lag l. The zero-lag line shows synchronous correlations computed separately for different seasons. The top half of the diagram shows seasonal correlations for x lagging y, and bottom half shows the same for x leading y. SL diagrams are convenient to suggest causalities; if x in some season affects values of y in the same and later seasons, it will be seen as structure of significant correlation values propagating from the zero-lag line to the bottom half of the diagram with slope -1.

We apply this type of analysis to compare the relationship between the NAO_{SLP} and NAO_{SST} indices. Figure 4a suggests that the SLP dipole of the NAO in January, February and March (JFM) influences SST in the five regions used to create the NAO_{SST} index until the onset of summer. The SL diagram can also be used to study autocorrelations of the climatic values separately for different seasons: Figure 4b demonstrates a lack of significant month-to-month correlation in the NAO_{SLP} index, while Figure 4c displays a 4–6 month decorrelation (depending on the season) for the NAO_{SST} index. All SL diagrams cover the period 1938–1984 (n = 47), therefore a correlation of ~0.23 is significant at the 90% confidence level and a correlation of ~0.27 is significant at the 95% confidence level according to the two-tailed t-test.

4.4. MIDDLE EAST TEMPERATURE AND PRECIPITATION

As ME temperature and precipitation are the natural drivers of river runoff and snowmelt, we begin with an investigation of these variables. First, we develop composite indices of ME winter precipitation and temperature anomalies to summarize the interannual to decadal climate variability spanning the interval 1938–1984. For the ME winter precipitation anomaly index, DJFM total precipitation was averaged over the region (26.25° E to 48.75° E) and (37.50° N to 42.50° N) using the precipitation dataset of Hulme (1994). This region captures the headwaters of the five rivers used in this study: the Tigris, the Euphrates, the Yarmouk as well as the Ceyhan and Karun. For the temperature anomaly index, DJFM temperature anomalies were averaged over the region (25.0° E to 50.0° E) and (35.0° N to 45.0° N) using the temperature dataset of Jones (1994).

In order to interpret the influence of the NAO on streamflow we first determine the influence of the NAO on ME precipitation and air temperature using the SL diagrams introduced in Figure 4. Results, presented in Figure 5, suggest that the

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Figure 4. (a) The seasonal lag (SL) correlation between NAO_{SLP} and the NAO_{SST} index, (b) the SL autocorrelation of NAO_{SLP} index and (c) the SL autocorrelation of the NAO_{SST} index. All SL diagrams show correlations for the period 1938–1984 (n = 47), a correlation of ~0.23 is significant at the 90% confidence level and a correlation of ~0.27 is significant at the 95% confidence level according to the two-tailed t-test.



Figure 5. The influence of the NAO_{SLP} index on (a) monthly Middle East (ME) gridded precipitation and, (b) monthly ME gridded temperature. The influence of the NAO_{SST} index on (c) monthly ME gridded precipitation and (d) monthly ME gridded temperature. Gridded precipitation data are from Hulme (1994) and are averaged over the region (26.25° E– 48.75° E; 37.50° N– 42.50° N) for the period 1938–1984. Gridded temperature data are from Jones (1994) and are averaged over the region (25.0° E– 50.0° E; 35.0° N– 45.0° N) for the period 1938–1984.

most robust influence of the NAO is its effect on December and January (DJ) precipitation and DJFM air temperature. This influence can be seen for both SLP and SST versions of the NAO index. However, since winter SLP influences the NAO_{SST} index at a variety of positive lags and this new index is in general much smoother and more persistent than the NAO_{SLP} index, the SL diagrams for the NAO_{SST} index show higher and more positive lag correlations than those for the NAO_{SLP} index. For example, while the NAO_{SLP} index shows no influence on June through October air temperature, the NAO_{SST} index shows continuous synchronous correlation from May through December.

Table II

River gauges used in the present analysis are from the NCAR data catalog (http://dss.ucar.edu/datasets/ds552.0/), covering, incompletely, the period of 1938 through 1984, location, flow, and length (n), in years, of the monthly time series. (Flow values are from Soffer (1999). NA means not available)

River	Location	Flow (BCM)	Length
Ceyhan	Misis, Turkey	NA	13
Euphrates	Keban, Turkey	20.1	36
Karun	Ahvaz, Iran	23.0	20
Tigris _{Mosul}	Mosul, Iraq	21.0	9
Tigris _{Baghdad}	Baghdad, Iraq	39.1	9
Yarmouk	Maqarin, Jordan	1.2	11

5. Optimal Interpolation of Streamflow Data

Although individual streamflow records are affected by both local climatic and basin management factors, we expect regional Middle Eastern streamflow variability to be influenced by large-scale circulation features as well. We analyze monthly streamflow for five ME rivers available from the NCAR data catalog, covering, incompletely, the period of 1938 through 1984. In Table II we show the location, flow (Soffer, 1999), and length (*n*), in years, of each streamflow record used in the analysis. Included in the analysis are only those rivers displaying distinct seasonality, with a significant portion of the discharge occurring in the period from December to July. Seasonal cycles of streamflow estimated for each of the six gauges are presented in Figure 6 (Yarmouk values are increased by a factor of 20 to make the curve more readable). The DJFM period displays an ~1.5 time smaller total water volume than April through July (AMJJ).

We then employ a statistical approach in order to establish connections of streamflow with local (precipitation and temperature) and large-scale (NAO) climatic indices. The scarcity of the streamflow data coupled with the strong seasonality of its climatic connections makes this task difficult. We begin by performing a reduced space optimal interpolation (OI) analysis (Kaplan et al., 1997; Evans et al., 2000) on the standardized monthly streamflow anomalies for each of the twelve months in the seasonal cycle. Standardization was performed by taking differences between the recorded values and the annual cycle of Figure 6 and dividing them by their standard deviations. Seasonal separation involves computing location by location (6×6) covariance matrices separately for each month in a year and prevents interseasonal mixing of dominant spatial patterns of streamflow covariability. Covariance matrices are then computed from these incomplete sets of



Figure 6. The monthly climatology of the six Middle East streamflow series used in this study, peak flows are in winter and spring. Yarmouk values are increased by a factor of 20 to make the curve more readable.

data with the covariance between each two locations estimated from the temporal subset where both locations are simultaneously available. As we are estimating different elements of a covariance matrix from different subsamples, the resulting matrix is not guaranteed to be positive-semidefinite. Indeed, each of our twelve covariance estimates had 1–2 negative eigenvalues. Because of this, we truncate the two trailing eigenmodes from each estimate thus making all matrices positive-semidefinite, before producing the OI analysis. This procedure is equivalent to the OI in a space of reduced dimension (Kaplan et al., 1997). In the OI procedure we assume that all available data has an observational error of 0.5 standard deviation, this ensures that the OI solution correlates with the majority of original records at 0.9 or higher.

5.1. STATISTICAL ANALYSIS OF OPTIMALLY INTERPOLATED DATA

We perform empirical orthogonal function (EOF) analysis on twelve seasonal subsamples of the interpolated record. It is well known that the choice of sign for an EOF pattern is arbitrary. In order to avoid month to month sign shifts and allow easier interpretation of the results, we restrict the sign of the sum of EOF loadings in each season to be positive, and change the sign of the corresponding principal components (PC) accordingly. Figure 7 describes the seasonal march of the the first EOF loadings (Figure 7a), as well as the percentage of variance (Figure 7b), and the total variance of monthly anomalies (Figure 7c). Throughout the period of significant streamflow (December through July) the 1st EOF loadings at all sites are almost always positive (except for Ceyhan in April and Yarmouk in July). This identifies a common signal in all records, represented by the corresponding PC, and accounts for 70% of total variance during DJFM and 60% to 90% during AMJJ. Note that the DJFM period is characterized by a more stable EOF1 loading profile than the AMJJ period.

We use these SL diagrams to substantiate connections between the first PC (PC1) of ME streamflow, local precipitation and temperature, and large-scale (NAO) climatic parameters (Figure 8). Seasonally inhomogeneous persistence in streamflow and local climate is illustrated by comparison of Figures 8a–c. While decorrelation times of precipitation (less than one month) and air temperature (1–2 months) are seasonally uniform, streamflow undergoes a transition in November, showing no correlation with earlier months. Also significant are two similar but more moderate transitions in April and July, thereby providing additional justification for separating streamflow into winter-spring (DJFM) and spring-summer (AMJJ) groups. We do not study streamflow in the remaining period (August to November) as total discharge in this period is negligible.

Figure 9 suggests that different processes may dominate variability in streamflow during the DJFM and AMJJ periods, with large-scale NAO dynamics influencing winter-spring (DJFM) and local processes influencing spring-summer (AMJJ). Figure 9a represents the influence of monthly precipitation on PC1 of reconstructed ME streamflow over the period 1938-1984. The DJFM period of PC1 shows synchronous and lagged positive correlations with DJ precipitation, as seen by the downward sloping region and 0 to -2 lag in months. This may be suggestive of some type of storage mechanism linking February and March streamflow to DJ precipitation. The AMJJ period of PC1 shows significant or nearsignificant positive correlation with precipitation during the previous half year from December to May, strongest correlation (r = 0.4) is between June streamflow and May precipitation (Figure 9a). This lagged relationship with temperature identifies snowmelt as a major factor in regulating AMJJ streamflow. Figure 9b represents the influence of monthly temperature on PC1. The DJFM period of PC1 displays a zero-lag positive correlation with February and March (FM) temperature which can be interpreted as an increase in streamflow via snowmelt during years in which



Figure 7. (a) Monthly loadings of the 1st EOF of Middle Eastern streamflow, (b) percent variance explained by the 1st EOF and, (c) total monthly variance in the optimal interpolation (OI) analysis of normalized streamflow.

air temperature during the cold season is anomalously high. We suggest that DJFM streamflow is most strongly influenced by the NAO, as streamflow over this period shows a synchronous NAO connection (centered on January and February) for the NAO_{SLP} index (Figure 9c). A similar phenomenon is seen for the NAO_{SST} index whereby significant correlation extends over lagged periods from 2 months prior to 4 months later (Figure 9d). The AMJJ period of PC1 shows no significant NAO correlation further supporting a more local connection.

In order to more closely examine these interpretations in a temporal domain, we compute total DJFM and AMJJ streamflow of ME rivers, and then repeat our OI analysis and EOF calculation for each of these two seasons. Figure 10a compares PC1 of DJFM streamflow (which accounts for 81% of total variance) with the dom-



Figure 8. (a) Seasonal lag (SL) correlation of PC1 of monthly Middle East (ME) streamflow, (b) SL correlation of monthly ME precipitation, (c) SL correlation of monthly ME temperature.



Figure 9. (a) The influence of Middle East (ME) precipitation on PC1 of reconstructed monthly streamflow, (b) same as in (a) except the influence of ME temperature, (c) same as in (a) except the influence of the NAO_{SLP} index, (d) same as in (a) except the influence of the NAO_{SST} index.

inant climatic factors presented in Figure 9: DJ precipitation, FM air temperature, and the NAO. Table III presents the corresponding correlation coefficients as well as the correlations for individual OI interpolated records. PC1 is well represented by all records and is significantly correlated with all four identified climatic factors, DJ precipitation being the strongest influence.

Figure 10b compares PC1 of AMJJ (which accounts for 59% of the total AMJJ variance) with December through May precipitation and March through May air temperature. Table IV presents the corresponding correlation coefficients and shows that the agreement in the records of the individual rivers in this period



Figure 10. (a) NAO_{SLP} and NAO_{SST} correlated with winter-spring (DJFM) PC1 of reconstructed Middle East (ME) streamflow, ME precipitation (December to January; DJ), and air temperature (February to March; FM). (b) PC1 of spring-summer (AMJJ) reconstructed ME streamflow correlated with ME precipitation (December through May; DJFMAM) and ME temperature (March through May; MAM). Correlation values are listed in Tables III and IV.

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River	PC1	${\rm T}_{\rm JF}$	P _{DJ}	NAO _{SLP}	NAO _{SST}
Ceyhan	0.81	0.37	0.57	-0.36	-0.46
Euphrates	0.97	0.41	0.57	-0.47	-0.48
Karun	0.75	0.15	0.38	-0.44	-0.29
Tigris _{Mosul}	0.99	0.37	0.62	-0.49	-0.50
Tigris _{Baghdad}	0.98	0.36	0.59	-0.49	-0.48
Yarmouk	0.73	0.03	0.46	-0.43	-0.32
PC1	1.00	0.36	0.61	-0.50	-0.49

Table III Correlations for DJFM period

Table IV					
Correlations for AMJJ period					
River	PC1	T _{MAM}	P _{DJFMAM}		
Ceyhan	-0.08	-0.02	0.26		
Euphrates	0.96	-0.46	0.63		
Karun	0.90	-0.47	0.61		
Tigris _{Mosul}	0.91	-0.44	0.39		
Tigris _{Baghdad}	0.96	-0.51	0.54		
Yarmouk	0.05	-0.12	0.13		
PC1	1.00	-0.50	0.59		

is not uniform. The Ceyhan and Yarmouk contribute very little towards PC1, while the Euphrates, the Tigris, and PC1 return a significant correlation for both climatic factors.

6. Conclusions

Water allocation issues have been ongoing since the emergence of early Mesopotamian civilization along the alluvial lowlands of the Tigris and Euphrates Rivers roughly 6,000 years ago (Weiss, 1997; Cullen and deMenocal, 2000b; Algaze, 2001). Water scarcity remains a concern in the modern Middle East, where regional population is now increasing by 3.5% each year and water storage/irrigation practices consume at least 80% of available water supply. In this paper, we further establish the link between changes in Middle Eastern water supply associated with natural variations in the climate system and SST variations in the Atlantic Ocean and eastern Mediterranean Sea. We stress that if the NAO, influenced by increased greenhouse gases, continues its upward trend, then future amounts of December through March precipitation and streamflow can be expected to be lower. It is quite possible that, in addition to non-linearity and complexity of climate connections in this period, active river management influences river data during the spring and summer. Both local and large-scale climate connections to streamflow established here show a way for possible climatic modeling of ME streamflow for the period 1985–present when data are difficult to obtain.

6.1. HOLOCENE CLIMATE VARIABILITY

In addition to their relevance for water resource management in the Middle East, these results may also be pertinent to understanding Holocene paleoclimate fluctuations and the paleoenvironments of early human civilizations in Mesopotamia (Weiss et al., 1993). Recent paleoclimate data from Greenland and the North Atlantic demonstrate that the past 11,700 years of the Holocene were punctuated by numerous millennial-scale $(1, 500 \pm 500 \text{ years})$ cooling events (Bond et al., 1997). Although these Holocene cooling events were evidently much lower amplitude (perhaps 1–2 °C) than the well-known Dansgaard-Oeschger cycles of the previous glacial period, they seriously challenge the view that the Holocene was climatically stable (Blunier et al., 1995; O'Brien et al., 1995; Keigwin, 1996). The linkage between North Atlantic and Middle Eastern climate described here suggests that these Holocene cooling events in the North Atlantic should have coeval climate change signatures in the Middle East. As controls on Middle Eastern winter climate are therefore dominated by North Atlantic sea surface temperature, and new evidence points to the likelihood of centennial scale variability in Atlantic sector SST's, we propose that Middle East climate change during the Holocene, might also be viewed from within the framework of North Atlantic climate variability.

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