The Relationship between Recent Arctic Amplification and Extreme Mid-Latitude Weather

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The Arctic has warmed more than twice as fast as the global average, a phenomenon known as Arctic amplification. The rapid Arctic warming has contributed to dramatic melting of Arctic sea ice and spring snow cover, at a pace greater than simulated by climate models. These profound changes to the Arctic system have coincided with a period of ostensibly more frequent events of extreme weather across the Northern Hemisphere mid-latitudes, including recent severe winters. The possible link between Arctic change and mid-latitude weather has spurred a rush of new observational and modeling studies. These studies can be broadly summarized as showing three potential dynamical pathways that link Arctic amplification to mid-latitude weather—changes in: storm tracks, the jet stream, and planetary-waves and their associated energy propagation. Through changes in these key atmospheric features, it is possible for sea ice and snow cover to jointly influence mid-latitude weather. However, large uncertainties remain owing to incomplete knowledge of how high-latitude climate change influences these phenomena, sparse and short data records, and imperfect models. This highlights the importance of improved process understanding, sustained and even new Arctic observations, and better coordinated modeling studies to advance our knowledge of mid-latitude weather and extreme events.

The Arctic cryosphere is an integral part of Earth's climate system and has undergone unprecedented changes within the past few decades. Rapid warming and sea ice loss has had significant impacts locally, particularly in late summer and early fall. September sea ice has declined at a rate of 12.4% per decade since 1979[1] so that by the summer of 2012, nearly half of the areal coverage had disappeared. This decrease in ice extent has been accompanied by an approximately 1.8m (40%) decrease in mean ice thickness since 1980[2] and a 75-80% loss in volume[3].

Though sea ice loss has received most of the research and media attention, snow cover in spring and summer has decreased at an even greater rate than has sea ice. June snow cover alone has decreased at nearly double the rate of September sea ice[4]. The decrease in spring snow cover has contributed to both the rise in warm season surface temperatures over the Northern Hemisphere (NH) extratropical landmasses and the decrease in summer Arctic sea ice[5]. The combined rapid loss of sea ice and snow cover in the spring and summer has played a role in amplifying Arctic warming. However, snow cover and sea ice trends diverge in the fall and winter with sea ice decreasing in all months while snow cover has exhibited a neutral to positive trend in fall and winter[6].

Climate Change and Arctic Amplification

While the global-mean surface temperature has unequivocally risen over the instrumental record[7], spatial heterogeneity of this warming plays an important role in the resulting climate impacts. In particular, the near-surface of the NH high latitudes is warming at rates double that of lower latitudes[8,9,10]. This observed phenomenon (see Figs. 1, 2a and 2b) is termed polar or Arctic amplification (AA). AA occurs in all seasons but is strongest in fall and winter. It is also a consistent feature in coupled climate model simulations of the recent past and future projections forced with increased greenhouse gas concentrations[11,12]. Several processes are thought to contribute to AA, including local radiative effects from increased greenhouse gas forcing[12,13], changes in the snow- and ice-albedo feedback induced by a diminishing cryosphere[14,15,16], aerosol concentration changes and deposits of black carbon on snow/ice surfaces[17], changes in Arctic cloud cover and water vapour content[18,19] and a relatively smaller increase in emission of longwave radiation to space in the Arctic compared to the tropics for the same temperature increase[20]. In addition to these local drivers of AA, Arctic temperature change is sensitive to variations in the poleward transport of heat and moisture into the Arctic from lower latitudes[16,21].

Rapid Arctic warming has been accompanied by extensive loss of sea ice[9]. Arctic sea ice strongly modulates near-surface conditions at high latitudes, which then influences regional and, potentially, remote climate. Because open water has a much lower albedo than ice, more sunlight is absorbed at the ocean surface, where sea ice has recently receded in the Arctic. More absorbed energy has resulted in 4-5°C sea surface temperature (SST) anomalies in these newly ice-free regions[22]. However, during autumn when the air cools to temperatures lower than the ocean surface, the excess heat absorbed during summer is transferred from the ocean to the atmosphere via radiative and turbulent fluxes, which strongly warms the lower Arctic troposphere. The additional heat in the system slows the formation of sea ice through winter, both in extent but especially in thickness[23,24]. Hence, winter sea ice has thinned[2], enabling easier melting, fracturing and/or mobility of the ice cover. The increased fraction of open water in winter generates warmer, moister air masses over the Arctic Ocean and nearby continents[15,25], weakening the meridional near-surface temperature gradient. Therefore, these feedbacks indicate that observed Arctic sea ice loss acts as both a response to and a driver of AA.

Mid-latitude Extreme Weather

A large number of extreme heat and rainfall events have been reported over the past decade, especially in the NH mid-latitudes[26-31]. Figure 3 illustrates that several standard extreme temperature and precipitation indices have increased in frequency and intensity over mid-latitude land-areas (20-50°N) with especially rapid changes since the 1990s. For example, the amount of precipitation on very wet days (exceeding the 95th percentile) has increased from 160 mm to 185mm and the percentage of warm days (exceeding the 90th percentile) increased from 10% before 1980 to 16% today[32].

Extreme weather has not been limited to heavy rainfall and warm temperatures and recently has included cold extremes as well. Winter temperatures have generally warmed since 1960 (see Fig. 2a), and the frequency of anomalously cold winter days has decreased over mid-to-high latitudes, but primarily north of 50°N, since 1979 in response to mean warming and decreased variability[33]. However, also evident in Figs. 3d and 3f is that the number of days continuously below freezing has increased and the minimum temperatures have decreased since 1990. Fig. 3h also indicates that the frequency of unusually cold winter months (colder than 2 standard deviations below the 1951–1980 mean[30]) has reversed its longer-term downward trend by the end of the 1990s. This trend reversal in cold extremes has coincided with an acceleration in the rate of warming at high-latitudes relative to the rest of the NH starting approximately in 1990 (Fig. 2b). As seen in Fig. 2c, continental winter temperature trends from 1990 for the continents exhibit cooling trends over the mid-latitudes, replacing the warming trends observed over the longer period (contrast to Fig. 2a). The winter temperature trends shown in Fig. 2c start in 1990 but are not sensitive to the exact start date. However, on average, daily winter cold extremes were less severe over this period than they have been historically[33]. The rapid Arctic warming implies that cold air outbreaks, when Arctic air moves south into the mid-latitudes, are becoming less severe[33].

The seven years between 2007 and 2013 have exhibited the lowest minimum sea ice extents recorded in September since satellite observations began, with an all-time record low in 2007 followed by another in 2012, when sea ice extent fell below 4 million km^2 for the first time in the observational record. Several of these seven winters following the low sea ice minima have been unusually cold across the NH extratropical landmasses[34-38]. The recent winter of 2013–2014 was characterised by record cold and widespread snowstorms across the eastern United States and Canada with the most intense cold-air outbreak in decades associated with the weakening of the polar vortex[39]. The persistent and harsh cold resulted in all-time record cold winters around the Great Lakes of the United States since record keeping began in the 1870s.

The media and public have been quick to make the connection between global, and in particular Arctic, warming and extreme weather[40]. While global warming theory is consistent with record warm temperatures and more intense precipitation events, it does not directly explain cold extremes. Coupled models project boreal winter amplification under greenhouse gas forcing, where the NH landmasses would warm faster in winter relative to the other seasons[11,41]. Warming in the Arctic has continued unabated since at least 1960. Longer-term observed temperature trends in mid-latitudes are consistent with these projections, while shorter-term trends are not. This highlights that results are sensitive to the spatial extent of the analysis, the exact definition used, and especially the duration of an extreme, as extremes of differing durations may be driven by different physical processes.

While cold extremes may be mostly due to natural variability, a growing number of recent studies argue that recent extreme winter weather is related to AA. Three possible dynamical pathways through which AA may influence mid-latitude weather, including extreme weather, are summarized below. We focus our discussion on Arctic linkages to mid-latitude weather in the winter season for two reasons. First, most studies that have linked AA to mid-latitude weather have focused on winter (a brief discussion of proposed linkages in other seasons, mainly summer is provided in the Supplementary Information). Second, winter is the season in which mid-latitude temperature trends have diverged most notably from both model projections and from the other seasons[42]. To provide a focused review, we limit our consideration to the literature concerning recent past (mid-20th century onwards) and present-day climate variability and trends. The implications of projected future AA (e.g. at the end of the 21st century) are likely large and wide-ranging, but are not considered here.

AA Influences and Uncertainties

Whether to attribute severe winter weather to AA or natural variability has emerged as a major debate among scientists[43-45]. In the observations, AA has separated from the noise of natural variability only in the past ~2 decades (see Fig. 2b) presenting a challenge for detection of robust atmospheric responses to AA including mid-latitude weather over such a short time period. In addition to the relatively short length of the observational record, the Arctic is poorly sampled. Another major caveat of any observational study is that correlation alone cannot demonstrate a causal link. Cause and effect can be established through sensitivity or perturbation studies using climate models, but models are subject to their own deficiencies. Known model errors include sea ice-atmosphere coupling[46,47], energy fluxes and cloud properties[47]. Furthermore, modeling studies of the effects of sea ice loss on the large-scale atmospheric circulation have produced conflicting results that make interpretation difficult. Finally our understanding of fundamental driving forces of mid-latitude weather is incomplete[48].

Given these sources of uncertainty, a consensus on whether and how AA is influencing mid-latitude weather is lacking. To facilitate advancement on this important issue, therefore, we synthesize key findings that argue for and against a significant link between AA and mid-latitude weather. All studies agree that the first order impact of sea ice melt is to modify the boundary layer in the Arctic[15,25]. However, if and how that signal propagates out of the Arctic to mid-latitudes differs and can be loosely grouped under three broad dynamical frameworks: 1) changes in storm tracks mainly in the North Atlantic sector, 2) changes in the characteristics of the jet stream and 3) regional changes in the tropospheric circulation that trigger anomalous planetary wave configurations. In Fig. 4 we show the known primary influences on mid-latitude weather including the three dynamical pathways introduced above, and which are described in more detail in the following sections. We recognize that these three pathways are not distinct as they involve dynamical features of the atmospheric circulation that are highly inter-connected. Whilst imperfect, our choice of this separation reflects the different dynamical frameworks that are commonly used—if not explicitly acknowledged—to study the dynamics of mid-latitude weather.

Storm Tracks

Large-scale and low-frequency variability in the extratropical atmosphere is dominated by shifts in storm tracks, often expressed by changes in large-scale atmospheric modes[49]. The dominant atmospheric or climate mode that explains the greatest percentage of the mid- to high-latitude atmospheric variability, including changes in the storm tracks, is the North Atlantic/Arctic Oscillation (N/AO). Changes in the storm tracks associated with the N/AO have a strong influence on the surface temperature and precipitation variability in the North Atlantic sector[50]. When the N/AO is in its positive phase, the storm tracks shift poleward and winters are predominately mild across northern Eurasia and the eastern United States but cold in the Arctic. When the N/AO is in its negative phase, the storm tracks shift equatorward and winters are predominantly more severe across northern Eurasia and the eastern United States, but relatively mild in the Arctic. This temperature pattern is sometimes referred to as the "warm Arctic-cold continents" pattern[51]. Recent observed wintertime temperature trends across the NH continents (see Fig. 2c) project strongly on this temperature anomaly pattern[37], reflecting a negative trend in the N/AO over the last two decades[37]. Given that climate models forced by regional and latitudinal variations in atmospheric heating also exhibit changes in the N/AO[50,52], it is plausible that variability in sea ice and/or snow cover can influence the phase and amplitude of the N/AO, and consequently the storm tracks.

The temperature pattern associated with variations in Eurasian snow cover projects strongly onto the temperature pattern associated with the N/AO and recent temperature trends[34,37,53]. October snow cover anomalies across Eurasia have been proposed as a skillful predictor of the winter N/AO[54,55], where extensive snow cover is associated with a the negative phase of the N/AO, though the relationship may lack stationarity[56]. Satellite-based data indicate a positive trend in Eurasian snow cover during October over the past two to three decades[6,37], though the veracity of these satellite-based increases has recently been questioned[57]. A proposed physical mechanism to explain increased snow cover is that a warmer Arctic atmosphere can hold more water vapour, which enhances precipitation over the Eurasian continent. Additionally, the loss of sea ice — and thus the increase in open water — has increased moisture fluxes to the atmosphere[9]. If near-surface atmospheric temperatures remain sufficiently cold — as is the case in Siberia during fall and winter — any additional precipitation will likely occur as snow[58,59]. Therefore, increasing October Eurasian snow cover may have contributed to the recent tendency towards negative N/AO and cold NH winters[37]. However, given that the N/AO has considerable internal variability on multiple timescales, the recent negative trend may be predominantly internally driven.

The strong decline in sea ice during recent decades has intensified interest in the interactions between sea ice conditions and the atmosphere[47,60]. Most sea ice-atmosphere coupled studies have discussed the atmospheric response in the context of N/AO variability. Observational analyses have shown significant correlation between reduced Arctic sea ice cover and the negative phase of the winter N/AO[35,37,61-64] although it is unclear whether late summer/early fall[35] or late fall/early winter[38] sea ice anomalies are more skillful at predicting the winter weather patterns.

Modeling studies have also examined the N/AO response to variations in Arctic sea ice[35, 65-74], by running simulations forced by past sea ice trends or case studies of years with large sea ice anomalies. These studies have shown a full spectrum of N/AO responses to reduced sea ice, from shifts toward the positive phase[68,71,73], the negative phase[35,65,74] or no significant change[73].

Furthermore, attributing N/AO changes and associated shifts in storm tracks to Arctic forcing has proved very difficult. The simulated atmospheric circulation response to sea ice loss is sensitive to differences in model physics, background atmospheric and oceanic states, and the spatial patterns and magnitude of sea ice anomalies. Additionally, it has proven difficult to separate forced change due to sea ice loss from internal model variability. Large numbers of model runs or ensembles are likely required to achieve statistically significant responses to forced sea ice changes[73]. While these disparities between studies preclude definitive conclusions, two general, results emerge. First, there are more studies that show a negative N/AO response than a positive N/AO response. Second, the simulated N/AO response to sea ice loss is relatively small compared to natural variability. This is consistent with the view that changes in the N/AO are predominately internally driven and do not necessarily require remote forcing[75].

Jet Stream

The second proposed dynamical pathway linking AA to increased weather extremes is through its effects on the behaviour of the polar jet stream. The difference in temperature between the Arctic and mid-latitudes is a fundamental driver of the polar jet stream; therefore a reduced poleward temperature difference could result in a weaker zonal jet with larger meanders. A weaker and more meandering flow may cause weather systems to travel eastward more slowly and thus, all other things being equal, AA could lead to more persistent weather patterns[76]. Furthermore, AA causes the thickness of atmospheric layers to increase more to the north, such that the peaks of atmospheric ridges may elongate northward and thus, increase the north-south amplitude of the flow[76]. Weather extremes frequently occur when atmospheric circulation patterns are persistent, which tends to occur with a strong meridional wind component[77,78].

Some aspects of this hypothesized linkage are supported by observations and in model simulations. A significant decrease in zonal-mean zonal wind at 500 hPa during fall is observed regionally[76,79]. This may be understood through the thermal wind relationship, which states that vertical wind shear is proportional to the meridional temperature gradient. Assuming that the winds do not increase at the surface, the zonal wind at the jet stream level should slacken with a weaker meridional temperature gradient. In other seasons when AA is weaker, no significant trend in zonal-mean zonal wind is observed.

However, challenges remain in linking AA directly to changes in the speed and structure of the jet stream. For example, other factors besides the near-surface meridional temperature gradient influence the zonal jet, including feedbacks from synoptic eddies or storms and the upper-level meridional temperature gradient. Indeed, although AA has weakened the near-surface meridional temperature gradient, the temperature gradient between the tropics and mid-latitudes at higher altitudes has strengthened[80], which would increase jet stream-level winds. Another challenge is identifying how much of the AA is driven by local changes compared to remote changes[16]. This distinction is highly relevant to the current debate on possible Arctic-mid-latitude linkages, because if a significant portion of AA is driven remotely, then AA may be partly viewed as a response to rather than a forcing of mid-latitude weather. This highlights the importance of considering the many ways in which mid-latitude jets are influenced, including the meridional temperature gradient, which are shown schematically in Fig. B1.

Observational support for the follow-on impacts of the hypothesis related to a weakening zonal component of the jet[76] is even less strong — namely, whether AA leads to larger amplitude waves, slower wave propagation speeds, and more persistent weather patterns. Statistically robust evidence of increasing north-south wave amplitude and slower propagation speed has not been established[79,81]. This is not surprising given the recent emergence of AA and the large natural variability of the atmosphere. Recent studies provide tentative evidence for increasing amplitude in summer and fall for some definitions of wave amplitude but not for others[81]. A significant reduction in 500 hPa wave speeds during autumn was reported[79] but the response was not apparent in higher-level winds. The frequency of blocking-high patterns is metric, region and time dependent, but as a whole the observations do not support a significant increase in blocking occurrence over recent decades[82].

The theory that AA is resulting in a slower zonal jet, increased meridional flow, amplified waves and more persistent extreme weather has received a lot of attention from the media, policy makers and climate sceintists[83]. In part due to the high profile, this hypothesis has been scrutinized in the scientific literature more extensively than other hypotheses linking Arctic climate change to mid-latitude weather. However, it is worth noting that other studies on related topics, especially other observational studies, share some of the same shortcomings[35,37,38,61-64] (lack of statistical significance, causality unclear, incomplete mechanistic understanding etc.).

Planetary Waves

Modification to large-scale Rossby waves over Eurasia is the third proposed dynamical pathway linking AA to mid-latitude weather. Both observational analyses and modeling experiments link more extensive snow cover across Eurasia, especially in October, to changes in wave structure at high latitudes. Extensive snow cover may lead to larger planetary waves that increase the vertical propagation of wave energy into the stratosphere, favouring a warmer and weakened stratospheric polar vortex[84-87]. It is proposed that the atmospheric response lags the snow cover changes by a few months because of the response time of the stratospheric circulation and subsequent feedback to the troposphere.

Observed reductions in autumn-winter Arctic sea ice, especially in the Barents-Kara seas (BK), are also correlated with strengthened anticyclonic circulation anomalies over the Arctic Ocean, which tend to induce easterly flow and cold air advection over northern Europe[38,88-90], a link that may be sensitive to the timing of the sea ice anomalies. Winter anomalies trigger an immediate, local and direct atmospheric response forced by increased turbulent heat fluxes locally over the BK, which in turn changes the baroclincity and affects large-scale planetary or Rossby waves in the atmosphere. Alternatively fall sea ice anomalies may force a delayed, remote and indirect and atmospheric response through increased Eurasian snow cover[46] or through altered baroclinicity and high pressure over the BK that force upward propagating planetary waves into the stratosphere. Sufficient wave breaking in the polar stratosphere weakens the stratospheric polar vortex and can trigger a stratospheric warming event. The circulation anomalies associated with a stratospheric warming event propagate back down to the surface in subsequent weeks, contributing to a persistent negative N/AO and cold continental conditions[90,91].

Several modeling studies have used prescribed BK sea ice reductions to examine how the atmosphere responds. Horizontal downstream propagation of the energy away from anomalous, sea ice-induced high pressure over the BK leads to the formation of a trough over Eurasia and subsequent cold continental temperatures[92]. Such model experiments have thus far only included the impact of sea ice changes and not the full extent of AA.

The proposed response of planetary waves to reductions in both snow cover and sea ice has inherent shortcomings. Free-running (i.e., without prescribed forcing) climate models do not simulate well the amplitude or the timing of wave changes to more extensive snow cover as in observations[86], resulting in a simulated weak relationship found between October Eurasian snow cover and the winter N/AO[93]. Regarding the response to sea ice loss, caution is urged because strong trends in the sea ice extent have made analyzing the co-variability between sea ice and the atmosphere difficult to interpret[46]. Furthermore the proposed atmospheric response to sea ice forcing is not robust and has yet to achieve statistical significance[46], in part due to the shortness of the data record.

To conclude, variability in both sea ice and snow cover has been hypothesized independently to force anomalously high geopotential heights in the BK. In Fig. B2, we provide a complementary perspective by proposing a synthesis of how extensive snow cover and reduced sea ice in the fall and early winter can force local changes that constructively interfere to force the same response in the planetary waves, which could influence winter weather patterns.

Synthesis of Arctic and Mid-latitude Linkages

Dramatic changes are occurring in the Arctic climate system, while at the same time, the frequency of mid-latitude extreme weather events appears to have increased. The potential link between AA and changes in extreme weather is clearly a critical one, especially as AA is robustly predicted to continue over the coming decades. The climate dynamics literature concerning Arctic-mid-latitude linkages is currently inconclusive, which may help explain the media portrayal of a polarized view among scientists[81]. Furthermore, the severe winter of 2013–14 across eastern North America focused the debate of whether extreme cold events can be attributed to climate change, including AA, or natural variability[43,44]. Cold winters like that experienced in 2013–14 have occurred before and are expected as part of normal weather variability even in a warmer planet[94]. Numerous studies have presented preliminary evidence that AA and continental weather are linked and a range of dynamical hypotheses for that link exists in the scientific literature. Other studies, however, have presented evidence demonstrating no robust statistical or dynamical link between AA and mid-latitude climate variability.

Nevertheless, dramatic changes to high-latitude sea ice and snow cover have occurred, along with profound impacts at least locally in the Arctic. The most robust atmospheric response to these changes is an altered near surface climate of the Arctic. All studies agree that sea ice loss enhances local warming, which weakens near-surface meridional temperature gradients, moistens the boundary layer, and decreases the near-surface static stability. A growing body of observational, modeling, and theoretical evidence suggests that the impact of high-latitude surface heating increases upper-level geopotential heights, which affects the large-scale atmospheric circulation beyond the Arctic. To first order, amplified warming in the Arctic and a decrease in the meridional temperature gradient should favour a weaker zonal jet. However, whether weaker upper-level zonal winds causes amplified and slower moving planetary waves remains unclear. Further evidence from modeling studies suggests that cryospheric anomalies can alter the stratospheric polar vortex, storm tracks and jet stream - all of which are key drivers of mid-latitude weather and extremes. These changes appear to be more likely in winter than other seasons owing to the large AA signal and divergence of winter temperature trends from the other seasons. The link between reduced Arctic sea ice and cold continental winters is currently the most studied and arguably the best-supported link between AA and mid-latitude extreme weather patterns.

Based on the research conducted to date, we offer a brief perspective on the challenges and research opportunities in the near future (a more detailed list is included in the Supplementary Information). Understanding the relative importance of different forcings mechanisms, and how they interact with internally generated variability, remains a key challenge. Increased and better observations (e.g., ocean-ice-atmosphere energy exchange, cloud cover, and troposphere-stratosphere coupling) would not only improve our understanding of the Arctic and its climate, but also help to elucidate the mechanisms of atmospheric response to AA and better constrain the models. Better standardization of metrics (extremes, blocking, wave amplitude, etc.) and coordination of modeling experiments would allow results to be more directly compared and the current disparities to be better understood. Finally, testing hypotheses in a hierarchy of models of increasing complexity, from simple dynamical models to state-of-the-art earth system models, would help to further our understanding and better equip us to untangle the complexity of Arctic - mid-latitude linkages.

Box B1

The different components of a generalized mid-latitude jet are illustrated in Fig. B1a. The proposed dynamical pathways linking AA to increased weather extremes are through the highly nonlinear interaction between the jet stream, the planetary waves and the storm tracks (see Fig. 4). The wintertime extratropical climate variability is affected by a complex set of interactions and feedbacks between components such as natural variability modes, diabatic heating anomalies due to variations in sea ice and snow cover, and atmospheric and oceanic heat transport from tropical and subtropical latitudes. However, recently it has been proposed that air-sea interaction in the Arctic could be forcing teleconnection patterns and influencing weather patterns remotely in the mid-latitudes by heating the Arctic relative to the rest of the globe[36,76].

A change in the meridional temperature gradient, which projects onto the thermally driven component of the jet may or may not result in a significant change in the jet depending on how the eddy driven part of the jet varies. Complex interactions between the mid-latitude wind jets, the planetary waves and baroclinic weather systems is a nonlinear two-way feedback process, where diabatic heating/cooling, orographic forcing and eddy wave breakings drive the jets and teleconnection patterns. The yellow arrow denotes the final influence, which is of synoptic variability (jet eddies) on midlatitude weather. The dynamical mechanisms associated with each green arrow are as follows:

A. The temperature gradient, in this definition, influences the thermally-driven jet (black dashed circle) via the thermal-wind balance (in combination with boundary conditions).

B. The temperature gradient influences the eddy-driven jet (solid black circle) via changes in baroclinicity. The eddy-driven jet influences the temperature gradient via horizontal heat fluxes. C. The eddy-driven jet affects stratospheric winds (black "U" shape) via vertical wave propagation. Stratospheric winds affect the eddy-driven jet by altering the vertical wave-guide. D. The thermally-driven jet affects stratospheric winds via generation of orographically-forced waves. Stratospheric winds affect the thermally-driven jet by altering the vertical wave guide. E. The thermally-driven jet affects the eddy-driven jet by acting as a wave guide (the role of baroclinicity here directly associated with the temperature gradient). The eddy-driven jet affects the thermally-driven jet via energy fluxes.

As can be seen from the figure, there are many feedbacks and interactions involving mid-latitude jets, with the temperature gradient being just one of them. Therefore a weakening in the temperature gradient may or may not result in a slowing down of the jet depending on the net effect of other factors.

The N/AO may be considered a paradigm for the debate within the climate community. Shown in Fig. B1b are the changes in the atmospheric circulation associated with the negative phase of the N/AO. Positive (negative) zonal wind anomalies associated with the negative N/AO are superimposed on the jet shown by a green solid (dashed) line. Also shown are the temperature changes with warmer temperatures in the Arctic (red) and colder temperatures in the mid-latitudes (blue), increased high-latitude blocking (represented by clockwise flow around a high) and a southward shift in the storm tracks (represented by a counterclockwise flow around a low) and increased meridional flow. All these dynamical changes are observed as the N/AO shifts from its positive phase to the negative phase. However, external forcing, such as a reduced thermal gradient due to AA, will project onto these dynamical patterns associated with the negative N/AO: an equatorward shift in the zonal jet, increased meridional flow, high latitude blocking and a southward shift in storm tracks. The yellow broken arrow denotes uncertainty whether a change in the meridional temperature gradient can force all the other changes depicted in the figure. Attributing observed changes in mid-latitude weather to either AA or internal variability has proven challenging to date.

Box B2

As a summary of the studies presented, in Fig. B2 we synthesize some common ideas about the atmospheric response to sea ice and snow cover variability that have until now been treated independently. All sea ice studies agree that sea ice loss heats and moistens the boundary layer of the Arctic atmosphere. It has also been shown that a surface heat source in the extratropics induces downward descent of air over the heat source, warming the atmospheric column and raising heights in the mid-troposphere, while a trough develops downstream inducing an equatorward flow of cold air[95]. This is consistent with the result that reduced sea ice favors an increase in mid-tropospheric heights in the Barents-Kara seas (BK) region in winter[51,88,92] with downstream troughing over Eurasia. Studies also agree that increased snow cover cools the boundary layer[54]. Therefore a snow-induced surface cooling can lower heights in the mid-troposphere, inducing enhanced ridging upstream.

In September and October, sea ice loss has been most pronounced in the Chukchi and East Siberian seas. Warming of the atmosphere due to increased heating from newly ice-free ocean causes geopotential heights to increase in the mid-troposphere, which suppresses the jet stream southward over east Siberia. This pattern, referred to as the Arctic Dipole, has strengthened during the era of sea ice loss[61]. A southward shift in the storm tracks over East Asia allows for a more rapid advance of Eurasian snow cover in October. Enlarged areas of open water north of Siberia also provide increased moisture flux to the atmosphere, which precipitates as snow as the air mass is advected southward over Siberia[58,71] (left globe).

In October a more extensive snow cover cools the surface leading to lower heights and a trough in the mid-troposphere. Increased troughing over East Asia favours upstream ridging near the BK and the Urals. Concurrently the large sea ice deficits and the associated strong surface heating anomalies migrate from the Chukchi and East Siberian seas in September and October to the BK in November and December. This favours mid-tropospheric ridging in the BK region with downstream troughing over East Asia. Therefore the extensive snow cover over Siberia in October and November and the sea ice loss over the BK in November and December produce same-signed mid-tropospheric geopotential height patterns over Eurasia. This planetary wave configuration is favourable for increased vertical propagation of Rossby waves from the troposphere into the stratosphere[96-98] (middle globe).

Increased vertical propagation of Rossby wave energy from the troposphere to the stratosphere weakens the polar vortex and resulting in a stratospheric warming event. Circulation anomalies associated with the warming event appear first in the stratosphere and subsequently appear in the troposphere in January and February. These circulation anomalies resemble those associated with the negative phase of the N/AO; i.e., ridging over the Arctic especially near Greenland, and a weaker, equatorward-shifted polar jet stream. As a result, warmer conditions prevail in the Arctic regions, but colder and more severe winter weather occurs across the mid-latitude continents with a greater likelihood of snowstorms in the population centers of the NH mid-latitudes (right globe).

We propose a chain of events where less sea ice and increased open water in the Arctic (that heats the atmosphere) and more snow cover (that cools the atmosphere) both force the same pattern, which results in a weakened polar vortex. Because the heating anomalies are displaced longitudinally, extensive Eurasian snow cover and reduced Arctic sea ice can constructively interfere to weaken the polar vortex and hence influence surface weather.

Methods

For Fig. 1 we used the monthly mean fields from the ERA-Interim reanalysis[99] to compute seasonal means for the period March 1979 to February 2014. These data were averaged around circles of latitude (at 1.5° resolution). Standard seasonal means were computed and used. We estimated trends using least-squares linear regression. The statistical significances of the regressions were calculated from a two-tailed t-test.

Surface temperature anomalies for Fig. 2 are taken from the NASA GISS temperature record[100]. The decadal linear trends in surface air temperature anomalies in Fig. 2a are based on a least-squares regression of the December-February (DJF)-mean of monthly-mean temperature anomalies from 1960/61–2013/14. The corresponding time series of DJF temperatures anomalies (middle panel) was constructed by weighting the anomalies by the cosine of latitude. The same convention is used for Fig. 2c except that the linear trends are calculated based on DJF values during the period 1990/91–2013/14.

Figures 3a-3f were created using the GHCNDEX global land gridded dataset of climate extremes[32] available at www.climdex.org. The online data-visualization tool was used to create linear trend maps and timeseries (over the period 1951–2014) for different extreme indices provided in the GHCNDEX global land gridded dataset. Timeseries are area-weighted averages of land regions within the latitudinal belt from 20-50°N. Figures 3g and 3h show the percentage of land in the mid-latitudes with unusually warm summer months or unusually cold winter months[30]. For this, we use monthly gridded data from the NASA-GISS surface temperature dataset with a base period of 1951–1980. First, we determine the local standard deviation due to natural variability at each grid point in the latitudinal belt from 20-50°N for each calendar month of the boreal winter (December-January-February) and boreal summer (June-July-August) seasons. To do so, we apply a singular spectrum analysis to extract the long-term (periods of 30 years or greater) non-linear trend over the 20th century. Next we detrend the original time series by subtracting the long-term trend, which gives the year-to-year variability. From this detrended signal, monthly standard deviations are calculated using the 1951–2010 period, which are then seasonally averaged. For boreal summer, we determine the percentage of land with temperatures warmer than 1 and 2 standard deviations beyond the mean (Fig. 3g). For boreal winter, we determine the percentage of land with temperatures colder than 1 and 2 standard deviations below the mean (Fig. 3h).

References

1. Stroeve, J. C. *et al.* Sea ice response to an extreme negative phase of the Arctic Oscillation during winter 2009/2010. *Geophys. Res. Lett.* **38,** L02502 (2011).

2. Kwok, R. & Rothrock, D. A. Decline in Arctic sea ice thickness from submarine and ICESat records: 1958–2008. *Geophys. Res. Lett.* **36,** L15501 (2009).

3. Overland, J. E., Wang, M., Walsh, J. E. & Stroeve, J. C. Future Arctic climate changes: Adaptation and mitigation timescales. *Earth's Future* **2,** 68–74 (2014).

4. Derksen, C. & Brown, R. Spring snow cover extent reductions in the 2008-2012 period exceeding climate model projections. *Geophys. Res. Lett.* **39,** L19504 (2012).

5. Matsumura, S., Zhang, X. & Yamazaki, K. Summer Arctic atmospheric circulation response to spring Eurasian snow cover and its possible linkage to accelerated sea ice decrease. *J. Clim.* **27,** doi:10.1175/JCLI-D-13-00549.1 (2014).

6. Mudryk, L. R., Kushner, P. J. & Derksen, C. Interpreting observed northern hemisphere snow trends with large ensembles of climate simulations. *Clim. Dyn.* **43**, 345–359 (2013).

7. IPCC. Summary for Policymakers. In: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* [Stocker, T.F., D. Qin, G.-K. Plattner, M. Tignor, S. K. Allen, J. Boschung, A. Nauels, Y. Xia, V. Bex and P.M. Midgley (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA (2013).

8. Serreze, M. C., Barrett, A. P., Stroeve, J. C., Kindig, D. M. & Holland, M. M. The emergence of surface-based Arctic amplification. *Cryosphere* **3,** 11–19 (2009).

9. Screen, J. A. & Simmonds, I. The central role of diminishing sea ice in recent Arctic temperature amplification. *Nature* **464,** 1334–1337 (2010a).

10. Cowtan, K. & Way, R. G. Coverage bias in the HadCRUT4 temperature series and its impact on recent temperature trends .*Q. J. R. Meteorol. Soc.* **133,** 459–77 (2013).

11. Holland, M. M. & Bitz, C. M. Polar amplification of climate change in coupled models. *Clim. Dyn.* **21,** 221–232 (2003).

12. Stroeve, J. C. *et al.* The Arctic's rapidly shrinking sea ice cover: A research synthesis. *Clim. Change* **110,** 1005–1027 (2012).

13. Gillett, N. P. et al. Attribution of polar warming to human influence. *Nature Geosci*. **1,** 750–754 (2008).

14. Winton, M. Amplified Arctic climate change: What does surface albedo feedback have to do with it? *Geophys. Res. Lett*. **33,** L03701 (2006).

15. Serreze, M. C. & Barry, R. G. Processes and impacts of Arctic amplification: A research synthesis. *Global and Planetary Change* **77,** 85–96 (2011).

16. Screen, J. A., Deser, C. & Simmonds, I. Local and remote controls on observed Arctic warming. *Geophys. Res Lett*. **39,** L10709 (2012).

17. Shindell, D. & Faluvegi, G. Climate response to regional radiative forcing during the twentieth century. *Nature Geosci.* **2,** 294–300 (2009).

18. Francis, J. A. & Hunter, E. New insight into the disappearing Arctic sea ice. *EOS Trans. Am. Geophys. Union* **87,** 509–511 (2006).

19. Graverson, R. G. & Wang, M. Polar amplification in a coupled climate model with locked albedo. *Clim. Dyn.* **33,** 629–643 (2009).

20. Pithan, F. & Mauritsen, T. Arctic amplification dominated by temperature feedbacks in contemporary climate models. *Nature Geosci.* **7,** 181–184 (2014).

21. Graversen, R. G., Mauritsen, T., Tjernstrom, M., Kallen, E. & Svensson, G. Vertical structure of recent Arctic warming. *Nature* **451,** 53–56 (2008).

22. Wood, K. R. *et al.* Is there a "new normal" climate in the Beaufort Sea? *Polar Res.* **32,** 19552 (2013).

23. Steele, M., Ermold, W. & Zhang, J. Arctic Ocean surface warming trends over the past 100 years. *Geophys. Res. Lett.* **35,** L19715 (2008).

24. Inoue, J., & Hori, M. E. Arctic cyclogenesis at the marginal ice zone: A contributory mechanism for the temperature amplification? *Geophys. Res. Lett.,* **38,** L12502 (2011).

25. Screen, J. A. & Simmonds, I. Increasing fall-winter energy loss from the Arctic Ocean and its role in Arctic amplification. *Geophys. Res. Lett*. **37,** L16707 (2010b).

26. Min, S. K., Zhang, X., Zwiers, F. W. & Hegerl, G. C. Human contribution to more-intense precipitation extremes. *Nature* **470,** 378–81 (2011).

27. Coumou, D. & Rahmstorf, S. A decade of weather extremes. *Nature Climate Change* **2,** 491–496 (2012).

28. Westra, S., Alexander, L. V. & Zwiers, F. W. Global increasing trends in annual maximum daily precipitation. *J. Clim*. **26,** 3904–3918 (2013).

29. Coumou, D. & Robinson, A. Historic and future increase in the global land area affected by monthly heat extremes. *Environ. Res. Lett.* **8,** 0–6 (2013)

30. Coumou, D., Robinson, A. & Rahmstorf, S. Global increase in record-breaking monthly-mean temperatures. *Climatic Change* **118,** 771 (2013).

31. Seneviratne, S. I., Donat, M. G., Mueller, B. & Alexander, L. V. No pause in the increase of hot temperature extremes. *Nature Climate Change* **4,** 161–163 (2014).

32. Donat, M. G. et al. Global land-based datasets for monitoring climatic extremes. *Bull. Am. Meteorol. Soc.* **94,** 997–1006 (2013).

33. Screen, J. A. Arctic amplification decreases temperature variance in northern mid- to high-latitudes. *Nature Climate Change* **4,** 577–582 (2014).

34. Cohen, J., Barlow, M. & K. Saito, K. Decadal fluctuations in planetary wave forcing modulate global warming in late boreal winter. *J. Clim.* **22,** 4418–4426 (2009).

35. Liu, J., Curry, J. A., Wang, H., Song, M. & Horton, R. Impact of declining Arctic sea ice on winter snow. *Proc. Natl. Acad. Sci.* **109,** 4074–4079 (2012).

36. Greene, C. H. & Monger, B. C. An Arctic wild card in the weather. *Oceanography* **25,** 7–9 (2012).

37. Cohen, J., Furtado, J., Barlow, J. M., Alexeev, V. & Cherry, J. Arctic warming, increasing fall snow cover and widespread boreal winter cooling. *Environ. Res. Lett.* **7,** 014007 (2012a).

38. Tang, Q., Zhang, X., Yang, X. & Francis, J. A. Cold winter extremes in northern continents linked to Arctic sea ice loss. *Environ. Res. Lett*. **8,** 014036 (2013).

39. Tollefson, J. US cold snap fuels climate debate. *Nature News*, Nature Publishing Co. (2014).

40. Hamilton, L. C. & Lemcke-Stampone, M. Arctic warming and your weather: public belief in the connection. *Int. J. Climatol*. **34,** 1723–1728 (2013).

41. Alexeev, V. A., Langen, P. L. & Bates, J. R. Polar amplification of surface warming on an aquaplanet in "ghost forcing" experiments without sea ice feedbacks. *Clim. Dyn.* **24,** 655–666 (2005).

42. Cohen, J., Furtado, J., Barlow, M., Alexeev, V. &. Cherry, J. Asymmetric seasonal temperature trends. *Geophys. Res. Lett.* **39,** L04705 (2012b).

43. Wallace, J. M., Held, I. M., Thompson, D. W. J., Trenberth, K. E. & Walsh, J. E. Global warming and winter weather. *Science* **343,** 729–730 (2014).

44. Palmer, T. Record-breaking winters and global climate change. *Science* **344,** 803–804 (2014).

45. Fischer, E. M. & Knutti, R. Heated debate on cold weather. *Nature Climate Change* **4,** 577-582 (2014).

46. Cohen, J., Jones, J., Furtado, J. C. &. Tziperman, E. Warm Arctic, cold continents: A common pattern related to Arctic sea ice melt, snow advance, and extreme winter weather. *Oceanography* **26,** 150–160 (2013).

47. Vihma, T. Effects of Arctic sea ice decline on weather and climate: a review. *Surveys in Geophysics,* 10.1007/s10712-014-9284-0 (2014).

48. Hoskins, B. Review Article - The potential for skill across the range of the seamless weather-climate prediction problem: a stimulus for our science. *Q.J.R. Meteorol. Soc.,* **139,** 573–584 (2013).

49. Woollings, T. & Blackburn, M. The North Atlantic jet stream under climate change and its relation to the NAO and EA patterns. *J. Clim.* **25,** 886–902 (2012).

50. Bader, J. *et al.* A review on Northern Hemisphere sea-ice, storminess and the North Atlantic Oscillation: Observations and projected changes. *Atmospheric Research* **101,** 809–834 (2011).

51. Overland, J. E., Wood, K. R. & Wang, M. Warm Arctic–cold continents: Impacts of the newly open Arctic Sea. *Polar Res.* **30,** 15787 (2011).

52. Wu, A., Hsieh, W. W., Boer, G. J. & Zwiers, F. W. Changes in the Arctic Oscillation under increased atmospheric greenhouse gases. *Geophys. Res. Lett.* **34,** L12701 (2007).

53. Mote, T. & Kutney, E. Regions of autumn Eurasian snow cover and associations with North American winter temperatures. *Int. J. of Climatology*, **32,** 1164–1177 (2012).

54. Cohen, J. & Entekhabi, D. Eurasian snow cover variability and Northern Hemisphere climate predictability. *Geophys. Res. Lett.* **26,** 345–348 (1999).

55. Cohen J. & Jones, J. A new index for more accurate winter predictions. *Geophys. Res. Lett.* **38,** L21701 (2011b).

56. Peings, Y., Brun, E., Mauvais, V. & Douville, H. How stationary is the Siberian snow - Arctic oscillation relationship over the 20th century? *Geophys. Res. Lett.* **40,** doi:10.1029/2012GL054083 (2013).

57. Brown, R. D. & Derksen, C. Is Eurasian October snow cover extent increasing? *Environ. Res. Lett.* **8,** 024006, doi:10.1088/1748-9326/8/2/024006 (2013).

58. Ghatak, D., Frei, A., Gong, G., Stroeve, J. & Robinson, D. On the emergence of an Arctic amplification signal in terrestrial Arctic snow extent. *J. Geophys. Res.* **115,** doi:10.1029/2010JD014007 (2010).

59. Ghatak, D. *et al.* Simulated Siberian snow cover response to observed Arctic sea ice loss, 1979–2008. *J. Geophys. Res.* **117,** D23108, doi:10.1029/2012JD018047 (2012).

60. Budikova, D. Role of Arctic sea ice in global atmospheric circulation. *Global and Planetary Change* **68,** 149–163 (2009).

61. Francis, J. A., Chan, W., Leathers, D. J., Miller, J. R. & Veron, D. E. Winter Northern Hemisphere weather patterns remember summer Arctic sea-ice extent. *Geophys. Res. Lett.* **36,** L07503, doi:10.1029/2009GL037274 (2009).

62. Overland, J. E. & Wang, M. Large-scale atmospheric circulation changes are associated with the recent loss of Arctic sea ice. *Tellus* **62A,** doi: 10.1111/j.1600-0870.2009.00421.x, 1–9 (2010).

63. Strong, C., Magnusdottir, G. & Stern, H. Observed feedback between winter sea ice and the North Atlantic Oscillation. *J. Clim*. **22,** 6021–6032 (2009).

64. Hopsch, S., Cohen, J. & Dethloff, K. Impact of anomalous fall Arctic sea ice concentration on atmospheric patterns in the following winter. *Tellus A* **64,** 18624 DOI: 10.3402/tellusa.v64i0.18624 (2012).

65. Magnusdottir, G., Deser, C. & Saravanan, R. The effects of North Atlantic SST and sea-ice anomalies on the winter circulation in CCM3. Part I: Main features and storm track characteristics of the response. *J. Clim*. **17,** 857–876 (2004).

66. Deser, C., Magnusdottir, G., Saravanan, R. & Phillips, A. The effects of North Atlantic SST and sea-ice anomalies on the winter circulation in CCM3. Part II: Direct and indirect components of the response. *J. Clim*. **17,** 877–889 (2004).

67. Alexander, M. A., *et al.* The atmospheric response to realistic Arctic sea ice anomalies in an AGCM during winter. *J. Clim*. **17,** 890–905 (2004).68. Strey, S. T., Chapman, W. L. & Walsh, J. E. The 2007 sea ice minimum: Impacts on the Northern Hemisphere atmosphere in late autumn and early winter. *J. Geophys. Res*. **115,** D23103, doi:10.1029/ 2009JD013294 (2010).

69. Porter, D. F., Cassano, J. J. & Serreze, M. C. Local and large-scale atmospheric responses to reduced Arctic sea ice and ocean warming in the WRF model. *J. Geophys. Res.* **117,** D11115, doi:10.1029/2011JD016969 (2012).

70. Bluthgen, J., Gerdes, R. & Werner, M. Atmospheric response to the extreme Arctic sea ice conditions in 2007. *Geophys. Res. Lett.* **39,** L02707 doi:10.1029/2011GL050486 (2012).

71. Orsolini, Y., Senan, R., Benestad, R. & Melsom, A. Autumn atmospheric response to the 2007 low Arctic sea ice extent in coupled ocean–atmosphere hindcasts. *Clim. Dyn.* **38,** 2437–2448 (2012).

72. Singarayer, J. S., Valdes, P. J. & Bamber, J. L. The atmospheric impact of uncertainties in recent Arctic sea-ice reconstructions. *J. Clim.* **18,** 3996–4012 (2005).

73. Screen, J. A., Deser, C., Simmonds, I. & Tomas, R. Atmospheric impacts of Arctic sea-ice loss, 1979-2009: separating forced change from atmospheric internal variability. *Clim. Dyn.* **43,** 333–344 (2013).

74. Peings, Y. & Magnusdottir, G. Response of the wintertime Northern Hemispheric atmospheric circulation to current and projected Arctic sea ice decline: a numerical study with CAM5. *J. Clim*. **27,** 244–264 (2014).

75. Tanaka, H. L. & Seki, S. Development of a three-dimensional spectral linear baroclinic model and its application to the baroclinic instability associated with positive and negative Arctic Oscillation indices. *J. Meteorol. Soc. Japan* **91,** 193–213 (2013).

76. Francis, J. A. & Vavrus, S. J. Evidence linking Arctic amplification to extreme weather in mid-latitudes. *Geophys. Res. Lett.* **39,** L06801, doi:10.1029/2012GL051000 (2012).

77. Petoukhov, V., Rahmstorf, S., Petri, S. & Schellnhuber, H. J. Quasiresonant amplification of planetary waves and recent Northern Hemisphere weather extremes. *Proc. Natl. Acad. Sci.* **110,** 5336–5341 (2013).

78. Screen, J. A. & Simmonds, I. Amplified mid-latitude planetary waves favour particular regional weather extremes. *Nature Climate Change* doi:10.1038/nclimate2271 (2014).

79. Barnes, E. A. Revisiting the evidence linking Arctic amplification to extreme weather in midlatitudes. *Geophys. Res. Lett.* **40,** 1–6 (2013).

80. Allen, R. J. & Sherwood, S. C. Warming maximum in the tropical upper troposphere deduced from thermal winds. *Nature Geosci*. **1,** 399–403 doi: 10.1038/ngeo208 (2008).

81. Screen, J. A. & Simmonds, I. Exploring links between Arctic amplification and mid-latitude weather. *Geophys. Res. Lett.* **40,** 959–964 (2013).

82. Barnes, E. A., Dunn-Sigouin, E., Masato, G. & Woollings, T. Exploring recent trends in Northern Hemisphere blocking. *Geophys. Res. Lett.* **41,** doi: 10.1002/2013GL058745 (2014).

83. Kintisch, E. Into the maelstrom. *Science* **344,** 250–253 (2014).

84. Fletcher, C., Hardiman, S. C., Kushner, P. J. & Cohen, J. The dynamical response to snow cover perturbations in a large ensemble of atmospheric GCM integrations. *J. Clim*. **22,** 1208–1222 (2009).

85. Allen, R. J. & Zender, C. S. Forcing of the Arctic Oscillation by Eurasian snow cover. *J. Clim*. **24,** 6528–6539 (2011).

86. Cohen, J., J. C. Furtado, J. Jones, Barlow, M., Whittleston, D. & Entekhabi, D. Linking Siberian snow cover to precursors of stratospheric variability. *J. Clim.* **27,** 5422–5432 (2014).

87. Peings Y., Saint-Martin, D. & Douville, H. A numerical sensitivity study of the influence of Siberian snow on the northern annular mode. *J. Clim*. **25,** 592–607 (2012).

88. Petoukhov, V. & Semenov, V. A. A link between reduced Barents-Kara sea ice and cold winter extremes over northern continents. *J. Geophys. Res*. **115,** D21111, doi:10.1029/2009JD013568 (2010).

89. Inoue, J., Hori, M. E. & Takaya, K. The role of Barents Sea ice in the wintertime cyclone track and emergence of a warm-Arctic cold-Siberian anomaly. *J. Clim*. **25,** 2561–2568 (2012).

90. Jaiser, R., Dethloff, K., Handorf, D., Rinke, A. & Cohen, J. Planetary- and synoptic-scale feedbacks between tropospheric and sea ice cover changes in the Arctic. *Tellus* **64,** 11595, doi:10.3402/tellusa.v64i0.11595 (2012).

91. Jaiser, R., Dethloff, K. & Handorf, D. Stratospheric response to Arctic sea ice retreat and associated planetary wave propagation changes. *Tellus* **65,** 19375, doi:10.3402/tellusa.v65i0.19375 (2013).

92. Honda, M., Inue, J. & Yamane, S. Influence of low Arctic sea-ice minima on anomalously cold Eurasian winters. *Geophys. Res. Lett.* **36,** L08707, doi:10.1029/2008GL037079 (2009).

93. Hardiman, S. C., Kushner, P. J. & Cohen, J. Investigating the ability of general circulation models to capture the effects of Eurasian snow cover on winter climate. *J. Geophys. Res*. **113,** D21123, doi:10.1029/2008JD010623 (2008).

94. Martin, S. & Diffenbaugh, N. S. Transient twenty-first century changes in daily-scale temperature extremes in the United States. *Clim. Dyn*. **42,** 1383–1404 (2014).

95. Hoskins, B. J. & Karoly, D. J. The steady linear response of a spherical atmosphere to thermal and orographic forcing. *J. Atmos. Sci.* **38,** 1179–1196 (1981).

96. Garfinkel, C. I., Hartmann, D. L. & Sassi, F. Tropospheric precursors of anomalous Northern Hemisphere stratospheric polar cortices. *J. Clim.* **23,** 3282–3299 (2010).

97. Kolstad, E. W. & Charlton-Perez, A. J. Observed and simulated precursors of stratospheric polar vortex anomalies in the Northern Hemisphere. *Clim. Dyn.* **37,** 1443–1456 (2010).

98. Cohen, J. & Jones, J. Tropospheric precursors and stratospheric warmings. *J. Clim.* **24,** 6562–6572 (2011a).

99. Dee, D. P., et al. The ERA-Interim reanalysis: configuration and performance of the data assimilation system. *Q. J. R. Meteorol. Soc*. **137,** 553–597 (2011).

100. Hansen, J. E. & Lebedeff, S. Global trends of measured surface air temperature. *J. Geophys. Res*. **92,** 13345–13372 (1987).

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Acknowledgements

We are grateful to Elizabeth Barnes for many helpful discussions and suggested revisions to the manuscript. JC is supported by the National Science Foundation grants BCS-1060323 and AGS-1303647. JS is funded by Natural Environment Research Council grant NE/J019585/1. MB received support from National Science Foundation grant ARC-0909272 and NASA NNX13AN36G. JO receives support from the Arctic Research Project of the NOAA Climate Program Office and the Office of Naval Research, Code 322.

Author Contributions

JC proposed and was the main author of the manuscript. All co-authors contributed to the writing of the manuscript. JS created Fig. 1, JF Figs. 2 & 4, DC Fig. 3, JF & JC Fig. 4, MB and JC Fig. B1 and DW & JC Fig. B2.

Competing Financial Interests

The authors declare no competing financial interests.

Figure Legends

Fig. 1. Polar amplification of temperature trends, 1979–2014. Zonally-averaged temperature trends averaged around circles of latitude for a) winter (December–February), b) spring (March–May), c) summer (June–August) and d) autumn (September–November). Trends are based on ERA-Interim reanalysis data[99] from March 1979 to February 2014. The black contours indicate where trends differ significantly from zero at the 99% (solid lines) and 95% (dotted lines) confidence levels. The line graphs show trends (same units as in colour plots) averaged over the lower part of the atmosphere (950–1,000 hPa; solid lines) and over the entire atmospheric column (300–1,000 hPa; dotted lines)[9].

Fig. 2. During the recent period of Arctic amplification warming is occurring across the Northern Hemisphere but with cold mid-latitude winters. a) (right) Linear trend (°C per 10 years) in December – February (DJF) mean surface air temperatures from 1960/61-2013/14. Shading interval every 0.1°C per 10 years. Dark gray indicates points with insufficient samples to calculate a trend. (left) The zonally-averaged linear trend (°C per 10 years). b) Area-average surface temperature anomalies (°C) from 0°-60°N (solid black) and 60°-90°N (solid red) along with 5-year smoothing (dashed black/red lines). c) As in a) but from 1990/91-2013/14. Shading interval every 0.2°C per 10 years. Also note different scales between a) and c). Data from the National Aeronautics and Space Administration Goddard Institute for Space Studies (NASA GISS) temperature analysis (http://data.giss.nasa.gov/gistemp)[100].

Fig. 3. Temperature and Precipitation Extremes. Extreme indices in the mid-latitudes: Trend maps for the 1951-2013 period and timeseries averaged over the land-area from 20°N to 50°N. a)

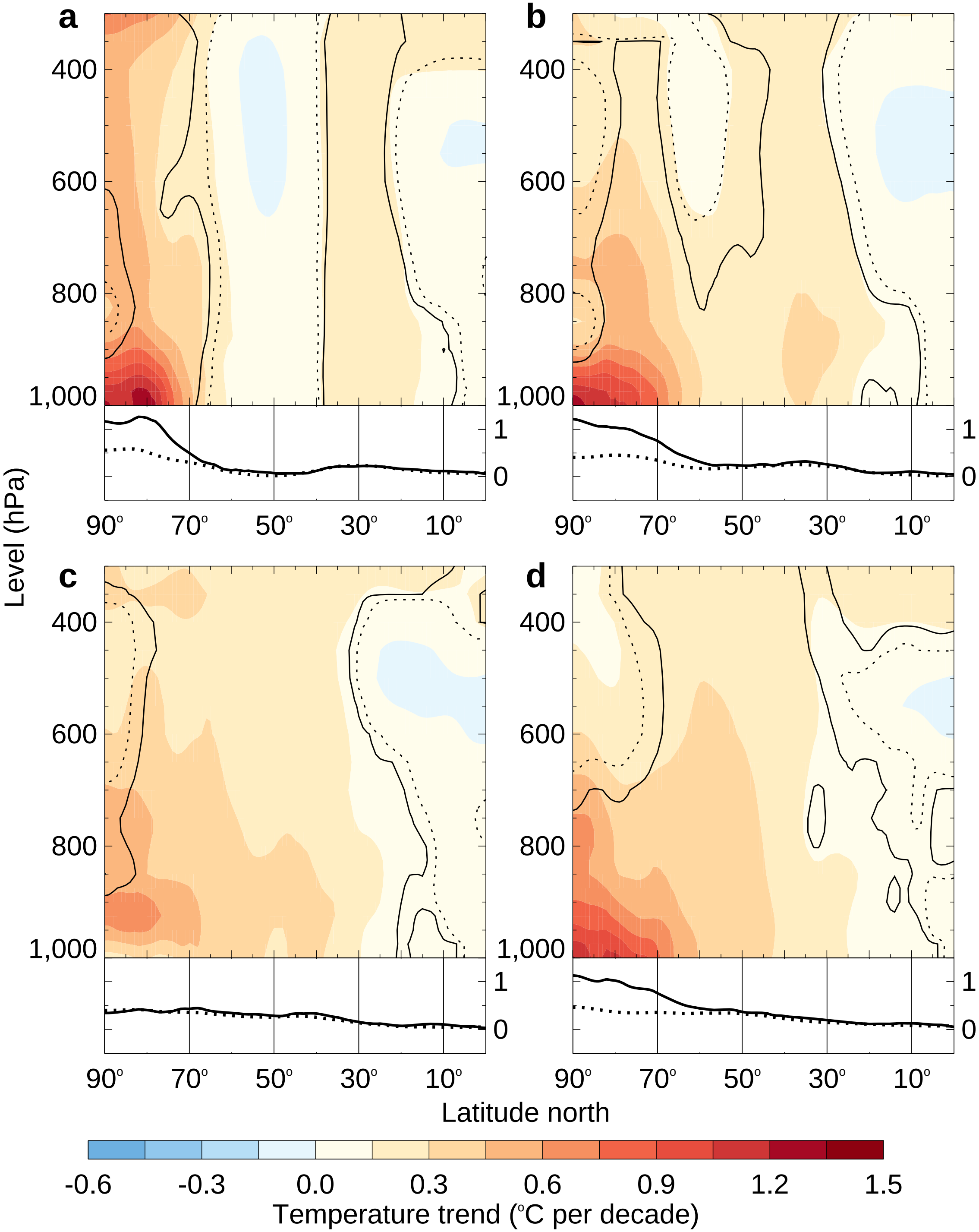
trend in annual total wet-day precipitation (PRCPTOT), b) annual very wet day precipitation (i.e. precipitation during days exceeding the 95th percentile, R95p), c) trend in annual very wet day precipitation (i.e. precipitation during days exceeding the 95th percentile, R95p), d) coldest daily minimum temperature (TNn), e) trend in annual warm days (i.e. percentage of days with temperatures exceeding the 90th percentile, TX90p), f) annual number of icing days (days with maximum temperature < 0^oC), g) percentage of land with summer months warmer than 1 standard deviation (solid) and 2 standard deviations (dashed) above the 1951–1980 mean, and h) percentage of land with winter months colder than 1 standard deviation (solid) and 2 standard deviations (dashed) below the 1951–1980 mean[30]. Stippling in the trend maps indicate significance at 95% confidence. The timeseries plot yearly values (thin grey curves) and the long-term non-linear trend (thick black curves). Panels a) to f) were created using the GHCNDEX global land gridded dataset of climate extremes[32] and definition of the extreme indices[32].

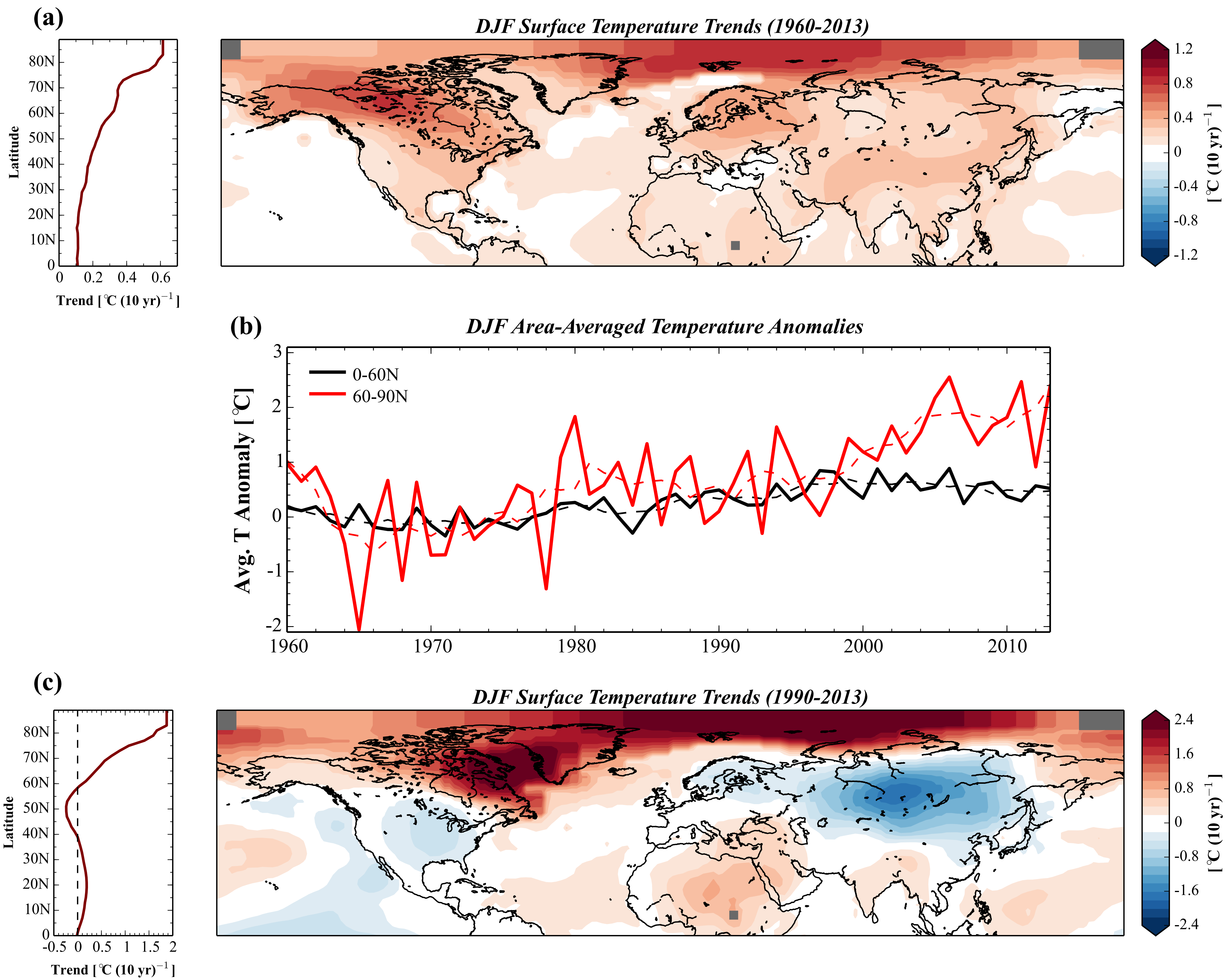
Fig. 4. Schematic of ways to influence NH mid-latitude weather. Three major dynamical features for changing Northern Hemisphere (NH) mid-latitude weather—changes in the storm tracks, the position and structure of the jet stream, and planetary wave activity—can be altered in several ways. The pathway on the left and highlighted by double boxes is reviewed in this manuscript. AA directly (by changing the meridional temperature gradient) and/or indirectly (through feedbacks with changes in the cryosphere) alters tropospheric wave activity and the jet stream in the middle and high latitudes. Two other causes of changes in the storm tracks, jet stream, and wave activity that do not involve AA are also presented: (1) Natural modes of variability and (2) the direct influence of global climate change (i.e., including influences outside

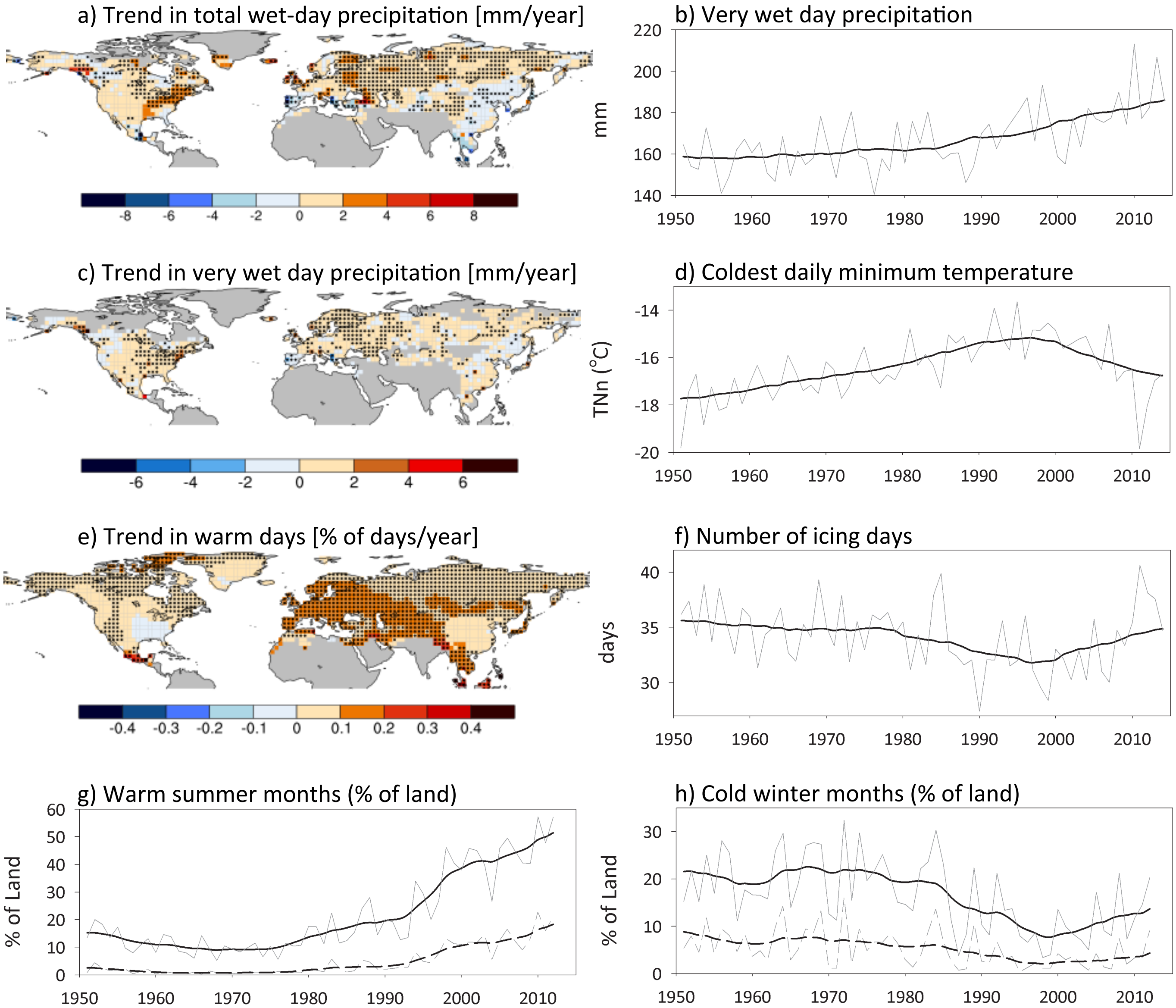
the Arctic) on the general circulation. The last two causes together present the current null hypothesis in the state of the science against which the influence of AA on mid-latitude weather is tested in both observational and modeling studies. Bi-directional arrows in the figure denote feedbacks (positive or negative) between adjacent elements. Stratospheric polar vortex is represented by “L” with counter-clockwise flow.

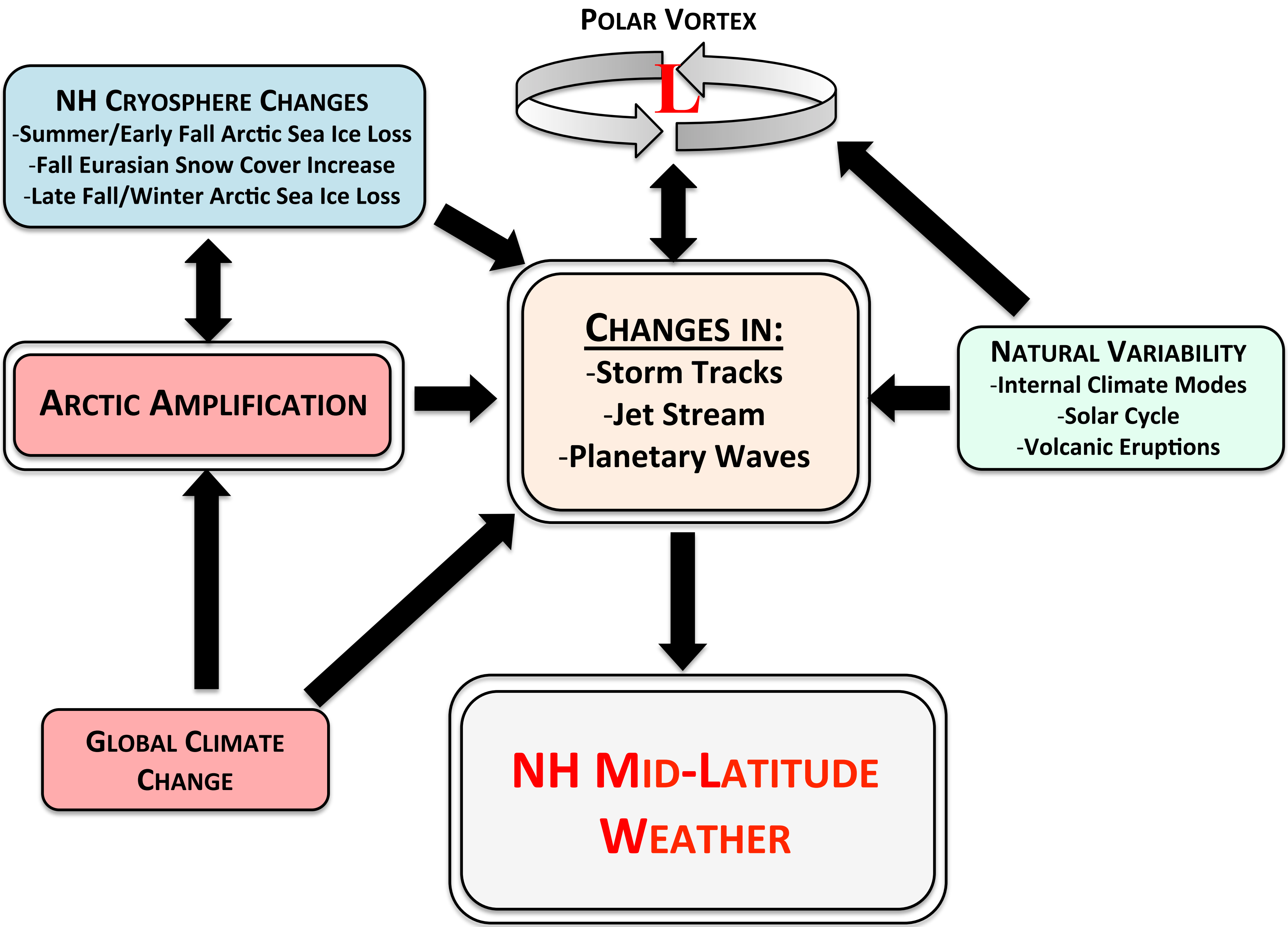
Fig. B1. Schematic view of jet-related and negative N/AO dynamics. a) Here, the tropospheric jet is divided into two parts, a “thermally-driven” part and an eddy-driven part. b) Changes in the atmospheric circulation associated with the negative phase of the N/AO.

Fig. B2. Synthesis of proposed cryospheric forcings. The schematic highlights a proposed way in which Arctic sea ice loss in late summer through early winter may work in concert with extensive Eurasian snow cover in the fall to force the negative phase of the N/AO in winter. Snow is shown in white, sea ice in white tinged with blue, sea ice melt with blue waves, high and low geopotential heights with red “H” (red represents anomalous warmth) and blue “L” (blue represents anomalous cold) respectively, tropospheric jet stream in light blue with arrows and stratospheric jet or polar vortex shown in purple with arrows. On the right globe cold (warm) temperature anomalies associated with the negative phase of the winter N/AO are shown in blue (orange).

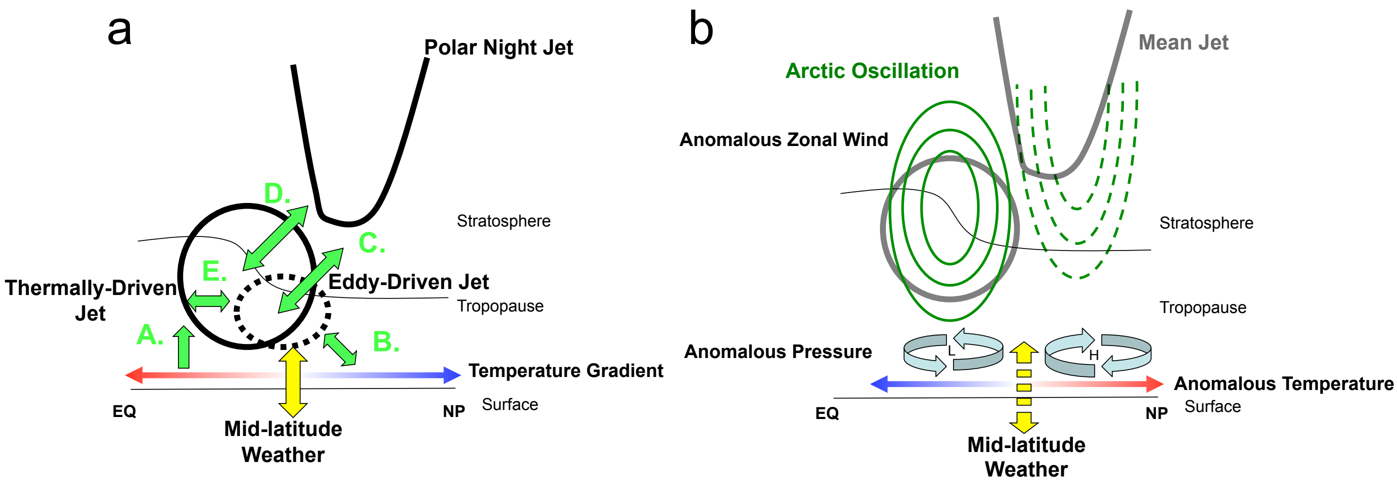

a
b
c
d
400
600
800
1,000
Level (hPa)
1
0
90°
70°
50°
30°
10°
Latitude north
-0.6
-0.3
0.0
0.3
0.6
0.9
1.2
1.5
Temperature trend (°C per decade)


(a)
DJF Surface Temperature Trends (1960-2013)
Latitude
80N
70N
60N
50N
40N
30N
20N
10N
0
0.2
0.4
0.6
Trend [℃ (10 yr)$^{-1}$]
1.2
0.8
0.4
0
-0.4
-0.8
-1.2
[℃ (10 yr)$^{-1}$]
(b)
DJF Area-Averaged Temperature Anomalies
0-60N
60-90N
Avg. T Anomaly [℃]
3
2
1
0
-1
-2
1960
1970
1980
1990
2000
2010
(c)
DJF Surface Temperature Trends (1990-2013)
-0.5
0
0.5
1
1.5
2
2.4
1.6
0.8
0
-0.8
-1.6
-2.4


a) Trend in total wet-day precipitation [mm/year]
-8 -6 -4 -2 0 2 4 6 8
b) Very wet day precipitation
mm
220 200 180 160 140
1950 1960 1970 1980 1990 2000 2010
c) Trend in very wet day precipitation [mm/year]
-6 -4 -2 0 2 4 6
d) Coldest daily minimum temperature
TNn (°C)
-14 -16 -18 -20
1950 1960 1970 1980 1990 2000 2010
e) Trend in warm days [% of days/year]
-0.4 -0.3 -0.2 -0.1 0 0.1 0.2 0.3 0.4
f) Number of icing days
days
40 35 30
1950 1960 1970 1980 1990 2000 2010
g) Warm summer months (% of land)
% of Land
60 50 40 30 20 10 0
1950 1960 1970 1980 1990 2000 2010
h) Cold winter months (% of land)
% of Land
30 20 10 0
1950 1960 1970 1980 1990 2000 2010


POLAR VORTEX
L
NH CRYOSPHERE CHANGES
-Summer/Early Fall Arctic Sea Ice Loss
-Fall Eurasian Snow Cover Increase
-Late Fall/Winter Arctic Sea Ice Loss
ARCTIC AMPLIFICATION
CHANGES IN:
-Storm Tracks
-Jet Stream
-Planetary Waves
NATURAL VARIABILITY
-Internal Climate Modes
-Solar Cycle
-Volcanic Eruptions
GLOBAL CLIMATE CHANGE
NH MID-LATITUDE WEATHER

Jet-Related Dynamics




a
Troposphere
H
L
Moisture Flux
b
Stratosphere
Troposphere
Vertical Wave Activity Flux
H
L
Constructive Interference
c
Stratosphere
Troposphere
L
H
H
L
L
Sept
Oct
Nov
Dec
Jan
Feb