Local increase of anticyclonic wave activity over Northern Eurasia under amplified Arctic warming

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Abstract

In an attempt to resolve the controversy as to whether Arctic sea ice loss leads to more mid-latitude extremes, a metric of finite-amplitude wave activity is adopted to quantify the midlatitude wave activity and its change during the observed period of the drastic Arctic sea ice decline in both ERA Interim reanalysis data and a set of AMIP-type of atmospheric model experiments. Neither the experiment with the trend in the SST or that with the declining trend of Arctic sea ice can simulate the sizable midlatitude-wide reduction in the total wave activity ($A_e$) observed in the reanalysis, leaving its explanation to the atmospheric internal variability. On the other hand, both the diagnostics of the flux of the local wave activity and the model experiments lend evidence to a possible linkage between the sea ice loss near the Barents and Kara seas and the increasing trend of anticyclonic local wave activity over the northern part of the central Eurasia and the associated impacts on the frequency of temperature extremes.
1. Introduction
Recent decades have seen accentuated warming and precipitous decline of sea ice in the Arctic, in keeping with the so-called Arctic amplification (AA) anticipated from the increasing greenhouse gas forcing [Holland et al., 2003; Screen and Simmonds, 2010; Graversen et al., 2008; Serreze et al., 2009]. Accompanying the Arctic change are the more frequently observed winter weather extremes like cold snaps and snow storms in the Northern Hemisphere midlatitudes since the early 1990s, especially over the eastern United States and central Asia [Min et al., 2011; Coumou and Rahmstorf, 2012; Coumou et al., 2013; Westra et al., 2013]. Some studies [Francis and Vavrus, 2012; Cohen et al., 2013, 2014] attempted to assign causation from the Arctic sea ice melting to the midlatitude extreme weathers, and thus a controversy ensues regarding whether the Arctic amplification has led to the more frequent midlatitudes extremes [Barnes and Screen, 2015; Overland et al., 2011; Liu et al., 2012; Gerber et al., 2014; Li et al., 2015; Sun et al. 2016; McCusker et al. 2016]. The proponents of the Arctic-midlatitude connection [Francis and Vavrus, 2012, 2015; Cohen et al., 2014] suggested that Rossby waves propagating in a weakened westerly jet tends to slow down and become meridionally amplified, thus favoring more extreme weather conditions in the context of AA. Other studies [Barnes, 2013; Screen and Simmonds, 2013] questioned their methodology and pointed out that the claimed increase in waviness might be an artifact due to the metric used for quantifying the wave amplitude and the interpretation thereof. Furthermore, their proposition was challenged by the notion that the relationship above between the wave activity and zonal jet holds only for internal variability, but may break down
under externally imposed thermal forcing [Hassanzadeh and Kuang, 2015; Chen et al. 2015].

Less controversial is the possible influence of the Arctic sea ice loss on weather and temperatures over the adjacent continents [e.g., Tang et al., 2013; Overland et al., 2015; Screen et al., 2013; Kretschmer et al., 2016]. Reduction in autumn-winter Arctic sea ice, especially in the Barents-Kara Sea, has been linked to more frequent and persistent anticyclonic circulation over northern Eurasia, inducing cold events to its southeastern flank [Mori et al. 2014; Petoukhov and Semenov, 2010; Inoue et al., 2011; Horton et al. 2015; Kug et al. 2015]. This is consistent with the finding that the amplification of quasi-stationary waves tends to preferentially occur in Eastern Europe and central Asia [Screen et al., 2014].

However, studies summarized above used disparate metrics for the midlatitude waviness and/or extremes; most of them are empirical and lacking an interpretation from the perspective of atmospheric dynamics; some explanation may lead to unnecessary confusion. In this study, we utilize an objective metric founded on fundamental geophysical fluid dynamical principles to quantify the changes in the midlatitude waviness and their possible attribution to Arctic sea ice melting through a suite of Atmospheric General Circulation Model (AGCM) simulations. The said metric is the local variant of the finite-amplitude wave activity, which can be readily partitioned into the mean gradient and eddy meridional scale,
thus resolving the caveat of the metrics that tend to confuse the mean temperature warming with the eddy meridional scale change in the wave activity.

2. Methodology and Model Experiments

2.1 Local Wave Activity (LWA) and its Flux

Finite-amplitude wave activity is developed as an objective measure for the areal displacement of a physical quantity that shows broad monotonicity in its spatial distribution. LWA is a natural extension of the finite amplitude wave activity towards locality. The exact definition of wave activity for potential vorticity and the related dynamical properties has been detailed in Nakamura and Solomon (2010) and Nakamura and Zhu (2010). Recently, Huang and Nakamura (2016) developed the concept of local wave activity and the related budget for local wave phenomena like wave breakings and blockings; Chen et al. (2016) further extended the concept of LWA to less conserved quantity (500hPa geopotential height, denoted by $z_{500}$) to facilitate its broader utilization. The latter approach is adopted for this study.

Specifically, for $z_{500}$ that has broad monotonic distribution with latitude, one can select a contour value $Z_{500}$ and define an equivalent latitude $\phi_e$ in the Northern Hemisphere such that the area $S$ bounded by the value $Z_{500}$ towards the North Pole is

$$S(Z_{500}) = \iint_{\pi z_{500}} a^2 \cos \phi \, d\lambda \, d\phi$$  \hspace{1cm} (1)$$

where $\lambda$ is longitude. A monotonic relationship between $\phi_e$ and $Z_{500}$ value can then be established:
\[ \phi_e(Z_{500}) = \arcsin\left[1 - \frac{S(Z_{500})}{2\pi a^2}\right] \]  

(2)

Introducing an eddy component \( \bar{z} = z_{500} - Z_{500} \), the southward and northward LWA at longitude \( \lambda \) and equivalent latitude \( \phi_e \) can be defined as

\[
A_S(\lambda, \phi_e) = \frac{a}{\cos\phi_e} \int_{2\leq 0, \phi \leq \phi_e} \bar{z}(\lambda, \phi) \cos \phi d\phi, 
\]

(3)

\[
A_N(\lambda, \phi_e) = \frac{a}{\cos\phi_e} \int_{2\geq 0, \phi \geq \phi_e} \bar{z}(\lambda, \phi) \cos \phi d\phi, 
\]

(4)

respectively. Defined as such, both \(-A_S\) and \(A_N\) are non-negative, with the former describing the cyclonic wave activity residing to the south of the equivalent latitude \( \phi_e \) and the latter the anti-cyclonic wave activity to the north of \( \phi_e \). Large and persistent \( A_N \) is often related to atmospheric blocking and the spatial correspondence between their climatological distributions has been noted (P. Martineau, personal communication). The sum of \(-A_S\) and \(A_N\) recovers the total wave activity \( A_e(\phi_e) = -A_S + A_N \), which is a function of \( \phi_e \) only, resuming the original meaning of wave activity measuring the total waviness in contour \( Z \). A simple dimensional analysis suggests that the total wave activity can be written as

\[ A_e = \frac{1}{2} l^2 \cdot \frac{dz}{dy}. \]

As such the wave activity can be thought of as the result of the stirring of tracer \( z \) with a background gradient \( dZ/dy \) by a meridional disturbance of scale \( l \), thus naturally partitioned into a thermodynamic factor (former) and a dynamical one (latter). For a typical meridional height gradient \( \frac{\partial Z_{500}}{\partial \phi_e} \sim 8 \, m/deg \), a meridional stirring by \( \delta \phi \sim 15^\circ \) latitude implies a wave activity of \( A \sim 10^8 \, m^2 \).

Defined as above, one can derive a budget for the local wave activity based on the temperature equation. The temperature tendency in the lower troposphere during
midlatitude large-scale extremes is the result of by the horizontal temperature
advection and diabatic heating/cooling, thus

\[ \frac{\partial T}{\partial t} = -\mathbf{v}_g \cdot \nabla T + \dot{T}, \quad (5) \]

where \( \mathbf{v}_g \) denotes the geostrophic flow and \( \dot{T} \) represents the total heating of all the
diabatic processes. Applying the hypsometric relation \( Z_{500} = \frac{R}{g} \int_{p_{500}}^{p_s} T \, d\ln p \) (Holton, 2004) and transformation \( \mathcal{A}_N(\cdot) \equiv \frac{a}{\cos \phi_e} \int_{\phi_e}^{\phi} (\cdot) \cos \phi \, d\phi \) to it, we
yield the budget equation for \( A_N \):

\[ \frac{\partial A_N}{\partial t} \approx -\mathcal{A}_N \left( \frac{R}{g} \int_{p_{500}}^{p_s} \nabla \cdot (\mathbf{v}_g T) \, d\ln p \right) + \mathcal{A}_N \left( \frac{R}{g} \int_{p_{500}}^{p_s} T \, d\ln p \right) \quad (6) \]

The first term on the right hand side represents effectively the convergence of
horizontal flux of \( A_N \), and the corresponding divergent flux vector \( \mathbf{F}_{AN} = (F_u, F_v) \)
can also be computed by inverting \( \nabla \cdot \mathbf{F}_{AN} \equiv -\mathcal{A}_N \left( \frac{R}{g} \int_{p_{500}}^{p_s} \nabla \cdot (\mathbf{v}_g T) \, d\ln p \right) \). The
second term is the sink/source of \( A_N \) due to diabatic heating. Over the regions
where diabatic processes are important, the balance is largely held between these
two terms at equilibrium, with a divergence (convergence) in the second term
balanced by a diabatic \( A_N \) source (sink).

2.2 AGCM Experiments and Reanalysis Dataset

For the purpose of this investigation, we make use of the daily output from two
groups of existing AGCM experiments described in Perlwitz et al. [2015] and Deser
et al [2015], respectively. The former comprises two sets of 30-member AMIP-style
simulations using ECHAM5: one denoted AMIP Historical and forced by the
observed/projected radiative forcing, observed monthly sea surface temperature (SST) and sea ice concentrations (SIC) (Hurrell et al. 2008), the other AMIP SST forced by SST and radiative forcings identical to AMIP Historical but a repeating climatological seasonal cycle of sea ice for 1979–1989 (see Perlwitz et al. [2015] for more details). The difference between the 30-member ensembles of AMIP Historical and AMIP SST, denoted as AMIP ΔSIC, can then be attributed to the variability of Arctic sea ice. The 1990-2014 linear trends of net upward surface turbulence heat flux (i.e., the sum of net short wave and long wave radiation, latent heat flux and sensible heat flux) averaged over the Arctic Ocean are compared among a reanalysis, AMIP Historical and AMIP ΔSIC experiments in Figure S1. There is a reasonable agreement among them in the seasonality of the trend of net surface heat flux, as well as the magnitude of the wintertime flux, which peaks at ~40 W m\(^{-2}\) in December in all the three cases. Despite the fact the sea ice and SST in the AMIP experiments are prescribed, this reasonable agreement lends us some confidence in the representativeness of the forcing to the atmosphere from the Arctic sea ice melting.

For the impact of future Arctic sea ice melting, we analyze a pair of time-slice experiments with CAM4, each 260 years long, forced by a repeating seasonal cycle of sea ice representing the 20\(^{th}\) century sea ice condition (averaged between 1980-1999 from CCSM4 historical runs) and the 21\(^{st}\) century condition (averaged between 2080-2099 from CCSM4 RCP8.5 runs), respectively. The two cases are denoted as ICE_CAM_20 and ICE_CAM_21 and their difference, denoted SIM, isolates
the direct atmospheric response to the Arctic sea ice loss in the absence of nonlocal oceanic feedbacks. Table 1 summarizes these two groups of experiments.

Daily mean surface air temperature and 500hPa geopotential height at spatial resolution of 1.125°lon×1.125°lat from the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim (ERA-I) project [Dee et al., 2011] are used in the present study as a reference for the observation.

3. Results

3.1. Trend of local wave activities in reanalysis and Model Simulations

To evaluate whether Arctic sea ice loss can induce a broad change in the midlatitude waviness, we start with examining the trend of the total wave activity $A_e$ in the ERAI reanalysis and AMIP-type experiments using ECHAM5. A significant upward trend (based on two-tailed Student’s $t$-test) of wintertime (December-February) $A_e$ during 1990/1991-2013/2014 is detected in ERA-Interim (Figure 1a, black line), especially for the latitude poleward of 50°N, indicating enhanced mid-latitude wave activities. This midlatitude increase occurs despite the decrease in the reduction of the Lagrangian gradient $dZ_{500}/dy$ (consistent with the overall weakening of the poleward gradient of the lower-tropospheric temperature), implicative of an even greater increase in the stirring length $l^2$ (Figure 1c). However, none of the features in ERAI can be replicated by the two AMIP experiments; their difference, AMIP $\Delta$SIC, even produces a negative $A_e$ trend in high latitudes. Inspecting all the ensemble members, we found that not a single member of these AMIP experiments can
produce as strong a magnitude of the midlatitude $A_e$ increase as seen in ERA-Interim, implying a potential model deficiency of ECHAM5 in capturing the magnitude of the wave activity response. An interesting exception is the subtropical increase of $A_e$, which appears attributable to the SST forcing in the AMIP experiments (Figure 1, green and magenta lines). Thus, we cannot reject the null-hypothesis that the large extratropical trend in $A_e$ seen in ERA-I is the result of internal variability, at least within the context of this set of experiments. It is interesting to note that the difference between ICE_CAM_21 and ICE_CAM_20 using CAM4 (not shown), representing the impact of the future Arctic sea ice loss, shows increases in $A_e$ and $l^2$ and decrease in $dZ/dy$ resembling the ERA-I trends in mid-to-high latitudes. Without further experiments, we cannot discern whether the apparent differing results between ECHAM5 and CAM4 experiments are due to the model difference or to the different forcing intensity. In summary, jury is still out as to whether Arctic sea ice loss can give rise to an overall increase in the midlatitude wave activity.

On the other hand, when it comes to the trend of the local wave activity, the observed two centers of significant increase over north central Eurasia and northeastern Pacific in the $A_N$ (representing anti-cyclonic wave anomalies) can be captured by the AMIP Historical experiment, but with reduced magnitude (Figure 2b, note that in Figure 2a the actual magnitude of the observed trend is scaled by a factor of $\frac{1}{2}$ to compare with the model trends). Decomposing the modeled signal into the parts due to SST and sea ice, we find that the increase over northeastern
Pacific can be largely attributed to the SST forcing, while both SST and sea ice play a
part in the $A_N$ trend over central northern Eurasia (Figure 2c,d). As a La Nina-like
SST anomaly can recurrently lead to a high $A_N$ anomaly over northeastern Pacific
(not shown), the $A_N$ trend there is likely the result of the La Nina-like SST trend in
the central and eastern equatorial Pacific since the early 1990s. Under the Arctic sea
ice condition of the late 21st century (which is much more severe and widespread
than the historical reduction), the $A_N$ increase over Eurasia simulated by CAM4 can
reach the same magnitude as the observation (Figure 2e), giving rise to detectable
signal on the temperature extremes over the impacts areas (to be elaborated in the
next subsection).

Focusing on the feature of the $A_N$ increase over Eurasian region, we further define a
local wave activity index ($I_{AN}$) by spatially averaging $A_N$ over (50°E-70°E, 50°N-
70°N) and the resultant time series for ERA-I reanalysis and AMIP simulations are
presented in Figure 3. Consistent with the $A_N$ trend pattern shown in Figure 2, both
ERA-I reanalysis and the two AMIP experiments exhibit an upward trend in $I_{AN}$.

Compared to the large internal variability of $I_{AN}$ in ERA-I reanalysis, the interannual
fluctuations of the model ensemble mean $I_{AN}$ is much muted, rendering a marginal
significance (at 95% level) to the $I_{AN}$ trend in AMIP Historical. This trend seems to
arise from both the SST (Figure 3, magenta line) and Arctic sea ice forcing (blue line)
during 1990-2013. Interestingly, a large positive $I_{AN}$ anomaly is often associated
with a La Niña SST condition in the central eastern Pacific and vice versa in both
ERA-I and AMIP SST experiment. This points to a possibility that ENSO
teleconnection can impact on the transient weather activity as remote as central
Eurasia, an aspect of ENSO worth further investigation.

3.2 Temperature impact of $A_N$ over Eurasia

Given the fact that large positive $A_N$ are often associated with anticyclonic blocking
[Chen et al., 2015] that advects cold Arctic air southward and warm air poleward, it
is conceivable that the increasing trend of $A_N$ over northern central Eurasia may
bring about detectable changes in the distribution of temperature extremes nearby.
Making use of the ERA-I reanalysis data, we first composite the surface temperature
anomalies that are congruent with the large $I_{A_N}$ and identify two locales that are
subject to its advective effect: an area near the Barents-Kara Sea ($50^\circ$-70$^\circ$E, 70$^\circ$N-
80$^\circ$N) and one near central Asia ($75^\circ$E-95$^\circ$E, 48$^\circ$N-58$^\circ$N). These are the two regions
where the large $I_{A_N}$ trend may be manifested in the respective temperature
distribution.

Indeed, a dipole with less frequent cold events and more frequent warm events is
found in the 1990-2014 trend of the PDF of the ERA-I surface temperature over the
Barents-Kara Sea region (Figure 4a), while an opposite dipole is found over the
central Asia region (Figure 4b). Both dipolar PDF trends are statistically significant
with respect to the PDFs constructed by randomly sampling the DJF daily
temperature over the two regions 10,000 times. The same analysis of the
temperature PDF over these two regions from experiment AMIP $\Delta$SIC shows similar
dipoles in character, but only the PDF dipole for the northern region is statistically
significant (Figure 4c). However, under Arctic sea ice loss of the late 21st century RCP8.5 scenario, the CAM4 time slice experiments do capture qualitatively similar PDF dipoles as observed (Figure 4e, f). The corollary is that the sea ice loss and the related warming trend over Barents-Kara Sea, if continuing and acting in isolation, can potentially bring more cold weather extremes to central Asia.

3.3 Flux budget for $A_N$

The evidence from the AMIP experiments for the teleconnection between the Arctic sea ice melting and the Eurasian $A_N$ trend is only suggestive at most. To identify the possible source and pathway for the Eurasian $A_N$ trend, we computed the convergence of the horizontal flux of $A_N$ (second term in equation (6)) and the corresponding divergent flux vector $F_{A_N} = (F_u, F_v)$ directly and their trends are displayed in Figure 5. According to equation (6), the change of the $A_N$ should be maintained jointly by the convergence term and the source term. Unfortunately, since ERA-I data set does not include diabatic heating, we can not quantify directly how much of the diabatic source term contributes to the budget. Nevertheless, the good spatial correspondence between the $A_N$ trend (contours) and its flux convergence (shading) is evident, vindicating this budget effort. In particular, the positive $A_N$ trend over central northern Eurasia is at least partially maintained by the large $A_N$ convergence, which itself may be traced back to the Arctic region following the $F_{A_N}$ vectors backward, hinting at a possible Arctic origin of the $A_N$ trend. It will remain elusive as to the relative importance of the local versus remote
source for the maintenance of $A_N$ trend over Eurasian continent until the availability of trustworthy diabatic heating data.

4. Conclusion

A novel metric for measuring the midlatitude wave activity is devised and used to evaluate the possible role of the precipitous Arctic sea ice melt during recent decades in the increasing frequency of midlatitude weather extremes. The said metric is the finite-amplitude wave activity $A_e$ developed recently from the Geophysical Fluid Dynamics community, which is computed from 500hPa geopotential height and can be readily decomposed into the changes in the mean gradient and eddy stirring scale. The local extension of it can be used to quantify the wave magnitude locally. Evaluating the trends of the both total and local wave activities from ERA-I reanalysis against those from a set of AMIP type AGCM experiments shows that the observed Arctic sea ice loss lead to no midlatitude-wide increase in the total wave activity, while locally the reduction of the sea ice over Barents-Kara Sea can induce an increase in the local northward wave activity over central northern Eurasia. This southward teleconnection is further supported by a diagnostic of the flux of the local wave activity. In another set of time slice experiments forced by the sea ice decline representing the condition at the end of the 21st century in the future RCP8.5 scenario, the local wave activity response over Eurasia is amplified and can lead to significant modulation on the distribution of the temperature extremes over the impacted areas.
As far as the linkage between Arctic amplification and midlatitude weather extremes is concerned, this study should not be regarded conclusive. The total wave activity response to a polar thermal forcing is a nuanced one depending on how the midlatitude eddy-mean flow interaction responds to the thermal forcing and redistributes between the zonal mean momentum and wave activity. One should not confuse the redistribution under external forcing with the internal exchange in the absence of external forcing.

At issue is the model-dependence of the atmospheric response to sea ice loss and further model inter-comparison study is in order. Meanwhile, the numerical experiments utilized in this study may not represent faithfully the true forcing from the sea ice melting [Cohen et al., 2014; Furtado et al., 2015]. Prescribing sea ice concentration and sea ice temperature may interfere the interactive nature of the air-ice heat exchange and prescribing SST artificially excludes the ocean dynamical feedbacks (whose importance was acutely underlined in Deser et al. 2015). More thoughtful, energetically consistent experimental designs with multiple models are needed to fulfill the goal of attributing the change in the midlatitude extremes in the past and future.

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The data for ECHAM experiments can be downloaded from

http://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/. The data for ECHAM experiments can be downloaded from

http://www.esrl.noaa.gov/psd/repository/entry/show?entryid=d0202b64-a9c5-4b65-9d7c-ef5b76a4dbe6.


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**Figure 1** Fractional linear trend in (a) $A_e$; (b) Lagrangian gradient $dZ_{500}/dy$, and (c) stirring length $l^2$ in ERA-I reanalysis (black), AMIP Historical (green), AMIP SST (magenta), AMIP ΔSIC (blue) experiments. The latitudinal range where the trend is statistically significant at 95% confidence level based on Student’s t-test is highlighted in bold.
Figure 2  Distributions of the background mean $A_N$ (contours, C.I.: $10^8$ m$^2$) and the trend of $A_N$ (shaded; Unit: $10^7$ m$^2$) during 1990-2013 in (a) ERA-I; (b) AMIP Historical experiment; (c) AMIP Sea Ice experiment; and (d) AMIP SST experiment. (e) shows the model climatological mean $A_N$ and the difference between experiments ICE_CAM_21 and ICE_CAM_20. The black dots in each panel stand for the significant trend or difference at 95% confidence level using Student's t-test. The green box demarcates the region where the $A_N$ index is taken.
Figure 3 Time series of $A_N$ anomaly (Unit: $10^8$ m$^2$) over Eurasian regions in ERA-I (black lines), AMIP Historical experiments (green lines), AMIP Sea Ice experiment (blue lines), and AMIP SST experiment. The dash lines stand for the trend of $A_N$ for 1990-2013. El Nino (La Nina) years (according to the Niño-3.4 index provided by the Climate Prediction Center) are marked by red (blue) small bars on the top of the box. The gray shading indicates the range of the $A_N$ index simulated by the 30 ensemble members of AMIP historical experiments.
Figure 4 The probability distribution of surface temperature over the regions to the northwest (a, c, e) and southeast (b, d, f) of $A_N$ center respectively for ERA-I (a, b), AMIP Sea Ice (c, d), and SIM (e, f). The blue bars stand for the probability increase (%). Only significant values according to a Monte-Carlo re-sampling method are shown (top and lower 5 percentiles).
Figure 5 Distributions of $F_{AN}$ (vector, Unit: $10^7$ m s$^{-1}$) and trend of $-\nabla \cdot F_{AN}$ (shaded; Unit: $10^3$ m s$^{-2}$) overlapped on the trend $A_N$ (contour; Unit: $10^6$ m$^2$). The white dots indicate the significant values at 0.05 confident level based on Student’s t-test.
Figure S1 The linear trend of the Arctic turbulence surface heat flux (sensible plus latent, W m$^{-2}$) during the period 1990-2014, computed from ERA-I (black), AMIP Historical (green), and AMIP ΔSIC (blue).