Arctic Sea Ice Loss in the Pacific Sector and Its Impacts on Sudden Stratospheric Warming Events

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Arctic Sea Ice Loss in the Pacific Sector and Its Impacts on Sudden Stratospheric Warming Events

By

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Submitted in Partial Fulfillment of the Requirements for the Degree of Doctor of Philosophy in Marine Science: Coastal and Marine Systems Science in the School of the Coastal Environment Coastal Carolina University 2022

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Abstract

The Arctic sea ice is a critical indicator of climate change. The extent of sea ice coverage over the Arctic Ocean has dramatically declined over the past few decades. The impact has been extensively studied through observations suggesting a linkage between the anomalously warm Arctic surface associated with the Arctic sea ice loss and the mid-latitude surface cooling in the subsequent boreal winter. This linkage could involve the wintertime stratospheric circulation by enhancing the upward planetary wave activity and weakening the polar vortex. With recent advances in climate model, more relevant studies relied on numerical simulations and some suggested that the effects of sea ice reduction on the atmospheric circulation and, in particular, on the warm Arctic-cold continent pattern at the surface are attributed to internal variability. Understanding the impact of sea ice changes on the atmospheric circulation is crucial for predicting and assessing climate changes in the coming decades as well as extreme weather.

The overarching goal of this thesis is to improve our basic understanding of the physical processes that link the large-scale atmospheric circulation, particularly for Sudden Stratospheric Warming (SSW) events, and Arctic sea ice loss. In addition, the roles of internal variabilities, namely, the Quasi-biennial Oscillation (QBO) and Madden-Julian Oscillation (MJO), in modulating the atmospheric response to Arctic sea ice loss are examined. To avoid conflating the effects of sea ice loss in different sectors, this dissertation solely focuses on the Chukchi-Bering Seas (i.e., the Pacific sector) where the observed autumnal Arctic sea ice extent shows the strongest decline in recent decades. Observational data record period is too short to provide statistically convincing conclusion
on how the underlying mechanism works in generating a global atmospheric response. Therefore, global climate model with well-resolved stratosphere (i.e., Whole Atmosphere Community Climate Model version 6 from the National Center for Atmospheric Research) is used. Since QBO and MJO are internally generated in the model, their roles on the responses are examined.

During the easterly QBO phase (EQBO), the prescribed sea ice loss, although culminating in autumn, induces a near-surface warming that persists into winter and deepens as the SSW develops. The resulting temperature contrasts foster a deep cyclonic circulation over the North Pacific, which elicits a strong upward wavenumber-2 activity into the stratosphere, reinforcing the climatological planetary wave pattern. The induced geopotential anomalies in the lower troposphere also project onto the anomalous patterns typically observed prior to SSWs. While not affecting the SSW occurrence frequency, the amplified wave forcing in the stratosphere significantly increases the SSW duration and intensity, enhancing thereafter cold air outbreaks over the Northern Hemisphere continents. However, for the westerly QBO phase (WQBO), the induced warming does not extend vertically into the middle troposphere but instead spreads horizontally prior to SSW onset. The resulting temperature contrasts weaken the precursor of SSW over the North Pacific. The other SSW precursor over the Europe is slightly strengthened. The cancelling effect leads to insignificant change of SSW duration and wind reversal intensity. The SSW occurrence frequency is significantly increased in response to Arctic sea ice loss. To this end, despite the prescribed sea ice loss being identical, the structure of temperature response is different between EQBO and WQBO.
The background state, conditioned by the QBO, influences the response of SSWs to the sea ice loss. While difficult, differentiating the QBO phases may be important in understanding the stratospheric response to sea ice loss.

In the climate system, there are additional factors that might modulate the atmospheric response to sea ice loss, namely the effects of El Niño-Southern Oscillation (ENSO), the solar cycle, and global warming. In this thesis, these factors are all excluded in the experimental design. Nevertheless, the tropical variability MJO is internally generated in WACCM6. Since the SSW precursor over North Pacific could be excited by MJO phase 7, the role of MJO is also examined. In general, MJO phase 7 occurs more frequently prior to SSW onset. However, in our experiment, there is very little difference in the MJO phase 7 response during EQBO, affirming the conclusion that strengthened SSW precursor over the North Pacific is induced by sea ice loss and autumnal sea ice loss extends the SSW duration and strengthens the accompanying stratospheric wind reversal. For WQBO, the weakened SSW precursor over the North Pacific is also mainly due to sea ice loss instead of MJO. Additionally, teleconnection patterns excited by MJO are enhanced in response to sea ice loss regardless of QBO phases.
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List of Symbols and Abbreviations

Roman Symbols

$x, y, z$: Local Cartesian Coordinate in the Zonal, Meridional, and Vertical Direction

$t$: Time.

$A$: Wave Amplitude

$\alpha$: Phase

$k, l$: Horizontal Wave Numbers

$m$: Vertical Wave Numbers

$\omega$: Angular Frequency

$c$: Phase Speed

$c_g$: Group Velocity

$c_i$: Intrinsic Phase Speed

$f$: Coriolis Parameter

$\beta$: Meridional Gradient of Coriolis Parameter

$u$: Zonal Wind

$v$: Meridional Wind

$V_g$: Geostrophic Horizontal Wind

$w$: Vertical Velocity

$\theta$: Potential Temperature

$U_c$: Rossby Critical Velocity for Stationary Waves

$\zeta$: Vertical Component of the Vorticity

$\psi$: Streamfunction
q: Quasi-Geostrophic Potential Vorticity

$\rho$: Density

$N$: Brunt-Vaisala or Inertial Frequency

$H$: Standard Scale Height (~7 km)

$n_k$: Refractive Index

$X$: Zonal Component of Drag Due to Small-Scale Eddies

$Q$: Diabatic Heating

$R$: Ideal Gas Constant for Dry Air

$\kappa$: Poisson Constant

$F$: Eliassen-Palm (EP) Flux

$E$: Plumb Flux

$\alpha$: Earth Radius

$\lambda$: Longitude

$\phi$: Latitude

$\Omega$: Earth’s Rotation Rate

$Z$: Geopotential Height

**Model Experiments**

CNTL: Control model experiment with climatological sea surface temperature and sea ice concentration (1 year with 75 ensemble members).

LOW: Perturbed model experiment with Arctic sea ice loss over Pacific sector (1 year with 75 ensemble members).
Abbreviations

AA: Arctic Amplification
AGCM: Atmospheric General Circulation Model
AMO: Atlantic Multidecadal Oscillation
AMOC: Atlantic Meridional Overturning Circulation
AO: Arctic Oscillation
CAM: Community Atmosphere Model
CAO: Cold Air Outbreak
CESM: Community Earth System Model
CLUBB: Cloud Layers Unified by Binormals
CMIP: Coupled Model Inter-comparison Project Phase
ECMWF: European Centre for Medium-Range Weather Forecasts
EMD: Empirical Mode Decomposition
ENSO: El Niño-Southern Oscillation
EOF: Empirical Orthogonal Function
EP flux: Eliassen-Palm Flux
ERA: ECMWF (European Centre for Medium-Range Weather Forecasts) Re-Analysis
FV: Finite Volume
JRA-55: Japanese 55-year Reanalysis
MAM: Modal Aerosol Module
MCA: Maximum Covariance Analysis
MEGAN: Model of Emissions of Gases and Aerosols from Nature
MERRA: Modern-Era Retrospective Analysis for Research and Applications

MJO: Madden-Julian Oscillation

MLT: Mesosphere and Lower Thermosphere

NAO: North Atlantic Oscillation

NASA: National Aeronautics and Space Administration

NCEP: National Center for Environmental Prediction

NCAR: National Center for Atmospheric Research

NOAA: National Oceanic and Atmospheric Administration

NSIDC: National Snow and Ice Data Center

PC: Principal Component

PDF: Probability Density Functions

PNA: Pacific North America pattern

QBO: Quasi-Biennial Oscillation

SIC: Sea Ice Concentration

SLP: Sea Level Pressure

SST: Sea Surface Temperature

SSW: Sudden Stratospheric Warming

TSMLT: troposphere, stratosphere, mesosphere, and lower thermosphere

WACCM: Whole Atmosphere Community Climate Model

WAF: Wave Activity Flux

WMO: World Meteorological Organization
1. Introduction

Changes in surface albedo, heat fluxes, and near-surface temperature gradient related to Arctic sea ice loss can impact the atmospheric circulation that couples the troposphere and stratosphere. This impact can vary depending on sea ice loss magnitude and region (Sun et al., 2015; McKenna et al., 2018) and can elicit a nonlinear atmospheric response (Semenov & Latif, 2015). A better understanding of the role of sea ice loss on the large-scale circulation is critical for improving prediction of weather and the assessment of climate change (Cohen et al., 2014; Kim et al., 2014; Peings & Magnusdottir, 2014).

The connection of sea ice loss and associated Arctic warming to the large-scale circulation can be attributed to atmospheric waves with very large horizontal wavelength. Referred to as planetary Rossby waves (PWs), these wave perturbations can readily propagate from the troposphere, where they are typically generated, into the stratosphere during the winter months when the prevailing stratospheric flow is eastward (Charney & Drazin, 1961). The decreased sea ice coverage between November and December over the Barents-Kara seas can lead to enhanced upward PW propagation, which weakens the stratospheric wintertime circumpolar flow (or the stratospheric polar vortex) in January and February (Kim et al., 2014). Cases studies based on observations (Zuev & Savelieva, 2019) or model simulations (Zhang et al., 2020) even suggested that the enhanced wave flux could be strong enough to cause the demise of the polar vortex associated with the sudden stratospheric warming (SSW) phenomenon. The SSW-induced temperature
anomalies could subsequently descend into the troposphere on time scales of weeks to months (Baldwin & Dunkerton, 2001), potentially leading to cold air outbreaks (CAOs) over the continents (Kolstad et al., 2010; Kretschmer et al., 2018).

The equatorial wind direction in the lower stratosphere (between 20-30 km) may alter how PWs interacts with the stratospheric polar vortex (Holton & Tan, 1980). Associated with the Quasi-Biennial Oscillation (QBO), this equatorial wind vacillates with a period of about 28 months. The QBO phase during which the equatorial wind is westward tends to confine PW activity in the winter latitudes, possibly leading to a more perturbed stratospheric polar vortex. However, past studies tend to neglect the role of the QBO in examining the stratospheric response to sea ice loss.

The Madden-Julian Oscillation (MJO) is the dominant component of atmospheric intraseasonal variability in the tropics. Diabatic heating anomalies induced by the MJO could excite Rossby wave trains, which propagate across the North Pacific, over North America, and into the Atlantic Ocean (Matthews et al., 2004). It has been found that the MJO can impact the evolution feature of SSWs, such as type (Liu et al., 2014) and frequency (Kang & Tziperman, 2017). However, the role of MJO in modulating the stratospheric response to sea ice loss was neglected in the previous studies.

While the observed autumn Arctic sea ice area has diminished by nearly half since 1980 (National Snow and Ice Data Center, 2018), this decline has occurred mostly in the Pacific sector, near the Bering Strait and the Chukchi Sea. To date, dynamical mechanisms linking this regional sea ice loss and the large-scale circulation (particularly for SSWs) are still unclear, especially the involvement of the stratosphere, QBO and tropical variability, MJO.
To this end, the goal of this thesis is to better understand the large-scale circulation in response to sea ice loss by asking:

- What is the impact of sea ice loss over the Bering-Chukchi Seas on SSWs characteristics?
- Does the QBO modulate the SSWs response to sea ice loss?
- What is the role of MJO in modulating the SSWs response to sea ice loss?

Related to these questions is how SSWs characteristics change and how those changes ultimately affect surface conditions.

To address these questions, this thesis attempts to elucidate the fundamental mechanisms linking the surface processes associated with sea ice loss to the wintertime stratospheric dynamics through numerical experiments using a global chemistry-climate model with a well-resolved stratosphere. To avoid conflating the effects of sea ice loss in different sectors, this study focuses solely on the Pacific sector where the observed autumnal Arctic sea ice extent shows the strongest decline in recent decades (Meredith et al., 2019; Simmonds & Li, 2021). The atmospheric response to sea ice loss in that sector has received much less attention than over the Barents-Kara Seas. We will develop an ensemble of 1-year control experiments in which the climatological sea ice and sea surface temperature are specified as part of the boundary conditions. A perturbation experiment will be conducted by explicitly forcing the model with an anomalously low sea ice loss over the Pacific sector with the QBO constrained to one phase throughout the winter. The differences between the control and the perturbed experiment will be used to address how the autumnal sea ice loss over the Pacific sector impacts the most basic
characteristics of SSWs. Through carefully designed model simulations, the role of QBO and MJO will be examined. In all these experiments, other potential factors influencing the stratosphere (like the El Niño-Southern Oscillation, varying solar conditions, increasing greenhouse gases, and ozone loss due to anthropogenic influence) are minimized.
2. Background On the Atmospheric Rossby Waves and Natural Variability

The atmospheric circulation central to this study is the global scale atmospheric flow that is nearly geostrophic (i.e., in balance between horizontal pressure gradient force and the Coriolis effect). This circulation can be separated into the zonally averaged component (i.e., averaging around each latitudinal circle across all longitudes) or the time-averaged component. Due to the presence of orography and land-sea heating contrast, the latter component varies with longitude. Overall, the circulation can evolve over a broad timescale with flows that are nearly constant in time (i.e., the quasi-stationary circulation), that change with the season (the inter-seasonal or monsoonal circulation), or that vary across various subseasonal-to-interannual time interval (collectively, the low-frequency variability).

Processes associated with the zonally averaged component of the circulation provide a useful framework for understanding the departures of the time-averaged flow from zonal symmetry. In examining the zonal averaged component, we typically split the circulation into the zonal-mean (or background) part and the longitudinally varying (or eddy) part, which may be related to atmospheric waves. We then try to understand the processes that maintain the background flow and those that lead to the development of waves. The most important atmospheric waves are large-scale ones known as planetary Rossby waves (PWs). They readily interact with the prevailing background flow that can evolve as part of the natural variability of the atmosphere and ocean.
As a background, we present here an overview of theories related to PWs and how they can influence the zonally averaged background flow. We then introduce key natural variability relevant to this thesis; namely, the Arctic Oscillation (AO), the Sudden Stratospheric Warming (SSW), the Quasi-Biennial Oscillation (QBO), and the Madden-Julian Oscillation (MJO).

2.1 Overview of Atmospheric Waves

An important property of the atmosphere is its ability to support wave motions. The oscillatory nature related to waves results from the competition between inertia of the air parcel and various restoring forces acting on the air parcel. Atmospheric waves impose perturbations on the wind, density, pressure or temperature fields and can be excited thermally or dynamically. This wave excitation tends to occur in the lower atmosphere. In propagating away from their sources, waves transmit energy and momentum from their source regions across the atmosphere without material transport. The transferred energy and momentum could then be deposited into the background flow elsewhere as the waves dissipate due to some form of damping in the atmosphere like radiative damping (Salby, 1996). Since waves could produce disturbances and impact the atmospheric circulation, it is necessary to properly account for the influence of wave propagation.

2.1.1 Linear Perturbation Theory

To qualitatively analyze atmospheric waves, linear perturbation theory is typically used. In perturbation method, all field variables are divided into two parts: a basic state part (i.e., the background or prevailing flow) and a disturbance or perturbation part (related to waves). The assumptions are that the basic state variables satisfy the dynamical
equations, evolve slowly compared to the waves, and the perturbation fields are small enough so that products of the perturbed amplitudes are negligible. When the disturbance fields attain large enough amplitudes during which nonlinear effects can no longer be neglected, wave breaking occurs, leading to a rapid, irreversible deformation of material contours. The waves are then greatly dissipated.

Applying linear perturbation theory, the nonlinear dynamical equations are reduced to linear equations of perturbation variables with basic state variables as coefficients. Solution of these equations helps determine the structure and characteristics of disturbances, such as vertical structure, propagation speed, and conditions for growth or decay of the waves.

2.1.2 Properties of Linear Waves

Since atmospheric disturbances are not purely sinusoidal, Fourier series of various sinusoidal components or modes are typically employed to represent the disturbances. However, for simplicity, we can initially consider a monochromatic three-dimensional wave of some field \( r \) as:

\[
r(x, y, z, t) = Ae^{i\alpha} = Ae^{i(kx + ly + mz - \omega t)}
\]  

(2.1)

Here, \( A \) is wave amplitude, \( \alpha \) is phase, \( k, l \) are horizontal wave numbers, \( m \) is vertical wave numbers, and \( \omega \) is angular frequency. Also, \( x, y, z \) are the local Cartesian coordinate, respectively, in the zonal, meridional, and vertical directions and \( t \) is time.

Wave motions are tied to its group velocity and phase velocity. The former describes the velocity of the wave energy and the latter how fast the wave phase moves. A disturbance moving the \( x \)-direction can be considered as the product of a carrier wave moving at a
certain phase speed \( (c = \frac{\omega}{k}) \) and an envelope that travels at a group velocity \( (c_g = \frac{\partial \omega}{\partial k}). \)

Waves whose phase speeds are independent of the wavenumber are nondispersive. These waves preserve their shape as they propagate in space at the phase speed of the wave. Dispersive waves have phase speeds that vary with wavenumbers, so the shape of various wave sinusoidal components will change as the waves propagate. The group velocity is usually different than the phase speed. As shown in Fig 2.1, individual wave components may move faster or slower than the wave group as the group propagates along for dispersive waves. The downstream development of a dispersive wave could occur when its group velocity exceeds the phase velocity. Therefore, through the relationship between phase speed and wavenumber, downstream or upstream development of the waves can be predicted.

\[\text{Fig 2.1. Schematic showing propagation of wave groups: (a) group velocity less than phase speed and (b) group velocity greater than phase speed. Heavy lines show group velocity, and light lines show phase speed. [From Holton and Hakim, 2012].}\]
2.1.3 Rossby Waves

The characteristics of atmospheric waves are known through the wave solution to the atmospheric primitive equations using the linear perturbation method. Atmospheric waves can be classified in various ways according to their physical properties like spatiotemporal scales and their restoring mechanisms. Rossby waves are large-scale motions with the Earth’s rotational effects as the restoring mechanism, specifically the potential vorticity gradient. Synoptic-scale Rossby waves play an important role in dynamics of the troposphere and planetary scale Rossby waves (PWs) the middle atmosphere. Thus, they are very relevant to our study.

Large-scale topography and thermal contrast (e.g., due to temperature difference between ocean and the continents) can excite Rossby waves. Since heating could be impacted by distribution of the orography, these two forcing are difficult to separate (Held et al., 2002; Chang, 2009). Disturbance could create imbalanced temperature distribution horizontally and vertically. Although the wind field can accommodate such distribution through changes in wind speed and direction, the imbalances could develop as long as forcing exists. Therefore, the wind will constantly change directions and develop into wave pattern. PWs could be stationary with phase surfaces fixed with respect to the earth or could be traveling with phase surfaces moving in longitude with periods of a few days to a few weeks.

Rossby waves can be seen by plotting the geopotential height of the atmosphere. Fig 2.2 shows the daily geopotential height at the constant 500-hPa surface. At this instance, the geopotential height shows strong longitudinal variation, which is due to Rossby waves.
Roughly four troughs and four ridges of geopotential height field are apparent on 1st January 2018, so the Rossby waves of zonal wavenumber 4 \((k = 4)\) is dominant. Few days later (7th January 2018), these troughs and ridges evolve and the geopotential height pattern adopts zonal asymmetries induced by the attendant wave disturbances.

Fig 2.2. 500hPa geopotential height \((Z)\) using NCEP/NCAR (National Center for Environmental Prediction/National Center for Atmospheric Research) reanalysis data. Left is on January 1st, 2018 and right is on January 7th, 2018. Unit is in decameters.

The simplest dispersion relationship can be derived by assuming a wave solution to the linearized vorticity equation as initially presented by Rossby (1939). For this equation, the atmosphere is assumed to have a constant density, a uniform zonal flow in the background, and no vertical motion. For the midlatitude region, the simplified vorticity equation is:

\[
\left( \frac{\partial}{\partial t} + u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} \right) \zeta + \beta v = 0 \tag{2.2}
\]

where \(u, v\) are the zonal (eastward) and meridional (northward) component of the horizontal wind, \(\zeta\) is vertical component of the vorticity, and \(\beta = \frac{df}{dy}\) is the meridional
gradient of the Coriolis parameter \((f)\). Equation (2.2) is applicable in the latitudinal range near a local Coriolis parameter value or in a \(\beta\)-plane.

According to linear perturbation method, the motion is assumed to consist of a constant basic, zonally averaged (or zonal-mean) state plus a small horizontal perturbation that departs from that state. For a uniform zonal background flow, we have 
\[ u = \bar{u} + u' \]
and 
\[ v = v' \]
with the overbar representing the zonal mean operation. We note that this operation is Eulerian as it is performed at each fixed location. Upon substituting these expressions into equation (2.2), the resulting linearized perturbation equation is:
\[
\left( \frac{\partial}{\partial t} + \bar{u} \frac{\partial}{\partial x} \right) \nabla^2 \psi' + \beta \frac{\partial \psi'}{\partial x} = 0
\] (2.3)

Here, \( \psi' \) is perturbation streamfunction related to \( u' \) and \( v' \). The del square \((\nabla^2)\) is the horizontal Laplacian.

After substituting wave solutions of the form \( \psi' = Ae^{i(kx + ly - \omega t)} \) in equation 2.3, the dispersion relationship is obtained from which the wave phase speed and zonal group velocity can be derived. Namely,
\[ \omega = \bar{u}k - \frac{\beta k}{k^2 + l^2} \] (2.4)
\[ c = \frac{\omega}{k} = \bar{u} - \frac{\beta}{k^2 + l^2} \] (2.5)
\[ c_g = \frac{\partial \omega}{\partial k} = \bar{u} + \beta \frac{k^2 - l^2}{(k^2 + l^2)^2} \] (2.6)

Here, \( c \) denotes the phase speed and \( c_g \) the group velocity along the zonal direction and relative to the ground.

The direction of movement for Rossby waves depends on two factors: the speed of mean
zonal flow and the wavenumbers. As $\beta$ is positive, equation (2.5) suggests that the wave phase always drifts westward relative to the mean zonal flow ($\bar{u}$) and that the zonal phase speed increases rapidly with increasing wavelength (or equivalently decreasing wavenumbers). That is, the intrinsic phase speed of PW, defined as $\hat{c} = c - \bar{u}$, is westward (i.e., negative). The westward propagating nature of Rossby waves can also be explained by conservation of potential vorticity. For a barotropic atmosphere (i.e., one with constant depth and density nearly independent of altitude), the absolute vorticity is conserved. A southward (northward) displacement results in a positive (negative) vorticity perturbation. The subsequent perturbation vorticity field represents westward propagation of Rossby waves as shown in Fig 2.3.

Relative to the ground, smaller synoptic-scale Rossby waves will appear to move eastward through advection by the predominantly eastward background zonal flow. For larger scale PWs (with small wavenumbers), the ground relative phase speed can approach zero and the PWs are said to be quasi-stationary. For perfect stationary wave ($c = 0$), we see from
equation (2.5) that \( \bar{u} = \bar{u}_s \equiv \frac{\beta}{k^2 + l^2} \). Stationary PWs could be strongly excited by flow over major mountain ranges.

Unlike phase speed, the zonal group velocity in equation (2.6) can be either eastward or westward relative to the mean flow, depending on the ratio of zonal and meridional wave numbers. Generally, group velocity tends to be eastward for PWs relative to the ground.

For synoptic-scale Rossby waves, their eastward group velocity is larger than their eastward phase speed so new disturbances tend to occur downstream of existing ones (similar to the diagram in Fig 2.2b). Due to their dispersive nature, stationary PWs also have a group velocity despite their stationary phase. Using \( \bar{u}_s \) defined above and equation (2.6), the zonal group velocity of stationary waves can be expressed as (where the \( s \) subscript represents stationary waves):

\[
\frac{c}{c_{gs}} = \frac{2\bar{u}_s k^2}{k^2 + l^2}
\]

(2.7)

Eastward zonal group velocities of stationary waves relative to the ground implies that new disturbances tend to develop downstream of existing disturbances. Charney and Eliassen (1949) used the topographic Rossby wave framework to model stationary Rossby waves and explain the winter mean longitudinal distribution of 500-hPa heights in the Northern Hemisphere midlatitudes.

After Rossby (1939), Charney and Drazin (1961) investigated the vertical propagation of Rossby waves and found its dependence on the vertical distribution of the background zonal-mean zonal wind. This dependence is quantified by the effective index of refraction.

Charney and Drazin (1961) considered the quasi-geostrophic potential vorticity equation
on the midlatitude $\beta$-plane.

$$\left(\frac{\partial}{\partial t} + \mathbf{V}_g \cdot \nabla\right) q = 0 \quad (2.8)$$

Here, $q$ is the quasi-geostrophic potential vorticity defined as

$$q = \nabla^2 \psi + f + \frac{f_0^2}{\rho N^2} \frac{\partial}{\partial z} \left( \rho \frac{\partial \psi}{\partial z} \right)$$

and $\mathbf{V}_g$ is the horizontal wind vector under geostrophic balance and is related to the streamfunction $\psi$ when $f$ is a constant. In defining $q$, $\rho(z)$ is the density vertical profile, and $N$ is the Brunt-Vaisala or inertial frequency (of a stable atmosphere). In assuming that the motion consists of a small-amplitude disturbance superposed on a constant zonal-mean zonal flow $\bar{u}$, the perturbation $q'$ field must satisfy:

$$\frac{\partial q'}{\partial t} + \bar{u} \frac{\partial q'}{\partial x} + \beta \frac{\partial \psi'}{\partial x} = 0 \quad (2.9)$$

, where $q'$ is the perturbed potential vorticity expressed as $q' = \nabla^2 \psi' + \frac{f_0^2}{\rho N^2} \frac{\partial}{\partial z} \left( \rho \frac{\partial \psi'}{\partial z} \right)$.

Assuming a wavelike form in $x, y, z$ with an angular frequency $\omega$, $\psi'$ may be written as:

$$\psi'(x, y, z, t) = \Psi_0 \frac{z}{H} e^{i(kx+ly+mz-\omega t)} \quad (2.10)$$

, where $H$ is a standard scale height (~7 km). The factor $\frac{z}{H}$ is introduced to account for density decreasing with height and to simplify the mathematics.

In substituting equation 2.10 into 2.9, a dispersion relation can be obtained, which is related to the vertical structure of PWs.

$$m^2 = \frac{N^2}{f_0^2} \left[ \frac{\beta}{(\bar{u} - c)} - (k^2 + l^2) \right] - \frac{1}{4H^2} \quad (2.11)$$

Since $m^2 > 0$ is required for vertical propagation, we can see that the zonal wind, zonal
phase speed, and horizontal wavenumbers impact the vertical structure of Rossby waves.

For stationary wave ($c = 0$), Fig 2.4 shows the regime transition from where they can propagate vertically to where they become evanescence as a function of zonal wind and horizontal wavenumber.

![Figure 2.4](image)

*Fig 2.4. The transition between vertically propagating and evanescent waves as a function of zonal wind $U$ and horizontal wavenumber $k$ and $l$. The calculations are for a latitude of $45^\circ$ and $N = 2 \times 10^{-2} \text{s}^{-1}$. [From James, I. N. (1995), Introduction to circulating atmospheres. Cambridge University Press].*

As a result, vertically propagation of stationary Rossby waves can exist only for mean zonal flows that satisfy the condition:

$$0 < \bar{u} < \frac{\beta}{\left[(k^2 + l^2) + f_0^2/(4N^2H^2)\right]} = U_c$$  \hspace{1cm} (2.12)

Here, $U_c$ is called the Rossby critical velocity for stationary waves. From Fig 2.4 and equation 2.12, we see that only ultra-long stationary PWs (small wavenumbers) can readily propagate into the stratosphere in middle and high latitudes during the wintertime when the flow is relatively strong and eastward. Vertical propagation of PWs
is inhibited for westward or very large eastward background flow. In other words, energy is trapped (reflected) in regions where the zonal winds do not fulfill the relationship in equation 2.12. During summer, the stratospheric mean zonal winds are westward, so stationary PWs are trapped vertically. Although zonal flows are not constant in real atmosphere, this theory still provides a qualitative guide for estimating vertical propagation of Rossby waves in observations.

Following the work of Charney and Drazin (1961), Matsuno (1970) first introduced the refractive index \( n_k \), which depends on latitude and height, as a diagnostic tool for investigating the impact of background zonal flow on the propagation of PWs as they penetrate into the stratosphere. By allowing for the background zonal wind to depend on latitude and altitude, i.e., \( \bar{u} = \bar{u}(y, z) \) and neglecting the small vertical variation of \( N^2 \), a linear perturbation method similar to those discussed above yield the dispersive relation:

\[
\omega = \bar{u}k - \frac{\bar{q}_y k}{\left\{ k^2 + l^2 + \frac{f_0^2}{N^2} \left( m^2 + \frac{1}{4H^2} \right) \right\}} \tag{2.13}
\]

Here, \( \bar{q}_y \) is the poleward gradient of the potential vorticity of the basic state. It can be expressed as \( \bar{q}_y = \beta - \bar{u}_y - \frac{1}{\rho \partial z} \left( \frac{\rho f^2}{N^2} \frac{\partial \bar{u}}{\partial z} \right) \). If the vertical and meridional variations of \( \bar{u} \) and \( \rho \) are neglected, equation 2.13 reduces to equation 2.4.

In rearranging equation 2.13 and requiring that the vertical group velocity be positive, the index of refraction is defined as:

\[
n_k^2(y, z) = \frac{\bar{q}_y}{(\bar{u} - c)} - k^2 - \frac{f^2}{4H^2N^2} \tag{2.14}
\]

Thus, PW propagation depends on more complex criteria than the simple Charney-Drazin
condition shown in equation (2.12), which applies when \( \bar{u} \) only varies with altitude. Generally, PWs propagate in regions where \( n_k^2 \) is positive and avoid the negative regions. At a critical surface where \( \bar{u}(y, z) = c, n_k^2 \) becomes infinite and reflects the simplest example of PW wave breaking (Holton & Hakim, 2012). For stationary PWs \( (c = 0) \) of low zonal wave number, \( n_k^2 \) is positive in the region with eastward winds and increases to infinity along a critical surface where the mean flow vanishes. For a representative zonal-mean zonal wind structure in the winter hemisphere, the refractive index associated with PW is positive and increases rapidly toward the equatorial zero wind line. Therefore, PWs activity tends to propagate upward and equatorward, and wave breaking occurs in the vicinity of the equatorial critical line (Holton & Hakim, 2012). We will illustrate PW propagation in the sub-section below.

### 2.1.4 Mean Flow and Wave Interaction

The evolution of the zonally averaged circulation can be understood as the interactions between the mean flow and atmospheric waves. The Eulerian mean equations are obtained by separating each variable in the relevant atmospheric equations (in near geostrophic balance) into a zonal-mean part and a disturbance part and, then, taking the zonal average. Equations 2.15 and 2.16 show the Eulerian mean equations respectively for the zonal momentum equation and the thermodynamic equation on the midlatitude \( \beta \) plane. As before, the overbar denotes zonal mean quantity and the asterisk denotes the departure from the zonal mean (i.e., the eddy part). The subscripts indicate partial derivatives and all asterisked quantities will be taken small.

\[
\ddot{u}_t - f_0 \ddot{v} - \dddot{X} = -(\nu^* u^*)_y \tag{2.15}
\]
$$\ddot{\theta}_t + \ddot{w}\ddot{\theta}_x - \ddot{Q} = -(v^*\theta^*)_y$$  \hspace{1cm} (2.16)

Here, $v$ is the meridional velocity, $w$ is vertical velocity, and $\theta$ is potential temperature. The term $\dddot{X}$ represents the zonal component of drag due to small-scale eddies and $Q$ the diabatic heating. The term on the right-hand side of equation 2.15 represents the meridional convergence of the meridional flux of zonal momentum by eddies, i.e., the eddy momentum flux. The term on the right-hand side of equation 2.16 describes the meridional convergence of the eddy heat flux.

Similarly, by taking the zonal average of the meridional momentum equation and assuming hydrostatic balance, we can obtain the thermal wind relationship.

$$f_0\dddot{u}_z + H^{-1}Re^-\kappa\dddot{\theta}_y = 0$$  \hspace{1cm} (2.17)

Here, $R$ is ideal gas constant for dry air and $\kappa$ is Poisson constant. This relationship imposes a strong constraint on the zonal-mean zonal wind structure and the temperature distribution in the meridional plane and couples the Eulerian mean meridional (overturning) circulation described by ($\dddot{v}, \dddot{w}$) to the eddy fluxes in equations (2.15) and (2.16). For example, if the mean meridional circulation is zero, then the eddy momentum flux would only accelerate the zonal-mean zonal wind if drag is negligible. Furthermore, the eddy heat flux would only change the temperature distribution if diabatic heat is absent. In the scenario without ($\dddot{v}, \dddot{w}$), the thermal wind balance would be violated.

Because the influence of atmospheric waves appears in the momentum and thermodynamics equations through eddy fluxes, the Eulerian mean equations (2.15) and (2.16) neither clearly diagnose the impact of waves on the zonal-mean zonal flow nor the mean meridional circulation associated with mass transport. In fact, for a typical Northern
Hemisphere winter, \((\bar{v}, \bar{w})\) derived from observed eddy fluxes in the troposphere describes a mean meridional circulation with a thermally direct cell (or the Hadley cell) along with a thermally indirect cell in the midlatitudes inconsistent with observed tracer transport (Holton & Hakim, 2012).

In practice, the Eulerian-mean equations are transformed to efficiently diagnose the impact of waves and appropriate mean meridional circulation. Andrews and McIntyre (1976) first introduced the transformed Eulerian mean (TEM) equation by defining the residual circulation \((\bar{v}^*, \bar{w}^*)\). The expressions for \(\bar{v}^*\) and \(\bar{w}^*\) on the midlatitude \(\beta\)-plane are shown in equations 2.18 and 2.19. With this transformation, the residual vertical velocity could represent that part of mean vertical velocity with a contribution to adiabatic temperature change that is not canceled by the eddy heat flux divergence (Holton & Hakim, 2012). Thus, the residual circulation is consistent with mass transport.

\[
\bar{v}^* = \bar{v} - \rho^{-1}(\rho \bar{v}' \bar{\theta}' / \bar{\theta}_z)_z \tag{2.18}
\]

\[
\bar{w}^* = \bar{w} + (\bar{v}' \bar{\theta}' / \bar{\theta}_z)_y \tag{2.19}
\]

Substituting these transformations in equations 2.15 and 2.16 yield the TEM formulations in which the eddy momentum and heat fluxes do not act separately to change zonal-mean circulation (see Equations 2.20 and 2.21). The divergence of \(F\) incorporates the impact of eddy heat and momentum fluxes together.

\[
\bar{u}_t - f_0 \bar{v}^* - \bar{X} = (\rho)^{-1} \nabla \cdot F \tag{2.20}
\]

\[
\bar{\theta}_t + \bar{w}^* \bar{\theta}_z = \bar{Q} \tag{2.21}
\]

, where

\[
F = (-\rho v' u', \rho f v' \bar{\theta}' / \bar{\theta}_z)
\]
Known as the Eliassen-Palm (EP) flux, the vector quantity \( F = (F^{(y)}, F^{(z)}) \) was first presented by Eliassen and Palm (1961). Neglecting \( \bar{X} \), the EP flux divergence, as shown on the right-hand side of equation 2.20, can manifest in accelerating the mean flow and driving the residual meridional circulation (\( \bar{v}^* \)). The contribution to the residual vertical circulation (\( \bar{w}^* \)) can be related the diabatic heating. The meridional and vertical residual circulation are linked together by the linearized continuity equation (not stated here).

Eliassen and Palm (1961) presented a basic theorem for wavelike disturbances to a zonal mean wind \( \bar{u}(y, z) \) without frictional or diabatic effects (i.e., \( \bar{X} = \bar{Q} = 0 \)). The Eliassen-Palm non-acceleration theorem states that the divergence of \( F \) (which depends on certain basic physical properties of the flow) is zero if the eddy disturbances are steady, linear, frictionless, and adiabatic and if the mean flow is conservative. In other words, steady and non-dissipative waves under adiabatic condition have no effects on the mean flow.

2.1.5 Physical Interpretation of the Eliassen-Palm (EP) Flux

While helping to quantify the impact of waves on the mean flow, the EP flux is also related to PW group velocity. To demonstrate this, we write the momentum and heat fluxes using the perturbation field shown in equation 2.10.

\[
v' u' = - \frac{\partial \psi'}{\partial x} \frac{\partial \psi'}{\partial y} = - \frac{k l}{2} |\Psi_0|^2 e^{z/H} \quad (2.22)
\]

\[
v' \theta' = \frac{\partial \psi'}{\partial x} \frac{f \theta}{g} \frac{\partial \psi'}{\partial z} = \frac{f \theta}{2g} km |\Psi_0|^2 e^{z/H} \quad (2.23)
\]

According to dispersion relation (2.13), the poleward and vertical component of group velocity can be obtained.
\[
c_{gy} = \frac{\partial \omega}{\partial l} = \frac{2q_y kl}{K_T^4} = \frac{4\tilde{q}_y e^{-z/H}}{K_T^4 |\Psi_0|^2} (-v'u')
\]  
(2.24)

\[
c_{gz} = \frac{\partial \omega}{\partial m} = \frac{2q_y f^2 km}{N^2 K_T^4} = \frac{4\tilde{q}_y e^{-z/H} f}{K_T^4 |\Psi_0|^2} \theta (v' \theta')
\]  
(2.25)

where \(K_T^4 = [k^2 + l^2 + \frac{l^2}{N^2} (m^2 + \frac{1}{4H^2})]^2\).

From equations 2.24 and 2.25, we can see that the direction of EP flux is parallel to the local group velocity. Therefore, patterns of the EP flux can be used to interpret wave propagation on the meridional plane (altitude versus latitude slice).

Fig 2.5 shows the meridional cross sections of EP flux and its divergence in January 1999. Above the tropopause (~12 km), the EP flux are associated with PWs as discussed above. Enhanced upward EP flux at middle latitude in the lower troposphere indicates intense poleward heat flux. Near the tropopause, the vectors veer toward the tropics, indicating the increasing poleward momentum flux. The EP flux convergence near 50km at about 70°N implies that the net effect of the eddies is to decelerate the mean flow. Since EP flux is related to group velocity, the arrow suggests that PW energy propagating in upward and equatorward direction above the tropopause.
2.2 Key Internal Variability in the Climate System

Climate variability refers to variations in the mean state and other statistics (e.g., standard deviations) of the climate on spatial and temporal scales beyond individual weather events (Change, 2007). Variability might be due to variations in natural and anthropogenic forcing external to the climate system (external variability) or natural internal processes within the climate system (internal variability). Natural external forcing, such as changes in volcanic eruptions and solar variability, contributes to the total natural variability of the climate system. Changing concentrations of greenhouse gases and aerosols caused by human activities are examples of anthropogenic external forcing. Excluding external forcing, internal variability may due to interactions within the atmosphere and coupling of the atmosphere with various components of the climate system (like the ocean). For our study, we will focus on key internal, low-frequency variability of subseasonal-to-

interannual time variations. Specifically, these are the Arctic Oscillation, Sudden Stratospheric Warming, Quasi-Biennial Oscillation, and Madden-Julian Oscillation.

2.2.1 Arctic Oscillation (AO)

Defined by the leading Empirical Orthogonal Function (EOF) pattern of the wintertime (November-April) sea level pressure anomalies in the domain poleward of 20°N latitude, the Arctic Oscillation (AO) accounts for 22% of the variance (Thompson & Wallace, 1998). A key AO feature is its high degree of zonal symmetry with a central anomaly over the Arctic regions and an opposing anomaly in the peripheral latitudes. Near the surface, AO is similar to the North Atlantic Oscillation (NAO), but with more zonal symmetry at high latitudes. Associated with the fluctuating sea level pressure (SLP) anomalies, the strong eastward circumpolar winds or the jet stream over the Pacific and Atlantic Ocean basin tends to shift north or south relative to their climatological position (Limpasuvan & Hartmann, 2000). As the jet stream tends to be collocated with storm systems, its meridional displacement alters the global weather pattern.

The near symmetric structure related to the AO appears in geopotential height variability well in the stratosphere during the winter. This pattern describes the changes of the stratospheric polar vortex or a planetary-scale eastward flow that encircles the winter pole in middle or high latitudes. Therefore, AO may be interpreted as the surface signature of modulations in the strength of the stratospheric polar vortex aloft (Fig 2.6). In its positive phase, the AO features negative geopotential height anomalies in the Arctic and positive anomalies in mid-latitudes. This pattern reverses during the negative AO phase along with the weakening of the stratospheric polar vortex and equatorward shift
of the jet stream, resulting in the displacement of cold Arctic air to lower latitudes and guiding storms to the Mediterranean region. In contrast, the stratospheric polar vortex is strong during a positive phase, which confines cold air and storms toward the north and causes drier in the Mediterranean.

Fig 2.6. Regression maps for 50hPa, 500hPa, and 1000hPa geopotential height based on AO index for 1947-1997. [From Thompson & Wallace, 1998].

Limpasuvan and Hartmann (2000) first pointed out that the zonal-mean zonal wind variations related to the AO are maintained by eddy fluxes in the free troposphere, while the Coriolis acceleration associated with the mean meridional circulation maintains the surface wind anomalies against frictional drag. In the Northern Hemisphere, stationary PWs provide most of the eddy momentum fluxes, although disturbances related to weather systems also make an important contribution. These authors also noted that during the negative AO phase, PWs tend to be refracted more toward the winter pole and, thereby, weakening the stratospheric polar vortex.

Baldwin and Dunkerton (1999) suggested that AO anomalies typically appear first in the stratosphere and propagate downward implying that stratospheric AO anomalies are precursors for tropospheric weather patterns. The stratospheric AO anomalies are initiated from the interaction between the stratospheric zonal mean flow and PWs
propagating from the troposphere into the polar stratosphere (Limpasuvan et al., 2004; Nishii et al., 2009; Garfinkel et al., 2010; Smith & Kushner, 2012). An example of wave-mean flow interaction is stratospheric sudden warming (SSW), which is described next.

2.2.2 Sudden Stratospheric Warming (SSW)

The wintertime stratosphere is characterized by a westerly and cold polar vortex, which is formed through radiative cooling and is characterized by a band of strong westerly winds at middle to high latitudes (Baldwin et al., 2021). When PW forcing becomes very intense, the stratospheric polar vortex can be weakened significantly. This weakening can eventually lead to the reversal of the zonal-mean zonal wind in the stratosphere (Matsuno, 1971; Andrews et al., 1987). Known as the sudden stratospheric warming (SSW), it was first detected in the 1950s when observations using radiosondes revealed that temperatures in the Northern Hemisphere wintertime stratosphere increased rapidly in several days (Scherhag, 1952). He referred it to as the “Berlin Phenomenon.” SSW events happen on average around once every two years in the Northern Hemisphere (Charlton & Polvani, 2007; Waugh et al., 2017). Fig 2.7 demonstrates the structure of one typical SSW event occurred on January 24th, 2009 as a time-altitude cross sections of the zonal-mean wind at 60°N. The effects of SSWs are not only in the middle stratosphere. They last much longer in the lower stratosphere and troposphere than in the upper stratosphere, which is due to the faster radiative time scale in the upper stratosphere (Baldwin et al., 2021).
For the mechanisms, numerous observational studies confirm that enhanced propagation of large scale PWs (of zonal wavenumber 1-2) from the troposphere is essential for SSW development (Matsuno, 1971; Andrews et al., 1987). As these waves dissipate according to equations 2.20 and 2.21, the resulting convergence of the PW EP flux ($\nabla \cdot F < 0$) decelerates the zonal-mean zonal wind and drives a poleward residual motion ($\bar{v}^+ > 0$) locally and residual descent ($\bar{w}^- < 0$) over the winter pole. The rapid adiabatic warming related to the polar descent helps maintain the thermal wind balance shown in (2.17). As shown in Limpasuvan et al. (2004), the zonal-mean zonal wind reversal leads to the formation of a critical layer for stationary PWs. Hence, subsequent PWs can no longer propagate beyond that level and dissipate leading to the gradual descent of the wind reversal toward the tropopause, as observed by Baldwin and Dunkerton (1999). As PW activity diminishes, radiative cooling in the polar stratosphere helps to reestablish the stratospheric polar vortex.

SSWs are often followed by a negative phase of the NAO and AO, which may persist for up to two months (Baldwin & Dunkerton, 2001), and are associated with colder weather.
over the northeast North America and northern Eurasia (Kolstad et al., 2010). However, the surface impact by SSW events is still under debate. Sigmond et al. (2013) showed that only 13 out of 20 major SSW events (65%) were followed by a negative AO after the onset. It has been suggested that the type of the wave forcing (wavenumber 1 or 2) into the stratosphere (Nakagawa & Yamazaki, 2006), the type of SSW events (resulting in the polar vortex being split or displaced off the Pole) (Mitchell et al., 2013; Seviour et al., 2016), and the persistence of the initial circulation anomaly in the lower stratosphere (Maycock & Hitchcock, 2015) prior to the onset could influence the downward propagation of SSW anomalies into the troposphere. In contrast, Charlton & Polvani (2007) did not find a statistically significant difference between tropospheric impacts by split and displacement SSW. Recently, Kodera et al. (2016) proposed the classification of SSW events depending on whether or not they are followed by a reflection of planetary waves in the stratosphere. They suggested that only the absorptive events are followed by the response of the tropospheric circulation via the AO. A better understanding of SSW dynamics is necessary for the surface weather prediction.

2.2.3 Quasi-Biennial Oscillation (QBO)

The Quasi-Biennial Oscillation (QBO) dominates the variability of the equatorial, low to middle stratosphere. It manifests as a vacillation in the zonal-mean zonal wind in the equatorial band with a variable period averaging approximately 28 months (Baldwin et al., 2001). Fig 2.8 demonstrates the QBO as a time-altitude cross sections of the zonal-mean wind averaged between 5°S and 5°N across the Equator. The westerly (easterly) QBO phase occurs during the time period when the zonal wind is eastward (westward). The
alternating wind regions develop at the top of the lower stratosphere and propagate downward at the rate of about 1 km per month until they are dissipated at the tropical tropopause (Lindzen & Holton, 1968). Similarly, QBO signature is present in the corresponding temperature field in the equatorial band.

![Fig 2.8. Time-height plot of monthly mean, zonal mean equatorial zonal wind in m s⁻¹ between about 20 and 35 km altitude above sea level over a ten-year period. Positive values denote eastward winds and the contour line is at 0 m s⁻¹. [From Morn. Graph made using data from FU Berlin].](image)

The QBO is another example of wave and mean-flow interaction (like SSW). It is thought that large-scale Kelvin and mixed Rossby-gravity waves drive the QBO. A detailed discussion of these equatorial waves is beyond the scope of this thesis. Wallace and Kousky (1968) showed that Kelvin waves produced an upward flux of eastward momentum, which could account for the eastward acceleration associated with the QBO. Bretherton (1969) found that a net westward acceleration of QBO is contributed by Rossby-gravity waves. However, observations imply that mixed Rossby-gravity and Kelvin waves cannot provide sufficient forcing to drive the QBO with the observed period. Dunkerton (1997) suggested that additional momentum flux is supplied by a broad spectrum of gravity waves to force a realistic QBO.
Although QBO is an equatorial phenomenon, it could affect the polar stratospheric flow through modulation of PW propagation. Observational studies suggest a stronger stratospheric polar vortex during the westerly QBO and a more disturbed stratospheric polar vortex during easterly phase. Holton and Tan (1980, 1982) proposed a mechanism relating PW propagation to the QBO. During the easterly QBO phase, the line of zero zonal-mean zonal wind extends to the subtropics of the winter hemisphere. This line serves as the critical surface for stationary PWs. Therefore, the meridional extent of the waveguide is narrowed and stationary PWs tends to be break or dissipate closer to the higher winter latitudes as they are refracted toward the critical surface. The dissipation of large PW amplitudes could lead to a weaker stratospheric polar vortex and a warmer polar stratosphere. In extreme cases, the forcing from PWs could lead to major SSWs. When QBO phase is westerly, PW propagation toward the equatorial region is less restricted as the waveguide widens. PW disturbance in the high winter latitudes can be potentially less, resulting in a more stable stratospheric polar vortex. Thus, fewer SSWs are expected during the westerly QBO.

2.2.4 Madden-Julian Oscillation (MJO)

The Madden-Julian Oscillation (MJO) is the dominant component of atmospheric intraseasonal variability in the tropics. It is characterized by eastward propagation (around 5 m s\(^{-1}\)) of tropical convection from the Indian Ocean to the central Pacific and has a recurrence period of 30-60 days (Madden & Julian, 1971, 1972) (Fig 2.9). The MJO is divided into eight phases as the oscillation propagates from the western Indian Ocean through the Maritime continent, and into the Pacific Ocean (Wheeler & Hendon, 2004).
At phase 3 (7), the convection is located in the Indian Ocean (Western Pacific Ocean). Diabatic heating anomalies induced by the MJO could excite Rossby wave trains, which propagate across the North Pacific, North America, and the Atlantic Ocean (Matthews et al., 2004). Therefore, MJO could impact some modes of climate variability in winter, such as the Pacific North America pattern (PNA; Seo & Lee, 2017), the NAO (Lin et al., 2009; Barnes et al., 2019), and the AO (Zhou & Miller, 2005; L’Heutreux & Higgins, 2008). Besides the tropospheric pathway, it can impact the tropospheric circulation via the stratosphere. Over the North Pacific, the induced Rossby waves by the MJO propagate upward into the stratosphere, which break down the stratospheric polar vortex and lead to the occurrence of SSWs (Garfinkel et al. 2012, 2014). It has been found that the MJO can impact the evolution feature of SSWs, such as type (Liu et al., 2014) and frequency (Kang & Tziperman, 2017). The anomalous signal of the stratosphere propagates downward and leads to a negative phase of the AO.
Multiple factors could modulate MJO, such as QBO (Yoo and Son, 2016; Hendon & Abhik, 2018; Densmore et al., 2019; Hood et al., 2020; Toms et al., 2020), El Niño-Southern Oscillation (ENSO; Son et al., 2017), 11-year solar cycle (Hood, 2017), and SSW (Wang et al., 2020). QBO can explain up to 40% of the interannual variation of the MJO during the
boreal winter (Son et al., 2017). During the easterly QBO (EQBO), the tropopause is higher and colder than during the westerly QBO (WQBO). Furthermore, the vertical zonal wind shear across the tropopause is weaker, and upper-tropospheric static stability is lower (Baldwin & Dunkerton, 2001). These contrasts are considered as the reasons for observed stronger tropospheric deep convection in EQBO than WQBO (Collimore et al., 2003). Son et al. (2017) found a stronger MJO and more pronounced MJO-induced teleconnections during EQBO winter. The MJO in boreal winter is observed to be more active and propagates farther eastward during EQBO (Hendon & Abhik, 2018; Hood, 2020). Since the background flow is a waveguide of Rossby waves, the QBO might impact the MJO teleconnections through the modulation of the extratropical background state (Toms et al., 2020). Recently, using reanalysis data, Song & Wu (2020) revealed that the MJO phase 6-7 and AO connection is modulated by the QBO, whereas the MJO phase 2-3 and AO relationship is favored in both QBO phases. Besides the QBO, 11-year solar cycle, ENSO and SSW could also impact the MJO. Hood (2017) found that the minimum of the 11-year solar cycle favors a strengthened wave train that propagates farther eastward. The MJO activity tends to extend further eastward to the Date Line during El Niño winters and contract toward the western Pacific during La Niña winters (Son et al., 2017). The occurrences of MJO phases 6 and 7 significantly increase in 20 days after the onset of SSWs (Wang et al., 2020).
3. Current Knowledge of Sea Ice Loss Impact on Atmosphere and Motivation

Over the past few decades, the Arctic region has warmed more than twice as fast as the global average (Serreze et al., 2009; Screen & Simmonds, 2010; Cowtan and Way, 2014). Referred to as the Arctic Amplification (AA), this unusual warming is evident in the lower-tropospheric temperatures along with the increased in atmospheric column thickness between the 1000 hPa and 500 hPa levels (Francis and Vavrus, 2012). Although occurring in all seasons, AA is stronger in autumn and winter. Based on current observed trend, the AA is expected to be stronger in the future, invoking changes in atmospheric circulation, vegetation and the carbon cycle with impacts in and beyond the Arctic regions (Serreze & Barry, 2011).

The rapid Arctic warming is accompanied by extensive loss of sea ice (e.g., Screen & Simmonds, 2010). Sea ice is frozen seawater that covers much of the Arctic and Antarctic Oceans and neighboring seas. Sea ice consists of two components: perennial ice and seasonal ice. Lasting more than 9 years, perennial ice can grow up to 4 m in thickness. Seasonal ice thickens and spreads during the fall and winter, then thins and shrinks during the spring and summer. The seasonal sea ice thickness is at most 2 m. Perennial sea ice generally contains less salt and is much stronger than seasonal sea ice (Polyak et al., 2010). Consistent with the observed Arctic warming, the extent of sea ice coverage over the Arctic Ocean have shown a dramatic decline over the past 30 years as illustrated in Fig 3.1. Observations also show a declining trend in sea ice concentration and increasing
trend in length of melting season (Vihma, 2014). Overland et al. (2014) estimated that the Arctic Ocean might be free of summer sea ice before 2050 if human activities continue to raise greenhouse gas concentrations.

The Arctic sea ice is a critical component of the climate system (Alexander et al., 2004). In addition to being a sensitive indicator of climate change, the Arctic sea ice has feedback effects on the other components of the climate system through the reduction in the exchanges of heat, momentum, water vapor, and other material between atmosphere and ocean and providing higher surface albedo than that of the open ocean (Vihma, 2014). Changes in the surface albedo, heat fluxes and near-surface temperature gradient related to sea ice loss are expected to have numerous consequences on climate patterns.

Fig 3.1. Arctic sea ice extent on September 23, 2018 and 1981-2010 averaged (top), and yearly Arctic sea ice area variations from 1979 to 2017 (bottom) as reported by NASA (National Aeronautics and Space Administration) Goddard Space Flight Center and National Snow and Ice Data Center (NSIDC).
Understanding the impact of sea ice changes on the atmospheric circulation is therefore crucial for predicting and assessing climate changes in the coming decades as well as extreme weather (Screen et al., 2015; Francis et al, 2017; Cohen et al., 2018).

In this chapter, we briefly review the current knowledge on the effects of Arctic sea ice loss on the atmosphere. These effects can directly impact the Arctic region and extend to influence the mid-latitudes. This review will allude to the background on Rossby waves and natural variability, discussed in Chapter 2.

3.1 Direct Impact of Sea Ice Loss on the Atmosphere: Local Effect

Local effects of Arctic sea ice loss occur over the Arctic Ocean and its marginal seas. With respect to near-surface AA warming, Serreze and Barry (2011) suggested the positive feedback related to sea ice loss in warming the atmosphere (Fig 3.2). Consider a climate forcing that warms the Arctic and melts sea ice. During the summer season, ice-free condition may appear in some regions and more solar radiation could be absorbed by open water. When summer ends and the sun set over the Arctic Ocean, large amount of heat will be transferred from the exposed ocean to the overlying atmosphere, via longwave (or infrared) radiation, latent heat flux (related to phase change), and sensible heat flux. Subsequently, the Arctic air becomes unusually warm. With diminishing solar radiation during autumn and winter, the ocean eventually loses heat gained during the summer and seasonal sea ice forms. Since it takes time for the ocean to lose its sensible heat, the formed sea ice in spring will be thinner (and weaker) than before and melt more readily in the following summer. This then leads to a positive feedback loop that further warms the Arctic air.
The anomalously large open water areas in September have resulted in a strong transfer of heat from the ocean’s mixed layer to the atmosphere, which could warm the overlying air (Serreze et al., 2009; Screen & Simmonds, 2010). Stroeve et al. (2012) observed that positive air temperature anomalies appear in the month (October) after the seasonal sea ice minimum (September). The increased heat flux from the ocean to the atmosphere can reduce the vertical static stability (Francis et al., 2009; Overland and Wang, 2010; Stroeve et al., 2011), which can make the atmosphere more baroclinically unstable (Jaiser et al., 2012) leading to the generation of atmospheric waves.

High-resolution models (like numerical weather prediction and large-eddy simulation models) have simulated enhanced convection over leads, polynyas, and the open ocean (Vihma, 2014) and point to the potential effects of these surface features on the atmospheric boundary layer (ABL) (e.g., Ledley, 1988; Vihma, 1995; Esau, 2007; Lüpkes et al., 2008, 2012; Ebner et al., 2011). Large-scale, lower resolution, climate models demonstrate that the atmospheric response to the sea ice loss is consistent with AA
Collectively, these numerical results illustrated that the impact of Arctic sea ice loss on the local atmospheric response is manifested in the enhancement of turbulent fluxes of sensible and latent heat, the increase in ABL height, air temperature, specific humidity, relative humidity, cloud cover, and precipitation, and the decrease in atmospheric stratification.

### 3.2 Global Impact of Sea Ice Loss: Remote Effect

Since the Arctic sea ice plays an important role in regulating the heat exchange between the ocean and the atmosphere, the Arctic sea ice loss could impact atmospheric circulation, not only in the marine Arctic, but also in the tropospheric midlatitudes as well as to the stratosphere. The remote effects are indirect and, therefore, much more complicated than local ones. In particular, the extended influence of Arctic sea ice loss appears to be intimately tied to the aforementioned AA warming, especially near the surface. While multiple intertwining factors on different spatiotemporal scales (e.g., cloud cover, water vapor, black carbon aerosols concentration, snow cover, and sea ice cover) may contribute to AA, the diminishing Arctic sea ice may be the leading role in recent AA (e.g., Dai et al., 2019).

The Arctic warming due to sea ice loss can have two large-scale effects that extend to the tropospheric mid-latitudes. First, the polar warming can reduce the climatological north-south atmospheric temperature gradient and thereby weaken the upper tropospheric prevailing eastward flow (i.e., the jet stream). Francis and Vavrus (2015) found robust relationships among seasonal and regional patterns of weaker poleward thickness
(atmospheric column) gradients, weaker upper-level zonal wind, and a more prevalence of the jet stream in the meridional direction. Second, the regional warming provides anomalous surface thermal forcing over the Arctic region that may excite stationary Rossby wave that interfere with the attendant climatological stationary wave.

Atmospheric Rossby waves over northern midlatitudes are proposed as a possible mechanism linking AA and midlatitudes extreme weathers (Screen & Simmonds, 2013a; 2013b; 2014; Cohen et al., 2014). In combination, the weakened (and meandering) eastward jet and altered stationary wave structure due to Arctic warming may induce a slower eastward advection of synoptic-scale Rossby wave propagation (as discussed in sub-section 2.1.3). As these synoptic-scale waves in the upper tropospheric level tend to organize weather systems, this unusually slow and variable jet can lead to more persistent weather systems and high probability of extreme weather events, such as drought, flooding, cold spells, and heat waves (Petoukhov et al. & Semenov, 2010; Francis & Vavrus, 2012; Liu et al., 2012; Tang et al., 2013; Cohen et al., 2014; Francis & Vavrus, 2015).

Persistent weather systems can be associated with atmospheric blocking (Berggren et al., 1949; Rex, 1950). Blocking is a regional, highly meandering state of the jet stream lasting between a few days to a week which may lead to weather extremes. The theoretical underpinning of blocking has been based traditionally on the idea of global quasi-resonant amplification in which a large-scale circulation, in the presence of topography, give rise to two equilibrium states: one with a typical zonal flow and the other with blocking (Charney & DeVore, 1979; Randall, 2015). Recently, Nakamura and Huang (2018) proposed a more localized view of blocking based on the theory of finite amplitude of
Rossby wave activity. Initially growing localized wave activity slows down the jet stream. This deceleration lowers the jet’s capacity to carry wave activity downstream. Subsequent wave flux from upstream begins to accumulate where the jet stream begins to stall, leading eventually to wave breaking and blocking. This theory is analogous to traffic congestion during which the influx of cars approaching a road with high-traffic volume or a slower speed limit. Hence, the weakened jet stream related to AA and sea ice loss may lead to more wave flux congestion and, therefore, blocking.

Current understanding suggests that AA associated with the declining summer sea ice extent may alter the atmospheric circulation and impact weather in the following autumn and winter seasons (e.g., Francis & Vavrus, 2015). This appears consistent with the high heat capacity of the open ocean when exposed due to sea ice loss during summer and autumn and observed positive air temperature anomalies in the month following the seasonal sea ice minimum in September (Stroeve et al., 2012). Using satellite and reanalysis data of sea ice and terrestrial snow areas, Liu et al. (2012) noted that the decreased Arctic sea ice area in autumn is linked to changes in the winter large-scale circulation resembling the negative surface AO phase (discussed in Chapter 2). However, Vihma (2014) suggested that the observed wintertime atmospheric patterns may be a combination of lagged responses to the sea ice decline in the previous seasons and fast responses to the sea ice decline in winter.

The impact of Arctic sea ice can extend into the stratosphere. Using reanalysis data, Jaiser et al. (2013) demonstrated the relationship between large-scale atmospheric circulation (throughout the troposphere and stratosphere) and sea ice loss in Chukchi Seas where
the seasonal ice extent decline is most pronounced (see Fig 3.1, top). The leading coupled response between the regional ice loss and the atmosphere consists of anomalous atmospheric pattern akin to AO in the troposphere and stratosphere. In particular, the first maximum covariance analysis (MCA) patterns describe the diminishing sea ice over the Siberian and Beaufort Sea in August/September (Fig 3.3a) with the 500-hPa anomalous geopotential pattern, which resembles the negative AO phase (Fig 3.3b). The remote effects extend barotropically into the stratosphere (Fig 3.3f).

*Fig 3.3. First (a, b and e, f) and second pair (c, d and g, h) of coupled patterns obtained by the MCA of August/September HadISST1 sea ice concentration (1979-2011) and ERA-Interim winter (DJF) geopotential height fields (1980-2012). Upper row displays the sea ice concentration anomaly maps in percent as heterogeneous regression maps. Lower row contains the corresponding anomaly maps for geopotential heights in 500 hPa (b, d) and 10 hPa (f, h) in gpm, shown as homogeneous regression maps. Black contours in (b) and (f) show the climatological mean (1980-2012) with 5100 gpm, 5200 gpm, and 5300 gpm isolines in (b) and 29000 gpm, 30000 gpm, and 30500 gpm isolines in (f) starting from the pole. Black contours in (d) and (h) show the climatological departures from the zonal mean (1989-2012) with 50 gpm contour interval in (d) and 150 gpm contour interval in (h). Negative contours are dashed. [From Jaiser et al., 2013].*
To date, evidence for Rossby waves as a link between AA, sea ice loss, and midlatitudes extreme weathers remains unclear (Barnes, 2013; McCusker et al., 2016; Francis, 2017). Screen and Simmonds (2013a) highlighted that the possible connections between AA and Rossby waves are sensitive to how wave amplitudes were conceptualized as meridional amplitudes (which measure of north-south meandering of the jet) or as zonal amplitudes (which measure the intensity of atmospheric pressure). Characterizing the changes in frequency and magnitude of blocking events due to climate change is likewise difficult (Francis et al., 2017). Since natural variabilities (like QBO, MJO) might blur the atmospheric response due to AA and sea ice loss, distinguishing the role of the natural variabilities from the atmospheric response to sea ice loss is necessary. With recent advances in climate model, more relevant studies relied on numerical simulations.

3.3 Simulated Atmospheric Response to Arctic Sea Ice Loss: Divergent Model Results

Although numerous observational studies exist, Hopsch et al. (2012) concluded that the connection between sea ice and atmosphere could not be considered statistically robust at present due to short observational time, which started in 1979 with satellite observations. Cohen et al. (2014) suggested that improved process understanding, additional Arctic observations, and better coordinated modeling studies will be needed to understand the impact of AA and sea ice loss on extreme weather events. Since model simulations can isolate the physical processes underlying the relationships found in observations, recent studies increasingly utilize model simulations to investigate the impact of sea ice loss on the large-scale circulation.

However, these simulations tend to disagree on the response of the AO to sea ice loss
Recall that the AO characterizes much of wintertime variance over Europe, North America, and parts of Asia. Some modeling studies concluded a negative AO response (Honda et al., 2009; Seierstad & Bader, 2009; Kim et al., 2014; Nakamura et al., 2015) while others a positive AO response (Strey et al., 2010; Orsolini et al., 2012; Rinke et al., 2013; Screen et al., 2014). Others found little to no response (Screen et al., 2013; Mori et al., 2014; Petrie et al., 2015; Blackport and Kushner, 2016). How the AO response may depend on the details of the forcing (Alexander et al., 2004; Petoukhov & Semenov, 2010; Peings & Magnusdottir, 2014; Pedersen et al., 2016).

Smith et al. (2017) suggested that the background state, impacted by different SSTs between atmosphere-only simulation and ocean-atmosphere coupling simulation, controls the sign of the winter AO response via the refraction of Rossby waves. The importance of ocean-atmosphere coupling for simulating the atmospheric response to sea ice has also been stressed in previous studies (e.g., Deser et al., 2015, 2016; Tomas et al., 2016). The coupling generally enables surface temperature responses to spread to the ocean. In absence of ocean coupling, the atmospheric response to Arctic sea ice loss is confined poleward of 30°N (Deser et al., 2015). The ocean-atmosphere coupling can magnify the response the Arctic sea ice loss but does not change its overall response structure (Deser et al., 2016). However, ocean dynamics can mediate the global climate response to the Arctic sea ice loss (Tomas et al., 2016). In particular, models with a full-depth ocean show that the freshening of the subpolar Arctic due to sea ice melting reduces the downwelling associated with the Atlantic Meridional Overturning Circulation (AMOC). By continuity, this weakening of the AMOC gyre decreases equatorial ocean
upwelling and promotes an anomalously warm equatorial SST through enhanced atmospheric deep convection and latent heat release. The resulting tropical upper tropospheric warming counters the reduced meridional gradient in atmospheric thickness due to the Arctic warming, “tugging” the jet back to the high latitude (Screen et al., 2018). The location of sea ice loss occurrence may also lead to different response. As suggested by Screen (2017), sea ice loss in some region could trigger large-scale dynamical responses but only induce local changes thermodynamically in other regions. Numerous studies found that sea ice anomalies over the Barents-Kara Seas in summer/autumn elicit large-scale atmospheric circulation that leads to a cold winter in Europe, North America, and parts of Asia via PWs (Honda et al., 2009; Petoukhov & Semenov, 2010; Kim et al., 2014; Mori et al., 2014; Kretschmer et al., 2016; Zhang et al., 2016). However, recent studies suggest that the warm Arctic and cold Asian pattern observed in recent winters may be due to natural internal variability (McCusker et al., 2016; Sun et al., 2016; Seviour et al., 2017; Ogawa et al., 2018; Blackport & Screen, 2020; Guan et al., 2020; Kretschmer et al., 2020) or covarying sea surface temperature (SST) forcing outside of the Arctic (Blackport & Screen, 2021).

The inclusion of the stratosphere in these models may also influence the atmospheric response to sea ice loss. As discussed in Chapter 2 and suggested by Fig 3.3, the AO pattern projects into the stratosphere. Furthermore, PWs generated with response to Arctic warming can penetrate into the stratosphere. There, the presence of PW activity in the polar stratosphere or lack thereof can influence the stratospheric polar vortex strength and steer the AO toward a particular phase (e.g., Limpasuvan et al., 2004). Using
an atmosphere only climate model with a well-resolved stratosphere, Zhang et al. (2018) found a robust cold Siberian pattern in the winter following sea ice loss over the Barents-Kara seas in late autumn.

The stratospheric circulation and its coupling to the AO phase response may be sensitive to the geographical location of sea ice loss (Screen, 2017; McKenna et al. 2018). Taken from Sun et al. (2015), Fig 3.4 demonstrates the opposing response to sea ice loss in the Atlantic sector versus that in the Pacific sector. Here, the Atlantic sector is defined as inside Arctic Circle (66.6°N) and Pacific sector outside. When considered together, the stratospheric response may cancel one another and explain the weakly negative and statistically insignificant response of the stratospheric polar vortex found in other studies. From Fig 3.4b, the sea ice loss in Pacific sector induces a strengthened stratospheric polar vortex, which contradicts with observations of Jaiser et al. (2013). The conclusion from Sun et al. (2015) may depend on how they define Atlantic and Pacific sectors. As the interaction between responses to regional sea ice losses is nonlinear (Screen, 2017), this cancelling effect proposed by Sun et al. (2015) may not work for other situations.
Fig 3.4. Winter (DJF) zonal-mean zonal wind response (shading) in (a) Atlantic sector and (b) Pacific sector, superimposed on the climatology from the control run (contours; contour interval of 5 m s\(^{-1}\)). Stippling indicates that the response is statistically significant at the 95% confidence level. [From Sun et al., 2015].

Using an atmosphere-only climate model, McKenna et al. (2018) also found the opposite stratospheric polar vortex response between sea ice loss in the Atlantic sector (Barents and Kara Seas) and Pacific sector (Chukchi and Bering Seas). Although using a different definition of the Atlantic and Pacific sectors, these authors confirm the results of Sun et al. (2015) but refute those of Jaiser et al. (2013). In the Atlantic (Pacific) case, the stratospheric polar vortex weakens (strengthens) in November-February (November-January) due to enhanced (suppressed) upward Rossby wave propagation into the stratosphere (Fig 3.5).
These authors also addressed the effects of sea ice loss magnitude. For large magnitude sea ice loss in the Pacific sector, the tropospheric response agrees with Jaiser et al. (2012, 2013) with negative AO response (Fig 3.5b, d). As the magnitude of sea ice loss increases, the tropospheric mechanisms become relatively more important than the stratospheric mechanisms. More recently, a numerical study (Ding et al., 2021) suggested that the sea ice loss over the Pacific sector is responsible for a weakened polar vortex, which agrees with Jaiser et al. (2012, 2013) and the large magnitude sea ice loss result in McKenna et al. (2018). We note that the experimental set up of McKenna et al. (2018) cast doubts on their conclusion. Instead of forcing their model with prescribed sea ice data, they parameterized sea ice loss with SST anomaly fields. Such indirect anomalous field in the boundary may induce unforeseen atmospheric response and make comparison with other studies difficult.

Upward wave flux could be strong enough to cause the demise of the polar vortex associated with the SSW phenomenon. Several studies have reported that SSW events are related with the recent Arctic sea ice loss (Hoshi et al., 2019; Zuev & Savelieva, 2019; Zhang et al., 2020). In analyzing reanalysis data, Zuev and Savelieva (2019) argued that autumnal Arctic sea ice loss in recent years could be a forcing factor for SSW (Fig 3.6). The small simple size of SSW events and short observational record however raises doubts on their conclusion. Hoshi et al. (2019) concluded that Arctic sea ice loss over the Barents-Kara Sea is a possible factor modulating the wave propagation during the weak polar
vortex events using reanalysis. The observed features are supported by an Atmospheric General Circulation Model (AGCM) experiment. In the presence of low Barents-Kara Sea sea ice, SSW can favor surface cold spells over North America based on observations and model experiments (Zhang et al., 2020).

Finally, the background state of the ocean may contribute to the divergent model results of the large-scale atmospheric response to Arctic sea ice loss. SST is a function of near-surface atmospheric fluxes of momentum and heat, the ocean’s response to those fluxes, and the underlying ocean dynamics. The large thermal capacity and the slowly evolving oceanic currents make the SST vary on timescales that are much longer than those of atmosphere. SST variation is thus often considered a prime candidate for inciting low frequency atmospheric changes (Lau, 1997). Consequently, these atmospheric changes can modulate the atmospheric response to sea ice loss. Osborne et al. (2017) tested the sensitivity of atmospheric response to Arctic sea ice loss to the phase of the Atlantic

Fig 3.6. Time series of monthly mean sea ice extent over the Beaufort Sea, the Canadian Arctic Archipelago and the Central Arctic (in total) for September, October and November from 1979 to 2018 (left); Time series of zonal mean zonal wind at 60°N 10hPa from November to March of 1984/1985 (red line), 1998/1999 (blue line) and 2012/2013 (brown line) in comparison with the 1979-2018 climatological means (green line) with ±1 standard deviations (right). [From Zuev & Savelieva, 2019].
Multidecadal Oscillation (AMO) with anomalous SST pattern varying on a multi-decadal time. They found that the modulation of the atmospheric response occurs prominently during the negative AMO phase and largely confined over the Pacific-North American sector with the southward shift of the North Pacific jet. Their study may provide an explanation to the ongoing debate about the relationship between Arctic sea ice and the AO.

Peings et al. (2017) found that the contribution of Siberian snow anomalies in fall to the wintertime extratropical atmospheric circulation seems to be dependent on the phase of the QBO. Labe et al. (2019) suggested that the stratosphere-troposphere pathway from sea ice loss is sensitive to the QBO phase. These two studies imply that different QBO phase may cause a different atmospheric response to same pattern and magnitude of sea ice loss.

The MJO could impact some modes of climate variability in winter (Zhou & Miller, 2005; L’Heutreux & Higgins, 2008; Lin et al., 2009; Seo & Lee, 2017; Barnes et al., 2019) (details in sub-section 2.2.4). Over the North Pacific, the induced Rossby waves by the MJO propagates upward into the stratosphere and leads to the occurrence of SSWs (Garfinkel et al. 2012, 2014). However, whether MJO modulating the atmospheric response to sea ice loss or not is still unknown.
4. Research Goal and Objectives

The Arctic sea ice loss and accompanying Arctic warming potentially have far-reaching effects on the large-scale atmosphere circulation in the troposphere and stratosphere. Observational data record period is still too short to provide statistically convincing conclusion on the how the underlying mechanism works in generating a global atmospheric response. Being more recent, observations also contain signatures of anthropogenic influence and natural variability of different spatiotemporal scales. This contamination makes understanding the underlying mechanism difficult.

Climate model simulations can potentially isolate the physical processes underlying the relationships found in observations. However, model results often disagree depending on experimental setup (e.g., Screen et al, 2018). As highlighted in Chapter 3, key sources for model disagreements are the magnitude and spatial pattern of the Arctic sea ice loss, the interaction of the troposphere with the overlying stratosphere and underlying ocean, and the background state that co-exists with the atmospheric response. Other possible sources include model treatment of varying sea ice thickness, SST as sea ice disappears, and inherent physics (e.g., microphysics, boundary layer) not resolved in the model.

Even with these difficulties, we see glaring contradictions and deficiencies worthy of examination. Observations reveal that the most sea ice retreat occurs in the Bering-Chukchi (Pacific) region (see Fig 3.1) and reanalysis results suggest a strong coupling between this sea ice loss region and a weakened stratospheric polar vortex (see Fig 3.3).
This large-scale response is confirmed by a numerical study by Ding et al. (2021) but is inconsistent with other model results (e.g., McKenna et al., 2018). In addition, several studies have reported that SSW events are related with the recent Arctic sea ice loss (Hoshi et al., 2019; Zuev & Savelieva, 2019; Zhang et al., 2020). However, they mainly focus on the Kara-Barents (Atlantic) region. The impact of the declining autumnal sea ice over Pacific sector on the stratospheric circulation (especially for SSWs) is still unknown and needs investigation.

While the stratosphere is evidently a key part of the atmospheric response, the role of the QBO in the literature related to sea ice loss is almost never considered, with the exception of Labe et al. (2019). Since the vertical propagation of PWs couples the troposphere to the stratosphere, the QBO phase can alter that propagation and the waves’ impact on the stratospheric polar vortex. A strongly perturbed vortex can then influence the troposphere and may be considered as the secondary effects of sea ice loss. In general, the QBO’s role of in modulating the atmospheric response to Arctic sea ice loss and how that portends to the large-scale atmospheric circulation have not received much attention.

Previous studies have suggested that MJO interacts with SSW characteristics (Liu et al., 2014; Kang & Tziperman, 2017; Wang et al., 2020) as well as QBO (Yoo and Son, 2016; Son et al., 2017; Hendon & Abhik, 2018; Song & Wu, 2020). However, in the literature, the role of MJO related to atmospheric response to sea ice loss is never considered. Overall, the gaps in understanding the detailed mechanisms associating the surface processes and their subsequent impact on the stratosphere contributes to the
uncertainties in distinguishing the potential sea ice-stratosphere connection from internal variability and other external forcings.

To this end, the overarching goal of this thesis is to improve our basic understanding of the physical processes that affect the large-scale atmospheric circulation in response to Arctic sea ice loss in the Pacific sector and whether internal variabilities (associated with the QBO and MJO) modulate the response. The thesis’ pertinent questions are: What is the sole impact of sea ice loss over the Pacific sector on the SSWs? How does the QBO modulate the response to sea ice loss? Does the MJO play a role in the response? Related to these questions is how the SSWs change in response to sea ice loss and how the changes in SSW characteristics ultimately affect surface conditions.

Answers to these questions may shed light on the inconsistencies between observations and model results pertaining to regional sea ice loss. They can also provide new insights on the fundamental coupling between the troposphere and the stratosphere during AA and internal varying factors such as the QBO and MJO that modulate that coupling. To address this goal, our research objectives are:

- To assess the dynamical response of SSWs exclusively to seasonal sea ice loss in the Pacific sector, and
- To quantify the impact of the QBO phase on the dynamical response, and
- To examine the role of MJO on the dynamical response.

To meet these objectives, we will utilize the atmospheric component of the latest state-of-the-art chemistry-climate model with a well-resolved stratosphere (released late fall 2018). In using an atmosphere-only model, the sea ice and SST fields can be prescribed in
a controlled fashion. To minimize the sources of numerical disagreements, we keep the prescribed forcing as simple as possible (nearly idealized) and control the background state and its associated atmospheric variability that might mask the direct response of the atmosphere to Arctic warming. Furthermore, the roles of solar influence will be fixed at the solar minimum condition and the impact of global warming eliminated.

As the summer/autumn Arctic sea ice loss is observed to elicit atmospheric response in the following winter, we forego long-term model simulations like most previous studies and focus on ensembles of short 1-year runs to focus on the seasonal atmospheric response and to build a statistically robust result. Over that 1-year simulation time span, the seasonal sea ice loss begins to recover and the winter polar stratosphere can experience strong wave perturbations leading possibly to SSWs. Yet, this time span is short enough for the preexisting background conditions related to QBO, solar cycle to remain nearly unchanged.
5. Methods

This chapter describes the methods used throughout this thesis. To address our goal and meet our research objectives, the study will conduct several numerical experiments using a global chemistry-climate model called the Whole Atmosphere Community Climate Model (WACCM), developed at the National Center for Atmospheric Research (NCAR). To affirm model results, we also analyze a reanalysis dataset. In addition, the data analysis methods used in the results are detailed including Empirical Orthogonal Function (EOF) analysis, wave activities diagnostics, sudden stratospheric warming (SSW) definition, statistical tests for statistical significance and detrend reanalysis data.

5.1. Numerical model

5.1.1 Description

WACCM is part of the NCAR Community Earth System Model (CESM) framework, comprising of land, ice, ocean and atmosphere components. Configured as an atmosphere-only component, WACCM is a fully coupled chemistry–climate model with vertical levels spanning from the surface to the lower thermosphere (~150km). WACCM has been used independently to study dynamic variability and the distribution of minor species in the stratosphere and mesosphere (Marsh et al., 2013). A fully resolved stratosphere improves the representation of stratospheric variability. As noted in Marsh et al. (2013), WACCM simulation results have produced key stratospheric behavior similar to observations (e.g., the occurrence frequency of the SSW phenomenon and the ozone
hole development).

The high-altitude range of WACCM effectively removes the artificial impact of the upper boundary condition as we study the response of the entire stratospheric to sea ice loss. Fig 5.1 shows the energy transfer in the mesosphere and lower thermosphere (MLT). About $10^{16}$J of energy propagates up daily from the atmosphere below in the form of waves and tides. A model with an upper lid below the MLT may reflect this energy downward or may require unwarranted artificial damping. However, we see that around $10^{17}$J is injected per day from space through solar activity and auroral processes during a geomagnetic storm which occurs about every 5 days (Jarvis, 2001). In using WACCM, we have to constrain the parameterized solar impact from above 150 km (as discussed below).

Fig 5.1. Energy transfer in the mesosphere and lower thermosphere. [From Jarvis, 2001]

At the time of this writing, the publicly supported version of WACCM is WACCM6, which
is part of CESM Version 2.1.0. With a 0.95° latitude by 1.25° longitude horizontal resolution, WACCM6 is subset of the Community Atmosphere Model version 6 (CAM6), and includes all of the CAM6 physical parameterizations. The chemistry processes in WACCM6 includes most species in the troposphere, stratosphere, mesosphere, and lower thermosphere (TSMLT). Through its Modal Aerosol Module (MAM4), the model also provides a prognostic representation of stratospheric aerosols from volcanic and non-volcanic source gases.

With a default horizontal resolution 4 times finer than its predecessor (i.e., WACCM4 of CESM1), WACCM6 has improved stratospheric variability including an internally generated QBO and an improved climatology of SSWs. Solar proton events, gravity wave (GW) drag deposition from vertically propagating GWs (generated by orography, fronts, and convection), and molecular diffusion are all contained in this latest version. Therefore, WACCM6 can simulate well the coupling between atmospheric layers, such as waves that transport energy and momentum from the lower atmosphere.

Three configurations are available to run WACCM6: free-running, specified dynamics, and specified chemistry. The free-running version contains interactions between chemistry and dynamics as predicted by the numerical integration. The specified chemistry and specified dynamics versions rely on observations to constrain the prognostic values by imposing small adjustments (or nudging) to the predicted dynamical fields. Since specified chemistry version could reduce the computational cost, some studies mentioned before recommended it (i.e., Zhang et al., 2018). However, the lack of interactive ozone in the specified chemistry version could lead to strong temperature difference in the upper
stratosphere and lower mesosphere, below the 65 km transition level where the chemistry is specified (Smith et al., 2014). Since we focus on the dynamical mechanism between sea ice loss and atmospheric circulation which intertwines with the chemistry, the free-running configuration will be used to avoid potential problems related to specified chemistry.

Post-Industrial revolution emissions are associated with global warming through increasing greenhouse gases and stratospheric ozone loss due to halocarbons (like chlorofluorocarbons), which may impact the model simulations and therefore mask the underlying mechanisms between sea ice loss and the large-scale circulation. To avoid this impact, we will use the CESM FW1850 component set to spin up the model. Based on a 1-degree finite volume (FV) dynamical core, this component set includes the TSMLT species distribution, an interactive land, and the Model of Emissions of Gases and Aerosols from Nature (MEGAN2.1). Most importantly, the gases and aerosol are specified with respect pre-industrial control emissions based on the Coupled Model Intercomparison Project Phase 6 (CMIP6). These control emissions remove the impact of global warming and ozone loss.

The initial condition data for FW1850 is specified for 1st January 1850 and has a horizontal resolution of 0.95° latitude by 1.25° longitude. Along with this initial condition, the default boundary condition values are derived from pre-industrial climatology through the merging of the monthly SIC and SST of the Hadley Center dataset version 1 (HadISST1) and the weekly optimum interpolation (OI) SST analysis of Hurrell et al. (2008) as provided by the National Oceanic and Atmospheric Administration (NOAA). Besides the pre-
industrial climatology data, WACCM database also contains historical monthly time series data from 1850 to 2012 and the present-day (1982-2001) climatology data.

Other relevant model parameterizations include microphysics, cumulus convection, and boundary layer (www.cesm.ucar.edu/models/cesm2/whatsnew.html). In WACCM6, the Cloud Layers Unified by Binormals (CLUBB) scheme is used to parameterize shallow convection, cloud macrophysics, and boundary layer turbulence, where the planetary boundary layer height is evaluated through effective roughness stress. CLUBB is a prognostic moist turbulence scheme that calculates joint higher-order moments of sub-grid vertical velocity, water content, and liquid water potential temperature. Through assumed joint binormal probability density functions (PDFs) of these quantities, equations for these moments are closed. In addition to calculating sub-grid vertical fluxes, CLUBB’s PDF closure is also used to calculate large-scale condensation and cloud fraction.

An improved two-moment prognostic cloud microphysics (MG2) has also been applied, which could carry prognostic precipitation species (rain and snow) in addition to cloud condensates. MG2 interacts with the MAM4 aerosol microphysics scheme to calculate condensate mass fractions and number concentrations. The process of deep convection is treated with a parameterization scheme developed by Zhang and McFarlane (1995) and modified with the addition of convective momentum transports by Richter and Rasch (2008) and a modified dilute plume calculation following Raymond and Blyth (1986; 1992). The scheme is based on a plume ensemble approach where it is assumed that an ensemble of convective scale updrafts (and the associated saturated downdrafts) may exist whenever the atmosphere is conditionally unstable in the lower troposphere. Since
all these physical parameterizations impact large-scale circulation, they are important in our study.

5.1.2 Specification of Boundary Conditions

For our experiments, we construct an annually repeating climatology of daily SST and SIC from the WACCM historical time series data (1850-2012) to form the bases of our WACCM6 boundary condition (BC) set, discussed below. Two different BC sets facilitate our investigation of the impact of sea ice loss on the large-scale atmospheric circulation. Prior to this construction, we initially trim the time series. For SIC, only historical time series data after 1979 is kept to avoid questionable sea ice variability before the satellite era, which has provided a continuous and nearly complete record of Earth’s sea ice cover. For SST, only time series after 1870 will be used since the data during first 20 years (1850-1870) was periodically repeated.

5.1.2.1 Neutral BC (BC1)

Monthly SST climatology is computed from 1870 to 2012 and correspondingly for SIC from 1980 to 2012. As shown in Fig 5.2, the monthly SIC evolution reveals the seasonal behavior with minimal values in September and peaking around March. We also see that changes in areal extent and concentration are most pronounced near the Bering Strait and Chukchi Sea. At low latitudes, the climatological SST is largest in the summer and fall. Around the same time, we also note large warming near the Bering Strait, with peak values in August and September.
Fig 5.2 Monthly climatology SIC (left) and SST (right). Each row represents a season starting from winter season (DJF). The unit of SIC is fraction and SST degree Celsius.

The computed monthly climatology SST and SIC are interpolated to daily values. Three interpolation methods: linear, Lagrangian, and cubic spline are compared (not shown). Lagrangian and cubic spine interpolations work well; however, the smoothing nature inherent in these methods tends lower the peak value. In order to keep the extreme signal, simple linear interpolation is adopted instead. As an illustration, Fig 5.3 shows the monthly climatology (triangle symbol) and linearly interpolated (dash line) daily SST and SIC at a model grid point. In comparing monthly climatology and daily interpolated value, we see that the interpolation works correctly. To this end, the BC1 set consists of neutral SST and SIC conditions and will be used to force WACCM6 to establish a baseline response of the large-scale atmospheric circulation.
5.1.2.2 Low Sea ice BC (BC2)

To assess the baseline atmospheric respond to sea ice change, we set up a second BC set (BC2) with sea ice loss region over the Pacific sector (Bering-Chukchi), defined between 120°E-220°E and 65-90°N. BC2 contains the annually repeating daily climatology SST and SIC values (noted above) but with a specified sea ice loss in the Pacific sector.

To define this regional ice loss, we utilize standard deviation (or sigma) values following the method of Screen and Francis (2016) and Screen (2017). Initially, 2-sigma SIC value for each month is determined relative to the corresponding 33-year mean (1980-2012) monthly averaged value. Within the defined Pacific sector, the 2-sigma SIC value is then subtracted from the corresponding annually-repeating climatological SIC value at each ice-covered grid box.

Fig 5.4 shows the SIC difference between the BC2 set relative to BC1 (i.e., BC2-BC1). A negative difference indicates that SIC values are smaller than the neutral (and climatological) condition. As expected, BC2 exhibits SIC loss over the defined Pacific sector. As large sea ice difference near the boundary of Pacific sector may cause misleading fluxes,
a smoothing of $10^\circ$ longitude and $5^\circ$ latitude is applied to the anomaly fields prior to subtraction from climatological values. Fig 5.4 shows the result after smoothing. Although sea ice loss region is different, the anomalously low SIC is comparable to Sun’s work (reproduced for reference in Fig 5.5).

Fig 5.4. SIC difference between BC1 and BC2. Each row represents a season starting from winter season (DJF). Unit is fraction.
Fig 5.5. Seasonal cycle of SIC (%) averaged over the (a) late twentieth century (1980-99), (b) late twenty-first century (2080-99), and (c) their difference. The dashed circle in (c) denotes the Arctic Circle (66.6°N). (d) Surface heat flux response (positive upward) in total sea ice loss (right). [Graph is from Sun et al., 2015].

Monthly low SIC is interpolated into daily and repeated for many years. Fig 5.6 shows regional averaged SIC over Pacific sector for one year (1850). The red solid line represents SIC from BC1 and blue dotted line from BC2. The difference between BC1 and BC2 reflects the sea ice loss magnitude, which maximize in September.
As noted by Screen (2017), SSTs can change when sea ice loss occurs. Hence, the corresponding SST of BC2 must be adjusted accordingly. Similarly, 2-sigma SST values are used to obtain the monthly SST anomaly, which is then added to the climatological SST value at each ice-covered grid box within the Pacific sector. The combined SST values are also linearly smoothed in space.

### 5.1.2.3 Adjustments of SST and SIC

Since the historical time series of SST and SIC are from different period and added on different anomalies, adjustments of SST and SIC are necessary to maintain realistic relationship between SST and SIC.

In CESM, the prescribed SST and SIC values are adjusted according to the empirical method proposed by Hurrell et al. (2008). To main consistency between the observed interrelated behavior of SST and SIC, these authors provided the following adjustment criteria:

- SSTs are set to -1.8°C when SSTs are colder than -1.8°C or for all ocean points in
regions where SICs are larger than 0.9.

- SICs are always set to be between 0 and 1.
- SICs are set to 0 when SSTs are warmer than 4.97°C.
- When SST exceeds \( \text{SST}_{\text{max}} \) (as defined by equation 5.1), SIC is reduced to a value that corresponded to \( \text{SST}_{\text{max}} \).

\[
\text{SST}_{\text{max}} = 9.328(0.729 - SIC^3) - 1.8 
\]

The -1.8°C is the freezing point of seawater. After this adjustment, the magnitude of the adjustments to SICs is small (~0.01) and much larger to the SSTs (~2°C). The adjustment impacts inland sea regions and Arctic regions covered by ice.

5.1.3 Model Spin-Up & Main Initial Condition

Whenever initial and boundary conditions are introduced to a numerical model, a spin-up time is needed for the model to remove possible inconsistencies and arrive at a balanced state. Changes in the atmospheric parameters impact the model spin-up time length (Simmonds, 1985). Since the atmosphere has a relatively short memory, an atmosphere-only model can spin up in an order of weeks. In our study, BC1 (neutral BC with annually repeating climatological SIC and SST) and CESM2 default initial condition in 1850 will be used to spin up WACCM6 in a free-running mode. Following the work of Mckenna et al. (2018) who studied the atmospheric response to sea ice loss using an atmosphere-only model (like ours), the spin-up time length is one year. The spin-up process allows the model’s atmospheric state to equilibrate from the initial condition and become adjusted to the boundary condition and various settings.

Since WACCM6 can internally generate the QBO, the model is run for multiple years to
establish a few QBO cycles. The exact year to establish our main initial condition is based on QBO phase. Fig 5.7 shows time-altitude cross section of zonal-mean equatorial ($5^\circ S$-$5^\circ N$) wind between 10-40km. QBO phase is defined as the zonal-mean equatorial wind over the altitude depth of 20-30 km. During model year 1850-1859, there are four QBO cycles. The first QBO cycle was discarded because the model is still spinning up. Since the second and third EQBO cycle start around January, we cannot branch from them. The fourth EQBO starts around June 1858 and serves as the initial condition for EQBO experiments. Similarly, the second WQBO starts around June 1852 and serves as the initial condition for WQBO experiments.

Fig 5.7. Time-altitude cross section of monthly-mean, zonal-mean equatorial ($5^\circ S$-$5^\circ N$) wind in m s$^{-1}$ between 10km and 40km from CNTL 1850-1859. Positive values denote eastward winds and the contour line is at 0 m s$^{-1}$.
5.1.4 Other External Forcings

As noted above, in using the CESM F1850 component set with pre-Industrial revolution emissions, we avoid complications associated global warming and stratospheric ozone loss. However, in using WACCM that extends vertically to ~150 km, the solar influence must be still properly addressed (see Fig 5.1). Solar activities could have a potential impact on the stratosphere. Camp and Tung (2007) found that the state of the eastward phase QBO during the 11-year solar cycle minimum emerges as a coldest state of the stratospheric polar vortex. To minimize the impact of solar activities on the stratospheric polar vortex, our experiments are prescribed with minimum solar forcing, which is obtained through historical solar forcing data. Fig 5.8 shows the 10.7-cm solar radio flux (F10.7) time series from 1850-2012. Minimum solar forcing years are defined when F10.7 is less than 73 solar flux unit (sfu). All variables related with solar activities and ion pair production rate from energetic particle precipitation will be averaged during those minimum solar forcing years, which will be supplied to WACCM6 as solar forcing input.
Fig 5.8. Time series of F10.7 from 1850 to 2012. Blue line is critical value (73sfu), which defines minimum solar forcing.

Table 5.1 shows a comparison of our derived minimum solar forcing parameters with those used in the default WACCM6 setting averaged from 1995-2005. We see that the minimum solar forcing specification in our model set up is much weaker. This suggests that potential atmospheric response to solar activities will be minimized. From Table 5.1, the geomagnetic activity in our model set up is also much weaker. The Ap index and Kp index describes geomagnetic activity. The former is a measure of the general level of geomagnetic activity over the globe for a given day. The latter measures solar particle radiation by its magnetic effects.
Table 5.1. Solar Parameter Difference between WACCM Default and Solar Min Specifications

<table>
<thead>
<tr>
<th>Solar Parameters</th>
<th>WACCM default</th>
<th>Specified Solar Minimum</th>
</tr>
</thead>
<tbody>
<tr>
<td>F10.7 solar radio flux (sfu)</td>
<td>124.48</td>
<td>66.42</td>
</tr>
<tr>
<td>Daily planetary Ap index</td>
<td>13.14</td>
<td>8.26</td>
</tr>
<tr>
<td>Daily planetary Kp index</td>
<td>2.18</td>
<td>1.54</td>
</tr>
<tr>
<td>Smoothed sunspot number</td>
<td>61.15</td>
<td>11.53</td>
</tr>
<tr>
<td>Reconstructed total solar irradiance at 1 AU (W m(^{-2}))</td>
<td>1361.16</td>
<td>1360.61</td>
</tr>
</tbody>
</table>

5.1.5. Experiments

With the boundary conditions and the main initial condition discussed above in WACCM6, we propose numerical experiments to address our research questions. The two boundary conditions are associated with the neutral and anomalous conditions of SIC and SST, related to sea ice loss over the Pacific sector. Two experiments will be performed as discussed below to assess the impact of autumn sea ice loss in the Pacific Sector on the large-scale atmospheric response in the following winter in a different background state (QBO).

As discussed in sub-section 5.1.3, for EQBO, two experiments will start on 1 June 1858 and comprise of a control experiment (CNTL) and a perturbation experiment (LOW) of different boundary conditions. Without the impact of sea ice loss, CNTL is prescribed with BC1, which contains annually repeating daily climatological SST and SIC. To assess the SIC impact, LOW is prescribed with BC2, which consists of annually repeating daily
climatological SST and sea ice loss in Pacific sector. Comparison between CNTL and LOW could reveal the impact of sea ice loss over Pacific sector on large-scale circulation during EQBO. Similar experiments are performed for WQBO, which will start on 1 June 1852. The impact of QBO could be assessed by comparing these four experiments.

The imposed boundary conditions (BC2) may not be consistent with the main initial condition on 1 June 1858 (EQBO) or 1852 (WQBO). Recall that the main initial condition will correspond to the state as provided by the spun-up control run. To resolve this inconsistency with the annually repeating boundary conditions, we will allow the perturbation experiments to additionally spin up for another year. Hence, the model arrives at a balanced state at 1 June 1858 (EQBO) or 1852 (WQBO) for all experiments. Fig 5.2 shows that SIC maximizes in March and minimizes in September. After three months, these experiments will capture the atmospheric response of the low seasonal sea ice and its seasonal recovery through winter. For consistency, the control (CNTL) experiment will also be set to have the same initial starting date. For all experiments, the total integration time will be one year, which is shorter than the QBO period (~28 months).

Atmospheric internal variability can be caused by non-linear interactions between different atmospheric components. Screen et al. (2014) stated the importance of the atmospheric impacts of sea ice loss partly depends on the relative magnitude of the sea ice forced change compared to internal variability. In an attempt to distinguish the effects of sea ice loss from the background variations, ensemble runs are performed and the combined results are used to derive the most probabilistic model response. Screen et al. (2014) concluded that, if only Arctic sea ice were changing, an ensemble size of around
50 members is desirable. Given our available computing resources and time constraints for this study, 75 is chosen as our ensemble size. Hence, each 1-year experiment (after a 1-year spin-up from the main initial condition) will be run 75 times with identical surface boundary conditions and an atmospheric initial state perturbed with atmospheric internal noise. As suggested by Kay et al. (2015), the initial temperature fields are randomly perturbated by order $10^{-14}$ K. Fig 5.9 illustrates the logistics of CNTL and LOW experiments during EQBO. Two experiments during WQBO will be similar but starts on 1 June 1852.

**Fig 5.9. Flow chart for two experiments, CNTL and LOW set-up during easterly QBO (EQBO).**

### 5.1.6 Model evaluation

Before running the LOW experiment, we evaluate model’s ability to adequately represent key features of the large-scale atmospheric circulation using the control experiment. The zonal mean zonal wind simulated in WACCM6 is evaluated for its representation of the tropospheric jet stream and stratospheric polar vortex. Fig 5.10 shows the zonal mean zonal wind during each season based on CNTL 150 ensemble members (top) and Japanese 55-year Reanalysis (JRA-55) averaged over 1979-2021 (bottom). The zonal mean wind is eastward through most of the troposphere and peaks at speeds around 30 m s$^{-1}$ in the
subtropical jet stream. The subtropical jet stream is usually centered near 30 degrees latitude and at an altitude of 12 km. It is strongest in the winter season. During winter, eastward zonal mean wind also peaks at speeds around 60 m s\(^{-1}\) in the polar jet stream, which is centered at mid-latitude and in stratosphere. Compared to subtropical jet stream, polar jet stream is much stronger during winter, which is caused by stronger temperature gradient between polar region and mid-latitude region, and the direction is reversed during summer. The zonal wind simulated in WACCM6 in Southern Hemisphere is slightly stronger. Overall, the structure and magnitude of zonal mean wind is comparable with that in the reanalysis data, which implies that the model’s characterizations of the global atmospheric circulation are realistic.

![Fig 5.10. Latitude-altitude cross section of zonal mean wind for Dec-Feb (a, e), Mar-May (b, f), Jun-Aug (c, g), and Sep-Nov (d, h) from CNTL 150 ensemble members (top) and Japanese 55-year Reanalysis (JRA-55) averaged over 1979-2021 (bottom). Contour interval is 10 m s\(^{-1}\). Negative values represent westward wind.](image)

Secondly, we check the PW and its forcing simulated in WACCM6. Fig 5.11 shows EP flux vector and its divergence in four seasons based on CNTL. During winter, enhanced upward EP flux at middle latitude in the lower troposphere indicates intense poleward heat flux.
Near the tropopause, the vectors veer toward the tropics, indicating the increasing poleward momentum flux. The EP flux convergence near 50 km at about 50°N indicates the decelerative effect of PWS on the mean flow. In the summer, the upward propagation at middle latitude is suppressed due to the westward mean flow. Given its relationship to group velocity, the shown EP flux illustrates that the PW energy propagation varies seasonally.

Next, we check the ability of WACCM6 to simulate SSWs. This suggests whether the model can adequately represent the impact of wave disturbances on the mean flow and the process by which the mean flow anomalies subsequently descend into the troposphere. SSW definition will be in sub-section 5.3.3. For CNTL experiments, there are 55 and 27 SSW events during EQBO and WQBO, respectively. The frequency simulated in WACCM6
is roughly 0.55, which compares well with the SSW frequency in reanalysis data. More frequent SSW events during EQBO is consistent with the observations of Butler and Polvani (2011). In terms of the descent of the stratospheric anomalies on the surface, we check the time-altitude cross sections of the zonal-mean wind evolution at 60°N during SSW. As an example, one SSW event is shown in Fig 5.12. Wind strengths are comparable with reanalysis (Fig 2.7). The strength of stratosphere-troposphere coupling is adequately represented by WACCM6.

![Time-altitude cross sections of zonal-mean zonal wind at 60°N for one SSW simulated in CNTL EQBO.](image)

As mentioned in Chapter 3, the atmospheric response to sea ice loss involves the interference of climatological stationary waves with anomalous forced waves. Next we check the climatological stationary waves simulated in WACCM6. Fig 5.13 shows zonal wavenumber 1 and 2 components of geopotential height averaged over 60°N-80°N during winter (DJF) from CNTL. The structure illustrates the climatological, stationary PW wavenumber 1 and 2. Fourier transform is applied to do spatial filtering. The wave amplitude and phase are comparable with reanalysis data (not shown).
Fig 5.13. Longitude-height cross section of spatially averaged (60°N-80°N) zonal wavenumber 1 (filled contours) and wavenumber 2 (line contours) geopotential height during winter (Dec-Feb) from CNTL 150 ensemble members. The line contour interval is 100 m.

In summary, WACCM6 does a good job of representing tropospheric and stratospheric processes and is appropriate for the purposes of this thesis.

5.1.7 Remarks on Model Set-Up

Compared to related studies on sea ice loss over the Pacific sector (Sun et al., 2015; Screen, 2017; McKenna et al., 2018; Ding et al., 2021), our study has some technical improvement. The model we use (WACCM6) contains full chemistry and has a much higher vertical lid to better represent the coupling between the troposphere and stratosphere. The finer resolution horizontally and vertically of WACCM6 could capture more heat flux from surface, which is very crucial for diagnosis. Different from McKenna et al. (2018), our study will prescribe model with sea ice data directly rather than using SST anomaly as proxy.

Yet, we must bear in mind model deficiencies as we analyze model output and interpret
our results. As described in sub-section 5.1.1, WACCM6 employs the default CAM6 physical parameterizations and adjustments of SST and SIC for ice covered and ice-free region based Hurrell et al. (2008). Our results will be sensitive to these parameterizations (e.g., related to microphysics or the boundary layer) and adjustments. Clearly, in using an atmosphere-only model, the coupling with the ocean will be discounted in our results. As noted by Screen et al., (2018), the atmospheric response using atmosphere-only model may be more robust and consistent with coupled model for large sea ice anomalies. In our study, two times standard deviation has been subtracted to represent sea ice loss in Pacific sector. The magnitude is comparable with sea ice loss in Sun et al. (2015). Therefore, our prescribed sea ice anomalies shown are large enough to elicit a robust atmospheric response.

As summarized in Fig 5.9, for one set of CNTL and LOW experiments, QBO is constrained to one phase throughout the winter. In most studies of the sea ice atmospheric response, with the exception of Labe et al. (2019), the impact of separate QBO phases is not considered, and the reported results are inferred for the combined QBO phases.

While sea ice anomalies can alter the overlying atmosphere, resulting atmospheric anomalies could in turn impact the sea ice. Smith et al. (2018) mentioned that stratospheric anomalies could change the sea ice evolution. In particular, the effect of SSW or anomalous strong stratospheric polar vortex on the Chukchi Sea can elicit a strong sea ice response in the following 6-8 months. Since we focus on the impact of autumn sea ice loss on following wintertime large-scale atmospheric circulation, the lagged atmospheric effect on sea ice may be negligible in our simulation with the prescribed sea
ice loss.

Finally, our experiments will be done with a fixed ice thickness of 2 m in the North Hemisphere and 1 m in the South Hemisphere. This fixed thickness is artificially imposed. In a simple sea ice model (ignoring snow coverage), the ice thickness along with the temperature difference across the sea ice determine the heat flux between the ocean and the near-surface atmosphere (Hartmann, 2015).

5.2. Reanalysis Data

A reanalysis dataset is essentially created by combining observations and numerical weather prediction simulations providing the most complete picture currently possible of past weather and climate. It typically extends over several decades or longer and covers the entire globe from the Earth’s surface to well above the stratosphere. The historical observational data across the globe are available from different sources including aircraft, buoys, meteorological stations, radiosondes (balloon-borne instruments which record vertical atmospheric profiles of temperature and winds), satellites, and ships. The incorporation of observations into the model is done by a process called data assimilation. The observations are introduced at different times to correct the model in order to ensure that the evolution of the atmosphere during the simulation is as close as possible to the observed atmosphere. The assimilation process transforms an irregularly distributed network of observations into a three-dimensional grid (Bengtsson et al., 2004) of the best estimate state of the atmosphere for any period and place (Thorne and Vose, 2010), at different vertical layers of the atmosphere and time steps.

Since 1979, the use of weather satellites has allowed more data to be incorporated in the
assimilation process. With improved observational systems, the model or the method of assimilation, several generations of global atmospheric reanalysis dataset have been produced by various meteorological centers. NCEP/NCAR (National Center for Environmental Prediction/National Center for Atmospheric Research) reanalysis spans from 1948 to 2021 and is a first generation product. Extending from 1857 to 2002, the ERA-40 data from the ECMWF (European Centre for Medium-Range Weather Forecasts) is a second generation using 3D-Var assimilation system (Uppala et al., 2005). The third generation includes well-established datasets such as ERA-Interim (Dee et al., 2011), National Aeronautics and Space Administration (NASA) Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2) (Gelaro et al., 2017), and Japanese 55-year Reanalysis (JRA-55) (Kobayashi et al., 2015). Available from January 1979 to August 2019, ERA-Interim is produced by the ECMWF using the Integrated Forecasting System model. The main difference from ERA-40 is the use of 4D-Var, which assimilates all available observations into the model over a 12-hour analysis period and optimizes them in both space and time. With coverage starting in 1980, MERRA-2 data was introduced to replace the original MERRA dataset because of the advances made in the assimilation system that enable assimilation of modern hyperspectral radiance and microwave observations, along with GPS-Radio Occultation datasets. JRA-55 is a comprehensive climate dataset that covers 55 years, extending back to 1958, coinciding with the establishment of the global radiosonde observing system.

Since our experiments are in 1980s, JRA-55 is used to verify model results. Atmospheric data of JRA-55 contains 37 vertical levels from 1000 to 1hPa with a 1.25° horizontal
resolution. The SIC data is from the Centennial In Situ Observation-Based Estimates of the Variability of SST and Marine Meteorological Variables (COBE-SST) product (Ishii et al. 2005), which provides boundary conditions for JRA-55. JRA-55 is the first comprehensive reanalysis that has covered the last half-century since the ECMWF ERA-40 and is the first one to apply four-dimensional variational analysis to this period (Kobayashi et al., 2015). It is suitable for the studies of multidecadal variability and climate change.

To quantify the simulation of QBO in WACCM6, MERRA-2 is used. Coy et al. (2016) showed that MERRA-2 produces a realistic QBO in the zonal winds, mean meridional circulation, and ozone during 1980–2015. In particular, the MERRA-2 zonal winds show improved representation of the QBO 50-hPa westerly phase amplitude at Singapore when compared to MERRA. The data record from January 1980 through January 2020 (40 years) is used. This gives 17 cycles of the QBO over the 480-month period with an average QBO period of about 28.2 month. The gridded longitude by latitude resolution used is 1.9° by 2.5° and the data has been interpolated into 72 pressure level form the surface up to 0.01hPa.

5.3. Data Analysis

We define the response of any field by its difference between CNTL and LOW experiments. To find the mechanism related to sea ice loss, we will quantify surface turbulent flux, which contains surface sensible heat flux and latent heat flux. From it, we can examine how the near-surface response to the imposed surface forcing and the time lag between them. To investigate the impact of AA, the air temperature and thickness will be quantified. Through eddy heat flux and eddy momentum flux, we can analyze wave
activities, their generation, and their impact on the background circulation. The circulation will be assessed through the time-mean of zonal wind, meridional wind, geopotential height, and temperature in the troposphere and stratosphere. This section sketches the tools used on this thesis.

5.3.1 Basic Structure and Internal Variability

Climate variations including internal variability can be diagnosed using Empirical Orthogonal Function (EOF) analysis (e.g., Thompson and Wallace, 1998). EOF can isolate structures that explain maximum variance among time series data. Therefore, it can extract coherent variations that are dominant and reduce the large number of variables of the original data to a few variables without compromising much of the explained variance (Hannachi, 2004).

5.3.1.1 EOF analysis

In EOF analysis, we assume a data matrix $D_{M \times N}$, where in this case $M$ represents time dimension and $N$ the space dimension. Through EOF, we try to find the state vector (leading EOF) that explains the largest possible fraction of total variance in $D_{M \times N}$. Let $e_1$ denote the state vector we are seeking. By definition, $e_1$ should have the maximum possible resemblance to the ensemble of state vectors that make up $D$ because $e_1$ explains the largest possible fraction of variance in data set $D$. The resemblance between $D$ and $e_1$ at one time ($t_1$) can be quantified as the projection between two vectors, that is:

$$\sum_{j=1}^{N} D(t_1, j) \cdot e_1(j)$$

(5.2)
Here $e_1$ is a unit vector, so only the direction and not its amplitude affects the projection.

To avoid the impact of mean, the mean is subtracted from the time series in $D_{M \times N}$ before projection. The projection at one time may be either positive or negative, which may cause the average of 5.2 over the entire data matrix be very small. Therefore, the net resemblance between $e_1$ and $D$ can be expressed as the average over all time steps of the squared projection of $e_1$ onto $D$:

$$
\lambda = \frac{1}{M} \sum_{i=1}^{M} \sum_{j=1}^{N} (D(i,j) \cdot e_1(j))^2 = \frac{1}{M} e_1^T D^T D e_1 \tag{5.3}
$$

$\lambda$ measures the resemblance between $e_1$ and $D$. In theory, the state vector $e_1$ can be found by maximizing $\lambda$. Since the means of columns in data matrix $D$ are zero, the diagonal of $\frac{1}{M} D^T D$ contains the variance of the time series at all grid points and the off diagonal contains the covariances between time series at different grid points in the data matrix. We rewrite 5.3 as:

$$
\lambda = \frac{1}{M} e_1^T D^T D e_1 = e_1^T C e_1 \tag{5.4}
$$

where $C_{N \times N}$ is defined as temporal covariance matrix. Multiply both sides by $e_1$:

$$
\lambda e_1 = C e_1 \tag{5.5}
$$

e_1$ is an eigenvector of the temporal covariance matrix $(\frac{1}{M} D^T D)$ which has the largest eigenvalue $\lambda$. The eigenvectors of the temporal covariance matrix are called the EOFs of the data. The eigenvector corresponding to the largest eigenvalue of the covariance matrix ($e_1$) is referred to as the leading EOF. The time series that describe the temporal evolution of the EOFs are called principal component (PC) time series. It can be found by
projecting the original data matrix onto the vectors corresponding to the EOFs.

### 5.3.1.2 AO loading pattern and index

To identify the AO loading pattern and its associated AO index, EOF is applied to the time series of geopotential height anomalies $Z'$ poleward of $20^\circ N$ for the Northern Hemisphere (Baldwin & Dunkerton, 2001).

The calculation is as follows. For each pressure level, we calculate the seasonally varying climatology $\bar{Z}$, which is defined as the 90-day low-pass filtered daily climatology, at each latitude, longitude, and day. The climatology is subtracted, leaving anomalies $Z'$ that retain daily to interannual time scale variations but without seasonal variation. Then 90-day low-pass filter is applied to the anomaly fields and retain only November to April $\bar{Z}'$. The anomaly fields are weighted by the square-root of the cosine of latitude to ensure equal area weighting towards the pole. The first PC time series obtained from EOF analysis are normalized by the standard deviation of the time series. Thus, it has unit variance. The associated AO loading pattern $e$ can be obtained by regressing the anomaly fields $\bar{Z}'$ onto the PC time series. The AO loading pattern represents the anomaly fields associated with an AO index of 1.

The daily AO index is constructed by projecting time series of daily geopotential height anomalies $Z'$ onto the AO loading pattern $e$. The daily anomalies are defined as deviations from the seasonally varying climatology $\bar{Z}$. Both the anomalies and loading pattern are area-weighted by the square-root of the cosine of latitude.

### 5.3.1.3 MJO phase and amplitude

The MJO activity is assessed through the daily real-time multivariate MJO (RMM) index,
which is defined as the leading two EOFs of the combined fields of the zonal wind at 850 and 200 hPa, and the outgoing longwave radiation meridionally averaged between 15°S and 15°N (Wheeler and Hendon 2004). The MJO phase is calculated as \( \tan^{-1}(\text{RMM2}/\text{RMM1}) \), where RMM1 and RMM2 are the first two principal components. It indicates the location of anomalous convection. A MJO event is termed active when the MJO amplitude \( \sqrt{\text{RMM}_1^2 + \text{RMM}_2^2} \) exceeds 1.

5.3.2 Wave Activities

Stationary and non-stationary Rossby waves will be examined to address their structure such as amplitude and phase. This can be done by examining the Hovmöller diagrams (longitude versus time plots at various latitudes or altitudes) or plots of the eddy field (i.e., departure from zonal symmetry) of key variables like geopotential height or temperature. This structure will elucidate the zonal wavenumbers, the wave phase propagation, and will be compared with theoretical expectations for Rossby waves discussed Chapter 2.

The wave energy propagation will be diagnosed through wave activity related to the wave fluxes. In addition to examining wave activity on the zonal-mean perspective, we will also consider their presence across different longitudes. The latter approach is not typically applied in most studies related to SIC loss. From wave activity, we will calculate the wave forcing related to the convergence of the wave fluxes and how that can alter the background flow and drive an overturning circulation which can spread the wave effects. The formulation of wave activity and forcing is discussed here.
5.3.2.1 The Traditional EP Flux

Dynamical changes in the atmosphere may result from the interaction between the mean flow and atmospheric waves. The meridional cross sections of EP flux and its divergence are useful as a standard diagnostic for propagation of PWs and their interactions with a zonal mean flow (Andrews and McIntyre, 1976; Edmon et al., 1980; McIntyre, 1982; Andrews et al., 1987). To analyze wave activities in the atmosphere, we extend the simple EP flux expression introduced in Chapter 2 for spherical coordinates:

\[
F(\phi) = \rho a \cos \phi \left( \bar{u}_z \bar{v}^* \bar{\theta}^* / \bar{\theta}_z - \bar{v}^* \bar{u}^* \right) \\
F(\zeta) = \rho a \cos \phi \left\{ \left[ f - (a \cos \phi)^{-1}(\bar{u} \cos \phi)_{\phi} \right] \bar{v}^* \bar{\theta}^* / \bar{\theta}_z - \bar{w}^* \bar{u}^* \right\}
\]

where \(a\) is earth radius and \(\phi\) is latitude. Here, we also consider vertical momentum fluxes \((\bar{u}^* \bar{w}^*)\) for completeness although they are expected to be theoretically small for Rossby waves. The wave forcing will be computed through flux convergence (formulation was discussed in Chapter 2). Preliminary, wave forcing computed from EP flux is shown in Fig 5.11.

As mentioned in 2.1.5, the direction of \(F\) indicates the relative importance of the principal eddy fluxes of heat and momentum. For example, an upward EP flux vector indicates poleward heat flux, whereas an equatorward EP flux vector the poleward flux of zonal momentum. The divergence of \(F\) is a direct measure of the total forcing of the zonal-mean state by the eddies and could indicate the wave dissipation. EP flux represents a snapshot of the wave propagation on the meridional plane. However, the wave propagation in the zonal direction is neglected (Takaya and Nakamura, 2001).
5.3.2.2 Three-dimensional wave flux

Most related studies investigated atmospheric response through analyzing wind pattern, geopotential field, and EP flux (like above). Although EP flux is a good tool to elucidate Rossby wave propagation and zonal-mean wave forcing, wave activities in longitude direction cannot be captured. The longitudinal (or localized) distribution of wave activity may allow us to better identify the regions of wave excitation. Therefore, our study will diagnose three-dimensional wave activities, in terms of longitude and latitude, as well as longitude and altitude. To do this, the Plumb flux will be calculated, which describe the full three-dimensional direction of energy propagation by stationary quasi-geostrophic Rossby waves (Plumb, 1986). The Plumb flux, $E$ is expressed as:

$$ E = \frac{p}{p_r} \cos \phi \left\{ \begin{array}{l}
\frac{1}{2a^2 \cos^2 \phi} \left( \frac{\partial \psi^*}{\partial \lambda} \right)^2 - \psi^* \frac{\partial^2 \psi^*}{\partial \lambda^2} \\
\frac{1}{2a^2 \cos^2 \phi} \left( \frac{\partial \psi^*}{\partial \lambda} \frac{\partial \psi^*}{\partial \phi} - \psi^* \frac{\partial^2 \psi^*}{\partial \lambda \partial \phi} \right) \\
\frac{2\Omega^2 \sin^2 \phi}{N^2 a \cos \phi} \left( \frac{\partial \psi^*}{\partial \lambda} \frac{\partial \psi^*}{\partial z} - \psi^* \frac{\partial^2 \psi^*}{\partial \lambda \partial z} \right)
\end{array} \right\} $$(5.8)

where $\lambda$ is the longitude, $\psi^*$ is the streamfunction zonal deviation (calculated from $\psi = gZ/2\Omega \sin \phi$, $Z$ is the geopotential height), $\Omega$ is the Earth’s rotation rate, $N^2$ is the squared Brunt-Väisälä buoyancy frequency, and all other symbols are as defined in the previous section for the EP flux.

5.3.2.3 Eddy Heat flux decomposition

The enhanced upward PW propagation from the troposphere weakens the stratospheric polar vortex (e.g., Polvani and Waugh, 2004). The anomalous wave itself and/or the amplification of the climatological PW induced by anomalous waves could enhance the
upward PW propagation. The relative role of anomalous and climatological wave on the upward PW propagation can be diagnosed by the poleward eddy heat flux decomposition (Nishii et al., 2009). The poleward eddy heat flux $v^*T^*$ is proportional to the local vertical wave activity. A positive (i.e., poleward) eddy heat flux corresponds to a local upward wave activity. The meridional wind and temperature eddies can be written as the sum of their climatological and anomalous components:

$$v^* = v_c^* + v_a^*$$

$$T^* = T_c^* + T_a^*$$

The subscript $c$ and $a$ denote the climatology and the anomaly, respectively. The poleward eddy heat flux then be decomposed as:

$$v^*T^* = v_c^*T_c^* + v_c^*T_a^* + v_a^*T_c^* + v_a^*T_a^*$$  \hspace{1cm} (5.9)

The first and fourth terms in the right-hand side represent the heat fluxes due solely to the background climatological PW and solely to anomalous associated with PW, respectively. The second and third terms represent the interaction between the background climatological PW and anomalous waves or the modulating effect on the climatological PW by anomalous waves (Nishii et al., 2009).

The climatological component of eddy heat flux is obtained by taking the climatological mean:

$$\langle v^*T^* \rangle_c = v_c^*T_c^* + (v_a^*T_a^*)_c$$  \hspace{1cm} (5.10)

The second term in the right-hand side represent the climatological eddy heat flux associated with the anomalous waves $v_a^*T_a^*$. The anomalous component is expressed as the subtraction of poleward eddy heat flux from the climatological component:
\[(v^*T^*)_a = v^*T^* - (v^*T^*)_c = v^*_c T^*_a + v^*_a T^*_c + v^*_a T^*_a - (v^*_a T^*_a)_c \tag{5.11}\]

The first two terms in the right-hand side of equation 5.11 refer to the linear terms. The sum of last two terms is the nonlinear term, which represents the contribution from the anomalous waves compared with its climatological value.

Hoshi et al. (2017) further decomposed the poleward eddy heat flux anomalies associated with sea ice loss. Following the method from Nishii et al. (2011) and Hoshi et al. (2017), we decompose the poleward eddy heat flux anomalies induced by sea ice loss throughout the SSW lifecycle. The daily poleward eddy heat flux in CNTL is expressed as:

\[(v^*T^*)_{CNTL} = v^*_{CNTL} T^*_{CNTL} + v^*_{CNTL} T^*_{CNTL} + v^* T^*_{CNTL} + v^* T^*_{CNTL} \tag{5.12}\]

where the subscript \(\text{CNTL}\) denotes the CNTL climatology, which is the composite average of CNTL, and the prime denotes the deviation from the CNTL climatology. The composite average (denoted by square brackets) is:

\[[v^*T^*]_{CNTL} = [v^*_{CNTL} T^*_{CNTL}] + [v^* T^*_{CNTL}] \tag{5.13}\]

The daily poleward eddy heat flux in LOW is expressed as:

\[(v^*T^*)_{LOW} = v^*_{CNTL} T^*_{CNTL} + v^*_{CNTL} T^*_{CNTL} + v^* T^*_{CNTL} + v^* T^*_{CNTL} \tag{5.14}\]

where the subscript \(\text{R}\) denotes the response to sea ice loss, defined by the difference between LOW and CNTL climatology. The composite average is:

\[[v^*T^*]_{LOW} = [v^*_{CNTL} T^*_{CNTL}] + [v^*_{CNTL} T^*_{CNTL}] + [v^* T^*_{CNTL}] + [v^* T^*_{CNTL}] \tag{5.15}\]

The poleward eddy heat flux response to sea ice loss is thus given by:

\[[v^*T^*]_R = [v^*T^*]_{LOW} - [v^*T^*]_{CNTL} = [v^*_{CNTL} T^*_{R}] + [v^*_{R} T^*_{CNTL}] + [v^*_{R} T^*_{R}] \tag{5.16}\]

where

\[[v^*_{R} T^*_{R}]_R = [v^*_{R} T^*_{R}] - [v^{*'} T^{*'}] \tag{5.17}\]
5.3.3 Detection of Sudden Stratospheric Warming

An SSW event is classified as either major or minor. The detection of SSW events is, however, sensitive to their definition and no unambiguous standard identification has been adopted (Butler et al., 2015). A major SSW is often defined as when the 10-hPa zonal-mean zonal wind reverses as a result of the switched meridional temperature gradient in the polar region. A minor SSW occurs if the stratospheric temperature increases by at least 25 degrees Celsius in a period of a week or less, but the 10-hPa zonal-mean zonal wind does not switch direction (McInturff, 1978).

In this thesis, only major SSW events are considered. The method to detect major SSW events is based on the World Meteorological Organization (WMO) definition. SSW is defined to occur when daily zonal mean zonal wind at 10hPa and 60°N becomes easterly and 10hPa zonal mean temperature difference between 50°N-70°N and 70°N-90°N is positive during the extended winter (November-March). Two events are separated by at least 20 days of westerly winds (Charlton and Polvani, 2007). To exclude the rapid polar warmings due to the transition of winter to spring (i.e., “final warming”), the 10-hPa zonal-mean zonal wind at 60°N must become westerly at least 10 days before April 30. The onset date of SSW is the first day of the wind reversal. Once the SSW events have been identified, composites are generated by averaging each SSW centered at the onset. The composite analysis smooths and weakens the amplitude but can reveal the robust response to SSW events.

Main characteristics of SSW events are examined, namely their frequency, duration, intensity, mean onset date, and type. The frequency of SSW event is the number of SSW
events divided by ensemble members. The duration is calculated by the number of consecutive days that the 10-hPa zonal-mean zonal wind at 60°N is easterly (Charlton et al., 2007). The intensity is defined as the peak westward zonal-mean zonal wind over 55°N-65°N and 40km-45km at SSW onset. The unit for intensity is m s⁻¹. The classification of SSW type is based on the geopotential height amplitude of zonal WN1 and WN2 averaged over 55°N-65°N at 10 hPa (Choi et al., 2019).

5.3.4 Comparison of Experiments and Their Significant Difference

The daily climatology is defined as the 31-day running average of the ensemble mean of all ensemble members in CNTL and LOW. Daily anomalies are calculated as deviations from that daily climatology. The composite difference between the perturbation experiment (LOW) and the control experiment (CNTL) will be assessed to discern the impact of the sea ice loss on SSW events. The LOW-CNTL difference might not only contain the model’s forced change due to the sea ice loss but also the influence of atmospheric internal variability. However, as the number of ensemble members increases, the influence of internal variability on the response will decrease, resulting in a more accurate estimate of the model response to sea ice loss.

We ensure that the response or an anomaly is statistically significant by employing two methods to assess significance. One method is the two-sided Student’s t-test with the null hypothesis that no difference exists between LOW and CNTL (for response) or between a variable and its climatology (for anomaly). The other method is the Monte Carlo simulation which makes no assumptions about test statistics’ probability distribution through mathematical parameterization. The response to sea ice loss or an anomaly will
be defined as statistically significant for a $p$-value smaller than 0.05. That is, the probability of the response or an anomaly occurring by chance is 5%.

The Monte Carlo simulation is based on the construction of artificial data sets from a given collection of real (experimental or observed) data (Wilks, 2011). One type of Monte Carlo simulation is done through permutation, in which samples are drawn without replacement, and works for multisampling problems (Mielke et al., 1981; Preisendorfer & Barnett, 1983). The other type is through bootstrapping, in which samples are drawn with replacement, and works better for a single sample problem. For our study, the bootstrapping technique is applicable. Rather than computing the test statistic using the pooled data, the pool is instead sampled some large number of times to reduce errors (Wilks, 2011).

We will perform the Monte Carlo significant test as follows using the anomaly or the response as the test statistics. As an example, we describe the procedure for testing the significance of the response. Assuming there are $m, n$ SSW events occurred in CNTL and LOW, respectively. Initially, the composite average in CNTL members ($M_{CNTL}$) and in LOW members ($M_{LOW}$) will be calculated separately. The difference of the two composite means constitutes the sampled mean difference ($M_\Delta \equiv M_{CNTL} - M_{LOW}$). Our null hypothesis is that CNTL and LOW are from one distribution, so changes related to low SIC in LOW are indistinguishable from the climatological SIC in CNTL. The alternative hypothesis is that they are different at a predefined 95% confidence level ($\alpha = 0.05$) based on 10,000 bootstraps.

For each bootstrapping with the drawn ($m + n$) SSW events, the composite average of $m$
and $n$ SSW events will be calculated to determine the mean difference. A histogram of
the mean difference ($\mu_\Delta$) will be constructed from 10,000 bootstraps. By comparing the
sampled mean difference ($M_\Delta$) to the mean difference ($\mu_\Delta$), we can decide to accept or
reject the null hypothesis.

When plotting the difference of an interested variable, this significant test will be done at
each data grid point and/or time as appropriate. The statistically significant portion will
be highlighted in the plot. The significances examined by the two-sided Student’s $t$-test
are similar with examined by the Monte Carlo simulation. To avoid the assumptions about
test statistics’ probability distribution, the Monte Carlo simulation will be employed
unless otherwise stated.

5.3.5 Detrend Analysis Data
To affirm model results, we will also analyze the reanalysis. The 3-hourly SIC dataset from
JRA-55 is used to identify years with anomalously low and high SIC. Fig 5.14a shows the
time series of original SIC over the Pacific sector from 1958-2020. SIC is averaged in
September-October for each year. Red dot represents low sea ice years that SIC is smaller
than mean minus 0.5 standard deviation. Since there is a declining trend of SIC, low sea
ice occurs in recent years due to much smaller SIC compared to the years before 1979.
Additional compounded influences from increasing greenhouse gases might be present.
To reduce these possible influences, the SIC data is detrended first before the composite
analysis. For detrended SIC, low sea ice years are when September-October averaged SIC
is smaller than zero.
Fig 5.14. Time series of September-October averaged sea ice concentration (SIC) over the Pacific sector from 1958-2020 (JRA55). (a) Original SIC. (b) Linearly detrended SIC. (c) 7-year high pass filtered SIC. (d) Reconstructed SIC based on IMFs.

Three detrending approaches are investigated. One method is assuming the trend is linear and then remove it from the time series. This method has been used in previous sea ice studies (i.e., Hopsch et al., 2012; Hoshi et al., 2017). The linear trend can be estimated by least-squares fit, which minimizes the sum of the squared errors in the data series. The linear detrended original SIC is shown in Fig 5.14b. Note that the distribution of low sea ice years is more evenly distributed from 1958 to 2020 after subtracting linear trend. One weakness of this method is that the trends inherently present in SIC might not be linear.

The second method is to apply a high-pass filter on the SIC time series to keep the interannual variability while excluding the decadal variability. The 7-year high pass filtered SIC is shown in Fig 5.14c. Identified low sea ice years using this method are different from using linear detrend, especially around 1975 and 2000. The marginal information is
missing due to the non-periodic nature of the SIC data.

The third approach is to use the Empirical mode decomposition (EMD), which is a useful tool to extract signal in time domain (Huang et al., 1998). With no assumption about linearity or stationarity, EMD is different from Fast Fourier Transform in that it decomposes a multi-component signal into a set of oscillation named intrinsic mode functions (IMFs). An IMF satisfies two conditions: 1) the number of extrema and zero-crossings must be either equal or differ at most by one in the whole data set, and 2) the mean value of the envelope denoted by the local maxima and the local minima is zero. The procedure of extracting an IMF is called sifting. Suppose we decompose a signal $y(t)$, the main steps of EMD are as following (Barbosh et al., 2020):

a) Identify all extrema of the original signal $y(t)$ and interpolate the local maxima (minima) using a cubic spline. It is the upper (lower) envelope.

b) Calculate the mean of upper and lower envelopes named $m_1(t)$.

c) Compute the difference between the signal $y(t)$ and the mean $m_1(t)$, $d_1(t) = y(t) - m_1(t)$.

d) Check if $d_1(t)$ is an IMF based on two conditions mentioned above. If $d_1(t)$ satisfies both conditions, then $d_1(t)$ is the first IMF of the original signal $y(t)$. The residual, which is the difference between the original signal and this IMF, is treated as original signal. The sifting process is repeated.

e) If $d_1(t)$ is not an IMF, the sifting process will be repeated by using $d_1(t)$ as original signal until it is an IMF.

Eventually, the original signal $y(t)$ is expressed as:
\[ y(t) = \sum_{i=1}^{n} d_i(t) + r_n(t) \]  

(5.18)

where \( d_i(t) \) denotes the IMFs of the original signal \( y(t) \) and \( r_n(t) \) is the residual.

The steps for the complete decomposition should be finite due to the number of extrema decreasing in going from one residual to the next. Moreover, the whole decomposition is only based on subtractions, which allow for a reconstruction of the original signal \( y(t) \) based on the collection of IMFs and the residual. By using this method, slow or fast oscillations can be selectively removed.

The decomposition of September-October SIC over Pacific sector from 1958 to 2020 produces three IMF modes and a trend (Fig 5.15). The first mode is the interannual cycle (panel b). The second ENSO-like mode has an average period around four years (panel c). The third mode is the decal oscillation (panel d). Lastly, there is a slowly varying trend in SIC (panel e). Therefore, the detrended SIC by EMD is reconstructed by summing IMF 1-3 to exclude the slowly varying trend. As shown in Fig 5.14d, the detrended SIC by EMD looks similar by linear detrend result shown in Fig 5.14b. With no assumption about linearity or stationarity, EMD is chosen to detrend SIC in this thesis.
Fig 5.15. The results of application of the EMD technique to the SIC time series averaged in September-October SIC over Pacific sector (JRA55). (a) Original SIC time series. (b) The first IMF. (c) The second IMF. (d) The third IMF. (e) The fourth IMF.
6. SSWs Modulated by the Arctic Sec Ice Loss

The investigation begins by analyzing the response of the characteristics of SSWs to Arctic sea ice loss during EQBO. During this QBO phase, SSWs tend to occur more often (Butler & Polvani, 2011), and the Arctic sea ice loss tends to produce a response that results to the weakening of the stratospheric polar vortex (Labe et al., 2019). In analyzing the large-member ensemble model experiments and reanalysis dataset, the results here aim to elucidate the fundamental mechanisms linking the surface processes associated with sea ice loss to the wintertime stratospheric dynamics.

The focus of this chapter is initially on the depth of the warming induced by Arctic sea ice loss. This was examined previously by He et al. (2020), who suggested that climate models can replicate the warm Arctic-cold Eurasia pattern, only when the warming signal extends deep into the upper troposphere. Even though anomalously warm Arctic surface associated with the Arctic sea ice loss has been linked to the mid-latitude surface cooling in the subsequent boreal winter (Honda et al., 2009; Mori et al., 2014; Kug et al., 2015; Cohen et al., 2018; Mori et al., 2019; Tachibana et al., 2019; Kim & Son, 2020), He et al. (2020) concluded that since deep warming is independent of sea ice forcing, internal natural variability instead of sea ice loss is the main driving forcing for deep Arctic warming-cold Eurasia pattern.

The second focus of this chapter is to identify the impact of the declining autumnal sea ice over the Pacific sector on the SSW characteristics. Several studies have argued that
SSW events are related to the recent Arctic sea ice loss (Hoshi et al., 2019; Zuev & Savelieva, 2019; Zhang et al., 2020). These studies focused on sea ice loss over Atlantic sector. Although observations reveal that the most sea ice retreat occurs in the Pacific sector, its impact there on the atmosphere has received much less attention than over the Atlantic region. In addition, its impact on the SSW characteristics is still unknown.

The last focus of this chapter is to unravel the exact dynamical mechanism linking the PW activity entering the stratosphere to the actual sea ice loss. Previous studies inferred that the localized shallow heating due to sea ice reduction could modify the upward PW propagation from the troposphere into the stratosphere following changes in synoptic eddies and their feedback on the mid-latitude jet (Jaiser et al., 2013; Ruggieri et al., 2017). Yet, the complete chain of events is not fully understood. The relative importance of the autumn or winter sea ice loss has also remained unclear.

6.1 Seasonal Surface Response to Sea Ice Loss

The stable QBO phase during the 1858–1859 boreal winter is crucial to allow the experiments to elucidate how SSW respond to autumn sea ice loss during EQBO. Fig 6.1 illustrates the one-year QBO evolution for all CNTL and LOW members. The illustrated time series is defined as the zonal-mean zonal wind averaged over 5°S–5°N, 50 hPa–30 hPa. Overall, EQBO is persistent and similar between CNTL and LOW. The initial QBO is also similar on 1 June 1858, suggesting the absence of notable wind drift during the one-year spin-up prior to the ensemble experiment.
To understand how sea ice loss affects the stratosphere prior to SSW onset, we examine the seasonal evolution of the surface air temperature, sensible heat flux, and latent heat flux responses (Fig 6.2). The stippling indicates areas of statistical significance at the 95% level. Based our experiment design, the prescribed sea ice loss peaks in autumn and begins to recover in winter (cf., Figs 5.4 and 5.6). The induced warming (left column) persists well into the winter, consistent with the findings of Francis et al. (2009). The wintertime surface warming is ~6 degrees (panels d & g). Associated with the local SST warming due to the sea ice loss, the sensible heat flux and latent heat flux are located over the sea ice loss region (65°N–90°N, 120°E–220°E), the former being more dominant (compare panel e & f). This enhanced upward turbulent heat flux (i.e., the sum of the sensible heat flux and latent heat flux) from the ocean to the atmosphere further warms the overlying near-surface air.
Fig 6.2. Seasonal surface warming induced by Pacific-sector Arctic sea ice loss during EQBO. Surface air temperature (a, d, g, j), sensible heat flux (b, e, h, k), and latent heat flux (c, f, i, l) difference (LOW-CNTL) averaged in September–October, November–December, January–February, and March–April. Line contours on top of surface air temperature are sea ice concentration difference (in percentage) between LOW and CNTL experiment (contour values are 2.5, 5, 10, and every 10 thereafter). Stippling indicates statistical significance at the 95% level.

The seasonal evolution of the warming depth is shown in Fig 6.3. Note the cross-section is averaged over the longitudinal band of 150E–210E, which covers the region of temperature extrema in Fig 6.2. The warming over sea ice loss region extends into the mid-troposphere during autumn when the sea ice loss peaks (panel a). As sea ice loss begins to recover in the winter, the warming remains confined below 850 hPa (panels b & c). Such wintertime surface warming is consistent with the shallow sea ice-induced
warming shown in observations (He et al., 2021). Although sea ice loss is minor during spring, the warming extends into the mid-troposphere (panel d). In the following section, we address the impact of Arctic sea ice loss on SSW characteristics.

During EQBO

![Fig 6.3. Seasonal air temperature difference (LOW-CNTL) averaged over 150°E-210°E in September–October, November–December, January–February, and March–April during EQBO. Stippling indicates statistical significance at the 95% level.]

6.2 SSW Characteristics in Response to Sea Ice Loss

During EQBO, the number of SSW events that occurred among the ensemble members for CNTL and LOW are 55 and 50, respectively (Fig 6.4). Most SSW onset dates are found between January and March, with the mean onset date being February 4th in CNTL and February 2nd in LOW. The SSW onset dates are more confined during Jan-Feb in LOW.
compared to CNTL.

*Fig 6.4. Onset dates of SSW events in the model simulations (EQBO). The number of SSW events that occurred among the ensemble members for CNTL (a) and LOW (b) are 55 and 50, respectively.*

Fig 6.5 summarizes the SSW characteristics in CNTL (gray) and LOW (red) during EQBO. The definitions are provided in the sub-section 5.3.3. The number of polar vortex splits relative to polar vortex displacements increases in LOW but not significantly. A linear relationship exists between the vertical EP flux component (EPz) at 100 hPa and the SSW intensity, as increasing upward wave propagation into the stratosphere further perturbs the polar vortex. Overall, the mean 100-hPa EPz and SSW intensity, as well as the SSW duration are significantly larger in LOW (Fig 6.6). Hence, the autumnal sea ice loss over the Pacific sector leads to more persistent SSW events with stronger zonal-mean zonal wind reversals in the stratosphere.
Fig 6.5. Changes in SSW characteristics induced by Pacific-sector Arctic sea ice loss (EQBO). Scatter plot of the 100-hPa $EP_z$ (vertical EP Flux component) averaged between 45°N–75°N, 5 days prior to the onset (x-axis unit in $10^{-2} \text{ m}^2 \text{s}^{-2}$) versus SSW intensity (measured as a zonal-mean zonal wind deceleration, y-axis unit in m s$^{-1}$). Bubble size indicates the SSW duration in days, with a reference size provided. The mean values along each axis are shown as dashed lines. The split SSW frequency is the ratio of split SSW events to the total number of SSW events. Bold text indicates statistical significance at the 95% level.
Fig 6.6. Histograms for SSW characteristics. Histograms of the 10000 bootstrapped SSW characteristics response (LOW-CNTL) in occurrence (a), duration (b), intensity (c), and 100-hPa vertical EP Flux component ($EP_z$) (d). The characteristics are the same as in Fig 6.5. The red dashed line shows the mean response.

Composites of the geopotential height at SSW onset (day 0) are shown in Fig 6.7 at 100 hPa (panels a & b) and 10 hPa (panels d & e) for CNTL, LOW, and their difference (i.e., the response to sea ice loss in panels c & f). The line contours indicate the distorted polar vortex at the SSW onset. The filled contours represent departures from climatology, defined as the 31-day running average of the ensemble mean of all ensemble members in CNTL and LOW. At 100 hPa, the polar vortex is displaced toward Northern Siberia. Significant trough deepening occurs in response to the prescribed sea ice loss (panel c).
Specifically, the geopotential height is significantly lower from Eurasia to the Date Line, along with a localized negative height response over the North Atlantic. At 10 hPa, the polar vortex has tilted westward and is centered over Scandinavia, with large positive height anomalies over the polar cap. In response to the sea ice loss, the polar vortex becomes significantly deeper during SSW (panel f).

Enhanced upward PW propagation from the troposphere into the stratosphere generally accompanies SSW events (Matsuno, 1971). As illustrated in Fig 6.5, the mean upward wave flux prior to SSW onset becomes even more pronounced in response to sea ice loss.
Fig 6.8 details the composite time evolution by the 100-hPa EPz anomalies (averaged between 45°N and 75°N) in panels a & b, along with the vertical distribution of the zonal-mean zonal wind anomalies at 60°N in panels d & e. Prior to SSW onset (from day -20 to day 0), the anomalous upward wave activity increases, weakening the stratospheric zonal-mean zonal wind. Following the onset, the negative wind anomalies reach the surface within a few weeks, consistent with SSW-related observed characteristics (Baldwin & Dunkerton, 2001; Polvani & Waugh, 2004). The zonal wavenumber decomposition of EPz reveals that the anomalous wavenumber 1 (WN1) dominates over wavenumber 2 (WN2) in the pre-onset period in both CNTL and LOW. The concurrent peaks of WN1 and WN2 greatly amplify the overall upward EPz anomaly in LOW.

The right column of Fig 6.8 reveals that the sea ice loss elicits a strong response in the zonal wind reversal and upward wave propagation during SSWs (panels c & f), in support of Fig 6.5. This response consists of a significantly stronger wind reversal in the extended period following onset that contributes to the aforementioned longer SSW duration. The stronger wind reversal follows a significantly enhanced upward PW activity between day -10 to day 0, initially by the WN2 component response. The persistence of significant and deep negative wind response to day 20 implies a stronger troposphere-stratosphere coupling in presence of sea ice loss. To further understand the dynamical mechanism linking sea ice loss and SSWs, we will focus on 10 days prior to onset.
Fig 6.8. Composite zonal-mean zonal wind (U) and vertical EP Flux component (EPz) during SSW (EQBO). The time series of 100-hPa EPz anomaly (averaged in 45°N–75°N), along with their decomposition in zonal wavenumbers 1 and 2 (a–c; in 10^2 m^2 s^-2), along with the time-height cross section of 60°N U anomaly (d–f; in m s^-1). Left column corresponds to the CNTL experiment, medium column LOW experiment, and right column the response (i.e., LOW-CNTL). Open circles (a–c) and grey shading (d–f) indicate statistical significance at the 95% level.

The upper panels of Fig 6.9 depict the latitude-altitude sections of the EP flux (as vectors) and associated wave forcing (i.e., EP flux divergence as filled contours) for the CNTL and corresponding response to the prescribed sea ice loss, averaged in the 10 days prior to onset. The remaining rows illustrate the corresponding WN1 and WN2 components. For SSW in CNTL, the anomalous wave activity propagates upward from the troposphere in the mid-latitudes, refracting more equatorward in the stratosphere. The anomalous wave forcing strongly decelerates the wind, weakening the polar vortex as expected prior to SSW. Comparing panels c and e reveals that WN1 contributes mainly to the anomalous upward activities and wave forcings, consistent with the 100-hPa EPz results in Fig 6.8a.

Shown in Fig 6.9b, the sea ice reduction enhances the upward wave propagation and the wave decelerative effects. The increased wave-activity response originates from the near-
surface and becomes concentrated toward higher latitudes in the stratosphere. The associated zonal-mean zonal wind response shows that the subtropical jet intensifies and that the mid-latitude flank of the polar night jet becomes stronger. An enhanced polar night jet acts as a more effective wave guide, steering the upward wave activity toward higher latitudes. The stronger wave forcing further weakens the polar vortex compared to CNTL, leading to more severe and persistent SSW events (as noted in Figs 6.5 and 6.8). Above 100 hPa, both WN1 and WN2 responses are large. However, throughout the troposphere, the upward wave activity response arises mainly from WN2 between 40°N and 60°N (panel f). In the next section, we attempt to examine near-surface location where the upward wave activity originates.
Fig 6.9. Planetary wave EP flux (EQBO). Zonal-mean zonal wind (line contours; in m s⁻¹), EP Flux anomaly (vector; in m² s⁻²), and divergence anomaly (filled contours; in m s⁻¹ day⁻¹) in CNTL (left column), and response (right column) for all wavenumbers (top row), WN1 (middle row), and WN2 (bottom row) averaged from day -10 to day -1 prior the onset. Stippling indicates statistically significant regions of divergence (filled contours in b, d, f) at the 95% level.

6.3 Near-Surface Upward Wave Activity

To understand how sea ice loss impact the wave activities prior SSW events, we first examine the global temperature response at 850 hPa, 500 hPa, and 100 hPa averaged prior to the onset (filled contours in Fig 6.10). The horizontal temperature advection is
overlaid as line contours, with cold advection as dashes. Below 500 hPa (panels a & b), warming occurs over the sea ice loss region, due apparently to warm advection. Central and Eastern Eurasia and the Northwestern Pacific experience a cooling response. Cold advection appears responsible for the significant cooling just off the coast of Asia. In the upper troposphere (panel c), significant warming occurs more toward the mid-latitudes of the North Pacific and Eastern U.S. Intense cooling occurs over Northern Europe and the North Atlantic and nearly coincides with region of enhanced cold advection.

![Fig 6.10. Changes in temperature induced by Pacific-sector Arctic sea ice loss (EQBO). The LOW-CNTL differences in temperature (filled contours; in K) and horizontal temperature advection (line contours; cold advection as dashes, in 10-6 K s\(^{-1}\)) at 850 hPa (a), 500 hPa (c), and 100 hPa (e). Stippling indicates statistically significant regions of the filled contours at the 95\% level. The missing 850-hPa values are due to topography.]

To identify the upward wave activity response in the troposphere arising essentially from WN2, we examine the global distribution of the eddy heat flux at 850 hPa during the 10-day period before the SSW onset. As a product between the eddy meridional wind and eddy temperature fields (i.e., \(V' T'\)), the meridional eddy heat flux is proportional to the local vertical wave activity. A positive (i.e., poleward) eddy heat flux corresponds to a local upward wave activity. Fig 6.11a shows the CNTL eddy heat flux as line contours and the eddy heat flux response to the sea ice loss, \((V' T')_R\), as filled contours. Before the SSW
onset, the CNTL poleward eddy heat flux is most dominant over the Northwestern Pacific between 40°N and 60°N. The imposed sea ice loss significantly enhances this dominant heat flux region, leading to an enhanced upward wave activity over the Northwestern Pacific near 130°E (panel a). Along the 60-degree latitude circle, other weaker patches of CNTL poleward eddy heat flux exist over the Euro-Atlantic and North America regions. The Euro-Atlantic patch near the Prime Meridian also becomes strongly enhanced due to $(V^+T^+)_R$ that extends eastward well in central Europe. When zonally averaged, the associated $(V^+T^+)_R$ contributes largely to the near-surface upward EP flux response in Fig 6.9b.
Fig 6.11. Components of poleward eddy heat flux. Poleward eddy heat flux in CNTL (line) and its response (R) to sea ice forcing (filled) (a), decomposed poleward eddy heat flux $V_R^* T_{CNTL}^*$ in (b), $V_{CNTL}^* T_R^*$ in (c), and $(V_R^* T_R^*)_R$ in (d) (EQBO). The underlying eddy fields of $T_{CNTL}^*$ (line; in K), $V_R^*$ (filled; in m s$^{-1}$) are in (e), $V_{CNTL}^*$ (line; in m s$^{-1}$), $T_R^*$ (filled; in K) are in (f) at 850 hPa. All variables are averaged from day -10 to day -1 prior to the onset. Line contour interval is 20 K m s$^{-1}$ in (a), 2 K in (e) and 2 m s$^{-1}$ in (f). Stippling in (a) indicate statistical significance at the 95% level.

Fig 6.12 shows the response in the three-dimensional structure of (stationary) PW activity (as Plumb flux) averaged between days -10 to -1 before SSW onset. The vertical activity is shown as contours at 850 hPa and the horizontal activity as vectors at 500 hPa. Alternatively, the global distribution of the product between the eddy meridional wind and eddy temperature fields (i.e., $V^* T^*$ or eddy heat flux) helps assess the local origin of vertical wave activity. Positive eddy heat flux corresponds to local upward wave propagation. It confirms the response in local upward wave activity over Northwestern Pacific near 130°E and the Euro-Atlantic patch near the Prime Meridian discussed in Fig 6.11a.
Fig 6.12. Changes in wave activity induced by Pacific-sector Arctic sea ice loss (EQBO). The response in vertical (contours; in m$^2$ s$^{-2}$) 850-hPa and horizontal 500-hPa wave activity flux (vectors; in m$^2$ s$^{-2}$). The latter are normalized and their colors show their magnitude. All variables are averaged from day -10 to day -1 prior to SSW onset. Stippling indicates statistically significant regions of the line contours at the 95% level. The missing values are due to the topography. The transparent red patches indicate the enhanced poleward eddy heat flux response shown in Fig 6.11a.

To relate the wind and temperature responses to the region of enhanced upward wave activity, we decompose ($V^* T^*$)$_R$ into its linear and nonlinear components (Nishii et al., 2011; Hoshi et al., 2017), as described in sub-section 5.3.2.3. The decomposition reveals that the amplified upward wave activity response arises mainly from the linear term $V^*_R T^*_{CNTL}$ (compare panels a & b).

Fig 6.13 details the time evolution of each component, confirming the predominance of the $V^*_R T^*_{CNTL}$ term prior to the onset. Over the Northwestern Pacific, the enhanced upward wave activity peaks at day -6 from the coupling between the CNTL cold temperature and southward wind response. Over the Euro-Atlantic region (panel c), the coupling between the CNTL warm temperature and northward wind response produces the strong total eddy heat flux response about 5 days later (day -1), which is consistent with a downstream propagation of wave activity seen in Fig 6.12 and ultimately the amplification of the WN2.
Fig 6.13. Components of poleward eddy heat flux during SSW (EQBO). The time series of 850-hPa poleward eddy heat flux difference (averaged over 120°E–140°E to represent the Northwestern Pacific region; 40°N–70°N) along with their linear components $V_{\text{CNTL}}^* T_R^*$, $V_{R}^* T_{\text{CNTL}}^*$ and nonlinear component $(V_{R}^* T_{R}^*)_R$ in (a), eddy fields of $V^*_R$, $T^*_R$ (in m s$^{-1}$) and $T_{\text{CNTL}}^*$, $T^*_R$ (in K) in (b). (c, d) are the same as (a, b) but averaged over 340°–360°E covering the Euro-Atlantic region; 40°N–70°N. Open circles and squares indicate statistical significance at the 95% level.

The upper panel of Fig 6.14 shows the 850-hPa geopotential height. For CNTL, significant cyclonic anomalies (bluish areas) appear over the Northeastern Eurasia and the North Pacific as well as over Eastern Canada (panel a). Similar patterns of height anomalies are found to be precursors of SSW events (Nishii et al., 2011; Orsolini et al., 2018). The prescribed sea ice loss greatly strengthens these cyclonic anomalies around the 60-degree latitude circle (panel b), resulting in the cyclonic response over the North Pacific (panel c). The circulation around this cyclonic response produces a southward flow near 130°E, consistent with Fig 6.11e, responsible for the enhanced $(V^* T^*)_R$ over the Northwestern Pacific region. This circulation likewise produces a northward flow near the Date Line that
could bring warmer-wetter air poleward from the lower latitudes. The prescribed sea ice loss also strengthens the anticyclonic anomalies over the Northeastern Eurasia, leading to a significant anticyclonic response within the Arctic circle between Greenland and Scandinavia. The associated northward flow response near the Prime Meridian (shown on Fig 6.11e) is responsible for the enhanced \((V^*T^*)_R\) over the Euro-Atlantic region (Fig 6.11a). The overall 850-hPa height response has a large WN2 component.

The longitude-altitude structure of the eddy geopotential height prior to the SSW onset is illustrated in the lower panels of Fig 6.14, along with the wave activity vectors. This structure is averaged between 55°N and 60°N, covering the region of strong upward wave activity (Fig 6.11a). Below 500 hPa, the geopotential height anomalies and responses exhibit a relatively weak phase tilt with altitude, suggesting a barotropic structure in the lower troposphere. Panel f clearly shows the WN2 characteristic in the height response throughout the troposphere, with a significant cyclonic response between 130°E and 180°E and anticyclonic response around 0°E with upward wave activity, consistent with the strong \((V^*T^*)_R\) region at 850 hPa (Fig 6.11a). In the stratosphere, the height anomalies have a pronounced westward tilt, associated with stronger wave activity flux at these altitudes prior to SSW (Fig 6.9).
Fig 6.14. Composite geopotential height and wave activity flux (EQBO). 850-hPa longitude-latitude cross section of geopotential height anomaly in CNTL (a), LOW (b), and response (c), 55°N–60°N longitude-height cross section of the eddy component of geopotential height anomalies (contours) and wave activity flux anomaly (vector) in CNTL (d), LOW (e), and response (f) averaged from day -10 to day -1 prior the onset. Stippling and shading indicate statistical significance at the 95% level. The black triangles mark the longitudinal edges of the sea ice loss region.

Fig 6.15 shows the 500-hPa geopotential height. Indeed, the height anomalies and the responses at 850 hPa extend to 500 hPa. In the following section, we attempt to examine the dynamical mechanism for the cyclonic response over the North Pacific.
6.4 Mechanism for the Cyclonic Response over the North Pacific.

To understand the cyclonic response responsible for the enhanced \((V^* T^*)_R\) over the Northwestern Pacific, we examine the tropospheric temperature and meridional wind structures prior to SSW onset. In Fig 6.16, the filled contours show the temperature anomalies in CNTL and LOW and the temperature response \(T_R\). At high latitudes (averaged between 60°N and 80°N), the longitude-altitude temperature distribution in CNTL indicates anomalous cooling westward of 190°E mainly below 7 km, with an overlying warm layer eastward of 170°E (panel a). In LOW, near-surface warm anomalies appear between 180°E and 240°E, consistent with the anticipated effect of sea-ice loss and extending upward toward the overlying warm layer (panel b). Averaged between 150°E and 210°E, the latitude-altitude temperature distribution (lower panel) reveals the upward extension of warm anomalies in LOW poleward of 70°N. Hence, the high latitude warming response to the prescribed sea ice loss near the Date Line deepens as the SSW develops, extending well into the lower troposphere (panels c & f). Additionally, a cooling
response occurs at high latitudes west of the Date Line and at mid-latitudes south of 60°N, due to cold air advection on the western flank of the cyclonic anomaly (Fig 6.10).

From the thermal wind relationship, a positive zonal (meridional) temperature gradient is associated with a meridional (zonal) wind increasing (decreasing) with altitude. Therefore, anomalous meridional wind vertical shear is positive (negative) where the zonal temperature gradient is positive (negative), and anomalous zonal wind vertical shear is positive (negative) where the meridional temperature gradient is negative (positive). As such, the existence of zonal and meridional temperature gradients along the lower boundary implies the existence of vertical shear of the horizontal winds. These horizontal winds drive the anomalous wave activity fluxes. Line contours in Fig 6.16 correspond to the meridional wind (V) and zonal wind (U) in purple and green,
respectively. The warming response triggers a significant northward flow response ($V_R$). At about the Date Line, the positive vertical shear in $V_R$ is consistent with the eastward temperature gradient (panel c). At about 130°E, the negative vertical shear in $V_R$ coincides with the westward temperature gradient. The induced warming also elicits a strong westward flow response ($U_R$; panel f). This negative vertical shear structure of $U_R$ is above the warming response where the underlying temperature gradient in the northward direction is positive. In summary, the near-surface warming response results in a deep cyclonic response over the North Pacific.

### 6.5 Planetary Wave Interference

To assess the tropospheric precursors to SSWs, linear stationary wave interference theory has been used in past studies (Nishii et al., 2009; Smith & Kushner, 2012). When the forced response is in-phase (out-of-phase) with the climatological PW pattern in the troposphere, upward wave propagation from the troposphere to the stratosphere is enhanced (suppressed). The left column of Fig 6.17 shows the climatological WN1 (line contours) and WN2 (filled contours) components of the extended wintertime 500-hPa geopotential height. The corresponding anomalous field in CNTL and LOW are illustrated in the remaining columns. For SSW events in CNTL, the anomalous WN1 and WN2 patterns tend to be in-phase with climatological wave patterns, albeit the patterns are now shifted poleward for SSW. The positive and negative geopotential height of the WN1 pattern at around 60°N correspond to the deep anticyclonic and cyclonic anomalies identified in Figs 6.14a and 6.15a. The wave structures in LOW are similar to those in CNTL but the amplitudes are much larger, especially for WN2 which has more than doubled.
Moreover, the wave fields are extending further equatorward. Thus, the warming caused by sea ice loss intensifies both the WN1 and WN2 components of the climatological geopotential height field in the troposphere, enhancing the vertical wave flux in the troposphere and lower stratosphere (as seen in Figs 6.8f, 6.9b, 6.11a, and 6.14f). In the next section, we examine the impact of Arctic sea ice loss on cold air outbreaks.

6.6 Cold Air Outbreaks

The time series of AO index and surface air response following the onset are shown in Fig 6.18. Consistent with the more intense and prolonged SSW, a reinforced bias toward the negative AO phase following the onset (panel a) leads to more intense CAOs over Eurasia and North America in presence of sea ice loss (panel b).
Fig 6.18. Cold Air Outbreaks. Time series of AO index composite in CNTL and LOW (a) and composite of surface temperature (K) response (CNTL-LOW) averaged from day 0 to day +30 after the onset (b). Stippling indicates statistical significance at the 95% level.

6.7 Observation

To affirm model results, we examine SSW composites in the JRA55 reanalysis dataset (1958-2021). Due to the poor quality of observation pre-satellite era, only dataset after 1979 is used. Same SSW identification method as used in the model finds 27 SSW events. Although the criterion for choosing SSW events is not exactly same with previous studies (Charlton and Polvani, 2007; Cao et al., 2019), identified onset dates for each event are close to those studies.

These 27 SSW events are then divided according to different QBO phase, based on the QBO index, defined as zonal-mean zonal wind. averaged over 5°S-5°N, 30hPa-50hPa, and 30 days prior to onset. In all, 16 SSWs are found during EQBO and further categorized based on the SIC time series detrended using EMD (details in sub-section 5.3.5). Years with low (high) autumnal Arctic sea ice are identified when the standardized, detrended Pacific-sector Arctic SIC values averaged in September are less (more) than zero.
Fig 6.19 summarizes the characteristics of these 16 EQBO SSWs in JRA55 with respect to the 100-hPa EPz for events corresponding to high-SIC years (represented by 8 gray bubbles) and low-SIC years (8 red bubbles). For low autumnal sea ice years, SSW events tend to be of longer duration (larger bubble), more intense (based on stronger wind reversal), and associated with larger upward wave flux. Given such few observed events, the changes in the SSW characteristics are not statistically significant. The SSW characteristics in the observational records are influenced by the several factors, such as ENSO, 11-year solar cycle and QBO. Nevertheless, the general response in SSW characteristics due to sea ice loss under the EQBO bear some similarity to the model results shown in Fig 6.5.

![Graph](image)

**Fig 6.19. SSW characteristics (JRA55). Scatter plot of 100-hPa EPz (vertical EP Flux component) averaged between 45°N–75°N, 5 days prior to the onset (x-axis unit in 10^2 m^2 s^-2) and SSW intensity (y-axis unit in m s^-1) (a). LOW**
(red) and HIGH (grey) correspond to SSW events during winters with low or high sea ice in the Pacific sector in autumn. The mean values along each axis are shown as dashed lines.

Analogous to Fig 6.8, the time evolutions of the zonal-mean zonal wind anomaly at 60°N for the high and low ice SSW composites during EQBO are shown in Fig 6.20. In low-SIC years, the observed wind reversal is stronger following the SSW onset compared to high-SIC years.

Fig 6.20. Composite zonal-mean zonal wind (U) during SSW. The time-height cross section of 60°N zonal-mean zonal wind anomaly (in m s\(^{-1}\)) composite with high (a) or low (b) sea ice in the Pacific sector in autumn. Grey shading indicates statistical significance at the 95% level.

The surface air response following the onset (JRA55) is shown in Fig 6.21. Note the surface air temperature has been detrended using EMD. The stronger wind reversal following the
SSW onset leads to more intense CAOs over North America in low-SIC years. The warming over Eurasia could be caused by sea ice loss over Atlantic sector and/or other interval variabilities in the climate system. Overall, the observed structures are similar to our model results.

Previous studies have shown that sea ice loss over Atlantic sector during early winter enhances upward PW propagation (Kim et al., 2014; Hoshi et al., 2019). We tried to address the potential impact of sea ice loss over Atlantic sector on SSW in an anomalous way that we did for the Pacific sector. However, in doing so, only one SSW case for high and one for low Atlantic SIC were found.

**6.8 Discussion and Summary**

Projections about the potential changes in SSWs resulting from climate change have
largely focused on the frequency of SSW occurrence. There is disagreement in the current literature (Mitchell et al., 2012; Schimanke et al., 2013; Ayarzagüena et al., 2018) which could reflect opposing Arctic stratosphere model responses to different aspects of climate change. Through idealized climate model experiments that exclude the effects of ENSO, solar activity, global warming, and ozone loss, we provide supporting evidence that autumnal sea ice loss over the Chukchi-Bering Seas (i.e., the Pacific sector) modifies the characteristics of SSWs rather than altering their frequency of occurrence. Namely, the enhanced upward wave activities (especially, those from WN2) and associated stronger westward deceleration in the stratosphere significantly extends the SSW duration and strengthens the stratospheric wind reversal. Consistent with the more intense and prolonged SSW, a reinforced bias toward the negative Arctic Oscillation phase following the onset leads to more intense CAOs over Eurasia and North America in presence of sea ice loss.

Through poleward heat flux decomposition, we demonstrate that the interaction of the sea ice loss with a precursory cyclonic anomaly over the North Pacific during the pre-onset stage of SSW is the critical factor in amplifying the near-surface wave-activity response (especially, over the Northwestern Pacific region), despite the seasonal-mean thermal response to the sea ice being shallow. The seasonal timing (i.e., autumn or winter) of the sea ice anomaly appears less relevant. As summarized in Fig 6.22, the near-surface warming response results in a deep cyclonic response over the North Pacific via thermal wind balance. This circulation of this cyclonic response extends into the lower troposphere and enhances the height anomalies before SSW onset.
Fig 6.22. The mechanisms associated with sea ice loss. A schematic of the wind response at surface (short arrows) and 850 hPa (long arrows) in colors and styles shown in Fig 6.16c & 6.16f, along with the CNTL 850-hPa geopotential height anomaly as line contours (given at -23 m and -20 m; same as filled contours in Fig 6.14a) and surface temperature response as filled contours (every 1 K interval).

We speculate that the amplified near-surface response of upward wave activity near the Northwestern Pacific region may induce a downstream wave activity across North America and into Eastern Europe, as suggested by zonal vectors in Fig 6.12. This downstream response may lead an increasingly WN2 characteristics in the geopotential height response and the amplified upward wave activity response of the Euro-Atlantic region that appears 5 days after the response over the Northwestern Pacific region, as evident in Fig 6.13. The greatly intensified WN2 components of the climatological 500-hPa geopotential height field and the WN2 vertical wave flux response in the troposphere may be associated with this downstream propagation.

The presented geopotential height and surface cooling patterns (Fig 6. 15c and Fig 18b)
are similar to another numerical model result (Cohen et al., 2021) that forced with sea ice loss over the Chukchi-Bering sea (see their Fig S13). Although these authors argued that sea ice loss over the Pacific sector has little impact on the stratospheric polar vortex stretching, it does seem to influence the characteristics of SSW events in our model. Given the expected reduction of Arctic sea ice in a warmer future climate (especially, in the Pacific sector), the future SSW characteristics could result in more extreme wintertime CAOs over continental areas. Hence, the autumnal sea ice may be regarded as a precursory factor for extreme weather events in the following winter.

These relationships between autumnal Arctic sea ice loss over the Pacific sector and SSW events may be model-dependent (e.g., regarding the inclusion of the stratosphere or the QBO) and forcing-dependent (e.g., how exactly the sea ice forcing is specified). As noted in the Chapter 3, various models and setups have been employed to understand the impact of sea ice loss on the stratosphere.

To isolate the regional impact of sea ice loss, our experiment setup focused solely on the impact of the reduced sea ice over the Pacific sector in a rather idealized manner. In reality, the atmospheric response over this region will likely interact with the response due to sea ice retreat over the Atlantic sector. Such interaction might contribute to the weak changes of SSW characteristics in JRA55. Sea ice retreat over the Atlantic sector can likewise trigger enhanced upward wave activities (Hoshi et al., 2019; Zhang et al., 2020). Since the sea ice over these two sectors varies relative to one another, how the interactions of these responses might impact SSW characteristics requires further studies.

In summary, we demonstrate that, during EQBO, autumnal sea ice loss over the Chukchi-
Bering Seas modifies the characteristics of SSWs by increasing their duration and intensity, yet without necessarily altering their frequency of occurrence. The results are based on idealized climate model experiments that exclude the effects of ENSO, solar activity, global warming, and ozone loss. Under these conditions, the near-surface warming induced by the sea ice loss persists into winter and deepens as the SSW develops. The resulting temperature contrasts foster a deep cyclonic circulation over the North Pacific that elicits an enhancement of WN2 upward wave activities into the polar stratosphere, which extends the SSW duration and strengthens the stratospheric wind reversal. Consistent with the stronger and prolonged SSWs, CAOs become intensified over Eurasia and North America following the SSW onset, with potentially strong societal impact.
7. The Impact of Internal Variabilities on Stratospheric Response to Arctic Sea Ice Loss

This chapter attempts to elucidate the impact of internal variabilities on the atmospheric response to sea ice loss. Considered internal variabilities are QBO and MJO. The focus is first on SSWs response to Arctic sea ice loss during WQBO. In the previous section, our experiment setup is strictly associated with an EQBO persisting during the entire winter to isolate the QBO influence. Alternatively, analogous experiments are performed for WQBO to understand the role of QBO. Labe et al. (2019) found that there is little-to-no stratospheric response during WQBO. However, it is the only study that separated the QBO phases when analyzing the atmospheric response to sea ice loss. The changes of SSWs characteristics in response to Arctic sea ice loss during WQBO warrant further examination.

The second focus of this chapter is to identify the impact of MJO on SSWs response to Arctic sea ice loss. In our idealized climate model experiments, the effects of ENSO, solar activity, global warming, and ozone loss have been excluded. However, tropical variability MJO is internally generated in WACCM6. Since SSW’s precursor over North Pacific could be excited by MJO phase 7, the role of MJO in modulating stratospheric response to sea ice loss will be explored.

7.1 SSWs in Response to Arctic Sea Ice Loss for WQBO

7.1.1 Seasonal Surface Response

We first examine the seasonal evolution of the surface air temperature, sensible heat flux,
and latent heat flux responses (Fig 7.1). The anomalous upward turbulent heat flux is mostly contributed by sensible heat flux over the sea ice loss region, which is similar with the pattern for EQBO. In addition, the magnitude of surface warming is comparable with that during EQBO (Fig 6.2). Overall, there is not much difference in the seasonal evolution of surface response during different QBO phase.

![Seasonal surface warming induced by Pacific-sector Arctic sea ice loss during WQBO. Surface air temperature (a, d, g, j), sensible heat flux (b, e, h, k), and latent heat flux (c, f, i, l) difference (LOW-CNTL) averaged in September–October, November–December, January–February, and March–April. Line contours on top of surface air temperature are sea ice concentration difference (in percentage) between LOW and CNTL experiment (contour values are 2.5, 5, 10, and every 10 thereafter). Stippling indicates statistical significance at the 95% level.](image)

However, the warming over sea ice loss region remains confined below 850 hPa
throughout the autumn to spring (Fig 7.2), which differs from the structure for EQBO (Fig 6.3). The different structure of induced warming might lead to a different SSWs response to sea ice loss during WQBO. In the following section, we attempt to examine the impact of Arctic sea ice loss on SSW characteristics during WQBO.

Fig 7.2. Seasonal air temperature difference (LOW-CNTL) averaged over 150°E-210°E in September–October, November–December, January–February, and March–April during WQBO. Stippling indicates statistical significance at the 95% level.

7.1.2 SSW Characteristics Response

During WQBO, the number of SSW events that occurred among the ensemble members for CNTL and LOW are 27 and 43, respectively (Fig 7.3). Most SSW onset dates are found between January and March, with the mean onset date being February 5th in CNTL and
February 9\textsuperscript{th} in LOW. The SSW onset dates are distributed more evenly during the extended winter (Nov-Mar) in LOW compared to CNTL.

\begin{figure}[h]
\centering
\includegraphics[width=0.5\textwidth]{figure7.3}
\caption{Onset dates of SSW events in the model simulations (WQBO). The number of SSW events that occurred among the ensemble members for CNTL (a) and LOW (b) are 27 and 43, respectively.}
\end{figure}

Similar with Fig 6.5 for EQBO, the SSW characteristics in CNTL (gray) and LOW (red) during WQBO are summarized in Fig 7.4. In contrast to EQBO, the frequency is significantly changed in LOW. Although the mean 100-hPa EPz, SSW intensity, as well as the SSW duration increase in response to sea ice loss, the changes of these characteristics are not significant. Overall, the autumnal sea ice loss over the Pacific sector triggers more frequency SSW events but does not impact its intensity during WQBO.
Fig 7.4. Changes in SSW characteristics induced by Pacific-sector Arctic sea ice loss (WQBO). Scatter plot of the 100-hPa EP$_z$ (vertical EP Flux component) averaged between 45°N–75°N, 5 days prior to the onset (x-axis unit in 10^{-2} m^2 s^{-2}) versus SSW intensity (measured as a zonal-mean zonal wind deceleration, y-axis unit in m s^{-1}). Bubble size indicates the SSW duration in days, with a reference size provided. The mean values along each axis are shown as dashed lines. The split SSW frequency is the ratio of split SSW events to the total number of SSW events. Bold text indicates statistical significance at the 95% level.

Fig 7.5 shows the composites of the geopotential height at 100 hPa and 10 hPa at SSW onset (day 0) for CNTL, LOW and their difference indicating the response to sea ice loss. In general, the shape of polar vortex is alike between CNTL EQBO and CNTL WQBO (compare Fig 6.7a and Fig 7.5a). In response to sea ice loss, the 100-hPa geopotential height is significantly increased from Eurasia to the Date Line, along with a localized negative height (albeit insignificant) response over the North Atlantic. At 10 hPa, the polar vortex is centered over Northern Atlantic, with large positive height anomalies over Europe. In contrast to EQBO, there is no deepening of the polar vortex during SSW in
response to the sea ice loss (panel f).

As illustrated in Fig 7.3 and 7.4, the number of SSW events is significantly increase in response to sea ice loss, whereas the increase of mean upward wave flux prior to SSW onset is insignificant. Fig 7.6 details the composite time evolution upward wave flux and wind structure. Comparing Fig 7.6a with Fig 6.8a, we notice that in CNTL prior to SSW onset (from day -20 to day 0), the anomalous upward wave activity is stronger during WQBO than EQBO, which leads to the stronger stratospheric wind reversal. In addition, following the onset, there is a stronger troposphere-stratosphere coupling during WQBO, as indicated by the persistence of significant and deep negative wind anomaly to day 40.
However, this enhanced coupling must be taken with caution since there are fewer SSW events in CNTL WQBO composite. At around day -10, the upward wave flux is significantly weakened in response to sea ice loss, which leads to the stronger stratospheric zonal-mean zonal wind in 10 days prior to onset (panels c & f). As an effective wave guide, more eastward wind enhances upward wave activity in 5-day period prior to onset, which perturbs the polar vortex and makes SSW events more likely to occur. The enhanced upward wave flux prior to SSW onset is small and only persists few days leading to an insignificant increase in SSWs duration and wind reversal intensity. To understand why the upward wave flux is weakened at around day -10, we examine the temperature, wind structures and precursors associated with SSWs in the next sub-section.

Fig 7.6. Composite zonal-mean zonal wind ($U$) and vertical EP Flux component ($EP_z$) during SSW (WQBO). The time series of 100-hPa $EP_z$ anomaly (averaged in 45°N–75°N), along with their decomposition in zonal wavenumbers 1 and 2 (a–c; in 10^{-2} m^2 s^{-2}), along with the time-height cross section of 60°N $U$ anomaly (d–f; in m s^{-1}). Left column corresponds to the CNTL experiment, medium column LOW experiment, and right column the response (i.e., LOW-CNTL). Open circles (a–c) and grey shading (d–f) indicate statistical significance at the 95% level.
7.1.3. Precursors Prior to SSW Onset

Temperature and horizontal temperature advection response at 850 hPa, 500 hPa, and 100 hPa averaged prior to the onset are shown in Fig 7.7. Unlike features during EQBO (Fig. 6.10), the warming over the sea ice loss region during WQBO does not extend vertically into mid-troposphere prior to SSW onset. Below 500hPa, the mid-latitudes experience warming responses, which is associated with warm advection. In the upper troposphere (100 hPa; panel c), significant warming occurs more toward Northern Eurasia and cooling is over central Pacific and Greenland. Enhanced warm/cold advection is responsible for these responses.

Fig 7.7. Changes in temperature induced by Pacific-sector Arctic sea ice loss (WQBO). The LOW-CNTL differences in temperature (filled contours; in K) and horizontal temperature advection (line contours; cold advection as dashes, in $10^{-6}$ K s$^{-1}$) at 850 hPa (a), 500 hPa (c), and 100 hPa (e). Stippling indicates statistically significant regions of the filled contours at the 95% level. The missing 850-hPa values are due to topography.

The tropospheric temperature structures prior to SSW onset are shown in Fig 7.8. At high latitudes (averaged between 60°N and 80°N), the longitude-altitude temperature distribution in CNTL indicates that anomalous cooling extends further north for WQBO compared to EQBO (Fig 6.16a). In LOW, near-surface warm anomalies appear over a
smaller region between 220°E and 240°E. Averaged between 150°E and 210°E, the latitude-altitude temperature distribution (lower panel) in CNTL reveals that anomalous cooling over 60°N-80°N is stronger for WQBO (panel d) compared to EQBO (Fig 6.16d). In LOW, there is weak upward extension of warm anomalies poleward of 85°N. In the lower troposphere, the temperature response induced by the prescribed sea ice loss spreads out horizontally, consistent with Fig 7.7. At high latitude, there is weak cooling at 180°E sandwiched between the two warming anomalies at 120°E and 220°E (panel c). Over these longitudes, there is weak cooling at 60°N sandwiched between the two warming anomalies at mid-latitude and high latitude (panel f). These temperature responses are associated with warm or cold air advection (Fig 7.7a). Even though the prescribed sea ice loss is the same for EQBO and WQBO, the structure of temperature response is different implying that the background state impacts the evolution of the warming induced by sea ice loss.

**Fig 7.8.** Composite temperature (T) prior to SSW (averaged between day -10 to day -1) (WQBO). Longitude-altitude cross section of T anomaly in CNTL (a), LOW (b), and the response (c) averaged between 60°N-80°N. The black
Triangles mark the longitudinal edges of the sea ice loss region. Latitude-altitude cross section of $T$ averaged over 150°E–210°E is given in (d–f). Stippling indicates statistical significance at the 95% level.

Temperature response could impact wind structure via thermal wind relationship. Next, we attempt to connect the wind response with temperature response. As noted in Fig 7.7 and Fig 7.8, the warming induced by the prescribed sea ice loss does not extend vertically into mid-troposphere but is distributed horizontally prior to SSW onset. The zonal wind response, meridional wind response and temperature response are shown in Fig 7.9. To capture the large significant temperature responses, the longitude-altitude cross section of temperature (panel a) is averaged over 40°N-60°N instead of the polar region. Filled contours in panel b are the same as Fig 7.8f. Overall, at mid-latitudes, strong warming responses are over 90°E-210°E with a weak cooling over 210°E-240°E (panel a).

At about 210°E, the negative vertical shear in $V_R$ is consistent with the westward temperature gradient (panel a). The positive vertical shear in $V_R$ coincides with the eastward temperature gradient at about 120°E. The warming response at around 40°N also elicits a strong westward flow response at around 30°N and a strong eastward flow response at around 50°N ($U_R$; panel b). The negative and positive vertical shear structure of $U_R$ appear where the underlying temperature gradient in the northward direction is positive and negative, respectively. In summary, the warming response prior to SSW onset is not confined over sea ice loss region but extends globally, which modifies the wind structures at midlatitudes.
Fig 7.9. Composite temperature (T), zonal wind (U) and meridional wind (V) response prior to SSW (averaged between day -10 to day -1) (WQBO). Longitude-altitude cross section of T and V response (a). T (averaged between 40°N–60°N) is shown as filled contours and V at 45°N as line contours (every 1 m s⁻¹ interval). The black triangles mark the longitudinal edges of the sea ice loss region. Latitude-altitude cross section of T and U (averaged over 150°E–210°E) are given in (b). Stippling indicates T statistical significance at the 95% level.

The 850-hPa geopotential height prior to the SSW onset is illustrated in Fig 7.10. The precursory patterns associated with SSWs are over the Europe and the North Pacific similar with EQBO shown in Fig 6.14a. The prescribed sea ice loss weakens the latter pattern resulting in the anticyclonic response over North Pacific (panel c). The circulation around this anticyclonic response produces a northward flow near 120°E and southward flow near 210°E, consistent with Fig 7.9. Another anticyclonic response appears just south of Greenland, which is also induced by zonal temperature gradient (not shown).
Fig 7.10. Composite geopotential height (WQBO). 850-hPa longitude-latitude cross section of geopotential height anomaly in CNTL (a), LOW (b), and response (c) averaged from day -10 to day -1 prior the onset. Stippling indicates statistical significance at the 95% level. The black triangles mark the longitudinal edges of the sea ice loss region.

7.1.4 Issues with the WQBO experiments

We note that our WQBO experiment contains some issues. First, the QBO amplitude in WACCM6 is weak. Zonal-mean zonal wind averaged over QBO region in WACCM6 and MERRA2 are shown in Fig 7.11 and Fig 7.12, respectively. As found in Gettelman et al. (2019), the simulated WQBO pattern in WACCM6 does not extend low enough into the lower stratosphere, when compared to observations. Furthermore, during late winter, notable WQBO “hiccup” occur in both CNTL and LOW experiments with the attendant eastward equatorial wind becoming unstable and reduced to the point of nearly reversing (Fig 7.11) – analogous to the observed WQBO during the 2015-2016 winter noted by Newman et al. (2016) (red line in Fig 7.12). Such hiccup tendencies might contribute to an increased frequency of SSW occurrence indicated in Fig 7.4. Finally, during the 1-year model spin-up (Fig 7.11), the equatorial wind tends to evolve quite differently between CNTL and LOW for the WQBO phase, leading to markedly different initial equatorial conditions when perturbations are introduced in the ensemble experiment (as shown in
Fig 5.9). As such, the comparison between the CNTL and LOW might be problematic. In most studies of the sea ice atmospheric response, with exception of Labe et al. (2019), the impact of separate QBO phases is not considered, and the reported results are inferred for the combined QBO phases.

![Figure 7.11](image1.png)

*Fig 7.11. Tropical lower stratospheric zonal-mean zonal wind simulated in WACCM6. Wind values averaged between 30 hPa–50 hPa, 5°S–5°N band in the CNTL (black line) and LOW (red line) experiments during WQBO. Vertical black line represents January 1st, 1853.*
7.1.5 The Role of QBO

Despite the caveats noted above, the influence of EQBO may nevertheless play an important role for our results, in that it favors geopotential anomalies in the lower troposphere that projects on precursory patterns attributed to SSWs. Through its influence on tropical convection and enhancement of upward PW propagation into the stratosphere, the QBO can influence the coupling between the troposphere and stratosphere (Yamazaki et al., 2020). Fig 7.13 shows the 500-hPa geopotential height in CNTL and the difference between two QBO phases in the CNTL simulation. The results are shown for November, when the tropospheric pathway of the QBO’s impact is strongest (Yamazaki et al., 2020). However, similar anomalies are apparent throughout the extended winter. We note that, in CNTL, the key precursory cyclonic patterns associated with SSWs (identified over Eastern Eurasia and the North Pacific) are enhanced in EQBO compared to WQBO. In addition, the structure of the temperature response is different even though the prescribed sea ice loss is identical for EQBO and WQBO. These results imply that the background state, conditioned by the QBO, may influence the response of SSWs to the sea ice loss. While difficult, differentiating the QBO phases (whether in numerical experiment setup or in composite based on models/observational results) is essential in understanding the stratospheric response to sea ice loss.
7.2 The Impact of MJO

After correcting for the convectively coupled coherent wind and precipitation features propagation direction, WACCM6 has markedly improved its simulation of the MJO variability (Danabasoglu et al., 2020). In this section, we examine the composites based on MJO phase to explore the impact of MJO on SSWs response to Arctic sea ice loss. To visualize the horizontal structure of MJO during its life cycle, composite outgoing longwave radiation (OLR) anomalies in CNTL EQBO for different MJO phases are shown in Fig 7.14. The anomalies here band-pass filtered for the 20-90 day period to extract intraseasonal variability associated with the MJO. Negative OLR anomalies correspond to regions of enhanced convection. MJO is asymmetric about the equator and disrupted by tropical continent. In phase 1, enhanced convection grows over Africa. In the subsequent phases, the enhanced convection further develops and moves eastward. The strongest
convection anomaly occurs in phase 3 and appears over the Indian Ocean. After the enhanced convection anomaly reaches the western Pacific in phase 5, the suppressed convection anomaly starts to grow over Africa and repeats the process of the enhanced convection. The OLR anomalies in CNTL WQBO are similar (not shown) but are weaker compared to CNTL EQBO consistent with the results of Son et al. (2017). Past studies have shown that the MJO phases with the dipole convection structures over eastern tropical Pacific (phase 3 and 7) have the most impact on the extratropics and stratosphere (Lin et al., 2009; Garfinkel et al. 2012). Therefore, phase 3 and phase 7 are mainly discussed.
First, we focus on the extratropical circulation anomaly associated with the MJO. To better show its evolution, lagged composites are performed. Shown in Fig 7.15 are the lag
composites of 300hPa geopotential height anomaly at day 0, day 5, day 10, day 15 after the MJO phase 3 for EQBO experiments. For CNTL (left column), the simultaneous composite (day 0; panel a) reveals the formation of a positive height anomaly over the North Pacific centered at 45°N 180°E, as found by Matthews et al. (2004) and Lin et al. (2009). Together with the negative height anomaly over Alaska and the positive anomalies over eastern North America, these circulation anomalies form a Rossby wave train propagating from the North Pacific to North America and trigger a negative phase of the Pacific North American (PNA) pattern. As the Rossby wave train propagates, at day 5, the wave pattern exhibits downstream dispersion, with the North Pacific, Alaska centers weakening and eastern North America center intensifying downstream. In addition, a negative anomaly develops over North Atlantic. After day 5, the anomalies in the North Pacific and Atlantic move eastward and weaken. In response to the prescribed sea ice loss (right panels), height anomalies are stronger implying that the Rossby wave train excited by the MJO convection is enhanced.
Fig 7.15. Composite 300hPa geopotential height anomalies (m) at day 0, day 5, day 10, day 15 after the MJO phase 3 for EQBO CNTL (a) and LOW (b) experiments.

Same figure for WQBO is shown in Fig 7.16. In CNTL, the height anomalies are stronger during EQBO (compare Fig 7.15a with Fig 7.16a), which is consistent with the observations of Hood et al. (2020). Possible mechanisms include reduced static stability in the upper troposphere lower stratosphere and weaker vertical wind shear across the tropopause during EQBO (Yoo and Son, 2016). However, the QBO-MJO relationship is not well simulated in some models, which may be due to weak QBO related temperature anomalies around the tropopause (Lee & Kingaman, 2018; Kim et al., 2020; Lim & Son, 2020; Richter et al., 2020; Martin et al., 2021). Another mechanism highlights the importance of cloud-radiative feedbacks owing to changes in high cirrus clouds induced by the QBO (Son et al., 2017; Sakaeda et al., 2020). The deficiencies in simulating clouds through parameterized convection in some models might lead to biases in the QBO-MJO relationships in simulations.

In addition, the height anomalies over Atlantic extend more eastward and are stronger for WQBO. These anomalies bear likeness to the pattern associated with the positive NAO and are consistent with the findings of Feng and Lin (2019). These authors suggested that the subtropical anomalous westerly wind in the North Pacific associated with WQBO provides a favorable environment for the MJO-induced extratropical Rossby wave to propagate. In response to sea ice loss, the Rossby wave train is enhanced.
For the MJO phase 7 (Fig 7.17 and Fig 7.18), the anomalies are similar with MJO phase 3, but with opposite signs. At day 0, the negative anomalies over North Pacific, eastern North America, and the positive anomaly over Alaska are akin to the positive PNA pattern. The MJO-PNA pattern features strong height signals near the subtropical jet exit (Mori & Watanabe, 2008; Zhou et al., 2020). The barotropic energy conversion near the jet exit regions in the Pacific Oceans allows waves to effectively extract energy from the zonally asymmetric climatology (Mori & Watanabe, 2008; Simmons et al., 1983). During EQBO, at day 5, the anomalies over North Pacific, Alaska, eastern North America, and North Atlantic develop as the Rossby wave train propagates. After day 5, the anomalies over the Pacific weaken and over the Atlantic intensify. During WQBO, the height anomalies over North Pacific are weaker relative to EQBO, whereas over Atlantic, the height anomalies are stronger and like a positive NAO. In response to sea ice loss, Rossby wave train is enhanced. In summary, the MJO phase 3 and 7 produce a stronger wave train during
EQBO compared to WQBO and in response to Arctic sea ice loss.

Garfinkel et al. (2012) found that, during the 12-day period preceding SSWs, the MJO phase 7 occurs more than twice compared to its climatological boreal winter frequency.

This is likely because the negative height anomalies over North Pacific after MJO phase 7
project onto the precursory patterns associated with SSWs. To demonstrate the impact of the MJO on SSWs response to sea ice loss, we examine the occurrence of eight MJO phases in 10 days prior to SSW onset (Fig 7.19). For EQBO, prior to SSWs, MJO phase 7 occurs most frequent. The frequency difference of phase 7 between CNTL and LOW is small, implying that cyclonic response over the North Pacific prior to onset (as detailed in section 6.4) is induced by the prescribed sea ice loss instead of by the MJO in the model. For WQBO, the prescribed sea ice loss weakens the precursory cyclonic anomaly over the North Pacific during the pre-onset stage of SSW (Fig 7.10). As mentioned before, positive height anomalies over the North Pacific are associated with MJO phase 3. If the anticyclonic response to sea ice loss over the North Pacific prior to onset is contributed by MJO, phase 3 is supposed to occur most frequent in LOW compared to CNTL. However, phase 7 and 8 occur most frequent in LOW compared to phase 3 and 4 during WQBO (panel b). Therefore, the anticyclonic response over the North Pacific prior to onset (Fig 7.10c) is not due to MJO during WQBO.

Fig 7.19. The frequency of eight MJO phases occurrences during 10 days prior to SSW onset. Only days with the amplitude of the RMM index exceeding 1.0 are included.
7.3 Discussion and Summary

Through idealized climate model experiments, we explored the impact of QBO on the atmospheric response to sea ice loss. During WQBO, autumnal sea ice loss over the Pacific sector significantly increases SSWs frequency of occurrence but appears to have no impact on SSWs duration and wind reversal intensity. Prior to SSW onset, the anomalous warming induced by Arctic sea ice loss extends globally through horizontal temperature advection. Via thermal wind balance, the near-surface warming response results in a weak anticyclonic response over the North Pacific. This circulation of the anticyclonic response extends into the lower troposphere and weakens the height anomalies before SSW onset. This might explain the weakened upward wave activity around day -10 in response to the prescribed sea ice loss. The precursory patterns associated with SSWs over Europe is slightly strengthened. The cancelling effect might lead to insignificant increase in SSW duration and intensity. These findings are consistent with those of Labe et al. (2019), who found little-to-no stratospheric response during WQBO.

However, our results might not be trustworthy. First, the QBO amplitude in WACCM6 is weak, and the WQBO pattern does not extend low enough into the lower stratosphere. Furthermore, during late winter, notable WQBO “hiccups” occur in both CNTL and LOW experiments, which might contribute to an increased frequency of SSW occurrence. Finally, during the 1-year model spin-up, there is a wind bias between CNTL and LOW for the WQBO phase, leading to markedly different initial equatorial conditions when perturbations are introduced in the ensemble experiment. As such, the comparison between the CNTL and LOW might be problematic.
The impact of MJO on SSWs response to Arctic sea ice loss is also explored in this chapter. We found that QBO and Arctic sea ice loss could impact MJO teleconnection patterns. With regards to QBO, MJO phase 3 and 7 could trigger a stronger wave train during EQBO compared to WQBO. It is probably associated with the reduced static stability in the upper troposphere lower stratosphere and weaker vertical wind shear across the tropopause during EQBO, both of which may result in stronger convective activities. In response to Arctic sea ice loss, MJO phase 3 and 7 also trigger a stronger wave train regardless of QBO phase. Since Arctic sea ice loss decreases the meridional temperature gradient, the subtropical jet shifts southward. We speculate that southward shifted subtropical jet might provide a preferable environment for the stronger extratropical response associated with the MJO. However, the detailed mechanisms require further studies. Finally, we confirmed that the changes of precursory patterns associated with SSWs over North Pacific prior to SSW onset are induced by sea ice loss rather than by MJO.
8. Conclusions

8.1 Overview

The extent of sea ice coverage over the Arctic Ocean has dramatically declined over the past few decades coinciding with cold air outbreaks at mid-latitude during winters. This has stimulated much research into the impact of Arctic sea ice loss on atmospheric circulation. Since the satellite record with high-quality is only available from 1979 (roughly 40 years) and contains the impact of decadal and multi-decadal natural variability, climate model experiments are preferred to help disentangle the dynamic mechanism behind the response to Arctic sea ice loss.

This thesis aimed at better understanding the impact of Arctic sea ice loss over Pacific sector on the wintertime stratospheric dynamics through large-member ensemble experiments using a climate model. While the perturbation experiments were idealized, they, nevertheless, might provide insight to physical mechanisms found within observations. In addition, the roles of internal variability such as QBO and MJO were explored.

8.2 Summary of Results

In chapter 6, the fundamental mechanisms linking the surface processes associated with sea ice loss to the wintertime stratospheric dynamics for EQBO were examined. The sea ice loss over the Chukchi-Bering Seas (i.e., the Pacific sector) modifies the characteristics of SSWs rather than altering their frequency of occurrence. Namely, the enhanced
upward wave activities (especially, those from WN2) and associated stronger westward deceleration in the stratosphere significantly extends the SSW duration and strengthens the stratospheric wind reversal. Consistent with the more intense and prolonged SSW, a reinforced bias toward the negative Arctic Oscillation phase following the onset leads to more intense CAOs over Eurasia and North America in presence of sea ice loss. Through poleward heat flux decomposition, we demonstrate that the interaction of the sea ice loss with a precursory cyclonic anomaly over the North Pacific during the pre-onset stage of SSW is the critical factor in amplifying the near-surface wave-activity response (especially, over the Northwestern Pacific region), despite the seasonal-mean thermal response to the sea ice being shallow. The seasonal timing (i.e., autumn or winter) of the sea ice anomaly appears less relevant. The circulation of this cyclonic response is in thermal wind balance with the warming induced by sea ice loss that extends into the lower troposphere and enhances the height anomalies before SSW onset. The results reveal that Arctic sea ice reduction acts as a factor in modulating upward propagating wave properties during SSWs and resulting in more intense CAOs following SSWs. The impact of sea ice loss on the wintertime stratospheric dynamics for WQBO was explored in chapter 7. For WQBO, in response to sea ice loss, the frequency of SSW occurrence is significantly increased but the SSW duration and intensity remains unchanged. Even though the prescribed sea ice loss is identical in EQBO and WQBO experiments, the structure of near-surface temperature response is different. Prior to SSW onset, the induced weak anticyclonic response over the North Pacific explains the weakened upward wave activity at around day -10 in response to the prescribed sea ice
loss. The precursory patterns associated with SSWs over Europe is slightly strengthened. The cancelling effect leads to insignificant increase in SSW duration and intensity. However, inherent issues related to the WQBO experiments (such as weak simulated WQBO in WACCM6 and the bias between CNTL and LOW from spin-up) might cause the comparison between the CNTL and LOW be problematic.

The important role of the stratospheric background condition related to the QBO was also examined. The EQBO phase in particular favors geopotential anomalies in the lower troposphere that project onto the precursory height patterns of SSWs. The background state, conditioned by the QBO, may likewise influence the response of SSWs to the sea ice loss. While difficult, differentiating the QBO phases (whether in numerical experiment setup or in composite based on models/observational results) may be important in understanding the stratospheric response to sea ice loss.

Finally, we quantified the role of tropical variability related to the MJO. We found that the Rossby wave trains excited by the MJO are enhanced during EQBO and in response to the prescribed sea ice loss. Moreover, we confirmed that the changes of precursor over North Pacific prior to SSW onset are induced by sea ice loss rather than by MJO.

8.3 Further Research

The relationships between autumnal Arctic sea ice loss over the Pacific sector and SSW events may be model-dependent (e.g., regarding the inclusion of the stratosphere or the QBO) and forcing-dependent (e.g., how exactly the sea ice forcing is specified). Thus, various models and setups need to be employed to understand the impact of sea ice loss on the stratosphere in the future.
To isolate the regional impact of sea ice loss, our experiment setup focused solely on the impact of the reduced sea ice over the Pacific sector in a rather idealized manner. In reality, the atmospheric response over this region will likely interact with the response due to sea ice retreat over the Atlantic sector. Such interaction might contribute to the weak changes of SSW characteristics in JRA55. Sea ice retreat over the Atlantic sector can likewise trigger enhanced upward wave activities (Hoshi et al., 2019; Zhang et al., 2020). Since the sea ice over these two sectors varies relative to one another, how the interactions of these responses might impact SSW characteristics requires further studies.

In this thesis, an atmosphere-only model was used and the coupling with the ocean was discounted in our results. Smith et al. (2017) suggested that the background state, impacted by different SSTs between atmosphere-only simulation and ocean-atmosphere coupling simulation, controls the sign of the winter AO response via the refraction of Rossby waves. Toward a better understanding of the linkage between Arctic sea ice loss and atmospheric circulation, using an ocean-atmosphere coupling model simulation is necessary for further studies.
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