Large-Scale Atmospheric Warming in Winter and the Arctic Sea Ice Retreat

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ABSTRACT

The ongoing shrinkage of the Arctic sea ice cover is likely linked to the global temperature rise, the pronounced warming in the Arctic, and possibly weather anomalies in the midlatitudes. By evaluating independent components of global atmospheric energy anomalies in winters from 1980 to 2015, the study finds the link between the sea ice melting in the Arctic and the combination of only three well-known atmospheric oscillation patterns approximating observed spatial variations of near-surface temperature trends in winter. The three patterns are the North Atlantic Oscillation (NAO), Scandinavian blocking (SB), and El Niño–Southern Oscillation (ENSO). The first two are directly related to the ongoing sea ice cover shrinkage in the Barents Sea and the hemispheric increase of near-surface temperature. By independent dynamical processes they connect the sea ice melting and related atmospheric perturbations in the Arctic either with the negative phase of the NAO or the negative trend of atmospheric temperatures over the tropical Pacific. The study further shows that the ongoing sea ice melting may often imply the formation of large-scale circulation patterns bringing the recent trend of colder winters in densely populated areas like Europe and North America.

1. Introduction

Observations show that the Arctic sea ice retreat is tied to global warming (e.g., Overland and Wang 2010; Cohen et al. 2014). Correlations and modeling experiments indicate that the loss of the ice cover in summer and autumn may modify large-scale atmospheric patterns in the following winter (Jaiser et al. 2013). The varying sea ice cover over the areas with the largest melting trend in the Arctic, like the Barents Sea, may further produce hemispheric-scale impacts in the atmosphere (e.g., Deser et al. 2007; Petoukhov and Semenov 2010; Screen et al. 2013; Mori et al. 2014). Other studies,

however, demonstrate that the relationship with some important weather structures in the midlatitudes may be weak or even nonexistent (e.g., Barnes and Dunn-Sigouin 2014). These contradictory findings suggest that the minimum set of independent atmospheric structures linking the nonuniform atmospheric warming with the contemporary sea ice melting in some small regions of the Arctic and weather changes in midlatitudes has not been established. It has been argued that a major impact on the societal perception of global warming may depend significantly on varying local weather conditions (e.g., Howe et al. 2012) that may be mainly modified by this set of large-scale atmospheric patterns and not so much directly by the change of the mean global temperature.

The retreat of the Arctic sea ice in winter is less dramatic than in summer, but there is a continuous reduction of the ocean surface covered by the ice. The sea ice retreat is especially pronounced in the Barents Sea (Fig. 1). During the polar night from December to January meridional gradients of temperature in the Northern Hemisphere are the largest, and the atmosphere over the Arctic may be heated only from the surface or by the northward advection. It has been recognized that because of these

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FIG. 1. The trend of December–February average of the sea ice cover from 1980 to 2015, expressed from 0 (no ice) to 1 (full ice). The magenta line shows the mean edge of the sea ice covered area from December to February between 1980 and 2015.

specific conditions in winter the sea ice melting in the Arctic may have an important influence on the large-scale dynamics in the atmosphere and impact the weather in the midlatitudes (e.g., [Cohen et al. 2014](#)). Previously, the impact of the observed melting trend in the Barents Sea on the atmospheric circulation has been studied by modeling or by the analysis of its correlations with some predefined weather structures (e.g., [Petoukhov and Semenov 2010](#); [Mori et al. 2014](#)). In particular the recently observed tendency for the negative North Atlantic Oscillation in winter has been explained by mechanisms that can be separated into three main groups: the Arctic drivers ([Cohen et al. 2012](#); [Francis and Vavrus 2012](#); [Screen et al. 2013](#); [Cohen et al. 2014](#); [Kim et al. 2014](#)) emphasizing the role of the sea ice melting and the snow cover in Siberia, tropical drivers ([Trenberth et al. 2014](#); [Ding et al. 2014](#)) stressing the impact of the El Niño–Southern Oscillation, and variations of extratropical SST ([Peings and Magnusdottir 2014](#); [Perlwitz et al. 2015](#)).

In this study we attempt to find the set of independent, large-scale structures that have significant linear trends and together approximate the spatial variability of near-surface temperature trends during winter. We further study their apparent relationship with the ongoing sea ice melting tendency in the Arctic. In [section 2](#) we describe a novel statistical methodology evaluating energy anomalies. The approximation of the spatial and temporal variability of near-surface temperature trends and the impact of each process are presented in [sections 3](#) and [4](#). The statistical estimation of the relationship between the sea ice melting and atmospheric patterns is given in [section 5](#), while [section 6](#) contains the discussion of the findings.

2. Data and methods

Spatial and temporal anomalies of monthly mean atmospheric energy estimated in the atmospheric reanalysis are studied to quantify their relationship with the sea ice

melting in the Arctic. By using energy as a common variable, multivariate samples containing temperature, geopotential, and velocity at all atmospheric levels are evaluated. Each sample spans the 3-month period from December to February and, therefore, resolves the slow seasonal evolution of dynamical processes in the atmosphere at the planetary scales.

a. Energy anomalies

Monthly averaged wind intensity, geopotential height, and temperature from the NCEP reanalysis (Kalnay et al. 1996) were available on 17 standard pressure levels from 1000 to 10 mb and topography with the horizontal resolution of 2.5° . At each horizontal point and pressure level kinetic, potential, and internal energies per unit mass are defined as $E_K = |\mathbf{v}|^2/2$, $E_P = gz$, and $E_I = c_V T$, respectively. Here $|\mathbf{v}|$ is the wind speed, z the height of isobaric surfaces, and T the temperature, while g is the acceleration due to gravity and c_V the heat capacity of air at constant volume.

State vectors are formed consisting of kinetic, potential, and internal energies at all levels. The statistical processing of energy components is justified by the fact that it processes multivariate state vectors without arbitrary scaling, while anomalies of geopotential height and temperature at different levels may be always recalculated after the division by known constants from $\delta z = \delta E_P/g$ and $\delta T = \delta E_I/c_V$. Although wind intensity cannot be easily recovered because of the quadratic form of the relationship, kinetic energy anomalies still may indicate areas with the largest change of wind speed.

After the statistical processing, vertically integrated energies may be diagnosed by integrating values at each level along the pressure coordinate from the pressure at the elevation of the topography to the pressure at the top of the atmosphere. In this way atmospheric column-integrated potential, internal, and kinetic energies, respectively, become closely related to the center of the atmospheric mass, column-averaged temperature, and wind intensity.

Energy anomalies are computed in December, January, and February from 1980 to 2015. The seasonal cycle is removed together with the long-term average at each point by subtracting long-term means calculated separately in each month. The whole set contains 36 samples (i.e., 36 years in the data). Each sample has anomalies of kinetic, potential, and internal energies on 17 atmospheric levels and during the three winter months. The number of grid points is reduced along latitudes to produce points with the same area preserving the variance with respect to horizontal geometry. Values at each level are multiplied by the square root of the pressure

thickness to preserve the variance with respect to the change of the layer mass.

Near-surface temperature anomalies in winter are evaluated by studying anomalies T1000 of temperature at the isobaric surface of 1000 mb ($\delta T_{1000} = \delta E_I|_{p=1000}/c_V$) that is positioned close to sea level. Anomalies H1000 of the height of the 1000-mb isobar surface ($\delta H_{1000} = \delta E_P|_{p=1000}/g$) closely following sea level pressure anomalies further indicate, by assuming geostrophy, the variation of near-surface dynamics and advection by winds. In addition, changes of vertically integrated internal (VIIE) and potential (VIPE) energies will describe the evolution of atmospheric column heat content and dynamics, respectively.

The processing spans from 1980 to 2015 during the so-called satellite era when T1000 is assumed to be well observed by in situ and satellite observations. In this period satellite observations of radiances monitor the temporal and spatial variability of VIIE (Kalnay et al. 1996; Vinnikov et al. 2006). On the other hand, the observational monitoring of the vertical distribution of temperature may be less accurate in remote areas (Vinnikov et al. 2006). Thus, in addition to the evaluation of VIIE, the study of trends is restricted to T1000 that is close to the surface and assumed to be well observed. In winter there are very few near-surface observations of temperature over the Arctic sea ice, and there the reanalysis of T1000 may be less accurate. Nevertheless, the sea ice coverage is well observed by satellites on a monthly frequency, and we assume that its connection with atmospheric conditions at large scales may be resolved by the reanalysis. Furthermore, over the ocean the sea surface temperature is also continuously observed by satellites at a monthly frequency providing an accurate boundary constraint on the near-surface temperature in the reanalysis over ice-free areas.

The additional uncertainty of statistical estimates from the reanalysis comes from the specific processing of indirect observations like the radiances and the use of a specific data assimilation scheme together with a particular atmospheric model. Extensive comparisons between several reanalysis products showed significant biases and large differences between state estimates even for the best-observed parameters like near-surface temperature. For example, in the Arctic all reanalyses underestimate the temperature inversion near the surface (Serreze et al. 2012) showing positive biases for near-surface temperature, and there is a large spread of estimates for near-surface temperature trends (Lindsay et al. 2014). During winter months, however, differences between reanalysis datasets appear to be smaller, and the NCEP reanalysis has a low bias (Lindsay et al. 2014).

The uncertainty linked to the particular reanalysis dataset is partly addressed by additionally processing anomalies from ERA-Interim (Dee et al. 2011) and comparing the results with those obtained from the NCEP reanalysis. This reanalysis dataset used the atmospheric model with a higher spatial resolution, and the data assimilation scheme applied additional treatment of biases estimated for observations and the atmospheric model. Anomalies are computed at the same horizontal resolution of 2.5° , but 38 levels are used between 1000 and 10 mb. Although the statistical processing was made on the same horizontal resolution, anomalies originate from a higher-resolution model and may contain a larger number of resolved scales and processes than the NCEP reanalysis.

b. Statistical processing

As most planetary-scale processes in the atmosphere do not follow normal probability distributions, Hannachi et al. (2009) proposed to apply the independent component analysis (ICA) method instead of the commonly used empirical orthogonal functions (EOFs) in climate studies. Here the Fastica algorithm for the ICA method (Hyvärinen and Oja 2000) will be applied to extract components sharing the minimum information without the Gaussianity assumption.

Starting from matrix $\mathbf{X}$ with rows containing temporal anomalies in the physical space, the ICA algorithm finds a linear mapping between a reduced-order subspace and anomalies in the physical space in the following form:

$$\mathbf{X} \cong \mathbf{AS}. \quad (1)$$

Rows of orthogonal matrix $\mathbf{S}$ contain temporally varying independent components, while columns of full rank matrix $\mathbf{A}$ represent spatially varying intensities. Hereafter columns of matrix $\mathbf{A}$ will be named atmospheric patterns or structures. The appendix describes in more detail the estimation of independent components by Fastica and relates ICA to the rotated EOF (REOF) method that is more frequently used in climate studies.

The chosen nonlinearity in the Fastica algorithm is the hyperbolic tangent. For computational reasons, first the eigenvalue decomposition is made on the correlation matrix originating from anomalies, and the solution is restricted to the subspace spanned by a small number of eigenvectors with the largest eigenvalues. Independent components are estimated from the subspace consisting of four EOFs. By visually inspecting the spatial variability of the T1000 trend it was estimated that only four EOFs were sufficient for the statistical processing. After the statistical processing spatial and temporal structures

of kinetic, potential, and internal energies present in atmospheric patterns were visualized in order to check their physical and dynamical consistency (see an example in supplementary Fig. S1). The same setup of the Fastica algorithm also is used for ERA-Interim except that, as a result of the higher number of resolved scales, six principal components were necessary to approximate the spatial distribution of near-surface temperature trends similarly to four principal components from the NCEP reanalysis. This minimum number of EOFs was again estimated by visually inspecting trends resolved by the reduced subspace.

Eventually it may be suspected that statistical processing of a large number of parameters may become noisy. Large-scale atmospheric dynamics, however, impose strong dependencies among parameters that may be detected by energy anomalies. Therefore, filtering large dynamically consistent structures with different parameters and many levels may define a smoother subspace for estimating independent components. This assumption further justifies the use of the NCEP reanalysis as a primary dataset produced with the coarse-resolution model because it has less small-scale features that may add noise to large-scale patterns.

The statistical processing was made over the whole troposphere and the lower stratosphere up to 10 mb in order to detect the eventual coupling between the troposphere and the stratosphere. The estimation of independent patterns was insensitive to the lowering of the top layer down to 100 mb (not shown). This can be explained by the fact that layers are scaled by the square root of the pressure thickness in order to preserve the variance with respect to the layer mass change. As a consequence the lower stratosphere is weighted by an order of magnitude less than the troposphere, and its variability does not significantly impact the statistical processing.

The statistical processing was also insensitive to a small change of the starting and the ending year. The reduction from 36 to the last 10 years significantly changed the results (not shown). This could be expected because at a short time period independent patterns reflect a different type of the variability, and the uncertainty of the statistical processing increases. The choice of 36 samples may result in a much better estimate of the variance while still satisfying the assumption on the linearity of the processes.

c. Linear trends and statistical significance

The linear slope m_k is computed for each independent component $\mathbf{s}_k$. The spatial distribution of trend associated with each independent component is represented as a column of matrix $\mathbf{T}$:

FIG. 2. Colors indicate (top) observed and (bottom) reconstructed trends of T1000 from 1980 to 2015 ($10^{-2} \text{ K yr}^{-1}$) for (a),(d) December, (b),(e) January, and (c),(f) February. Dots indicate areas with statistically significant trends.

$$\mathbf{T} = \mathbf{AM}, \quad (2)$$

where $\mathbf{M}$ is a diagonal matrix with linear slopes m_k on the diagonal. Only columns $\mathbf{t}_k$ corresponding to linear slopes with large statistical significance are selected for the further analysis.

The full reanalysis trend t_{Fi} of a variable at each geographical point i is then approximated by the reconstructed trend t_{Ri} that is estimated from approximated anomalies:

$$\tilde{\mathbf{x}}_i = \sum_k a_{ki} \mathbf{s}_k \Delta_k, \quad (3)$$

where Δ_k equals one if the trend associated with the k th independent component is statistically significant and equals zero otherwise.

Trends are calculated by the Sen–Kendall method (Sen 1968). The statistical significance for all trends is set to the 0.05 level and is estimated by the Mann–Kendall test (Mann 1945). Areas with significant trends are marked in figures, except for atmospheric patterns that have spatially uniform slopes.

3. Distribution of trends

a. Approximation of reanalysis trends by atmospheric patterns

Figures 2a–c show that during winter linear tendencies of T1000 from the reanalysis vary significantly between positive values over the Arctic Ocean, the North Atlantic, and the North Pacific and negative values over Siberia and the tropical Pacific. In particular there is a prominent dipole between the strong warming over the Arctic and the strong cooling over Siberia. Another large dipole is between the cooling in the eastern Pacific and the warming in the western Pacific over the tropics. There are also smaller spots with enhanced warming over North America, North Africa, and Asia. The general structure is present in all winter months, while the interseasonal variations are localized on small scales. An exception to this picture is an increasing warming trend spreading from the eastern Mediterranean to China in February that was less evident in December and January. All areas with strong warming and cooling tendencies also have statistically significant trends.

FIG. 3. As in Fig. 2, but for VIIIE (kJ yr^{-1}).

The ICA analysis showed that three independent components had statistically significant trends. Anomalies of T1000 were then approximated by these three atmospheric patterns by using Eq. (3). The trend of reconstructed anomalies (Figs. 2d–f) depicts all dominant features shown in Figs. 2a–c. It also resolves many small-scale features like warming spots over North America and Asia. The evolution from December to February is very similar, depicting the increase of the warming trend from the eastern Mediterranean to China. There are also some small-scale differences like opposite weak trends over central Europe in January or the slightly stronger cooling over the equatorial Pacific and the slightly weaker warming over the equatorial Atlantic. The strong warming in the Arctic is localized to the average position of the edge of the sea ice cover, while in the observed trend it also spreads over the ice-covered Arctic Ocean. Another difference is that in the approximated estimate the statistical significance of trends covers a much larger area, including many places where the trend is weak. This may be due to the fact that the reconstruction is made in a reduced space with only three independent

components alleviating the impact of statistically less significant processes that may add noise to linear trends.

The distribution of VIIIE trends estimated by the reanalysis (Figs. 3a–c) is similar to the distribution of T1000 trends. The same large-scale features are present in the signal spanning the whole atmosphere. They are, however, smoother with less small-scale spots. One interesting feature is that in December the atmospheric column warming above the Arctic is much weaker than in January and February. The reconstruction by only three atmospheric structures resolves even more closely the spatial distribution of VIIIE trends (Figs. 3d–f). It also shows the increase of the warming trend over the Arctic in January and February. As for T1000 the statistical significance of reconstructed trends covers a much larger area and even weak trends are often statistically significant.

By visualizing the three structures we recognized their similarity to three well-known large-scale weather patterns. Accordingly we named them the North Atlantic Oscillation (NAO), Scandinavian blocking (SB), and El Niño–Southern Oscillation (ENSO). In

FIG. 4. (top) T1000 and (bottom) H1000 patterns related to the NAO for (a),(d) December, (b),(e) January, and (c),(f) February. Multiplication by 4.9 gives T1000 ($10^{-2} \text{ K yr}^{-1}$) and H1000 ($10^{-2} \text{ m yr}^{-1}$) trends that are statistically significant.

the following subsections each structure will be described in more detail.

b. NAO

The NAO is characterized by the difference between the sea level pressure over the Azores and Iceland (Walker and Bliss 1932). Its positive phase is manifested by the low sea level pressure anomaly over the Arctic and Iceland with mild and humid winters in northern Europe and the east coast of North America contrasted by warm and dry winters over the Mediterranean.

In January and February T1000 and H1000 trends associated with the first atmospheric structure clearly depict the tendency toward the negative phase of the NAO (Fig. 4). H1000 has the maximum positive tendency close to Iceland, which is accompanied with the negative tendency over the North American coast, the North Atlantic Ocean, and the Mediterranean. T1000 increases over the Arctic with the maximum located along the western coast of Greenland. It also decreases over northern Europe and the northern part of Siberia.

On the other hand, in December H1000 associated with this structure has a tendency to create the anticyclonic anomaly over Scandinavia (Fig. 4d). It is further associated with the T1000 dipole having warming tendency maxima over the western coast of Greenland and the Barents Sea and the minimum over central Siberia (Fig. 4a). The anticyclone is apparently related to the sea ice retreat in the Barents Sea because by assuming the geostrophy it can be deduced that it sustains the advection of the warm air to the Arctic. It also precedes the formation of the negative phase of the NAO in January (Fig. 4b). Vertically integrated energy tendencies also indicate the formation of negative NAO in January (Fig. 5). However, already in December VIIE and VIPE trends reassemble the negative phase of the Arctic Oscillation (AO) (Thompson and Wallace 1998), and in January the negative phase of AO intensifies (Figs. 5d,e). In January and February tendencies of vertically integrated kinetic energy in the NAO pattern are consistent with the VIPE trend, indicating the southward shift of the jet stream above the Atlantic Ocean (supplementary Figs. S2b,c).

FIG. 5. (top) VIIIE and (bottom) VIPE patterns related to the NAO for (a),(d) December, (b),(e) January, and (c),(f) February. Multiplication by 4.9 gives the corresponding trends (kJ yr^{-1}) that are statistically significant.

A sea level pressure pattern very similar to the NAO already appeared in Fig. 8 in Hannachi et al. (2009). By using energy anomalies in the statistical analysis here it is shown how the near-surface pattern is consistently connected to other levels and parameters.

The hemispheric mean of the T1000 linear tendency associated with the NAO is $0.70 \times 10^{-2} \text{ K yr}^{-1}$. The warming trend is also present in the atmospheric average of VIIIE, being equal to 3.09 kJ yr^{-1} .

Consistent with previous observations (Walker and Bliss 1932), in January and February the negative NAO tendency is associated with the cold T1000 trend over northern Europe (Figs. 3b,c). In February the cold trend is also present over the east coast of North America. Negative NAO further produces a cyclonic trend over the North Atlantic and Europe (Figs. 3e,f), which advects the moist air from the North Atlantic enhancing the precipitation over the Mediterranean.

c. ENSO

The second structure is ENSO (Walker 1924) which is described by wind, sea surface temperature, and

convective precipitation anomalies over the equatorial Pacific. Its trend shown in Fig. 6 reduces T1000 along the equatorial Pacific toward a more La Niña-like state. The impact of the change of the surface temperature on the whole troposphere is visible in VIIIE fields (Figs. 7a–c). There is a negative trend of VIIIE over a broad area of the equatorial Pacific with the minimum positioned at 15°N and 150°W . Theory and observations (e.g., Wallace et al. 1998) relate the internal energy production in ENSO to the latent heat release in the upper troposphere with a similar distribution of heat anomalies along the surface and in the troposphere as in the structure shown in Figs. 6 and 7.

The corresponding VIPE trend (Figs. 7d–f) has a form of a planetary standing wave in the midlatitudes. In accordance with numerical experiments studying the interaction between tropical and extratropical waves (Jin and Hoskins 1995) it is initiated in the Pacific and propagates with decay coinciding with alternating positive and negative linear tendencies of VIIIE.

The hemispheric mean of the T1000 trend in winter of $0.03 \times 10^{-2} \text{ K yr}^{-1}$ is negligible in comparison to the one

FIG. 6. (top) T1000 and (bottom) H1000 patterns related to ENSO for (a),(d) December, (b),(e) January, and (c),(f) February. Multiplication by 2.6 gives T1000 ($10^{-2} \text{ K yr}^{-1}$) and H1000 ($10^{-2} \text{ m yr}^{-1}$) trends that are statistically significant.

by the NAO. On the other hand, the VIIE trend of -1.5 kJ yr^{-1} has a significant contribution to its total linear tendency. The discrepancy between the intensity of near-surface and total atmospheric column trends of temperature associated with ENSO is due to the fact that in ENSO a small change of ocean surface temperature reflects a large change of heat stored in the ocean's mixed layer and, therefore, strongly impacts the deep atmospheric convection and the anomaly of VIIE (e.g., Wallace et al. 1998).

The impact of this atmospheric structure is mainly localized to the tropics and midlatitudes. There seems to be no evidence of a strong connection with the Arctic sea ice melting. It should be noted that here the ENSO pattern does not represent the full variability in the tropics. The next section will describe a pattern linking the tropics and the Arctic.

In February the structure forms a weak positive phase of the NAO and AO (Figs. 6f and 7f). Eventually, midlatitude planetary waves shown in Fig. 7 excite the stratospheric vortex and the cooling of the tropical

Pacific imposes the positive phase of the AO in the stratosphere (e.g., Manzini et al. 2006) that is then reflected in the VIPE and H1000 fields over the mid- and high latitudes. In February the vertical integral of kinetic energy indicates the trend of the northward shift of the jet stream above the Atlantic Ocean (supplementary Fig. S2f) in agreement with the positive NAO trend.

d. SB

The T1000 linear tendency from the third structure (Figs. 8a–c) shows a strong warming in the Barents Sea throughout winter. Maxima are present in January and February when there is also a strong cooling over Siberia. The near-surface circulation tendency indicated by H1000 (Figs. 8d–f) shows a dipole between the high pressure over the western coast of North America and low pressure over the Pacific Ocean in December that evolves into the high pressure over the Pacific Ocean in January and February. In January and February the high pressure anomaly also develops and remains over Scandinavia and the northwestern part of Siberia.

FIG. 7. (top) VIIIE and (bottom) VIPE patterns related to ENSO for (a),(d) December, (b),(e) January, and (c),(f) February. Multiplication by 2.6 gives the corresponding trends (kJ yr^{-1}) that are statistically significant.

Because of the prevailing anticyclonic anomaly over Scandinavia we name this structure Scandinavian blocking (Miles 1961).

The anticyclone over Scandinavia links the warming of T1000 over the Barents Sea with the cooling over Siberia (e.g., Mori et al. 2014). In a dynamically consistent way it advects the warm air from the south to the Barents Sea and the cold air from the Arctic into Siberia. The warming over the Barents Sea is, however, present in December, before the development of the anticyclone and the cooling in Siberia in January and February. An important feature of this structure is the cold tendency of T1000 along the equatorial Pacific (Figs. 8a–c) coexisting with the warm tendency over the rest of the Pacific Ocean. The cooling over the equatorial Pacific is even more visible in the VIIIE tendency (Figs. 9a–c), indicating the similarity with the negative ENSO trend leading to cooler near-surface temperatures and less intense latent heat release owing to the weaker deep convection in the equatorial Pacific. In February this trend is related to the anticyclonic VIPE trend (Figs. 9d–f)

and the negative trend of the vertically integrated kinetic energy over the Pacific Ocean, indicating the weakening of the jet stream (supplementary Fig. S2i).

Again it should be noted that a sea level pressure pattern similar to SB already appeared in Fig. 8 in Hannachi et al. (2009). Even in this case, the application of energy anomalies in the statistical analysis consistently links near-surface pressure anomalies with other atmospheric levels and parameters.

The hemispheric mean tendency of T1000 explained by SB is $0.17 \times 10^{-2} \text{ K yr}^{-1}$. Although there is a large cooling area along the equatorial Pacific the warming over the Arctic is more significant. On the other hand, the large latent heat release anomaly over the equatorial Pacific in the upper troposphere results in the overall negative VIIIE trend of -1.01 kJ yr^{-1} .

The anticyclonic trend over Scandinavia (Figs. 9d–f) brings colder T1000 over Europe (Figs. 8a–c) (Miles 1961). In January and February the cold tendency is associated with the cyclonic circulation trend that advects the moist air from the North Atlantic toward the

FIG. 8. (top) T1000 and (bottom) H1000 patterns related to SB for (a),(d) December, (b),(e) January, and (c),(f) February. Multiplication by 4.2 gives T1000 (10^{-2} K yr $^{-1}$) and H1000 (10^{-2} m yr $^{-1}$) trends that are statistically significant.

Mediterranean, eventually impacting the precipitation in the area.

e. Approximation of reanalysis trends in ERA-Interim

T1000 trends from ERA-Interim are very similar to those from NCEP (supplementary material Figs. S3a–c). They show maximum positive values over the Arctic and negative over the equatorial Pacific and Siberia. The three patterns in ERA-Interim are selected as those with the minimum root-mean-square difference of energy on all atmospheric levels from the three patterns found in the NCEP reanalysis. Trends of T1000 reconstructed by the three patterns (supplementary material Figs. S3d–f) also approximate the spatial distribution of reanalysis estimates, although, as for NCEP, the intensity is mostly underestimated.

Over the Arctic and the midlatitudes, estimates of VIIE trends by ERA-Interim are very similar to those from NCEP, but over the tropics ERA-Interim shows stronger positive values (supplementary material Figs. S3g–i). The reason for this difference is that in the

lower stratosphere between 100 and 50 mb NCEP has a negative trend of temperature, while ERA-Interim has no trend and, therefore, does not compensate the positive temperature trend in the troposphere present in both reanalyses (not shown). The disagreement in the lower stratosphere over the tropics may be due to the different processing of radiance observations in the two reanalyses. The reconstruction by the three patterns captures the major spatial features of VIIE trends over the Arctic and the midlatitudes but does not reproduce the large-scale warming over the tropics estimated by ERA-Interim (supplementary material Figs. S3j–l).

The three patterns from ERA-Interim that are the most similar to the NAO, ENSO, and SB patterns from NCEP are shown in Figs. S4–S6. They show the same structure and planetary-scale waves as NCEP patterns, but there are some differences. First, the T1000 warming has a weaker maximum in the Barents Sea in December as represented by the NAO pattern. Second in ERA-Interim only NAO has a statistically significant trend with the hemispheric mean of T1000 trend of 0.24×10^{-2} K yr $^{-1}$. The SB

FIG. 9. (top) VIIIE and (bottom) VIPE patterns related to SB for (a),(d) December, (b),(e) January, and (c),(f) February. Multiplication by 4.2 gives the corresponding trends (kJ yr^{-1}) that are statistically significant.

trend becomes significant at the 0.08 level with the hemispheric mean of $0.07 \times 10^{-2} \text{ K yr}^{-1}$, while ENSO has no trend at all. Nevertheless, the spatial distribution of T1000 is again almost exclusively resolved by NAO and SB because among all six independent components only NAO and SB have trends that are statistically significant above the 0.1 level. The value of 0.08 by the SB trend is also much lower than the minimum of 0.27 reached by all remaining independent components in both NCEP and ERA-Interim.

Apparently the higher spatial resolution of the model and the different data assimilation scheme in ERA-Interim result in somewhat different estimates of the main processes carrying the hemispheric warming in winter. The atmospheric pattern with the major trend is NAO, followed by a much weaker SB, while the trend of ENSO is statistically insignificant.

f. NAO and SB evolutions and the Scandinavian anticyclone

The lagged correlation between the formation of the negative NAO phase and the presence of the anticyclone

over the Barents Sea has been found by [Feldstein and Lee \(2014\)](#) and explained by the impact of the anticyclone on the weakening of the polar vortex by the wave flux into the stratosphere. In the NAO pattern the anticyclonic anomaly over Scandinavia in December ([Fig. 4d](#)) extends up to the top of the troposphere ([Fig. 10c](#)). In agreement with numerical experiments on the generation of stratospheric warming events ([Matsuno 1971](#)) the anticyclone, in the form of a planetary wave with a low wavenumber that extends to the top of the troposphere, may impact the stratospheric vortex and induce the stratospheric warming. In particular it has been found that the anticyclone over eastern Europe coupled with the anticyclone over the North Pacific, as in [Figs. 4d](#) and [10c](#), is significantly correlated with the stratospheric vortex weakening because the Eliassen–Palm flux associated with this type of the tropospheric perturbation may effectively penetrate into the stratosphere ([Garfinkel et al. 2010](#)). In accordance with previous observational studies (e.g., [Baldwin and Dunkerton 1999](#)), the warming tendency propagates downward into the troposphere ([Fig. 10d](#)) as the

FIG. 10. Geopotential height patterns related to the NAO at (top) 100 and (bottom) 200 mb for (a),(c) December and (b),(d) January. Multiplication by 4.9 gives trends ($10^{-2} \text{ m yr}^{-1}$).

negative AO structure. The evolution of the NAO pattern from December to February is, thus, in agreement with findings by [Feldstein and Lee \(2014\)](#) indicating the same type of the process linking the troposphere with the stratosphere.

On the other hand, the Scandinavian anticyclone anomaly is present also in the SB pattern where it does not evolve into the AO anomaly ([Figs. 8d–f](#) and [9d–f](#)). In December there is a planetary wave initiated by the negative phase of ENSO that decays from the Pacific to Europe ([Fig. 8d](#)). In January, while being in phase with the anticyclone developed over Scandinavia, the standing planetary wave becomes remotely amplified ([Fig. 8e](#)). In February the amplified

anticyclonic perturbation over the Barents Sea is further in phase with the anticyclone over the Pacific Ocean ([Fig. 8f](#)).

An explanation for the absence of the AO anomaly emerging from the Scandinavian anticyclone in the SB pattern is again given by the dynamical interaction between the polar troposphere and the stratosphere. With the presence of ENSO the stratospheric vortex is remotely forced to be in phase with it (e.g., [Manzini et al. 2006](#)). It is evident in [Fig. 11](#) that in December and January the negative phase of ENSO is associated with the negative polar vortex anomaly, while the tropospheric anomaly of geopotential over Scandinavia has the opposite sign ([Fig. 9d](#)). It is, therefore, more difficult

FIG. 11. Geopotential height patterns related to SB at (top) 50 and (bottom) 150 mb for (a),(c) December and (b),(d) January. Multiplication by 4.2 gives trends ($10^{-2} \text{ m yr}^{-1}$).

to trigger the AO anomaly as in the case when the tropospheric perturbation is in phase with the anomaly of the stratospheric vortex. Instead, opposite signs of perturbations in the equatorial Pacific and in Scandinavia maintain a planetary wave trapped in the troposphere.

In summary, in the NAO pattern the AO anomaly is preconditioned in December by the preexisting stratospheric vortex anomaly of the same sign (Fig. 10). Only under these specific conditions does the tropospheric anomaly activate the barotropic AO response in the later part of winter. On the other hand, in the SB pattern tropospheric and stratospheric anomalies at

high latitudes have opposite signs (Fig. 11), and the tropospheric anomaly does not penetrate into the stratosphere.

The anticyclone over Scandinavia is apparently linked to the warming trend and the sea ice melting trend over the Barents Sea. The correlation between the sea ice melting in early autumn and both tropospheric and stratospheric wave activity in the following winter has also been found previously (Jaiser et al. 2013). Here, as in Feldstein and Lee (2014) and the modeling experiment by Deser et al. (2007), atmospheric processes are connected to the sea ice melting in the Arctic at shorter time scales. The nature of the link between the

FIG. 12. Hemispheric mean of anomalies of reanalyzed (blue) and reconstructed (red) (a) T1000 (K), (b) VIIIE (kJ), (c) T1000 estimated by each structure, and (d) VIIIE estimated by each structure. The reanalyzed trend of T1000 is $2.0 \times 10^{-2} \text{ K yr}^{-1}$, while the reconstructed is $0.9 \times 10^{-2} \text{ K yr}^{-1}$. Both reanalyses and reconstructed trends of VIIIE are statistically insignificant.

sea ice melting and the Scandinavian anticyclone will be discussed in section 5.

4. Temporal evolution of hemispheric averages

The observed mean hemispheric value of T1000 shows a high interannual variability and a clear trend of $2.00 \times 10^{-2} \text{ K yr}^{-1}$ (Fig. 12a). The high-frequency variability of the signal reconstructed by the three independent structures does not always coincide with the reanalyzed one. On the other hand, the reconstructed trend is positive and equal to $0.90 \times 10^{-2} \text{ K yr}^{-1}$ (i.e., about one-half of the reanalyzed trend).

There may be several possible reasons for the discrepancy at the high frequency. First, our statistical analysis was made for the whole three-dimensional structure of the atmosphere, while the surface is only a two-dimensional slice of this field. Second, the explanation for the intensity of the mean surface forcing may require the study of other processes that are not governed by the atmospheric dynamics at large scales, like for example the impact of ocean or soil temperature anomalies that remain inside the mixed layer and have a small influence on the whole atmospheric column. Furthermore, the hemispheric mean value is much smaller than its magnitude in areas with the largest temperature variability and, although the statistical method reconstructs the spatial structure well, it may be less accurate for the small hemispheric mean. It may be further noted that a major underestimation of the T1000 trend is over the ice-covered Arctic (Fig. 2) where the reanalysis is mainly a modeling estimate lacking in situ and satellite observations.

The reconstruction of the temporal variability of the mean hemispheric VIIIE by the three independent

patterns is much closer to the variability of the observed signal (Fig. 12b). A reason for the better correspondence may be that in this case the signal is the vertical integral of the three-dimensional field, which may be better approximated. Another reason may be that this signal is fully described by the atmospheric dynamics making it easily detectable by the atmospheric energy variation. Both the original and the reconstructed signals of VIIIE have statistically insignificant trends.

The hemispheric mean of the NAO structure shows the high-frequency variability of T1000, while SB and ENSO have much smaller amplitudes (Fig. 12c). The variability of the hemispheric mean of VIIIE attributed to the NAO is also very significant, but this time, as a result of the latent heat release over the equatorial Pacific, ENSO and SB also show strong variabilities at high frequency (Fig. 12d).

Table 1 shows correlations between reanalyzed and reconstructed NAO indices. The correlation between the NAO index and the NAO structure is 0.5. The combination mainly with the ENSO structure increases the correlation to 0.6. Clearly there may also be many other manifestations of the NAO index that are not connected to the dynamical processes described by the three structures approximating the near-surface temperature trend. The correlation with the ENSO index of 0.9 is exclusively accomplished by ENSO and SB structures because the NAO has a negligible correlation. The high correlation coefficient indicates that the ENSO index is probably well captured by atmospheric structures that carry a significant part of T1000 and atmospheric energy trends.

Recently the negative trend of the ENSO index has been observed, and modeling studies demonstrated that it could have an impact on the trend of the hemispheric

TABLE 1. Correlations between the three structures, the NAO index in January and February, and the ENSO index from December to February (correlation coefficients equal to or larger than 0.3 are statistically significant). The NAO index is computed as the gradient of H1000 between grid points nearest to the Azores and Iceland corresponding to the original definition with the sea level pressure by Walker and Bliss (1932). The ENSO index is obtained as the H1000 gradient between Tahiti and Darwin also corresponding to the original definition with the sea level pressure by Walker (1924).

	NAO structure	ENSO structure	SB structure	NAO + ENSO + SB
NAO index	0.5	0.3	0.0	0.6
ENSO index	0.1	0.7	0.5	0.9

mean of the near-surface temperature (Kosaka and Xie 2013; Trenberth and Fasullo 2013). Here it is shown that the negative ENSO index trend has two parts. The first part is due to the ENSO structure described in section 3c. This structure is not apparently connected with the Arctic. The second part is due to SB, which in the equatorial Pacific has an equally important cooling trend and is apparently strongly coupled with winter sea ice melting in the Barents Sea and the anticyclone formation over Scandinavia.

The ERA-Interim estimate of the hemispheric mean of trends is statistically significant with values equal to $1.11 \times 10^{-2} \text{ K yr}^{-1}$ for T1000 and $5.6 \times 10^{-2} \text{ kJ yr}^{-1}$ for VIIE. It has a lower estimate for the T1000 trend and a higher estimate for the VIIE trend than the NCEP reanalysis. The hemispheric mean of the reconstruction of the ERA-Interim trend by the three components is also statistically significant and is equal to $0.28 \times 10^{-2} \text{ K yr}^{-1}$ for T1000 and $2.8 \times 10^{-2} \text{ kJ yr}^{-1}$ for VIIE. In ERA-Interim the trend estimated by the three independent components is almost identical to the one obtained by the reduced subspace with six principal components (not shown), indicating that in ERA-Interim about two-thirds of the T1000 trend is described by processes that are not related to the most significant part of the variance of the atmospheric energy at the global scale.

5. Relationship between the sea ice melting and the anticyclone formation

The sea ice reduction along the northern edge of the Barents Sea is often coupled with the formation of the anticyclonic anomaly over Scandinavia (e.g., Sato et al. 2014; Screen et al. 2013; Cohen et al. 2014). This is clearly visible in Fig. 4 in December and in Fig. 9 in January and February. However, Fig. 9 shows that the Barents Sea ice melting in December is not always connected with the anticyclone over Scandinavia. The sea ice melting and the northward advection of warm air seem to be dynamically consistent, indicating a positive feedback. By only visualizing patterns it is, however, not possible to evaluate whether the remotely forced anticyclone determines the sea ice melting by advecting the

warm air to the northern part of the Barents Sea, as proposed, for example, by Sato et al. (2014), or the sea ice melting defines the near-surface pressure gradient anomalies imposing the formation of the anticyclone, as proposed, for example, by Screen et al. (2013) and Cohen et al. (2014).

On the other hand, the ICA method finds independent components by evaluating the full statistical distribution and not only covariances. Because of this property, unlike covariance-based methods, ICA may estimate implications between observed variables. In the case of two variables x_1 and x_2 with non-Gaussian distributions, the statistical model may be estimated by performing the ICA decomposition on the two variables and by evaluating their mixing matrix (Hyvärinen 2013). If x_1 implies x_2 the ICA model should have the following form:

$$\begin{bmatrix} x_1 \\ x_2 \end{bmatrix} = \begin{bmatrix} 1 & 0 \\ b & 1 \end{bmatrix} \begin{bmatrix} s \\ e \end{bmatrix}. \quad (4)$$

If x_2 implies x_1 its form is as follows:

$$\begin{bmatrix} x_1 \\ x_2 \end{bmatrix} = \begin{bmatrix} b & 1 \\ 1 & 0 \end{bmatrix} \begin{bmatrix} s \\ e \end{bmatrix}. \quad (5)$$

Variable s is the independent component, e is the residual, and b is the mapping coefficient. When implications exist in both directions models will not have exact forms like in Eqs. (4) and (5), but the similarity to the model in Eq. (4) would indicate that x_1 mostly implies x_2 , and the similarity to the model described by Eq. (5) would indicate that x_2 mostly implies x_1 .

Variable x_1 will be defined as the anomaly of internal energy at 1000 mb at 80°N and from 37.5° to 42.5°E , and variable x_2 as the anomaly of potential energy at 1000 mb from 67.5° to 72.5°N and from 45° to 50°E (supplementary Fig. S7). They will represent the near-surface temperature change over the sea ice melting area in the Barents Sea and the geopotential change over Scandinavia, respectively.

Mixing matrices estimated by ICA for each winter month are shown in Table 2. The mapping between variables is positive, indicating that the temperature

TABLE 2. Mixing matrices for 1000-mb internal energy over the area with the sea ice retreat (variable x_1) and 1000-mb potential energy east of Scandinavia (variable x_2).

	December	January	February
Mixing matrix	$\begin{vmatrix} 1.0 & -0.6 \\ 1.5 & 1.0 \end{vmatrix}$	$\begin{vmatrix} 1.0 & -0.6 \\ 1.5 & 1.0 \end{vmatrix}$	$\begin{vmatrix} 0.9 & 1.0 \\ 1.0 & -0.3 \end{vmatrix}$

anomaly over the sea ice melting area is in phase with the anticyclone. The remaining upper and lower off-diagonal elements in matrices are not exactly equal to zero as they should be if the impact is only in one direction. Instead they indicate that both features may imply each other. Nevertheless, the relative size of matrix elements indicates that in December and January the near-surface warming and the ice melting in the Barents Sea mostly imply the anticyclone formation, while in February the anticyclone trend mostly enhances the near-surface warming and the sea ice melting.

It can be further seen in Figs. 4 and 9 that the near-surface temperature anomaly in the Barents Sea often has the opposite sign from the anomaly over Siberia. Even in this case there may be a positive feedback. The anticyclone may sustain the advection of cold air to Siberia, while it has been proposed that the snow cover change, originating from the sea ice cover change in autumn and impacting the temperature anomaly over Siberia, may also influence the anticyclone formation over Scandinavia (e.g., Cohen et al. 2014). An ICA model that considers the near-surface temperature anomaly over Siberia may, therefore, also provide information on the relationship between the Arctic sea ice melting and the Scandinavian anticyclone. To test this hypothesis, in the second model x_1 was defined as the internal energy at 1000 mb from 57.5° to 60°N and from 87.5° to 92.5°E (i.e., over the area with the negative near-surface temperature trend in Siberia), while x_2 remained the same as in the first model (supplementary Fig. S7).

Table 3 shows that this time the linear relationship between x_1 and x_2 is negative, linking the negative temperature anomaly with the anticyclone. The comparison of mixing matrices shown in Table 3 with their forms shown in Eqs. (2) and (3) indicates once again that in December and January the temperature anomaly over Siberia mostly implies the geopotential anomaly over Scandinavia, while in February the geopotential anomaly implies the temperature anomaly.

6. Discussion

The study shows that in the NCEP reanalysis only three large-scale atmospheric patterns by combined action may approximate the spatial variability of reanalysis trends of near-surface temperature over the Northern Hemisphere in winter. Two of them, the North Atlantic

TABLE 3. Mixing matrices for 1000-mb internal energy over the area with the snow cover anomaly in Siberia (variable x_1) and 1000-mb potential energy east of Scandinavia (variable x_2).

	December	January	February
Mixing matrix	$\begin{vmatrix} 1.0 & -0.3 \\ -1.0 & 1.0 \end{vmatrix}$	$\begin{vmatrix} 1.0 & 0.4 \\ -2.5 & 1.0 \end{vmatrix}$	$\begin{vmatrix} -1.6 & 1.0 \\ 1.0 & 0.0 \end{vmatrix}$

Oscillation (NAO) and Scandinavian blocking (SB), are directly connected to the sea ice melting in the Barents Sea, while the third, El Niño–Southern Oscillation (ENSO), is mainly limited to the lower and midlatitudes.

The study further offers an insight into the dynamical mechanism linking the sea ice anomaly in the Barents Sea with NAO and SB patterns. When in December the sea ice melting and the tropospheric anticyclone are in phase with the stratospheric vortex anomaly, the anticyclone may initiate the negative phase of the NAO in January and February that propagates throughout the troposphere and the stratosphere. This process is similar to the one found by Feldstein and Lee (2014), but here it is limited to the case when geopotential height anomalies have the same sign in the troposphere and the stratosphere. On the other hand, when the tropospheric anticyclone and the sea ice melting have opposite signs of the stratospheric vortex, the anticyclone remains trapped in the troposphere. In that case they also have the opposite sign of the oscillation over the tropical Pacific.

By determining the set of hemispheric patterns linked to the near-surface warming trend we, therefore, find that the NAO and SB are governed by the difference in the phase between the tropospheric height anomaly over Scandinavia and the anomaly of the stratospheric vortex. Depending on this difference, the tropospheric height anomaly either induces the negative phase of the NAO or eventually sustains the planetary wave initiated over the tropical Pacific.

The third atmospheric structure, ENSO, also has a negative tendency over the tropical Pacific, but it does not seem to be connected with the dynamical process happening over areas with the largest sea ice retreat in winter.

In NAO and SB the sea ice anomaly in the Barents Sea is directly linked to the occurrence of the geopotential height anomaly over Scandinavia. As dynamically any of these two features may initiate and maintain the other one, it may be important to estimate which one of them is independent and which one is implied. The statistical evaluation assuming a linear relationship has shown that, although both features have forced each other during the last three and a half decades, in December and January the sea ice melting in the Arctic mostly implied the anticyclone over Scandinavia, while in February the anticyclone mostly implied the sea ice melting.

It should be emphasized that in the absence of the direct atmospheric forcing the sea ice melting in the Barents Sea is driven by the warm ocean anomaly. The ocean anomalies last longer and should be formed before winter either owing to heating anomalies during the preceding months or years or owing to the increased advection of warm Atlantic waters from the south (e.g., [Smedsrud et al. 2013](#)). Both of these processes may be driven by anomalies in the form of other atmospheric patterns or may be connected to oscillations at longer time scales, but these lagged implications were not considered in the study.

In the future, in order to further describe the nature of remote relationships in the large-scale atmospheric patterns between the tropics and the Arctic, it may be useful to study latitudinal wave fluxes of full atmospheric fields. It may be also possible to find patterns of energy anomalies from an ocean reanalysis corresponding to atmospheric independent components.

Understanding processes connecting the sea ice retreat in winter with large-scale atmospheric structures is especially important for interpreting recent weather trends over the eastern coast of North America and Europe. The more frequent and stronger negative phase of NAO may bring colder temperatures to the eastern coast of North America. It may further produce drier conditions in northern Europe and more humid conditions in southern Europe and the Mediterranean. The positive phase of SB may also bring colder and more humid weather over western Europe and the Mediterranean. On the other hand, in December and January both the NAO and SB appear to be frequently implied by the sea ice melting anomaly in the Arctic. In this way, although NAO and SB tendencies often initiated by the sea ice melting carry hemispheric increases of near-surface temperature, they may also strongly impact the negative societal perception of global warming by bringing colder weather events to highly populated areas of Europe and North America.

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APPENDIX

Estimation of Independent Components and Their Relation to REOFs

Although, by requiring the statistical independence of components and avoiding the assumption on the Gaussianity, ICA appears to be more general than the EOF method, it has not been applied frequently in climate

studies. [Hannachi et al. \(2009\)](#) note that ICA is closely related to rotated EOFs (REOFs) and estimate independent components with the algorithm based on squared fourth-order statistics.

After performing the partial decomposition on EOFs represented by matrix $\mathbf{V}$, REOF rotates them by orthogonal matrix $\mathbf{R}$ in a way that

$$\mathbf{B} = \mathbf{VR} \quad (\text{A1})$$

optimizes a specific predefined criterion $f(\mathbf{B})$. Usually the criterion is some kind of simplicity imposed on the REOFs. Each EOF in $\mathbf{V}$ can be weighted by different coefficients before the rotation.

On the other hand, it can be shown that independent components $\mathbf{S}$ may be decomposed by the following:

$$\mathbf{S} = \mathbf{YR}, \quad (\text{A2})$$

where $\mathbf{Y} = \mathbf{XVA}^{-1}$, and $\mathbf{\Lambda}$ is a diagonal containing singular values corresponding to EOFs. Even if instead of EOFs $\mathbf{Y}$ contains principal components, Eqs. (A1) and (A2) have the same form involving a rotation matrix $\mathbf{R}$. The ICA estimate, therefore, may be obtained by rotating principal components in a way to produce $\mathbf{S}$ with maximized independency between components instead of applying the simplicity criterion.

[Hannachi et al. \(2009\)](#) further propose using the algorithm by [Jennrich and Trendafilov \(2005\)](#) to perform ICA. This algorithm finds $\mathbf{R}$ as a rotation that minimizes off-diagonal contributions to the Frobenius norm of the covariance matrix resulting from the matrix with elementwise squares formed by all possible orthogonal rotations of $\mathbf{Y}$. The algorithm is particularly interesting as it directly links ICA with REOF, which is a common method in climate studies.

On the other hand, this study applies the Fastica algorithm by [Hyvärinen and Oja \(2000\)](#). In Fastica independent components are selected in a way that they also maximize the approximation of the negentropy of the random variable J defined as follows:

$$J = H(y_{\text{gauss}}) - H(y), \quad (\text{A3})$$

where y_{gauss} is a random variable with the Gaussian distribution and the same variance as y and, H is entropy. It is demonstrated by the central limit theorem that negentropy is always positive because the Gaussian variable has the largest entropy of all random variables with the same variance. A simple approximation of the negentropy is the following:

$$J \propto \{E[G(y)] - E[G(v)]\}^2, \quad (\text{A4})$$

where ν is a random variable with the Gaussian distribution and unit variance, $E[\]$ is the expectation, and G is an arbitrarily selected nonquadratic function such that its expectation produces a robust approximation of the negentropy (Hyvärinen and Oja 2000). Fastica estimates $\mathbf{S}$ by iteratively minimizing the gradient of the cost function defined in Eq. (A4) (Hyvärinen and Oja 2000). At the maximum of the negentropy random variables will have the most non-Gaussian distributions that, as demonstrated by the central limit theorem, correspond to the most independent components (Hyvärinen and Oja 2000).

The ICA method assumes the absence of the noise in Eq. (1). This may impact the application of ICA in the atmosphere because of the presence of anomalies that are not directly related to the slow temporal change and may mix with estimates of independent components. Here the impact of high-frequency signals is reduced by studying anomalies of monthly averages. Furthermore, the noise is alleviated by performing ICA on a reduced subspace defined by several leading principal components. The number of principal components is chosen arbitrarily as a trade-off between suppressing the noise and resolving as much as possible sources of observed warming patterns.

Together with other methods that do not assume Gaussianity, like self-organizing maps (SOM) (e.g., Feldstein and Lee 2014) based on the neural network learning theory and probabilistic graphical models (PGMs) based on Bayesian and Markov networks (e.g., Runge et al. 2014), ICA may contribute to the better understanding of climate processes. Compared to SOM and PGM the ICA method may eventually produce unphysical patterns because its outcome depends on the ability of the partial EOF decomposition to correctly select the dynamically relevant subspace without the contribution of the noise. On the other hand, ICA has the advantage of providing a generalization of the well-known EOF method, and consequently it may be directly applied with large dimensions in the physical space on three-dimensional and multivariate atmospheric anomalies.

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