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Physics and Chemistry of the Arctic Atmosphere

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Physics and Chemistry of the Arctic Atmosphere

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This book is dedicated to the memory of
Prof. Wolfgang von Hoyningen-Huene
(1947–2019)

Preface

Our planet experiences significant climatic change due to increasing anthropogenic influences and also due to natural variations. The changes have accelerated in recent decades, especially in the polar regions, where substantial warming of both the atmosphere and the underlying surface has been reported. This leads to impacts on terrestrial and freshwater species, communities, and ecosystems. Climate change in the Arctic will continue, with implications for biological resources and globally important feedbacks to climate.

With this in mind, a survey of modern research related to properties of the Arctic atmosphere has been conducted and is presented in this book. The book consists of eleven chapters that survey the dynamics and thermodynamics of the Arctic atmosphere, Arctic aerosols, fog, clouds (tropospheric, stratospheric, and mesospheric), radiation, and trace gases. In addition, recent results from remote sensing of Arctic atmospheric aerosols are outlined, and observed and projected changes in the Arctic climate are described. We believe that this book, aimed at a better understanding of various processes and trends characteristic of the Arctic atmosphere, will be useful both for scientists dealing with various aspects of Arctic atmospheric research and for graduate and undergraduate students.

August 30, 2019

Contents

1 **Dynamical Processes in the Arctic Atmosphere** 1
Marius O. Jonassen, Dmitry Chechin, Alexey Karpechko, Christof Lüpkes, Thomas Spengler, Annick Tepstra, Timo Vihma, and Xiangdong Zhang

2 **Thermodynamics of the Arctic Atmosphere** 53
Claudio Tomasi, Boyan H. Petkov, Oxana Drofa, and Mauro Mazzola

3 **Trace Gases in the Arctic Atmosphere** 153
Kimberly Strong, William R. Simpson, Kristof Bognar, Rodica Lindenmaier, and Sébastien Roche

4 **Arctic Aerosols** 209
Roberto Udisti, Rita Traversi, Silvia Becagli, Claudio Tomasi, Mauro Mazzola, Angelo Lupi, and Patricia K. Quinn

5 **A Climatological Overview of Arctic Clouds** 331
Abhay Devasthale, Joseph Sedlar, Michael Tjernström, and Alexander Kokhanovsky

6 **Arctic Ice Fog: Its Microphysics and Prediction** 361
Ismail Gultepe, Andrew J. Heymsfield, and Martin Gallagher

7 **Polar Stratospheric Clouds in the Arctic** 415
Francesco Cairo and Tiziana Colavitto

8 **Noctilucent Clouds: General Properties and Remote Sensing** 469
Christian von Savigny, Gerd Baumgarten, and Franz-Josef Lübken

9 Remote Sensing of Arctic Atmospheric Aerosols 505
Alexander Kokhanovsky, Claudio Tomasi, Alexander Smirnov, Andreas Herber, Roland Neuber, André Ehrlich, Angelo Lupi, Boyan H. Petkov, Mauro Mazzola, Christoph Ritter, Carlos Toledano, Thomas Carlund, Vito Vitale, Brent Holben, Tymon Zielinski, Simon Bélanger, Pierre Larouche, Stefan Kinne, Vladimir Radionov, Manfred Wendisch, Jason L. Tackett, and David M. Winker

10 Radiation in the Arctic Atmosphere and Atmosphere – Cryosphere Feedbacks 591
Claudio Tomasi, Boyan H. Petkov, Angelo Lupi, Mauro Mazzola, Christian Lanconelli, and Ismail Gultepe

11 Climate Change in the Arctic 673
Torben Koenigk, Jeff Key, and Timo Vihma

Index 707

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Chapter 1
Dynamical Processes in the Arctic Atmosphere

Marius O. Jonassen, Dmitry Chechin, Alexey Karpechko, Christof Lüpkes, Thomas Spengler, Annick Tepstra, Timo Vihma, and Xiangdong Zhang

Abstract The scales of dynamical processes in the Arctic atmosphere range from turbulence in the atmospheric boundary layer (ABL) via interactive mesoscale processes, such as orographic flows and Polar lows, to synoptic-scale cyclones, and further to hemispherical-scale circulations characterized by the Polar front jet stream and planetary waves. Specific boundary conditions for tropospheric dynamics in the Arctic include (a) sea ice and snow, which strongly affect the surface energy budget, (b) large transports of heat and moisture from lower-latitudes, and (c) the wintertime stratospheric Polar vortex, which has a large impact on tropospheric large-scale circulation and synoptic-scale cyclones. Knowledge on dynamics of the Arctic atmosphere is advancing but, compared to mid- and low-latitudes, still limited due to lack of process-level observations from the Arctic. The dynamics of the Arctic atmosphere poses a challenge for numerical weather prediction (NWP) and climate models, in particular in the case of ABL, orographic flows, Polar lows, and troposphere-stratosphere coupling. More research is also needed to

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better understand how the atmospheric dynamics affects and is affected by climate warming.

Keywords Atmospheric boundary layer · Cold-air outbreak · Downslope wind · Orographic flow · Planetary waves · Polar low · Polar front jet stream · Polar vortex · Synoptic-scale cyclones · Troposphere-stratosphere interactions

1.1 Introduction

Dynamical processes in the Arctic atmosphere cover a broad range of spatial and temporal scales. The hemispherical-scale circulation is characterized by the stratospheric Polar vortex, the tropospheric Polar front jet stream, the Arctic Oscillation, and planetary waves (Sect. 1.2). The synoptic-scale dynamics are dominated by transient cyclones (Sect. 1.3) and anticyclones, including high-pressure blockings. The synoptic-scale flow often results in marine cold-air outbreaks, which in winter are favourable conditions for generation of Polar lows – the most vigorous mesoscale dynamical features in the Arctic (Sect. 1.4). Various orographic effects on the flow include patterns in synoptic spatial scale, such as the Greenland and Ural Blockings, as well as meso- and local-scale features, such as katabatic and other downslope winds, barrier winds, tip jets, gap flows, and downslope wind storms (Sect. 1.5). In all spatial scales, the flow features occurring in the lower troposphere are affected by atmospheric boundary-layer (ABL) dynamics, which control the exchange of momentum, heat, and moisture between the atmosphere and Earth surface (Sect. 1.6). This couples the Arctic atmosphere, both dynamically and thermodynamically, with the open ocean, sea ice, land ice, and terrestrial surfaces.

The broad range of spatial scales of Arctic atmospheric dynamics is associated with temporal variability in scales of seconds (ABL turbulence), hours (some orographic flow features), days (transient cyclones), weeks (jet stream patterns, planetary waves, high-pressure blockings), seasonal (troposphere – stratosphere interactions), and inter-annual/decadal (large-scale coupling between the atmosphere and ocean). In addition to the natural variability, the Arctic climate system is rapidly changing. A major challenge is to better understand how the atmospheric dynamics will respond to the Arctic Amplification of climate warming.

Our knowledge on dynamical processes in the Arctic atmosphere is based on theoretical understanding of atmospheric dynamics, observations, and model experiments. Theories on atmospheric dynamics have mostly been developed on the basis of observations from mid and low latitudes, and the applicability of these theories for the Arctic environment has been evaluated on the basis of limited observations from the Arctic. The regular network of in situ observations is limited to coastal and archipelago stations, most of them only equipped for near-surface observations. In addition there are stations for regular radiosonde soundings; a total of approximately 36 stations are in operation north of 65 °N, but there are no regular soundings north of 80 °N. These regular observations are complemented by field expeditions based on research vessels, drifting ice stations or aircraft. Research vessel and aircraft expeditions are mostly carried out during summer and spring, allowing process-level observations among others on dynamics of the atmospheric boundary layer and its interaction with cloud physics, radiative transfer, and the ocean and sea ice surface. Year-round drifting ice stations have been operated in the Arctic Ocean regularly from 1950 to 1991 and from 2003 to 2012, mostly by Russian colleagues. In addition, data relevant for atmospheric dynamics, such as surface pressure and near-surface air temperature, are collected by automatic stations deployed on sea ice, above all by the International Arctic Buoy Program since 1979. The data collected at regular stations and field expeditions are mostly point observations, except of research aircraft observations. Even point observations are useful for understanding ABL dynamics but not enough from the point of view of meso- and synoptic-scale dynamics. Satellite data are typically better than in situ observations with respect to spatial and temporal resolution, but usually worse with respect to accuracy and vertical resolution. Satellite observations are also hampered by the Polar night and high latitudes (geostationary satellite data not available).

Dynamics of the Arctic atmosphere poses a challenge for numerical weather prediction (NWP) and climate models, in particular in the case of ABL (Sect. 1.6), mesoscale processes (Sects. 1.4 and 1.5), and troposphere-stratosphere coupling (Sect. 1.2). Operational forecasting of Polar lows is a challenge due to their fast development and small spatial scale, and in climate models Polar lows are often not resolved at all, although they sometimes have effects that are also important for climate, e.g., when generating deep convection in the ocean (Sect. 1.4). In high latitudes, NWP and climate models as well as atmospheric reanalyses typically have their largest temperature, humidity and wind errors in the ABL (Sect. 1.6). These errors are critical, as improved modelling capabilities are needed to better understand the recent rapid Arctic amplification of climate warming and to reliably project the future. In addition, increasing navigation, aviation, and offshore activities in the Arctic calls for more accurate weather forecasts for the Arctic. To improve modelling of the Arctic troposphere-stratosphere coupling, ABL and mesoscale processes, we need improved understanding on the relevant dynamical processes.

1.2 Large-Scale Circulation in the Troposphere and Stratosphere

1.2.1 Stratospheric Circulation

The stratosphere is a region in the atmosphere extending approximately from 10 km to 50 km above the surface and mostly known for being home to Earth's ozone layer. The stratosphere is also characterised by a distinct circulation which is capable of affecting weather and climate down to the surface. In this sub-chapter we first describe general characteristics of the Arctic stratosphere and factors affecting its variability and then describe the variability and recent changes in the stratosphere.

1.2.1.1 General Characteristics of the Arctic Stratosphere and Its Circulation

In the stratosphere air temperature increases, or remains almost constant with altitude. The temperature profile is mainly controlled by radiative balance between heating due to absorption of solar ultraviolet (UV) radiation by ozone and cooling due to radiative emission mostly by carbon dioxide. Such thermal stratification means that the stratosphere is very stable, i.e. vertical motion is restricted. As a result stratospheric dynamics differs considerably from that of the troposphere where convection plays significant role in the dynamics.

Outside of the equatorial belt the stratospheric characteristics experience pronounced seasonal cycle. In summer, solar irradiance peaks over the North Pole and the Arctic stratosphere warms with respect to lower latitudes. This leads to a formation of a pole-centred summer anticyclone (Fig. 1.1b). During summer, the Arctic stratosphere remains very close to radiative equilibrium with little interannual and interdecadal variability.

During autumn the polar air starts to cool faster than that in mid-latitudes and the summertime anticyclone is replaced by a wintertime pole-centred cyclone, called polar vortex (Fig. 1.1a). The edge of the polar vortex is surrounded by strong westerly winds, called polar night jet, with typical wind speed in the upper stratosphere (10–1 hPa) of 30–40 m/s and occasionally up to 60–70 m/s during mid-winter. Unlike in summer, stratospheric winds and temperatures in winter experience large variability at synoptic, intraseasonal and interannual time scales. The variability is linked to interaction between the zonal mean circulation and planetary waves, and is driven by variability in the strength of planetary wave sources as well as variability in the transmitting properties of zonal mean stratospheric circulation, as discussed in more detail below.

The difference between summer and winter circulation arises from the fact that planetary-scale stationary waves generated in the lower atmosphere by orography and ocean-land contrasts can only propagate their energy upward when background zonal winds are westerly (Charney and Drazin 1961). Thus, as zonal wind direction changes from summer easterlies to winter westerlies the stratosphere becomes supportive of planetary waves propagating from below. Charney and Drazin (1961) showed that the wave propagation is strongly controlled by zonal mean zonal winds: upward wave propagation is only possible when zonal winds are positive (i.e. westerly) and weaker than a critical wind speed, which is inversely proportional to zonal wave number. Under typical wind speed of the polar night jet, only longest waves with zonal wave numbers one and two can propagate to the stratosphere.

Propagation of the planetary waves in the atmosphere is best described by the Eliassen-Palm flux (Edmon et al. 1980) which is, under realistic approximations, proportional to wave's group velocity multiplied by a scalar called wave activity,

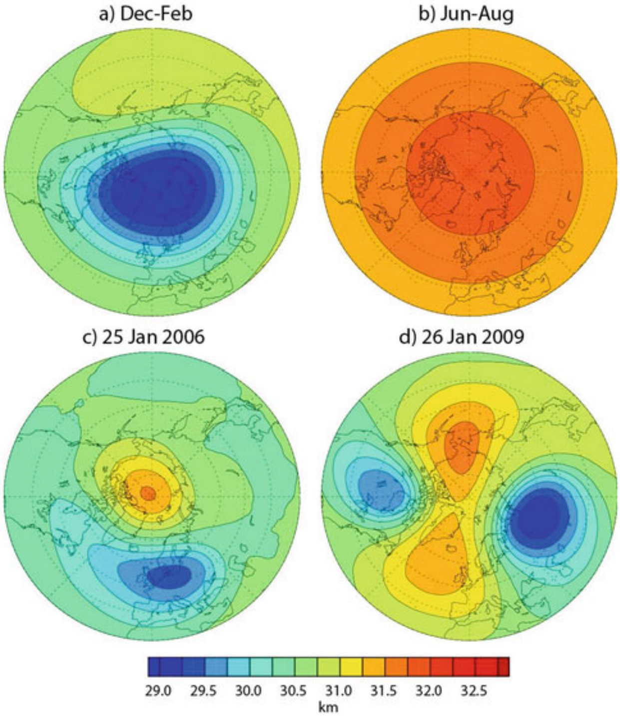

Fig. 1.1 Geopotential height at 10 hPa with four panels: (**a**) climatological winter mean, (**b**) climatological summer mean, (**c**) example of polar vortex displacement during an SSW; (**d**) example of polar vortex split during an SSW

which is quadratic in wave's amplitude. Thus upward wave propagation is associated with transfer of wave activity to the stratosphere. Wave activity can either be dissipated in the stratosphere or reflected back to the troposphere. The dissipation characterised by a convergence of the EP flux is the most typical situation. The dissipation is often associated with planetary wave breaking, which happens when wave approaches zero wind line, acting as a critical line for the stationary waves with zero phase speed. While details of wave breaking and dissipation are not well understood (Andrews et al. 1987), their effect on the zonal mean circulation is a deceleration of westerly winds. The deceleration destroys the thermal wind balance. In order to restore the balance, the vertical motion is induced in both polar and low latitudes. The descending motion in the polar region warms the polar atmosphere through adiabatic compression, while the ascending motion in the low latitudes cools the atmosphere through adiabatic expansion. The adiabatic heating results in a reduced meridional temperature gradient, which again satisfies the thermal wind balance. The fact that stratospheric temperatures are affected by dynamical heating also means that the winter stratosphere is not in a radiative equilibrium. Comparison of observed temperatures with those calculated under the assumption of no dynamical heating reveals that the dynamics heats the Arctic stratosphere by 30 K (lower stratosphere) to 100 K (upper stratosphere) in mid-winter.

While upward wave propagation and deceleration of the westerly winds is a permanent process, the intensity of the wave forcing varies in time because it depends on strength of wave sources and transmitting properties of the atmosphere. As a result, there is strong variability in the wintertime stratosphere. In an extreme case, wave propagation and dissipation in the stratosphere can lead to a strong warming of the polar atmosphere in a process called Sudden Stratospheric Warming (SSW) when the meridional temperature gradient between mid-latitudes and the pole reverses its sign. During such episodes westerly winds may reverse to easterly and the polar vortex becomes strongly disturbed and relocated. Typically, the vortex either becomes displaced from the pole or split into two smaller vortices (Fig. 1.1c, d). If such an event happens during mid-winter, an SSW is called a major SSW. It may take from a few days to a few weeks before a combination of dynamical processes and radiative relaxation towards radiative equilibrium restores the polar vortex. When such event occurs near the end of winter, or during spring, the polar vortex may not restore at all, and a transition towards the summer type of the circulation occurs. In this case, which is a typical ending of stratospheric winter, the SSW is called the final SSW.

The opposite extreme case can arise when wave propagation from the troposphere weakens. Under such conditions radiative cooling dominates thermal balance, polar vortex strengthens, and lower stratospheric polar temperatures can decrease down to 185–190 K. When such low temperatures are reached, water vapour and nitrous acid present in small amount in the stratospheric air start to freeze forming Polar Stratospheric Clouds (PSCs). PSCs play a key role in chemical ozone depletion (see Chap. 8).

Zonal wind deceleration and warming of the Arctic stratosphere are not the only effects of planetary wave dissipation in the stratosphere. Vertical motion generated to maintain thermal wind balance requires existence of meridional circulation to ensure mass conservation, with poleward motion in the stratosphere and returning flow beneath. This overturning meridional circulation is slow, with typical velocities of several centimetres per second; however it plays an important role in transport of chemicals. In particular, atmospheric ozone mostly produced in the tropical atmosphere is transported to extratropics, including the Arctic, by this circulation. The meridional circulation responsible for the transport of chemicals in the stratosphere is known as Brewer-Dobson circulation. Two branches of the Brewer-Dobson circulation are usually distinguished: a deep branch affecting the whole depth of the stratosphere, which is driven by the dissipation of the planetary waves propagating upward along the polar night jet, and a shallow branch, which extends up to ~ the 50 hPa level and is driven by planetary and synoptic waves propagating to the lower stratosphere in sub-tropics along the sub-tropical jet stream. While both branches contribute to the ascending motion in the tropics, it is the deep branch that is associated with the polar downwelling and adiabatic warming of the Arctic stratosphere.

1.2.1.2 Arctic Stratosphere Variability and Change

There is strong variability in the characteristics of the Arctic polar vortex on all timescales, from synoptic to decadal. This variability is mostly driven by the interaction between planetary waves and zonal mean flow. Sources of the variability are associated with factors affecting generation of the waves as well as modulating propagation of the waves within the stratosphere. These factors are considered below.

Blocking in the tropospheric circulation is a remarkable example of amplification of wave amplitude at synoptic scale. There is strong statistical connection between occurrence of blocking and SSWs (Martius et al. 2009). Although blocking is associated with synoptic-scale waves, which are not able to propagate to the stratosphere, blocking can enhance the amplitude of planetary waves through constructive interference (Garfinkel et al. 2010) and thus affect planetary wave flux to the stratosphere.

The Madden-Julian Oscillation (MJO) in the tropical atmosphere-ocean system can modulate the stratospheric variability at monthly scales. There is a significant connection between phases of MJO and state of the Arctic stratospheric vortex, which is attributed to MJO's ability to generate planetary waves that propagate poleward and consequently upward to the stratosphere (Garfinkel et al. 2012a). For example, enhanced convection over the western Pacific associated with MJO triggers enhanced planetary wave flux to the stratosphere, which is followed by a weakening of the polar vortex approximately 10 days later.

Another tropical phenomenon, El Niño, which is known to affect weather and climate globally, also affects the Arctic stratosphere. During El Niño winters, there is increased wave activity flux due to wavenumber one to the stratosphere. Composite analysis shows that weakening of stratospheric winds and warming of the Arctic stratosphere is observed during January–February of El Niño winters (Cagnazzo and Manzini 2009). El Niño variability also modulates frequency of SSWs, but in a somewhat unexpected way. The frequency of SSWs is enhanced during both El Niño and La Niña winters, while SSWs are rather rare during neutral winters (Butler et al. 2011).

Anomalies in the sea ice and terrestrial snow can also affect generation of planetary waves and thus stratospheric circulation. It has been reported that large-scale cooling associated with enhanced snow cover over Eurasia during boreal autumn is followed by an enhanced upward propagation of planetary waves and weakening of the polar vortex (Cohen et al. 2014). A similar connection was found between polar vortex weakening and sea ice loss over Barents and Kara sea region (Kim et al. 2014). The later effect is of interest because the rapid sea ice decline observed since the last decade of the twentieth century can potentially lead to long-term changes in the Arctic stratospheric circulation.

The Quasi-Biennial Oscillation (QBO) in the stratospheric equatorial winds is known to affect the strength of the polar vortex at inter-annual scale (Baldwin et al. 2001). During the easterly phase of QBO, when zonal winds in the equatorial lower stratosphere at about 30–50 hPa blow west, the polar vortex is weaker than during the westerly QBO phase. It has been hypothesized (Holton and Tan 1982) that the shift of the zero wind line to the Northern Hemisphere sub-tropical stratosphere during the easterly phase makes planetary waves to break further north and thus decelerate the polar night jet more effectively than during the westerly phase, when the zero wind line moves to the southern hemisphere allowing planetary waves to propagate deeper to the tropics and less affect the polar vortex. It has also been suggested (Garfinkel et al. 2012b) that other factors affecting planetary wave propagation can play a more important role than the zero wind line shift in modulating planetary wave propagation by QBO.

The main source of variability on the decadal scale is the 11-year solar cycle. The modulation of the polar vortex by the solar cycle is linked to differential absorption of solar radiation between the tropical and high latitudes. Larger solar UV radiation during years of solar maximum leads to a stronger meridional temperature gradient and consequently, via thermal wind balance, to a stronger polar vortex (Gray et al. 2010). Weaker UV radiation during years of solar minimum results in an opposite situation. The other potential sources of the decadal Arctic stratosphere variability are the long modes of ocean circulation variability, such as Atlantic Multidecadal Oscillation (Omrani et al. 2014) and Pacific Decadal Oscillation (Woo et al. 2015); however, the exact mechanisms of influence and importance of these factors remain poorly understood.

In addition to the regular and quasi-regular factors outlined above, explosive eruptions of tropical volcanoes can also affect the atmospheric circulation (Timmreck 2012). The aerosol formed in the tropical stratosphere from erupted gases absorbs the terrestrial infrared radiation and thus warms the lower tropical stratosphere. The absorption increases the meridional temperature gradient and strengthens the polar vortex, i.e. the mechanism is similar to that occurring during solar maximum.

Like the rest of the climate system, the Arctic stratosphere is expected to change in response to ongoing anthropogenic emission of greenhouse gases and associated global warming. Increased concentration of CO_2 leads to increased longwave radiation emission to space, which is expected to cool the stratosphere globally. However, the response of the Arctic lower stratosphere, whose temperatures are strongly controlled by the dynamics, will strongly depend on the response of the atmospheric circulation. Climate models consistently predict strengthening of the Brewer-Dobson circulation under global warming scenarios. While stronger Brewer-Dobson circulation is often associated with a weaker polar vortex and warmer Arctic stratosphere, the response will depend on whether or not the anomalous downwelling associated with strengthened Brewer-Dobson circulation will extend to the Arctic stratosphere. Since theory predicting the response of atmospheric circulation to global warming is currently lacking, there is a large uncertainty in the future of the Arctic stratosphere with climate models simulating both strengthening and weakening of the polar vortex as a response to climate change (Manzini et al. 2014).

1.2.2 Stratosphere-Troposphere Coupling

For a long time it was assumed that stratospheric dynamical variability is driven by tropospheric variability, while the stratosphere has little influence on the processes below. Such a view is intuitively understandable since the mass of the stratosphere is much smaller, only about 15% of that of the troposphere. Furthermore there are no obvious sources of variability in the stratosphere, such as dynamical instabilities, which are present in the troposphere. However, it gradually became widely accepted that the stratosphere plays an active role in the stratosphere-troposphere dynamical coupling. Experiments with simple atmospheric models revealed that the stratosphere behaves like a chaotic dissipative dynamical system, and may develop its own variability even if tropospheric wave forcing is constant (Holton and Mass 1976; Yoden 1987). Moreover, it was demonstrated that the atmospheric adjustment to a mechanical force applied within the stratosphere (which in real world can result, for example, from convergence of planetary or gravity wave activity) involves a meridional circulation extending downward to the boundary layer, the effect known as a downward control principle (Haynes et al. 1991). Research during the last decade of the twentieth century has confirmed that large stratospheric anomalies, such as SSWs or polar vortex strengthening usually propagate downward in time and sometimes affect tropospheric circulation (e.g. Baldwin and Dunkerton 1999). For example, an SSW and associated weakening of the stratospheric zonal winds can induce a southward shift of the tropospheric jet and lead to cold anomalies over northern Eurasia and eastern North America. Composites of temperature and precipitation anomalies following downward propagation of geopotential height anomalies are illustrated in Fig. 1.2. Although the downward control principle predicts weakening of the tropospheric zonal winds following SSWs, it does not explain all important details of the observed response, such as the shift of the tropospheric jet position. A feedback from the tropospheric synoptic-scale eddies is understood to play an important role, but the details of this feedback remain a subject of ongoing research.

The other mechanism of downward effects is the reflection of planetary waves from the stratosphere. The reflection does not affect the zonal mean circulation in the way described above. Rather, observations suggest that it can amplify tropospheric stationary waves, sometimes resulting in blocking.

Presently, there is a strong interest in stratosphere-troposphere coupling associated with the fact that weather patterns following the stratospheric anomalies are remarkably persistent, and therefore potentially predictable, at scales up to several weeks, i.e., beyond the reach of state-of-art numerical weather prediction models (Sigmond et al. 2013; Scaife et al. 2016). The opportunity to improve climate predictability at sub-seasonal to seasonal scales attracts considerable research efforts, which means that our understanding of the stratosphere-troposphere coupling will likely advance in the years to come.

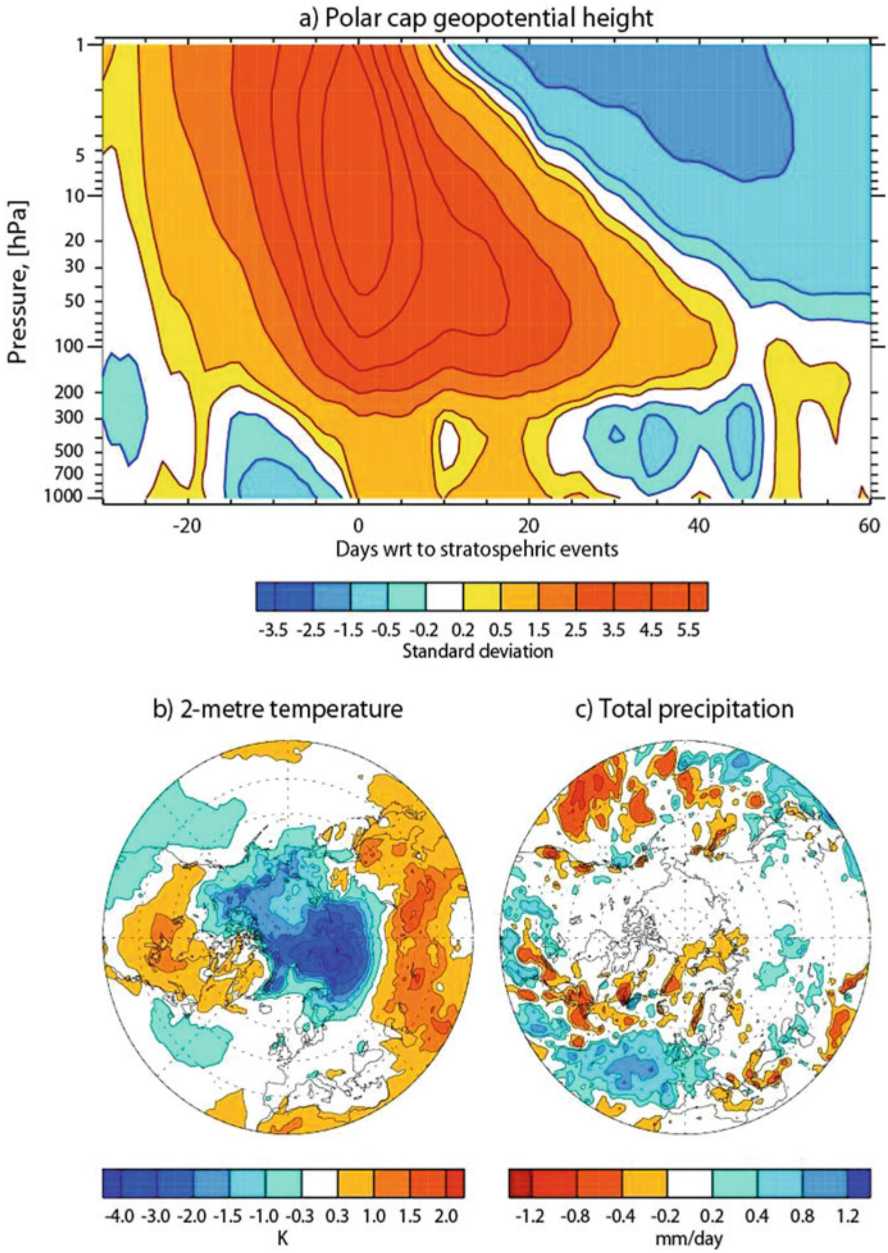

Fig. 1.2 (**a**) Composite of downward propagation of polar cap geopotential height anomalies; (**b**) composite of temperature anomalies during days 11–60 following stratospheric events; (**c**) composite of precipitation anomalies during days 11–60 following stratospheric events

1.2.3 Tropospheric Large-Scale Circulation

The state and variations of the Arctic troposphere strongly depend on the large-scale circulation in high- and mid-latitudes of the Northern Hemisphere. Important features of the circulation include the Polar front jet stream in the mid- and upper-troposphere, the Arctic Oscillation, planetary waves, and high-pressure blockings. These features interact with synoptic-scale cyclones (Sect. 1.3) and affect the large-scale boundary conditions for the occurrence of Polar lows (Sect. 1.4).

The Polar front jet stream guides the cyclone tracks due to the strong wind shear below the jet core. Particularly important properties of the jet stream are its latitude, speed, meandering, and the occurrence of events when the jet is split in two branches (Hall et al. 2015; Vavrus 2018). The split jet occurs typically in summer and the events are often associated with quasi-resonance of planetary waves with thermal and orographic forcing (Coumou et al. 2017). Hence, the split jet favours persistency of weather patterns, often resulting in floods, droughts, and heat waves in mid-latitudes. Arctic effects on the jet stream properties have recently received a lot of attention (e.g., Francis and Vavrus 2015), but also the jet stream has a strong influence on weather conditions in the Arctic. A meandering jet stream allows advection of warm, moist air masses from mid-latitudes to the Arctic, often resulting in extreme anomalies in weather and sea ice conditions (Vihma 2017; Rinke et al. 2017).

The Arctic Oscillation Index (AOI) characterizes the strength of the circumpolar zonal circulation. The AOI is calculated on the basis of the loading pattern of the first principal component of the mean-sea-level pressure (MSLP) or 1000 hPa geopotential height in the Northern Hemisphere north of 20 °N. The North Atlantic Oscillation (NAO) represents the role of the Atlantic sector on the circumpolar Arctic Oscillation. The NAO index (NAOI) has several definitions, but is always associated with a north–south- dipole structure in the MSLP (or 1000 hPa geopotential) field over the Atlantic, approximately between the Icelandic low and the Azores high (Ambaum et al. 2001). Both AOI and NAOI vary irregularly in time. Large positive (negative) values of AOI are related to strong (weak) westerly winds, and high (low) winter temperatures in mid-latitudes, but low (high) winter temperatures in the Arctic (Fig. 1.3). The effects of NAO in the Atlantic sector and Europe are qualitatively similar to those of AO. NAO has is often regarded as a cause of weather variations, but can also be seen as a consequence of variations in the jet stream and storm tracks (Vallis and Gerber 2008). NAOI is indeed significantly correlated with the jet stream latitude, a negative NAOI being associated with southward displacement of the jet stream (Woollings and Blackburn 2012). Several studies have suggested that the recent Arctic Amplification of climate warming has favoured a negative NAO (Kim et al. 2014; Nakamura et al. 2015; Francis and Skific 2015; Deser et al. 2016), but different conclusions have also been presented (Singarayer et al. 2006; Screen et al. 2014; Smith et al. 2017).

However, AO and NAO alone do not characterize all important aspects of large-scale circulation in northern mid- and high-latitudes. Several indices have been used

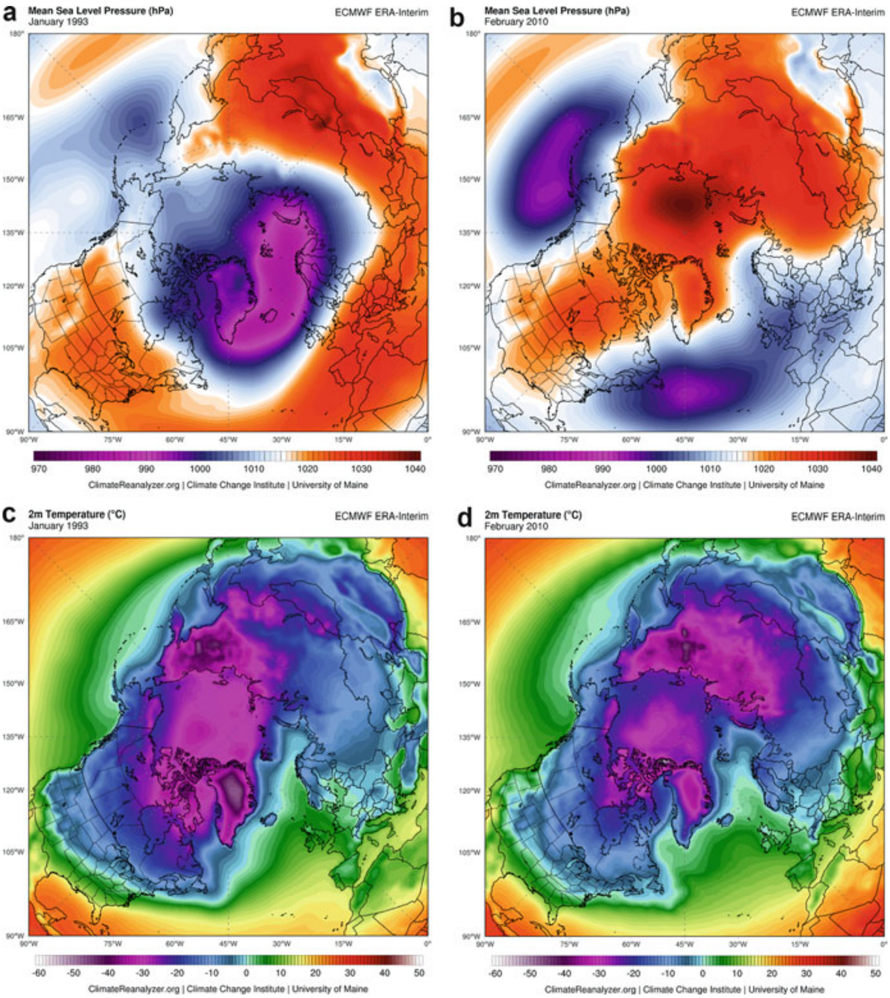

Fig. 1.3 Monthly mean fields of mean-sea-level pressure (in hPa, **a**, **b**) and 2-m air temperature (in °C, **c**, **d**) in January 1993 (left panels), when the Arctic Oscillation Index was highest in the record during 1900–2018, and in February 2010 (right panels), when the index was lowest in the record

to quantify the meridionality of the circulation. These include the Arctic Dipole (AD, Overland and Wang 2005), the closely related Dipole Anomaly (DA, Wu et al. 2006; Wang et al. 2009), the Central Arctic Index (CAI, Vihma et al. 2012), and the Meridional Circulation Index (Francis and Vavrus 2015). Although the exact definitions of AD, DA, and CAI differ, they all are related to the MSLP difference between the North American and Eurasian parts of the Arctic Ocean, and are therefore used to characterize the transport of marine air masses from the Pacific or Atlantic to the central Arctic and the atmospheric forcing on sea ice drift along the Transpolar Drift Stream. Unlike, AD, DA and CAI, MCI is calculated separately for each location, as MCI = v|v| / (v^2 + u^2), where u and v are the zonal and meridional wind components, respectively. Hence, MCI is applicable in quantifying the waviness of the jet stream. In general, strong meridionality of the flow is related to negative values of AOI and NAOI. Further, important regional features of the circulation include the Siberian High, Ural Blocking, Greenland Blocking, the Scandinavian Pattern, and Atlantic Ridge.

The Siberian High (SH), usually centered approximately around Lake Baikal, is the most prominent feature of atmospheric circulation over Eurasia from September until April (Gong and Ho 2002). The SH affects weather patterns over Eurasia and farther in the Northern Hemisphere. The local effects in the region under the high pressure pattern are related to clear skies and weak winds, favouring extremely low winter temperatures, whereas remote effects in winter are felt via advection of cold northerly air masses to Southeast Asia (Wu et al. 2011) and advection of cold, continental air masses to Europe. The recent strengthening of SH has been associated with a negative trend in winter air temperatures over large parts of Eurasia (Cohen et al. 2014). Some studies have suggested that this is at least partly forced by the sea ice decline in the Barents and Kara seas via troposphere-stratosphere coupling (Kim et al. 2014; Jaiser et al. 2016; Furtado et al. 2016), whereas some studies suggest that this is due to natural variability (McCusker et al. 2016).

The Ural Blocking (UB) is a quasi-stationary anticyclonic circulation feature, which occurs around the Ural Mountains. It is considered a part of the negative phase of the East Atlantic / West Russia wave train, which is often associated with the positive phase of NAO (Luo et al. 2016a). There is a positive feedback between the sea ice loss in the Barents and Kara seas and UB (Gong and Luo 2017). When UB occurs, the associated clockwise circulation advects warm air masses to over the Barents and Kara seas, resulting in or enhancing the sea ice loss. The sea ice loss and related atmospheric warming regionally reduces the north-south temperature gradient in the sector around 60°E. This reduces the zonal flow, and favours the occurrence and strengthening of UB. Comparing periods 1979–1999 and 2000–2013, UB events have indeed become more persistent, contributing to the Warm Arctic – Cold Eurasia pattern in winter (Luo et al. 2016a, b). When UB merges with the Siberian High, anomalously cold winters occur over particularly large areas in central Asia.

The Scandinavian (SCA) pattern is seen, e.g., in the 700 hPa height anomalies. The pattern has the primary centre of action around Scandinavia, with two other centres of action with the opposite sign, one over the north-eastern Atlantic and the other over central Siberia. The positive phase of SCA is characterized by prominent anticyclonic anomalies around Scandinavia, in winter giving rise to below-normal temperatures across central Russia and western Europe, above-normal precipitation across southern Europe, and dry conditions over Scandinavia (Bueh and Nakamura 2007). A closely related patters, characterized by a positive pressure anomaly over the north-eastern Atlantic Ocean and a negative one over Scandinavia, is often called as the Atlantic Ridge (Barrier et al. 2014). Accordingly, the Atlantic Ridge favours north-westerly winds in Scandinavia, and is also associated with remarkable sea surface temperature anomalies.

The Aleutian low pressure center is a dominant atmospheric pattern over the North Pacific in winter. The Aleutian low appears in temporally averaged pressure fields, but the instantaneous conditions in the region are characterized by transient cyclones (Sect. 1.3) and a background stationary wave pattern. The wintertime variability of the Aleutian Low is associated with the Pacific – North American Pattern and the AO (Overland et al. 1999).

Due to its high orography, orientation perpendicular to westerly winds, and large extent in north-south direction, Greenland ice sheet tends to block the lower-tropospheric westerly flow. The strength of the blocking depends on both thermodynamic and dynamic factors (Rajewicz and Marshall 2014). The Greenland Blocking Index (GBI, Fang 2004; Hanna et al. 2013, 2016) is defined as the mean 500 hPa geopotential height for the 60–80 °N, 20–80 °W region, but also slightly different indices can be used to characterize the Greenland Blocking (Scherrer et al. 2006; Davini et al. 2012). Greenland Blocking has recently varied a lot with a lot of extreme values since 2001 (Hanna et al. 2016, 2018), probably related to the strongly meandering Polar front jet stream (Overland et al. 2015). The recent rapid climate warming over Greenland has contributed to thermodynamic forcing of Greenland Blocking, with warmer air expanding and raising geopotential heights (Hanna et al. 2016). According to Hanna et al. (2016), decadal variations of the GBI are related to variations in NAO and the Atlantic Multidecadal Oscillation (AMO).

1.3 Synoptic-Scale Cyclones

1.3.1 Theoretical Background

Extratropical synoptic-scale cyclones are fundamental atmospheric circulation system occurring on daily basis in the mid and high latitudes, including the Arctic region. Genesis, development, dissipation, and spatial structure of cyclones in midlatitudes have been extensively investigated (e.g., Bluestein 1993; Hoskins et al. 1985; Davis and Emanuel 1991; Lackmann 2002). However, studies particularly on cyclone dynamics over the Arctic have been very limited (e.g., Aizawa et al. 2014; Tao et al. 2017b). Based on the existing statistics, cyclones generally take 3–6 days to develop, reaching a sea level pressure (SLP) about 970–1000 hPa at cyclone centers, and 3–6 days to dissipate, and move eastward or northeastward at a speed of about 50 km/h. The cyclone diameter is roughly 1000–2500 km on average. Cyclones play predominant role in driving variation of weather elements or occurrence of weather events along their travelling pathways and adjacent areas. Intense cyclones can cause wind gusts, heavy rainfall or snowfall, and large fluctuations of air temperatures, possibly resulting in extreme events such as ocean wave surge, flooding, costal erosion, blizzard, and cold air outbreak. All of these extreme events certainly impact daily life and may cause severe consequences including destruction of infrastructures and loss of properties and even life.

Theoretical study of extratropical cyclones can be traced back to the First World War. A comprehensive understanding about the structure and developing mechanisms of cyclones in the mid-latitudes have been established along with development of atmospheric dynamics and observational network (e.g., Bluestein 1993; Holton 2004). The genesis of cyclones is fundamentally driven by baroclinic instability, i.e., perturbed thermal contrast between warm and cold air masses and resulting conversion of available potential energy to kinetic energy by secondary atmospheric circulation. The coupling between upper level waves and surface cyclones enhances the secondary circulation, in particular when the left exit of upper-level jet ahead of wave trough is located above the surface cyclone. This increases warm and cold temperature advection and, in turn, strengthens baroclinicity, leading to an increased conversion of available potential energy to kinetic energy and intensifying cyclone development. Once the upper level waves catch the surface cyclone and in particular when the surface cyclone is relocated beneath the left entrance of the jet behind the wave trough, the secondary circulation and baroclinicity weaken and then the cyclone decays. Due to the nature of the driving mechanism, cyclones in the mid-latitudes are predominantly characterized by baroclinic structures over their majority of lifetime.

To quantitatively describe details of the cyclone structures and driving mechanisms mentioned above, various theories have been developed, such as the Quasi-Geostrophic (QG) theory (e.g., Bluestein 1993). During recent decades, the potential vorticity (PV) has been increasingly employed to diagnostically analyze and interpret mid-latitude cyclone development (e.g., Hoskins et al. 1985; Davis and Emanuel 1991; Hirschberg and Fritsch 1991; Lackmann 2002; Liberato 2014). In particular, the PV perspective provides an insight into impacts of upper troposphere - lower stratosphere planetary waves on surface cyclones. Associated with deepening and southward amplification of the waves, positive PV anomaly occurs in mid-latitudes. The PV anomaly intensifies surface cyclone through enhancing troposphere baroclinicity ahead of upper-level PV anomaly or wave trough.

Although cyclones in the Arctic, in particular those originating from the mid-latitudes, can share many common features with the their counterparts in the midlatitudes, recent studies found that at least some intense and long-lived Arctic cyclones can be very different in their structures and driving mechanisms (Tanaka et al. 2012; Aizawa et al. 2014; Aizawa and Tanaka 2016; Tao et al. 2017a, b). Tao et al. (2017a) investigated a long-lived Arctic cyclone in September 2010. This cyclone demonstrates barotropic structures almost throughout its lifetime (Fig. 1.4), rather than baroclinic structures found in mid-latitude cyclones as discussed above. The baroclinic instability only occurred near the surface and in the low troposphere at the initial state of the surface cyclone genesis and development. After the generation of the surface cyclone, a synoptic-scale, symmetric stratospheric vortex quickly moved to be just above the surface cyclone, showing a barotropic structure with a warm core in the upper troposphere-lower stratosphere and a cold core in the low-mid troposphere. The positive PV anomaly induced by the downward intrusion

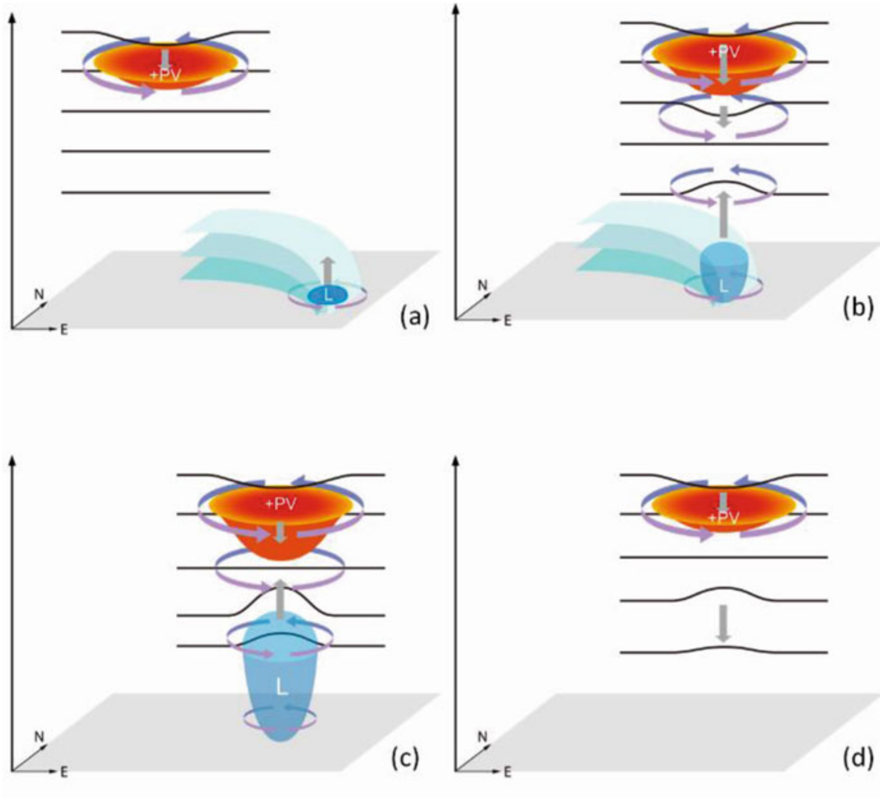

Fig. 1.4 The concept model showing Arctic cyclone structures and driving mechanisms in (**a**) Pre-generation phase; (**b**) development and intensification phase; (**c**) mature and persistence phase; and (**d**) decay and dissipation phase. The red shading represents the lower stratosphere positive PV anomaly associated with the downward intrusion of the stratospheric vortex; the blue shading the troposphere cyclone; the black solid lines the air temperature contours from the mid-troposphere to the lower stratosphere; and the light blue surfaces the air temperature isosurface associated with a surface and low-troposphere front. The vertical and horizontal curved vectors indicate the vertical motions and the horizontal circulations. (Adapted from Tao et al. 2017a, © American Meteorological Society. Used with permission)

of stratospheric vortex played a decisive role in intensifying the surface cyclone and maintaining its intensity over an extended time period. The strengthening and persistence of the PV anomaly resulted from the out-of-phase occurrence of the upper level warm core and the lower level cold core.

1.3.2 Observed Arctic Cyclones

Due to the nature of their driving mechanisms, extratropical cyclones are mainly generated over specific areas and moving along preferred tracks. Cyclones occurring in the Arctic region can originate from the mid-latitudes or be generated locally. There are two major climatological cyclones tracks in the Northern Hemisphere (Fig. 1.5; Zhang et al. 2004). One ranges from the northwest Pacific coastal area and shelf seas to the Bering Sea and Aleutian Islands and the other from the northeast coast of the North American continent to the Icelandic Sea and Barents Sea. The cyclone tracks show obvious seasonality, being more confined over the North Pacific and North Atlantic oceans during winter. During summer, there are more numerous cyclones occurring over the landmasses. Cyclones also have a lower central SLP,

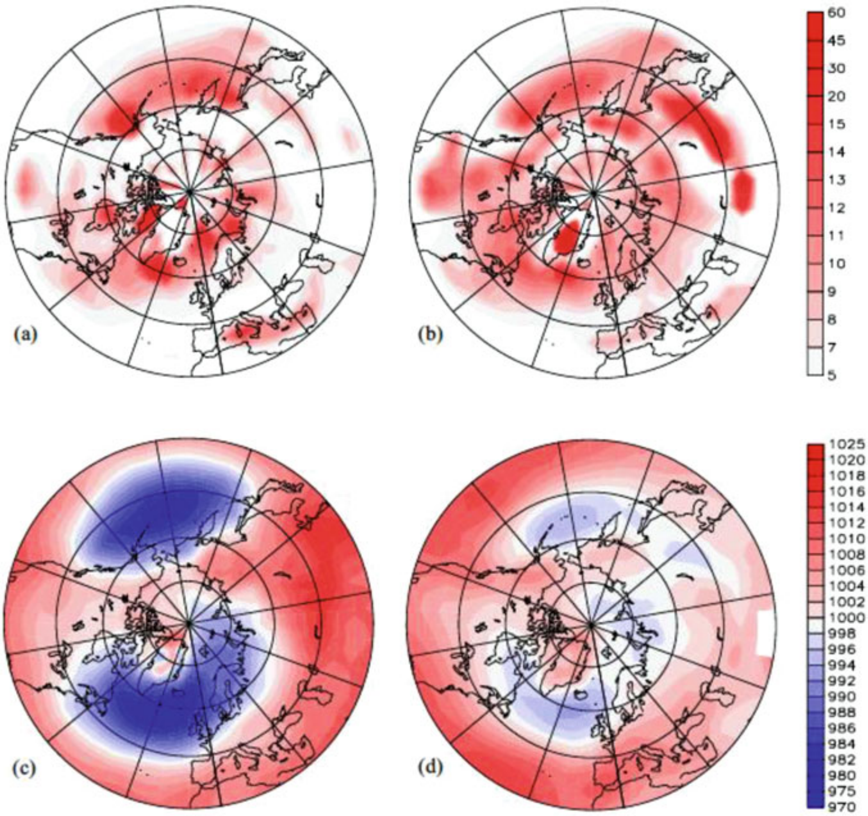

Fig. 1.5 The climatological count (unit: count per 10^5 km^2) of cyclone centers in (**a**) winter and (**b**) summer; and the composite climatological mean sea level pressure (unit: hPa) of cyclone centers in (**c**) winter and (**d**) summer. (Zhang et al. 2004; their Fig. 2; © American Meteorological Society. Used with permission)