Amplified winter Arctic tropospheric warming and its link to atmospheric circulation changes

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ABSTRACT

The physical cause of amplified deep Arctic tropospheric warming in winter in the Barents–Kara Seas (BKS) is examined. The authors propose that changes in the atmospheric circulation patterns are important for deep Arctic tropospheric warming in winter. It is found that the retrograde Urals blocking (UB) event concurrent with a negative North Atlantic Oscillation (NAO−) that arises from a prior negative Arctic Oscillation (AO −) is not favorable for tropospheric warming because of less water vapor over the BKS. Such UB events are related to more winter BKS sea ice associated with the negative sea surface temperature (SST) anomaly in the BKS. In contrast, a UB occurring together with a positive North Atlantic Oscillation (NAO+) shows less movement and can significantly enhance tropospheric warming over the BKS through increasing tropospheric sensible heat energy due to a persistent BKS water vapor increase. This type of quasi-stationary UB event is related to prior less BKS sea ice associated with a positive BKS SST anomaly that coexists with the North Atlantic SST tripole structure. In summary, because warm, wet and low sea-ice winters in the BKS are related to UB events with an NAO+, and depend on the winter prior sea-ice condition, the tropospheric warming is to some extent a manifestation of the sea-ice–blocking–moisture feedback in the BKS.

1. Introduction

In recent decades, the Arctic temperature in the troposphere, especially at the Earth’s surface, has increased more rapidly than in other regions outside the Arctic, which is referred to as Arctic amplification (Johannessen et al. 2004; Gillett et al. 2008; Grant, Brönnimann, and Haimberger 2008; Graversen et al. 2008; Thorne 2008). The underlying cause of the Arctic amplification is an important research topic (Graversen et al. 2008; Screen and Simmonds 2010; Chung and Räisänen 2011; Perlwitz, Hoering, and Dole 2015).

Various mechanisms have been proposed to explain the amplified Arctic warming in recent decades. For example, many studies have revealed that the Arctic warming is likely due to anthropogenic forcing (Gillett et al. 2008), surface-ice–albedo feedback (Holland and Bitz 2003; Screen and Simmonds 2010), and changes in cloud cover, atmospheric water vapor and increased poleward heat and moisture transport (Winton 2006; Graversen et al. 2008; Graversen and Wang 2009; Woods, Caballero, and Svensson 2013, 2016). A recent study by Burt, Randall, and Branson (2016) also revealed that positive ice–insulation feedback plays an important role in winter Arctic warming. However, it is not clear what has led to the recent winter amplification of Arctic tropospheric warming, although the atmospheric response to the ice decline in the Barents–Kara Seas (BKS) is a Urals blocking (UB) with a positive North Atlantic Oscillation (NAO+) (Gong and Luo 2017; Luo, Xiao, Diao et al. 2016; Luo, Xiao, Yao et al. 2016). Some studies have also indicated that the Arctic warming
is related to the intrusion of warm ocean water into the Arctic (Lien et al. 2017) and sea surface temperature (SST) anomalies in the Arctic (Perlwitz, Hoerling, and Dole 2015). This perhaps hints toward the intensified Arctic warming not only depending on changes in atmospheric circulation, but also to the basic Arctic sea-ice state associated with SST change in the Arctic. In this paper, we propose a new viewpoint that the amplified deep Arctic warming in winter in the BKS is a consequence of the interaction between the background sea ice linked to the SST change and the occurrence of a UB with an NAO⁺.

2. Data and method

This study uses daily data obtained from the ERA-Interim data-set (Dee et al. 2011) on a 2.5° × 2.5° grid in winter (December–January–February; DJF hereafter) during the period from December 1979 to February 2016 (1979–2015 hereafter). These daily data include sea-ice concentration (SIC), SST, multi-level geopotential height, air temperature, specific humidity, surface air temperature (SAT), total column water vapor (TCWV), and downward infrared radiation (IR). The daily anomalies at each grid point are defined as the deviation from its long-term mean for that day, and are linearly detrended.

The one-dimensional blocking index of Tibaldi and Molteni (1990; TM index hereafter), which is constructed based on the reversal of the meridional gradient of the 500-hPa geopotential height at three given reference latitudes, is used to identify UB events confined to the longitudinal zone of 40°–80°E. Details of the TM index definition can be found in Luo, Xiao, Yao et al. (2016). Linear regression is used to obtain the SIC, SAT, and 500-hPa geopotential height anomalies, and their statistical significance is tested using the Student’s t-test and a Monte Carlo simulation.

3. Results

3.1. Regional features of Arctic tropospheric warming

As noted by many investigators (Fang and Wallace 1994; Deser, Walsh, and Timlin 2000; Luo, Xiao, Diao et al. 2016; Luo, Xiao, Yao et al. 2016), the decline of the sea ice in the BKS in winter is most pronounced relative to other regions. As such, it is useful to examine the linear trends of tropospheric temperature anomalies over the whole Arctic and BKS in winter (DJF), spring (March–April–May; MAM), summer (June–July–August; JJA), and autumn (September–October–November; SON) (Figure 1). It is found that the tropospheric warming in the Arctic is strongest in winter relative to other seasons (Figure 1(a)). This result is consistent with the findings of Graversen et al. (2008) and Screen and Simmonds (2010). However, we can see that the winter Arctic tropospheric warming exists mainly in the BKS (Figure 1(b)). Moreover, the Arctic warming can penetrate deeper to the middle level of the troposphere in winter than in other seasons. Clearly, such winter Arctic tropospheric warming cannot be explained by local evaporation, because the role played by evaporation is mainly

![Figure 1](image-url)

**Figure 1.** Height–latitude profiles of temperature trends during 1979–2015 averaged over (a) the whole Arctic and (b) the Barents–Kara Seas for winter (December–January–February), spring (March–April–May), summer (June–July–August), and autumn (September–October–November). The black solid contour line indicates the region where the trends are statistically significant above the 95% confidence levels for a two-sided Student’s t-test.
confined to the lower troposphere (Figure 1(b)). This suggests that other atmospheric processes in winter may be important for winter tropospheric warming.

### 3.2. Link between the BKS warming and atmospheric circulation patterns

To explain why the winter Arctic warming can reach the middle level of the Arctic troposphere, it is first necessary to choose a region (here, 600–400 hPa) to represent this level of the atmosphere. Accordingly, we show the time series of domain-averaged winter SIC, SAT, and tropospheric mid-level air temperature (MLAT) anomalies in the region (70°–85°N, 600–400 hPa) in Figure 2(a–c). Interestingly, the SAT anomaly in the BKS (Figure 2(b)) exhibits a high positive correlation of 0.87 with the MLAT anomaly (Figure 2(c)) for a detrended case, while it has a high opposite correlation of −0.85 with the SIC in the BKS. We also see that the BKS

![Figure 2](image_url)

**Figure 2.** Time series of domain-averaged winter (December–January–February)-mean (a) sea-ice concentration (SIC) and (b) surface air temperature (SAT) anomalies in the Barents-Kara Seas (70°–85°N, 20°–80°E; red line), and (c) the tropospheric mid-level (70°–85°N, 600–400 hPa) mean air temperature (MLAT) anomaly for non-detrended (red line) and detrended (blue line) cases. Linear trends of (d) SIC and (e) SAT anomalies during 1979–2015. (f) Regressed fields of 500-hPa geopotential height (contours) and SAT (color shading) anomalies against the SIC time series in the BKS. Here, $r_{\text{SAT, SIC}} = -0.85$, $r_{\text{MLAT, SIC}} = -0.69$, and $r_{\text{MLAT, SAT}} = 0.87$ represent the correlation coefficients among the SAT, MLAT, and SIC. In (a), the black line denotes the linear trend. In (d–f), the color shading indicates the statistically significant region above the 95% confidence levels for a two-sided Student’s $t$-test.
SAT shows a strong warming after 2000 (Figure 2(b)), and corresponds to a large decline in the BKS sea ice (Figure 2(a)). The regression further shows that the atmospheric response to the BKS sea-ice decline is characterized by a UB pattern with an NAO+ (Figure 2(f)). This implies that a UB with an NAO+ may play an important role in amplified tropospheric warming in the BKS. Next, we examine whether the mid-level warming in the BKS results from the interaction between the BKS ice decline and the blocking pattern with an NAO+.

To understand the role of the UB, we first define the normalized value of the domain-averaged, winter-mean detrended MLAT anomaly (70°–85°N, 600–400 hPa; blue line) having a ±0.5 (−0.5) standard deviation as a strong (weak) tropospheric warming – referred to as a warming (cooling) winter hereafter. It is easy to identify that the tropospheric warming winters are 1979, 1983, 1984, 1989, 1994, 2004, 2005, 2007, 2008, 2011, and 2013, whereas the tropospheric cooling winters correspond to 1981, 1985, 1986, 1993, 1996, 2002, 2003, 2010, 2012, and 2014. In the 11 warming winters, we have 22 UB events, but only 12 UB events in the 12 cooling winters. The occurrence dates of each UB event are shown in Table 1. The mean period of UB events is 20.1 days in warming winters, but 9.3 days in cooling winters. Thus, it is clear that the tropospheric warming is likely related to an increased frequency of UB events.

To see the effect of UB events on tropospheric warming, we present the composite 500-hPa geopotential height anomalies for UB events in Figure 3 for warming and cooling winters. It is apparent that, for a cooling winter, an anticyclonic height anomaly dominates over the whole Arctic at lag −10. Also apparent are large-scale ridges over northern Eurasia and the western North Atlantic. This Eurasian ridge intensifies and shifts eastward to form a UB until lag 0, before moving rapidly westward (from lag 0 to lag 10; Figure 3(a)), although a ridge appearing over North Atlantic (Figure 3(a)) does intensify into a typical negative NAO (NAO−) with the UB’s decay. In particular, its retrogression is dominant during the UB’s life cycle. Thus, the winter cooling is related to a retrograde UB with an NAO−. It is also easy to see (Figure 3(b)) the establishment of a quasi-stationary UB pattern with an NAO+ in the warming winter (Luo, Xiao, Diao et al. 2016; Luo, Xiao, Yao et al. 2016). Luo et al. (2017) found that a UB with an NAO− (NAO+) tends to suppress (favor) the intrusion of midlatitude moisture to the BKS. Thus, it is concluded that a UB with an NAO− originating from an AO−-like pattern is not favorable for BKS warming. Next, we further examine the role of the UB in the winter tropospheric warming.

### Table 1. Occurrence date and period of UB events

| Years       | Date       | Period |
|-------------|------------|--------|
| Cooling(12) | 1981/12–1982/1,2 | 2/14–2/18 | 5 |
|            | 1985/12–1986/1,2 | 1/25–2/7 | 13 |
|            | 1986/12–1987/1,2 | 0 | 0 |
|            | 1988/12–1989/1,2 | 0 | 0 |
|            | 1993/12–1994/1,2 | 12/6–12/17 | 12 |
|            | 1996/12–1997/1,2 | 0 | 0 |
|            | 1997/12–1998/1,2 | 1/16–1/21 | 6 |
|            | 2002/12–2003/1,2 | 2/5–2/11 | 7 |
|            | 2003/12–2004/1,2 | 1/24–2/4 | 12 |
|            | 2010/12–2011/1,2 | 12/28–1/5 | 9 |
|            | 2012/12–2013/1,2 | 12/7–1/9 | 14 |
|            | 2014/12–2015/1,2 | 1/21–2/8 | 8 |
| Warming(11) | 1979/12–1980/1,2 | 12/23–1/30 | 8 |
|            | 1983/12–1984/1,2 | 1/27–2/11 | 16 |
|            | 1984/12–1985/1,2 | 1/24–2/9 | 6 |
|            | 1989/12–1990/1,2 | 1/25–2/20 | 6 |
|            | 1994/12–1995/1,2 | 1/13–2/17 | 5 |
|            | 2004/12–2005/1,2 | 1/20–2/5 | 5 |
|            | 2005/12–2006/1,2 | 1/20–2/5 | 17 |
|            | 2007/12–2008/1,2 | 1/1–1/11 | 11 |
|            | 2008/12–2009/1,2 | 12/12–1/24 | 13 |
|            | 2011/12–2012/1,2 | 12/15–2/23 | 9 |
|            | 2013/12–2014/1,2 | 1/18–2/12 | 26 |

#### 3.3. Roles played by the UB and increased persistence of BKS water vapor in the winter Arctic tropospheric warming

Figure 4 shows height–latitude distributions of winter-mean geopotential height, air temperature, and specific humidity anomalies averaged over 20°–80°E, with and without UB events, for warming and cooling winters, and their differences. The results show that the BKS warming is weaker even in the presence of UB events (Figure 4(a–c)). Thus, the UB-induced tropospheric warming is weaker and less persistent under a background condition of more sea ice, as noted below. This indicates that the role of the UB depends on the background condition of the BKS sea ice. For a warming winter, the tropospheric warming in the BKS is evidently stronger for the case with a UB (Figure 4(d)) than without (Figure 4(e)), which can be clearly seen from their difference field (Figure 4(f)). Thus, the tropospheric warming is related to large-scale circulation patterns. Further calculation shows that, in the near surface layer (800–1000 hPa) of the high-latitude Arctic region (70°–85°N, 20°–80°E), the ratio of the UB-induced winter-mean air temperature anomaly relative to the winter-mean air temperature anomaly is 0.45, but becomes 0.83 in the upper level (500–800 hPa) of the lower troposphere.
Thus, the UB can significantly amplify the warming of the lower troposphere to produce an amplified tropospheric warming in the BKS. Figure 4(g–l) show the corresponding height–latitude distributions of specific humidity anomalies in the BKS. We can clearly see that the BKS has more water vapor in a warming winter (Figure 4(l)) than in a
anomalies during the life cycle of UB events (Figure 5). Here, we use the definition of sensible heat energy presented by Lu, Gao, and Zhai (1996), in the form

\[ E_s = c_p \int T dp, \]

where \( c_p = (1 + 0.837q) \times 1.005 \), \( q \) is the specific humidity, \( T \) is the temperature, and \( p \) is the pressure.

cooling winter (Figure 4(f)). This reflects the likelihood that the UB tends to induce a winter tropospheric warming through water vapor change (Luo et al. 2017).

To further reveal the effect of water vapor on the air temperature in the troposphere, it is useful to show the height–time evolution of the domain-averaged composite daily temperature, specific humidity, and sensible heat energy anomalies during the life cycle of UB events (Figure 5). Here, we use the definition of sensible heat energy presented by Lu, Gao, and Zhai (1996), in the form \( E_s = c_p \int T dp \), to calculate the sensible heat energy in the whole troposphere, where \( c_p = (1 + 0.837q) \times 1.005 \), \( q \) is the specific humidity, \( T \) is the temperature, and \( p \) is the pressure.

**Figure 3. (Continued).**
winter (Figure 5(a)). Such a more persistent warming is related to longer persistence of water vapor in the BKS in the warming winter (Figure 5(d)) than in the cooling winter (Figure 5(a)).

It is interesting to see that, during the UB life cycle, the BKS tropospheric warming is much more intense and long-lived in a warming winter (Figure 5(b)) than cooling winter (Figure 5(a)). Such a more persistent warming is related to longer persistence of water vapor in the BKS in the warming winter (Figure 5(d)) than in the cooling winter.
toward the BKS by weakened horizontal winds under an NAO+ without a UB event (not shown). It is also apparent that a persistent increase in BKS water vapor (Figure 5(d)) can lead to persistently enhanced sensible heat energy in the Arctic troposphere (Figure 5(f)). Thus, it is concluded that the tropospheric warming in the BKS is amplified by

winter (Figure 5(c)). As noted by Luo et al. (2017), the UB can lead to a persistent water vapor increase in the BKS when a UB occurs together with an NAO+. However, the water vapor can weakly intrude into the BKS when an NAO+ does not accompany a UB. Thus, the BKS water vapor is lacking due to reduced advection of water vapor toward the BKS by weakened horizontal winds under an NAO+ without a UB event (not shown). It is also apparent that a persistent increase in BKS water vapor (Figure 5(d)) can lead to persistently enhanced sensible heat energy in the Arctic troposphere (Figure 5(f)). Thus, it is concluded that the tropospheric warming in the BKS is amplified by
important role played by the UB in tropospheric warming through increased moisture intrusion (Woods, Caballero, and Svensson 2013, 2016; Gong and Luo 2017). However, the pattern of a UB with an NAO+. Moreover, it is found that winters with more UB events exhibit significantly greater warming in the BKS (not shown), thus demonstrating the important role played by the UB in tropospheric warming through increased moisture intrusion (Woods, Caballero, and Svensson 2013, 2016; Gong and Luo 2017). However,
what prior condition determines these circulation patterns is unclear. Below, we seek to identify the prior condition of the UB-with-NAO+ pattern and associated tropospheric warming in the BKS.

3.4. Prior condition of tropospheric warming in the BKS

To see whether there is a prior condition for the UB-induced tropospheric warming, it is useful to calculate the temporal evolution of the domain-averaged SIC, SAT, TCWV, and downward IR anomalies in the BKS. The results during the life cycle of a UB event and its extended period are shown in Figure 6(a–d). It can be seen that, prior to the blocking onset (before lag −10), the SIC in the BKS, when associated with a UB, is lower for a warming winter (red line in Figure 6(a)) than for a cooling winter (blue line in Figure 6(a)). Here, we define the sea ice during the period from lag −30 to lag −10 as the prior sea-ice condition of the UB. Also, the time interval from lag −30 to lag −10 is referred to as the prior period of the UB onset. We also see that the positive TCWV anomaly during the period from lag −30 to lag 0 (Figure 6(c)) corresponds to positive SAT (Figure 6(b)) and downward IR (Figure 6(d)) anomalies in the BKS. This suggests that the larger quantity of water vapor during the prior period of the UB is likely related to increased evaporation due to sea-ice loss before the UB onset, even though the UB’s amplitude is so small before lag −10 that the southwestward horizontal winds in the BKS are weak (not shown). However, we need to explain why there is less prior sea ice in a BKS warming winter. To examine this problem, we show the SST anomalies averaged from lag −30 to lag −10 in Figure 6(e) and (f). An interesting result is found in that, prior to the blocking onset, the SST anomaly shows a tripole pattern that coexists with a positive SST anomaly in the BKS for a warming winter (Figure 6(f)); however, the prior SST anomaly of the UB has an opposite structure for a cooling winter (Figure 6(e)). Thus, the prior sea-ice state is mainly determined by the horizontal distribution of the prior SST tripole pattern that corresponds to more water vapor near the Gulf Stream extension region (not shown). In this case, a combined UB and NAO+ pattern can transport more warm and moist air to the BKS through the UB–NAO+ ‘relay’ (Luo et al. 2017). Thus, the increased persistence in the BKS water vapor associated with a UB-with-NAO+ pattern can persistently warm the troposphere in the BKS through upward sensible heat energy to generate an amplified tropospheric warming in the BKS.

4. Conclusion and discussion

The present study has examined the physical cause of tropospheric warming in the BKS. It is found that tropospheric warming not only depends on the prior sea-ice state in the BKS, but also on changes in atmospheric circulation patterns. For a strong tropospheric warming winter, the corresponding atmospheric circulation shows a UB pattern with an NAO+, which is related to prior positive SST anomalies and less prior sea ice over the BKS. However, for a cooling winter, the atmospheric circulation looks like a UB pattern with an NAO−, which originates from an AO− pattern. Such a circulation pattern is related to a negative SST anomaly and more prior sea ice in the BKS. For the UB with an NAO+, warm moisture is transported in large quantities to the BKS to produce a persistent increase in water vapor through the UB–NAO+ ‘relay’ (Luo et al. 2017). In this case, the persistent increase in water vapor can lead to amplified tropospheric warming through persistent enhanced upward sensible heat energy. The UB can then induce an upward enhancement of the tropospheric warming in the BKS. Thus, the warm winters are wet winters with low sea ice in the BKS and tend more toward an NAO+ with a UB. For a UB with an NAO−, the moisture intrusion into the BKS is suppressed to produce a less persistent water vapor increase. In this case, the tropospheric warming is weaker because of a less persistent increase in the sensible heat energy. Thus, cooling winters are drier and tend more toward an NAO− with a UB.

In conclusion, the above results reveal that, while the tropospheric warming in the BKS depends on the sea-ice state prior to the UB onset, it also depends on the subsequent UB and phase of the NAO. This also indicates that the tropospheric warming is likely a result of the interaction between the sea-ice decline and atmospheric circulation change through the moisture change (Woods, Caballero, and Svensson 2013, 2016).

Disclosure statement

No potential conflict of interest was reported by the authors.

Funding

This work was supported by the National Natural Science Foundation of China [grant number 41430533], [grant number 41375067].

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