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The Arctic Oscillation and Its Impact on Temperature and Precipitation in Northern Eurasia in the 20th Century

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Abstract—Presented is the review of the modern knowledge of the Arctic Oscillation (AO). Demonstrated is the relation of air temperature and precipitation in Northern Eurasia to this dominant type of wintertime atmospheric variability at northern extratropical latitudes. It is demonstrated that AO is a result of the coupling between the troposphere and stratosphere. The attention is paid to the long-range forecasting of AO index and to the factors complicating the forecasting. Given are the new results of the authors' research. Used is the wintertime AO index computed by the authors from the 20th Century Reanalysis dataset. The high- and low-frequency components of AO index variability and the periods of statistically significant trends are analyzed using the 112-year series (1901–2012). Demonstrated is the key impact of wintertime AO phase on the anomalies of air temperature and precipitation in Northern Eurasia at the time scale of years and decades. This impact is manifested in the northern part of Northern Eurasia in the prevalence of warmer and wetter winters at the positive AO phase and of colder and drier winters at the negative AO phase. The precipitation anomalies of opposite sign prevail in the southern part of Northern Eurasia. It is demonstrated that the winter AO phase affects the terms of the springtime air temperature transition to positive values.

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1. INTRODUCTION

The term "Arctic Oscillation" (AO) was proposed by D.W.J. Thompson and J.M. Wallace [48] for describing the form (mode, state) of variability of atmospheric processes that is corresponded by the variability pattern of air pressure and geopotential height characterized by the anomalies of the same sign in the polar region and by the anomalies of the opposite sign in the zone of 40 –50 N. This variability pattern is formalized as the leading empirical orthogonal function (EOF) of the anomalies of pressure and geopotential height northward of 20 N that makes up about 20% of the total variance of the series of pressure and geopotential fields. The positive AO phase is characterized by the negative anomalies of pressure and geopotential in the polar region and by their positive anomalies in the zone of 40 –50 N. If the AO phase is negative, the signs of anomalies are opposite.

The variability pattern typical of AO for the first time had been detected by E.N. Lorenz in 1951 [39] and had been considered afterwards in a number of studies [4, 36, 51, 52, 54] long before paper [48] was published. However, D.W.J. Thompson and J.M. Wallace revealed for the first time that AO is the dominant type of variability of wintertime extratropical atmosphere in the Northern Hemisphere that has the equivalent-barotropic structure at the seasonal scale of averaging and mainly defines atmospheric conditions from the Earth surface to the middle stratosphere. The authors of [49, 50] demonstrated that this variability pattern manifested in the fields of wind, air temperature, snow and ice characteristics, and many other meteorological parameters, mainly defined the trends in meteorological parameters at the middle and high latitudes of the Northern Hemisphere which were observed in the last decades of the 20th century.

D.W.J. Thompson and J.M. Wallace [48] defined the basic pattern of AO as the first EOF of monthly mean anomalies of sea-level air pressure field northward of 20 N for the period of 1979–1995 in winter months from November to April inclusive. The AO index was defined as the projection of monthly mean pressure fields to this pattern with the subsequent normalization. The series of this index for the period of 1899–2002 can be found at the website of University of Washington (jisao.washington.edu/ao/). In their next paper [49] D.W.J. Thompson and J.M. Wallace replaced the sea-level pressure fields by the fields of H_{1000} northward of 20 N for all months of the year and defined the AO index as the projection of monthly mean anomaly of H_{1000} to this base. Nowadays the AO index is computed in the Climate Prediction Center (CPC) with the basic period of 1950–2000. The AO indices from 1950 till now are published at the CPC website (www.cpc.noaa.gov/products/precip/CWlink/daily_ao_index/monthly.ao.index.b50.current.ascii.table).

The physical nature of the Arctic Oscillation has become a subject of numerous studies in the recent decade. Currently the scientific community shares a common conception of AO. It is understood as the result, to a significant degree, of the coupling between the troposphere and stratosphere. The planetary Rossby waves with the wave numbers 1 and 2 (wintertime climatic waves caused by contrasts between sea surface temperature and land temperature and by terrain inhomogeneity) propagate from the troposphere to the stratosphere; there they interact with the circumpolar vortex, disturb and weaken it to a certain degree [35, 37, 38, 46, 49]. The positive AO phase is associated with the positive anomaly of the circumpolar vortex intensity that is favored by increase in the temperature gradient between the sun-heated and shadowed parts of the atmosphere, by the intensification of the mean zonal flow, and by the smaller amplitude of planetary waves. This makes the circumpolar vortex less sensitive to the impact of waves because their refraction towards the equator takes place [46]. The negative AO phase is associated with the weakening of the circumpolar vortex that is caused by the lower temperature gradient between the sunheated and shadowed parts of the atmosphere, by the weakening of the mean zonal flow, and by increase in the amplitude of planetary waves. This intensifies meridional circulation and increases the probability of the collapse of waves that is accompanied by the sudden stratospheric warming and by the circumpolar vortex destruction [22, 38, 55]. Stratospheric variations of AO are transferred to the troposphere [16–18, 28, 29]. The propagation of Rossby waves in the stratosphere is possible only in the western flow [20]; therefore, AO is a wintertime phenomenon. According to the definition given by D.W.J. Thompson and J.M. Wallace [49], the typical active season of AO is January–March.

At present, intensive studies have been carried out in the area of long-range forecasting of winter AO index [21, 26, 43, 44]. The most successful model forecasts [12, 14, 47] describe up to 40% of the variance of the AO index interannual variability. On the one hand, this is quite high percentage; on the other hand, it is not high enough to make these forecasts meet the requirements of practical use.

The long-range forecasting of the winter AO index is complicated by the multi-stage nature of the process that defines the AO polarity and consists of at least two stages. At the first stage the generation of the anomalies of long planetary waves takes place in the troposphere. At the second stage the interaction of waves with the zonal flow in the stratosphere and the signal translation to the troposphere occur.

The intensification of long tropospheric waves is observed in winters with El Niño, and the weakening takes place in winters with La Niña [25]. However, the correlation between the wintertime indices of El Niño–Southern Oscillation (ENSO) and AO is statistically insignificant. It is demonstrated in [34] that the statistical insignificance of this relation is caused by its instability. El Niño affects greatly the AO polarity only in the case of the weak circumpolar vortex when different AO phases are almost equiprobable. In the case of the strong circumpolar vortex, El Niño does not affect AO and the positive AO phase is observed, this is manifested in the weakness of the linear relation.

According to the data of [31], along with El Niño the intensification of long climatic waves is favored by the advection of cold Arctic air to East Asia along the eastern periphery of the zone of positive anomalies of pressure and geopotential located over the Taymyr Peninsula. This anomaly induces warm advection to the polar region resulting in geopotential increase and the circumpolar vortex weakening. The intensification of long waves and the circumpolar vortex weakening result in the negative polarity of AO [25, 31, 37].

It should be noted that the AO pattern over the North Atlantic is similar to the pattern of the North Atlantic Oscillation (NAO). In spite of the fact that NAO is, to a significant degree, a result of the interaction between the North Atlantic and tropospheric circulation [40, 41] and AO is the stratosphere-troposphere phenomenon [18, 49], NAO is often represented as the regional North Atlantic display of AO [6, 53]. The physical substantiation of closeness of NAO and AO indices is presented in [15]. However, the coefficient of correlation between the mean wintertime AO and NAO indices is about 0.8, hence, these are two different indices. The difference between these indices is clearly manifested if one analyzes the impact of AO and NAO on meteorological parameters in Northern Eurasia.

2. DATA AND METHODS

The present paper uses the grid data on sea-level air pressure, air temperature, and precipitation for the period of 1901–2012 (112 years) from the 20th Century Reanalysis dataset, v.2 [23, 24, 56] published at the Earth System Research Laboratory website (www.esrl.noaa.gov). To provide the synoptic interpretation of AO, the data were used on the position and intensity of cyclonic centers at the middle and high latitudes of the Northern Hemisphere in 1958–2008 [45] from the website of the US National Snow and Ice Data Center (nsidc.org). The AO index was computed in accordance with the definition by D.W.J. Thompson and J.M. Wallace [48]. The base was computed as the first EOF of the fields of monthly (December–March) mean anomalies of sea-level pressure northward of 20 N in 1950–2000. The monthly mean values of the index were computed as the projections of the fields of monthly mean anomalies of pressure in 1901–2012 to this base. The average wintertime (December–March) AO index was computed as the mean value of non-normalized monthly mean indices for this season. After that the normalization of the index with the base period of 1950–2000 was carried out. The winter years in the graphs and text are given for January, for example, the year 2010 corresponds to the winter 2009/10.

The present research is mainly based on the analysis of linear relations. The interannual (high-frequency) and decadal (low-frequency) variability is clearly pronounced in the data series with the duration of more than a century. The uncertainty in the correlation coefficient associated with the low-frequency variability is not resolved by the simple removal of a linear trend. Therefore, the series of the AO index and the series of air temperature and precipitation at the grid points were divided into low-frequency (moving 9-year average) and high-frequency (the difference between the values of the initial series and of the corresponding 9-year average) components characterizing decadal and interannual variability, respectively. The length of the analyzed series decreased from 112 to 104 years (1905–2008). The correlation analysis was carried out separately for the high- and low-frequency components of variability of AO index and fields of air temperature and precipitation. The significance of correlation coefficients and linear trends was estimated using the *t*-test taking into account the effective number of degrees of freedom [19]. The correlation coefficients are significant at the level of 5% if their absolute value exceeds 0.2 for the high-frequency component and 0.5 for the low-frequency component.

The analysis of composites was also used. The values of AO index depend on the base period of the analysis of principal components and normalization. To avoid this dependence, the composites were constructed based on the upper and lower quintiles of the AO index (each equals 22 years of the total 112-year series, 1901–2012). The statistical significance of difference between the composites was assessed using the Monte Carlo method [57].

3. ANALYSIS AND RESULTS

3.1. The Arctic Oscillation

The positive polarity of AO (Fig. 1a) is associated with the negative anomalies of sea-level pressure and geopotential in the polar region and with the positive anomalies of these parameters in the zone of 40 –50 N. In the field of sea-level pressure anomalies, this zone is divided into two vast areas of positive anomalies: Atlantic and Pacific. The positive anomaly of zonal wind in the zone northward of 45 N and the negative anomaly southward of 45 N correspond to this pattern of the field of pressure and geopotential anomalies. The Hadley and Ferrel cells intensify and the polar cell weakens due to the air pressure drop in the polar region. In the case of AO negative polarity, the anomalies have the opposite sign. It should be noted that the AO pattern becomes more zonally symmetric as height increases and the zone of the positive anomalies of geopotential in the middle stratosphere has almost no breaks.

The intensification of the westerlies at middle and high latitudes in the Atlantic-European sector results in the intensification of the advection of warm and humid Atlantic air to Northern Eurasia and exerts significant influence on wintertime temperature and precipitation (see below).

The synoptic interpretation of AO is obvious from the analysis of anomalies of the number of cyclonic centers at the middle and high latitudes of the Northern Hemisphere (Fig. 1b). In winters with the positive polarity of AO in the area of the Icelandic low the intensification of cyclogenesis takes place and developed cyclones move to the east-northeast mainly through the northern part of the North Atlantic and further along the northern coast of Eurasia. The number decreases of polar-frontal cyclones moving from the east-

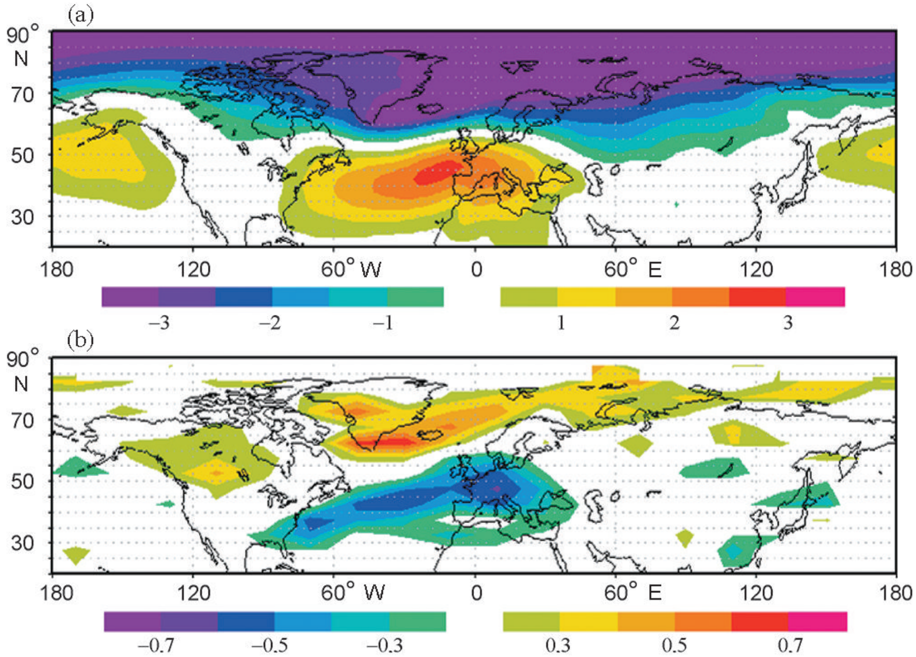

Fig. 1. (a) Leading empirical orthogonal function (EOF_1) of the mean monthly sea level pressure anomalies (December–March) shown as a regression of sea level pressure fields on the AO index. (b) Rank correlations between wintertime numbers of cyclonic centers with sea level pressure below 1010 hPa in 5 latitude 20 longitude trapezoids and the AO index (1958–2008).

ern coast of the USA to the east-northeast through the North Atlantic to Central and Southern Europe. In winters with the negative polarity of AO, the spatial distribution of anomalies in the number of cyclonic centers is opposite.

In terms of climatologic fronts [13], the positive AO phase is associated with the intensification of the Arctic front and with the weakening of the polar front; on the contrary, the negative AO phase is associated with the weakening of the Arctic front and with the intensification of the polar front (hereinafter, “intensification” and “weakening” should be considered as anomalies). This concept explains well the interrelation between temperature and precipitation, on the one hand, and the AO index, on the other hand. This interrelation is analyzed below.

Both the interannual variability and the periods of long-term variability are strongly pronounced in the long-term variations of the average wintertime (December–March) AO index (Fig. 2a). The periods of long-term variability include the following: the decrease in the index from the early 20th century to the early 1940s, its relative stabilization in the 1940s–1960s, the dramatic increase in the 1970s–1980s with the peak (+2.49 SD) in winter 1988/89, and the subsequent decrease in the 1990s–2000s with minimum (–2.71 SD) in winter 2009/10 (SD is the standard deviation). The mentioned values correspond to the absolute maximum and minimum of AO index according to the observational data since the beginning of the 20th century.

Figure 2b presents the moving estimates of 20-year trends of AO index. The points in the graph correspond to the central years of 20-year periods. Negative trends in AO prevailed in the 1900s–1950s. The minimum negative trend equal to –1.79 SD/20years (the point corresponding to the year 1930) was observed in the 1920s–1930s. The increase in the AO index with the maximum of 1.90 SD/20 years in 1987 prevailed from the 1960s to the 1990s. After that the dramatic transition to the descending trends was observed with the minimum value of –1.56 SD/20 years in 2000. All above extremes are statistically significant at the level of 5%.

One should notice the following feature of the long-term variations of trends that is manifested in the dramatic passage from the ascending trend in the 1980s–1990s to the descending trend in the 1990s–2000s. One of the possible reasons for such dramatic change is associated with the significant warming observed in the Arctic in the last third of the 20th century [8, 30]. The warming was accompanied by geopotential increase in the polar region and by the circumpolar vortex weakening that made it more sensitive to the impact of planetary waves [34, 36] and lead to dramatic decrease in the AO index.

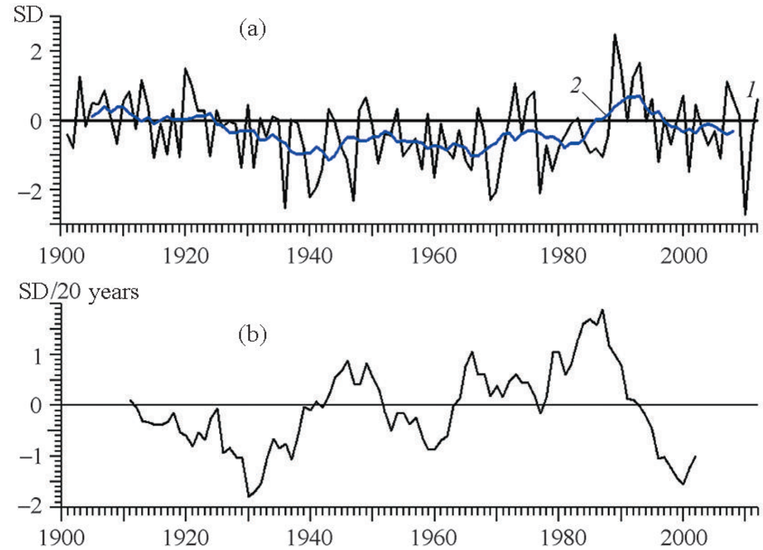

Fig. 2. (a) Time series of the normalized wintertime (December–March) AO index (*1*), and its 9-year running mean (*2*). (b) Running estimates of 20-year linear trends of the wintertime (December–March) AO index. Trends exceeding 1.4 (the absolute value) SD/20 years are significant at the 5% level.

3.2. Synchronous Relations between Wintertime Values of Air Temperature and Precipitation and the Arctic Oscillation

Air temperature. The statistically significant synchronous relations between the wintertime air temperature in Northern Eurasia and the AO index are considered in [48–50] and other papers. In [9–11, 32, 33, 42] the effects of the winter polarity of AO on air temperature anomalies in different regions of the Northern Hemisphere in next seasons were demonstrated along with synchronous relations. However, the authors of the above papers came to nothing more than the analysis of the situation in the second half of the 20th century. The given paper considers initial series of 112 years (1901–2012) that enabled analyzing interrelation between air temperature and total precipitation and the AO index at different time scales with the insignificant loss of the series length.

Figure 3a presents the map of coefficients of correlation between the high-frequency components of variability (less than nine years) of average winter values of air temperature and the AO index computed for the period of 1905–2008. The zone of positive statistically significant correlations covers almost the whole territory of Northern Eurasia. As a whole, correlation coefficients exceed 0.5 for the huge zone from the British Isles to the Sea of Okhotsk and exceed 0.7 for Scandinavia, the most part of Eastern European Plain, and some regions in East Siberia. On the vast territory of Northern Eurasia, 25–50% of the variance of average wintertime air temperature associated with its interannual variability is related to AO in the 20th century. The positive AO phase is associated with the positive anomaly of air temperature almost on the whole territory of Northern Eurasia. The negative temperature anomaly is observed if the AO phase is negative.

The field of air temperature anomalies associated with the AO phase is a quadrupole that agrees well with the circulation pattern of AO presented in Fig. 1. In the case of the positive AO phase, the Icelandic low intensifies, and cold advection from the Canadian Arctic Archipelago intensifies at its rear. This results in the negative anomaly of temperature in the northeast of North America and in Greenland. The advection of the warm sea air to Northern Eurasia where the positive anomalies of air temperature are observed, intensifies in the warm sectors of the cyclones moving from the Icelandic low to the east-northeast along the Arctic front. The warm sea air flows to the southeast of North America along the southern periphery of the Azores high ridge. At the same time, the intensification of cold air advection from Central Asia results in the negative anomaly of temperature in the Eastern Mediterranean and Northern Africa. The increase in the positive anomalies of pressure in the Pacific region results in the positive anomaly of temperature in Eastern Asia and in the negative anomaly of temperature in Alaska and in the coastal areas of North America. The distribution of anomalies is opposite in the case of the negative AO phase. It should be especially noted that the weakening of the Arctic front and the intensification of the polar front take place at the

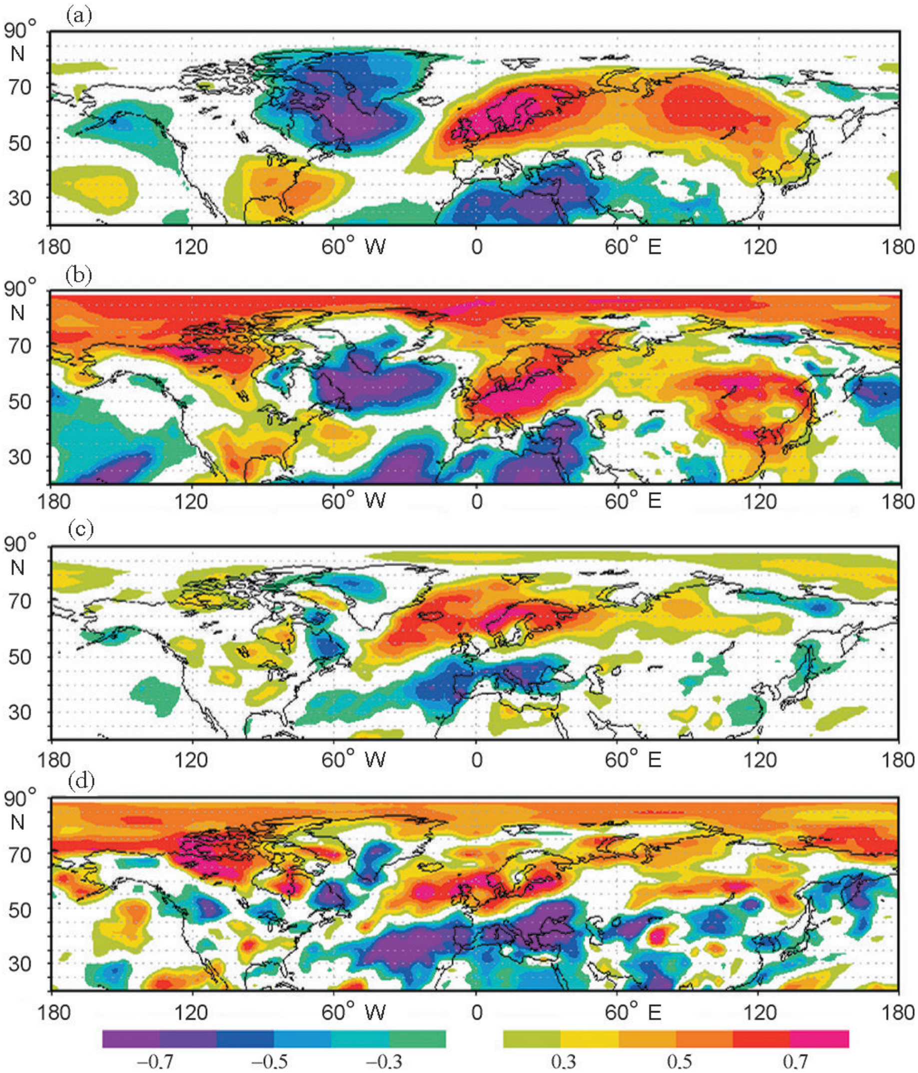

Fig. 3. Correlations between the high frequency (a, c) and low frequency (b, d) components of the mean wintertime (December–March) values of (a, b) air temperature, (c, d) total precipitation, and the AO index (1905–2008).

negative AO phase (Fig. 1b). This means that the prevalence of the negative anomalies of air temperature over the northern part of Northern Eurasia at the negative AO phase is associated with the fact that there meridional Arctic intrusions with the trend towards anticyclogenesis and blocking occur more frequently.

The absence of significant correlation between temperature and the AO index in the Central Arctic should be noted. The positive anomalies of air temperature in the Central Arctic are associated with the interlatitudinal exchange, namely, with warm advection from the south [8] as it was, for example, in the 1930s. The intensive zonal transport at the positive AO phase isolates the Central Arctic from the impact from the south that should have resulted in the cooling [8]. However, the significant destruction of ice cover in the Arctic Ocean under the influence of wind strengthening at the positive AO phase [5, 42], evidently compensates this isolation and results in the insignificance of relations.

The field of correlation coefficients for low-frequency variability components is presented in Fig. 3b. In Northern Eurasia the zone of correlations of above 0.5 mainly covers Western and Central Europe and only the northwestern regions of Eastern Europe. However, the values of correlation are kept statistically significant at the level of 5% even outside this zone. On the whole, increase is observed in the area of significant correlations over water areas that is explained by the high interannual inertia of sea surface temperature anomalies formed under the influence of AO. The correlation coefficients exceeding 0.5 for low-frequency components are observed over the most part of the Arctic that is evidently associated with decrease in the ice concentration at the positive AO phase and with the long-term inertia of ice cover characteristics [1, 2, 42].

The analysis of correlation coefficients between temperature and the AO index in Northern Eurasia enables the important conclusion: the successful forecast of AO polarity is needed for the accurate forecast of wintertime air temperature anomaly in Northern Eurasia. This conclusion is true at least for the territory of Eastern Europe where more than a half of the variance of average wintertime air temperature associated with its interannual variability is related to AO index.

Precipitation. The impact of AO on precipitation in Northern Eurasia is not as unambiguous as its impact on air temperature; therefore, the literature pays less attention to this problem. Meanwhile, if one analyzes the relations between precipitation in Northern Eurasia and AO as well as the relations between AO and air temperature and other meteorological parameters, the results agree well so illustrating the mechanisms of these relationships.

Two zones of significant correlations are clearly observed in the field of correlations between high-frequency components of wintertime total precipitation and AO index (Fig. 3c): the zone of positive correlations from Newfoundland to Chukotka and the zone of negative correlations from Florida to the Black Sea. These zones are still more pronounced for low-frequency components (Fig. 3d), and the zone of negative correlations spreads to Kamchatka with small breaks. These zones of significant correlations agree well with the correlations between the anomalies of the number of cyclonic centers and the AO index presented in Fig. 1b which indicate the position of main climatologic fronts at extratropical latitudes (the Arctic and polar fronts). The intensification of the climatologic Arctic front takes place in the winters with the positive polarity of AO, and the positive anomaly of precipitation is observed in the northern part of Northern Eurasia. The positive anomalies of precipitation over the North Atlantic and Eurasia in winters with the negative polarity of AO are mainly associated with the polar-frontal cyclones and are observed in the southern part of Northern Eurasia (the zone of 40 –50 N).

The most dramatic meridional contrast between the positive and negative anomalies of precipitation is observed over the North Atlantic and Western Europe. Though not so dramatic, this contrast is also clearly observed over Eastern Europe. The zones of positive anomalies of total precipitation tend to be displaced to the north of Eastern Europe in winters with the positive polarity of AO and to the southwest in winters with the negative polarity.

The comparison of correlation maps for the low-frequency components of the variability of precipitation (Fig. 3d) and air temperature (Fig. 3b) demonstrates that the zone of the maximum positive correlations of temperature over Europe is concentrated between climatologic fronts that are clearly manifested in precipitation. It should also be noted that the signs of correlation between AO and air temperature and precipitation mainly coincide over the northern part of Northern Eurasia (northwards of about 50 N). The positive anomalies and values of air temperature and total precipitation are observed at the positive AO phase, and the negative ones, at the negative AO phase. This is explained by the dominant contribution of circulation to the climate formation in this region in winter: the sign of anomalies of air temperature and precipitation is defined by the sign of the anomalies in advection of warm and humid Atlantic air to the continent. The signs of correlations mainly differ in the southern part of Northern Eurasia (southwards of 40 –50 N). The positive anomalies of air temperature and the negative anomalies of precipitation prevail at the positive AO phase, and opposite anomalies are registered at the negative AO phase. This happens because the contribution of the radiative climate-forming factor increases as the latitude decreases. At the positive AO phase the anticyclonic pressure field, the negative anomaly of precipitation and cloudiness, and the positive anomaly of insolation caused by the latter and defining the positive anomaly of temperature, prevail in these regions. The intensification of the climatological polar front accompanied by the positive anomaly of precipitation and cloudiness, by the negative anomaly of insolation, and by the negative anomaly of air temperature, occurs at the negative AO phase.

It is noteworthy that the relationship between the interannual variability of precipitation and AO is significantly manifested on the windward (western) slope of (quasi)meridionally oriented mountain systems, the Scandinavian Mountains and the Ural Mountains, and is not pronounced for the variability of the decadal scale.

The analysis of correlations between the wintertime total precipitation and winter AO index indicates the significant dependence of precipitation anomalies on the AO phase. Up to 30% of interannual variability and up to 50% of decadal variability of wintertime precipitation in Eastern Europe are associated with the AO index.

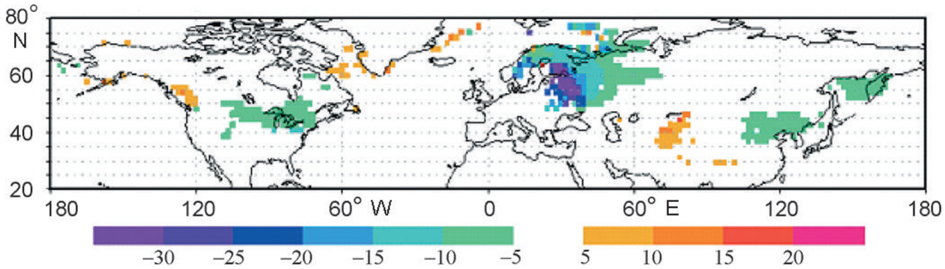

Fig. 4. Difference between the composites of the dates of spring transition of air temperature to positive values after the winters with the AO index in the upper and lower quintiles.

3.3. The Effects of Wintertime Arctic OScillation on the Terms of Springtime Temperature Transition to Positive Values

The effects that AO produces in winter on the climatic characteristics of spring are especially interesting. These characteristics include the period of snow melting, river ice breakup, flood beginning, and vegetation start. The statistically significant correlation between the springtime and summertime air temperature in different areas of Eastern Europe and the wintertime AO index was noted in [9, 10, 32, 33]. The territory of Northern Eurasia is huge and the terms of the beginning of spring vary considerably from region to region. So, we present here the estimates of the impact of the winter polarity of AO on the terms when air temperature steadily crosses 0 C (they are defined as the dates when the zero isotherm from minus to plus is crossed by the straight line connecting the monthly mean values of air temperature assigned to the middle of the month). For this purpose, the terms of the crossing (dates) were averaged for 22 years with the maximum (upper quintile) and 22 years with the minimum (lower quintile) values of the AO index from the period of 1901–2012, and the differences (days) between them were computed (Fig. 4). The differences exceeding five days are statistically significant at the level of 5% and are typical of almost the whole northern part of Eastern Europe. These differences exceed 30 days in the center and northwest of the European part of Russia. This means that the trend towards earlier spring is observed in the European part of Russia after the winters with the positive polarity of the AO index.

The earlier transition to the positive values of air temperature on the Arctic coast of Eastern Europe and Western Siberia is associated with the earlier melting of the coastal ice and with the positive correlation between the springtime temperature in the Barents Sea and the winter AO index [9]. The asynchronous impact of the wintertime circulation on the air temperature in continental Europe in spring is explained by the positive feedbacks between air temperature and snow cover [27]; the spring position of the southern boundary of snow cover depends on the intensity of zonal circulation in the preceding winter [3, 7].

After the winters with the positive AO phase the earlier ice breakup should be expected on the rivers in the central and northern part of Eastern Europe. In the north of the European part of Russia (Figs. 3c and 3d) such winters accompanied with large total wintertime precipitation also cause earlier and higher floods. Later and lower floods should be expected after the winters with the negative polarity of the AO index.

4. CONCLUSIONS

The Arctic Oscillation is the dominant type of wintertime atmosphere variability at the extratropical latitudes of the Northern Hemisphere characterized by different signs of the anomalies of air pressure and geopotential in the polar region and in the zone of 40 –50 N.

The positive AO phase is characterized by the intensification of the zonal circulation northwards of 40 –50 N, by the displacement of the trajectories of the Atlantic cyclones to the north of Eurasia, and by the intensification of the Arctic front and advection of the warm Atlantic air to the continent. In the case of the negative AO phase, the meridional circulation prevails, the trajectories of the Atlantic cyclones displace to the south, the blocking is observed in the Atlantic-European sector, and the advection of the Arctic air to the north of Eurasia intensifies.

The mean value of the Arctic Oscillation index in winter in the 20th century is characterized by significant high- and low-frequency variability. The most significant 20-year trends in the average wintertime AO index were observed in the 1920s–1930s (up to –1.79 SD/20 years), 1980s–1990s (up to 1.90 SD/20 years), and 1990s–2000s (up to –1.56 SD/20 years).

More than 25–50% of the variance of average winter air temperature in Northern Eurasia in the 20th century was associated with AO. The positive anomalies of air temperature prevail in the winters with the positive AO phase, and the negative anomalies of air temperature prevail in the winters with the negative AO phase.

Along with air temperature, the Arctic Oscillation significantly affects the spatial distribution of wintertime precipitation in Northern Eurasia. The intensification of the Arctic front in winters with the positive AO phase is accompanied by increase in precipitation in the north of the European part of Russia, in Siberia, and over the Arctic Ocean. In the case of the negative AO phase, the zone of precipitation increase is displaced to the south, to Southern Europe, Central Asia, and Mongolia.

The earlier transition to the positive air temperature in the northern and central parts of Eastern Europe associated with the positive AO phase in combination with the positive anomalies of wintertime precipitation, results in earlier and higher spring flood on the rivers of this region. The results of studying the relationship between the terms and height of the spring flood and the winter AO phase will be presented in the next paper.

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REFERENCES

1. G. V. Alekseev, L. V. Bulatov, V. F. Zakharov, and V. V. Ivanov, “Thermal Expansion of Atlantic Water in the Arctic Basin,” Meteorol. Gidrol., No. 7 (1998) [Russ. Meteorol. Hydrol., No. 7 (1998)].
2. G. V. Alekseev, V. F. Zakharov, A. N. Smirnov, and N. P. Smirnov, “Long-term Variations of Ice Conditions and Atmospheric Circulation in the Atlantic Arctic and North Atlantic,” Meteorol. Gidrol., No. 9 (1998) [Russ. Meteorol. Hydrol., No. 9 (1998)].
3. R. M. Vil’fand, V. P. Sadokov, and V. A. Tishchenko, “Relationship between the Snow Cover Boundary and Intensity of Cyclonic Activity in the Northern Hemisphere,” Meteorol. Gidrol., No. 9 (2002) [Russ. Meteorol. Hydrol., No. 9 (2002)].
4. E. M. Volodin and V. Ya. Galin, “Investigation of the First Mode of Low-frequency Variability of the Winter Atmospheric Circulation in Midlatitudes of the Northern Hemisphere,” Meteorol. Gidrol., No. 9 (1998) [Russ. Meteorol. Hydrol., No. 9 (1998)].
5. V. M. Kattsov, V. P. Meleshko, G. V. Alekseev, et al., “Effect of Sea Ice Compaction on High Latitude Atmospheric Variability,” Meteorol. Gidrol., No. 4 (1997) [Russ. Meteorol. Hydrol., No. 4 (1997)].
6. A. V. Kislov, *Climatology* (ITs Akademiya, Moscow, 2011) [in Russian].
7. L. M. Kitaev, “Spatial and Temporal Variations of Snow Depth in the Northern Hemisphere,” Meteorol. Gidrol., No. 5 (2002) [Russ. Meteorol. Hydrol., No. 5 (2002)].
8. *The Climatic Regime in the Arctic at the Turn of the 20th and 21st Centuries*, Ed. by B. A. Krutskikh (Gidrometeoizdat, St. Petersburg, 1997) [in Russian].
9. V. N. Kryzhov, “Connection between Monthly Mean, Seasonal, and Annual Air Temperatures in Northern Russia and Winter Zonal Circulation Indices,” Meteorol. Gidrol., No. 2 (2003) [Russ. Meteorol. Hydrol., No. 2 (2003)].
10. V. N. Kryzhov, “Relationship between Annual Mean Air Temperature in Northwestern Eurasia and the Arctic Oscillation,” Meteorol. Gidrol., No. 1 (2004) [Russ. Meteorol. Hydrol., No. 1 (2004)].
11. V. N. Kryzhov, “Causes of November Cooling of the 1980s–1990s in European Russia,” Meteorol. Gidrol., No. 1 (2008) [Russ. Meteorol. Hydrol., No. 1, **33** (2008)].
12. M. A. Tolstykh, D. B. Kiktev, R. B. Zaripov, et al., “Simulation of the Seasonal Atmospheric Circulation with the New Version of the Semi-Lagrangian Atmospheric Model,” Izv. Akad. Nauk, Fiz. Atmos. Okeana, No. 2, **46** (2010) [Izv., Atmos. Oceanic Phys., No. 2, **46** (2010)].
13. S. P. Khromov, *Introduction into Synoptic Analysis* (Gidrometeoizdat, Moscow, 1937) [in Russian].
14. J. B. Ahn and H. J. Kim, “Improvement of 1-month Lead Predictability of the Wintertime AO Using a Realistically Varying Solar Constant for a CGCM,” Meteorol. Appl., **21** (2014).
15. M. H. Ambaum and B. J. Hoskins, “The NAO Troposphere-stratosphere Connection,” J. Climate, **15** (2002).
16. M. P. Baldwin and T. J. Dunkerton, “Propagation of the Arctic Oscillation from the Stratosphere to the Troposphere,” J. Geophys. Res., **104** (1999).
17. M. P. Baldwin and T. J. Dunkerton, “Stratospheric Harbingers of Anomalous Weather Regimes,” Science, **294** (2001).

18. M. P. Baldwin, D. W. J. Thompson, E. F. Shuckburgh, et al., “Weather from the Stratosphere?”, Science, **301** (2003).
19. C. Bretherton, M. Widmann, V. Dymnikov, et al., “The Effective Number of Spatial Degrees of Freedom of a Time-varying Field,” J. Climate, **12** (1999).
20. J. G. Charney and P. G. Drazin, “Propagation of Planetary-scale Disturbances from the Lower into the Upper Atmosphere,” J. Geophys. Res., **66** (1961).
21. J. Cohen, M. Barlow, P. J. Kushner, and K. Saito, “Stratosphere-troposphere Coupling and Links with Eurasian Land Surface Variability,” J. Climate, **20** (2007).
22. J. Cohen, J. Foster, M. Barlow, et al., “Winter 2009–2010: A Case Study of an Extreme Arctic Oscillation Event,” Geophys. Res. Lett., **37** (2010).
23. G. P. Compo, J. S. Whitaker, P. D. Sardeshmukh, et al., “The Twentieth Century Reanalysis Project,” Quart. J. Roy. Meteorol. Soc., **137** (2011).
24. G. P. Compo, J. S. Whitaker, and P. D. Sardeshmukh, “Feasibility of a 100 Year Reanalysis Using Only Surface Pressure Data,” Bull. Amer. Meteorol. Soc., **87** (2006).
25. C. J. Fletcher and P. J. Kushner, “The Role of Linear Interference in the Annular Mode Response to Tropical SST Forcing,” J. Climate, **24** (2011).
26. C. I. Garfinkel, D. L. Hartmann, and F. Sassi, “Tropospheric Precursors of Anomalous Northern Hemisphere Stratospheric Polar Vortices,” J. Climate, **23** (2010).
27. P. Y. Groisman, T. E. Karl, R. W. Knight, and G. L. Stenchikov, “Changes of Snow Cover, Temperature, and Radiative Heat Balance over the Northern Hemisphere,” J. Climate, **7** (1994).
28. D. L. Hartmann, J. M. Wallace, V. Limpasuvan, et al., “Can Ozone Depletion and Global Warming Interact to Produce Rapid Climate Change?”, Proc. Nat. Acad. Sci. USA, **97** (2000).
29. P. H. Haynes, C. J. Marks, M. E. McIntyre, et al., “On the “Downward Control” of Extratropical Diabatic Circulations by Eddy Induced Mean Zonal Forces,” J. Atmos. Sci., **48** (1991).
30. *IPCC Fifth Assessment Report. Climate Change 2013: The Physical Science Basis Working Group I Contribution to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* (Cambridge Univ. Press, Cambridge, United Kingdom and New York, USA, 2013).
31. V. N. Kryjov, “October Circulation Precursors of the Wintertime Arctic Oscillation,” Int. J. Climat., **35** (2015).
32. V. N. Kryjov, “Searching for Circulation Patterns Affecting Northern Europe Annual Temperature,” Atmos. Sci. Lett., No. 1, **5** (2004).
33. V. N. Kryjov, “The Influence of the Winter Arctic Oscillation on the Northern Russia Spring Temperature,” Int. J. Climate, **22** (2002).
34. V. N. Kryjov and C.-K. Park, “Solar Modulation of the El Niño/Southern Oscillation Impact on the Northern Hemisphere Annular Mode,” Geophys. Res. Lett., **34** (2007).
35. Y. Kuroda and K. Kodera, “Role of Planetary Waves in the Stratosphere-troposphere Coupled Variability in the Northern Hemisphere Winter,” Geophys. Res. Lett., **26** (1999).
36. J. E. Kutzbach, “Large-scale Features of Monthly Mean Northern Hemisphere Anomaly Maps of Sea-level Pressure,” Mon. Wea. Rev., **98** (1970).
37. V. Limpasuvan, D. L. Hartmann, D. W. J. Thompson, et al., “Stratosphere-troposphere Evolution during Polar Vortex Intensification,” J. Geophys. Res., **110** (2005).
38. V. Limpasuvan, D. W. J. Thompson, and D. L. Hartmann, “The Life Cycle of Northern Hemisphere Sudden Stratospheric Warmings,” J. Climate, **17** (2004).
39. E. N. Lorenz, “Seasonal and Irregular Variations of the Northern Hemisphere Sea-level Pressure Profile,” J. Meteorol., **8** (1951).
40. T. J. Mosedale, D. B. Stephenson, M. Collins, and T. C. Mills, “Granger Causality of Coupled Climate Processes: Ocean Feedback on the North Atlantic Oscillation,” J. Climate, **19** (2006).
41. L. L. Pan, “Observed Positive Feedback between the NAO and the North Atlantic SSTA Tripole,” Geophys. Res. Lett., **32** (2005).
42. I. G. Rigor, J. M. Wallace, and R. L. Colony, “Response of Sea Ice to the Arctic Oscillation,” J. Climate, **15** (2002).
43. A. A. Scaife, D. Copsey, C. Gordon, et al., “Improved Atlantic Blocking in a Climate Model,” Geophys. Res. Lett., **38** (2011).
44. A. A. Scaife, A. Arribas, E. Blockley, et al., “Skillful Long-range Prediction of European and North American Winters,” Geophys. Res. Lett., **41** (2014).
45. M. C. Serreze, *Northern Hemisphere Cyclone Locations and Characteristics from NCEP/NCAR Reanalysis Data. 1958–2008* (National Snow and Ice Data Center, Boulder, CO, 2009).
46. D. T. Shindell, G. A. Schmidt, R. L. Miller, and D. Rind, “Northern Hemisphere Winter Climate Response to Greenhouse Gas, Ozone, Solar, and Volcanic Forcing,” J. Geophys. Res., **106** (2001).
47. J. Sun and J. B. Ahn, “Dynamical Seasonal Predictability of the Arctic Oscillation Using a CGCM,” Int. J. Climat., Wiley Online Library (2014).

48. D. W. J. Thompson and J. M. Wallace, “The Arctic Oscillation Signature in Wintertime Geopotential Height and Temperature Fields,” Geophys. Res. Lett., **25** (1998).
49. D. W. J. Thompson and J. M. Wallace, “Annular Modes in the Extratropical Circulation, Part I: Month to Month Variability,” J. Climate, **13** (2000).
50. D. W. J. Thompson, J. M. Wallace, and G. C. Hegerl, “Annular Modes in the Extratropical Circulation, Part II: Trends,” J. Climate, **13** (2000).
51. K. E. Trenberth and D. A. Paolino, “Characteristic Patterns of Variability of Sea Level Pressure in the Northern Hemisphere,” Mon. Wea. Rev., **109** (1981).
52. E. M. Volodin and V. Ya. Galin, “The Nature of the Northern Hemisphere Winter Troposphere Circulation Response to Observed Ozone Depletion in Low Stratosphere,” Quart. J. Roy. Meteorol. Soc., **124** (1998).
53. J. M. Wallace, “North Atlantic Oscillation/Annular Mode: Two Paradigms—One Phenomenon,” Quart. J. Roy. Meteorol. Soc., **126** (2000).
54. J. M. Wallace and D. S. Gutzler, “Teleconnections in the Geopotential Height Field during the Northern Hemisphere Winter,” Mon. Wea. Rev., **109** (1981).
55. L. Wang and W. Chen, “Downward Arctic Oscillation Signal Associated with Moderate Weak Stratospheric Polar Vortex and the Cold December,” Geophys. Res. Lett., **37** (2010).
56. J. S. Whitaker, G. P. Compo, X. We, and T. M. Hamill, “Reanalysis without Radiosondes Using Ensemble Data Assimilation,” Mon. Wea. Rev., **132** (2004).
57. D. S. Wilks, “Resampling Hypothesis Tests for Autocorrelated Fields,” J. Climate, **10** (1997).