Recent temperature and precipitation increases in West Siberia and their association with the Arctic Oscillation

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Surface air temperature and precipitation records for the years 1958–1999 from ten meteorological stations located throughout West Siberia are used to identify climatic trends and determine to what extent these trends are potentially attributable to the Arctic Oscillation (AO). Although recent changes in atmospheric variability are associated with broad Arctic climate change, West Siberia appears particularly susceptible to warming. Furthermore, unlike most of the Arctic, moisture transport in the region is highly variable. The records show that West Siberia is experiencing significant warming and notable increases in precipitation, likely driven, in part, by large-scale Arctic atmospheric variability. Because this region contains a large percentage of the world's peatlands and contributes a significant portion of the total terrestrial freshwater flux to the Arctic Ocean, these recent climatic trends may have globally significant repercussions. The most robust patterns found are strong and prevalent springtime warming, winter precipitation increases, and strong association of non-summer air temperatures with the AO. Warming rates for both spring (0.5 - 0.8 °C/decade) and annual (0.3 - 0.5 °C/decade) records are statistically significant for nine of ten stations. On average, the AO is linearly congruent with 96 % (winter), 19 % (spring), 0 % (summer), 67 % (autumn) and 53 % (annual) of the warming found in this study. Significant trends in precipitation occur most commonly during winter, when four of ten stations exhibit significant increases (4 - 13 %/decade). The AO may play a lesser role in precipitation variability and is linearly congruent with only 17 % (winter), 13 % (spring), 12 % (summer), 1 % (autumn) and 26 % (annual) of precipitation trends.

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The sensitivity of Arctic climate is documented in the instrumental record of recent decades and is evident in the large temperature increases seen over Northern Hemisphere land areas from about 40° N to 70° N (Nicholls et al. 1996). In particular, sharp temperature increases from 1966–1995 are found in Eurasia and northwest North America, with pronounced warming during winter and spring (Serreze et al. 2000). Similarly, Fallot et al. (1997) find significant positive trends in winter temperatures across much of the former Soviet Union (FSU) since the early 1950s or mid-1960s. More recently (1979–1997), winter temperatures have warmed at an exceptionally high rate of 1 °C/decade in the Eurasian Arctic, which is contrasted by winter cooling (–1 °C/decade) in the North American Arctic (Rigor et al. 2000). Although Serreze et al. (2000) show relatively large winter temperature increases over widespread areas in Eurasia and western North Amer-

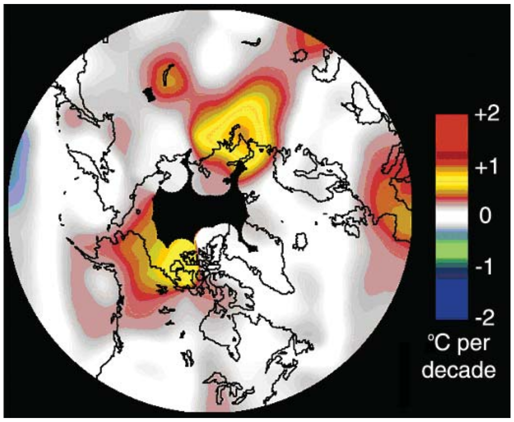


Fig. 1. Trends in summer mean surface air temperature (°C per decade) from 40° N to 90° N for the years 1966–1995 (modified from Serreze et al. 2000; printed with kind permission of Kluwer Academic Publishers, M. Serreze and J. Walsh).

Fig. 2. West Siberia, showing locations of the ten meteorological stations used in this study.

ica, warming during spring and summer is amplified over West Siberia and the North Slope of Alaska (Fig. 1). Northern Eurasia has also experienced slight increases in precipitation over the last 50 years (Groisman et al. 1991), concurrent with significant increases in cyclone density since the mid-1960s for regions north of 60° N (Serreze et al. 1997) and large reductions in sea level pressure over the central Arctic (Walsh et al. 1996; Serreze et al. 2000).

A primary mode of climate variability in the Arctic is the Arctic Oscillation (AO; Thompson & Wallace 1998). The AO index is determined by the variation of the wintertime leading empirical orthogonal function (EOF) of sea level pressure northward of 20° N and is dependent on the fluctuation of atmospheric pressures between the central Arctic and two weaker centres at about 45° N over the Atlantic and Pacific basins. The AO is highly correlated with atmospheric phenomena throughout the Northern Hemisphere, accounting for ca. 50 % of the winter warming observed over Eurasia and ca. 30 % of the winter warming seen over the whole Northern Hemisphere for varying record lengths (Thompson & Wallace 1998; Thompson et al. 2000). AO trends are also well correlated with variability in sea level pressure, storm tracks and precipitation (Thompson & Wallace 2001). Since the late 1960s, the AO has exhibited a pronounced positive phase (Thompson & Wallace 1998, 2001; Thompson et al. 2000), which may be partly responsible for the recent climate change found in the High latitude Northern Hemisphere. It is uncertain, however, whether this recent AO trend is reflective of natural climate variability and/or symptomatic of anthropogenic forcing such as lower stratosphere ozone depletion (Volodin & Galin 1998) and greenhouse gas emissions to the stratosphere (Fyfe et al. 1999; Shindell et al. 1999).

Although recent changes in atmospheric variability are associated with broad Arctic climate change, West Siberia appears particularly susceptible to warming (Fig. 1). Furthermore, the prevailing direction (north or south) of moisture flux in the region is highly unpredictable each year, unlike most other parts of the Arctic (Rogers et al. 2001). This regional temperature amplification and highly unstable direction of moisture flux are particularly relevant to the global carbon cycle, owing to the presence of the West Siberian peatlands that occupy nearly 400 000 km^2 in the region (Zhulidov et al. 1997). Northern peatlands are a major pool of stored carbon in the form of undecomposed plant matter and are a significant component of global carbon sequestration and emission calculations. For example, total carbon stocks of northern peatlands are currently estimated at 455 Pg C (ca. one-third of the global pool of soil C; Gorham 1991). In addition, northern peatlands are currently a significant source of global atmospheric methane (Roulet et al. 1992; Panikov 1999), which has been modelled at ca. 20 Tg CH_4/year (Christensen et al. 1996). While currently a net sink or small source of CO_2, northern peatlands may become a significant source of atmospheric CO_2 under a warming climate, owing to associated reduction in wetness and aerobic decomposition of peat (Gorham 1991, 1994; Oechel et al. 1993). Emissions of CH_4 are also expected to take dramatic shifts with changes in wetness (e.g. Laine et al. 1996; Moore et al. 1998). The observed recent climatic trends in West Siberia may therefore have critical consequences for global carbon cycle dynamics.

West Siberia (Fig. 2) generates large volumes of freshwater runoff to the Arctic Ocean. Therefore, any perturbations in temperature and/or precipitation across West Siberia may also affect global-scale processes through hydrology. The Ob' and Yenisey rivers of West Siberia alone account for about 35 % of the total terrestrial freshwater flux to the Arctic Ocean (Aagaard & Carmack 1989). Warming temperatures and increasing precipitation in West Siberia may alter the timing, volume and distribution of freshwater transport to the Arctic Ocean, thereby potentially modifying Arctic Ocean and global ocean circulation by triggering changes in the volume of North Atlantic Deep Water (NADW) formation, salinity distribution and sea ice formation (Rahmstorf 1995; Vörösmarty et al. 2001).

In sum, the data of Serreze et al. (2000; Fig. 1) suggest amplification of warming in West Siberia, a region of global significance with respect to hydrology, carbon storage and greenhouse gas exchange. Here, we present meteorological data for the years 1958–1999 from ten stations located throughout West Siberia (Fig. 2, Table 1). The stations are geographically located within the area of maximum warming in Eurasia (Fig. 1). Seasonal and annual time series of temperature and precipitation are investigated with the Mann-Kendall test to determine the presence of long-term trends. The meteorological records are then compared with the AO index to determine the components of the observed trends that are linearly congruent with the AO.

Data and methods

Meteorological and Arctic Oscillation data

Meteorological data for the years 1958–1999

Table 1. Meteorological station records used in this study. Average surface air temperature and precipitation over the record period are given for December–February (DJF), March–May (MAM), June–August (JJA), September–November (SON) and December–November (annual).

Met. station	Location	Elev. (m)	Record	No. years	Temperature (°C)					Precipitation (mm)				
					DJF	MAM	JJA	SON	Annual	DJF	MAM	JJA	SON	Annual
Mys Kamennyj	68.47° N, 73.60° E	2	1958–1993	36	–24.1	–14.0	6.7	–5.7	–9.2	75.8	70.3	119.1	107.6	372.4
Berezovo	63.93° N, 65.05° E	27	1958–1999	42	–20.5	–4.3	13.6	–3.3	–3.5	78.2	96.8	197.0	143.7	516.1
Tarko-Sale	64.92° N, 77.82° E	27	1958–1999	42	–23.6	–8.5	12.5	–5.5	–6.2	75.8	88.2	193.0	149.3	504.0
Turuhansk	65.78° N, 87.95° E	37	1960–1999	40	–24.8	–8.0	13.1	–6.5	–6.4	102.6	93.2	186.0	178.3	560.6
Khanty-Mansiysk	60.97°N, 69.07° E	45	1958–1999	42	–18.4	–1.4	15.4	–1.4	–1.3	82.1	92.9	215.0	152.9	540.5
Aleksandrovskoe	60.43° N, 77.87° E	47	1958–1999	42	–19.6	–2.6	15.3	–2.1	–2.1	63.8	84.7	213.0	133.1	494.6
Bor	61.60° N, 90.00° E	62	1959–1999	41	–22.7	–3.6	15.2	–3.6	–3.6	109.9	103.6	194.1	183.2	591.9
Tobol'sk	58.15° N, 68.18° E	48.5	1958–1999	42	–16.7	0.9	16.5	0.3	0.3	62.8	74.7	192.4	119.7	449.7
Tara	56.90° N, 74.38° E	73	1958–1999	42	–16.9	0.9	16.7	0.4	0.4	58.1	75.9	187.7	109.6	433.0
Kolpasev	58.30° N, 82.90° E	80	1958–1999	42	–18.4	–1.0	16.0	–1.0	–1.0	71.2	89.2	198.0	134.7	493.1

from ten West Siberian stations (Mys Kamennyj, Berezovo, Tarko-Sale, Turuhansk, Khanty-Mansiysk, Aleksandrovskoe, Bor, Tobol'sk, Tara and Kolpasev; Fig. 2, Table 1) are presented. These stations are physiographically similar with the exception of Mys Kammenyj, which may experience a greater maritime influence than other stations owing to its coastal location. Data for the years 1958–1995 were obtained from the *Daily temperature and precipitation data for 223 USSR stations* dataset (Razuvaev et al. 1993), available from the National Climatic Data Center (NCDC). Daily temperature and precipitation data for the years 1996–99 were obtained from the NCDC Climate Visualization website (http://lwf.ncdc.noaa.gov/oa/climate/onlineprod/drought/xmgr.html/). Six-hourly surface data for Tarko-Sale were purchased directly from the NCDC as individual station data. Records after 1999 were not examined owing to the abrupt increase of missing data after this year. Monthly AO indices (Thompson & Wallace 2000) were obtained from the Colorado State University Annular Modes website (http://horizon.atmos.colostate.edu/ao). Temperature and AO time series were averaged and precipitation time series were summed into winter (December–February, DJF), spring (March–May, MAM), summer (June–August, JJA), autumn (September–November, SON) and annual (December–November) time series for the years 1958–1999.

The physical relevance of the AO compared to the North Atlantic Oscillation (NAO; e.g. Hurrell 1995) has been subject to considerable debate (Deser 2000; Ambaum et al. 2001; Rogers & McHugh 2002). The NAO describes a large-scale fluctuation of atmospheric mass in the North Atlantic region and is a major source of seasonal to interdecadal variability in atmospheric circulation throughout the Northern Hemisphere (Hurrell 1995). Although the NAO and AO are highly correlated, the AO is found to be more strongly coupled to surface air temperature fluctuations over the Eurasian continent than the NAO (Thompson & Wallace 1998). Thompson & Wallace (1998) also suggest that the AO captures more hemispheric variability in atmospheric circulation, as opposed to more regional North Atlantic/European variability. In recognition of these observations, we focus this study on the AO.

Empirical studies show that up to three correction factors (K1, K2 and K3) should be applied to precipitation records from the FSU in order to homogenize the time series (e.g. Groisman et al. 1991; Fallot et al. 1997). K1 compensates for a change in gauge shield design implemented in the late 1940s to early 1950s. Because the time series used in this study start in 1958, this correction is not needed. K2 corrects for wind effects on precipitation measurements, which is particularly important during winter months. K2 coefficients rely on annual mean wind speeds averaged over several years, introducing potentially large errors when applied to monthly time series. K2 coefficients were therefore not applied in this study. K3 corrects for a change in FSU measurement protocol implemented in 1966, after which moisture adhering to gauge walls was included in the precipitation estimate. Therefore, this K3 correction was applied to all precipitation data in this study prior to 1966, using values for "raw" precipitation time series as recommended by Groisman & Rankova (2001). The addition of these recommended K3 coefficients can significantly alter observed trends in precipitation and can increase measured values by up to 6.4 % in summer months and 9.6 % in winter months.

Assessing potential errors in temperature and precipitation records associated with urbanization, land use change and data collection methods is a challenging task. Despite industrialization that began in the early 1960s throughout much of the region, the ten stations chosen for this study are located in relatively small cities, all with fewer than 100 000 inhabitants and most with significantly fewer than 50 000 inhabitants. Jones et al. (1990) found little evidence for significant urban influence on temperature from 1930 to 1987 in the western part of Russia, including West Siberia. Urbanization and land use change can also affect precipitation measurements, although these effects are not normally apparent until ca. 20 - 40 km downwind of urban centres (e.g. Oke 1987). Precipitation collection methods are inherently more problematic than those for temperature. In 1965, precipitation measurements within the FSU increased from two to four times daily, although this probably did not significantly alter the total amount of precipitation measured. Errors introduced by applying the generalized K3 wetting correction coefficients or not applying the wind-compensating K2 correction coefficients (as described above) are also a consideration. Precipitation in the form of snow may also be more difficult to measure than rain, owing to the potential for significant problems of undercatch of solid precipitation (e.g. Woo et al. 1983; Yang & Ohata 2001).

Trend analysis and linear congruence with the AO

Seasonal and annual time series of surface air temperature and precipitation are analysed using the nonparametric Mann-Kendall test for monotonic trend (Mann 1945; Kendall 1975; Maidment 1993), which makes no hypothesis about the value of a parameter in a statistical density function. The Mann-Kendall statistic S is given by:

$$S=\sum_{t'=1}^{n-1}\sum_{t=t'+1}^{n} z_k \qquad (1)$$

where the ranked series z_k is generated by first considering the annual time series y_t, t = 1,…, n and comparing each value $y_{t'}$, t = 1, n – 1 with all subsequent values y_t, $t = t' + 1, t' + 2, \ldots, n$ and applying the following conditions:

$$\begin{aligned} z_k&=1 \quad \text{if } y_t>y_{t'} \\ z_k&=0 \quad \text{if } y_t=y_{t'} \\ z_k&=-1 \quad \text{if } y_t<y_{t'} \end{aligned} \qquad (2)$$

The S statistic therefore represents the number of positive differences minus the number of negative differences found in y_t. For $n>40$, the standardized test statistic z is obtained using a normal approximation:

$$z=\frac{S+m}{\sqrt{Var(S)}} \qquad (3)$$

where $m=1$ if $S<0$, $m=0$ if $S=0$, and $m=-1$ if $S>0$. Because the Mann-Kendall test is based on ranks of the data only, a correction is needed for the effect of data ties on the variance of S. Data ties occur when adjacent entries have the same value or when two or more years of data are absent (missing values are replaced with the series mean). The correction is as follows:

$$Var(S)=\frac{1}{18}\left[n(n-1)(2n+5)-\sum_{i=1}^{n} t_i(t_i-1)(2t_i+5)\right] \qquad (4)$$

where n is the number of tied groups and t_i is the number of data in the ith (tied) group. The null hypothesis (no trend) is rejected at the α significance level if $|z| > z_{(1-\alpha/2)}$, where $z_{(1-\alpha/2)}$ is the $1-\alpha/2$ quantile of the standard normal distribution. This study uses an exceedance probability of $p=0.90$ ($\alpha=0.2$) to establish trend. A slope estimator is not used (e.g. Sen 1968), since the aim of this study is simply to establish the presence or absence of trend.

To estimate the potential contribution of the AO to observed trends in temperature and precipitation, we apply trend analysis similar to Rigor et al. (2000), Thompson et al. (2000), Kryjov (2002), Rigor et al. (2002) and Wallace & Solomon (2002). The analysis is as follows: 1) all time series are linearly detrended; 2) for each resulting time series, values of temperature or precipitation are regressed onto associated AO indices (where AO indices are normalized by the series standard deviation); 3) the resulting regression coefficient is then multiplied by the linear trend in the associated AO index time series (in units of standard deviations per decade); 4) this product is the component of the decadal temperature or precipitation trend that is "linearly congruent" with the AO. Linear congruence does not necessarily imply that the AO is driving the observed variance in temperature or precipitation, but it does identify likely connections between them.

Results

Departures of surface air temperature, precipitation and AO index from their associated long-term record means (Table 1) are shown in Fig. 3a. Positive values indicate above average anomalies and negative values indicate below average anomalies. Both seasonal and annual temperature anomalies for the ten stations are relatively similar to one another for a given year (Fig. 3a), indicating a strong correlation in temperature variability over a relatively broad geographic area. This pattern is much less evident in records of precipitation departures, where little correlation exists between stations for a given year (Fig. 3b). Interannual variability in temperature is greatest during DJF (ca. 18 °C) and least during JJA (ca. 4 °C). In contrast, interannual variability in precipitation is greatest during JJA (ca. 200 mm) and least during DJF (ca. 100 mm). This largely exists owing to the significantly greater amount of precipitation falling in JJA as compared with DJF (e.g. JJA precipitation averages more than twice that of DJF precipitation; Table 1). For this reason, July precipitation variability can drive up

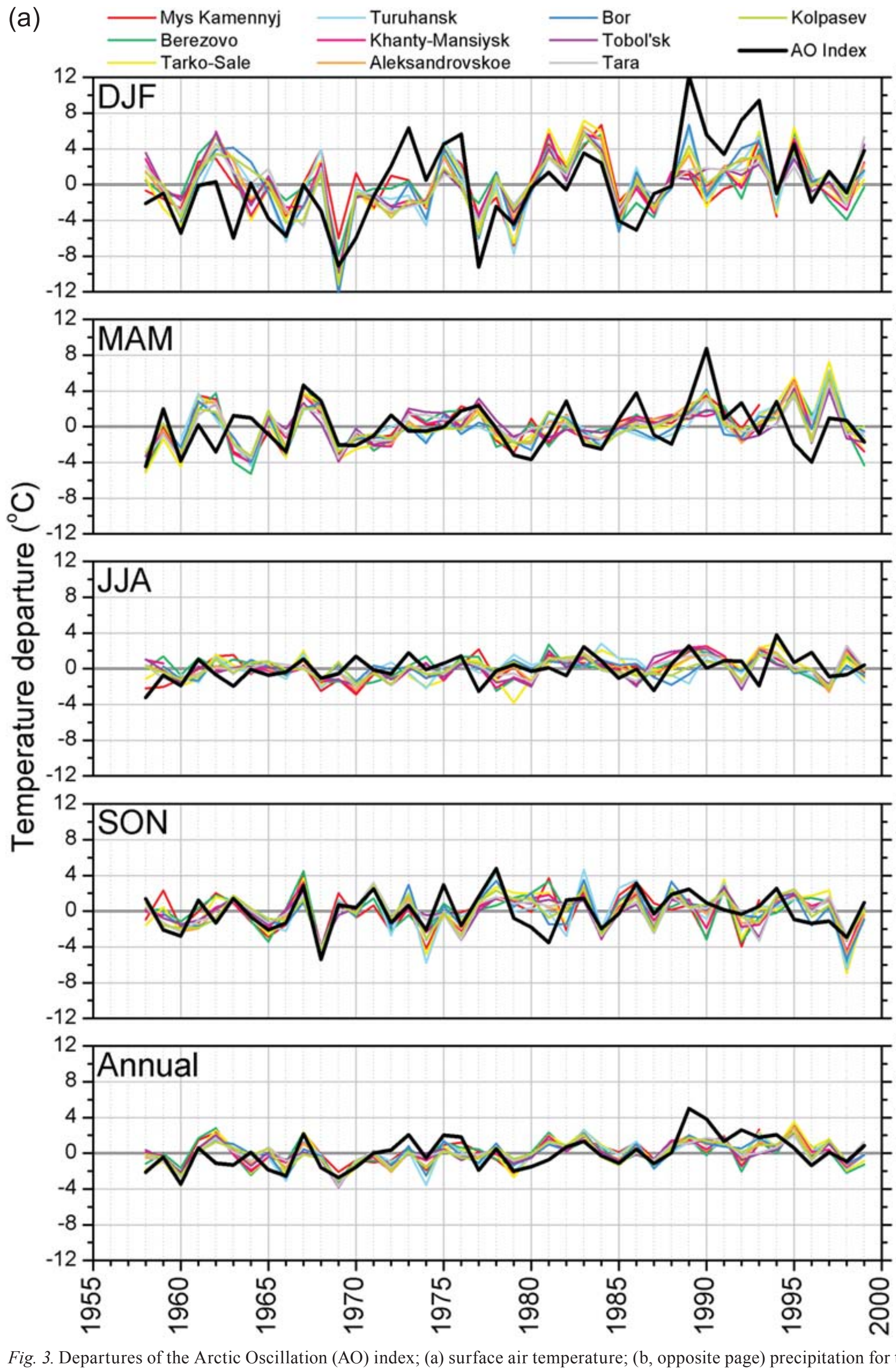
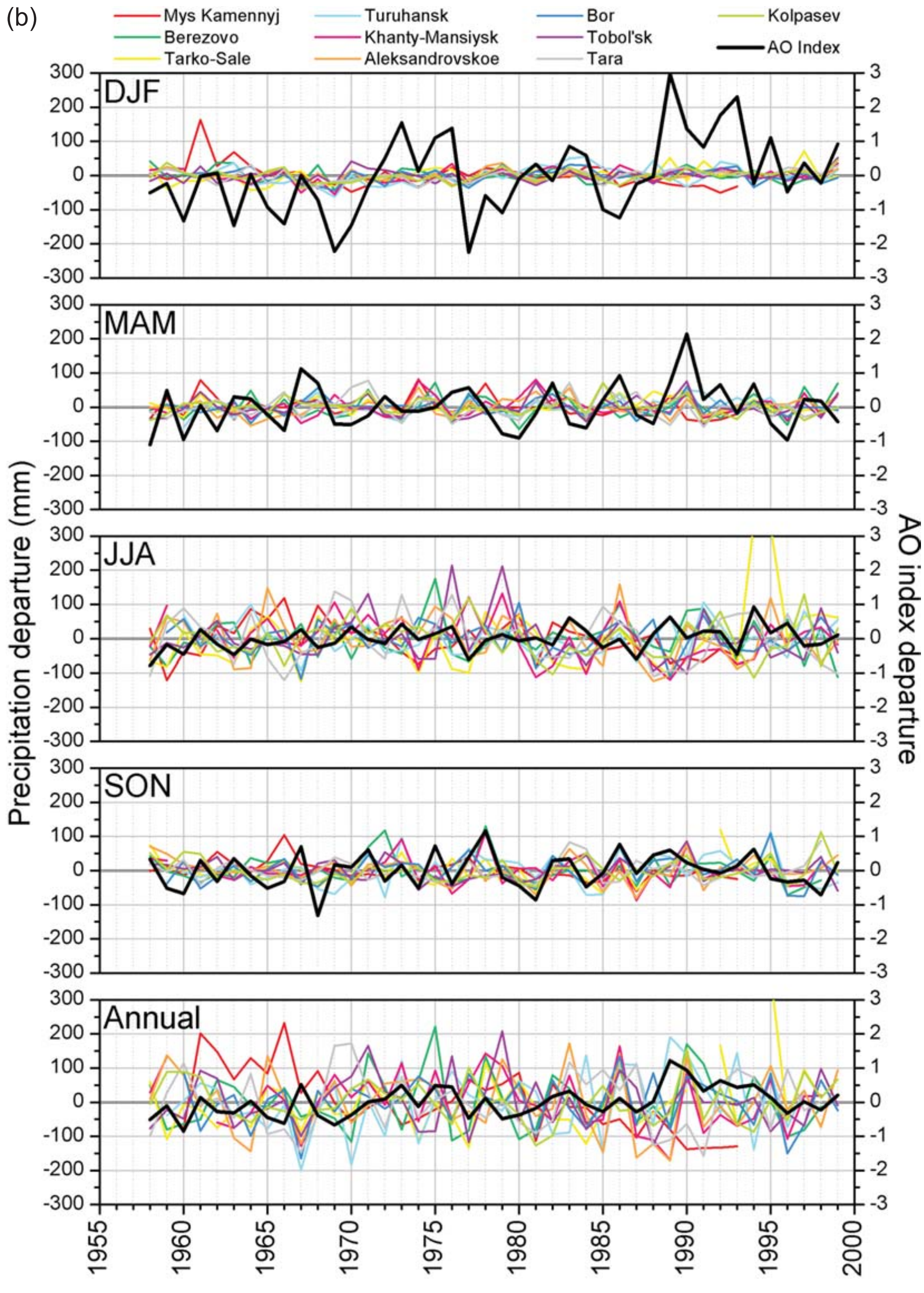


Fig. 3. Departures of the Arctic Oscillation (AO) index; (a) surface air temperature; (b, opposite page) precipitation for ten West Siberian stations. Plots are shown for December–February (DJF), March–May (MAM), June–August (JJA), September–November (SON) and December–November (annual). Departures are calculated from associated long-term record means.


(b)
Mys Kamennyj
Turuhansk
Bor
Kolpasev
Berezovo
Khanty-Mansiysk
Tobol'sk
AO Index
Tarko-Sale
Aleksandrovskoe
Tara
DJF
MAM
JJA
SON
Annual
Precipitation departure (mm)
AO index departure
300
200
100
0
-100
-200
-300
3
2
1
0
-1
-2
-3
1955
1960
1965
1970
1975
1980
1985
1990
1995
2000

to a 70 % difference in annual poleward moisture flux in this region (Rogers et al. 2001).

Of the 50 seasonal and annual temperature records analysed, 28 (56 %) display trends at significance levels of $p \geq 0.90$, all of which display positive (warming) trends (Table 2). Of these 28 records, 14 records display trends at significance levels of $p = 0.90$, 9 records at $p = 0.95$, and 5 records at $p = 0.99$. Table 3 presents corresponding decadal temperature trends for these 28 stations, calculated from ordinary least-squares regression over the associated length of each record. Nine of ten stations exhibit significant warming trends for both MAM and annual temperature records (Tables 2, 3), while statistically significant trends in DJF, JJA and SON temperatures are much less common (Tables 2, 3). Trends in annual time series range from 0.3 - 0.5 °C/decade and trends in seasonal time series range from 0.1 - 0.8 °C/decade. On average, the steep-

Table 2. Mann-Kendall statistics for seasonal and annual time series (ca. 1958–1999) of surface air temperature and precipitation for ten meteorological stations. Negative values indicate negative trends and positive values indicate positive trends. Significance levels: $p = 0.90$ (boldface), $p = 0.95$ (italicized boldface) and $p = 0.99$ (underlined, italicized boldface).

Met. station	Temperature (°C)					Precipitation (mm)				
	DJF	MAM	JJA	SON	Annual	DJF	MAM	JJA	SON	Annual
Mys Kamennyj	0.37	**1.62**	<u>***2.47***</u>	0.57	***1.76***	<u>***–2.52***</u>	<u>***–2.41***</u>	<u>***–2.48***</u>	<u>***–2.85***</u>	<u>***–3.99***</u>
Berezovo	0.02	***1.91***	-0.07	0.41	1.02	–0.61	0.91	0.48	–1.13	–0.19
Tarko-Sale	1.01	<u>***2.58***</u>	***1.98***	**1.33**	***2.31***	***2.25***	1.06	**1.52**	–0.84	1.09
Turuhansk	1.02	**1.57**	0.80	0.97	**1.50**	***1.77***	***1.70***	–0.10	0.37	**1.48**
Khanty-Mansiysk	0.98	**1.59**	1.14	1.06	**1.61**	0.41	0.74	**–1.38**	–1.15	–1.12
Aleksandrovskoe	**1.32**	**2.17**	1.24	1.13	<u>***2.34***</u>	0.79	***1.72***	–1.13	–1.01	–0.80
Bor	***1.64***	**1.52**	0.33	0.51	***1.69***	0.03	–0.86	–0.42	0.62	0.15
Tobol'sk	1.08	0.91	1.13	**1.41**	**1.50**	0.64	0.52	–0.67	–0.41	0.07
Tara	**1.30**	**1.37**	***1.80***	***1.65***	<u>***2.43***</u>	<u>***2.49***</u>	–0.43	–0.72	0.00	0.12
Kolpasev	**1.39**	***2.08***	1.02	1.15	<u>***2.54***</u>	**1.39**	–1.00	–0.87	0.26	–0.56

Table 3. Estimated decadal trends in surface air temperature and precipitation (ca. 1958–1999) for those records displaying statistically significant ($p \geq 0.90$) Mann-Kendall statistics. Values in parentheses indicate the components of the trends that are linearly congruent with the AO index (as defined in the section "Trend analysis and linear congruence with the AO").

Met. station	Temperature (°C / decade)					Precipitation (% / decade)				
	DJF	MAM	JJA	SON	Annual	DJF	MAM	JJA	SON	Annual
Mys Kamennyj		0.7 *(0.2)*	0.6 *(0.0)*		0.5 *(0.2)*	–15 *(–1)*	-10 *(-1)*	–9 *(–1)*	–13 *(0)*	–12 *(0)*
Berezovo		0.6 *(0.1)*								
Tarko-Sale		0.8 *(0.1)*	0.4 *(0.0)*	0.2 *(0.1)*	0.5 *(0.3)*	13 *(4)*		18 *(2)*		
Turuhansk		0.5 *(0.1)*			0.4 *(0.2)*	9 *(2)*	9 *(2)*			4 *(2)*
Khanty-Mansiysk		0.5 *(0.1)*			0.3 *(0.1)*			–6 *(–1)*		
Aleksand-rovskoe	0.6 *(0.6)*	0.6 *(0.1)*			0.4 *(0.2)*		7 *(1)*			
Bor	0.8 *(0.8)*	0.5 *(0.1)*			0.4 *(0.3)*					
Tobol'sk				0.1 *(0.1)*	0.3 *(0.2)*					
Tara	0.5 *(0.5)*	0.5 *(0.1)*	0.3 *(0.0)*	0.2 *(0.1)*	0.4 *(0.2)*	11 *(2)*				
Kolpasev	0.7 *(0.6)*	0.6 *(0.1)*			0.4 *(0.2)*	4 *(1)*				
Average trend[1]	0.7	0.6	0.4	0.2	0.4	4	2	1	–13	–4
Average trend[2]						9	8	6		4
Average % of trends congruent with the AO	96 %	19 %	0 %	67 %	53 %	17 %	13 %	12 %	1 %	26 %

[1] Average incorporating all statistically significant trends.

[2] Average excluding the negative precipitation trends at Mys Kamennyj.

est trends are found during DJF (0.7 °C/decade) and the shallowest trends are found during SON (0.2 °C/decade). However, statistically significant trends are most consistently found in MAM (average 0.6 °C/decade) and annual records (average 0.4 °C/decade).

Of the 50 seasonal and annual precipitation records analysed, only 14 (28 %) display trends at significance levels of $p \geq 0.90$ (Table 2). Of these 14 records, 4 records display trends at significance levels of $p = 0.90$, 4 records at $p = 0.95$, and 6 records at $p = 0.99$. Statistically significant trends are positive for all station records (i.e. precipitation increase) except for those at Mys Kammenyj, where negative trends are found for all seasonal and annual records (Table 3). Significant trends in precipitation are most commonly found during DJF (five stations) and least commonly found during SON (one station). Seasonal trends range from –15 to +18 %/decade (–21.2 to +25.6 mm/decade). Annual trends range from –12 to 4 %/decade (–59.5 to +20.2 mm/decade) (Table 3). Statistically significant trends in precipitation are most commonly found at higher latitudes, with all but two of the significant trends occurring at stations north of 60° N.

The components of observed trends in temperature and precipitation that are linearly congruent with the AO are also presented in Table 3. Correlations between the AO and temperature and precipitation time series are qualitatively apparent in Fig. 3. The most robust pattern found is the high linear congruence of the AO with non-summer air temperatures. On average, the AO is linearly congruent with 96 % of DJF, 19 % of MAM, 67 % of SON, and 53 % of annual warming seen in this study (Table 3). In contrast, none of the summer warming found in this study can be attributed to the AO. The potential contribution of the AO to precipitation trends is relatively weak (Table 3) and is largest for DJF and annual time series, but smallest for SON time series. On average, the AO is linearly congruent with 17 % of DJF, 13 % of MAM, 12 % of JJA, 1 % of SON and 26 % of annual precipitation trends.

Discussion and conclusions

The most robust findings of this study are strong and prevalent springtime warming, increases in winter precipitation and strong association of non-summer air temperatures with the AO. Surface air temperatures in West Siberia have increased strongly since 1958 (ca. 0.4 °C/decade annually; Table 3) as compared with global mean surface air temperature increases over the past 40 years (0.050 - 0.075 °C/decade; Nicholls et al. 1996). West Siberia may therefore be considered a region of amplified warming. Annual temperature records show statistically significant warming trends at nine of ten stations, ranging from 0.3 to 0.5 °C/decade (Table 3). However, these may be primarily driven by even stronger warming during MAM, the season exhibiting the most prevalent warming trends (Table 2). Rates of MAM temperature increase average 0.6 °C/decade (ranging from 0.5 to 0.8 °C/decade; Table 3). This finding is consistent with a broader pattern of warming Eurasian springs. Using Russian North Pole drifting station records for the years 1961–1990, Martin et al. (1997) find statistically significant increases in May temperatures of 0.89 °C/decade. Krijov (2002) finds even stronger springtime warming (up to ca. 1.4 °C/decade) from 1968–1997 in northern Russia. Rigor et al. (2000) also observe spring warming as high as 2 °C/decade in the eastern Arctic. In fact, Groisman et al. (1994) find that Northern Hemisphere temperatures have increased more during spring than any other season. These springtime temperature increases have been shown to parallel the retreat of Northern Hemisphere spring snow cover extent and are likely related with a strong positive feedback (Groisman et al. 1994; Brown 2000). Our finding of strong springtime warming is corroborated by time shifts (ca. 1 - 3 weeks) in river ice melt onset dates in western and eastern Siberia for the years 1917–1994 (Smith 2000). Additionally, since the early 1960s, the annual amplitude of the seasonal CO_2 cycle has increased by 40 % in the Arctic, likely resulting from an increase in growing season length (Keeling et al. 1996). Randerson et al. (1999) suggest that these recent high latitude amplitude increases in the seasonal CO_2 cycle are responding to growing seasons that are lengthened because of warmer springs, not warmer autumns. Time series of the satellite-derived normalized difference vegetation index (NDVI) provide additional evidence of changes in the magnitude and duration of growing seasons in the Arctic and across Eurasia, brought about by warming spring temperatures (e.g. Myneni et al. 1997; Zhou et al. 2001).

The most evident result from the precipitation records examined is a general increase in winter precipitation throughout West Siberia. It is possible that since station gauges measure rain with more accuracy than snow, a portion of the precipitation increases found during DJF could be due to a transition from snowfall events to more rain events, thus leading to erroneous observations of increasing winter precipitation. However, even for stations exhibiting the strongest increases in DJF precipitation, average air temperatures during these months are still well below freezing throughout the entire record, suggesting that a significant increase in rain events is unlikely. Although the trends found in DJF precipitation may be small in terms of magnitude, they are substantial in terms of percent increase. For instance, the observed ca. 8.3 mm/decade increase in DJF precipitation at Tarko-Sale represents a 55 % increase in DJF precipitation over the 42-year record. Similar findings of increased winter precipitation have been found over the past 20 years in northern Eurasia (Serreze et al. 2000). In West Siberia, Aizen et al. (2001) find a positive trend in annual precipitation over the past 100 years and Vinnikov et al. (1990) find increases of 14 %/100 years. Winter precipitation increases found in this study contradict the results of Fallot et al. (1997), who find no evidence of an increasing trend in cold season precipitation in the former Soviet Union over the last century. However, time series used in their study end in 1984 and thus do not capture the most recent positive anomalies in winter precipitation (Fig. 3b). Also, the majority of stations in their study fall south of 60° N, where winter precipitation is not expected to increase in response to positive AO conditions (e.g. McCabe et al. 2001). In this study, significant precipitation trends are found most commonly at higher latitudes, with 12 of the 14 significant trends occurring at stations north of 60° N. Significant precipitation trends are positive for all stations with the sole exception of Mys Kammenyj, where all records display negative trends. Mys Kammenyj also tends to be much drier than any other station (Table 1), which is likely due to the influence of relatively cold and dry Arctic maritime air masses that have not yet gained moisture through evaporation enhanced by forests and wetlands farther south on the continent (Shahgedanova 2003).

Strong association of West Siberian air temperatures with the AO suggests that the recent dominance of the positive AO phase may potentially contribute to about half (53 %; Table 3) of the annual temperature increase, which is consistent with other studies (Thompson & Wallace 1998; Rigor et al. 2000; Thompson et al. 2000). The AO is also linearly congruent with 19 % of MAM warming and may thus play a role in creating earlier dates of melt onset. In contrast, none of JJA warming is attributable to the AO, an expected result since the AO is generally considered to be a wintertime phenomenon (Thompson & Wallace 1998). The AO is also linearly congruent with 17 % of precipitation trends during winter, when statistically significant trends are more prevalent than any other season.

Recent intensification of the AO has established climatic conditions that are consistent with the strong warming and precipitation increases over West Siberia found in this study. In the past 30 years, Arctic atmospheric circulation patterns have shown large deviations from normal with a cyclonic pattern of circulation persisting over the polar region, resulting in unusually low pressure, strong subpolar westerlies, and warm high latitude temperatures over land (Serreze et al. 2000). Although the strong anticyclone circulation system known as the Siberian High is the most pronounced feature of atmospheric circulation in the lower troposphere over continental Asia during winter, it has experienced pronounced weakening during the last ca. 20 years (Gong & Ho 2002). Gong et al. (2001) show that there is a significant out-of-phase relationship between the AO and Siberian High, where the negative phase of the AO might dynamically strengthen the Siberian High and vice versa. During positive AO conditions, cyclone activity in the Northern Hemisphere shifts poleward, leading to anomalously strong subpolar westerlies, the conditions of which are characterized by westerly geostrophic surface winds along ca. 55° N and strong zonal flow extending into Europe, resulting in abnormally high surface air temperatures as far east as Siberia (Rogers 1997; Clark et al. 1999; Thompson & Wallace 2001). The position of the increased cyclone activity in Siberia favors stronger and more frequent warm, southerly winds, which are consistent with recent temperature anomalies and early melt (Serreze et al. 1995). This northward shift in cyclone activity is important as cyclones are one of the main factors setting the variance of temperature, pressure, and moisture in the troposphere on timescales of 2.5 - 10 days (Paciorek et al. 2002). The warm winter temperatures in Siberia have in fact been linked to stronger westerly flow and stronger intrusions of cyclone warm sectors into this region (Rogers & Mosley-Thompson 1995; Rogers 1997). A recent increase in cyclone activity has been observed over high latitude Northern Hemisphere regions and Eurasia, coincident with the recent increase in the AO and northward shift in Northern Hemisphere storm tracks. Paciorek et al. (2002) find an increase in winter cyclone intensity over Eurasia over the past ca. 50 years. Serreze et al. (1997) and McCabe et al. (2001) find a significant increase in high latitude cyclone activity in the Northern Hemisphere, although only for regions northward of 60° N. This is relatively consistent with the findings in this study, where 12 of the 14 statistically significant precipitation trends occur at stations north of 60° N.

The recent persistence of Arctic cyclone activity has been linked to relatively large reductions (particularly along the Siberian sector) in Northern Hemisphere sea ice cover, which may in part occur because enhanced southerly winds advect the ice poleward away from the coasts (Serreze et al. 1995; Maslanik et al. 1996). Rigor et al. (2002) suggest that these changes in sea ice may in turn be responsible for recent observed trends in surface air temperatures, by way of increased latent heat released during formation of new ice in diverging leads and increased heat flux through thinner ice. Maslanik et al. (1996) additionally suggest that because an increase in cyclone activity favors divergence and shear within the sea ice pack, resulting open water areas are increasingly heated owing to decreased albedo. These phenomena are probably most important in the winter and spring warming of coastal areas, such as at Mys Kammenyj.

Factors other than the AO are important in driving the observed temperature and precipitation trends. The strong interannual and spatial variability in precipitation may in part be due to the contribution of convection. Convective weather patterns are much more spatially variable than synoptic-scale patterns and are much less likely to correlate well with the AO. Sun et al. (2001) find a statistically significant increase in the frequency of convective cloudiness for all seasons over the FSU, which could contribute to the portion of precipitation trends not congruent with the AO. Increased convection could also contribute to observed warming through enhanced downward longwave radiation (e.g. Stone 1997), particularly during summer when the occurrence of convective cloudiness is most frequent (Sun et al. 2001). Factors other than the AO should be driving observed summer warming trends, when linear congruence with the AO is zero. June through August is termed an "inactive" season by Thompson & Wallace (2000), when the AO is much weaker and the subpolar zonal wind anomalies are displaced poleward of their wintertime position. Therefore, congruence between the AO and climate phenomena during summer months is not expected here.

The strong springtime warming and winter precipitation increases of the past few decades may have important implications for the carbon storage, flux of greenhouse gases, and hydrology in West Siberia. The ca. 40-year warming trends in springtime temperatures have undoubtedly led to an earlier occurrence of spring thaw and an increase in growing season length. The effect of spring warming on melt season length has been quantified by Rigor et al. (2000), who find that spring warming is associated with a lengthening of the melt season by 2.6 days/decade over the eastern Arctic Ocean. The timing of melt onset, and thus the duration of both the growing and the dormant seasons for plants, are strongly linked to the carbon balance of Boreal and Arctic systems (Goulden et al. 1997) and may be important for peat accumulation potentials (Nicholson et al. 1996). As there is a large range in interannual variability (up to 6 weeks or more) in the timing of freeze and thaw at a given location (e.g. Frolking et al. 1996), these events not only serve as climate indicators, but they are crucial for the understanding of possible feedback mechanisms in the carbon cycle.

Greenhouse gas exchange in peatlands is directly related to temperature and wetness conditions. Small changes in water table, temperature, or timing of thaw and senescence all have the potential to facilitate modification of CO_2 dynamics in peatlands (Bubier et al. 1998). One probable effect of climate change is that warmer spring temperatures and an earlier start to the growing season will result in increased plant production and uptake of CO_2 (Moore et al. 1998). A warming climate and associated reduction in wetness could also cause peatlands to become a significant source of CO_2 to the atmosphere (Gorham 1991, 1994; Oechel et al. 1993; Botch et al. 1995; Alm et al. 1999). In contrast, CH_4 emissions are expected to decrease with soil drying (Roulet et al. 1993; Laine et al. 1996), as its production is optimal under anaerobic waterlogged conditions (Valentine et al. 1994; Bubier 1995; Bubier et al. 1995). Under climate change scenarios, CH_4 emissions from northern peatlands are more sensitive to changes in wetness and water table position than in temperature (Roulet et al. 1992), although climate-induced changes in CH_4 emissions may be dependent on the initial position of the water table (e.g. Heikkinen et al. 2002) or additional factors such as peatland type (Nykänen et al. 1998). Given similar wetness regimes, however, CH_4 emissions are enhanced by warmer temperatures (Christensen et al. 1999) and in permafrost regions, greater active layer depths (Christensen & Cox 1995; Heyer et al. 2002). CH_4 production is also expected to be enhanced by permafrost degradation, owing to resulting warm, saturated conditions and shifts in vegetation (Bubier et al. 1995). However, from a greenhouse forcing point of view, the radiative effect of continued warming of West Siberian peatlands is difficult to predict. Because CO_2 and CH_4 dynamics are negatively coupled in wetlands (e.g. Whiting & Chanton 2001), one might expect the radiative effects of CH_4 to be tempered by those of CO_2 and vice versa. For example, earlier warming may enhance sedge production, simultaneously enhancing CO_2 uptake and CH_4 emissions (Bubier et al. 1995). Should drier conditions enhance peat decomposition and CO_2 production, Laine et al. (1996) suggest that this new source of CO_2 would be completely compensated for by decreasing CH_4 emissions and increasing tree-stand biomass storage. Although uncertainties in the specific response of northern peatlands to climate change may exist, there is no question that perturbations in temperature and precipitation have the potential to profoundly affect the exchange of atmospheric CO_2 and CH_4.

Hydrologically, an earlier transition from frozen to thawed conditions leads to the earlier occurrence of ice breakup and flooding in rivers, thus altering the timing of freshwater delivery to the Arctic Ocean. Potentially of more importance, however, is that an increase in winter precipitation may expand the snow pack water equivalence, thus increasing river discharge during the following spring and summer months. Whether evaporation losses from increased temperature will offset this increase in discharge is unclear. Peterson et al. (2002) found a 7 % increase in Eurasian river discharge over the past 70 years that is consistent with large-scale hemispheric climate patterns, likely resulting from associated warming temperatures and an increase in cyclone abundance and intensity. Although the predicted timing is uncertain, sensitivity experiments have identified a threshold of freshwater import to the Arctic Ocean at which point NADW will cease to form (e.g. Clark et al. 2002; Rahmstorf 2002). The increases in winter precipitation over West Siberia found in this study should affect the volume of freshwater input to the Arctic Ocean and may therefore have consequences of global importance.

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References

Aagaard, K. & Carmack, E. C. 1989: The role of sea ice and other fresh water in the Arctic circulation. *J. Geophys. Res. 94(C10)*, 14 485−14 498.

Aizen, E. M., Aizen, V. B., Melack, J. M., Nakamura, T. & Ohata, T. 2001: Precipitation and atmospheric circulation patterns at mid-latitudes of Asia. *Int. J. Climatol. 21*, 535−556.

Alm, J., Schulman, L., Walden, J., Nykänen, H., Martikainen P. J. & Silvola J. 1999: Carbon balance of a Boreal bog during a year with an exceptionally dry summer. *Ecology 80*, 161–174.

Ambaum, M. H. P, Hoskins, B. J. & Stephenson, D. B. 2001: Arctic Oscillation or North Atlantic Oscillation? *J. Clim. 14*, 3495–3507.

Botch, M. S., Kobak, K. I., Vinson, T. S. & Kolchugina, T. P. 1995: Carbon pools and accumulation in peatlands of the former Soviet Union. *Glob. Biogeochem. Cycles 9*, 37–46.

Brown, R. D. 2000: Northern Hemisphere snow cover variability and change, 1915–97. *J. Clim. 13*, 2339–2355.

Bubier, J. L. 1995: The relationship of vegetation to methane emission and hydrochemical gradients in northern peatlands. *J. Ecol. 83*, 403−420.

Bubier, J. L., Crill, P. M., Moore, T. R., Savage, K. & Varner, R. K. 1998: Seasonal patterns and controls on net ecosystem CO_2 exchange in a Boreal peatland complex. *Glob. Biogeochem. Cycles 12*, 703−714.

Bubier, J. L., Moore, T. R., Bellisario, L. & Comer, N. T. 1995: Ecological controls on methane emissions from a northern peatland complex in the zone of discontinuous permafrost,

Manitoba, Canada. *Glob. Biogeochem. Cycles 9*, 455–470.
Christensen, T. R. & Cox, P. 1995: Response of methane emission from Arctic tundra to climatic change: results from a model simulation. *Tellus 47B*, 301–309.
Christensen, T. R., Jonasson, S., Callaghan, T. V., Havstrom, M. & Livens, F. R. 1999: Carbon cycling and methane exchange in Eurasian tundra ecosystems. *Ambio 28*, 239–244.
Christensen, T. R., Prentice, I. C., Kaplan, J., Haxeltine, A. & Sitch, S. 1996: Methane flux from northern wetlands and tundra. *Tellus 48B*, 652–661.
Clark, M. P., Serreze, M. C. & Robinson, D. A. 1999: Atmospheric controls on Eurasian snow extent. *Int. J. Climatol. 19*, 27–40.
Clark, P. U., Pisias, N. G., Stocker, T. F. & Weaver, A. J. 2002: The role of the thermohaline circulation in abrupt climate change. *Nature 415*, 863–869.
Deser, C. 2000: On the teleconnectivity of the "Arctic Oscillation." *Geophys. Res. Lett. 27*, 779–782.
Fallot, J. M., Barry, R. G. & Hoogstrate, D. 1997: Variations of mean cold season temperature, precipitation and snow depths during the last 100 years in the former Soviet Union (FSU). *Hydrol. Sci. J. 42*, 301–327.
Frolking, S., Goulde, M. L., Wofsy, S. C., Fan, S. M., Sutton, D. J., Munger, J. W., Bazzaz, A. M., Daube, B. C., Crill, P. M., Aber, J. D., Band, L. E., Wang, X., Savage, K., Moore, T. & Harriss, R. C. 1996: Modeling temporal variability in the carbon balance of a spruce/moss Boreal forest. *Glob. Change Biol. 2*, 343–366.
Fyfe, J. C., Boer, G. J. & Flato, G. M. 1999: The Arctic and Antarctic Oscillations under global warming. *Geophys. Res. Lett. 26*, 1601–1604.
Gong, D.-Y. & Ho, C.-H. 2002: The Siberian High and climate change over middle to high latitude Asia. *Theor. Appl. Climatol. 72*, 1–9.
Gong, D.-Y., Wang, S.-W. & Zhu, J.-H. 2001: East Asian winter monsoon and Arctic Oscillation. *Geophys. Res. Lett. 28*, 2073–2076.
Gorham, E. 1991: Northern peatlands: role in the carbon cycle and probable response to global warming. *Ecol. Appl. 1*, 182–195.
Gorham, E. 1994: The future of research in Canadian peatlands: a brief survey with particular reference to global change. *Wetlands 14*, 206–215.
Goulden, M. L., Daube, B. C., Fan, S.-M., Sutton, D. J., Bazzaz, A., Munger, J. W. & Wofsy, S. C. 1997: Physiological responses of a black spruce forest to weather. *J. Geophys. Res. 102(D24)*, 28987–28996.
Groisman, P. Y., Karl, T. R. & Knight, R. W. 1994: Observed impact of snow cover on the heat balance and the rise of continental spring temperatures. *Science 263*, 198–200.
Groisman, P. Y., Koknaeva, V. V., Belokrylova, T. A. & Karl, T. R. 1991: Overcoming biases of precipitation measurement: a history of the USSR experience. *Bull. Am. Meteorol. Soc. 72*, 1725–1733.
Groisman, P. Y. & Rankova, E. Y. 2001: Precipitation trends over the Russian permafrost-free zone: removing the artifacts of pre-processing. *Int. J. Climatol. 21*, 657–678.
Heikkinen, J. E. P., Maljanen, M., Aurela, M., Hargreaves, K. J. & Martikainen, P. J. 2002: Carbon dioxide and methane dynamics in a sub-Arctic peatland in northern Finland. *Polar Res. 21*, 49–62.
Heyer, J., Berger, U., Kuzin, I. L. & Yakovlev, O. N. 2002: Methane emissions from different ecosystem structures of the subarctic tundra in western Siberia during midsummer and during the thawing period. *Tellus 54B*, 231–249.
Hurrell, J. W. 1995: Decadal trends in the North Atlantic Oscillation: regional temperatures and precipitation. *Science 269*, 676–679.
Jones, P. D., Groisman, P. Y., Coughlan, M., Plummer, N., Wang, W.-C. & Karl, T. R. 1990: Assessment of urbanization effects in time series of surface air temperature over land. *Nature 347*, 169–172.
Keeling, C. D., Chin, J. F. S. & Whorf, T. P. 1996: Increased activity of northern vegetation inferred from atmospheric CO_2 measurements. *Nature 382*, 146–149.
Kendall, M. G. 1975: *Rank correlation methods. 4th edition*. London: Charles Griffin.
Kryjov, V. N. 2002: The influence of the winter Arctic Oscillation on the northern Russia spring temperature. *Int. J. Climatol. 22*, 779–785.
Laine, J., Silvola, J., Tolonen, K., Alm, J., Nykanen, H., Vansander, H., Sallantaus, T., Savolainen, I., Sinisalo, J. & Martikainen, P. J. 1996: Effect of water-level drawdown on global climatic warming: northern peatlands. *Ambio 25*, 179–184.
Maidment, D. R. (ed.) 1993: *Handbook of hydrology.* New York: McGraw-Hill.
Mann, H. B. 1945: Non-parametric test against trend. *Econometrica 13*, 245–259.
Martin, S., Munoz, E. & Dreucker, R. 1997: Recent observations of a spring–summer warming over the Arctic Ocean. *Geophys. Res. Lett. 24*, 1259–1262.
Maslanik, J. A., Serreze, M. C. & Barry, R. G. 1996: Recent decreases in Arctic summer ice cover and linkages to atmospheric circulation anomalies. *Geophys. Res. Lett. 23*, 1677–1680.
McCabe, G. J., Clark, M. P. & Serreze, M. C. 2001: Trends in Northern Hemisphere surface cyclone frequency and intensity. *J. Clim. 14*, 2763–2768.
Moore, T. R., Roulet, N. T. & Waddington, J. M. 1998: Uncertainty in predicting the effect of climatic change on the carbon cycling of Canadian peatlands. *Clim. Change 40*, 229–245.
Myneni, R. B., Keeling, C. D., Tucker, C. J., Asrar, G. & Nemani, R. R. 1997: Increased plant growth in the northern high latitudes from 1981 to 1991. *Nature 386*, 698–702.
Nicholls, N., Gruza, G. V., Jouzel, J., Karl, T. R., Ogallo, L. A. & Parker, D. E. 1996: Observed climate variability and change. In J. T. Houghton et al. (eds.): *Climate change 1995: the science of climate change. Contribution of Working Group I to the Second Assessment of the Intergovernmental Panel on Climate Change.* Pp 137–192. Cambridge: Cambridge University Press.
Nicholson, B. J., Gignac, L. D. & Bayley, D. E. 1996: Peatland distribution along a north–south transect in the Mackenzie River Basin in relation to climatic and environmental gradients. *Vegetatio 126*, 119–133.
Nykänen, H., Alm, J., Silvola, J., Tolonen, K. & Martikainen P. J. 1998: Methane fluxes on Boreal peatlands of different fertility and the effect of long-term experimental lowering of the water table on flux rates. *Glob. Biogeochem. Cycles 12*, 53–69.
Oechel, W. C., Hastings, S. J., Vourlitis, G., Jenkins, M., Riechers, G. & Grulke, N. 1993: Recent change of Arctic tundra ecosystems from a net carbon dioxide sink to a source. *Nature 361*, 520–523.
Oke, T. R. 1987: *Boundary layer climates. 2nd edition*. London: Routledge.
Paciorek, C. J., Risbey, J. S., Ventura, V. & Rosen, R. D. 2002:

Multiple indices of Northern Hemisphere cyclone activity, winters 1949–99. *J. Clim. 15*, 1573–1590.
Panikov, N. S. 1999: Fluxes of CO_2 and CH_4 in high latitude wetlands: measuring, modeling, and predicting response to climate change. *Polar Res. 18*, 237–244.
Peterson, B. J., Holmes, R. M., McClelland, J. W., Vörösmarty, C. J., Lammers, R. B., Shiklomanov A. I., Shiklomanov, I. A & Rahmstorf, S. 2002: Increasing river discharge to the Arctic Ocean. *Science 298*, 2171–2173.
Rahmstorf, S. 1995: Bifurcations of the Atlantic thermohaline circulation in response to changes in the hydrological cycle. *Nature 378*, 145–149.
Rahmstorf, S. 2002: Ocean circulation and climate during the past 120,000 years. *Nature 419*, 207–214.
Randerson, J. T., Field, C. B., Fung, I. Y. & Tans, P. P. 1999: Increases in early season ecosystem uptake explain recent changes in the seasonal cycle of atmospheric CO_2 at high northern latitudes. *Geophys. Res. Lett. 26*, 2765–2768.
Razuvaev, V. N., Apasova, E. G., Martuganov R. A., Vose, R. S. & Steurer P. M. 1993: *Daily temperature and precipitation data for 223 USSR stations, ORNL/CDIAC NDP-040.* Oak Ridge National Laboratory, Oak Ridge, TN: Carbon Dioxide Information Analysis Center.
Rigor, I. G., Colony, R. L. & Martin, S. 2000: Variations in surface air temperature observations in the Arctic, 1979–97. *J. Clim. 13*, 896–914.
Rigor, I. G., Wallace, J. M. & Colony, R. L. 2002: Response of sea ice to the Arctic Oscillation. *J. Clim. 15*, 2648–2663.
Rogers, A. N., Bromwich, D. H. & Sinclair, E. N. 2001: The atmospheric hydrologic cycle over the Arctic Basin from reanalyses. Part II: interannual variability. *J. Clim. 14*, 2414–2429.
Rogers, J. C. 1997: North Atlantic storm track variability and its association to the North Atlantic Oscillation and climate variability of northern Europe. *J. Clim. 10*, 1635–1647.
Rogers, J. C. & McHugh, M. J. 2002: On the separability of the North Atlantic Oscillation and Arctic Oscillation. *Clim. Dyn. 19*, 599–608.
Rogers, J. C. & Mosley-Thompson, E. 1995: Atlantic Arctic cyclones and the mild Siberian winters of the 1980s. *Geophys. Res. Lett. 22*, 799–802.
Roulet, N. T., Ash R. & Quinton, W. 1993: Methane flux from drained northern peatlands: effect of a persistent water table lowering on flux. *Glob. Biogeochem. Cycles 7*, 749−769.
Roulet N., Moore, T., Bubier, J. & Lafleur, P. 1992: Northern fens: methane flux and climatic change. *Tellus 44B*, 100–105.
Sen, P. K. 1968: Estimates of the regression coefficient based on Kendall's Tau. *J. Am. Stat. Assoc. 63*, 1379–1389.
Serreze, M. C., Carse, F., Barry, R. G. & Rogers, J. C. 1997: Icelandic low cyclone activity: climatological features, linkages with the NAO, and relationships with recent changes in the Northern Hemisphere circulation. *J. Clim. 10*, 453–464.
Serreze, M. C., Maslanik, J. A., Key, J. R. & Kokaly R. F. 1995: Diagnosis of the record minimum in Arctic sea ice area during 1990 and associated snow cover extremes. *Geophys. Res. Lett. 22*, 2183–2186.
Serreze, M. C., Walsh, J. E., Chapin III, F. S., Osterkamp, T., Dyurgerov, M., Romanovsky, V., Oechel, W. C., Morison, J., Zhang, T. & Barry, R. G. 2000: Observational evidence of recent change in the northern high-latitude environment. *Clim. Change 46*, 159–207.
Shahgedanova, M. (ed.) 2003: *The physical geography of northern Eurasia*. Oxford: Oxford University Press.
Shindell, D. T., Miller, R. L., Schmidt, G. & Pandolfo, L. 1999: Simulation of recent northern winter climate trends by greenhouse-gas forcing. *Nature 399*, 452–455.
Smith, L. C. 2000: Trends in Russian Arctic river-ice formation and breakup, 1917 to 1994. *Phys. Geogr. 21*, 46–56.
Stone, R. S. 1997: Variations in western Arctic temperatures in response to cloud radiative and synoptic-scale influences. *J. Geophys. Res. 102(D18)*, 21769−21776.
Sun, B., Groisman, P. Y. & Mokhov, I. I. 2001: Recent changes in cloud-type frequency and inferred increases in convection over the United States and the former USSR. *J. Clim. 14*, 1864−1880.
Thompson, D. W. J. & Wallace, J. M. 1998: The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett. 25*, 1297–1300.
Thompson, D. W. J. & Wallace, J. M. 2000: Annular modes in the extratropical circulation. Part I: month-to-month variability. *J. Clim. 13*, 1000–1016.
Thompson, D. W. J. & Wallace, J. M. 2001: Regional climate impacts of the Northern Hemisphere annular mode. *Science 293*, 85–89.
Thompson, D. W. J., Wallace, J. M. & Hegerl, G. C. 2000: Annular modes in the extratropical circulation. Part II: trends. *J. Clim. 13*, 1018–1036.
Valentine, D. W., Holland, E. A. & Schimel, D. S. 1994: Ecosystem and physiological controls over methane production in northern wetlands. *J. Geophys. Res. 99(D1)*, 1563−1571.
Vinnikov, K. Y., Groisman, P. Y. & Lugina, K. M. 1990: Empirical data on contemporary global climate changes (temperature and precipitation). *J. Clim. 3*, 662−677.
Volodin, E. M. & Galin, V. Y. 1998: Sensitivity of midlatitude Northern Hemisphere winter circulation to ozone depletion in the lower stratosphere. *Russ. Meteorol. Hydrol. 8*, 23–32.
Vörösmarty, C. J., Hinzman, L. D., Peterson, B. J., Bromwich, D. H., Hamilton, L. C., Morison, J., Romanovsky, V. E., Sturm, M. & Webb, R. S. 2001: *The hydrologic cycle and its role in Arctic and global environmental change: a rationale and strategy for synthesis study.* Fairbanks, Alaska: Arctic Research Consortium of the U.S.
Wallace, W. J. & Solomon, S. 2002: Interpretation of recent Southern Hemisphere climate change. *Science 296*, 895–899.
Walsh, J. E., Chapman, W. L. & Shy, T. L. 1996: Recent decrease of sea level pressure in the central Arctic. *J. Clim. 9*, 480–486.
Whiting, G. J. & Chanton, J. P. 2001: Greenhouse carbon balance of wetlands: methane emission versus carbon sequestration. *Tellus 53B*, 521–528.
Woo, M., Heron, R., Marsh, P. & Steer, P. 1983: Comparison of weather station snowfall with winter snow accumulation in High Arctic basins. *Atmosphere–Ocean 21*, 312–325.
Yang, D. Q. & Ohata, T. 2001: A bias-corrected Siberian regional precipitation climatology. *J. Hydrometeorol. 2*, 122–139.
Zhou, L., Tucker, C. J., Kaufmann R. K., Slayback, D., Shabanov, N. V. & Myneni, R. B. 2001: Variations in northern vegetation activity inferred from satellite data of vegetation index during 1981 to 1999. *J. Geophys. Res. 106(D17)*, 20 069–20 083.
Zhulidov, A. V., Headley, J. V., Robarts, R. D., Nikanorov, A. M. & Ischenko, A. A. 1997: *Atlas of Russian wetlands.* Saskatoon: National Hydrology Institute, Environment Canada.