Climate change in the Arctic: causes and mechanisms

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Abstract. Atmospheric heat and moisture transfers from the North Atlantic make the main contribution to the Arctic warming in winter. The increase in transfer is associated with changes in atmospheric circulation in the Northern Hemisphere under the influence of the sea surface temperature (SST) in low latitudes, where the bulk of the heat influx from the Sun accumulated. The mechanism of influence includes the interaction between the circulation of the ocean and atmosphere, which enhances the oceanic heat influx into the Norwegian and Barents seas, and atmospheric transport to the Arctic. SST rises with participation of orbital-forced increase of the solar insolation. Changes in insolation are small, but their effect is enhanced by feedbacks between temperature, water vapour content and downward long-wave radiation in low latitudes. An increase in water inflow, heat and moisture transfers to the Atlantic Arctic lead to increase in air temperature, water vapour content, downward long-wave radiation, and a reduction of ice thickness growth and its extent in the Barents and Greenland Seas in winter.

1. Observed changes

The Arctic is around Pole part of the North Polar Area, but its southern boundary is defined in different ways [1]. Of particular interest is the region of the marine Arctic, including the Arctic Ocean and adjacen seas covered by ice in winter. Changes of air temperature in this area affect the winter growth and summer melting of the sea ice. To assess this influence the data from 41 stations located on the islands and the coast of the Arctic Ocean and adjacent seas (figure. 1.1b) was used [2]. The average winter and summer surface air temperatures (SAT) at these stations from 1951, are shown in figure 1.1a and 1.1c. The curves in figure 1 show a rapid decrease in negative temperatures after 1998 and an increase in positive temperatures after 1996.

To compare warming in the Arctic in relation to the other regions of the Earth, the ratio of trends in surface air temperature is used [3]. As a result, there is a manifold excess of the warming rates in the Arctic, called the "Arctic amplification." The value of the amplification depends on the size of the compared areas and on season. Most often, the trends in average annual temperature in the region north to 60° N are compared with trends in average annual temperature in the Northern Hemisphere or around the globe (Table 1).

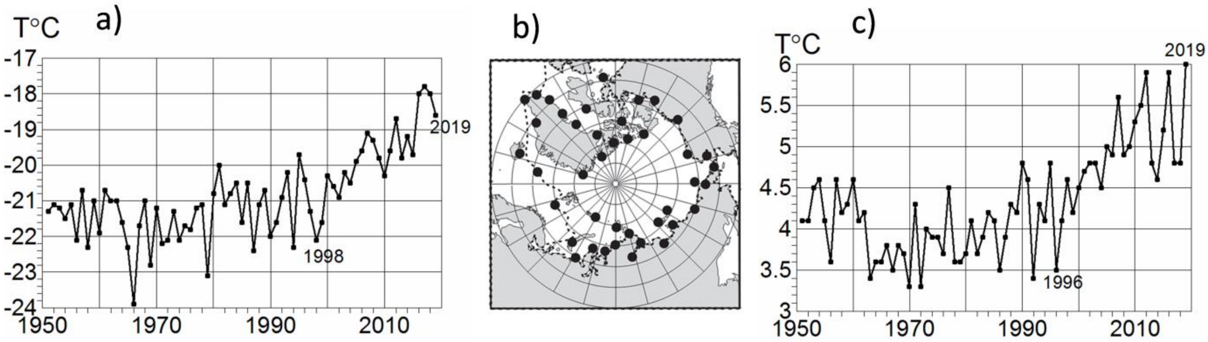

Figure 1. Average surface air temperature in winter (DJF, a) and in summer (JJA, c) at meteorological stations in the marine Arctic b) for 1951–2019.

Table 1. The ratio of the SAT trends in the region of 60–90° N to trends in the Northern Hemisphere and in the Globe for 1989–2018 on average for year and seasons. SAT trends calculated from the ERA Interim 2m reanalysis data.

Region averaging	Annual	Winter	Spring	Summer	Autumn
Northern Hemisphere	2.91	4.26	2.28	1.43	3.25
Globe	4.32	6.76	3.22	2.10	4.74

The amplification of warming in the marine Arctic is much stronger (Table 2) compared with the region of 60–90° N.

Table 2. The ratio of the SAT trends in the marine Arctic to the trends in the Northern Hemisphere and in the Globe for 1989–2018 on average for year and seasons. SAT trends calculated from the ERA Interim 2m reanalysis data.

Region averaging	Annual	Winter	Spring	Summer	Autumn
Northern Hemisphere	4.05	6.28	3.13	1.99	4.05
Globe	6.02	9.97	4.42	2.93	5.91

If we compare the temperature trends in the circumpolar regions with the trends in the adjacent parts of the Northern Hemisphere or the globe, then the effect of air exchange between the regions on the ratio of trends is revealed [4]:

$$\frac{a_A}{a_{nA}} = \frac{T_A}{T_{nA}} \frac{S_{nA}}{S_A},$$

where a_A, a_{nA} – trend coefficients of temperature in the Arctic and in the surrounding area, T_A ,T_{nA} – mean temperatures, S_A, S_{nA} – square of areas.

In this case the Arctic amplification is greater (Table 3) then the amplification in Table 1.

Table 3. The ratio of the SAT trends in the region of 60–90° N to SAT trends in the region of 0–60° N for 1989–2018 on average for year and seasons. SAT trends calculated from the ERA Interim 2m reanalysis data.

Averaged for	Annual	Winter	Spring	Summer	Autumn
Ratio	4.16	8.79	2.9	1.52	5.03

The effect of the air exchange between high and adjacent latitudes is seen from negative correlation between their average temperatures (Table 4). This correlation amplified with the warming.

Table 4. Correlation coefficients between deviations from the SAT trend in February 1979–2018 in the Arctic and adjacent middle latitudes. Significant coefficients at 95% level are bold.

North latitudes,°	Period	North latitudes,°				
		60	55	50	45	40
	1979-2018	-0.08	**-0.38**	**-0.46**	**-0.43**	**-0.39**
80	1979-1998	0.10	-0.21	-0.34	-0.36	-0.42
	1999-2018	-0.30	**-0.53**	**-0.54**	**-0.47**	-0.38
	1979-2018	0.03	**-0.41**	**-0.52**	**-0.47**	**-0.43**
75	1979-1998	0.29	-0.22	**-0.46**	**-0.46**	**-0.49**
	1999-2018	-0.35	**-0.60**	**-0.58**	**-0.49**	-0.39
	1979-2018	0.42	-0.15	**-0.37**	**-0.34**	**-0.32**
70	1979-1998	0.59	-0.00	-0.32	-0.31	-0.33
	1999-2018	0.07	-0.35	-0.44	-0.40	-0.34

Warming is accompanied by the Arctic sea ice extent reduction which has a good agreement with summer temperature increase (figure 2).

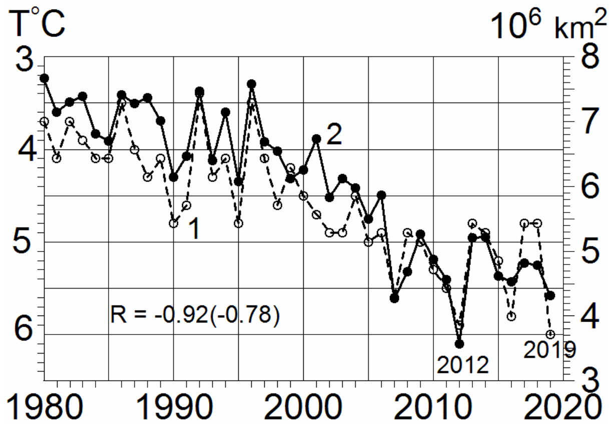

Figure 2. Summer surface air temperature in the marine Arctic (1) and September sea ice extent in the Arctic (2). R – correlation between (1) and (2). In brackets – without quadratic trend.

2. Reasons of warming and Arctic amplification

The largest contribution to climate warming in the Arctic compared with the climate that would be observed under conditions of radiation equilibrium in the static atmosphere and ocean is made by advection of heat towards the pole as a result of atmospheric and ocean circulation. Due to these processes, the Arctic climate "warms up", air temperature increases by several tens of degrees compared with the climate in the absence of advection, and greenhouse effect increases the surface temperature in the Arctic by a much smaller amount [5]. During the polar night, when there is no heat influx from the Sun, the fluctuations of average surface air temperature in the polar region are determined by the fluctuations of advection [6].

Meridional atmospheric heat and moisture transfers to the Arctic calculated from ERA Interim reanalysis data [7] showed that the main influx of sensible and latent heat into the high-latitude Arctic in winter comes through the Atlantic part (from 0 ° to 80 ° E) of 70° N circle in a layer from the surface to 750 hPa. The contribution of this influx to the interannual variability of the average winter air temperature at the surface is more than 50% and its value increases. Recently Cao et al, [8] showed that an increase in the greenhouse effect caused by water vapor in winter, slows down ice growth and accelerates the beginning of melting.

In summer, the main contribution to warming made by radiative heat influxes to the surface, including long-wave radiation, which increases because of an increase in the water vapor content in the atmosphere with snow and ice melting and the appearance of open water. The reduction in ice extent in the summer months due to an increase in water vapor content and an increase in downward long-wave radiation is up to 40% [7, 9].Under this moisture transfer through 70° N does not affect on the vapor content in the lower troposphere, where outflow from the Arctic prevails [7].

2.1. Influence from low latitudes

Recent studies [10] showed that the source of oceanic and atmospheric heat inflows into the Arctic is located in low latitudes (figure 3), where SST anomalies are formed, which appear after 2–3 years in heat influx to the Arctic, variations of SAT and sea ice extent in the Arctic Ocean.
The influx of warm and salty water from the North Atlantic to the Barents and Greenland Seas affects also to the warming in the Arctic and the reduction of sea ice extent. The close relationship between the temperature of the water coming from the North Atlantic to the Barents Sea and the ice extent during the period of ice growth from December to the beginning of melting in June (Table 5) was confirmed in work [11]. Interannual changes in the oceanic heat influx from the North Atlantic well reflected in fluctuations of water temperature at the Kola section in the Barents Sea [12].

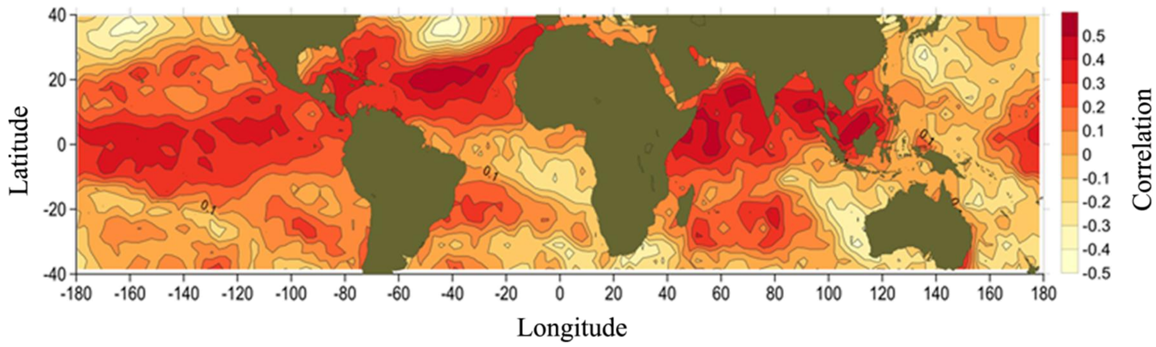

Figure 3. Correlation between SST anomalies in 40°S–40°N zone in October and anomalies of the meridional atmospheric transport of sensible heat in December–February through the Atlantic gate in 70°N on 27 months later for 1982–2015.

Table 5. Correlation between the water temperature in a layer of 50–200 m at the Kola section and the ice extent in the Barents Sea 1979–2014. Bold figures are |70| and more.

Month	1	2	3	4	5	6	7	8	9	10	11	12
correlation coefficient	**–0.83**	**–0.82**	**–0.70**	**–0.78**	**–0.87**	**–0.83**	–0.67	–0.48	–0.26	–0.28	–0.44	**–0.70**

The correlation coefficient between the water temperature in the Kola section and the ice extent in the Arctic Ocean in May is -0.92 (-0.83). In brackets there is a correlation coefficient after the trend is removed.

2.2. Warming mechanism

The interannual fluctuations of winter atmospheric heat transports to the Arctic are delayed relatively SST anomalies in low latitudes from 23 to 30 months [10]. The previous studies show the transfer of low-latitude SST influence to the Arctic by the means of atmospheric circulation with a delay of 1–2 weeks [13, 14]. Delays of several years can occur only with participation of the ocean circulation.

SST anomalies affect to atmospheric circulation modes and the ocean circulation system in the North Atlantic. They reinforce the Headley and Ferrel atmospheric circulation cells, increase the meridional atmospheric transport, weaken the NAO, which contributes to a decrease in the heat loss by the ocean and all together increase the oceanic heat transport by the Gulfstream, North Atlantic, West Spitsbergen and Norwegian currents. The final link is the increase of SST in the Norwegian and Barents Seas and atmospheric transport to the Arctic (figure 4). At the negative NAO phase, the zonal component of the wind over the North Atlantic is weakened and positive SST anomaly is formed (figure 4, a). With a positive NAO index, the zonal wind amplified with a maximum on 50° N, which contributes to the cooling of the ocean and the formation of a negative SST anomaly (figure 4 b). Hoerling et al. [15] earlier already noted a negative relationship between SST anomalies at low latitudes and the NAO index.

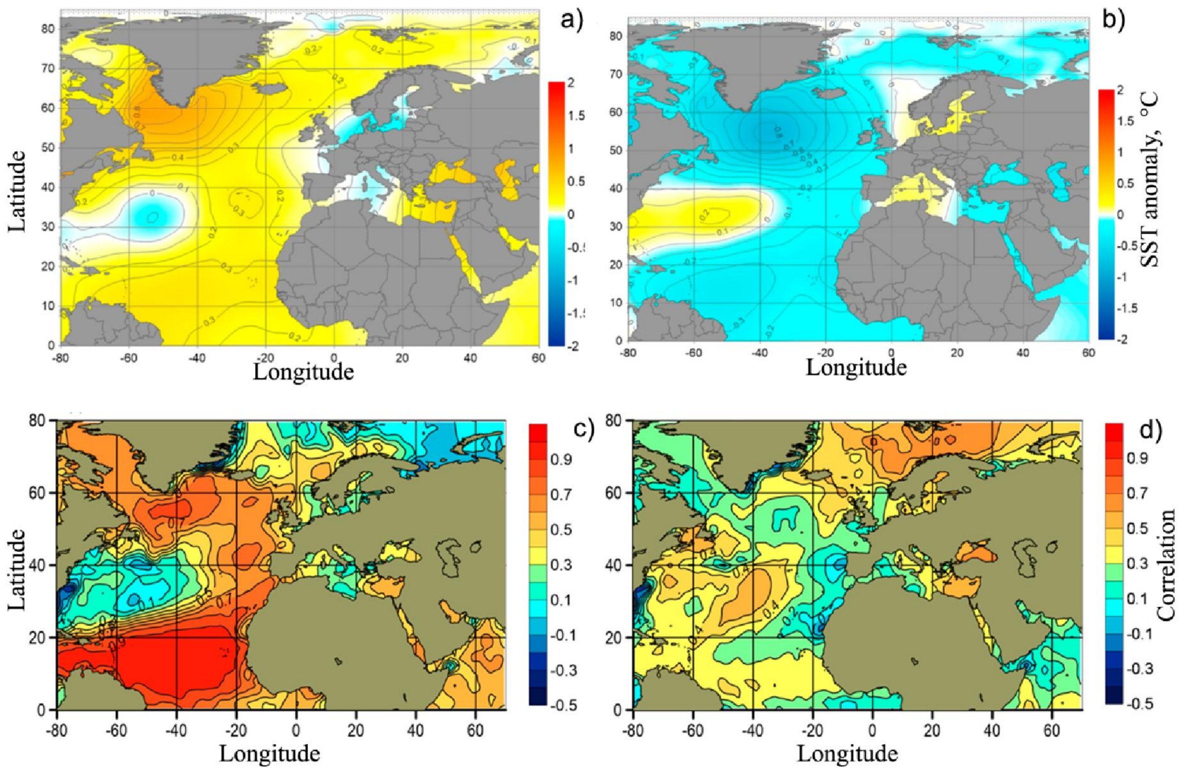

Figure 4. Influence of SST anomalies in low latitudes of the North Atlantic on the atmosphere and ocean to the north.

Upper panel: composites of annual SST anomalies for large (> σ) anomalies of mean annual NAO indexes in 1950–2015: (a) – for negative NAO indexes, (b) – for positive NAO indexes.

Low panel: the distribution of the correlations between the anomalies of mean annual SST averaged in the tropical region of the North Atlantic and the anomalies of mean annual SST in every grid point of entire area in 1980–2015: (c) – synchronous correlation, (d) – with a delay of 3 years relative to SST in the tropics. SST data from HadlSST [16]

2.3 The role of insolation

Climatic areas with maximal SAT, maximal SST, maximal water vapor content and maximal influx of solar insolation are situated in low latitudes (figure 5). The main part of the heat absorbed by the ocean accumulated here [17].

The increase concentration of carbon dioxide and other greenhouse gases in the atmosphere, according to calculations at global climate models, is the main cause of the global temperature rise [18]. Slow changes in insolation caused by astronomical reasons also can influence on increase of SST, total content of water vapor (TCWV) and SAT in low latitudes. These changes are small, but their effect enhanced by the accumulation of the heat in the ocean, which occupied most of the low latitude area and has low surface albedo. Effect enhanced also by feedbacks between temperature, water vapor content and downward long-wave radiation (DLR).

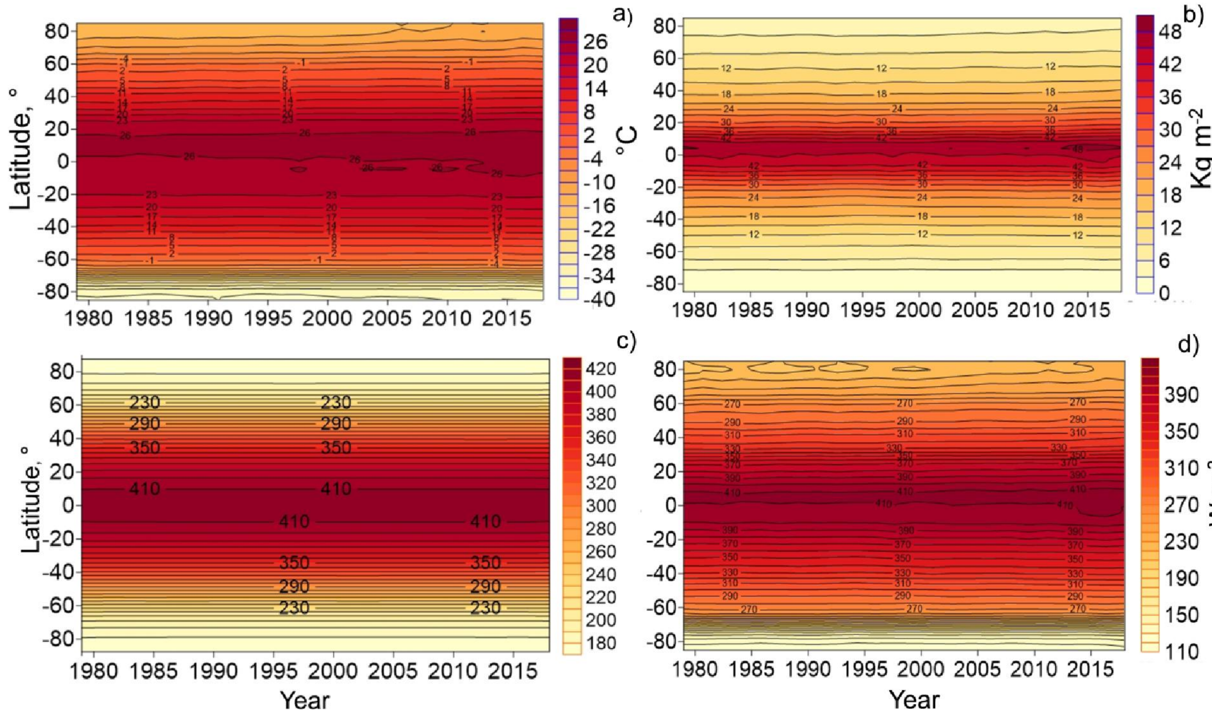

Figure 5. Zonal values of mean annual atmospheric parameters.
(a) – surface air temperature; (b) – total content of water vapor; (c) – insolation on top of atmosphere on base data from [20]; (d) – downward longwave radiation on surface. (a), (b), (d) – on base ERA5 data [21].

In order to evaluate the contribution of insolation, we used the calculations of A.A. Kostin and V.M. Fedorov [20] where is presented insolation of the globe surface without the atmosphere from 3000 BC until 2999 AD. They noted increase in incoming solar radiation to the equatorial regions of the Earth and a decrease in the polar regions [22].

To compare insolation and parameters of the atmosphere calculated according to ERA5 data [21], we used data from 1979 to 2018. The average values of insolation and atmospheric parameters in low latitudes were calculated in the region of 0–25° N for every month, season and year. The highest increase of insolation for 1979-2018 is found in the spring and in average for the year is noted weak growth (Table 6).

Table 6. Trends of mean insolation in 0–25°N ($Wm^{-2}yr^{-1}$)×100 for 1979–2018.

Month	January	February	March	April	May	June	July
Trend	0.056	0.23	0.344	0.422	0.398	0.26	0.04
Month	August	September	October	November	December	Spring	Year
Trend	-0.19	-0.359	-0.388	-0.282	-0.109	0.352	0.025

A strongest correlation found between insolation in spring and SAT, SST (figure 6 a, b) and other parameters (water vapor content, downward long-wave radiation) in autumn. This relationship explained by the cumulative effect of the insolation positive trend in spring-summer months , which forms the autumn maximum of SST trend in low latitudes. The calculations of spectral density and coherence shown that a high correlation between insolation and SAT is provided by periods of more than 16 years (more than 50% of dispersion). Periods of less than 3 years, which removed from the temperature series by smoothing with 3 years window, contributed about 10% of the variance.

Temperature changes in the region of 0–25° N in autumn appear in winter SAT changes in the region of 70–87.5° N three years later (after 27 months for the central months of the seasons) (figure 6 c, d).

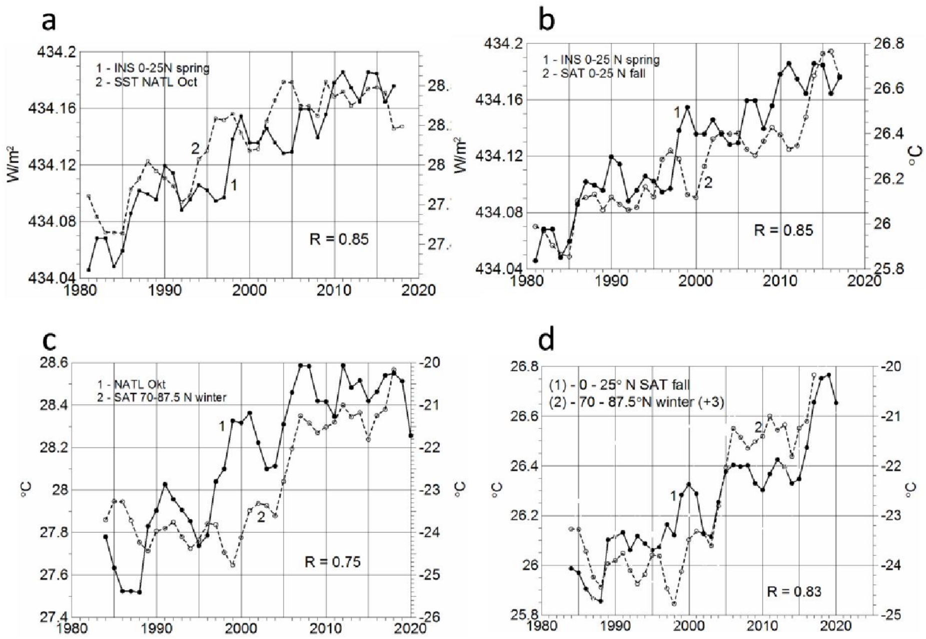

Figure 6. Change of insolation, surface air temperature and sea surface temperature.
(a) – insolation in spring in 0–25°N and SST in October in the NATL area (5–20°N, 60–30°W) [http://www.cpc.ncep.noaa.gov/data/indices]; (b) – insolation in spring and SAT in fall in 0–25°N area; (c) – SST in NATL area in October and SAT in winter in 70–87.5°N area 27 months later; (d) – SAT in fall in 0–25°N and in winter in 70–87.5°N on 27 months later. SAT were smoothed by 3 year window. SAT in 0–25°N lead 3 year SAT in 70–87.5°N. R – correlation coefficient between 1 and 2.

According to the Figure 6 more than 70% of the variability of autumn SST and SAT at low latitudes associated with interannual changes in insolation. In turn, changes in temperature at low latitudes are responsible for more than 50% of the variability of winter SAT at high latitudes of the Northern Hemisphere. If we compare the non-smoothed values, then the estimates will be about 55 and 45%, respectively.

Another estimate of the insolation effect on changes of SAT and SST in low latitudes can obtained from the analysis of the influx of solar radiation on the ocean surface. From Table 6 we find an estimate of the cumulative influx of insolation from January to September, which ensures the accumulation of heat in the near-surface layer of the ocean until the autumn maximum of SST trend in October. We summarize the monthly trend estimates and obtain an increase in insolation influx to the ocean surface (ΔI_Σ), part of which is absorbed in the surface layer with depth (h) in the area S and increases the layer heat storage by ΔI_h:

$$\Delta I_h = \alpha_h(1-A)\Delta I_\Sigma\, S\, \tau = C_p \rho\, \Delta T\, h\, S,$$

from which we find the increase in SST caused by an increase in insolation:

$$\Delta T_I = \frac{\alpha_h(1-A)\Delta I_\Sigma\, \tau}{C_p \rho h}. \quad (1)$$

Where $\alpha_h -$ part of insolation absorbed in the surface water layer with depth h (m), A is the albedo of the ocean surface, C_p, ρ – heat capacity and density of water, S – area, and τ – number of seconds in a month.

Assuming $A = 0.1$, $C_p = 4022$ J kg^{-1} K^{-1}, $\rho = 1028$ kg m^{-3}, $\alpha_1 = 0.55$, $\Delta I_\Sigma = 0.01201$ Wm^{-2} yr^{-1} (based on Table 6), we obtain $\Delta T_I = 0.00373$ K yr^{-1} for the "insolation" trend of SST in October. The observed SST trend in October, according to the data in Table 7, is 0.023, which is 6 times more than ΔT_I. But SST also increases as a result of the influx of DLR caused by an increase in the content of water vapour and the outgoing longwave radiation.

The outgoing longwave radiation increment corresponding to ΔT_I will $\Delta I_U = 4\varepsilon\sigma T_0^3\, \Delta T_I$, $\Delta I_U =$ 0.0208 W m^{-2} yr^{-1} for $\sigma = 5.67\ 10^{-8}$ W m^{-2} K^{-4}, $\varepsilon = 0.9$, $T_0 = 301.15$ K (Table 7). The I_U trend in October (Table 7) is 0.13738 W m^{-2} yr^{-1}, and the trend $I_D = 0.20506$ W m^{-2} yr^{-1}

Table 7. Mean values and trends of longwave radiation, atmosphere and ocean parameters in NATL area (5–25°N, 60–20°W) for 1979–2018.

Month,	SAT, °C		TCWV, kg m^{-2}		STRU, Wm^{-2}		STRD, Wm^{-2}		SST, °C	
season	Mean	Trend	Mean	Trend	Mean	Trend	Mean	Trend	Mean	Trend
Jan	23.96	0.0156	29.12	0.0158	446.89	0.0764	381.41	0.1569	26.05	0.0152
Feb	23.60	0.0076	27.60	0.0474	443.96	0.0368	378.15	0.1896	25.62	0.0112
March	23.61	0.0085	27.94	0.0043	443.90	0.0355	378.60	0.1236	25.62	0.0085
April	24.01	0.0054	31.09	-0.0004	445.99	0.0237	384.68	0.0839	25.96	0.0082
May	24.58	0.0083	35.51	-0.0176	449.19	0.0341	392.69	0.0821	26.41	0.0066
June	25.16	0.0060	38.62	-0.0060	452.71	0.0203	399.77	0.0875	26.81	0.0054
July	25.60	0.0114	40.60	0.0298	455.95	0.0622	404.84	0.1387	27.25	0.0107
Aug	26.12	0.0157	42.93	0.0571	459.72	0.0919	408.80	0.1824	27.81	0.0154
Sept	26.42	0.0172	43.39	0.0646	462.42	0.1076	409.67	0.1819	28.21	0.0206
Oct	26.27	0.0225	41.90	0.0673	462.26	0.1374	406.78	0.2051	28.15	0.0230
Nov	25.67	0.0192	38.83	0.0428	458.52	0.1148	399.99	0.1711	27.66	0.0186
Dec	24.84	0.0164	33.15	0.0271	452.67	0.0878	389.86	0.1351	26.86	0.0182
Year	24.99	0.0128	35.89	0.0277	452.85	0.0691	394.60	0.1448	26.87	0.0135
Winter	24.13	0.0145	30.01	0.0373	447.85	0.0768	383.22	0.1786	26.18	0.0167
Spring	24.07	0.0074	31.51	-0.0046	446.36	0.0311	385.32	0.0965	26.00	0.0078
Summ	25.63	0.0110	40.72	0.0269	456.13	0.0581	404.47	0.1362	27.29	0.0105
Full	26.12	0.0197	41.37	0.0582	461.07	0.1199	405.48	0.1860	28.01	0.0207

Here STRU –surface thermal radiation upward, STRD – downward.

Assuming that the ratio of downward and upward radiation trends tr (I_D)/tr (I_U) = 1.46 is also true for $\Delta I_D/\Delta I_U$ increments, we found that the increment $\Delta I_D = 0.0304$ Wm^{-2} yr^{-1} in October. According to the equation (1), this increase will lead to increase in SST by ΔT_D = 0.0094 K yr^{-1}. Total SST increment $(\Delta T_I + \Delta T_D)$ = 0.01315 K yr^{-1} = 57% of the observed SST trend in October (Table 7). If we consider that the STRD albedo is less than 0.1, and the absorption in the near-surface layer is greater than for the incoming insolation, the impact of insolation heating will increase.

The calculated heating from insolation followed by an increase in the total content of water vapor:

$$\Delta Q_I = \frac{Q_8 - Q_2}{T_8 - T_2} \times (\Delta T_I + \Delta T_D), \quad (2)$$

where Q_8, Q_2 is the total water vapor content in August and February, T_8, T_2 – air temperature in August and February.

Using data from Table 7 and estimating of insolation heating, we obtain ΔQ_I = 0.0731 kg m^{-2} yr^{-1}. This value exceeds the trend Q_{10} = 0.0673 kg m^{-2} yr^{-1}, which may be the result of an overestimated sensitivity by equation (2).

3. Conclusions

An analysis of the studies performed earlier and fulfilled in this work shown that:

The warming in the marine Arctic accelerated in the late 1990s.

The index of Arctic amplification depends on the size of compared areas, average air temperature and season.

The main contribution to winter warming in the Arctic belongs to the atmospheric and oceanic influx of heat and moisture mostly from the Atlantic, which also affect to the reduction of the extent and thickness of ice.

The sources of atmospheric and oceanic heat for the Arctic locate in low latitudes, where the heat of solar radiation accumulated.

Maximal trends of SAT, SST and TCWV are formed here in the autumn at more than 40% under influence of positive insolation trends in spring and early summer months. In turn, autumn trends in the tropics affect to winter trends in the Arctic.

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