RELATIONSHIP BETWEEN THE NORTH ATLANTIC OSCILLATION, EURO-ASIAN CLIMATE ANOMALIES AND PACIFIC VARIABILITY

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The aim of this paper is to discuss the manifestations and generating mechanism of the interannual-to-interdecadal variability of the coupled ocean-atmosphere system in the North Atlantic Ocean, Equatorial and North subtropical Pacific Ocean resulting in climate change in the Euro-Asian region. It is argued that El Niño-Southern Oscillation and North Atlantic Oscillation is a complex interactive system. It manifests a definite pattern of Euro-Asian climate variability and variability of Euro-Asian rivers’ run off and impacts on the El Niño-monsoon interaction. This system is characterized by different governing mechanisms for quasi-biennial, interannual and interdecadal scales which are analyzed separately.

INTRODUCTION

El Niño-Southern Oscillation (ENSO), Pacific decadal oscillation (PDO) and North Atlantic Oscillation (NAO) are the coupled ocean-atmosphere phenomena believed to be responsible for much of the low-frequency climatic variability experienced in several parts of Europe and Asia. As recent events suggest, the health of the physical environment and the resources it supports, incidence and trends in severe weather conditions, e.g. storms, floods and droughts, are linked to a significant degree to those phenomena. There are also clear evidences of interaction between different climatic signals including the ENSO-monsoon and NAO-monsoon links. It follows that prediction of ENSO, PDO, monsoon variability and NAO extremes with a reasonable lead time can be an effective first step in securing human life and property (e.g. CLIVAR, 1995; Fraedrich and Muller, 1992; Glowienka-Hense, 1990; Hurrell, 1995; Polonsky, 2001; Rogers, 1997). That is why the study of mechanisms of relationship between ENSO, PDO, East Asian monsoon and NAO is extremely important. These mechanisms are discussed in the present paper using historical hydrometeorological data and NCEP re-analysis output.

GENERAL DESCRIPTION OF PHENOMENA AND PRINCIPAL GOAL

ENSO and PDO. Interannual variability in the global coupled ocean-atmosphere system induced by the processes in the Pacific Ocean is quite completely documented. It is well known that the magnitudes of the Southern Oscillation index (SOI, which is defined usually as the normalized difference of sea level pressure between Tahiti and Darwin) and of zonal temperature gradients in the equatorial Pacific increase before a typical ENSO event as a result of the enhanced intensity of the Pacific Walker cell. The warm pool in the western equatorial Pacific is widespread and deepens at that time. The typical ENSO begins in the early boreal spring. The first ENSO manifestation is the Walker cell weakening in the Pacific Ocean. This results in the lowered SOI magnitude that persists in boreal spring and summer. The rapid oceanic response occurs in the equatorial Pacific. The spread of warm surface water over the whole equatorial region is usually observed during at least half of an year. The positive feedback between the zonal gradient of the equatorial sea surface temperature (SST) and the intensity of the Walker cell is very important in the generation of coupled high-amplitude anomalies. During the mature ENSO stage the large-scale SST anomaly reaches 4–5°C. Then, negative feedback between the SST meridional gradient and the Hadley cell intensity causes the trade wind intensification (Bjerknes, 1969; Rasmusson and Carpenter, 1982; Rasmusson, 1989; Wyrtki, 1975).

A typical temporal interval between consequent ENSO events is of 3 to 5 years. At the same time, long-term time series of SOI and Equatorial Pacific SST show that there are significant quasi-biennial and ~6 yr ENSO fluctuations (Schneider and Sconwiese, 1989). Besides, the ENSO characteristics are modulated by interdecadal-to-multidecadal oscillations (Enfield and Mestas-Nunez, 1999). PDO is a principal regional player on the interdecadal scale. Its space pattern resembles the ENSO in the Equatorial Pacific. However, PDO involves the subtropical and mid-latitude processes (Trenberth and Hurrel, 1994; Zhang et al., 1997; Enfield and Mestas-Nunez, 1999). The most recent results (Picaut et al., 2004) summarize the mechanisms of influence of these processes upon the equatorial Pacific SST as follows:

• atmospheric bridge (“decadal variability of the wind generated at mid-latitudes extends to the tropics to affect ENSO”)
over the North Atlantic and Europe. The increase indicates the dominant zonal atmospheric circulation. Thompson and Wallace (1999) considered the NAO as manifestation of the large-scale atmospheric and oceanic fields in the North/Tropical Atlantic. At the same time, we would like to draw attention to the fact that the NAO (which will be discussed in the next section) is the most significant climatic signal on the quasi-decadal and multidecadal scales and it can affect the ENSO-monsoon interaction and, hence, the low-frequency variability of ENSO itself (Polonsky, 2001; Polonsky et al., 2004a, b).

**NAO.** The North Atlantic Oscillation causes the most climatically significant fluctuations in the atmosphere and ocean of the Northern Hemisphere (Barnston and Livezey, 1987; Hurrell, 1995, 1996; Machel et al., 1998; Marshall et al., 2001a, b; Polonsky, 2001; Polonsky et al., 2004a). The NAO manifests a quasi-simultaneous increase (decrease) in the atmospheric pressure in the Azores High (Icelandic Low). It accounts for the significant portion of total variability of the large-scale atmospheric and oceanic fields in the North Atlantic and over the surrounding continental regions. Walker and Bliss were the first who described this phenomenon (Walker and Bliss, 1932). The pressure gradient between the Azores High and the Icelandic Low determines the strength of mid-latitude westerly transport and, hence, characterizes the movement of relatively warm and humid air from the North Atlantic to Europe. Thompson and Wallace (1998) considered the NAO as manifestation of the large-scale variability of the angular momentum in the Northern Hemisphere, which was called the Arctic Oscillation (AO).

A measure of oscillation is the NAO index (NAOI), which is defined as the normalised difference between the atmospheric pressure anomalies over the Azores and Iceland (or modified Rossby index, which is equal to sea level pressure (SLP) difference between Azores High and Iceland Low, see (Polonsky and Sizov, 1991)). A NAOI increase in a positive NAO phase indicates the dominant zonal atmospheric circulation over the North Atlantic and Europe. The increase is accompanied by strengthened zonal winds within 50–70°N in the troposphere and by weakened blocking activity there (Glowienka-Hense, 1990; Werner and von Storch, 1992; Deser and Blackmon, 1993; Kozuchowski, 1993; Latif and Barnett, 1994; Polonsky, 1997; Rogers, 1997). This leads to the positive temperature anomalies over the most West, Central and North Europe (Werner and von Storch, 1992; Kozuchowski, 1993; Polonsky, 1997; Drevillon et al., 2001). At the same time as reported by Nesterov (1998), cyclonic activity along the North Atlantic storm track is enhanced. Results by McCabe et al. (2001) and Polonsky and Basharin (2002) confirm that. Other data, however, do not support the strong relationship between the intensity of North Atlantic cyclones and the NAO phase (Rogers, 1997; Serreze et al., 1997). These authors only noted that the number of North Atlantic cyclones increases during the NAOI rise period. The overwhelming majority of data suggest the cyclone pathways (and the atmospheric centres of action) in this period shift to the north-east as compared to the years with the intermediate NAOI magnitudes. The reverse tendency is observed during negative NAO phases.

Broad spectrum of NAO index is more close to white noise than SOI spectrum. However, the former is characterized by two significant peaks on the quasi-biennial and quasi-decadal scale (T = 2–3 and 6–8 years, QBO and QDO, respectively). There are also 60–80 yr variations, which are visible in the long-term instrumental data and extracted with reasonable statistical significance from the paleodata (Schneider and Senonwies, 1989; Marshall et al., 2001b; Polonsky and Semiletova, 2002; Polonsky et al., 2004a). Polonsky (1997) showed that NAO multidecadal variability is characterized by the displacement of Icelandic Low and Azores High to the south-west when NAOI is rising. In the other words, the interannual and multidecadal NAO-related tendencies of displacement of the North Atlantic centres of action are opposite to one another. This means that the NAO-ENSO links differ for the interannual and multidecadal scales. We are going to consider below these links and governing mechanisms.

**NAO-ENSO links.** Certainly, ENSO event is a phenomenon on a globe (Walker, 1924; Hastenrath, 1991; Harrison and Larkin, 1998; Alexander et al., 2002; Mitchell and Wallace, 2003). Walker (1924) and Walker and Bliss (1932) first defined the Southern Oscillation as a global phenomenon as follows: when pressure is high in the Pacific Ocean, it tends to be low in the Indian Ocean from Africa to Australia; these conditions are associated with low temperature in both areas. Trenberth and Shea (1987) showed that the SLP anomalies in Darwin correlate with the SLP anomalies in the subtropical North Atlantic and subtropical North Pacific (correlation coefficient is about -0.2 and -0.5, respectively) and the SLP.
anomalies in the high latitudes of the North Atlantic (correlation coefficient is about 0.2). Thus, ENSO should manifest in the North Atlantic in spite of rather small correlation.

Walker and Bliss (1932) first studied the NAO and SO teleconnections. They found that the connection between the SO and the NAO is very weak. This result was confirmed by the numerous studies (see e.g., Hastenrath, 1991; Schneider and Seneviratne, 1989). At the same time, Polonsky and Sizov (1991) have analyzed the NAO and SO teleconnections for the 20th century and showed that the seasonal cycle of the NAO tends to be stronger during ENSO events. Rogers (1984) and Polonsky and Sizov (1991) have documented the intensified zonal atmospheric circulation in the North Atlantic prior to and for the mature phase of typical El Niño event and its relaxation during the transient ENSO phase occurring in year “0” of El Niño event. They argued that the NAO and SO indices tend to vary in phase (accurate in year “0” of El Niño event. They argued that the NAO and SO indices tend to vary in phase (accurate in year “0” of El Niño event.

Another possible impact of the ENSO event on the NAO is due to the tropical Atlantic ↔ tropical Pacific interaction and further tropical ↔ extratropical link in the Atlantic Ocean. The ENSO-induced North and tropical Atlantic pattern of SLP/SST anomalies resembling the NAO tripole was found by Nath and Nath (2001). North tropical Atlantic response to the ENSO forcing through the atmospheric bridge was analyzed by Klein et al. (1999) and Dzhiganshin and Polonsky (2001). They found that the Equatorial Pacific anomalies lead by 3 to 6 months. Klein et al. (1999) argued that the ENSO-induced surface heat flux anomaly in the North tropical Atlantic accounts for the SST anomaly there, while Dzhiganshin and Polonsky (2001) paid attention to the advective anomalies of SST and heat content associated with the ENSO-induced trade wind changes. At the same time, Servain (personal communication, 2001) reported that the tropical Atlantic SST anomaly south off meteorological equator leads to SOI with an average time lag of 6 months and the two signals share 14% of total variance. The similar delay between SST anomaly in the Equatorial Atlantic and Equatorial Pacific was documented for the ENSO event of 1991–1993 by Polonsky (1994). He argued that this delay is a result of faster adjustment of the Equatorial Atlantic due to its smaller size.

Possible influence of the NAO on the ENSO is due to the changes in temperature and snow conditions over Euro-Asia in different NAO phases and associated ENSO-monsoon relationship (Bamzai, 2003; Bamzai and Shukla, 1999; Chang et al., 2001; Kumar et al., 1999). Gong and Ho (2003) found the signature of spring AO (or NAO) in the East Asian summer monsoon precipitation. According to their result, the correlation coefficient between May AO index and August Far East precipitation is about -0.45. So, East Asian monsoon in late summer tends to be weaker after strong NAO during preceding May.

Thus, there is some evidence of the both-sided atmospheric bridge between the North/Tropical Atlantic and Pacific Ocean. There is also multidecadal variability of ENSO-NAO links (Kinnpertz et al., 2003; Mitchell and Wallace, 2003). This could be the result of opposite interannual and multidecadal tendencies of displacement of the North Atlantic centres of action for different NAO phases (Polonsky, 1997). That is why we would like to discuss once more the mechanisms of NAO-ENSO interaction for different temporal scales using mostly the re-analysis data. This is a principal goal of the present study.

**DATA SETS AND PROCESSING PROCEDURE**

The following historical data sets have been used:

- Rossby/NAO/SO monthly indices in 1876–2000
- monthly SST/SLP data at 2° by 2° grid points in 1950–1997 from COADS
- following NCEP re-analysis data at regular 1.825° by 1.825° grid points since 1950 to 2001:
  - 1000 mbar pressure (in fact, SLP)
  - 2 m air temperature (in fact, surface air temperature – SAT)
  - total precipitation
  - latent (LE) and sensible (H) surface heat flux

Data sets at each grid point, SOI and NAOI were filtered using band-pass (2 to 5/10 yr and 5/10 to 20/30 yr) and low-pass (10 and 30 yr cut off) filters. Then, the correlation coefficient between each time series and filtered NAOI was calculated in each grid. Correlation coefficients between (H+LE) and NAO/SO indices have been calculated for the globe. Method of empirical orthogonal functions (EOF) was used to analyze the COADS data in the North Atlantic (0–70°N, 0–70°W) and NCEP re-analysis data sets on SLP, SAT and precipitation over Euro-Asian region (25–80°N, 10°W–150°E). Before the processing, monthly data at each grid point were both time averaged for consecutive two months (January–February, March–April and so on) and space averaged at the 5.475° by 5.475° grid points. Then, first five EOFs for the corresponding anomalies (defined relative to detrended climatic mean) were calculated, and the correlation between NAO/SO indices and time coefficients of different EOFs have been evaluated. To analyze the connection between Euro-Asian rivers’
discharge and NAOI/SOI fluctuations, the correlation matrix between monthly series has been calculated for different lags. Then, the long-term time series were band-passed filtered and correlation procedure has been applied once again.

RESULTS AND DISCUSSION

**Global NAO/SO/PDO manifestations.** Figures 1 and 2 clearly demonstrate the global features of NAO, SO and PDO. They display the zero-lag March correlation of NAO/SO indices and (H+LE) for interannual-to-quasi-decadal and interdecadal-to-multidecadal scales, respectively. March was chosen for the following reasons. On the one hand, this is a month when there is quite a strong NAO influence on the Euro-Asian hydrometeorological conditions including spring rivers’ run off. On the other hand, March is a month when the onset of typical El Nino occurs.

The most extended areas of significant correlation between interannual-to-quasi-decadal variations of NAOI and (H+LE) occur in the Euro-Asian region, North Atlantic and Northwestern subtropical Pacific (Figure 1a). The space patterns of zero-lag winter (January–February)

![Figure 1a](image_url)

**Figure 1a.** Significant (at a 5% confidence level) correlation between March NAO and (H+LE). NCEP re-analysis data for 1950–2001 after high-pass (10 yr) filtering have been used.
correlation of NAOI and (H+LE), as well as lagged correlation between winter NAOI and early spring (H+LE) resemble the zero-lag correlation pattern for March. At the same time, the April correlation patterns differ (Figures for January, February and April are not presented). That is why, when we analyze the spatial variability of bi-monthly fields we will mostly choose January–February and May–June periods for consideration.

On the interdecadal-to-multidecadal scale, only March (H+LE) in the Northwestern Pacific and Central Siberian region correlates significantly with the March NAO index (we say nothing about small areas of formally significant correlation in different parts of the globe, which look mostly like random NAO manifestations, see Figure 2a). This demonstrates that only the Asian and Northwestern Pacific regions are mostly influenced by the low-frequency NAO.

The most extended areas of significant zero-lag correlation between interannual-to-quasi-decadal variations of March SOI and (H+LE) occur in the Tropical Pacific and North/Tropical Atlantic (Figure 1b). North Atlantic correlation pattern resembles the well-known tripole associated with the NAO (Deser and

Figure 2. Significant (at a 5% confidence level) correlation between March NAO (a) / SO (b) indices and March (H+LE). NCEP re-analysis data for 1950–2001 after low-pass (10 yr) filtering and de-trending have been used. Positive correlations are dotted, while negative ones are shaded.
Blackmon, 1993; Marshall et al., 2001a, b; Polonsky et al., 2004a). Again, the space patterns of zero-lag winter correlation of SOI and (H+LE), as well as lagged correlation between winter SOI and early spring (H+LE) resemble the correlation pattern for March as opposed to April. This confirms that the winter ENSO signal impacts the NAO through the atmospheric bridge, which at first initiates the North/Tropical Atlantic response and, then, maintains the NAO (see also, “NAO-ENSO links” and “Euro-Asian EOF in winter to early summer” sections). It should also be noted that the most extended area of significant correlation of SOI with (H+LE) on the interdecadal-to-multidecadal scale occurs in the Northwestern subtropical Pacific (Figure 2b). This points out to potential modulation role of PDO for the East Asian monsoon-ENSO interactive system.

Certainly, the statistical significance of correlation between interdecadal-to-multidecadal signals is quite poor because of relative shortness of the analyzed data sets. Discussed correlation for low-pass filtered series demonstrates the tendency in interdecadal-to-multidecadal signal rather than the significant result. However, this is a remarkable demonstration.

**Euro-Asian variability associated with NAO and ENSO.** Now we are going to consider the leading winter–spring Euro-Asian EOF, which has been extracted from the detrended data sets. Linear trends of all analyzed parameters are significant over extended Euro-Asian regions. That is why first of all, we give short description of the linear trends in the Euro-Asian region.

**Linear trends.** Linear SAT trends for winter 1950–2001 are positive (negative) over the North Euro-Asia, Arctic and Far East (Mediterranean region, Central and South Asia) (Figure 3a). These trends are accompanied by the corresponding trends of precipitation. Precipitation became more abundant over the northwestern Europe and northeastern China region, while it weakened over Mediterranean region, Central Asia and Far East (Figure 3b). Numerous studies show that first of all, these trends are due to the NAO intensification in the 20th century, which was especially strong during 1960–1990s (Machel et al., 1998; Marshall et al., 2001b; Polonsky et al., 2004a). The linear SLP trends confirm that. They show that in high latitudes winter SLP was decreasing by 1 to 1.5 mbar per a decade, while to the south of ~50°N it was increasing by about 1 mbar per a decade during the second half of the 20th century (Figure 3c). As a result, the NAO intensified. This result is also confirmed by the COADS data. They show that winter SLP trends reached about 1 mbar per 10 years and -1.5 mbar per 10 years during 1950–1997 in the vicinity of Azores High and Icelandic Low, respectively.

Polonsky and Semiletova (2002) showed that century-scale SLP trends are accompanied by displacement of Azores High/Icelandic Low to the North-West/South-West. Using historical data since 1890 till 1990, they found that linear trend of winter Azores High/Icelandic Low shift was about 60 km/75 km per 10 years. Thus, the distance between the Atlantic centres of action decreased. This is just as important for the Euro-Asian climate tendencies as the intensification of zonal circulation over the Atlantic Ocean due to rising of SLP difference between centres of action (see also, “North Atlantic QDO and interdecadal-to-multidecadal modes” section).

The space pattern of SLP linear trends in the Euro-Asian region for March–April is close to the January–February pattern. However, the magnitudes of spring trends are typically 1.5–2 times as little as winter ones except the Far East region, where the spring positive trend exceeds essentially the winter trend. These tendencies become even more distinctive in May–June and July–August. Maximum SLP linear trends (exceeding 2.0 mbar per decade) occur over the Japan sea region. It should be emphasized that the spring–summer SLP trends over the Far East are not a direct result of NAO intensification, which is quite weak in March–April and absent in May–June. Space pattern of spring–summer SAT/precipitation trends confirms that. For instance, May–June SAT trends are characterized by significant negative magnitudes over Northwestern Europe and more complex structure in the Central Asia and Far East than in winter. At the same time, the most prominent feature of May–June precipitation trends is their maximum negative magnitude in the Northeastern China region (Figures for spring and summer are not presented). So, the spring–summer trends in Asian hydrometeorological fields are not significantly related with the NAO trends in that time as opposed to winter.

**Euro-Asian EOF in winter to early summer.** NAO in winter accounts for the significant proportion of interannual-to-multidecadal fluctuations of SAT, precipitation and (especially) SLP over the most Euro-Asia in winter and spring. Actually, NAOI correlates significantly with the leading winter–spring EOF of Euro-Asian SAT, precipitation and SLP and this EOF is responsible for the large proportion of total variance (Figures 4–6). Maximum zero-lag correlation occurs for the January–February leading SLP EOF and NAOI. It reaches 0.6 (Figure 5c). Space pattern of leading EOF shows that in winter NAO impacts mostly the European precipitation, North Euro-Asian SAT and Arctic SLP (Figure 4). Spring–early summer and winter patterns of leading SLP/SAT EOF are similar for the most Euro-Asia (cf., Figures 4a, c and 6a). However, there is a following remarkable feature in the spring SLP pattern. Zonal SLP gradient (which, roughly speaking, characterizes the meridional wind) is opposite in the Central Asia and Far East regions. Significant concurrent correlation between May–June NAO/SO indices and leading SLP mode shows that
strong NAO/SO phase is accompanied by more intense early summer Far East monsoon (as opposed to southern Asian region). This is a result of enhanced temperature contrast between Northeastern Asia and Northwestern subtropical Pacific during strong NAO phase (see below). Spring–summer pattern of leading precipitation EOF is characterized by the additional area of large negative magnitudes in the northeastern China region. This EOF is negatively correlated with the winter NAO. This means that positive winter NAO phase is accompanied by more abundant early summer precipitation there (and, hence, more intense East Asian summer monsoon). Other features of leading precipitation EOF are close for winter and spring–early summer patterns (Figures for spring–early summer SAT and precipitation are not displayed). It should be emphasized that the sign of correlation between spring NAOI and leading EOF of East Asian precipitation in spring–early summer is opposite to the sign of correlation received by Gong and Ho (2003) for May NAO index and August Far East precipitation.

![Figure 3](image-url)

**Figure 3.** Spatial distribution of linear SAT (Celsius degrees per 100 years, precipitation (mm/day per 100 years) and SLP (mbar per 100 years) trends (a, b and c, respectively). NCEP re-analysis data for January–February of 1950–2001 were used. Areas of significant (at a 5% confidence level) trend are dotted. Positive contours are shown by thick lines.
Strong phase of NAO (large positive magnitudes of the NAO index) in winter is accompanied by warming of the most Euro-Asia to the north of 40–45°N and its cooling to the south of about 40°N in winter and early spring (Figures 4a, 5a). The lead-lag correlation coefficient is at maximum between January–February NAO index and March–April leading EOF of Euro-Asian SAT. It reaches 0.62. Such correlation is a result of intensified/weakened zonal circulation during strong/weak NAO phase in winter. Thus, the leading winter–spring Euro-Asian EOF for SLP, SAT and precipitation can be considered as a manifestation of winter NAO mode.

It should be noted that the signs of SAT/precipitation/SLP fluctuations are opposite in the Mediterranean region and high latitudes of Euro-Asia (Figures 4, 6a). This is also true for the correlation between NAOI and surface heat flux for the interannual-to-quasi-decadal fluctuations (Figure 1a). The above result is mostly the manifestation of NAO-related shift of the Atlantic centres of atmospheric action (and storm-tracks). Because of differences in such a shift for interannual and interdecadal-to-multidecadal variations, the correlation between NAOI and SAT, SLP and (in part) precipitation tends to increase after removing the interdecadal-to-

Figure 4. Space leading EOF for SAT (a), precipitation (b) and SLP (c). NCEP re-analysis data for January–February of 1950–2001 were used. The explained variance (%) of corresponding fields is shown in the lower right corners.
multidecadal variations (Figures 5, 6b). This confirms the result by Krishnamurthy and Goswami (2000).

Results of study of rivers’ run off variability confirm the above conclusions. In fact, for the most European rivers (except the northern Europe) the negative correlation between winter NAOI and spring run-off is typical, while spring run-off of North Dvina, Ob and Yenisei correlates positively with the winter NAOI. It should be emphasized, the coefficient of correlation between January NAO index and April discharge of North Dvina was persistently increasing from 0.25 in 1881–1985 up to 0.6 in 1966–1985 as a result of long-term NAO trend discussed above. As a result of NAO intensification and shift of Azores High and Icelandic Low, the NAO impact on the Asian rivers’ run-off became stronger in the last 30 yr period of the 20th century and coefficient of correlation between interannual variability of spring Ob/Yenisei run-off and winter NAO index reached 0.56. All these assessments are significant at a 99% confidence level.

An important role of NAO-related land-ocean interaction manifests also as a significant positive/negative correlation between spring NAO index and spring surface heat flux in the northwestern subtropical Pacific/northeastern Euro-Asia (Figure 1a). In fact, temperature contrast in Eastern Asia and Western Pacific and associated monsoon intensity depend significantly on the surface heat flux anomalies in the northwestern subtropical Pacific because the variance of monthly heat flux in the region of our study is at maximum just there. This exceeds the heat flux variance over the Euro-Asia by about one order. Significant proportion of this variance is due to NAO. Enhanced spring turbulent heat flux in the northwestern subtropical Pacific during strong NAO phase leads to the ocean surface cooling in spring–early summer and, hence, intensifies the land-ocean temperature contrast in the Far East region for the early summer monsoon period.

Thus, winter–spring interannual NAO intensification leads to warmer/cooler spring–early summer conditions in the North/South Euro-Asia, cooler early summer conditions in the northwestern subtropical Pacific and, hence, earlier and more intense summer East Asian monsoon. As a result, the La Niña conditions or El Niño relaxation in the Equatorial Pacific tend to develop after that. On the other hand, the winter–spring NAO weakening is favorable for the El Niño conditions next spring–early summer. This is one branch of NAO-ENSO interannual interactive system.

The other branch was described by Polonsky and Sizov (1991) as follows. The Pacific ENSO cycle is accompanied by weakening/strengthening of NAO during transient/mature ENSO phase due to the Hadley cell variations. This manifests itself as weakening/strengthening of northwest trade wind and

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**Figure 5.** Lead-lag correlations between bi-monthly SO(1)/NAO(2) indices and temporal coefficients of leading EOF for SAT (a), precipitation (b) and SLP (c) in January–February. Dashed curves show the same coefficients but after the high-pass (10 yr cut off) filtering. Horizontal dashed lines specify a 5% confidence level. Spatial patterns of associated leading EOFs are shown in Figure 4.
mid-latitude westerly in the North Atlantic during summer and early autumn/winter. NAO intensification during mature ENSO phase (typically since September–October), leads to the enhanced early winter temperature over the most northern and central Euro-Asia and reduced temperature in the southern Asia. As a result, the anomalous spring monsoon occurs in the northwestern Pacific-East Asian region and this impacts the ENSO retreat in that time.

Semi-annual NAO fluctuations and QBO. Our results permit to specify the above scheme of NAO-ENSO link. In fact, Figures 5 and 6b show that described tendencies are significant only for the restricted period between late autumn and late spring. At the same time, they show that the NAO-induced winter Euro-Asian anomalies are accompanied by the reverse spring NAO phase. This maintains the Pacific ENSO-related anomalies in that time and leads to semi-annual NAO fluctuations.

The likely reason for visible semi-annual variability of correlation between NAOI and time coefficient of leading EOF is due to interaction between Atlantic Ocean and surrounding continents (Polonsky and Sizov, 1991). In this context, a delayed response of the Icelandic Low in the North Atlantic to the variations of Pacific Aleutian Low and associated changes of hydrometeorological conditions over North America and North Atlantic Ocean may be also important (Moron and Gouirand, 2003).

It should be emphasized that the entire cycle of the described NAO-ENSO interaction takes just about two years (Figure 5). Additional NAO-induced spring heat flux from the ocean to the atmosphere in the northwestern subtropical Pacific also causes delayed monsoon-ENSO response to NAO and accounts in part for quasi-biennial SOI fluctuations. This means that significant quasi-biennial signal in the NAO and SO indices can stem from an interaction between these two global atmospheric modes involving the ocean upper layers in the North/Tropical Atlantic and northwestern subtropical Pacific. As the most recent results show, this signal may be also affected by the heat content variability in the Indian Ocean and resulting SST anomalies over multiple seasons (Meehl et al., 2004). However, the relative importance of this mechanism, at least in generating the NAO quasi-biennial signal, is unclear.

North Atlantic QDO and interdecadal-to-multidecadal modes. Recently published results show that inherent Atlantic quasi-decadal scale variations can be responsible for the 6–8 yr peak in the NAO index spectrum (Latif and Barnett, 1994; Polonsky, 1998; Marshall et al., 2001a, b; Dzhiganshin and Polonsky, 2003). Latif and Barnett (1994) argued that decadal-scale coupled variability in the North Pacific and Atlantic is controlled by subtropical gyre adjustment depending on the phase speed of the first baroclinic Rossby mode. Dzhiganshin and Polonsky (2003) showed that the advection of SST and heat content anomalies by
mean large-scale currents and the following impact of these anomalies on the atmospheric circulation is more important (see also, Figure 7). This subject is under intense discussions (see e.g., Marshall et al., 2001b; Polonsky et al., 2004a). However, anyway it is clear that the Atlantic and Pacific Oceans have different inherent temporal scales due to difference in their geometric sizes, physical and geographical features. NAO is characterized by the quasi-decadal scale, while PDO is the longer-term (~bi-decadal) signal. This means that the winter anomalies of NAO (due to inherent Atlantic quasi-decadal variations) before the ENSO onset and associated changes in the Asian hydrometeorological conditions may be one of the provocative factors for the ENSO event and just NAO may be responsible for ~6 yr peak in the SOI spectrum. At the same time Table 1 shows that decadal-scale variability of Euro-Asian rivers’ run-off depends not only on quasi-decadal NAO fluctuations, but PDO as well. The joint influence of decadal-scale NAO and PDO can explain about 50% of total decadal-scale variability of the most Euro-Asian rivers’ run-off in spring (if one considers them as independent signals, see Polonsky et al., 2004a).

Figure 7. Typical NAO index/SST (solid curves) and NAO index/heat content (0–200 m) (dotted) correlation functions for 8 regions of the North Atlantic specified at the right panel. Sign "+" means that NAO index leads. 10% confidence level is marked by the horizontal dashed lines at the left panel. The average surface circulation by Fratantoni (2001) is shown at the right panel. Star/crest shows the West African upwelling/Guinea cupola region. Historical data since 1950 till 1998 have been used (after Dzhiganshin and Polonsky, 2003).

Table 1

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<th>River</th>
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<td>Dniester</td>
<td>-0.60</td>
<td>3</td>
</tr>
<tr>
<td>Danube</td>
<td>-0.77</td>
<td>3</td>
</tr>
<tr>
<td>Garonne</td>
<td>-0.72</td>
<td>2</td>
</tr>
<tr>
<td>Luara</td>
<td>-0.72</td>
<td>2</td>
</tr>
<tr>
<td>Ob</td>
<td>0.53</td>
<td>2–3</td>
</tr>
<tr>
<td>N. Dvina</td>
<td>0.41</td>
<td>3</td>
</tr>
</tbody>
</table>

Note: The data sets have been filtered by a band-passed (5–20 yr) Parzen filter to extract a decadal-scale signal. Significant (5% confidence level) correlation coefficients are bold.
As mentioned above, NAO is characterized by century-scale trend and interdecadal-to-multidecadal variability. Their magnitude is at maximum in winter (see also, Enfield and Mestas-Nunez, 1999; Marshall et al., 2001b; Polonsky, 2001; Polonsky et al., 2004a). Interdecadal-to-multidecadal variability of SLP difference between the North Atlantic centres of action and distance between them are in phase as opposed to the century-scale trends discussed above (cf., Figures 8 and 9). Thus, the concurrent interannual-to-multidecadal variations of SLP and changes in location of centres of action reduce their joint influence on meridional SLP gradient and intensity of zonal circulation as a result of mutual compensation (as opposed to the century-scale trends). This means that interannual-to-multidecadal changes in the location of Azores High and Icelandic Low in winter and associated variations of hydro-meteorological conditions in Euro-Asian/Pacific Ocean region are the principal manifestations of low-frequency North Atlantic variability.

The signs of correlation between NAO index and surface heat flux in the northwestern subtropical Pacific are opposite on the interannual-to-quasi-decadal and interdecadal-to-multidecadal scales (cf., Figures 1a, 2a). On the interdecadal-to-multidecadal scale this correlation is at maximum for the entire studied region. This means, the NAO-related Pacific SST change may be a crucial factor regulating low-frequency temperature contrast between Euro-Asia and Pacific Ocean.

The origin of multidecadal NAO variations is likely due to the change of global thermohaline circulation generated in the realm of North Atlantic deep water sinking (Chen and Ghil, 1995; Delworth and Dixon, 2000; Polonsky, 2001). Thus, the slow (multidecadal) variations of NAO-related thermohaline circulation can impact the ENSO event not only directly (through the change of the basic state of equatorial ocean), but also indirectly (through the change of the temperature Asia-Pacific contrasts and following ENSO-monsoon interaction).

Figure 8. Interannual variability of modified yearly Rossby index (bold curve, 1), distance between Icelandic Low and Azores High along the longitude circle (thin curve, 2) and linear trends.
CONCLUSIONS

NAO-ENSO-PDO is a complex interactive system. It manifests itself as certain patterns of Euro-Asian SAT/precipitation/SLP leading EOF and influences variability of the Euro-Asian rivers’ run-off and the ENSO-monsoon interaction. This system is characterized by different conditions and governing mechanisms for interannual, quasi-decadal and interdecadal-to-multidecadal scales. QBO stems from inherent interaction in this system, while quasi-decadal and bi-decadal modes are likely due to Atlantic and Pacific inherent fluctuations, respectively. Multidecadal variations are likely due to the change of global thermohaline circulation generated in the realm of North Atlantic deep water sinking. In particular, this variability manifests a change in the location of Azores High and Icelandic Low and associated variations of hydrometeorological conditions in Euro-Asian region and Pacific Ocean and impacts the East Asian monsoon. So, multidecadal variations of NAO-related thermohaline circulation can impact the ENSO event not only directly (through the change of the basic state of equatorial ocean), but also indirectly (through the change of the temperature Asia-Pacific contrasts and following ENSO-monsoon interaction). Century-scale NAO trend and superimposed multidecadal variability causes the enhanced influence of geophysical processes in the North Atlantic coupled system (especially in winter) on the Euro-Asian hydrometeorological conditions including the spring river run-off during the last ~40 years.

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