ABSTRACT

LONG, XIAOYU. Dynamical heating of the Arctic during the springtime transition and its response to climate change. (Under the direction of Dr. Walter A. Robinson.)

The seasonal transition from winter to spring in the Arctic is observed as an abrupt transition from quasi-steady winter to rapid warming spring. In this thesis research, the author aims to understand the dynamic and thermodynamic mechanisms that are responsible for the abrupt transition and their response to climate change.

An objective two phase linear regression model is used to define the Arctic springtime transition. Through heat budget analysis, it is found that the rapid warming during spring transition is due to the strong eddy heat flux into the Arctic. The variation of eddy heat flux controls the Arctic SAT variability on the sub-seasonal time scale. The variability of eddy heat flux decreases dramatically as the season advances into spring.

The anomalous flow pattern associated with eddy heat flux events is a dipole-like pattern spanning the North Atlantic and the Greenland Sea with a low in the west and a high in the east. The dipole pattern is related to the interaction between a traveling barotropic Rossby wave (wave-1) and the thermodynamic condition at lower levels. The variability of eddy heat flux on sub-seasonal time scales is mainly affected by dynamical factor: Rossby wave dynamics. The inter-seasonal variability and climate change of eddy heat flux are controlled by thermodynamic conditions at lower levels: the land-sea contrast. The dipole pattern extracts energy from the zonal flow through baroclinic conversion. The maintenance of the dipole pattern is by a combination of linear wave dynamics, nonlinear dynamics of synoptic eddies, low frequency eddies and local diabatic processes.

A projection of the future climate shows a less abrupt springtime transition in the
Arctic, because the eddy heat flux is weaker and less variant under climate change. This, in turn, is caused by a change in the land-sea contrast. The overall energy transport into the Arctic is weaker in nearly all latitudes in the simulated future climate.
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Dynamical heating of the Arctic during the springtime transition and its response to climate change

by
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DEDICATION

To my family.
BIOGRAPHY

Xiaoyu Long is originally from Chongqing, China. He received his bachelor degree in Atmospheric Science in 2003 at Lanzhou University and master degree in Meteorology in 2006 at Peking University. After working as a weather forecaster for 6 years in Chongqing Meteorological Observatory, he joined the Department of Marine, Earth, and Atmospheric Sciences at North Carolina State University in August 2012 to start his doctoral study.
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Chapter 1

Introduction

1.1 The Arctic climate system

The Arctic climate system is characterized by its low thermal energy state, close interaction between the atmosphere, the ocean and the land surface, and its sensitive response to climate change, both in the last few decades’ observations and in climate model projections. The most common definition of the Arctic is the region north of the Arctic Circle (66.5°N), which is related to Earth’s obliquity. Some research also uses the 70°N parallel as the boundary of the Arctic, north of which is occupied mostly by the Arctic Ocean.

The Arctic is an energy sink and it loses more energy to the space than the solar radiation it receives (Trenberth and Caron, 2001). Poleward energy transport is required to maintain energy balance in the Arctic. In the high latitudes, atmospheric transport dominates. The atmospheric transport is usually divided into two parts: transport by the zonal mean flow and by the eddies.

Under a changing climate, the Arctic climate system and its coupling with mid-latitudes changes accordingly. How other components contribute to the changing Arctic
is an unsolved scientific question. Recent studies found that even the tropical systems have a potential role in influencing the Arctic climate (Lee, 2012; Lee and Yoo, 2014; Yoo et al., 2014, 2012a,b)

1.2 Seasonality and spring transition in the Arctic

In addition to changes in the mean, changes in the seasonality are of at least equal importance. The annual march of seasons is the most important externally driven cycle in the climate system. The resulting seasonal transition, especially from winter to spring, is a key event for all living beings. A small change in the seasonal cycle of temperature would cause huge impacts on the environment, which is particularly true in high latitude regions. For example, a change in the timing and strength of the spring transition would affect snow-cover and sea-ice melting, and would change the length of the growing season for plants (Cayan et al., 2001; Linderholm, 2006). In the literature of climate science, only a few papers are dedicated to the subject of seasonal cycle.

There is no consensus on how to define a seasonal cycle. There are two popular approaches to determine the timing of seasons. The more common one is a threshold based model wherein seasonal transitions are defined as the times of year when the temperature rises above or drops below some specific value. Under this model, the warming season onset (Spring) will be earlier and the cooling season onset (Fall) will be later if the mean temperature is increased. This approach makes more sense for seasonality problems in biological systems (blooming dates, timing of bird migrations and growing season length) (Schwartz et al., 2006; Sparks and Menzel, 2002) and in components that respond to absolute temperature (snow-cover and sea-ice melting). However, this threshold based method inevitably blends changes in the phase of annual cycle with changes in the annual
mean. An alternative method is proposed by Thomson (1995), who describes the annual cycle by phase and amplitude of the one-year sinusoidal component ($1^{st}$ harmonic) of temperature time series for each year, which has nothing to do with the change in the annual mean. Another, more fancy, method has been proposed to define the seasonal cycle (Straus, 1983), in which the atmospheric temperature time series is projected onto a predefined subset of orthogonal basis functions, but this method is not widely used. However, none of the above approaches is suitable for the purpose of my research, because the high latitude seasonal cycle is far from a sinusoid and my emphasis here is only on the springtime transition. An objective method will be used to define the spring onset; this is discussed in next chapter.

Research using different definitions of the seasonal cycle usually lead to different conclusions. Cayan et al. (2001) found earlier spring onsets in terms of phenological events exist since the late 1970s, in which the phenological definition is equivalent to a temperature threshold definition. On the other hand, Thomson (1995) found a delayed phase of the seasonal cycle and ascribed this to variability of the earth-sun distance and the tilt of earth rotation axis, as well as the possible role of increased greenhouse gas concentration because the delayed phase is beyond the range of natural variability. With the same methodology, Mann and Park (1996) also found a delay in the seasonal phase, and they ascribe it to sea ice loss, which increases open ocean exposure and hence the thermal inertial of the climate system. This mechanism explains some of the phase delay in high latitude ocean simulated by a GCM (Dwyer et al., 2012) but is contradictory to observed early phase on land. Stine et al. (2009) ascribe the early phase on land partly to the trends of decreases in soil moisture based on model simulations (Seager et al., 2007), which is not well verified in observation. Stine and Huybers (2012) further connect the change in the phase of the annual cycle to variability of large scale atmospheric circulation
modes (NAM and PNA). They conclude that the phase is strongly influenced by the spring atmospheric circulation while the amplitude is affected by the winter circulation. Abatzoglou and Redmond (2007) found that, in western North America, change in the atmospheric circulation will produce an asymmetry in seasonal warming, which leads to the change of amplitude of the seasonal cycle. Paluš et al. (2005) found that the variability of seasonality in Europe is correlated with the North Atlantic Oscillation (NAO) and anti-correlated with and El Nino/Southern Oscillation (ENSO) index. In these previous studies, there is more and more evidence showing that the seasonality and its change are heavily modulated by dynamical factors.

In the last two decades, the Arctic has attracted much research interest since it has warmed almost twice as much as the global average under climate change (Graversen et al., 2008). What accompanies the warming in the annual mean is the asymmetry in seasonal mean warming rate. Warming in the winter is more prominent than in summer, which changes the amplitude of seasonal cycle. Some researches ascribe most of the asymmetry of seasonal cycle to differences in boundary conditions (local feedback) and to the ocean’s role in taking heat away from the atmosphere.

The Arctic seasonal cycle itself comprises a complex interplay between external radiative forcing and internal dynamics. It has a large seasonal cycle in solar radiation, from zero in winter to 300 – 400 W/m² in summer, and the seasonal variation of sea ice conditions amplify the seasonality of energy budgets and hence the annual cycle of temperature. Simultaneously, the Arctic thermodynamic state is maintained by a combination of solar radiation, infrared radiation and dynamically driven poleward energy transport, which are all subject to climate change. Though the modification of radiation budgets by local boundary conditions is very important, dynamical heating might also have a crucial role in modulating the seasonal cycle.
The climatology of zonal mean surface air temperature (SAT) shows that the Arctic generally has a larger seasonal cycle than mid-latitudes (Fig.1.1a). Another prominent feature of the Arctic seasonal cycle is that the transition from winter to spring is not gradual (He and Black, 2015; Rogers et al., 1997): the Arctic winter has a steady thermodynamical state and the SAT just fluctuates due to internal variability. The SAT starts to increase rapidly during the spring transition (Fig.1.1). Even in the climatology, it is still clear that the transition from winter to spring in the Arctic is abrupt compared to more gradual transition in mid-latitudes and the subtropics (Fig.A.1).

1.3 Research objectives

This thesis research will start from trying to understand the unique feature of springtime transition in the Arctic shown in Fig.1.1, to investigate the possible dynamical mechanisms underlying it, and finally to make a projection for how the Arctic springtime transition will respond to climate change. The main scientific questions and objectives I will focus on in this research are:

1. Since there is no consensus on how to define a seasonal cycle and the seasonal transition therein, and the existing methodologies are not suitable for current research, the first step is to invent an objective metric to define the springtime transition. The metric should represent the salient characteristics of the Arctic springtime transition as are shown in Fig.1.1.

2. The sharp transition from winter to spring in the Arctic indicates the role of dynamics. The second part of this research is aimed at isolating the possible dynamic and thermodynamic mechanisms that are responsible for the abrupt transition. The
initial attempt will use heat-budget analysis to quantify the direct cause of rapid warming during the transition. Then, possible dynamic or thermodynamic factors that are associated with those important heat budget terms will be investigated. These factors might include systems outside the Arctic.

3. Numerous studies have focused on the response of the Arctic climate system to global warming. Only a few involve changes in the seasonal cycle. None of them has touched the springtime transition problem to my best knowledge. Therefore, the third part will be to make a projection about how the Arctic springtime transition will respond to future climate change. The projection will be based on the mechanisms revealed in last question.
Figure 1.1: Climatology of surface air temperature (SAT): (a) Zonal average and (b) area average north of 70°N. The calculation uses ERA-interim reanalysis.
Chapter 2

Data, model and method

2.1 Data

Although the main theme of this research is using model simulation to investigate the role of atmospheric internal variability in the Arctic springtime transition, the very first step is to verify that the model has the capability to reproduce the salient features of the Arctic springtime transition. Observations are needed.

Due to the poor coverage of traditional observation in the Arctic, which is especially true for the upper air observations, a reanalysis product is chosen as a replacement for the observation. **ERA-Interim** (Dee et al., 2011) produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) is used. Lindsay et al. (2014) compared seven reanalysis products in the Arctic region over the period from 1981 to 2010, and they concluded that three models stand out as being more consistent with independent observations: CFSR, MERRA, and ERA-Interim. He and Black (2015) showed that there is only little difference between MERRA and ERA-Interim when considering the Arctic springtime transition. In this research, thirty years of ERA-Interim data from 1979 to
2008 is used, and the data were downloaded from NCAR/UCAR research data archive website (http://rda.ucar.edu/datasets/ds627.0/).

2.2 Model configuration

The community earth system model (CESM 1.0.5) is used because, as a community model, it is available, supported, and well documented. CESM is a coupled climate model for simulating Earth’s climate system. It is composed of five separate models which are capable of simultaneously simulating the Earth’s atmosphere, ocean, land, land-ice, and sea-ice, plus a central coupler component. For the purpose of this research, the model is configured as F components set (equivalent to CAM5 stand-alone configuration) which includes active atmosphere and land model forced with data ocean and sea-ice model.

The model components used in this research are: CAM-5.1.1 for atmosphere model, CLM4 for land model, and Data Model v8 for ocean and sea-ice model. The grid spacing is 0.9X1.25 realized in a finite volume grid. The parameterization schemes in CESM1.0.5 (CAM5) are pre-set and are not an option for the user to change. The summary of the schemes is list in table 2.2, and the details of the formulation of these schemes can be found in the references therein.

2.2.1 Control and NO-Greenland simulation

To compare with the reanalysis and future projection, a 30-years control simulation is conducted. The stand-alone atmosphere model is forced by monthly mean climatology of SST and sea-ice from 1940 – 1969. The SST and sea-ice data set is a combination of NCAR and Hadley center SST and sea-ice data, and is the default data used as the boundary condition in the CESM community. The tracer gas values are chosen as common
20th century values, and are pre-set in the CESM F2000 component set. I choose these 30 years instead of more recent years, because the influence of global warming signal is not wanted in the control simulation. Caution should be taken when comparing the control simulation results with reanalysis because there are some distinctions between them. First, in the control simulation, the model is not forced by real time SST and sea-ice but by climatology; secondly, the SST and sea-ice used to force the atmosphere are from 1940 – 1969 instead of 1979 – 2008 which is the time period of the reanalysis data. The spring transition, by definition, is on the intra-seasonal time scale, while the global warming signal operates on a much longer time scale. Meanwhile, He and Black (2015) found there is no significant trend in the spring onset date. Therefore, it is still feasible to compare the control simulation with reanalysis if the emphasis is on the sub-seasonal dynamics. This experiment can also serve as a test for the capability of the model to reproduce internal atmosphere variability of interest. By comparing the leading EOF patterns of surface pressure and the envelope function (Nakamura and Wallace, 1990), calculated from reanalysis and control simulation output, the model indeed can generate atmospheric variability (storm-track variability and NAO), which compare with the variability in the reanalysis (not shown).

To investigate the potential role of Greenland topography in the Arctic Springtime transition, an experiment with the Greenland topography removed is conducted. Except for the removal of Greenland topography, other boundary conditions are the same as in the control simulation.
2.2.2 Perpetual simulation

Besides the simulations with full seasonal cycle, three perpetual simulations for each climate state (current and RCP8.5) are carried out. The specific dates for the three perpetual simulation are: January 31\textsuperscript{st}, March 16\textsuperscript{th} and April 30\textsuperscript{th}. As will be shown in the next chapter, the average spring onset date is March 16\textsuperscript{th}, so these three dates represent mid-winter, spring onset and spring respectively. In each of the perpetual simulation, the seasonal cycle of solar radiation is fixed for that day while the diurnal cycle is retained. Other boundary conditions, like SST, sea-ice and aerosol, are also set as fixed values on that day. The benefit of the perpetual simulations is that I can get a clean anomaly field from the seasonal cycle and more data points without running the model for too long.

2.2.3 RCP8.5 simulation

The purpose of the RCP8.5 simulations is to investigate how the spring transition would respond to climate change. The future climate state is set as the last decade of the 21\textsuperscript{st} century. The differences between the RCP8.5 and control simulation are only in the boundary conditions. The SST and sea-ice conditions are averaged over five CMIP5 model projections (Table 2.1). These CMIP5 models are chosen from the best 7 group described in Colle et al. (2013). These 7 models are found to be the best at reproducing mid-latitude variability over the Atlantic region, and we found, in the current climate, that storm-track activity is important to the Arctic springtime transition. Other boundary conditions, including aerosol, and concentrations of radiatively active gases, are all adjusted to corresponding values in the last decade of 21\textsuperscript{st} century. One simulation with the full seasonal cycle and three perpetual simulations are conducted under RCP8.5 scenario.
2.3 Analysis method

The seasonal cycle in the Arctic is most profound at the surface and in the lower troposphere. Here the surface air temperature is chosen as the index to represent the seasonal march of the Arctic climate. Then the areal average of SAT north of 70\(^\circ\)N (Arctic SAT) is used to characterize the seasonality of the Arctic region as a whole.

As was shown in Fig.1.1, a salient feature of the Arctic seasonal cycle is that the onset of spring is abrupt. In the climatology, mid-to-late winter (Jan~Feb) exhibits a quasi-steady thermodynamic state, then SAT abruptly increases and spring onset occurs. Based on the behavior of the Arctic SAT, a modified version of a two phase linear regression model (LRM) (Cook and Buckley, 2009) is used to define the spring onset date. The model is applied to the Arctic SAT series spanning the period from January 1st to May 31st. The first segment is constrained to be a flat line, which represents the wintertime quasi-steady state, and the second segment starts from the end point of first segment. The model minimizes the root mean square error. The intersection point thus captures the spring onset, the time of the start of spring season. The LRM is used instead of the ROC (radius of curvature) metric in He and Black (2015) because the LRM is more suitable for capturing the climate transition from a quasi-steady winter to the rapid warming of spring, while the ROC method emphasizes more the local temperature variations. An example of the application of the model is shown in Fig.2.1. It can be seen from the example that the two phase linear regression model can capture the abrupt spring transition from a quasi-steady winter to a rapid warming spring very well.

The thermodynamic equation is used to calculate the heat budget on pressure levels, which helps to identify what contributes to heating the Arctic and how these factors change during the springtime transition. The heat budget analysis is carried out on the
850hPa level instead of at the surface for three reasons: First, the temperature at 850hPa is highly correlated with SAT (Fig. 2.2). The correlation between SAT and T850 is above 0.97. A noticeable feature of Fig. 2.2 is that the linear slope between T850 and SAT changes with range of temperature. In the low range of temperatures, the linear slope between SAT and T850 approximates to 1.0, which indicates that, in the winter, SAT and T850 change with the same rate with time and the lapse rate remains the same. In the middle range, SAT changes faster than T850; this range corresponds to spring warming. In the top range, the SAT tends to change slower than T850. A possible reason is that during summer SAT is synchronized with slowly changing SST and its variation tends to be decoupled from the lower tropospheric air temperature. In the springtime transition research, the temperature range is basically in the low end regime. Second, it is easier to calculate the heat budget on a pressure level than at the surface, where the topography makes the calculation complicated. Third, climate models have a positive bias in the strength of the wintertime low-level temperature inversion over the high-latitude Northern Hemisphere (Medeiros et al., 2011). So air temperature on 850hPa is a good indicator of the seasonality in the Arctic and meanwhile suffers much less from the bias of low level temperature inversion in the model.

Lag composites and linear regression (Wilks, 2006) are extensively used in this research. The regression approach provides a useful framework for weighting the data points according to the numerical values of the reference time series (here it is the standardized eddy heat flux at 70N) while, in contrast, the composite method usually ascribes equal weights to all data points entering the composites. Additionally, regression approach utilizes all data points, so the results represent the covariability between the regressed field and the reference time series and have larger degree of freedom. The composite method includes fewer data points, which exceeds a certain threshold or form complete events.
However, the composite method has its advantages, especially in this research. Most of the regressions are onto the eddy heat flux at 70$^\circ$N, which, as will be shown in next chapter, has a skewed distribution and has a mode value close to zero. Linear regression, therefore, will ascribe negative weights to data points corresponding to near-zero eddy heat flux. These data points do not have interesting physical meaning, if the emphasis is on the strong heat flux into the Arctic. On the other hand, the composite based on each eddy heat flux event includes only the data points corresponding to strong eddy heat flux. The composite method can also retain the time mean field, which is desirable sometimes, while linear regression results lose time mean information. However, detailed comparison between the composite and linear regression results show only minor differences.

In order to isolate the contribution of different processes and eddies of different time scales to maintaining the anomalous flow pattern that is responsible for fluxing heat into the Arctic and hence controls the variability of Arctic air temperature on sub-seasonal time scale, the Quasi-geostrophic (QG) geopotential tendency equation (Holton, 2004) is used. The equation is of the form:

\[
\left[ \frac{1}{f_0} \nabla^2 + \frac{\partial}{\partial p} \left( f_0 \frac{\partial}{\sigma \partial p} \right) \right] \chi = -\mathbf{V} \cdot \nabla (\zeta + f) + \frac{\partial}{\partial p} \left[ \frac{f_0 R}{\sigma p} \mathbf{V} \cdot \nabla T \right] \tag{2.1}
\]

$\chi$ is the geopotential tendency which is the wanted solution, $f$ and $\zeta$ is the vertical component of planetary vorticity and relative vorticity respectively. $f_0$ is a constant taking the value of $f$ at 45$^\circ$N, I choose the value at 45$^\circ$N because it is in the middle of Northern hemisphere and the southern boundary of the domain is chosen at equator when solving the equation. $\sigma = -\frac{R T_0}{p} \frac{d \ln \theta_0}{dp}$ is a stability parameter and the subscripts 0 therein represent the basic state which only depends on pressure. Other symbols are used with their conventional meteorological meanings.
The Quasi-geostrophic (QG) geopotential tendency equation is a variant of QGPV conservation, and the RHS of above equation is actually the advection of QGPV. This is a diagnostic linear partial differential equation. The right hand side of the equation can be treated as forcing and can be readily calculated using reanalysis or model output, and the left side is the response of the atmosphere to the forcing in terms of geopotential tendency. The advection of QGPV is decomposed into two parts which represent the barotropic and baroclinic forcings respectively. The barotropic forcing (1st term on RHS) is due to the (horizontal) advection of the absolute vorticity (barotropic part of PV), and the baroclinic forcing (2nd term) is related to advection of the baroclinic PV. It should be noticed that the usual subscript \( g \) is dropped here, and the full variables instead of geostrophic ones are used. This is because it requires differentiation to get the geostrophic winds from the geopotential fields, and these procedures create noisy results when full geopotential fields are used, which is especially true near to topography. For the same reason, the barotropic forcing is calculated from the wind field and the baroclinic forcing is calculated from the temperature field to reduce the order of differentiation. In addition, the flow pattern of interest have planetary scale (more details in next Chapter), so the ageostrophic components of the winds are much less important than the geostrophic ones. A combination of iteration and spectral methods are used to solve this PDE. A detail description is in Appendix A.1. Because its LHS is a linear operator, the equation can be solved with different forcing (RHS) and then each of the solutions are summed to get the total solution. I can, therefore, solve the height tendency response to forcing from different frequency and different processes separately.

An energetics calculation is used as an alternative point of view to understand how the system is sustained. The calculation follows Peixoto and Oort (1974); Baggett and Lee (2014)
Table 2.1: CMIP5 models used to create the SST and sea-ice ensemble projection

<table>
<thead>
<tr>
<th>Model</th>
<th>Horizontal resolution (lon x lat)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Community Earth System Model, version 1 (CESM1)</td>
<td>1.25° x 0.94°</td>
</tr>
<tr>
<td>Meteorological Research Institute Coupled Atmosphere-Ocean General Circulation Model, version 3 (MRI-CGCM3)</td>
<td>1.125° x 1.12°</td>
</tr>
<tr>
<td>Centre National de Recherches Meteorologiques Coupled Global Climate Model, version 5.1 (CNRM-CM5)</td>
<td>1.4° x 1.4°</td>
</tr>
<tr>
<td>Hadley Centre Global Environmental Model, version 2 - Earth System (HadGEM2-ES)</td>
<td>1.0° x 1.0°</td>
</tr>
<tr>
<td>Max Plank Institute Earth System Model, low resolution (MPI-ESM-LR)</td>
<td>1.9° x 1.9°</td>
</tr>
</tbody>
</table>
Table 2.2: Parameterization schemes in CAM5 and related references

<table>
<thead>
<tr>
<th>Scheme</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shallow Convection</td>
<td>Park and Bretherton (2009)</td>
</tr>
<tr>
<td>Deep Convection</td>
<td>Zhang and McFarlane (1995)</td>
</tr>
<tr>
<td>Cloud Microphysics</td>
<td>Morrison and Gettelman (2008)</td>
</tr>
<tr>
<td>Cloud Macrophysics</td>
<td>Park (2010 unpublished)</td>
</tr>
<tr>
<td>Radiation</td>
<td>Iacono et al. (2008)</td>
</tr>
<tr>
<td>Aerosols</td>
<td>Ghan and Easter (2006)</td>
</tr>
</tbody>
</table>
Figure 2.1: Example of fitting the two phase linear regression model to the Arctic SAT series. The data source is the ERA-interim reanalysis for 2008.
Figure 2.2: Scatter plot (blue circles) of daily surface air temperature (SAT) and air temperature at 850hPa (T850), areally averaged over latitudes north of 70N, for (a) ERA-interim reanalysis and (b) control simulation. The red straight line is a linear fit to the data.
Chapter 3

Dynamical heating of the Arctic

3.1 Spring onset

As was introduced in last chapter, the two phase linear regression model is applied to the Arctic SAT time series from day 1 to day 151 of each year. The distributions of the spring onset dates for reanalysis and control simulation are shown in Fig. 3.1. Despite the difference between reanalysis and model simulation, they both have a mode value between 70 ∼ 80 (mid-March), and the average spring onset date is 77 and 73 for reanalysis and control simulation respectively. The variability in the spring onset dates in the control simulation is comparable to that in the reanalysis, even without the variation of SST. That the control simulation has a slightly earlier spring onset compared with reanalysis is consistent with the fact that the spring onset under a warmer climate is delayed, which will be discussed in more detail in next chapter. Here the control simulation is conducted under a background climate colder than the reanalysis data time period (1979 ∼ 2008) during which significant global warming and sea-ice loss in the winter have occurred.

Another control simulation experiment is conducted, in which the SST and sea-ice
climatology is calculated from 1979-2008, the same period as ERA-interim reanalysis. The comparison of the transition dates distribution is shown in Fig.A.2. The spread of the transition dates is similar in these two simulations, which again confirms that the spring transition is controlled mainly by atmospheric internal variability.

The composite of SAT is shown in Fig.3.2, together with the climatology of SAT. The two phase linear regression model can capture the abrupt transition from a quasi-steady winter to a rapidly warming spring. The control simulation can reproduce the salient features of the Arctic springtime transition, which indicates the role of atmospheric internal variability and justifies the experiment design applied here. The composites of SAT show rapid warming during the transition and are in contrast to the more gradual transition in the climatology (Fig.3.2). Fig.3.3 indicates that the Arctic springtime transition is not confined at the surface but extends into middle troposphere. Actually the Arctic springtime transition, as will be shown in following sections, is driven by lower troposphere processes.

The warming is not spatially uniform during the spring onset (Fig.3.4 and 3.5). The initial warming occurs around the Barents Sea and the Kara Sea, and then spreads downstream and into the deep Arctic. In the ERA reanalysis (Fig.3.4), the warming pattern is more concentrated over the Kara Sea and north Russia, while the control simulation (Fig.3.5) shows more downstream spread of the warming. Fig.3.4 and 3.5 are composites based on original SAT, so it might include some of the climatological warming which is assumed to be caused only by solar radiation. The composites of anomalous SAT are shown in Fig.A.4 and A.5. The anomaly is defined as the deviation of the variable from its climatology. The climatology is calculated as follows: first, calculate the daily 30 years average and get a raw climatological seasonal cycle; then perform a Fourier transform on the raw seasonal cycle and retain only its first three harmonics and mean. It is assumed
that the smoothed seasonal cycle is caused only by solar radiation and that the dynamics act on shorter time scales and hence are not included in the first three harmonics. The resemblance between the original composite and anomaly composite indicates that the change of SAT during spring transition is mainly driven by dynamical heating.

3.2 Heat budget

The rapid increase of air temperature in early spring transition is equivalent to strong positive temperature tendency in that period. To disentangle what contributes to the rapid warming, a heat budget analysis is conducted at the 850hPa level. Because, in the definition of springtime transition, the Arctic is treated as a whole system, the thermodynamic equation is integrated (or averaged) over the Arctic region (north of 70\( \text{N} \)),

\[
\left\{ \frac{\partial T}{\partial t} \right\} = \{-V_h \cdot \nabla T\} + \left\{ \left( \frac{1}{c_p \rho} \frac{\partial T}{\partial p} \right) \cdot \omega \right\} + \{Resi\} \tag{3.1}
\]

Curly bracket represents areal integral over the Arctic region. \( \left\{ \frac{\partial T}{\partial t} \right\} = \iint_A \frac{\partial T}{\partial t} dA \), \( A \) is the area of polar cap. The term on the left side is just the temperature tendency which should have large values during the rapid warming period. The first term on the right side is the horizontal advection of temperature (HADV), the second term includes the vertical advection of temperature and temperature change due to adiabatic expansion or compression because of vertical motion (S-OMEGA). The last term on the right side represents the temperature tendency from physics and calculation error (RESI). The HADV term can be related to the heat flux across the Arctic boundary after applying
the divergence theorem and the chain rule of differentiation,
\[
\int\int (-V_h \cdot \nabla T) dA = \int\int \nabla \cdot (-V_h T) dA + \int\int (T \nabla \cdot V_h) dA
\]
\[
= \oint (V_h T \cdot n) dl + \int\int (T \nabla \cdot V_h) dA
\]
\(\oint (V_h T \cdot n) dl\) is the zonal integral of the heat flux cross the 70\(N\) latitude circle, which is defined as the boundary of the Arctic region. The 2\(-D\) version of divergence theorem is used in the above equation so it is assumed that the area bounded by the 70\(N\) latitude circle is a flat surface. This assumption is reasonable if considering only a small polar cap. Detailed calculation of each term gives an approximate equal sign for the above equation.

The heat flux integral can be further written, by decomposition of variables into zonal mean part and eddy part, as:
\[
\oint (V_h T \cdot n) dl = L[vT] = L[v][T] + L[v^\ast T^\ast]
\]
\(\) represents of zonal mean of the quantity, \(L\) is the perimeter of the latitude circle at 70\(N\). Superscript (\(^\ast\)) represents the deviation of the quantity from its zonal mean. \([v^\ast T^\ast]\) is the zonal mean of eddy heat flux, \([v][T]\) is the mean heat flux by the mean meridional wind, \(\{T \nabla \cdot V_h\}\) is due to the mass divergence. The re-arranged thermodynamic equation is:
\[
\left\{ \frac{\partial T}{\partial t} \right\} = L[v^\ast T^\ast] + L[v][T] + \{T \nabla \cdot V_h\} + \{S\omega\} + \{\text{residual}\} \quad (3.2)
\]
\(S = -\frac{T \nabla \theta}{\theta \partial p}\) is the static stability parameter in the isobaric system. The temperature tendency integrated over the polar cap is the sum of the eddy heat flux (EHF) and the zonal heat flux (ZHF) across the latitude boundary, tendency due to mass divergence, vertical motion induced tendency (S-OMEGA) and tendency from physics.
The heat budget composite based on the spring onset date is shown in Fig. 3.6. As is expected, the temperature tendency (dT) is anomalously high during the early springtime transition (Lag0 ∼ Lag15). The variance of dT is mainly controlled by variations in the horizontal advection of temperature (HADV), while the S-OMEGA term is slightly anti-correlated with dT and HADV. This is consistent with the large scale atmospheric response to anomalous heating under the framework of the quasi-geostrophic omega equation. When anomalous heat is advected into the Arctic region, the secondary circulation generated to bring the flow back into balance will adiabatically cool the atmosphere. Additionally, the HADV itself is highly correlated with eddy heat flux term (EHF), while the zonal heat flux (ZHF) and mass divergence (MDV) are anti-correlated with each other (not shown), and the sum of these two does not contribute much to the variance of dT. These qualitative results also apply to the original heat budget time series before compositing and also to the perpetual simulation results. Thus, the abrupt warming in the early springtime transition corresponds to a period of strong temperature tendency, which is caused by the horizontal advection of temperature. Meanwhile, HADV averaged over the Arctic is highly correlated with the eddy heat flux into the Arctic on the subseasonal time scale. It is necessary to investigate the mechanism of the eddy heat flux into the Arctic to understand the Arctic springtime transition dynamically. Adams et al. (2000) shows that the transient eddy heat flux is important in the energy budget in the Arctic when transitioning from fall to winter.

### 3.3 Eddy heat flux

It was shown in Fig. 3.6 that EHF explains much of the variance of the temperature tendency over the Arctic. Further inspection of Fig. 3.6 answers the question why there
is rapid warming during the springtime transition. Given an initial value, the integral of $dT$ (blue curve) gives the trajectory of $T$. Before lag 0, $dT$ bounces around its mean value of zero, so the integral of $dT$ should also bounce around the initial $T$ value, which corresponds to the quasi-steady winter state. During the early springtime transition, a pulse of strong positive $dT$ and EHF cause the rapid warming. After this stage, $dT$ is constantly above zero and has much less variance. The persistently positive $dT$ is mainly due to the fact that solar heating starts to weigh in, and the reduced variance is due to the less varying EHF. The variance of the zonal average of the eddy heat flux at $70^\circ N$ is calculated as following: first calculate the variance of EHF within an 11-day moving window; the variance value is recorded at the center point of the window; then the 30-year climatology of the variance obtained in last step is calculated. The climatology of EHF variance is shown in Fig.3.7 as the blue curve. The three dots are calculated using a similar method for each perpetual simulation. The variance of the perpetual simulations follows the trend in the control simulation: highest variance appears in the winter, and it decreases with time and reaches its minimum in the summer. However, the decreasing trend is not gradual, the variance decreases in a stepwise manner. This sudden decrease in the variance of the eddy heat flux is part of the reason that the springtime transition is abrupt.

The eddy heat flux at $70^\circ N$ is regressed onto its zonal average (Fig.3.8) to show its zonal distribution. The regression is maximized at lag 0, as is expected, and it does not show any propagation in the longitude-time space. This indicates that the EHF is a localized system. Near the Greenwich Meridian at lag 0, there is a local maximum of the regression coefficient, and there is some hint that EHF in this region has some periodicity which appears as a second maximum at around lag 25. The region near the Greenwich Meridian is a critical region for EHF, and EHF averaged over this region explains more than half
of the variance of the zonal mean EHF. A similar regression calculated using reanalysis
data shows similar results in Fig.A.3a. One caveat for the regression shown here is that it
covers data from all seasons of the 30 years, and it was shown (Fig.3.7) that the variance
of EHF is strongest in the winter and changes dramatically with different seasons. It
is, therefore, possible that the regression pattern in Fig.3.8 is mostly from the winter
anomalies. Similar regressions are calculated in each perpetual season, and the results at
lag 0 are shown in Fig.3.9. The distribution of EHF in January is similar to the regression
pattern over the whole year, which shows a concentrated EHF over the Greenland Sea. In
March, EHF exhibits a similar pattern but with reduced magnitude, which is consistent
with the decreasing variance of EHF with time (Fig.3.7). In April, the distribution of
EHF has totally changed. There is weak EHF over the Greenland Sea, while the center of
strong EHF is located in Northern Russia. The mean air temperature at 850hPa (contour
in Fig.3.9) indicates that the location of center of EHF might be largely controlled by
the lower boundary thermodynamic forcing which is the consequence of land-sea thermal
contrast. In the winter and early spring, the Greenland Sea is relatively warmer than
north Eurasia at the same latitude. However, in the middle to late spring, the continent
has warmed and is warmer than the ocean. Recall that EHF is $v^*T^*$, if $v^*$ and $T^*$ are
separated into their time-mean and deviation from time-mean components, then:

$$vT = \bar{vT} + v'T + \bar{v}'T'$$

Note that the superscripts *, which represents the deviation from the zonal mean is
dropped from here to the end of this section for the sake of clarity. The overbar represents
the time mean of the quantity and ’ represents the deviation from its time mean. The
variance of eddy heat flux scales with:

\[
\frac{1}{N-1} \sum (vT - \bar{vT})^2 = \frac{1}{N-1} \sum (v'T + \bar{v}T')^2 \\
= \frac{1}{N-1} \sum [(v'T)^2 + (\bar{v}T')^2 + 2v'T\bar{v}T']
\] (3.3)

So the variance of (eddy) heat flux is composed of three terms: transient meridional wind fluxing climatology temperature, climatology meridional wind fluxing transient temperature, and the cross-interaction between them. Fig.3.10 shows the distribution of the decomposed variance of EHF for three perpetual simulations. For both January and March (Fig.3.10a,b), the main factor is \(v'T\) (red curve), which is consistent with the fact that the superposition of climatological temperature (\(\bar{T}\)) with transient meridional wind component (\(v'\)) produces the variation of EHF. In April (Fig.3.10c), however, the other two components also contribute significantly to the total variance of EHF, in which case the dynamics is less clear than January and March. Notably, the magnitude of the variance decreases with time from January to April (different scales used in the plot).

The decrease of EHF variance from January to March is caused by the reduction in the \(v'\) (actually its variance) and \(\bar{T}\) (its square) (Fig.3.11). \(v'\) contributes about 1/4 of reduction while \(\bar{T}\) contributes more than half of the reduction. The total reduction in the EHF variance is about 2/3. The reduction in \(\bar{T}\) is due to the change in land-sea contrast (contour in Fig.3.9). The reduction in \(v'\) is from the weaker Rossby wave-1 in March than in January (Fig.3.12, and details in next section).

Based on the fact that EHF over the critical region explains most of the variance of its zonal average, and it potentially has rich dynamics (related to transient \(v''\)), from here on, I focus on the EHF average over the critical region, and EHF automatically refers to its average over 20W ∼ 70E. Fig.3.14 is the PDF of EHF. It has a skewness of 1.0 and

27
a long positive tail. This indicates the importance of extreme events for the EHF. The contribution of extreme events to the total EHF is listed in table 3.1: the top 5% of strong events contribute about 1/4 of the total EHF, and the top 10% contribute almost half of the EHF. This result is not as dramatic as that in Messori and Czaja (2013), because their calculation includes all data points in the Northern Hemisphere and, hence, more non-extreme values make the extremes stand out. The skewness of the EHF distribution and the percentage of contribution from extreme events are significant.

3.3.1 EHF in other latitudes

One question arises: whether the concentration of EHF is a unique feature in the Arctic. The regression of EHF onto the zonal average EHF in the same latitude is shown in Fig. 3.13. A local maximum indicates a concentrated EHF in that region. The concentration of EHF is truly a unique feature in the Greenland Sea between 60°N ∼ 70°N. This is because the thermodynamic lower boundary condition there is unique.

3.4 Anomalous flow pattern

In this section, I present the anomalous flow patterns associated with the anomalous EHF. Due to the sporadic nature of EHF, the linear regression method might not be suitable for extracting the pattern associated with EHF. Fig 3.14 shows that EHF has a positive mean value and a close-to-zero mode, and these close-to-zero data points would contribute to the regression as negative anomalies. Instead, a composite method is used to isolate the common patterns associated with strong EHF events. The lagged composite of geopotential height at 500hPa (Z500) is shown in Fig 3.15. The most prominent pattern at lag 0 is a dipole-like pattern spanning the North Atlantic and the Greenland Sea,
which is consistent with anomalous southerly flow \( (v''') \) there. Again, similar calculation with reanalysis data shows similar results, as is shown in Fig.A.3b.

In the sequence of composite maps with lags, there is some indication that the pattern propagates westward and the propagation slows when the pattern is strongest (lag - 3 ∼ lag 3). The vertical profile at 70N shows that the anomalous pattern associated with EHF is quasi-barotropic (shading in Fig.3.16) while it has a baroclinic structure when superposed on the time mean field (contour in Fig.3.16). The barotropic structure is consistent with Rossby wave dynamics (details in the rest of this section), and the baroclinic total structure indicates the role of baroclinic conversion in the energetics (next section).

A Hovmoller diagram (Fig.3.17) confirms this westward propagation. The period estimated from the diagram is about 25 days, which is consistent with the phase speed of Hough mode \((1, 4)\) in Lindzen et al. (1984). The geopotential height fields are decomposed into different wave numbers and similar composites are calculated for each wave number component; the results are shown in Fig.3.18. The wave 1 component shows clear westward phase propagation and its phase speed is consistent with estimates from Fig.3.15 and 3.17. For wave 2, there is no apparent precursor before lag 0. The emergence and amplification of wave 2 at lag 0 is not due to propagation from upstream. One possible mechanism for wave 2 growth might be the non-linear interaction of wave 1 components of geopotential height (inducing meridional wind through geostrophy) and the temperature wave through baroclinic process. The details of this interaction are beyond the scope of this research. Wave 3 has only a weak contribution to the dipole pattern, and it shows clear eastward propagation even in the composite. The sum of wave 1 and wave 2 can reproduce most of the shape and magnitude of Fig.3.17.

Early studies have shown evidence of regularly propagating large-scale waves (Mad-
den, 1978; Lindzen et al., 1984). Here I repeat some of the analyses and apply the methods to model output. Cross-spectrum analysis results of Z500 at 70N for wave number 1, 2 and 3 are shown in Fig.3.19. For a traveling wave of certain wave number, the fluctuation in the cosine coefficient would be one quarter of a cycle out of phase with that of the sine coefficient (Deland, 1964). For wave 1, significant coherence squared with phase of $-90$ covers a wide frequency band, period of about $12 \sim 50$ days, and has a maximum value at a period of 16 days, which confirms that there is a westward propagating wave 1 component. For wave 2, significant coherence squared with phase $90$ indicates eastward propagation, and the period is $2 \sim 5$ days. It should be noted that in the cross-spectrum calculation, the phase is averaged over a frequency band and the phase has meaning only if the coherence squared is significant.

Another method to investigate the wave propagation is using the space-time spectral analysis technique (Hayashi, 1979; Straus and Shukla, 1981). Any variable depending on time and longitude can be decomposed into westward and eastward propagating part for a certain wave number. However, after this decomposition, the sine and cosine coefficient of this wave number will be in perfect quadrature with each other, so the cross-spectral analysis will give a perfect 90 or $-90$ phase and the coherence squared will be 1.0 exactly (not shown).

Recall that in Fig.3.17 and 3.18, the wave 1 amplitude also fluctuates with time and amplifies at a certain phase. Indeed, wave 1 is amplified at phase 90 and $-90$ (Fig.3.20(b) and (c)). A bi-modal distribution is apparent in the phase distribution, specially when considering only strong wave 1 (amplitude exceeding one standard deviation from the mean, Fig.3.20(c)). The distribution of wave 1 amplitude has a skewness of 0.63 and deviates far from a normal distribution, which partially explains the skewed distribution of EHF (Fig.3.14). For westward and eastward propagating waves only, the amplitudes
have even higher skewness. However, the bi-modal distribution is not as dramatic as in Fig.3.20, when the westward and eastward propagating wave are separated. A possible reason is that when wave 1 is amplified at certain phase, say −90, the phase speed is slowed and tends to be stationary, and this amplitude is ascribed to neither westward nor eastward propagating wave. The scatter plot of wave-1 phase and amplitude also shows that wave-1 prefers certain phases (90 and −90), and stronger wave-1 clusters around these two phases.

We do not expect a one-to-one correspondence between the wave 1 amplitude and EHF, because both have a highly skewed distribution and close-to-zero mode. However, it is reasonable to expect that the extreme EHF events have a preference for wave 1 with larger amplitude. Comparing Fig.3.24 with Fig.3.20, it is clear that, when considering only these data points corresponding to strong EHF events (above one standard deviation), the mode of wave 1 amplitude shifts to the right, which means stronger waves are more frequent. The bi-modal distribution is replaced by a single modal distribution with a skewness of −1.5. The phase of 90 corresponds to the dipole pattern in Fig.3.15.

The composite of geopotential height at 500hPa along 70N, based on the spring onset date of each year, is shown in Fig.3.25. Though it is noisier than the composite based on EHF events, this diagram also shows westward propagation of wave-1. At lag 0 (spring onset date), the SAT starts to increase rapidly, so it is in the lowest point of EHF event and corresponds to the reversed phase of dipole pattern. As the pattern propagates westward, the dipole pattern spans the North Atlantic and the Greenland Sea, and induces strong EHF in that region. This occurs about 8 days after the spring onset. Recall that the two phase linear regression model is not designed to capture the exact turning point of air temperature time series. However, the composite based on the spring onset dates still shows westward propagation of the wave-1 pattern.
In the NO-Greenland experimental simulation, the dipole pattern is also related to the EHF, as is shown in Fig.3.26. The role of Greenland topography in regulating the EHF and related flow pattern is not critical, at least in this model simulation.

### 3.5 Energetics

When the dipole pattern spans the North Atlantic, the anomalous \( v^* \) induces northward heat flux \( (v^*T^*) \) into the Arctic. In classical baroclinic conversion theory, this behavior will convert zonal available potential energy (ZAPE), which is stored as the meridional gradient of zonal temperature, into eddy available potential energy (EAPE) and eddy kinetic energy (EKE). Meanwhile, this energy conversion is necessary in order to maintain the anomalous dipole against dissipation by friction and radiative damping. The energetics are calculated and integrated over the Northern hemisphere horizontally and from 1000hPa to 100hPa vertically, and then is composited based on the EHF events. Compositing, instead of linear regression, is used for the same reason discussed before. The meaning of each term is: ZAPE is zonal available potential energy, EAPE is eddy available potential energy, EKE is eddy kinetic energy, BC is baroclinic conversion, BT is barotropic conversion; TOT ENERGY is the sum of zonal and eddy available potential energy and kinetic energy, GZAPE is the generation rate of ZAPE, GEAPE is the generation of EAPE, DISS is the dissipation of energy by boundary and radiative processes. The generation rates are multiplied by \( 24 \times 3600 \) in order to convert the units to energy generation per day. The results are shown in Fig.3.27. The energetics composite has an apparent periodicity of about 25 days. The conversion between ZAPE and EAPE is consistent with the classical baroclinic growth theory. When the anomalous \( v^* \) fluxes heat down the zonal temperature gradient, the ZAPE decrease, and the EAPE and
EKE are increased. These eddy energies are important for maintaining the dipole pattern and anomalous $v^*$. Such a positive feedback, in an energetics view, exists, and it is the mechanism under which wave 1 is amplified at phase $+90$.

To form a quasi-periodic oscillation, there must be some negative feedback. If the Northern Hemisphere is treated as a whole system, and it is assumed the energy flux cross the lateral boundary is negligible after compositing, the only sources of total energy are GZAPE and GEAPE, and the sink of energy (DISS) can be estimated by subtracting GZAPE and GEAPE from the energy tendency. Neither GZAPE nor GEAPE contribute much to the variation of total energy (Fig.3.27b), and the total energy tendency (blue dash line) closely follows the dissipation. However, the dissipation might have an important role in forcing the oscillation. DISS is roughly anti-correlated with EKE and EAPE, and is in quadrature with the total energy cycle. Thompson and Barnes (2014) showed that the negative feedback between eddy heat flux and meridional temperature gradient can, with a stochastic forcing of eddy heat flux, generate a periodic oscillation of eddy heat flux. It is not necessary to have a nonlinear interaction (Ambaum and Novak, 2014) to generate the oscillation, if a stochastic forcing exists. A similar negative feedback between total energy (TE) and EKE can be used to explain the oscillation feature of the energetics. To simplify the model, I applied two assumptions: (1) Dissipation is anti-correlated with EKE; (2) generation of EKE is proportional to the total energy (TE). The first assumption is supported by Fig.3.27 and the idea that most of the dissipation is due to boundary process and scales with the strength of motion. The second assumption is based on the fact that baroclinic conversion rate is roughly constant and
the total energy is dominated by ZAPE. The simplified model is:

\[
\frac{\partial TE}{\partial t} = -\alpha EKE + \varepsilon(t) \\
\frac{\partial EKE}{\partial t} = \beta TE
\]

(3.4)

\(\varepsilon(t)\) represents stochastic forcing of system which is the sum of GZAPE and GEAPE and prevents the system from reaching an equilibrium. By eliminating EKE in the above equation set, it becomes an oscillator equation for TE, with additional stochastic forcing \(\varepsilon(t)\). What is happening physically is just as: Starting from a state with surplus of ZAPE, BC converts ZAPE into EAPE and EKE, strong EKE causes strong dissipation through boundary processes, so the total energy decreases, and the system switches to a state of ZAPE deficit. Then the dissipation anomaly changes sign, so ZAPE starts to accumulate and prepare for the next cycle.

The previous energetics is integrated over the domain (Northern hemisphere), so a question arises whether those composites represent the action of the dipole and EHF events. Because there are only 32 EHF events included in the composites, it is possible the variation of energetics in other places has not been smoothed out. Fig.3.28 shows the composite without integrating in the horizontal dimensions. The EAPE, EKE and BC are indeed associated with the dipole pattern and EHF events. The EAPE maximum is located in the Norwegian Sea and Barents Sea, which is the consequence of heat being fluxed into the Arctic; the EKE maximizes in the Greenland Sea, which captures the anomalous \(v^*\) associated with the dipole; the BC also maximizes in the same region while the pattern is complicated by local topography.
3.6 Geopotential height tendency

From a dynamic point of view, the dipole pattern must be maintained by a height tendency against dissipation. The quasi-geostrophic geopotential height tendency equation is used to diagnose the tendency from different processes and motions of different time scales. To construct a budget of the geopotential height tendency from different terms, the tendencies are areally averaged over two boxes. Both boxes have a meridional range from 60°N ∼ 80°N. The west box extends from 90°W ∼ 0°W while east box covers 0°E ∼ 90°E. Fig.3.29 shows the geopotential height anomaly composite averaged over these two boxes at the 850hPa and 300hPa levels respectively. The west average and east average are out of phase with each other, which is consistent with the traveling wave-1 hypothesis. The 300hPa curves track the 850hPa ones, which indicates that the structure associated with EHF events is quasi-barotropic, and, again, it is consistent with linear wave theory and its propagation. Before showing how different terms contribute to maintain the dipole height anomaly, it is necessary to ensure that the QG model can reproduce the original height tendency reasonably (Fig.3.30). The inverted height tendency includes the forcing of large scale condensation and long wave radiation in addition to the RHS of equation A.1. The QG model indeed can reproduce the timing and magnitude of the height tendency reasonably well even in the lower troposphere where boundary layer processes are important. The inverted solution is smoother and slightly leads the real solution, because in the inversion I keep only the first 10 wave numbers and no friction is included.

The contribution from long-wave radiation forcing is small, so, in the following results, this term is omitted. Fig.3.31 shows the inverted tendencies from different forcing terms averaged over the west and east boxes. The largest contribution is from the linear term (Linear, red curve). The linear term includes both the phase propagation and group
propagation of linear wave. However, the role of the linear term is different in the west and east centers of the dipole. In the west center, the linear term damps the height anomaly, such that the tendency is most positive when the height anomaly is negative (lag 0) and *vice versa*. In the east center, the linear term reinforces the height anomaly. Its peak is located between the ridge and trough of the geopotential anomaly (about lag 5). This is verified, in terms of wave activity flux, in Fig. 3.32. At lag 0, the west center is a source of wave activity, and wave activity propagates eastward and equator ward. The divergence of the wave activity flux weakens the upstream low and its convergence strengthens the downstream high.

The contribution from high frequency eddies (Hi-Hi) tends to correlate with the height anomaly itself. This can be explained by the interaction between the high-frequency eddies and the low-frequency mean flow. For example, at lag 0, the west center is low, and the westerly flow in its southern flank is strengthened. Consequently, positively anomalous synoptic eddies originate there, and they flux heat and vorticity northward which converge into the west center. This convergence of heat and vorticity flux reinforce the negative height anomaly and produce a positive feedback (Lau and Nath, 1991). However, the roles of the heat flux and the vorticity flux differ. The anomalous vorticity flux reinforces the negative height anomaly barotropically, while heat flux reinforces the height anomaly at lower levels and damps the height anomaly at higher levels. The results shown here indicate that vorticity flux forcing dominates the heat flux forcing, because this term has the same sign both at higher and lower levels. The condensation forcing is most important in the lower troposphere, because moisture is concentrated at low levels. When there is anomalous EHF, warm, moist air is fluxed into the Arctic and making it easier to condense and produce precipitation near the Greenland east coast. This released heat intensifies the negative height anomaly in the lower troposphere (Fig. 3.31c). This also acts as a
positive feedback and reinforces the west low at lower levels.

The contribution from low frequency eddies has a comparable magnitude with the linear terms in some locations. This is consistent with the results in Blackmon et al. (1977) who found that time-scales between 2.5 to 6 days and motions with periods longer than 10 days contribute comparable amounts of transient-eddy heat transport, with the high frequency eddies dominating in the storm-track region and the low frequency motions at higher latitudes. Another source of low frequency interaction might be from anomalous wave breaking events in these regions.
Table 3.1: Contribution of extreme events to total EHF

<table>
<thead>
<tr>
<th>Percentile</th>
<th>Overall%</th>
<th>Positive%</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>26.13</td>
<td>22.61</td>
</tr>
<tr>
<td>10</td>
<td>45.07</td>
<td>39.01</td>
</tr>
</tbody>
</table>

Figure 3.1: Histogram of the spring onset date calculated using (a) ERA-interim reanalysis and (b) CTRL simulation.
Figure 3.2: Composite of SAT based on the spring onset date calculated using ERA-interim reanalysis and CTRL simulation. The climatology of SAT is shifted so that day 73 in the climatology aligns with day 0 in this plot. The mean value for each curve over the 101 composite period is subtracted.
Figure 3.3: Composite of zonal mean Air Temperature at different levels based on the spring onset date using (a) ERA-interim reanalysis and (b) CTRL simulation. Units are $K$. 
Figure 3.4: Composite maps of the change in SAT between each lag day and lag 0. Units are $K$. The data source is the ERA-interim reanalysis.
Figure 3.5: Composite maps of change in SAT between each lag day and lag 0. Units are $K$. The data source is the control simulation.
Figure 3.6: Composite of heat budget terms in equation 3.2 based on the spring onset date. All curves are smoothed with a 5-day running mean. Units are $K/s$. The data source is the ERA-interim reanalysis.
Figure 3.7: Variance of the zonal average of the eddy heat flux at 70N; the solid line is the climatology of variance within an 11-day moving window (control simulation); the blue dot, orange dot and red dot are the mean 11-day variance for perpetual January, March and Apr simulation respectively. Units are $K^2 \cdot m^2/s^2$.
Figure 3.8: Regression of the eddy heat flux at 70N onto its zonal average: (a) with different lags and (b) at lag 0 only. The regression is calculated with anomaly fields only. Units are $K \cdot m/s$. The data source is the control simulation.
Figure 3.9: Shading: Regression of the eddy heat flux onto the its zonal average at 70N (units: $K \cdot m/s$); contour: mean air temperature at 850hPa (units: K), the contour interval for the temperature is 4K. The data source is perpetual simulation in (a) January, (b) March and (c) April.
Figure 3.10: Decomposition of the variance of EHF into three terms as shown in equation 3.5, for the perpetual simulation in (a) January, (b) March and (c) April.
Figure 3.11: Comparison of the variance of $v^*$ and $T^2$ between January and March. The data sources are perpetual simulation in January and March.
Figure 3.12: Histogram of the amplitude of wave-1 of geopotential height at 500hPa along 70N, for current January, current March and future January simulations. The data sources are perpetual simulation in January, March and RCP8.5 January.
Figure 3.13: Regression of EHF onto its zonal average at the same latitude. Units are $K \cdot m/s$. The data source is the perpetual January simulation.
Figure 3.14: PDF of EHF averaged over $20W \sim 70E$ (bar plot) and normal distribution with the same mean and standard deviation (curve) as the EHF. The data source is the perpetual January simulation.
Figure 3.15: Composite of geopotential height at 500hPa (Z500) based on 32 strong EHF events. Units are m. The data source is the perpetual January simulation.
Figure 3.16: Composite of geopotential height along 70N based on EHF events at (a) Lag -12, (b) Lag 0 and (c) Lag 12. Shading is based on the anomaly field (time mean removed); contours show the full field (including the time mean). Units are m. The data source is the perpetual January simulation.
Figure 3.17: Hovmöller diagram: composite of geopotential height at 500hPa along 70N, based on EHF events. Units are m. The data source is the perpetual January simulation.
Figure 3.18: *Hovmöller* diagram: composite of geopotential height at 500hPa along 70N, based on EHF events for (a) wave 1, (b) wave 2, (c) wave 3 and (d) sum of wave 1 and wave 2. Units are m. The data source is the perpetual January simulation.
Figure 3.19: Coherence squared and phase of cross-spectra between the sine and cosine coefficients of wave-1, wave-2 and wave-3 respectively. The red horizontal line indicates the 95% significance level. The calculation uses geopotential height at 500hPa along 70N. The data source is the perpetual January simulation.
Figure 3.20: Histogram of wave 1 of geopotential height at 500hPa along 70N: (a) Amplitude, (b) Phase and (c) Phase when amplitude exceeds one standard deviation. The data source is the perpetual January simulation.
Figure 3.21: Histogram of westward propagating wave 1 of geopotential height at 500hPa along 70N: (a) Amplitude, (b) Phase and (c) Phase when amplitude exceeds one standard deviation. The data source is the perpetual January simulation.
Figure 3.22: Histogram of eastward propagating wave 1 of geopotential height at 500hPa along 70N: (a) Amplitude, (b) Phase and (c) Phase when amplitude exceeds one standard deviation. The data source is the perpetual January simulation.
Figure 3.23: Scatter plot of wave-1 amplitude and phase. Only data points for amplitudes greater than one standard deviation above the mean are retained. The wave-1 component is calculated using geopotential height at 500hPa along 70N. The data source is the perpetual January simulation.
Figure 3.24: Histogram of wave 1 of geopotential height at 500hPa along 70N, when EHF is at least one standard deviation above its mean: (a) amplitude and (b) phase. The data source is the perpetual January simulation.
Figure 3.25: *Hovmöller* diagram: composite of geopotential height at 500hPa, based on spring onset date of each year. Units are m. The data source is the control simulation.
Figure 3.26: Regression of geopotential height at 500hPa onto zonal mean eddy heat flux at 70N. Units are m. The data source is the NO Greenland simulation.
Figure 3.27: Domain averaged energetics composite based on EHF events, see text for the meaning of each term. The data source is the perpetual January simulation. The domain average is calculated over the Northern hemisphere and from 1000hPa to 100hPa. The time mean of energetics terms are subtracted before compositing.
Figure 3.28: Vertically averaged energetics composite averaged over certain days, (a) EAPE average from day $0 \sim 5$, (b) EKE average from day $-1 \sim 4$ and (c) BC average from day $-6 \sim 0$. The vertical average is calculated from 1000hPa to 100hPa. The time mean of energetics terms are subtracted before compositing.
Figure 3.29: Geopotential height composite averaged over the west box and east box at (a) 850hPa and (b) 300 hPa. The west box covers 60N -80N and 90W-0W; the east box covers 60N-80N and 0E-90E. Units are m. The data source is the perpetual January simulation.
Figure 3.30: Comparison of observed (model output) geopotential height tendency and inverted tendency using the QG geopotential height tendency equation. Units are m/s. The data source is the perpetual January simulation.
Figure 3.31: Inverted geopotential height tendency averaged over the west and east boxes at 850hPa and 300hPa. Units are m/s. The data source is the perpetual January simulation. See text for details of the meaning of each term.
Figure 3.32: Composite of wave activity flux (vector) and its divergence (shading) at lag0 based on EHF events. Units for the vector are $m^2/s^2$, units for shading are $m/s^2$. The data source is the perpetual January simulation.
Chapter 4

Response to future climate change

This chapter includes the study of how the Arctic springtime transition and associated dynamic heating will change under future climate change. As was introduced in Chapter 2, the time period is the last decade of the 21st century. The purpose of this chapter is not only to project a possible change in the Arctic springtime transition and dynamic heating within, it also serves as a testing ground for the mechanisms proposed in Chapter 3. Under the climate change, each of the factors that is found to affect the Arctic springtime transition will have potential change. Accordingly, a test of whether the change in the Arctic springtime transition and associated dynamic heating are consistent is useful to give us more confidence in the previously proposed mechanisms.

4.1 Change in the Spring onset and seasonal cycle

As is shown in Fig.4.1, the climatological seasonal cycle is delayed in the future projection. The arrival of the hottest day in the summer is delayed by about 17 days while the coldest day in the winter is delayed by about 50 days. This is mainly due to sea ice loss: as sea ice
melts during the twenty-first century, the previously unexposed open ocean increases the effective heat capacity of the surface layer, slowing and damping the temperature response (Dwyer et al., 2012). The amplitude of the seasonal cycle is reduced significantly in the RCP8.5 projection. In addition, the variability of SAT, especially in winter, in the future projection is highly reduced (Fig.4.2).

The spring onset is also delayed using the definition in this research (Fig.4.3), and the average onset date is day 94 of year in the future projection compared with day 73 in the control simulation. Meanwhile, the spring onset in the future projection also has a small range. The dates of spring onset in the future projection range from day 80 to day 105, and, in contrast, the dates in the control simulation span the time from day 40 to day 100. The composite of SAT shows that the springtime transition in the future projection is not abrupt compared with the current climate (Fig.4.4). Two possible reasons might account for the lack of abruptness in the springtime transition in the future climate: first, the EHF variation is weaker in the future projection than in the current climate. It was shown that it is the strength of EHF variation that directly causes the rapid warming during the springtime transition. Secondly, the delayed springtime transition has weaker EHF at that time, because the EHF variance decreases with time rapidly as the season advances. The next section will investigate these possible reasons and associated physical mechanisms.

4.2 Change in Eddy Heat Flux

Fig.4.5 shows a comparison of the EHF variance in the control and RCP8.5 simulations. The sudden decrease of the EHF variance in the control simulation no longer exists in future projection. There is some stepwise decrease of the variance in the future projection,
but its magnitude is much weaker than in the control simulation. In the control simulation, the average spring onset date is day 73 (mid-March) which generally precedes the landslide decrease in the EHF variance. The mean spring onset date for future simulation is day 94 (early April) which is also right before the stepwise decrease of the EHF variance, but with a much weaker magnitude. These differences partially explain why the spring transition in the future projection is not as abrupt as in the control simulation.

Another feature in Fig.4.5 is that the variance of wintertime EHF in the future projection is about 1/3 that in the control simulation. Again, the EHF variance can be decomposed as in equation 3.5. Fig.4.6 shows that the difference of EHF variance comes mainly from the difference in \( v^* T^* \). The two centers with most change are in the Greenland Sea and North America, and these regions are exactly where the EHF manifests itself most strongly (Fig.3.9a). The change of \( v^* T^* \) is possibly due to the change of dynamics \( (v^*) \) or thermodynamics \( (T^*) \) or a combination of both. Fig.4.7a shows that the transient meridional wind does not change significantly, especially in the Greenland Sea region, while mean \( T^* \) explains most of the change in the variance of EHF (Fig.4.7b). The decrease of \( T^* \) over the critical region is mainly due to the change in land-sea contrast with global warming. Fig.4.8 shows that warming occurs everywhere, but the magnitude differs a lot spatially. The enhanced warming over North America and reduced warming over the North Atlantic are most striking. This will reduce the land-sea contrast. The land-sea warming contrast is a phenomenon of both equilibrium and transient simulations of climate change. Areas of the land surface undergo temperature changes whose amplitude is more than that of the surrounding oceans (Joshi et al., 2008; Sutton et al., 2007). Joshi et al. (2008) proposed that the main cause of this contrast is from the difference in lapse rate over land and ocean. However, this mechanism does not explain the contrast between Eurasia and North America. The full mechanism of land-sea contrast
change under global warming is beyond the scope of this thesis research, but its role in dynamical heating of the Arctic has been verified here. Fig.4.8 (bottom) shows the mean T850 along 70N with its zonal mean subtracted. North America has an enhanced T anomaly (compared to its zonal mean) while the Greenland Sea region has a reduced T anomaly. The enhanced T anomaly in North America also corresponds to the reduction in EHF variance, because there is northerly flow there when there is southerly flow in the Greenland Sea in the wave-1 configuration. So, under the change of the climatology of $T^*$, there is basically a reduction in the variance of EHF everywhere. (Fig.4.9b),

The variance of EHF is affected only by $v'\overline{T}$ and $\pi T'$ and their interaction as was shown in equation 3.5, in which the subscripts $*$ were dropped. The time mean of EHF is not affected by these two terms, instead it is composed of:

$$\overline{vT} = \overline{vT} + \overline{v'T'} \quad (4.1)$$

the change in the mean EHF is shown in Fig.4.9a. There is also reduction in the mean EHF in the North Atlantic and the Greenland Sea while there is an increase of the mean EHF in the west coast of North America.

### 4.3 Energy transport

The previous chapters show that the variation of the Arctic lower tropospheric temperature is controlled by the eddy heat flux into the Arctic in the critical region, and this eddy heat flux becomes weaker and less variant under global warming. However, the Arctic is a region with an energy deficit, and it irradiates more energy out to space under global warming which make the deficit worse, so it requires more energy transported into the
Arctic either through the lateral or lower boundary. Total heat flux, instead of eddy heat flux, might give more hints about how these energy transports will change under climate change. Fig. 4.10 shows the mean total heat flux at 850hPa in the control and RCP8.5 simulations and their difference. The North Atlantic region sees a reduction in total heat flux, which is consistent with the reduction in eddy heat flux. A region with increased heat flux is near the west coast of North America, which is also the region with the most eddy heat flux increase. The zonal average of Fig. 4.10c is shown in Fig. 4.11; most of the latitudes see a reduction of total heat flux with the exceptions in the tropics and high latitudes. The change in the mean meridional wind and stationary wave (Fig. 4.12) explain the change in the total heat flux. It is the strengthened low in the middle Pacific that is responsible for the increase of heat flux in that region.

A more precise way to depict the energy transport is using the moist static energy (MSE) which is an approximately conserved quantity. Fig. 4.13 shows that all the three components in the MSE flux have a similar pattern, while the sensible heat flux dominates the other two components. The pattern of changes in the sensible heat flux resembles that of total heat flux at 850hPa (Fig. 4.10). In Fig. 4.13, the change of the potential energy flux has a similar pattern with that of the latent heat flux, which indicates that the change of flow pattern is a barotropic, instead of an overturning circulation. The zonal average of Fig. 4.13d is shown in Fig. 4.14. Only the mid-latitudes experience a slight increase of the MSE flux, while other places see uniform decreases in the MSE flux.
Figure 4.1: Climatology of SAT area averaged north of 70N (a) full cycle and (b) anomaly with annual mean subtracted. The data sources are the RCP8.5 simulation and the control simulation.
Figure 4.2: SAT averaged north of 70N for each year, (a) RCP8.5 simulation and (b) Control simulation.
Figure 4.3: Histogram of Springtime onset date, (a) RCP8.5 simulation and (b) control simulation
Figure 4.4: Composite of SAT averaged north of 70N. The data sources are the RCP8.5 simulation and the control simulation.
Figure 4.5: Climatology of EHF 11-day variance for the future projection and the current climate simulation. The blue curve is the same in Fig.3.7. The data sources are the RCP8.5 and control simulations.
Figure 4.6: Decomposition of variance of eddy heat flux, for the future projection and the current climate simulation; solid lines are future, dash line are control. Red: $u'^*T^*$, Blue: $v^*T'^*$, Black: cross term. See details in equation 3.5. Data sources are RCP8.5 and control perpetual January simulations.
Figure 4.7: Comparison of the variance of $v^*$ and $T^*_{\text{bar}}^2$. The data sources are the RCP8.5 and control perpetual January simulations.
Figure 4.8: Top: change in air temperature at 850hPa (future minus current); bottom: mean $T^*$ in ctrl and future simulations. The data sources are the RCP8.5 and control perpetual January simulations.
Figure 4.9: Change in the EHF (future minus current), (a) time mean and (b) variance. The data sources are the RCP8.5 and control perpetual January simulations.
Figure 4.10: Mean total heat flux from (a) ctrl simulation, (b) rcp8.5 simulation and (c) the difference rcp8.5 - ctrl. Units are $K \cdot m/s$. The data sources are the RCP8.5 and control perpetual January simulations.
Figure 4.11: Zonal average of change in total heat flux (future minus current). Units are $K \cdot m/s$. The data sources are the RCP8.5 and control perpetual January simulations.
Figure 4.12: Mean meridional wind (shading) and geopotential height (contour, zonal mean subtracted), in (a) ctrl simulation, (b) rcp8.5 simulation and (c) the difference: rcp8.5-ctrl.
Figure 4.13: Change of vertically integrated moist static energy (MSE) flux and each component, (a) sensible heat flux, (b) potential energy flux, (c) latent heat flux and (d) MSE flux. The vertical integral is conducted from 1000hPa to 100hPa. Units are $\text{J} \cdot \text{m/s}$. The data sources are the RCP8.5 and control perpetual January simulations.
Figure 4.14: Change (future minus current) of zonal average of MSE flux, vertically averaged from 1000hPa to 100hPa. Units are $J \cdot m/s$. The data sources are the RCP8.5 and control perpetual January simulations.
Chapter 5

Discussion

5.1 Summary

In this thesis research, I invent an objective two-phase linear regression to define the Arctic springtime transition. Different from traditional metrics used in seasonality studies, this new method captures the unique characteristics of the springtime transition in the Arctic, which is featured as a transition from a quasi-steady winter to a rapidly warming spring.

Through heat budget analysis, I find that it is the eddy heat flux (EHF) entering the Arctic that controls the Arctic SAT variability on sub-seasonal scales. What accompanies the springtime transition is the dramatic reduction of the variability in EHF and the Arctic SAT. The variability of the Arctic SAT has been studied in a few investigations. However, the dynamical mechanism of this variability and its relation to EHF has not been thoroughly investigated. I find that the variability of EHF decreases dramatically as the season advances, and this reduction shapes the rapid transition in the Arctic SAT.

The anomalous flow pattern associated with the anomalous EHF, identified through
composites based on EHF events, is a dipole like pattern spanning the North Atlantic and Greenland Sea. This dipole pattern also resembles a leading EOF pattern of the geopotential height field north of 50N, which indicates that this pattern is dynamically driven and recurring.

The dipole pattern is a result of the interaction between large-scale dynamics and lower boundary thermodynamic condition. Westward propagating Rossby waves (mostly the wave 1 component) interacts with the temperature ridge at lower levels in the North Atlantic, and create a strong eddy heat flux into the Arctic. The sub-seasonal variability of EHF is mainly affected by the dynamics, the propagating wave, and it has a quasi-periodicity of about 25 days. The seasonal change of EHF is basically controlled by the thermodynamic factor, the location and strength of temperature ridge at lower levels, which is the consequence of land-sea contrast. As the season advances into spring, the continent warms faster than the ocean, so the temperature ridge over the North Atlantic is weakened and disappears gradually.

An energetics analysis reveals that the dipole pattern in a certain phase can extract energy from the zonal flow through the baroclinic conversion, and this energy conversion helps to maintain the anomalous flow pattern. The negative feedback between eddy kinetic energy and dissipation also provide an explanation for the quasi-periodicity in the EHF and energetics.

I also use the quasi-geostrophic geopotential height tendency equation as a framework to investigate the dynamical mechanisms for the dipole pattern. The results confirm that the propagating Rossby wave is important to the dipole pattern because the linear advection of QGPV reinforces the east part of the dipole. The nonlinear interaction of low frequency eddies is one of the most important contributors to the growth of the dipole. However, the mechanism of this interaction is not fully understood in this research. In
addition, the interaction between synoptic eddies and the low frequency mean flow also reinforces the dipole pattern. Diabatic processes in lower levels are also important for maintaining the west center of the dipole.

A future projection shows that the Arctic springtime transition become less abrupt or not abrupt at all under climate change. This is mainly due to the change that the EHF is weaker and has less variation. The dynamics has only little change, while the change in land-sea contrast causes the EHF change. Both the variance and mean value of EHF see reductions in the North Atlantic region. The total heat flux at the 850hPa level and moist static energy flux integrated vertically show reductions in the North Atlantic region and increases on the west coast of North America. The increase of energy flux in these region is mainly due to changes in the stationary wave. The overall northward energy flux shows reduction in most latitudes. However, in my model configuration, the Arctic ocean serves as an infinite energy source to the atmosphere in the almost ice-free condition. It is necessary to use fully coupled model to investigate the energy budget precisely.

5.2 Future work

The current research has revealed the dynamic and thermodynamic mechanisms underlying the Arctic springtime transition. However, a few questions are not thoroughly understood.

The amplitude of planetary wave 1 has a bimodal distribution, the +90 phase corresponds to positive EHF and extracts energy from the zonal mean flow. However, what causes the wave to grow in the −90 phase is not clear. Additionally, further work is needed to fully understand what contributes to the growth of wave-2 component.
As was discussed in the last section, the energy budget can not be closed in the current model configuration, so it is necessary to use a fully couple model or use CMIP5 output with a full seasonal cycle to continue the study of energy transport into the Arctic.

Greenland is the largest topography at high latitudes. I have conduct a simulation with Greenland removed. A detailed comparison is needed to investigate how the Greenland topography affects the Arctic climate.

Current simulations are with a resolution of about 1X1. If computation resources allow, it is desirable to repeat these simulations at higher resolutions, which might improve the representation of the transient eddies and moist processes.
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APPENDIX
Appendix A

Appendix

A.1 Quasi-geostrophic geopotential equation

The equation I need to solve is a linear diagnostic PDE:

\[
\left[ \frac{1}{f_0} \nabla^2 + \frac{\partial}{\partial p} \left( \frac{f_0}{\sigma} \frac{\partial}{\partial p} \right) \right] \chi = -\mathbf{V} \cdot \nabla (\zeta + f) + \frac{\partial}{\partial p} \left[ \frac{f_0 R}{\sigma p} \mathbf{V} \cdot \nabla T \right] \tag{A.1}
\]

\( \chi \) is the solution aimed, the RHS of above equation can be considered as forcing field and calculated with the observation or model output variables, and \( \chi \) is the response to these forcing.

Using center finite difference, the LHS of equation A.1 can be written as:

\[
\frac{1}{f_0} \nabla^2 \chi = \frac{1}{f_0} \left( \frac{\partial^2 \chi}{\partial x^2} + \frac{\partial^2 \chi}{\partial y^2} \right) = \frac{\chi(i+1,j,k) + \chi(i-1,j,k) - 2\chi(i,j,k)}{f_0 \Delta x^2} + \frac{\chi(i,j+1,k) + \chi(i,j-1,k) - 2\chi(i,j,k)}{f_0 \Delta y^2}
\]
The vertical term is:

\[
\frac{\partial}{\partial p} \left( \frac{f_0 \partial \chi}{\sigma \partial p^2} \right) = \frac{f_0 \partial^2 \chi}{\sigma \partial p^2} - \frac{f_0 \partial \chi \partial \sigma}{\sigma^2 \partial p \partial p} \\
= \frac{f_0 \chi(k + 1) + \chi(k - 1) - 2\chi(k)}{\Delta p^2} \\
- \frac{f_0 \chi(k + 1) - \chi(k - 1)}{\sigma^2} \cdot \frac{\sigma(k + 1) - \sigma(k - 1)}{2\Delta p}
\]

note that I drop the \(i,j\) subscripts here.

After combining these terms and group the terms associated with the same grid data point (same \((i,j,k)\)),

\[
LHS = A\chi(i + 1) + A\chi(i - 1) + C\chi(j + 1) + C\chi(j - 1) \\
+ E\chi(k + 1) + F\chi(k - 1) + G\chi(i, j, k)
\]

\[
A = \frac{1}{f_0 \Delta x^2} \\
C = \frac{1}{f_0 \Delta y^2} \\
E = \frac{f_0}{\sigma_k(\Delta p)^2} - \frac{f_0}{\sigma_k^2} \cdot \frac{\sigma_{k + 1} - \sigma_{k - 1}}{(2\Delta p)^2} \\
F = \frac{f_0}{\sigma_k(\Delta p)^2} + \frac{f_0}{\sigma_k^2} \cdot \frac{\sigma_{k + 1} - \sigma_{k - 1}}{(2\Delta p)^2} \\
G = -\frac{2}{f_0 \Delta x^2} - \frac{2}{f_0 \Delta y^2} - \frac{2f_0}{\sigma \Delta p^2}
\]

if the subscripts in the parenthesis is \(i + 1\), it refers to grid point \((i + 1, j, k)\), same reasoning applies to other subscripts. The above formula are used in the interior of the domain; at the boundaries, specified boundary conditions are applied.

There are two ways to apply the boundary conditions. The first one is to use the above formula to calculate \(\chi\) in the interior of the domain, then update the boundary value after each iteration. An alternative way is to incorporate the boundary grid points
into the interior points, then do the iteration together. Here I adopt the second method because it might be more efficient.

A.1.1 Iterative process

For each forcing $\Lambda(i, j, k)$,

$$A\chi(i + 1) + A\chi(i - 1) + C\chi(j + 1) + C\chi(j - 1) + E\chi(k + 1) + F\chi(k - 1) + G\chi(i, j, k) = \Lambda(i, j, k)$$

Give the $\chi$ field a first guess, say all zero, then calculate a guessed $\Lambda_g$,

$$A\chi(i + 1) + \cdots + G\chi(i, j, k) = \Lambda_g$$

$$A\chi(i + 1) + \cdots + G\left[\chi(i, j, k) + \frac{\Lambda - \Lambda_g}{G}\right] = \Lambda$$

If a adjustment term $\frac{\Lambda - \Lambda_g}{G}$ is added to $\chi(i, j, k)$, then at this point $(i, j, k)$, the above equation is perfectly valid; this process is repeated to all points. However, when doing the adjustment to neighbouring points, the preceding point solution is perturbed (the equality is no longer valid because the neighbouring points value is being adjusted). Therefore, the entire grid needs to be scanned many times, and the final result converges to the solution.

A.1.2 Spectral method

Above discussion give the general iteration method to solve the QG height tendency equation. However, in this research, the domain extends from equator to the pole, so
the grid spacing in x dimension approaches to zero in the pole. The iteration method with \( \Delta x \) in it is hard to converge. Instead, a combination of spectral method and iteration is used.

Starting from the equation,

\[
\frac{1}{f_0} \left[ \frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2} + \frac{\partial}{\partial p} \left( f_0^2 \frac{\partial}{\partial p} \right) \right] \chi = f(x, y, p)
\]

Fourier transform the above equation in \( x \) dimension, and equate the coefficient based on linear independence,

\[
\frac{1}{f_0} \left[ -k^2 + \frac{\partial^2}{\partial y^2} + \frac{\partial}{\partial p} \left( f_0^2 \frac{\partial}{\partial p} \right) \right] \hat{\chi}_k = \hat{f}_k(y, p)
\]

\[
\left[ \frac{1}{f_0} \frac{\partial^2}{\partial y^2} + \frac{\partial}{\partial p} \left( f_0 \frac{\partial}{\partial p} \right) - \frac{k^2}{f_0} \right] \hat{\chi}_k = \hat{f}_k(y, p)
\]

\( k = \frac{2n\pi}{L} \) is the zonal wave number, \( n \) is the planetary wave number, and \( L(y) \) is the length of the latitude circle. Here the pole point is still a issue, where \( L \) is zero there. However, the value at pole is not solved, but using boundary condition to update it. Then the boundary conditions (in \( y \) and \( p \) dimensions) is transformed similarly:

\[
\chi_{eq} = 0
\]
\[
\left( \frac{\partial \chi}{\partial y} \right)_{90N} = 0
\]

which leads to:

\[
\hat{\chi}_{eq} = 0
\]
\[
\left( \frac{\partial \hat{\chi}}{\partial y} \right)_{90N} = 0
\]
The bottom boundary condition is provided by the thermodynamic equation.

The lateral boundary condition at pole need to considered carefully. Even though in the physical space, we can arbitrarily impose boundary condition; in the wave number space, some constrains must be satisfied. For wave number 0 (zonal mean), the gradient must be zero at pole. The gradient cross pole must be symmetrical and after taking zonal mean, it ends up as zero. For other wave number (1, 2, 3, ...), the value at pole must be zero (pole must be a node for any wave number other than zero), otherwise, there will be infinite gradient cross pole. So in the spectral method, the lateral boundary condition at pole must be treated separately for wave number 0 and other wave numbers.

Back to the transformed equation, and use finite difference to express the derivatives,

$$\frac{1}{f_0} \frac{\partial^2 \hat{\chi}_k}{\partial y^2} = \frac{\hat{\chi}_k(j + 1) + \hat{\chi}_k(j - 1) - 2\hat{\chi}_k(j)}{f_0 \Delta y^2}$$

The vertical term is:

$$\frac{\partial}{\partial p} \left( \frac{f_0}{\sigma} \frac{\partial \hat{\chi}_k}{\partial p} \right) = \frac{f_0}{\sigma} \frac{\partial^2 \hat{\chi}_k}{\partial p^2} - \frac{f_0}{\sigma^2} \frac{\partial \hat{\chi}_k}{\partial p} \frac{\partial \sigma}{\partial p}$$

$$= \frac{f_0}{\sigma} \frac{\hat{\chi}_k(l + 1) + \hat{\chi}_k(l - 1) - 2\hat{\chi}_k(l)}{\Delta p^2} - \frac{f_0}{\sigma^2} \frac{\hat{\chi}_k(l + 1) - \hat{\chi}_k(l - 1)}{2\Delta p} \cdot \frac{\sigma(l + 1) - \sigma(l - 1)}{2\Delta p}$$

Here I use $l$ as the index in the $p$ dimension because $k$ is used as the wave number in Fourier transform.
Combine these terms, and put terms associated with the same grid point together,

\[
\left[ \frac{1}{f_0 \triangle y^2} \frac{\partial^2}{\partial y^2} + \frac{\partial}{\partial p} \left( \frac{f_0}{\sigma} \frac{\partial}{\partial p} \right) - \frac{k^2}{f_0} \right] \hat{\chi}_k = C \hat{\chi}_k(j + 1) + C \hat{\chi}_k(j - 1) \\
+ E \hat{\chi}_k(l + 1) + F \hat{\chi}_k(l - 1) + G \hat{\chi}_k(i,j,l)
\]

\[
C = \frac{1}{f_0 \triangle y^2} \\
E = \frac{f_0}{\sigma_l(\triangle p)^2} - \frac{f_0}{\sigma_l^2} \cdot \frac{\sigma_{l+1} - \sigma_{l-1}}{(2\triangle p)^2} \\
F = \frac{f_0}{\sigma_l(\triangle p)^2} + \frac{f_0}{\sigma_l^2} \cdot \frac{\sigma_{l+1} - \sigma_{l-1}}{(2\triangle p)^2} \\
G = -\frac{2}{f_0 \triangle y^2} - \frac{2f_0}{\sigma \triangle p^2} - \frac{k^2}{f_0}
\]

Using general iteration method to solve \( \hat{\chi}_k \) for each wave number \( k \), and finally inverse transform \( \hat{\chi}_k \) into \( \chi \), which is the solution of the original equation.

Follow Lau and Holopainen (1984), I truncate at wave number 10, so I need to solve the equation 21 times (one for zonal mean, and two for each of the first ten wave numbers).

A.2 Chapter 1

A.3 Chapter 3
Figure A.1: SAT climatology areal averaged over sub-tropics (0 ∼ 20N), mid-latitude (35 ∼ 55N) and the Arctic (70 ∼ 90N), the annual mean of each of the climatology is subtracted, data source is ERA-interim reanalysis.
Figure A.2: Histogram of transition dates for two control simulations: (a) Forced with 1979-2008 SST and sea-ice climatology and (b) Forced with 1940-1969 climatology.
Figure A.3: Regression of (a) eddy heat flux (unit is $K \cdot m/s$) and (b) geopotential height at 850hPa (units is $m$) ($Z_{850}$) onto the zonal average of eddy heat flux at 70N. Data are from ERA reanalysis, and the smoothed climatology is subtracted before regression. (regression is calculated using anomaly fields)
Figure A.4: Composite map of change in SAT between each lag day and lag 0. Units is K. Data source is ERA-interim reanalysis. Climatology is subtracted before composite, so composite is calculated with anomaly fields.
Figure A.5: Composite map of change in SAT between each lag day and lag 0. Units is K. Data source is Control simulation. Climatology is subtracted before composite, so composite is calculated with anomaly fields.