IMPACTS OF THE HEATING–CIRCULATION RELAY MECHANISM ON
STATIONARY WAVE AND STORM TRACK DYNAMICS

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ABSTRACT

Tropical diabatic heating plays a vital role in driving extratropical circulation variability by exciting extratropical circulation anomalies that redistribute heat and moisture. With moisture being advected to mid-latitudes, extratropical diabatic heating anomalies can arise and excite additional extratropical circulation anomalies downstream. However, this possibility of heating–circulation relay has not yet been explored in the literature. In this dissertation, this heating–circulation relay process is investigated in the context of Northern Hemisphere (NH) stationary waves and storm tracks. First, I investigate the contribution of the proposed relay-circulation sequence to wave interference between stationary waves and transient eddies, which is associated with the amplitude and sustenance of Arctic temperature anomalies. It is found that tropical heating and extratropical heating are not independent of each other, and that Arctic warming has larger magnitudes and persists longer when both tropical and extratropical heating anomalies are large.

Next, I investigate the role of tropical convection in driving the extratropical storm track variability during midwinter, from a global perspective. A comparison of the strong and weak North Pacific eddy kinetic energy years reveals that a midwinter minimum of the North Pacific storm track intensity is in part caused by enhanced tropical convection over the western warm pool. Because tropically excited poleward propagating planetary-scale waves grow by tapping zonal available potential energy of the background state, the planetary-scale wave growth hinders the growth of synoptic-scale eddies, resulting in the midwinter minimum in storm track intensity. Weakened latent heating over the North Pacific storm track promotes a further suppression of the North Atlantic storm track intensity through the heating–circulation relay mechanism.

For stationary waves in climate model simulations, the importance of the heating–circulation relay process in the model representation of stationary waves is further investigated
using a suite of state-of-the-art climate model output. Regarding the upper-tropospheric stationary eddy meridional wind, I show that there are substantial differences between multi-model ensemble and reanalysis data. A set of idealized model experiments with climate model latent heating bias which is estimated from model precipitation field revealed that the systematic model bias in the meridional wind field can be attributed to latent heating bias over the North Pacific, off the coast of the Pacific Northwest, and the heating bias in turn can be attributed to heating biases over the tropical Pacific. This result demonstrates that the heating–circulation relay mechanism also plays an important role in propagating model systematic error in one location and one variable to far-field locations and other variables.

For the interdecadal variability of the relay process, I address a research question as to whether and to what extent tropical and extratropical heating play a role in driving the extratropical isentropic slope variability over the recent few decades. The hypothesized relationship is investigated by observational analyses and idealized model simulations. The results support the hypothesis that the recent trend of the extratropical isentropes is driven by planetary-scale eddy heat fluxes generated by tropical and extratropical heating, and those same diabatic heating anomalies contribute to an occurrence of a warm-Arctic–cold-continent-like temperature pattern.
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Chapter 1

Introduction

1.1 Motivation

A defining characteristic of Earth’s climate is its distinct zonal asymmetry. Stationary waves are zonally asymmetric components of the atmospheric circulation that shape geographical variations of climate, and variation in the structure of stationary waves is an important driver of atmospheric phenomena in the current climate and regional climate variability. Although stationary waves can be forced by zonally asymmetric topography, land-sea contrast, diabatic heating, and transient eddy fluxes, stationary wave studies showed that climatological diabatic heating plays a critical role in shaping the climatological stationary waves throughout the seasonal cycle (Ting and Yu 1998; Ting and Wang 1999; Held et al. 2002; Chang 2009), with the largest amplitude during boreal winter. Because a localized tropical convective heating generates a poleward-propagating Rossby wave train through vortex stretching and absolute vorticity advection by the divergent horizontal flow (Hoskins and Karoly 1981; Sardeshmukh and Hoskins 1988), an increase of zonal asymmetry in tropical heating would augment poleward heat transports driven by planetary-scale waves, which may lead to warming over high-latitudes and enable equable climate conditions – the relatively weak equator-to-pole temperature gradient.

Prompted by the relationship between tropical convective heating and extratropical circulation change, a dynamic mechanism that links tropical convection, poleward propagating planetary-scale waves, and Arctic warming has been proposed. The mechanism is referred to as the tropically excited Arctic warming (TEAM; Lee 2014) mechanism. The TEAM mechanism builds upon the findings that a large, untapped reservoir of zonal available potential energy
(ZAPE; Lorenz 1955) forms during wintertime (Peixoto and Oort 1992), and this quantity can be
tapped by disturbances outside the extratropical baroclinic zone, such as the tropically forced
planetary-scale waves (Lee 2012, 2014). Previous studies found that the TEAM mechanism
operates on various time scales, including the intraseasonal (Yoo et al. 2012a; Baggett and Lee
2015; Goss et al. 2016; Baggett and Lee 2017), the interannual (Lee 2012), and the interdecadal
time scales (Yoo et al. 2011; Gong et al. 2017; Clark and Lee 2019).

From the viewpoint of the Lorenz energy cycle, Baggett and Lee (2015) tested the TEAM
mechanism by examining the planetary- and synoptic-scales of the eddy life cycle. They showed
that during the planetary-scale eddy kinetic energy (EKE) life cycle, enhanced convection over
the western Pacific warm-pool excites a Rossby wave train that taps the ZAPE. The amplification
of climatological stationary waves through constructive interference results in enhanced poleward
heat transports and a longer duration of Arctic warming than that found during the synoptic-scale
EKE life cycle. Moreover, as moisture is channeled by planetary-scale waves toward high-
latitudes during the planetary-scale EKE life cycle (Baggett et al. 2016), additional latent heating
can be generated over the extratropics and act as stationary wave forcing; namely, the growth of
planetary-scale eddies can be triggered by tropical convection and amplified by extratropical
heating. Considering that extratropical heating is a stationary wave forcing as important as
tropical heating (Chang 2009), one may hypothesize that there is a sequence of anomalous
diabatic heating and the associated circulation anomalies as follows: 1) Tropical heating induces
extratropical circulation anomalies. 2) These circulation anomalies then drive extratropical latent
heating anomalies through moisture transport. 3) The extratropical heating anomalies in turn
induce circulation anomalies farther downstream, which influence broader extratropical regions.

A heating–circulation relay process can be also useful for understanding variations in
mid-latitude synoptic-scale eddies. In the extratropical troposphere, baroclinic eddies primarily
develop over the North Pacific Ocean and the North Atlantic Ocean (Blackmon et al. 1977), and
these concentrated regions of synoptic-scale eddies are referred to as storm tracks. Because the regional temperature and hydroclimate variability are closely linked to storm track variability, understanding the dynamic mechanism that regulates the meridional displacement and intensity of storm tracks is critical. A long-standing research question in storm track dynamics is that the North Pacific storm-track intensity reaches a temporal minimum during January despite its maximum baroclinity, upon which baroclinic eddies feed (Nakamura 1992). Although mechanisms to date mostly consider local processes that can influence the North Pacific storm track, recent studies found that the signature of the midwinter minimum is also observed in the North Atlantic storm track intensity (Afargan and Kaspi 2017) or zonal-mean storm-track intensity (Shaw et al. 2018). From the perspective of the hemispheric energetics, the depletion of ZAPE owing to tropical convection results in the growth of planetary-scale EKE (Baggett and Lee 2015). In the meantime, the same process also results in the reduced growth of synoptic-scale EKE, because less ZAPE is available for the growth of synoptic-scale EKE. Thus, the midwinter minimum can occur on a hemispheric scale and be explained by the global perspective.

In this dissertation, I explore the importance of the heating–circulation relay process on diverse research topics in large-scale atmospheric dynamics. Specifically, this dissertation aims to understand the role of planetary-scale eddies excited from localized tropical diabatic heating in driving stationary-transient wave interference over the extratropics (Chapter 2), the midwinter minimum of storm track intensity (Chapter 3), upper-level stationary wave bias in climate model simulations (Chapter 4), and an interdecadal trend of extratropical isentropic slopes (Chapter 5). Although each chapter consists of a stand-alone study, I provide a brief introduction and key research questions of each study in the following subsections. Answers to key research questions in each subsection will be provided in the corresponding subsection of Chapter 6, together with a summary and suggested future work for each study.
1.2 Introduction to Chapter 2: *Relationship between Tropical and Extratropical Diabatic Heating and their Impact on Stationary-transient Wave Interference*

Stationary waves are forced by topography, tropical and extratropical diabatic heating, and transient eddy vorticity fluxes, with their strongest amplitude in winter when both the equator-to-pole temperature gradient and land-sea thermal contrasts are enhanced. According to Northern Hemisphere winter stationary wave theory, zonal asymmetry in climatological tropical diabatic heating is a key forcing for the structure of wintertime stationary waves (Held et al. 2002 and references therein), and recent studies investigated the role of tropical heating in forcing amplification of climatological stationary waves through wave interference (Fletcher and Kushner 2011; Garfinkel et al. 2012; Goss et al. 2016). Specifically, Goss et al. (2016) showed that the western Pacific warm-pool convection triggers stationary-transient wave interference in midlatitudes and leads to Arctic warming through enhanced poleward heat transport. Moreover, Baggett et al. (2016) showed that the same localized tropical convection can drive substantial moisture transports over the extratropics, which may result in additional latent heat release. Prompted by these previous findings, I investigate how anomalies in tropical and extratropical diabatic heating contribute to stationary wave interference, as well as the subsequent impact of this interference on Arctic temperature anomalies, by performing observational data analyses and idealized model experiments. The goal of this study is to address the following two key questions.

1. During stationary-transient wave interference, to what extent are constructive interference and Arctic temperature anomalies driven by latent heating in the tropics, and by latent heating in the extratropics?

2. During stationary-transient wave interference, how might tropical latent heating and extratropical latent heating be related to each other? Does the circulation driven by the tropical heating organize extratropical latent heating?
1.3 Introduction to Chapter 3: A Mechanism for the Midwinter Minimum in North Pacific Storm Track Intensity from a Global Perspective

According to the linear baroclinic instability theory, the growth rate of baroclinic eddies is contingent upon the magnitude of the meridional temperature gradient and static stability. The climatological storm track intensity over the North Pacific, however, exhibits a minimum during January when lower-tropospheric baroclinity is at a maximum and the upper-tropospheric zonal wind is strongest over the North Pacific (Nakamura 1992). This counterintuitive feature of North Pacific storm tracks is referred to as a midwinter minimum and has been a subject of active research in storm track dynamics over the decades. Recently, Afargan and Kaspi (2017) found that the midwinter minimum is also observed for North Atlantic storm tracks when the Atlantic jet is relatively strong. Also, Shaw et al. (2018) used a moist static energy framework to diagnose zonal-mean storm-track intensity and found a midwinter minimum in the Northern Hemisphere seasonal storm-track intensity, which is mostly driven by contributions from stationary eddies. These findings in turn raise the possibility that the midwinter minimum can be understood from a global perspective, rather than a local perspective that many proposed mechanisms have adapted.

One way to approach the global midwinter minimum is an investigation of the seasonal cycle of the energetics within the theoretical framework of the Lorenz energy cycle. Baggett and Lee (2015) showed that the growth of planetary-scale EKE on a hemispheric scale can be achieved by the forced tapping of ZAPE, and it is tropically forced planetary-scale waves that can tap a large reservoir of ZAPE during boreal winter without any dependence on the flux-gradient relationship. Because the depleted ZAPE state is less favorable for the growth of synoptic-scale storm-track eddies, it is possible that the midwinter minimum occurs globally during winters with reduced ZAPE and enhanced planetary-scale EKE. Thus, I test this hypothesis by analyzing seasonal variations and lead-lag relationships of ZAPE, planetary-scale EKE, synoptic-scale
EKE, and western Pacific warm-pool convection in two different subsets separated by the midwinter North Pacific storm track intensity. The following key questions are addressed in this study.

1. Does the midwinter minimum in storm track intensity occur on a hemispheric scale?
2. Is there any relationship between tropical convection and suppression of global storm-track intensity, and how is this relationship explained using the Lorenz energy framework?

1.4 Introduction to Chapter 4: Is the Stationary Wave Bias in CMIP5 Simulations Driven by Latent Heating Biases?

Under global warming, changes in atmospheric circulation are closely linked to the variability of surface temperature and precipitation extremes on a regional scale (e.g., Horton et al. 2015; Agel and Barlow 2020). For instance, using projections in the Representative Concentration Pathway 8.5 (RCP 8.5) emissions scenario, Simpson et al. (2016) showed that changes in the stationary eddy meridional wind field are associated with changes in the precipitation minus evaporation field over North America during boreal winter. However, there is still substantial uncertainty in regional circulation responses to climate change, partly due to uncertainties in the stationary wave response (Shepherd 2014; Wills et al. 2019). Using phase 5 of the Coupled Model Intercomparison Project (CMIP5) models, Lee et al. (2019) found that the models have systematic stationary eddy meridional wind biases (i.e., the difference between models and reanalysis) poleward of 60°N during boreal winter. Specifically, the 250-hPa eddy meridional wind is underestimated (overestimated) over the East Siberian Sea and Greenland (the Beaufort Sea and Europe). The projected changes in RCP 8.5 exhibit a spatial pattern that resembles the bias field, suggesting that the projection in part reflects an amplification of the
model bias. These stationary eddy biases contribute to the development of regional biases in moisture flux into the Arctic, which lead to regional biases in Arctic warming.

A key finding of Chapter 2 is that tropical latent heating and extratropical latent heating are not independent of each other, because extratropical circulation anomalies induced by tropical heating can lead to additional latent heat release locally. It is shown that this latent heating anomaly could further induce circulation anomalies and affect stationary wave interference, resulting in a heating–circulation relay process. In light of the heating–circulation relay mechanism, I hypothesize that biases in atmospheric latent heating may be an important contributor to the stationary wave bias in CMIP5 models. To test this hypothesis, I examine 250-hPa stationary eddy meridional wind and total precipitation during boreal winter in both reanalyses and CMIP5 historical simulations. To explore the causality between the latent heating and stationary wave biases, I also conduct idealized model experiments and examine model circulation response to CMIP5 precipitation bias forcing. In this study, I assume precipitation bias can be treated as latent heating bias in climate model simulations, and address the following key questions.

1. What are spatial patterns of stationary wave and precipitation biases in CMIP5 simulations during boreal winter?
2. Can precipitation biases induce the stationary wave bias in climate models?
3. Which regional precipitation bias contributes to the development of the meridional wind bias pattern?
1.5 Introduction to Chapter 5: *The Role of Planetary-scale Eddies on the Recent Isentropic Slope Trend during Boreal Winter*

Baroclinic eddies are known to arise from baroclinic instability in the region where the meridional slope of isentropic surfaces is steep. According to the baroclinic adjustment theory, these eddies in turn neutralize the atmosphere by flattening the isentropes such that the isentropic slope is maintained at its marginal state for baroclinic instability (Stone 1978). However, contrary to this theory, according to which the isentropic slope should remain constant, the recent trend of Arctic warming raises the possibility that there has been a systematic trend in the extratropical isentropic slope over the same time period. Recently, Thompson and Birner (2012) examined the relationship between the extratropical isentropic slope and lower tropospheric eddy fluxes during boreal winter. They found that low-frequency eddy fluxes play a significant role in driving the extratropical isentropic slope variability. Because low-frequency, planetary-scale eddies mostly contribute to poleward heat and moisture transports, which lead to Arctic warming (Baggett and Lee 2015; Graversen and Burtu 2016; Gong et al. 2017; Lee et al. 2019; Papritz and Dunn-Sigouin 2020), it is reasonable to hypothesize that both interdecadal trends of Arctic warming and the extratropical isentropes are driven by the same process that excites planetary-scale eddy activities.

In Chapter 2, the relay process of latent heating and circulation anomalies is explored, which links planetary-scale eddies excited from tropical convection to Arctic warming within the TEAM mechanism framework. In light of this dynamic picture, I test the hypothesis that the recent wintertime trend of the extratropical isentropic slopes is driven by planetary-scale eddy heat fluxes generated by tropical and extratropical latent heating. A relevant index to capture the extratropical isentropic slope variability is constructed. Composite analyses are performed to investigate lead-lag relationships among isentropic slopes, temperature, synoptic- and planetary-scale eddy heat flux anomalies. Initial value calculations with an idealized model are followed to
explore the causality between latent heating and extratropical isentropic slope anomalies. The following questions are addressed in this study.

1. Does the extratropical isentropic slope exhibit a significant trend over the past few decades? If so, is this recent trend mainly driven by an enhancement of planetary-scale eddy activity?

2. Is there a physical linkage between tropical convection and extratropical isentropic slopes?

3. Does an idealized model perturbed by latent heat forcing reproduce a warm-Arctic–cold-continent-like temperature pattern associated with the extratropical isentropic slope trend pattern?
Chapter 2

Relationship between Tropical and Extratropical Diabatic Heating and their Impact on Stationary-transient Wave Interference¹

2.1 Abstract

During boreal winter, the climatological stationary wave plays a key role in the poleward transport of heat in mid- and high latitudes. Latent heating is an important driver of boreal winter stationary waves. In this study, the temporal relationship between tropical and extratropical heating and transient-stationary wave interference is investigated by performing observational data analyses and idealized model experiments. In line with stationary wave theory, the observed heating anomaly fields during constructive interference events have a spatial structure that reinforces the zonal asymmetry of the climatological heating field. The observational analysis shows that about ten days prior to constructive interference events, tropical heating anomalies are established, and within one week North Pacific and then North Atlantic heating anomalies follow. This result suggests that constructive interference involves a heating–circulation relay: tropical latent heating drives circulation anomalies which transport moisture in such a manner to increase latent heating in the North Pacific; circulation anomalies driven by this North Pacific heating similarly lead to enhanced latent heating in the North Atlantic. This heating–circulation relay picture is supported by initial-value model calculations in which the observed heating anomalies are used to drive model circulations. Our results also show that the constructive interference

driven by both tropical and extratropical diabatic heating generates a relatively large amplitude wave in high latitudes and leads to particularly prolonged Arctic warming episodes, whereas when both the tropical and extratropical diabatic heating are weak, constructive interference is confined to midlatitudes and does not lead to Arctic warming.

2.2 Introduction

The planetary-scale stationary wave plays an important role for regulating weather and climate in the Northern Hemisphere extratropics during the winter. Constructive interference with transient waves accentuates the effect of the stationary wave. In particular, it has been shown that stationary-transient wave constructive interference plays an important role in the extratropical response to tropical forcing (Fletcher and Kushner 2011; Smith et al. 2011; Garfinkel et al. 2012; Goss et al. 2016). It has been proposed that constructive interference is also likely to play a crucial role in maintaining equable climate conditions, i.e., where the equator-to-pole temperature gradient is relatively weak, even during winter (Lee 2012, 2014). The basis of this equable climate theory is that the stationary wave is a major contributor to the poleward energy flux in mid-to-high latitudes during boreal winter (Peixoto and Oort 1992); zonal asymmetry in the climatological tropical latent heating, while independent of the equator-to-pole temperature gradient, is important for forcing the stationary wave (Hoskins and Karoly 1981; Ting and Held 1990; Ting 1996; Held et al. 2002; Chang 2009); and there is a large untapped reservoir of zonal available potential energy (Lorenz 1955) which in principle can be unleashed by disturbances from outside of the extratropical baroclinic zone, such as the tropics (Lee 2014).

Goss et al. (2016; GFL, hereafter) hypothesized that constructive interference would be preceded by tropical heating anomalies that reinforce the climatological zonal asymmetry, and that it would be followed by anomalous warming in high latitudes. To test this hypothesis, GFL
constructed an index, referred to as the stationary wave index (SWI), which is positive (negative) for constructive (destructive) interference that measures the degree to which the instantaneous daily 300-hPa streamfunction field matches with its winter climatology. It was found that the SWI tends to be positive 7–10 days after an enhanced warm pool convection, which enhances the zonal asymmetry of the climatological heating field in the tropics. Positive SWI days are typically followed by Arctic warming 8–10 days later. Examining the stationary wave responses to individual forcing terms, presented by Held et al. (2002), we find that the responses to diabatic heating fields—tropical and extratropical heating individually—exhibit circulation patterns that promote warmth in the Arctic. Specifically, the wave solutions are comprised of a southerly flow over the Bering Strait and Greenland and Norwegian Seas where most of the warm, moist air intrusions occur (Woods et al. 2013; Baggett et al. 2016; Woods and Caballero 2016). The southerly flow is especially prominent in the response to the extratropical heating.

Extratropical heating and tropical heating are unlikely to be independent of each other, however. The results of Baggett et al. (2016) raise the possibility that tropical heating leads to Arctic warming not just by its direct excitation of poleward and upward propagating planetary-scale waves, but also by amplification of these planetary-scale waves through tropically induced extratropical diabatic heating anomaly. Therefore, we hypothesize that during SWI days, there is a latent heat–circulation relay process that takes place as follows: tropical latent heating anomalies excite extratropical circulation anomalies; the extratropical circulation then drives extratropical latent heating anomalies by transporting moisture; the resulting extratropical latent heating, in turn, drives additional extratropical circulation. This hypothesized sequence is somewhat akin to how a hybrid vehicle operates in the sense that the tropical latent heating is analogous to gasoline fuel put into the vehicle and the extratropical latent heating is analogous to electric power generated by running the vehicle. This hypothesis of a latent heat–circulation
relay is also supported by the studies of Willison et al. (2013) and Papritz and Spengler (2015) who showed that enhanced synoptic-scale wave activity is often associated with additional latent heat release.

In this study, we test the relay hypothesis in the context of the SWI by performing observational analyses and initial-value model calculations. Specifically, we address the following two questions: (1) During SWI events, what circulation and temperature anomalies are induced by the individual heating anomalies over the tropics and extratropics? (2) During SWI events, is there any evidence that tropical heating anomalies lead the extratropical heating anomalies, and then the stationary-transient wave interference? In Section 2.3, we describe the data, analysis methods, and the setup of the idealized model experiments. The results of our observational analyses are presented in Section 2.4, while the model results are discussed in Section 2.4. Lastly, Section 2.5 provides a discussion and summary.

2.3. Data and Methods

2.3.1 Data

Daily data of zonal wind, meridional wind, temperature, specific humidity, vertically integrated eastward/northward moisture flux, and surface pressure have been acquired from European Centre for Medium-Range Weather Forecasts (ECMWF) Interim reanalysis (ERA-Interim; Dee et al. 2011). All variables are on a grid with 2.5°×2.5° horizontal resolution and with 23 pressure levels, except for the vertically integrated moisture flux and surface pressure. The time period examined here is December–February (DJF) for years 1979–2014. For our study, latent heating generated by convection and large-scale condensation are important variables. Because these individual latent heating variables are unavailable in the ERA-Interim reanalysis
data, we use diabatic heating data provided by the Japanese 55-year reanalysis (JRA-55; Kobayashi et al. 2015) with a 2.5°×2.5° horizontal resolution and 37 pressure levels.

There are differences in the climatological diabatic heating among different reanalysis datasets (Ling and Zhang 2013; Wright and Fueglistaler 2013). However, Clark and Feldstein (2020) show excellent agreement in non-radiative diabatic heating between the ERA-Interim and JRA-55 Reanalyses in their Northern Atlantic Oscillation (NAO) composites. This comparison is possible because ERA-Interim provides total diabatic heating and radiative heating; non-radiative diabatic heating was computed by subtracting radiative heating from the total diabatic heating. In the JRA-55 data, the non-radiative diabatic heating was computed by summing the convective heating $Q_{\text{conv}}$, large-scale condensational heating $Q_{\text{lrg}}$, and subgrid-scale vertical diffusion. Also, Zhang et al. (2017) calculated the composite values of non-radiative diabatic heating against outgoing longwave radiation over the Western Pacific at 400 hPa from ERA-Interim, JRA-55, and The Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2), and found that the agreement between ERA-Interim and JRA-55 stood out. As a further test to ascertain that our results are insensitive to the choice of the data set, we computed the stationary wave index (defined in Section 2.3.2) using the JRA-55 reanalysis and compared it with the same index computed using the ERA-Interim data. The correlation coefficient between the two time series turned out to be 0.964. In addition, the 300-hPa streamfunction anomaly composites (not shown) from the JRA-55 dataset are essentially indistinguishable from those from the ERA-Interim (Fig. 2.1). These results lend confidence that JRA-55 $Q_{\text{conv}} + Q_{\text{lrg}}$ can be used to investigate the relationship between the latent heating and the circulation.
2.3.2 Stationary wave index and diabatic heating index

Following GFL, we employ the projection method of Feldstein (2002) to quantify the intensity of wave interference:

\[ P_{SW}(t) = \frac{\sum_i \sum_j \Psi'(\lambda_i, \theta_j, t) \Psi^* (\lambda_i, \theta_j, d) \cos \theta}{\sum_i \sum_j [\Psi^* (\lambda_i, \theta_j, d)]^2 \cos \theta}, \]  

(1)

where \( P_{SW}(t) \) represents the projection time series of stationary wave for a specific day \( t \), and \( d \) is determined by converting \( t \) to the corresponding day of the year. \( \Psi \) indicates the annual cycle, obtained by calculating the calendar-day mean of the 300-hPa streamfunction \( \Psi \), and \( \Psi^* \) denotes the zonal-mean removed annual cycle, thus the 300-hPa climatological stationary wave. The variables \( \lambda \) and \( \theta \) represent the longitude at zonal grid point \( i \) and the latitude at meridional grid point \( j \), respectively. We calculate the daily 300-hPa streamfunction anomalies \( \Psi' \) by subtracting the annual cycle \( \Psi \) from the daily 300-hPa streamfunction \( \Psi \). If there is constructive (destructive) interference at a given day, the projection value is positive (negative). More details on calculating the climatological streamfunction can be found in GFL. Finally, \( P_{SW}(t) \) is normalized by its standard deviation during DJF to generate the daily SWI.

In a manner similar to that for the SWI construction, a daily projection time series for tropical, North Pacific, and North Atlantic latent heating anomalies are computed by using the following equation (2). For the sake of conciseness, henceforth, we use the term heating to refer to the summation of \( Q_{\text{env}} \) and \( Q_{\text{irg}} \), vertically averaged (pressure weighted) from 950 hPa to 150 hPa, and denoted by \( Q \).

\[ P_{Q}(t) = \frac{\sum_i \sum_j Q'(\lambda_i, \theta_j, t) \overline{Q_{SWI}(\lambda_i, \theta_j)} \cos \theta}{\sum_i \sum_j [\overline{Q_{SWI}(\lambda_i, \theta_j)}]^2 \cos \theta}, \]  

(2)

where \( Q' \) is the daily diabatic heating anomaly field obtained by subtracting the smoothed calendar-day climatology of \( Q \) to remove the seasonal cycle. \( \overline{Q_{SWI}} \) is the time average of the
anomalous heating field during the days when either constructive interference (SWI > 1.0, hereafter SWI+) or destructive interference (SWI < −1.0, hereafter SWI−) occurs.

For the projection domains (λᵢ, θⱼ), we define the tropics as 30°S–30°N, 0°E–360°E, and the North Pacific (North Atlantic) domain as latitudes between 30°N and 70°N (30°N and 80°N) and longitudes between 150°E and 260°E (280°E and 360°E). The North Pacific (North Atlantic) domain is indicated by the yellow (green) box in Fig. 2.1j. As was shown by GFL, tropical diabatic heating associated with the SWI tends to peak during lag days −10 to 0 relative to the peak of SWI, where lag day 0 corresponds to the days when SWI exceeds one standard deviation. Therefore, for the tropical domain, Q_{SWI} is obtained by averaging the anomalous heating field from lag days −10 to 0. Figure 2.1 shows that organized extratropical heating anomalies tend to become established several days after the Warm Pool heating. For the North Pacific (North Atlantic) domain, Q_{SWI} is obtained by averaging from lag days −6 to 0 (lag days −6 to +3). Once again, the resulting daily projection time series from equation (2) are normalized, and the results are referred to as the tropical heating index which is denoted as T⁺(T⁻) if Q_{SWI} is the tropical heating anomalies during the SWI+ (SWI−) days. Similarly, if Q_{SWI} is the North Pacific heating anomalies during the SWI+ (SWI−) days the normalized projection is referred to as the North Pacific heating index, and denoted as P⁺(P−); for the North Atlantic heating anomalies, the resulting indices are denoted as A⁺ (A−) for SWI+ (SWI−) days. The time evolution of daily heating indices relative to the daily SWI is presented in Fig. 2.2. Consistent with the aforementioned temporal behavior of the regional heating, which is based on a visual inspection of Fig. 2.1, Fig. 2.2a (for SWI+) and Fig. 2.2b (for SWI−) show that tropical heating anomalies lead the extratropical anomalies by about 5 days.
2.3.3 Binning procedure based heating indices

Lastly, to answer the question (1) posed in Section 2.2, we divided the SWI days into multiple bins by sorting them based on the magnitude of the three heating indices. Because tropical heating anomalies occur first (Fig. 2.2), the tropical heating index is used as the first criterion for creating the bins. For example, all SWI+ days are divided into three bins, with the first bin corresponding to those SWI+ days with the top one-third of the $T^+$ values, and the second and third bin corresponding to SWI+ days with the middle and bottom one-third of the $T^+$ values, respectively. We denote the first bin as $T^+_E$ and the third bin as $T^+_F$, where the subscript T (B) stands for top (bottom) one-third. The results for the second bin show characteristics that are between those of the first and third bins, hence they are not presented.

The North Pacific (Atlantic) domain is then used as the second (third) criterion. For these extratropical heating criteria, only two bins are used to retain a sufficient number of SWI days in each of the bins; within the $T^+_E$ bin, the SWI+ days are divided into two additional bins ranked by $P$. This procedure generates two secondary bins $P_T|T^+_E$ and $P_B|T^+_E$, where A|B denotes condition A at given condition B. For example, $P_T|T^+_E$ denotes the top one-half Pacific heating days among the top one-third tropical heating SWI+ days. The $P_T|T^+_E$ bin is further divided into two additional bins according to the Atlantic heating index: $A_T|P_T|T^+_E$ and $A_B|P_T|T^+_E$, where the former (latter) denotes the top (bottom) one-half Atlantic heating days in the $P_T|T^+_E$ bin. Ultimately, this procedure yields four tertiary bins for the $T^+_E$ bin: $A_T|P_T|T^+_E$, $A_B|P_T|T^+_E$, $A_T|P_B|T^+_E$, and $A_B|P_B|T^+_E$. The behavior of the SWI+ days in all four bins are worthy of investigation, but to address the questions raised in Section 2.2, our analysis will be limited to $A_T|P_T|T^+_E$ and $A_B|P_B|T^+_E$ bins. If the Atlantic heating index is used as the second criterion to create $P_T|A_T|T^+_E$ and $P_B|A_B|T^+_E$ bins, the SWI+ days that belong to these bins overlap with the SWI+ days in $A_T|P_T|T^+_E$ and $A_B|P_B|T^+_E$ by 95% and 87.5%, respectively, indicating that the
membership of the SWI+ days is qualitatively insensitive to the order that the Pacific and Atlantic heating criteria are applied. Therefore, henceforth we denote $A_T | P_T | T_T^+$ as $E_T | T_T^+$ and $A_B | P_B | T_T^+$ as $E_B | T_T^+$, where $E$ stands for extratropics. The same procedure is applied to $T_B^+$, yielding four tertiary bins. For our purpose, again, $A_T | P_T | T_B^+$ and $A_B | P_B | T_B^+$ bins are analyzed, which will be denoted as $E_T | T_B^+$ and $E_B | T_B^+$ bins, respectively.

Summarizing the binning procedure, SWI+ days are divided into multiple bins and, among these bins, we analyze $E_T | T_T^+$, $E_B | T_T^+$, $E_T | T_B^+$, and $E_B | T_B^+$ bins. These four bins, respectively, represent the SWI+ days when both tropical and extratropical heating are relatively strong, the SWI+ days when tropical heating is strong and extratropical heating is weak, the SWI+ days when tropical heating is weak and extratropical heating is strong, and the SWI+ days when both tropical and extratropical heating are weak. Using the same procedure, SWI− days are also divided into multiple bins and four of those bins are examined: $E_T | T_T^-$, $E_B | T_T^-$, $E_T | T_B^-$, and $E_B | T_B^-$. The criteria of the 8 bins analyzed in this study are summarized in Table 2.1.

For the purpose of analyzing time sequence of events, we identify individual events based on the following procedures. For each bin, an SWI event is defined as a 15-day time interval that contains at least one SWI day that satisfies the binning criteria and the day of the maximum (or minimum for destructive interference) SWI value. The eighth day within the 15-day interval coincides with the maximum or minimum SWI value, and is defined as the SWI lag-0 day. With this definition, consecutive SWI events are separated from each other by at least 7 days. For the statistical significance of the composite values presented in Figs. 2.1–2.5, Monte Carlo simulations are performed. For composites based on N number of events, we generated 1000 composites, with each composite consisting of N randomly chosen events with each event having a time interval extending from lag -10 to lag +10 days (Fig. 2.1) and from lag −20 to lag +20 days (Figs. 2.2–2.5). A null distribution was then constructed from the 1000 random composites for...
each lag day, and the $p$ value of the observed composite was computed based on the distribution. Again following GFL, to account for the fact that the consecutive days are not independent of each other, for Fig. 2.1, we divided N number of days by 3.38 and rounded to the nearest integer. In Fig. 2.1, the stippled region indicates statistical significance at the $p < 0.10$ level, while in Figs. 2.2–2.5, the $p < 0.05$ level is indicated.

2.3.4 Model experiment setup

To test the causal relationships that emerge from our diagnostic analyses, we utilize the spectral dynamic core from the NOAA Geophysical Fluid Dynamics Laboratory (GFDL). The same model setup as in Baggett et al. (2016) is used in this study: a horizontal resolution of triangular 42, a vertical resolution of 28 sigma levels, a damping timescale of 0.1 days at its smallest scale for a fourth-order horizontal diffusion, Newtonian cooling and Rayleigh friction parameterized as in Held and Suarez (1994). The climatological DJF values of zonal wind, meridional wind, temperature, and surface pressure are used as the initial background state. For passive tracer experiments, we use the zonal mean climatological specific humidity field. In order to ensure that the zonally varying DJF climatological state is a solution to the model equations, we add a forcing term which is obtained by integrating the model by one time step starting from the climatological state. This forcing term prevents the model from drifting from the initial state (the DJF climatological state) unless additional forcing, e.g., diabatic heating, is added. Additional details on the model setup and limitations can be found in Franzke et al. (2004). In the control experiment, the model is integrated without any additional forcing, whereas for perturbation experiments, the model is forced by time-dependent (lag days $-10$ to $+7$) $E_T |\mathcal{F}_T^+|$ composite heating (Fig. 2.3a): at model day 1, the forcing incorporated into the model integration is the lag day $-10$ composite; at model day 2, the forcing is lag day $-9$ composite, and so on. Each
model experiment was run for 25 days. The diabatic heating composites, computed at the JRA55’s 37 pressure levels, are interpolated into model’s 28 sigma levels.

2.4. Observation analysis

2.4.1 Relationship between circulation and heating anomalies during SWI

We first examine the 300-hPa streamfunction evolution for SWI+ days (Figs. 2.1a-g). Because GFL provides a detailed description of the same fields, we present just a few key highlights pertinent to this study. It can be seen that the streamfunction anomalies (shading) in Fig. 2.1 grow and decay over a period of about two weeks. For the SWI+, by construction, the positive (negative) anomalies coincide with the climatological ridges (troughs). The opposite is the case for the SWI– composites (see Fig. B1). Poleward of 50°N, there are two ridges, one centered at the eastern end of Gulf of Alaska, and the other centered over the British Isles. As such, associated with these ridges, there are southerly flows over the two ocean corridors to the Arctic Ocean, one over the Bering Sea and the other over the Norwegian Sea.

Figures 2.1h–n illustrate the composite heating anomalies during SWI+ days. When there is constructive interference, consistent with GFL, there is enhanced warm-pool convection and suppressed convection in central tropical Pacific, reinforcing the zonal asymmetry of the climatological heating field. In the extratropics, which was not examined by GFL, a positive anomaly is seen over the central North Pacific, while a negative anomaly is found over the eastern North Pacific and western North America. These heating anomalies appear at lag day −6 and dissipate during positive lag days (Figs. 2.1i–m). We also see heating anomalies over the northeastern Atlantic and cooling over subtropical Atlantic shortly after the North Pacific heating
is excited. In the case of destructive interference, SWI–, streamfunction and heating anomalies with the opposite signs are found over the tropics and extratropics (see Fig. B1).

Synthesizing the results of the streamfunction and heating field composites, we highlight two findings: Firstly, the SWI+ (SWI–) heating anomaly field reinforces (dampens) the zonally asymmetric component of climatological diabatic heating (e.g., Fig. 8 of Held et al. 2002). This relationship suggests that a strengthening of the zonally asymmetric component of the climatological diabatic heating leads to an amplification of the climatological stationary wave. Because diabatic heating is a major driver of the climatological stationary wave, this transient heating–SWI relationship is consistent with stationary wave theory (e.g., Held et al. 2002; Chang 2009). Secondly, as was briefly discussed in the previous section, the tropical heating anomalies tend to precede the extratropical heating anomalies by several days, suggesting that the SWI extratropical heating anomalies are, at least in part, driven by the circulation driven by the tropical heating anomalies. As a first step to explore this possibility, we next examine lead-lag relationships between tropical and extratropical heating.

2.4.2 Relationship between tropical and extratropical heating anomalies during SWI

Figure 2.2 shows the heating index composites associated with SWI+ (left column) and SWI– (right column). We first examine composites over all SWI+ and SWI– days, that is, $T^+, P^+, A^+$ (Fig. 2a) and $T^-, P^-, A^-$ (Fig. 2.2b). It can be seen that a strengthening of the tropical heating indices, $T^+$ and $T^-$, starting lag day –20, reaching a peak at around lag day –5, while the North Pacific heating indices, $P^+$ and $P^-$, show statistically significant values between lag day –8 and +6, and peaking at lag day –1. Although there are some differences in the time evolution of the North Atlantic heating indices, $A^+$ and $A^-$, both indices slightly lag their respective North Pacific indices. However, even during the SWI+ days, not all tropical heating anomalies would be
followed by the extratropical heating anomalies, and not all extratropical heating anomalies would be preceded by the tropical heating anomalies.

As a way to evaluate the extent of the association amongst these three heating anomalies, the aforementioned composite analysis was repeated, first with the condition that \( T^+ > 1 \) (Fig. 2.2c) and \( T^- > 1 \) (Fig. 2.2f); second with the condition that \( P^+ > 1 \) (Fig. 2.2d) and \( P^- > 1 \) (Fig. 2.2g); and third the condition that \( A^+ > 1 \) (Fig. 2.2e) and \( A^- > 1 \) (Fig. 2.2h). The results show that in all three cases, the onset of tropical heating precedes that of extratropical heating by nearly 10 days. Focusing on the SWI+ days, we find that the statistically significant \( P^+ \) values also precede statistically significant \( A^+ \) values by approximately 3 days. Therefore, we hypothesize the following causal relationships during SWI+ days; tropical heating tends to excite a circulation that enhances extratropical heating, and the North Pacific heating further enhances the North Atlantic heating. This relay hypothesis will be addressed in Section 2.5 using initial-value calculations.

2.4.3 Four types of heating anomalies during SWI and associated circulation pattern

Before presenting results from the model calculations, we first examine the structure of the heating and circulation anomalies for the following cases: \( E_T | T_T^+ \), \( E_B | T_T^+ \), \( E_T | T_B^+ \), and \( E_B | T_B^+ \) for the SWI+ days, and \( E_T | T_T^- \), \( E_B | T_T^- \), \( E_T | T_B^- \), and \( E_B | T_B^- \) for the SWI– days (see Section 2.3.3 for details). Figure 2.3 shows the composite heating anomalies averaged from lag days –10 to +3 in each bin during SWI+ (left column) or SWI– (right column). Every panel displays statistically significant anomalies in the tropics and the midlatitude storm track regions. In the two \( T_T^+ \) bins, we see La Niña-like features — enhanced convection over the Maritime Continent and suppressed convection over the central tropical Pacific (Figs. 2.3a,c). The \( T_B^+ \) bins, on the other hand, show El Niño-like heating pattern (Figs. 2.3b,d). The \( E_B | T_B^+ \) bin (Fig. 2.3d) shows that
stationary wave interference can occur without significant diabatic heating anomalies. Destructive interference composites show opposite sign anomalies, suggesting a linear relationship between the heating and the SWI anomalies.

The zonal asymmetry in the heating anomalies suggests that the SWI events might be influenced by the Madden–Julian Oscillation (MJO; Madden and Julian 1971) which is a 30–60 day tropical variability with eastward propagating convection. By calculating the composites of the real-time multivariate MJO (RMM; Wheeler and Hendon 2004) indices for each bin, we found that in the $J_T^+$ events MJO phase 3–4 signal can be detected starting lag day −20 and reaches phase 6 by lag day 0. For the $E_T|J_T^+$ events, the RMM values are statistically significant only briefly at lag day −5 when it resides in phase 5. The $E_T|J_B^+$ events tend to occur during the same MJO phases as the $E_T|J_T^+$ events, but with smaller amplitudes. The $E_B|J_B^+$ events typically start with phase 6 (lag day −20) and end with phase 2 (lag day +20), with little statistical significance (not shown).

Figure 2.4 shows the 300-hPa streamfunction anomalies in each bin, averaged from lag days −1 to +1. We overlay vectors of vertically integrated moisture flux north of 30°N. In all four bins of SWI+, $Ψ'$ (shading) reinforces $Ψ$ (black contours). However, there are some differences. Figs. 2.4a–d reveal that intense moisture transport into the Arctic is present in the $J_T^+$ bins (Figs. 2.4a,c), whereas it is generally weaker and trapped in the midlatitudes in the two $J_B^+$ bins (Figs. 2.4b,d). The largest moisture flux is found in the $E_T|J_T^+$ bin (Fig. 2.4a), while the weakest vectors are found particularly over the North Pacific in the $E_B|J_B^+$ bin (Fig. 2.4d). During the SWI– days, there are generally negative moisture flux anomalies propagating out of the Arctic, and also the flux anomalies are strongest in $E_T|J_T^-$ bin (Fig. 2.4e).

Figure 2.5 presents temperature anomalies averaged vertically between 700 hPa and 1000 hPa. Within the five days following the SWI events (left column), every bin shows warming at
least either over the Pacific or the Atlantic sector of Arctic Ocean. In later lag days, however, Arctic warming is observed only in the $\mathcal{T}_E^+$ bins (right column). Moreover, the $\mathcal{T}_E^+$ bins show a ‘Warm Arctic–cold continents’ pattern (Overland et al. 2011). The SWI– cases show features that are essentially the same but opposite in sign (not shown). To further quantify the relationship between SWI and Arctic warming in each heating bin, we constructed lag composites of lower tropospheric temperature anomaly averaged poleward of 70°N and the corresponding SWI (Fig. B2). During SWI+, the Arctic warming persists out to +16 days in the $\mathcal{T}_E^+$ bins for the zonal-mean anomalies, but not in the $E_T | \mathcal{T}_E^+$ cases. The warming in the $E_T | \mathcal{T}_E^+$ bin persists only until lag day +7, and the $E_B | \mathcal{T}_E^+$ bin shows no significant zonal-mean temperature anomaly throughout the period. This result suggests that tropical heating is important for prolonged Arctic warming. For the SWI– days, the most prolonged Arctic cooling occurs in the $E_B | \mathcal{T}_E^-$ bin.

### 2.5. Model results

In this section, we use the model (see Section 2.2.4 for details) to test the causal relationships suggested by the observational analysis presented in the previous section. Specifically, we first ask the extent to which the SWI+ heating anomalies can explain the associated circulation and temperature anomalies. Next, we ask if the tropical heating anomalies can lead to the North Pacific and North Atlantic heating anomalies. Because water vapor is the central ingredient of latent heating, to make headway toward addressing the second question, we use a passive tracer to represent moisture. This approach is supported by the earlier studies which show that large-scale advection can account for much of the water vapor distribution in the free troposphere (Pierrehumbert et al. 2007 and references therein; Baggett et al. 2016; Ming and Held 2018).
2.5.1 Circulation and temperature response

For brevity, we only present results for the SWI+ experiments because the model responses to the SWI– heating anomalies are by and large opposite to those of the SWI+ heating. We also confine our analysis to the $E_T | T_T^{+}$ case because both the tropical and extratropical heating anomalies are strong. We first examine the model $\Psi^\prime$ response to the time-dependent $E_T | T_T^{+}$ composite heating during model days 10 to 12, which corresponds to lag day $-1$ to $+1$. Figures 2.6ae depict individual contributions to the $E_T | T_T^{+}$ response from tropical heating, North Pacific heating, and North Atlantic heating, respectively. The summation of these three solutions, shown in Fig. 2.6d, closely resembles Fig. 2.6e where diabatic heating is forced over the entire globe. The result indicates that the model response is mostly linear and that the heating outside of these three domains has a negligible effect on the model solution. Comparing the individual panels in Fig. 2.6, we see that most of the extratropical response is forced by tropical heating and North Pacific heating. Wave trains forced by tropical heating emanate from the western warm-pool and eastern tropical Pacific, reaching eastern Siberia and the eastern subpolar North Atlantic (Fig. 2.6a), while the North Pacific heating drives a wave train that propagates equatorward and downstream of the heating, traversing North America (Fig. 2.6b). In Fig. 2.6c, the North Atlantic heating excites a wave train that spans from the subpolar North Atlantic Ocean to South Asia and another wave train from central Siberia to the western North Pacific. These modeling results suggest that for the $E_T | T_T^{+}$ events, North Pacific heating plays the central role in driving anomalies over the Arctic Ocean. Turning our attention to Figs. 2.6d and 2.6e, it can be seen that a preeminent ridge develops over Alaska and the subpolar North Atlantic which would reinforce the climatological ridge at the same location (Fig. 2.1). It is also worth noting that this subpolar North Atlantic ridge is connected to the subtropical ridge centered over the Gulf of Mexico in Fig. 2.6e. The result is a southwest–northeast tilted ridge spanning from the Gulf of Mexico to the
Greenland Sea. This tilted structure is an important characteristic of the North Atlantic stationary wave and climatological jet.

Figure 2.7 shows the vertically averaged (700–1000 hPa) temperature response averaged from model day 11 to 16, which corresponds to the lag days from 0 to +5 in observation (Fig. 2.5a). In Fig. 2.7a, tropical heating causes warming over eastern Siberia and Greenland and cooling over northern Canada and the Norwegian Sea. The North Pacific heating leads to warming over a broad swath of area poleward of 50°N, ranging from northern North America to Scandinavia, and cooling over eastern Siberia and much of the contiguous United States (Fig. 2.7b). Figure 2.7c shows that the North Atlantic heating results in warming (cooling) over the Barents–Kara Seas and northern Europe (eastern Greenland and the Mediterranean Sea). Figure 2.7d shows the sum of the three model solutions. Compared with the observation (Fig. 2.5a), we see a reasonable agreement except over northeastern Canada and Greenland where the model solution shows warming while the observations show cooling. We found that horizontal temperature advection is the primary contributor to the temperature anomalies (not shown).

Overall, the model temperature solutions indicate that tropical heating is important for warming the Pacific sector of the Arctic, while extratropical heating is more important for warming the Atlantic sector.

2.5.2 Passive tracer response

We next address the question of whether tropical heating anomalies can lead to the North Pacific and North Atlantic heating anomalies. Figures 2.8a and 2.8b show passive tracer fields simulated by the model. The initial condition of the tracer is the three-dimensional climatological specific humidity during boreal winter. During model day 10 to 12, it can be seen that the circulation driven by tropical heating transports anomalous tracer over the central North Pacific
and western North America, as well as over the east coast of North America and the northern subtropical North Atlantic (Fig. 2.8a). Figure 2.8b shows that North Pacific heating also contributes to the positive tracer anomaly over northwestern North America and the northeastern North Atlantic.

Because this passive tracer transport represents moisture transport, the model condensational heating can be estimated (see Appendix A for the method). The results (Figs. 2.8c and 2.8d) show horizontal structures that resemble the composite heating field (Fig. 2.3a), supporting that the lead-lag relationships in Fig. 2.2 are indeed causal. The model result also suggests a positive feedback process where the North Atlantic heating drives a circulation that can reinforce the heating that drove the circulation in the first place. This positive feedback is reminiscent of the mechanism of Hoskins and Valdes (1990), but it differs in the sense that the feedback suggested here is through the interplay between latent heating and circulation while the mechanism of Hoskins and Valdes (1990) is self-maintaining through latent heating, circulation, and baroclinity. When tropical and extratropical heating are imposed together, anomalous tracer transport poleward of 60°N is found mostly over the two ocean corridors where anomalous moisture transport into the Arctic Ocean occurs (not shown). This result shows that when forced by latent heating, moisture transport, via an increase in downward infrared radiation, is likely to reinforce temperature changes caused by dry dynamics (Yoo et al. 2012a).

To quantify the tropics–North Pacific–North Atlantic linkage that involves heating and ‘moisture’ transport, we compute area-weighted averages of the net tracer anomaly over the North Pacific and North Atlantic domains forced by tropical, North Pacific and North Atlantic heating. Figure 2.9a shows that, for the Pacific domain, tropical heating is the main driver of tracer transport, whereas the net effect of extratropical heating is nearly zero. For the Atlantic domain illustrated in Fig. 2.9b, the tracer anomaly forced by North Pacific heating is predominant. To further evaluate the resemblance between the model tracer anomaly and the
reanalysis heating anomaly, we compute the projection of the condensational heating estimated from tracer anomalies, shown in Figs. 2.8c and 2.8d, onto the composites of heating anomalies (\(Q_{SWI}\)) in the same manner as constructing the heating indices. The result again supports the proposed causality.

2.6 Summary and Conclusions

In this study, we addressed the following two questions: (1) During SWI events, what circulation and temperature anomalies are induced by the individual heating anomalies over the tropics and extratropics? (2) During SWI events, is there any evidence that extratropical heating is excited by waves forced by tropical heating?

To address the first question, we binned the SWI+ days by ranking the magnitude of tropical and extratropical latent heating. Our analyses reveal that there are different flavors of stationary wave interferences. Simultaneous enhancements in zonal asymmetries in both tropical and extratropical heating (\(E_T | T_T^+\)) generate circulation interferences not only in the midlatitudes but also in the Arctic, hence the strongest moisture flux into the Arctic. When the zonal asymmetry in the heating field is suppressed (\(E_T | T_T^-\)), stationary wave interference is mostly confined to midlatitudes, and its impact on Arctic temperature is substantially weaker.

To address the second question, we first examined the lead-lag relationships amongst the tropical and the two extratropical heating indices. The result indicates that while the latent heating anomalies in these three domains can occur by themselves, they tend to occur together within 7–10 days of each other, with the tropical heating anomaly leading the North Pacific heating anomaly which in turn is followed by the North Atlantic anomaly. This finding suggests that not only the answer to this second question be positive, but also that the circulation driven by the North Pacific heating enhances latent heating over the North Atlantic domain. We tested this
heating–circulation relay hypothesis with the model by computing condensational heating from the model’s passive tracer which represents specific humidity. This initial-value passive tracer experiment indeed supports the heating–circulation relay hypothesis that emerged from the observational analysis: tropical latent heating $\rightarrow$ circulation anomalies $\rightarrow$ latent heating in the North Pacific $\rightarrow$ circulation anomalies $\rightarrow$ latent heating in the North Atlantic.

The relay picture is a reminder that diabatic heating is not only an important driver of the atmospheric circulation (e.g., Sutcliffe 1951; Hoskins and Valdes 1990), but it also reveals that the extratropical diabatic heating is dependent on tropical heating. The implication is that whilst the atmospheric response to the individual components of the heating is linear, they are not independent of each other. Therefore, the impact of the tropical heating should not be readily dismissed even for a circulation feature attributable to extratropical heating.
### Table 2.1: Acronym definitions after binning procedures, and number of DJF days in each bin out of 3249 all DJF days.

<table>
<thead>
<tr>
<th>Acronyms</th>
<th>Definition</th>
<th>Number of days</th>
<th>Number of events</th>
</tr>
</thead>
<tbody>
<tr>
<td>SWI+</td>
<td>SWI greater than 1.0</td>
<td>502</td>
<td>—</td>
</tr>
<tr>
<td>SWI−</td>
<td>SWI less than −1.0</td>
<td>468</td>
<td>—</td>
</tr>
<tr>
<td>$E_T^+</td>
<td>T_T^+$</td>
<td>Top one-half of extratropical heating bin among top one-third of tropical heating bin during SWI+ days</td>
<td>41</td>
</tr>
<tr>
<td>$E_B^+</td>
<td>T_T^+$</td>
<td>Bottom one-half of extratropical heating bin among top one-third of tropical heating bin during SWI+ days</td>
<td>41</td>
</tr>
<tr>
<td>$E_T^+</td>
<td>T_B^+$</td>
<td>Top one-half of extratropical heating bin among bottom one-third of tropical heating bin during SWI+ days</td>
<td>41</td>
</tr>
<tr>
<td>$E_B^+</td>
<td>T_B^+$</td>
<td>Bottom one-half of extratropical heating bin among bottom one-third of tropical heating bin during SWI+ days</td>
<td>41</td>
</tr>
<tr>
<td>$E_T^−</td>
<td>T_T^+$</td>
<td>Top one-half of extratropical heating bin among top one-third of tropical heating bin during SWI− days</td>
<td>39</td>
</tr>
<tr>
<td>$E_B^−</td>
<td>T_T^+$</td>
<td>Bottom one-half of extratropical heating bin among top one-third of tropical heating bin during SWI− days</td>
<td>39</td>
</tr>
<tr>
<td>$E_T^−</td>
<td>T_B^+$</td>
<td>Top one-half of extratropical heating bin among bottom one-third of tropical heating bin during SWI− days</td>
<td>39</td>
</tr>
<tr>
<td>$E_B^−</td>
<td>T_B^+$</td>
<td>Bottom one-half of extratropical heating bin among bottom one-third of tropical heating bin during SWI− days</td>
<td>39</td>
</tr>
</tbody>
</table>
Fig. 2.1: (a)–(g) DJF total 300-hPa streamfunction (contours with interval of $15 \times 10^6 \text{ m}^2\text{s}^{-1}$) and anomalies (shading), and (h)–(n) vertically averaged latent heating anomalies during the SWI+ days. Dotted areas indicate statistical significance at the 10% level. Statistical significance is evaluated by employing a Monte Carlo simulation with 1000 random samples.
Fig. 2.2: (a)–(g) Lag composites of daily heating indices (black for tropical, red for North Pacific, and blue for North Atlantic) during SWI+ (left) and SWI– (right) days. Each row denotes composites during (a), (b) entire SWI days, (c), (f) SWI and tropical heating index > 1σ days, (d), (g) SWI and North Pacific heating index > 1σ days, (e), (h) SWI and North Atlantic heating index > 1σ days. A 3-day running mean was applied to the heating fields shown in (c)–(h). A Monte Carlo simulation with 1000 random samples is performed for the statistical significance test, and the thick lines indicate statistical significance at the 5% level.
Fig. 2.3: Composites of latent heating anomalies averaged from lag -10 to +3 in (a) $E_T|J^+_T$, (b) $E_T|J^+_B$, (c) $E_B|J^+_T$, (d) $E_B|J^+_B$, (e) $E_T|J^-_T$, (f) $E_T|J^-_B$, (g) $E_B|J^-_T$, and (h) $E_B|J^-_B$. (See Table 2.1 for details on each acronym). A Monte Carlo simulation with 1000 random samples is performed for the statistical significance test, and the dotted areas indicate statistical significance at the 5% level. Prior to the statistical significance test, a nine-point local smoothing was applied twice to the composites.
**Fig. 2.4:** Composites of 300-hPa streamfunction anomalies (shading) averaged from lag days $-1$ to $+1$ in the same subsets as in Fig. 2.3, DJF total 300-hPa streamfunction (contours with interval of $15 \times 10^6$ m$^2$s$^{-1}$). Overlaid vectors represent vertically integrated moisture flux north of 30°N. The vectors whose magnitude is less than 10 kg m$^{-1}$s$^{-1}$ are omitted, and reference vector is 70 kg m$^{-1}$s$^{-1}$. A Monte Carlo simulation with 1000 random samples is performed for the statistical significance test, and the dotted areas indicate statistical significance at the 5% level.
Fig. 2.5: Composites of the lower tropospheric (700-1000 hPa) temperature anomalies during SWI+ days averaged (a)–(d) from lag days 0 to +5 and (e)–(h) from lag days +10 to +20 in the same subsets for SWI+ days as in Figs. 2.3a–d. Gray shading denotes the region where surface pressure is below 700 hPa. A Monte Carlo simulation with 1000 random samples is performed for the statistical significance test, and the dotted areas indicate statistical significance at the 5% level.
Fig. 2.6: The 300-hPa streamfunction anomaly averaged from model day 10 to 12. The model is forced with the heating composite of the $E_T \left| T^+_T \right.$ subset—both tropical and extratropical heating is anomalously large during SWI+ days—for (a) the tropics, (b) the North Pacific domain, (c) the North Atlantic domain, (d) the sum of (a)–(c), and (e) the entire globe.
Fig. 2.7: The lower-tropospheric (700–1000 hPa) temperature anomaly averaged from model days 11 to 16. The same heating composites as in Fig. 2.6 are used to force the model, but only over (a) the tropical, (b) the North Pacific, and (c) the North Atlantic domain. (d) Summation of (a)–(c). Gray shading denotes the region where surface pressure is below 700 hPa.
Fig. 2.8: The vertically integrated tracer anomaly averaged over model days (a) 10–12 and (b) 11–13. The same heating composites as in Fig. 2.6 are used to force the model, but only over (left) the tropical and (right) the North Pacific domain. (c),(d) The vertically integrated condensational heating computed from the tracer anomalies in (a) and (b), respectively (See Section 2.5.2 and Appendix A for details).
Fig. 2.9: (a),(b) Time series of the area-weighted average of the vertically integrated tracer anomaly. (c),(d) Time series of projection of the model condensational heating anomaly computed from the model tracer onto the observed latent heating anomaly field in the $E_T |j_T^+ \text{subset}$ where both tropical and extratropical heating are anomalously large during SWI+. The projection domain is the North Pacific for (a) and (c), and the North Atlantic for (b) and (d). The black, red, and blue lines denote the model response to the tropical, North Pacific, and North Atlantic heating composites, respectively.
Chapter 3

A Mechanism for the Midwinter Minimum in North Pacific Storm Track Intensity from a Global Perspective

3.1 Abstract

The midwinter minimum in North Pacific storm-track intensity is a perplexing phenomenon because the associated local baroclinity in the North Pacific is maximum during midwinter. Here, a new mechanism is proposed wherein the midwinter minimum occurs in part because global planetary-scale waves consume the zonal available potential energy (ZAPE), limiting its availability for storm-track eddy growth. During strong midwinter suppression years, the midwinter minimum is preceded by anomalously large planetary-scale eddy kinetic energy (EKE) and subsequent reduction in ZAPE and global baroclinity. Consistent with previous studies, this large planetary-scale EKE takes place after enhanced Pacific warm-pool convection, which peaks during winter. These results indicate that the midwinter minimum is in part caused by heightened warm pool convection which, through excitation of planetary-scale waves, leads to a weaker storm-track. This finding also helps explain the existence of the midwinter North-Atlantic storm-track minimum.

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3.2 Introduction

During boreal wintertime, the development of baroclinic waves is localized in two regions, one over the North Pacific Ocean and the other over the North Atlantic Ocean (Blackmon et al. 1977). Commonly referred to as storm tracks, these regions can be characterized by relatively high eddy kinetic energy (EKE) at the synoptic timescale (e.g., Chang and Orlanski 1993; Shaw et al. 2016). Linear baroclinic instability theory shows that baroclinic eddy growth hinges on the magnitude of the meridional temperature gradient and static stability (e.g., Charney 1947; Eady 1949; Lindzen and Farrell 1980), and the maximum growth rate is concentrated over the two storm tracks (e.g., Hoskins and Valdes 1990). However, Nakamura (1992) showed that North-Pacific storm-track intensity exhibits a local minimum during January when the lower-tropospheric baroclinity is maximum and the upper-tropospheric zonal wind is strongest. The anticorrelation between the strength of the upper-tropospheric jet and the storm track intensity has been supported by later studies (e.g., Afargan and Kaspi 2017; Chang 2001; Christoph et al. 1997; Zhang and Held 1999; Nakamura and Sampe 2002; Schemm and Schneider 2018; Yuval et al. 2018).

There are a number of theories for the North Pacific midwinter minimum. Chang (2001) found fast eastward energy propagation over the Pacific storm track region, suggesting that storm-track eddy growth is limited by a relatively short transit time over the baroclinic zone. By computing eddy energy budget, Chang (2001) also showed that the generation of eddy available potential energy (EAPE) caused by condensational heating is greater in October than in January. Nakamura et al. (2002) showed that energy conversion from EAPE to EKE is weaker when the North Pacific jet is stronger. Other studies considered changes in the characteristics of the upper-tropospheric jet such as jet width or position (Nakamura and Sampe 2002; Harnik and Chang 2004; Yuval et al. 2018), an enhanced local barotropic damping (e.g., Deng and Mak 2005), a
reduction in eddy diffusivity (e.g., Schemm and Schneider 2018), a sensitivity of eddy activity to the vertical structure of baroclinity (e.g., Yuval and Kaspi 2016), and the vertical tilt of the eddies (e.g., Schemm and Rivière 2019). A common thread of these mechanisms is a local perspective in the sense that the background state confined to the North Pacific plays the central role. Somewhat deviating from this local perspective, it has been proposed that atmospheric conditions upstream of the North Pacific storm track could help explain the midwinter minimum (Nakamura 1992; Ren et al. 2010; Penny et al. 2010; Park et al. 2010).

In this study, we present a hypothesis that rests on a global perspective of storm track dynamics. This hypothesis was motivated by two groups of studies. First, there is evidence that the midwinter minimum is not limited to the North Pacific storm track, but also occurs for the North Atlantic storm track especially when the North Atlantic subtropical jet is anomalously strong (Penny et al. 2013; Afargan and Kaspi 2017). Nakamura et al. (2002) showed that stationary wave anomalies originating from North Atlantic are associated with North Pacific storm track anomalies. This finding was corroborated by Chang and Guo (2007) who showed that both Pacific and Atlantic storm track activities are enhanced during weak Pacific jet years, and that enhanced eddy forcing over the Atlantic Ocean and Asia might have contributed to force Pacific stationary wave and storm track anomalies. Shaw et al. (2018) found that stationary waves play a critical role in modulating storm track intensity. These results suggest that the midwinter minimum can be viewed as a global phenomenon.

Second, it has been shown that during boreal winter, the growth of planetary-scale waves depletes zonal available potential energy (ZAPE) (Lorenz 1955). [The term waves and eddies are used interchangeably throughout this section.] Baggett and Lee (2015) examined the planetary-scale wave life cycle with global reanalysis data, and in contrast to synoptic-scale waves, they found that anomalously large ZAPE is not required for anomalously strong growth of planetary-scale waves. Instead, they showed that the growth of planetary-scale waves is typically preceded
by enhanced tropical convection over the western tropical Pacific warm-pool. The occurrence of warm pool convection prior to planetary-scale wave growth is not surprising given that tropical convection is known to excite poleward propagating planetary-scale waves (e.g., Hoskins and Karoly 1981; Sardeshmukh and Hoskins 1988; Yoo et al. 2012a). The net impact of the planetary-scale wave life-cycle on the zonal mean flow is thus a state of depleted ZAPE. Baroclinic instability theory tells us that this depleted ZAPE state would hinder the growth of synoptic-scale storm track eddies. It is therefore reasonable to hypothesize that an enhancement in warm pool convection during midwinter, which can correspond to an increase in the zonal gradient of tropical latent heating, could contribute to the occurrence of the midwinter minimum by exciting planetary-scale waves that leave behind an atmospheric state less favorable for storm track eddy growth. To test this hypothesis, we analyzed seasonal cycle of, and lead-lag relationships amongst, Pacific warm pool convection, planetary-scale EKE, synoptic-scale EKE, and ZAPE. The lead-lag relationships are then contrasted between the years when the North Pacific midwinter minimum is particularly pronounced, and those years when the North Pacific midwinter storm track is slightly weakened, or even strengthened.

3.3 Data and Methods

For the purpose of our study, we use daily European Centre for Medium Range Weather Forecasts’ ERA-Interim Reanalysis data (Dee et al. 2011). The study period is 1979–2017, and all variables analyzed are on a horizontal resolution of 2.5° × 2.5°. Equations used to calculate ZAPE and EKE are derived from Peixóto and Oort (1974; their equations 1 and 4) and modified in the same manner as was done in Baggett and Lee (2014); their Appendix A) to retain the spatial eddy component. The growth of synoptic-scale EKE is achieved through energy conversion from
ZAPE to EAPE, and then from EAPE to EKE (Lorenz 1955). As discussed in the introduction, we hypothesize that ZAPE is depleted by the growth of planetary-scale EKE, and as a result energy conversion from ZAPE to synoptic-scale EKE is suppressed.

In order to test the hypothesis, the EKE is separated into planetary-scale (zonal wavenumbers 1-3) and synoptic-scale (zonal wavenumbers 5-8) contributions by applying Fourier analysis on zonal wind ($u$) and meridional wind ($v$). The planetary- and synoptic-scale EKE and ZAPE are integrated over all longitudes, latitudinally from 20°N to 90°N, and vertically from the surface to 300 hPa. The tropical latitudes are excluded in order to evaluate the energetics in the extratropics (e.g., Bowley et al. 2018). For additional storm track diagnosis, in addition to the synoptic-scale EKE, we also employ a 24-hr difference filtered variance of meridional wind (e.g., Wallace et al. 1988; Chang and Fu 2002) multiplied by the cosine of latitude. This metric will be referred to as storm-track activity. As a proxy for warm pool convection, we use ERA-Interim outgoing longwave radiation (OLR) averaged over the domain of 15°–15°N, and 90°E–150°E. Our results remain unchanged if we use NOAA (National Oceanic and Atmospheric Administration) OLR dataset (Liebmann and Smith 1996).

As a measure of midwinter Pacific storm track intensity, we compute January-mean synoptic-scale EKE averaged over the domain of 20°N–60°N, and 160°E–160°W. This domain is chosen based on Figs. 1 and 2 in Afargan and Kaspi (2017) which show that suppression of vertically integrated synoptic-scale EKE occurs mostly over this domain. In our analysis, January 1979 is excluded because the previous December data does not exist in ERA-Interim. The leap days are also excluded in our analysis. The resulting midwinter minimum metric (based on January synoptic-scale EKE) is positively skewed (Fig. C1). This result indicates that strongly suppressed years are more typical than the years of little-or-no suppression.

This skewed distribution therefore indicates that the time-mean January condition is not representative, and that important insight can be gained by examining strongly suppressed years.
Given the limited length of the data set, to obtain a large enough sample size, we chose the top (bottom) 11 (13) to represent years when the midwinter storm track is strongest (weakest). As we will see, for the bottom years, the midwinter storm track strength is weaker than that prior to and after midwinter, and vice versa for the top years. We refer to the former years as suppression years and the latter as enhancement years. These criteria correspond approximately to a 0.5 standard deviation threshold (see Fig. C1). Multiple sensitivity tests have been performed: (1) increasing the threshold to 1.0 standard deviation, (2) increasing the upper level for the vertical integral to the 200-hPa level, (3) enlarging the domain to 150°E–150°W, and (4) using zonal wavenumbers 5-to-9 for the definition of synoptic-scale EKE. We found that the main results do not change qualitatively.

For testing statistical significance, we performed a two-sided Monte Carlo simulation with 5000 random subsets (e.g., Wilks 2011). Specifically, to test the statistical significance of the composites for the suppression and enhancement years, each of the 5000 sample composites was computed by randomly selecting 11 (13) years out of the 38 winters. The resulting 5000 random composites are then used to construct a null distribution and the $p$ values.

3.4 Results

3.4.1 Temporal relationships amongst extratropical ZAPE, planetary- and synoptic-scale EKE, and warm pool OLR

Figure 3.1 illustrates the October-to-April seasonal cycle of the NH extratropical ZAPE, planetary-scale EKE, synoptic-scale EKE, and the warm-pool OLR. Because ZAPE is an order of magnitude greater than EKE, to facilitate comparison, their time averages from October to April are removed. The climatology (Fig. 3.1a) shows a rapid increase in ZAPE up until the beginning
of December, followed by a mild increase during the months of December, January, and February (DJF). The synoptic-scale EKE also shows nearly constant values during the same time period. In contrast, planetary-scale EKE is maximized during January.

For the suppression years (Fig. 3.1b), the DJF ZAPE is more level and lower compared with the climatology. The suppression years are selected based on Pacific storm track strength (Section 3.3), yet the signature of the midwinter minimum can be seen in the entire NH extratropical synoptic-scale EKE (Fig. 3.1b). This result is consistent with previous studies that examined zonal-mean NH storm track (e.g., Shaw et al. 2018). Unlike the synoptic-scale EKE, the planetary-scale EKE shows a distinct maximum during the midwinter compared with the climatology (the difference between the planetary-scale EKE of the suppression years and that of the climatology will be shown below to be statistically significant). For the enhancement years (Fig. 3.1c), the ZAPE shows a larger increase throughout the DJF period. As mentioned earlier, during enhancement years, the synoptic-scale EKE also continues to increase during January. The planetary-scale EKE shows a local minimum during late December. These results support our hypothesis that strong planetary-scale wave growth contributes to the midwinter minimum.

The climatological inverted OLR (Fig. 3.1a) over the warm-pool shows a gradual increase from October to December, indicating that warm pool convection intensifies during these months. (The inverted OLR is shown because in the tropics lower [higher] value of OLR indicates more [less] convection. Therefore, in the inverted OLR, the peak [trough] corresponds to convection maximum [minimum].) Once again, consistent with our hypothesis, during the suppression years (Fig. 3.1b), warm pool convection is maximized during mid-December. For the enhancement years, midwinter warm pool convection is not as strong, and instead shows evidence of suppressed convection in mid-December.

The sequence of events discussed above, i.e., enhanced warm pool convection followed by the growth of planetary-scale EKE, the decline of ZAPE, and then finally suppressed synoptic-
scale EKE, can be better evaluated by examining the evolution of anomalies relative to the climatological seasonal cycle (Fig. 3.2). To examine the relationships between these variables, the ZAPE, both EKEs, and OLR are smoothed by Lanczos filter (Duchon 1979) with 10-day cutoff. For ease of comparison, the anomalies are also normalized. The time evolution of the anomalous synoptic-scale EKE shown in Fig. 3.2a captures the midwinter minimum presented in Fig. 3.1b. (In Fig. 3.2, ZAPE is multiplied by −1 to help facilitate comparison between suppressed ZAPE and enhanced planetary-scale EKE.)

In the suppression years (Fig. 3.2a), the anomalous ZAPE shows a minimum during early January, and this ZAPE minimum is preceded several days by anomalously large planetary-scale EKE. The positive planetary-scale EKE anomaly is in turn preceded 10-to-14 days earlier by enhanced warm-pool convection (Fig. 3.2a). This time lag is consistent with a timescale of convectively driven Rossby wave packet propagation (e.g., Hoskins and Karoly 1981). The opposite sequence of events is found during enhancement years, although the January ZAPE anomaly is not statistically significant at the 90% level (Fig. 3.2b). These lead-lag relationships – enhanced warm pool convection → growth of planetary-scale EKE → suppressed ZAPE → suppressed synoptic-scale EKE – are consistent with our hypothesis. Although not shown, for the suppression years, planetary-scale baroclinic energy conversion term is enhanced, but synoptic-scale baroclinic energy conversion term is suppressed. Also, the former precedes the latter. These results further support the proposed hypothesis. During late November in the suppression years, similar lead-lag relationships are found with opposite signs, although synoptic-scale EKE anomaly lasts very briefly. This short persistence could be due to the weaker tropical convection anomaly in early November, moist processes that are influential to storm track variability (e.g., Hoskins and Valdes 1990; Shaw et al. 2016), and/or other local processes (see Section 3.2 for details) that could be more important during shoulder months.
The enhanced warm pool convection does not necessarily indicate, however, that the zonal gradient in tropical convective heating, necessary for exciting waves, is also enhanced prior to the midwinter storm track minimum. To discern whether the zonal gradient in the tropical Pacific convection is enhanced (reduced) prior to the midwinter suppression (enhancement), the longitude-time plots of OLR anomalies are also examined. Figure C2 shows that during mid-December, in suppression (enhancement) years, significantly enhanced (suppressed) tropical convection is located over the western tropical Pacific, but not in the eastern tropical Pacific. Therefore, during the years of suppression (enhancement), there is an increased (reduced) zonal asymmetry in tropical convective heating, which would promote (deter) the excitation of Rossby waves (Hoskins and Karoly 1981; Sardeshmukh and Hoskins 1988; Yoo et al. 2012a) and subsequent planetary-scale eddy growth (e.g., Baggett and Lee 2015, 2017).

3.4.2 Storm track structure, and global and local baroclinity

The results thus far indicate that even when the midwinter suppression (enhancement) years are selected based on the Pacific storm track amplitude, the suppression (enhancement) can be seen in the entire NH extratropical synoptic-scale EKE. Therefore, it is of interest to examine the spatial structure of the storm track to evaluate whether the suppression of extratropical synoptic-scale EKE is dominated by the Pacific storm track, or the suppression is more widespread.

Figure 3.3a shows that the suppression of synoptic-scale eddy activity (24-hr difference filtered variance of meridional wind; see Section 3.3) is widespread, although the suppression is more pronounced over the two oceanic storm track regions than over the continents. A similar spatial structure is present in the \( k = 5-8 \) EKE field (Fig. C4i). This finding is in line with the earlier results that a midwinter minimum can be observed upstream of the Pacific storm track.
(e.g., Nakamura 1992; Penny et al. 2010; Ren et al. 2010) and in the Atlantic storm track (e.g., Afargan and Kaspi 2017). Figure 3.3b shows that during enhancement years, the storm track activity is elevated, and once again the anomalies occur at most longitudes in midlatitudes. As a measure of baroclinity, the Eady growth rate is computed using the expression shown in Hendricks et al. (2014). Because eddy activity is found to be more sensitive to changes in upper level baroclinity in idealized models (e.g., Yuval and Kaspi 2016), the mid-tropospheric Eady growth rate is shown. The contours in Fig. 3.3 shows composites of Eady growth rate anomaly at 500 hPa, averaged over a 7-day period starting from the day of extremum in raw planetary-scale EKE. As can be seen in Fig. 3.1, this 7-day period typically occurs from late December to early January. We see that positive (negative) Eady growth rate anomalies are collocated over the two oceanic storm track activity anomalies in enhancement (suppression) years.

The results described indicate that, during the suppression years, baroclinity is anomalously negative where the storm track anomalies are also negative (Fig. 3.3a), consistent with previous studies (e.g., Penny et al. 2013). This finding seems to be at odds with the well-known fact that baroclinity associated with the North Pacific jet is anomalously strong during the time of the midwinter minimum. From Figure C3a, for the suppression years, it can be seen that the Eady growth rate is indeed anomalously positive in the region where the Pacific jet is anomalously strong, but it is anomalously negative over most of the midlatitudes. (The area-weighted average of the Eady growth rate anomaly between 20°N and 70°N is \(-3.67 \times 10^{-3}\text{day}^{-1}\) \([1.75 \times 10^{-3}\text{ day}^{-1}]\) during suppression [enhancement] years, indicating that baroclinity is overall anomalously negative [positive].) The zonal wind anomalies shown in Fig. C3 are the contribution by the planetary-scale eddies only. Because these planetary-scale zonal wind anomalies capture the well known strengthening of the Pacific jet, we interpret the strengthening of the Pacific jet as being a local manifestation of global-scale planetary-scale wave growth. As a result, baroclinity increases locally (e.g. the subtropical North Pacific where the planetary-scale
through is strengthened), but because the planetary-scale waves tap the ZAPE (Baggett and Lee 2015), baroclinity decreases globally.

3.5 Discussion and Conclusions

In this study, we have shown that the midwinter minimum of the Pacific storm track can be understood from the perspective of hemispheric-scale energetics. Figure 3.4 presents a schematic diagram summarizing the physical picture: 1) Tropical convection over the warm-pool is enhanced during December, and the resultant increase in the zonal gradient of convective heating leads to the excitation of planetary-scale waves that propagate into the extratropics. 2) As the planetary-scale waves amplify, the zonal wind and baroclinity increase locally, including the equatorward flank of the Pacific subtropical jet. 3) However, at the expense of the planetary-scale wave growth, when averaged over the entire NH midlatitudes, the ZAPE is reduced, and so is the hemisphere-wide baroclinity. This reduction in ZAPE and baroclinity typically occurs from late December to early January. 4) Following this period of reduced hemispheric ZAPE and baroclinity, the January mean synoptic-scale storm track activity declines. The above time sequence is also supported by Figs. C3, C4 and C5.

One of the main findings of this study is that warm-pool convection contributes to the midwinter storm track minimum. During the midwinter suppression years, the warm-pool convection reaches its maximum strength during December. In this regard, it is interesting that the Niño-3 SST anomalies for El Niño and La Niña reach their peak values during December (Li and Lau 2012). This behavior implies that during December the zonal gradient in tropical convective heating would be particularly small (large) for the El Niño (La Niña) years. According to the mechanism that we present here, the midwinter storm track would be enhanced (suppressed) during El Niño (La Niña) years. To test this possibility, we examined the values of
the Oceanic Niño Index (ONI) during November-December-January, which uses a 3-month running mean of the ERSSTv5 sea surface temperature anomalies in the Niño 3.4 region (5°–5°N, 120°–170°W). Among the 13 suppression years, there are six La Niña years (ONI ≤ −0.5), and only three El Niño years (ONI ≥ 0.5). For the 11 enhancement years, there are five El Niño years, while only one year belongs to La Niña. This finding lends further support to the midwinter minimum mechanism presented here.

The hemispheric-scale energetics employed in this study provides a conceptual framework to understand the cause of the midwinter minimum that occurs not only over the North Pacific, but also over the entire midlatitudes. This perspective is in line with the finding by Shaw et al. (2018) that the growth of stationary waves during early winter might contribute to the hemispheric-wide midwinter minimum. The global perspective presented in this study is not the only cause, however, and rather complements the previously proposed mechanisms. In their dry idealized model calculations, Chang and Zurita-Gotor (2007) found that Pacific midwinter minimum is not evident even though stationary waves are reasonably well simulated. Also, it remains to be seen if the early winter tropical heating anomalies can have a lasting impact throughout midwinter. Chang and Guo (2007) found that tropical heating anomalies during strong jet Januaries potentially contribute to the stationary wave anomalies over the Pacific, but their heating anomaly structure is different from the early winter heating anomalies shown in this study. While a delayed extratropical response is expected because it takes about 10 days for the forced planetary-scale waves to reach extratropics (e.g., Hoskins and Karoly 1981), further investigation is necessary to fully address this question. As a future study, it would be worthwhile to build a more complete picture of the midwinter minimum by combining the two—local and global—perspectives.
Fig. 3.1: Seasonal cycle of ZAPE (black), EKE (planetary scale, red; synoptic scale, blue), and OLR (green) starting from the preceding 1 October to 30 April for (a) climatology, (b) suppression years and (c) enhancement years. For ZAPE and EKE, their time averages from October to April are removed for the ease of comparison. The y axis for OLR is reversed so that the peaks correspond to enhanced tropical convection. The dates 1 January and 31 January are denoted by black dashed lines.
Fig. 3.2: Normalized, anomalous ZAPE (black), EKE (planetary scale, red; synoptic scale, blue), and OLR (green) anomalies starting from the preceding 1 November to 31 March during (a) suppression years and (b) enhancement years. Lanczos filter with 10-day cutoff is applied to each term for smoothing. ZAPE and OLR anomalies are multiplied by $-1$ to aid comparisons. Thin lines are the deviation from the climatological seasonal cycle, and the thick lines (darker colored dots) indicate anomalies that are statistically significant at the 10% (5%) level.
Fig. 3.3: Composites of 300-hPa January storm track activity (shading) for (a) suppression and (b) enhancement years. Red (Blue) contours denote positive (negative) values of the 500-hPa Eady growth rate anomaly composited from (a) 1 January to 7 January and (b) preceding 26 December to 1 January, dates that correspond to a week starting from the day of extremum in raw planetary-scale EKE (Fig. 3.1). The contour interval is 0.02 day⁻¹, and growth rate anomalies larger than ±0.03 day⁻¹ are shown. To bring attention to midlatitudes, anomalies between 20°N–70°N are plotted.
Fig. 3.4: Schematic diagram of the key processes in the proposed mechanism for the midwinter minimum. We refer readers to section 3.5 for details. The proposed sequence of events is indicated by (1) – (4) in chronological order.
Chapter 4

Is the Stationary Wave Bias in CMIP5 Simulations Driven by Latent Heating Biases?\(^3\)

4.1 Abstract

Atmospheric stationary waves play an important role in regional climate. In phase 5 of the Coupled Model Intercomparison Project (CMIP5), a prior study found that there are systematic biases in Arctic moisture intrusions caused by stationary eddy meridional wind biases. In this study, using initial-value model calculations, it is shown that CMIP5 latent heating biases in the tropics and midlatitudes play a substantial role in generating the systematic meridional wind bias poleward of 50°N. It is further shown that the midlatitude heating biases are in part driven by the circulation caused by the tropical and subtropical heating biases. These results indicate that the systematic stationary meridional wind biases poleward of 50°N can be traced to systematic model biases in tropical and extratropical latent heating. Therefore, reliable regional climate projections likely hinge on accurate representations of moist processes upstream of the region of interest and in the tropics.

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4.2 Introduction

Climatological stationary waves influence regional climate variability through their impact on moisture transport (Baggett et al. 2016; Li et al. 2012; Simpson et al. 2016), storm track intensity and shift (Chang 2009; Park and Lee 2020; Shaw et al. 2016, 2018), and transient eddies (Cai et al. 2007; Goss et al. 2016; Park and Lee 2019). Therefore, an accurate representation of stationary waves is essential for climate models to predict future climate change (Wills et al. 2019). Using model ensembles from the Coupled Model Intercomparison Project Phase 5 (CMIP5; Taylor et al. 2012), recent studies have shown that circulation change plays a critical role in driving regional precipitation and temperature changes under global warming scenarios (Garfinkel et al. 2020; Shepherd 2014; Simpson et al. 2014, 2016; Tamarin-Brodsky et al. 2020; Vallis et al. 2015; Xie et al. 2015; Zappa and Shepherd 2017).

The climatological stationary waves play an important role in warming the Arctic by advecting warm, moist air from lower latitudes to the Arctic (Baggett and Lee 2017; Baggett et al. 2016; Gong et al. 2017; Park et al. 2015; Woods and Caballero 2016). Woods et al. (2017) found that CMIP5 models systematically underestimate (overestimate) moisture intrusions into the Atlantic (Pacific) sector of the Arctic Ocean due to weak (strong) southerly wind biases, and Lee et al. (2019) showed that this bias is mostly due to a systematic misrepresentation of the stationary meridional wind in the CMIP5 models. Because warm, moist air carried by a southerly wind warms the Arctic, the underestimated moisture intrusions through the Greenland and the Barents Seas lead to a cold surface temperature bias over the Arctic Ocean, while overestimated moisture intrusions in the Pacific sector contribute to a warm bias over the Russian Far East.

One of the possible causes of this CMIP5 stationary wave bias is latent heating bias. According to stationary wave theories, diabatic heating is the main driver of boreal wintertime stationary waves (e.g., Chang 2009; Held et al. 2002). Because tropical heating excites poleward
propagating planetary-scale waves (Hoskins and Karoly 1981; Sardeshmukh and Hoskins 1988; Ting 1996), tropical heating biases can result in a misrepresentation of stationary waves. Indeed, CMIP5 models are known to have systematic biases in the tropical sea surface temperature (SST), tropical precipitation, and static stability (Adam et al. 2016, 2018; Li et al. 2015, 2016; Luo et al. 2018; Seager et al. 2019; Sohn et al. 2016; Zhou and Xie 2015). Lee et al. (2019) found that meridional wind biases across 60°N are correlated with tropical temperature biases. Park and Lee (2019) showed that extratropical latent heating also plays an important role in amplifying the stationary waves, but that extratropical heating is in part driven by the circulation that was excited by tropical heating.

In this study, we test a hypothesis that tropical and midlatitude latent heating biases make substantial contributions to high latitude stationary meridional wind bias in CMIP5 models. Orographic drag is also a possible source of circulation bias (Pithan et al. 2016), but not considered in this study.

4.3 Data and Method

4.3.1 CMIP5 and Reanalysis Data

We use daily-mean data from the historical simulations of 28 CMIP5 models (Table D.1) from 1979-2005 December to February (DJF). Only the first ensemble member (r1i1p1) is selected from each CMIP5 model, except for CCSM4 where only r6i1p1 is available. For consistent horizontal resolution, CMIP5 data are linearly interpolated to a 2.5° × 2.5° grid. CMIP5 data are available on eight pressure levels (1000, 850, 700, 500, 250, 100, 50, 10 hPa), and 250-hPa meridional wind is used for the analysis of stationary waves. As a reference dataset to evaluate the performance of CMIP5 models, following Woods et al. (2017), the European
Centre for Medium Range Weather Forecasts interim reanalysis data (ERA-Interim; Dee et al. 2011) is used. We use daily-mean data with a horizontal resolution of $2.5^\circ \times 2.5^\circ$. The stationary eddy meridional wind is obtained by first averaging over the December-January-February period, and then by subtracting the zonal mean values. Hereafter, the stationary eddy meridional wind is denoted by $\vec{v}^*$, where the overbar indicates a time average and the asterisk indicates the deviation from the zonal mean.

4.3.2 Initial value calculations

To test the hypothesized relationship between the model latent heating bias and stationary wave bias, idealized model experiments are performed by using the dry spectral dynamical core of the NOAA Geophysical Fluid Dynamics Laboratory (GFDL). Because the model setup is identical to that in Baggett et al. (2016), only a brief summary is provided here. Our model has a horizontal resolution of triangular 42 and a vertical resolution of 28 sigma levels with the topography. The background state of the model is initialized by the 1979-2005 DJF climatological fields in ERA-Interim. In the absence of a perturbation, the initial state is maintained through a forcing term (Franzke et al. 2004), which is acquired by integrating the model forward by one time step after the initialization (see Appendix D.1 for details). The perturbations in this study are latent heating biases. Because daily latent heating data is not available in the CMIP5 archive, we estimated the heating using the climatological precipitation biases of CMIP5 models (e.g., Fig. 4.2c) by assuming that all condensed water precipitates out and that the resulting latent heating takes on an idealized vertical profile that has a maximum at 500 hPa. We refer the reader to Yoo et al. (2012a) for details. The model solution reaches a quasi-steady state by model day 11. The circulation response to the heating is defined as the difference between the model solution and the climatological state. While the results of this study are
qualitatively insensitive to the choice of maximum heating altitude (not shown), the Gaussian profile may not be appropriate for representing heating profile at some locations. Therefore, our model calculations need to be revisited if and when three-dimensional heating output becomes available.

4.4 Results

4.4.1 Systematic biases of stationary eddy meridional wind and precipitation in CMIP5 models

Figures 4.1a and 4.1b show $\bar{v}^*$ in CMIP5 models and ERA-Interim reanalysis, respectively. In both the multimodel mean and reanalysis, there are two centers of southerly winds, one over the North Pacific and the other over the North Atlantic. Although the CMIP5 models capture the overall spatial structure of the stationary waves, there are substantial regional $\bar{v}^*$ biases poleward of 50°N (Fig. 4.1c); there are negative biases over the East Siberian Sea and the Greenland Sea, while positive biases are prevalent over North America and northeastern Europe. Because the negative (positive) biases coincide with the climatological southerly (northerly) wind, this bias field indicates that the CMIP5 models systematically underestimate the strength of $\bar{v}^*$. To test if the bias pattern is sensitive to the choice of reanalysis dataset, the same analysis is repeated using the daily-mean Japanese 55-year Reanalysis (JRA-55; Kobayashi et al. 2015) data. The result (Fig. D1) agrees with that shown in Fig. 4.1c, indicating that the bias field is insensitive to the choice of the reanalysis dataset.

The climatological precipitation in CMIP5 models (Fig. 4.2a) captures the main features seen in ERA-Interim (Fig. 4.2b). However, their difference, shown in Fig. 4.2c, reveals substantial regional precipitation biases. In the tropics, for example, the models overestimate precipitation over the Indian Ocean and Maritime Continent, and underestimate precipitation over
the western and central equatorial Pacific. In midlatitudes, the models tend to overestimate precipitation over most of the continents and coastal regions.

4.4.2 Which regional precipitation biases cause the stationary meridional wind biases?

Given that the CMIP5 models have systematic tropical and extratropical precipitation biases (Fig. 4.2c), we next test the hypothesized causal relationship between the precipitation and the stationary wave biases. Figure 4.1d shows the GFDL model day 10-13 response to global CMIP5 latent heating biases. Although not perfect, the GFDL model $\bar{v}^*$ response captures the negative (positive) biases over the East Siberian Sea and Greenland Sea (North America and Europe). Within one week of the model integration, the GFDL model $\bar{v}^*$ response resembles the CMIP5 stationary wave biases, which is consistent with Ma et al. (2014) who showed that some features in the mean precipitation and circulation bias fields emerge by day 2 in their hindcast experiments. In order to test if the latent heating bias pattern is sensitive to the choice of precipitation data, the same analysis is performed using the daily Global Precipitation Climatology Project (GPCP) data (Huffman et al. 2001). The GPCP climatological precipitation is calculated over the period of 1997–2005 DJF due to data availability. The result, shown in Fig. D2, reveals that the precipitation bias and $\bar{v}^*$ response are qualitatively consistent with the ERA-Interim-based result (see Figs. 4.2d and 4.1c). Notable differences are that the $\bar{v}^*$ response to the CMIP5–minus–GPCP precipitation field is generally stronger, and that over the western and central North Pacific, the $\bar{v}^*$ response agrees less with the $\bar{v}^*$ bias. Therefore, in the rest of the analysis, closer attention will be given to the Greenland-Scandinavia sector (the black box in Fig. 4.1c).

Because regional precipitation biases are widespread (Fig. 4.2c), to identify the most important heating bias features, we next investigate the model response to a forcing that has the
form of the heating bias over different geographical locations. The approach we adopt is similar to SST patch experiments (Branstator 1985; Barsugli et al. 2006), except that instead of SST we use the heating bias fields. Each of the heating patches spans $30^\circ \times 30^\circ$ grid boxes, with an increment of $10^\circ$ in longitude and latitude; the patch domains overlap with neighboring patches. Inside the patch, the forcing is set equal to the heating bias, and the forcing outside of the patch in question is set to zero. There are 360 patches covering $30^\circ$S–$90^\circ$N and all longitudes. The extent to which the individual $\vec{v}^*$ solution to each patch forcing resembles the bias $\vec{v}^*$ field is quantified by projecting the former onto the latter over the two Arctic domains indicated in Fig. 4.1c. The resulting projection values are summarized in a map of ($10^\circ \times 10^\circ$) resolution (Figs. 4.3a and 4.3f). In Fig. 4.3a, for example, the value assigned to the grid point ($75^\circ$W, $15^\circ$N) represents the projection of the $\vec{v}^*$ response to a heating patch over $0^\circ$–$30^\circ$N and $90^\circ$W–$60^\circ$W (coinciding with the Tropical Atlantic domain) onto the $\vec{v}^*$ bias field over the Greenland-Scandinavia domain (the black box in Fig. 4.1c). Similarly, Figure 4.3f shows projection values of $\vec{v}^*$ fields over the East Siberian-Beaufort Sea domain.

Figure 4.3a shows that there are three centers of heating bias that are effective at generating the $\vec{v}^*$ bias over the Greenland-Scandinavia domain which is highlighted in Fig. 4.3e. They are the eastern North Pacific (NP), tropical Atlantic (TA), and western North Atlantic (NA). Figure 4.3b shows that the NP heating bias generates a negative $\vec{v}^*$ anomaly over Baffin Bay and positive $\vec{v}^*$ anomaly over northern Europe; a similar dipole structure is generated by the TA heating bias, but with a larger magnitude (Fig. 4.3c); in response to the NA heating bias, Fig. 4.3d shows that negative and positive $\vec{v}^*$ anomalies develop over Greenland and the Barents Sea, respectively. Outside of the NP, TA, and NA domains, Figure 4.3a shows weak positive or negative values, indicating that heating biases outside of those three domains do not contribute to the $\vec{v}^*$ bias over the Greenland-Scandinavia domain. For the East Siberian-Beaufort Sea (Fig. 4.3f), two local maxima of $\vec{v}^*$ projection values are identified: the central tropical Pacific (CTP)
and western subtropical Pacific (WSP). While both CTP and WSP precipitation biases generate a negative $\bar{v}^*$ anomaly over the East Siberian Sea, CTP forcing drives a positive $\bar{v}^*$ anomaly centered over northern Canada, and WSP forcing drives a positive $\bar{v}^*$ anomaly centered over western North America.

4.4.3 Are the regional precipitation biases independent of each other?

The model results described above indicate that the NP, NA, TA, WSP, and CTP biases play a pivotal role in generating the stationary wave biases. These heating biases are generally removed from the stationary wave biases in question, with tropical heating biases generating relatively large $\bar{v}^*$ responses. The findings by Park and Lee (2019) suggest that the NP and NA biases may not be independent from tropical biases. In the context of stationary wave interference, they showed that anomalously strong tropical heating generates an anomalously strong circulation which, by transporting moisture, causes latent heating to increase over the midlatitude storm track regions. Therefore, it is plausible that some of the key extratropical heating biases may be caused by tropical heating biases. Because large-scale advection plays an important role in determining the distribution of tropospheric water vapor (Pierrehumbert et al. 2007), one approach to test this hypothesis is to mimic moisture using a passive tracer (see Appendix D.2 for details). Therefore, we next address this question by examining the vertically integrated passive tracer ($q_{tr}$) response to individual heating patches.

Figures 4.4a and 4.4e show projection values computed by projecting the model $q_{tr}$ response to the same heating patches used for Fig. 4.3, but onto the precipitation biases over the NA and NP domain (red boxes in Figs. 4.4d and 4.4h), respectively. These two domains are considered because, as Fig. 4.3a shows, precipitation/heating biases over those regions are
effective at generating the $\bar{v}^*$ biases over the Greenland-Scandinavia sector (Fig. 4.1c). The initial value of the $q_{tr}$ field is the climatological specific humidity.

Figure 4.4a indicates that the negative NA precipitation bias is caused by precipitation/heating biases over the NP and TA domains. Interpretation of Fig. 4.4a can be aided visually by examining Figs. 4.4b and 4.4c respectively, which show that the NP and TA precipitation/heating biases drive negative $q_{tr}$ anomalies that resemble the CMIP5 precipitation bias over the NA domain (Fig. 4.4e). The positive $q_{tr}$ anomalies with northeast-southwest tilt are also consistent with overestimated precipitation spanning from the east coast of North America to Europe (Fig. 4.4d).

4.5 Discussion and Conclusions

In this study, we tested the hypothesis that precipitation (latent heating) biases in CMIP5 models play an important role in causing stationary wave biases in subarctic and Arctic regions during boreal winter. Our results reveal that systematic stationary meridional wind biases in the CMIP5 models over the subarctic and Arctic can be traced to tropical and extratropical latent heating biases. Tropical biases have both direct and indirect influences on the Greenland-Scandinavia $\bar{v}^*$ bias. As for the indirect effect, Fig. 4.4 shows that the NP bias can be generated by the EA/WTP biases, and the NA bias by the NP/TA biases. Therefore, the extratropical NP and NA precipitation biases are in part produced by tropical and upstream precipitation/heating biases through their impact on moisture transport. This relationship is summarized in Fig. 4.5: (1) Western Pacific precipitation bias induces the wind biases (blue and red arrows). In addition, it leads to enhanced moisture transport (green arrow) into the eastern North Pacific, amplifying the eastern North Pacific precipitation bias. (2) The resulting precipitation bias over the eastern North Pacific, and the dry bias over the western tropical Atlantic both generate a northerly flow bias.
over Greenland and a southerly flow bias over northern Europe. These precipitation biases contribute to the dry bias south of Greenland by reducing moisture transport (brown arrows), which further amplifies the stationary wave biases.

These findings highlight that in order to improve model performance in predicting regional climate change, it is important to accurately model the magnitude and location of latent heating, not only over the region of interest, but also farther upstream and even in the tropics. Possible biases in the height of the latent heat release may also be consequential, but this effect was not investigated in this study because the heating profile is unavailable in the CMIP5 archive. If the tropical convective heating field continues to be misrepresented, stationary wave biases are likely to remain and influence other model fields including Arctic surface temperatures (e.g., Woods et al. 2017; Lee et al. 2019). We show that the latent heating biases can generate stationary wave biases at high latitudes within 10 days. This fast emergence of model bias is consistent with the findings of previous studies (Xie et al. 2012; Ma et al. 2014). The findings of this study not only call for close attention to the causes and impacts of model biases, but also suggest caution in interpreting regional climate model projections.
Fig. 4.1: (a) CMIP5 multimodel mean of 250-hPa $\bar{v}^*$ over the period of 1979–2005 DJF. (b) As in (a) but for the ERA-Interim reanalysis. (c) CMIP5 multimodel mean biases in 250-hPa $\bar{v}^*$. Stippling indicates mean biases where at least 80% of the CMIP5 models have the same sign of bias. The red box denotes the domain of the East Siberian-Beaufort Sea (50°–90°N, 135°E–90°W), and the black box denotes the domain of Greenland-Scandinavia (50°–90°N, 70°W–50°E). (d) GFDL model 250-hPa $\bar{v}^*$ response averaged over model days 10–13 forced by CMIP5 latent heating biases converted from precipitation biases.
Fig. 4.2: (a) CMIP5 multimodel mean of precipitation over the period of 1979–2005 DJF. (b) As in (a) but for the ERA-Interim reanalysis. (c) CMIP5 multimodel mean biases in precipitation. Stippling indicates mean biases where at least 80% of the CMIP5 models have the same sign of bias. Boxes represent domains of precipitation biases used for the perturbation in Figure 4.3. See text for details.
Fig. 4.3: (a) Projections of the GFDL model $\bar{v}^*$ response averaged over model days 10–13 onto CMIP5 250-hPa $\bar{v}^*$ bias over Greenland-Scandinavia (black dashed box in Fig. 4.1c). Each projection value derived from the corresponding heating patch experiment is located at the center of the patch. See text for details. (b)–(d) $\bar{v}^*$ response to forcing over (b) eastern North Pacific (NP; 20°–50°N and 140°–110°W, blue box), (c) tropical Atlantic (TA; 0°–30°N and 90°–60°W, purple box), and (d) western North Atlantic (NA; 40°–70°N and 70°–40°W, cyan box). (e) CMIP5 250-hPa $\bar{v}^*$ bias centered at the North Atlantic Ocean. (f) As in (a) but for the bias over the East Siberian-Beaufort Sea (red dashed box in Fig. 4.1c). (g)–(h) Model $\bar{v}^*$ response to forcing over (g) western subtropical Pacific (WSP; 10°–40°N and 130°–160°E, brown box) and (h) central tropical Pacific (CTP; 10°S–20°N and 160°E–170°W, green box). (i) As in (e) but centered at the North Pacific Ocean.
Fig. 4.4: (a) Projections of the GFDL model $q_{tr}$ response averaged over model days 10–13 onto CMIP5 precipitation bias over the western North Atlantic (45°–75°N and 70°–30°W) denoted by the red box in (d). (b)–(c) Model $q_{tr}$ response to forcing over (b) eastern North Pacific (NP; blue box) and (c) tropical Atlantic (TA; purple box). (d) CMIP5 precipitation bias centered at the Atlantic Ocean. (e) As in (a) but for bias over the eastern North Pacific (30°–60°N and 150°–110°W) denoted by the red box in (h). (f)–(g) Model $q_{tr}$ response to forcing over (f) the East Asia (EA; 20°–50°N and 120°–150°E, yellow box) and (g) western tropical Pacific (WTP; 10°S–20°N and 110°–140°E, green box). (h) As in (e) but centered at the Pacific Ocean.
Fig. 4.5: Schematic diagram of relationships between CMIP5 model precipitation biases and stationary eddy meridional wind biases. See text for details.
Chapter 5

The Role of Planetary-scale Eddies on the Recent Isentropic Slope Trend during Boreal Winter

5.1 Abstract

According to baroclinic adjustment theory, the isentropic slope maintains its marginal state for baroclinic instability. However, the recent trend of Arctic warming raises the possibility that there could have been a systematic change in the extratropical isentropic slope. In this study, global reanalysis data is used to investigate this possibility. The result shows that tropospheric isentropes north of 50°N have been flattening significantly for the recent 25-yr winters. This trend pattern fluctuates at intraseasonal time scales. An examination of the temporal evolution indicates that it is the planetary-scale (zonal wavenumbers 1–3) eddy heat fluxes, not the synoptic-scale eddy heat fluxes, that flatten the isentropes; synoptic-scale eddy heat fluxes instead respond to the subsequent changes in isentropic slope. This extratropical planetary scale wave growth is preceded by an enhanced zonal asymmetry of tropical heating and poleward wave activity vectors.

A numerical model is used to test if the observed latent heating can generate the observed isentropic slope anomalies. The result shows that the tropical heating indeed contributes to the isentropic slope trend. The agreement between the model solution and the observation improves substantially if extratropical latent heating is also included in the forcing. The model temperature response shows a pattern resembling the warm-Arctic–cold-continent pattern. From these results,

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it is concluded that the recent flattening trend of isentropic slope north of 50°N is mostly caused by planetary scale eddy activities generated from latent heating, and that this change is accompanied by a warm-Arctic–cold-continent pattern that permeates the entire troposphere.

5.2 Introduction

In baroclinic adjustment theory, the meridional temperature gradient of the mean state is that of the marginal state with respect to baroclinic instability (Stone 1978). The primary region of baroclinic growth of synoptic-scale eddies coincides with the region of the strongest horizontal temperature gradient (Charney 1947; Eady 1949). If the meridional temperature gradient were to exceed that of the marginal state, the resultant eddy heat transport would weaken the temperature gradient, stabilizing the background state (Simmons and Hoskins 1978; Thorncorft et al. 1993). This process of stabilization by the eddies, according to the theory, leads to the mean state of the atmosphere being a neutral state. The meridional slope of isentropic surfaces, or the isentropic slope, has been considered as supporting evidence of the theory. Stone (1978) showed that the observed time-mean isentropic slope corresponds to a baroclinically neutral state of a two-layer quasi-geostrophic model. Using a dry quasi-geostrophic β-plane channel model with a higher vertical resolution, Solomon and Stone (2001) examined the sensitivity of the equilibrated state of the midlatitude troposphere to various changes in radiative forcing. From their idealized model experiment, it is found that the vertical profile of the equilibrated potential vorticity gradient is insensitive to a wide range of radiative forcing. However, their model lacks tropical convective heating which serves as a source of wave activity in the atmosphere. Stone and Nemet (1996) examined the isentropic slope for each season, and concluded that baroclinic adjustment occurs primarily between 400 to 800 hPa in midlatitudes in all seasons. Modeling studies with a dry
dynamical core found that the extratropical isentropic slope remains nearly constant with respect to variations in imposed radiative-equilibrium baroclinity (Schneider 2004; Zurita-Gotor 2008).

In a more recent observational study, Thompson and Birner (2012; Hereafter, TB12) investigated the relationship between the isentropic slope and eddy fluxes in the lower troposphere, separating eddies into high frequency (less than 10 days) and low frequency (longer than 10 days) components. Their result shows that the high frequency eddy heat flux peaks a few days after extratropical isentropes are anomalously steepened, but they deviate from the baroclinic adjustment theory, as appreciable flattening of isentropes does not occur following the heat flux maximum. Somewhat opposite behavior was found for the low frequency eddies; the low-frequency eddy heat flux maximum is followed by a sizable flattening of the isentropes, but prior to the heat flux maximum, there is a disproportionally weak steepening. Their finding suggests that low frequency eddy fluxes play a more important role in regulating the extratropical isentropic slope than previously recognized.

What drives those low-frequency eddies that are only weakly dependent on the extratropical baroclinity, yet regulate the isentropic slope? One such source of extratropical low-frequency eddies is tropical convective heating. Localized tropical heating is an important driver of low frequency, planetary-scale eddies that can propagate poleward into the extratropics (Hoskins and Karoly 1981; Sardeshmukh and Hoskins 1988). Recently, a number of studies investigated the impact of localized tropical heating on generating low frequency, planetary scale eddies that drive temperature changes over the Arctic (Lee 2012, 2014; Ding et al. 2014; Baggett and Lee 2015; Flournoy et al. 2016; Goss et al. 2016; Hu et al. 2016; Johnson and Kosaka 2016; Baggett and Lee 2017; Park and Lee 2019). With the exceptions of Ding et al. (2014) and Hu et al. (2016), these studies have shown that La Niña-like heating conditions preferentially excite planetary scale eddies that advect heat and moisture poleward. While the La Niña-like convective heating can excite planetary-scale waves that affect zonal-mean temperature over the extratropics,
the direct effect of this zonally asymmetric heating on the isentropic slope is negligible, because it does not increase the zonal-mean temperature in the tropics (e.g., Fig. 12 in Baggett and Lee 2017). Therefore, the localized heating can excite the extratropical planetary-scale waves without steepening the zonal-mean isentropic slopes within the tropics. Moreover, Park and Lee (2019) found that the circulation driven by tropical heating plays a role in generating localized extratropical latent heating which further amplifies the planetary scale eddies. Therefore, both tropical and extratropical latent heating could play an important role in regulating the isentropic slope by exciting the low-frequency, planetary-scale eddies. This indirect effect of latent heating on isentropic slopes is different from the direct effect by zonally symmetric heating, for example, considered by Butler et al. (2011).

Given that there has been a La Niña-like trend over the tropical Pacific (Clark and Lee 2019; Seager et al. 2019) and warming of the Arctic, in this study, we investigate (1) whether the extratropical isentropic slopes have been declining over the past few decades and (2) whether an enhancement of planetary scale eddy activity has been a key driver of the recent extratropical isentropic slope variability.

We provide a brief description of the data, analysis tools, and the idealized model configuration in Section 5.3. Next, we discuss our results from the analyses of the extratropical isentropic slope variability from observations in Section 5.4. The mechanism identified in Section 5.4 was tested using initial-value calculations, and the result is shown in Section 5.5. The conclusions are presented in Section 5.6.
5.3 Data and Methods

5.3.1 Data

We use daily-mean data of zonal wind, meridional wind, temperature, specific humidity, potential vorticity, outgoing longwave radiation, and surface pressure acquired from the European Centre for Medium-Range Weather Forecasts (ECMWF) Interim reanalysis (ERA-Interim; Dee et al., 2011) over the time period ranging from 1979 to 2017, December–February (DJF). All variables have a 2.5°×2.5° horizontal resolution, and those with vertical dimension have 23 pressure levels.5 We use diabatic heating data provided by the Japanese 55-year reanalysis (JRA-55; Kobayashi et al. 2015) because the individual diabatic heating variables are unavailable in the ERA-Interim. Diabatic heating data has a 2.5°×2.5° horizontal resolution and 37 pressure levels, and we define latent heating as a summation of convective heating and large-scale condensational heating. One may be concerned about the usage of the JRA-55 diabatic heating data in conjunction with the ERA-Interim data, because these two reanalysis datasets are produced by different data assimilation and parameterization schemes (e.g., Ling and Zhang 2013; Wright and Fueglistaler 2013). However, in a recent study examining the thermodynamic energy budget associated with the North Atlantic Oscillation, composites of the total non-radiative heating (summation of convective heating, large-scale heating, and vertical diffusional heating) show close agreement between the ERA-Interim and JRA-55 datasets (Clark and Feldstein 2020). Also, Gong et al. (2020) showed that the JRA-55 latent heating anomaly pattern associated with its recent inter-decadal trend is physically consistent with the wave activity flux calculated by ERA-Interim data over the Arctic Ocean. To further ensure the robustness of our results to the choice of

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5 23 pressure levels used in this study are listed as follows: 1000, 925, 850, 775, 700, 600, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, 10, 7, 5, 3, 2, 1 hPa.
reanalysis dataset, we calculated the linear trend of the extratropic isentropic slopes using both ERA-Interim and JRA-55. The pattern correlation between these two trends in the Northern Hemisphere troposphere domain (20°–90°N, 300–1000 hPa) is 0.78 and that in the projection domain (50°–80°N, 300–1000 hPa; see Section 5.3.3) is 0.89. In addition, we found that the correlation coefficient between the isentropic slope index (details in Section 5.3.3) computed using JRA-55 and that computed using ERA-Interim is 0.958. These analyses lend further confidence that the results of the study are qualitatively robust with respect to the choice of reanalysis dataset.

5.3.2 Isentropic slope and zonal-mean eddy heat fluxes

In this study, we use the isentropic slope defined as the meridional potential temperature gradient at constant pressure divided by the vertical potential temperature gradient at constant latitude.

\[ S_\theta = \frac{\left( \frac{\partial \theta}{\partial y} \right)_p}{\left( \frac{\partial \theta}{\partial p} \right)_y} \]  

(1)

where \( S_\theta \) represents the isentropic slope, \( \theta \) the potential temperature, \( p \) the pressure, and \( y \) the latitude. A qualitatively similar result is found in the trend analysis using isentropic slopes defined in isentropic coordinates.

For the calculation of the linear trend, we averaged the zonal-mean isentropic slopes for each DJF spanning from 1992/93 to 2016/17, and then computed the linear trend at every grid point by the Theil-Sen’s trend estimate method (Sen 1968), as shown in Fig. 5.1b. The statistical significance of the trend is tested using the Mann-Kendall test (Mann 1945). The recent 25 years are chosen because the Arctic warming has been accelerating since early 1990 (e.g., Gong et al. 2017).
The zonal-mean planetary-scale eddy heat flux (PF) and synoptic-scale eddy heat flux (SF) are calculated as $[v^i \theta^i]$ and $[v^s \theta^s]$ respectively, where brackets represent the zonal mean, asterisks represent the deviation from the zonal mean, $v$ represents meridional wind, and $\theta$ represents potential temperature. Subscript $l$ ($s$) represents planetary-scale (synoptic-scale) components of the variable, wherein zonal wavenumbers 1–3 (4–72) are retained for each variable via a Fourier transform. The definition of synoptic-scale in this study is unconventional, as it includes all scales shorter than zonal wavenumber 3. However, a prior study shows that the above definitions of planetary-scale and synoptic-scale depict clearly distinguishable energy life cycles and associated eddy fluxes (Baggett and Lee 2015).

The definition of eddy used in this study is different from that of TB12 whose definition is based on time scale. As will be shown in Section 5.4.2, our results turn out to be insensitive to whether or not the eddy is defined using space or time. Following TB12, the low-frequency and high-frequency time scale eddies are defined as $[v^M \theta^M]$ and $[v^N \theta^N]$, where subscript L and H denote time scale longer than 10 days and time scale less than 10 days, respectively. Eddy fluxes are calculated at 6-hourly time increments prior to daily averaging.

### 5.3.3 Isentropic slope index

To quantify how similar a given day’s $S_\theta$ anomaly field is to the recent 25-yr trend pattern, we construct an index using a projection method (e.g., Feldstein 2002; Gong et al. 2020):

$$ P_{IS}(t) = \frac{\sum_j \sum_k S_{\theta'}(\varphi_j, p_k, t) \tilde{S}_\theta(\varphi_j, p_k, d) \cos \varphi}{\sum_j \sum_k [\tilde{S}_\theta(\varphi_j, p_k, d)]^2 \cos \varphi}. $$

Here, $P_{IS}(t)$ is the projection value at day $t$, $S_{\theta'}$ is a daily anomaly, and $\tilde{S}_\theta$ denotes the 25-yr trend of $S_\theta$ (Fig. 5.1b). The daily anomaly $S_{\theta'}$ is defined as the deviation from its seasonal cycle which is constructed by retaining the first two harmonics of $S_\theta$. The latitude and the
pressure level are denoted by \( \varphi \) and \( p \) at the grid point \( j \) and \( k \), respectively, and \( d \) represents the day of the year that corresponds to \( t \). Figure 1b shows that \( S'_{\varphi} \) is negative north of 50°N, which is associated with temperature variability north of 60°N (TB12). To focus on the relationship between the declining trend of the extratropical isentropic slopes and the warming trend of the Arctic, the projection domain is chosen to cover 50°–80°N and from 1000 hPa to 300 hPa (the black dashed box in Fig. 5.1b). The main results of this study are found to be insensitive to the choice of the projection domain (e.g., 30°–70°N, 50°–70°N). The time series derived from (2) is normalized by its DJF standard deviation. The resulting daily time series is hereafter referred to as isentropic slope index (ISI).

Figure 5.2 shows the time series of the seasonal ISI values over the 1979/80–2016/17 period, wherein the seasonal ISI value is defined as the summation of all daily ISI values for which ISI > 1.0 in each DJF. In Fig. 5.2a, there is a sharp increasing trend of the seasonal ISI value with a \( p \) value of 0.012, indicating that the trend pattern has been occurring more frequently over time. In order to investigate whether the upward trend is caused by large-amplitude events, we binned all of the ISI > 1.0 days into three groups by ranking them based on their magnitude. Indeed, we see that the upward trend in Fig. 5.2a is essentially driven by the upward trend in the top bin of ISI > 1.0 days with a \( p \) value of 0.054 (Fig. 5.2b). The middle and bottom bins make very small contributions to the trend (Figs. 5.2c,d).

To examine intraseasonal fluctuations of \( S'_{\varphi} \), composite analysis is performed for ISI event days, which are defined below. To define a positive ISI event, we first identify the days when the ISI value is larger than one standard deviation. From these days, we next identify the days of local maxima within a 21-day time interval, and then define these days as lag 0 of the ISI event days. Thereby, the eleventh day within the 21-day window has the maximum ISI value, resulting in the isolation of event days by at least 10 days from each other. Negative event days are defined in the same manner but with the opposite sign. For the composite analysis, daily
anomalies of fluxes and meteorological variables are calculated by subtracting the annual cycle of each variable from its raw value.

Statistical significance of the composites (Figs. 5.4, 5.5, 5.7 and 5.8) is tested using two-sided Monte Carlo simulations (Wilks 2011). Specifically, for the composites of events with sample size N, we construct a composite consisting of N randomly chosen events that are also separated from each other by at least 10 days. This procedure is repeated 1000 times to build a null distribution for each lag day, and the p value of the observed composite is determined from the distribution.

5.3.4 Model experiment configuration

The dry spectral dynamic core model from the NOAA Geophysical Fluid Dynamics Laboratory is used to investigate the causal relationships associated with the linear trend of the isentropic slope. The model setup is identical to that used in Baggett et al. (2016), which has a horizontal resolution of triangular 42 and a vertical resolution of 28 sigma levels with realistic topography, fourth-order horizontal diffusion with a damping timescale of 0.1 days at its smallest scale and Held and Suarez parameterizations (Held and Suarez 1994) for Newtonian cooling and Rayleigh friction. The initial background state of the model is the 1979–2017 DJF climatology of zonal wind, meridional wind, temperature, and surface pressure. In the model, these zonally varying initial states are not balanced. Therefore, the three-dimensional climatological state does not remain as the solution of the model, which is written as

$$\frac{\partial \bar{X}_c}{\partial t} = \mathcal{L}(\bar{X}_c) + \mathcal{N}(\bar{X}_c, \bar{X}_c) \neq 0, \quad (3)$$

where $\bar{X}_c$ is the climatological background state, $\mathcal{L}$ is the model linear operator, and $\mathcal{N}$ represents model nonlinear processes. In order to prevent model fields from drifting from the initial states,
we follow the methodology used by Franzke et al. (2004)—adding a forcing term \( F \) which is equal to \(-[\mathcal{L}(\bar{X}_c) + \mathcal{N}(\bar{X}_c, \bar{X}_c)]\) to the model equations. This forcing term is obtained numerically by integrating the model equations one-time step forward. With this forcing term \( F \), the model equations can be written as

\[
\frac{\partial \bar{X}_c}{\partial t} = \mathcal{L}(\bar{X}_c) + \mathcal{N}(\bar{X}_c, \bar{X}_c) + F = 0. \tag{4}
\]

In doing so, we ensure that the model deviates from the initial condition only if additional forcing is included. We refer readers to Franzke et al. (2004) for further details on the forcing term \( F \). In this study, the additional forcing is the anomalous latent heating composites on lag day \(-20\) to \(0\) imposed on model days \(1\) to \(21\), respectively. The diabatic heating composites from the \(37\) pressure levels are interpolated into the model’s \(28\) sigma levels. The resultant model solution minus the initial state is defined as the model response.

### 5.4 Observational analysis

#### 5.4.1 Boreal wintertime climatology and the recent trend of isentropic slopes

We begin our analysis by examining the DJF climatology of the isentropic slopes and their recent 25-yr trend. Figure 5.1a shows the Northern Hemisphere (NH) DJF climatology of the zonal-mean isentropic slopes (shading) overlaid with the climatological zonal-mean zonal wind (green contour) and potential temperature (black contour). As expected, the largest isentropic slope values are found in midlatitudes. It is also seen that in the extratropics, as expected, the isentropic slope changes sharply at the dynamic tropopause (red contour) which is identified by the 2-potential vorticity unit (PVU; \(10^{-6}\) K kg\(^{-1}\) m\(^2\) s\(^{-1}\)) isoline.
In Fig. 5.1b, the linear trend of $S_\theta$ over the recent 25 years (1992/93–2016/17 DJF) is shown. Black contours denote the climatological $S_\theta$ depicted in Fig. 5.1a. As stated earlier, this time period is chosen because Arctic warming has been particularly rapid during this time period. Although not shown, we also calculated the linear trend for a longer period (1979/80–2016/17). The overall structure is comparable, but the recent 25-year trend is stronger. Figure 5.1b shows that the regions with significant flattening of the isentropes are 1) the extratropical troposphere poleward of 50°N and 2) the subtropical mid and lower troposphere equatorward of 30°N. Steepening isentropes are found in 1) the region poleward of 85°N, 2) the troposphere between 30°N and 50°N, and 3) the tropical upper troposphere centered at 250 hPa.

The fact that $S_\theta$ has undergone significant trends questions the baroclinic adjustment theory which proposes that the extratropical $S_\theta$ in the troposphere should remain constant. To quantify the deviation, we computed the linear trend averaged over the NH troposphere domain (20°–90°N, 300–1000 hPa), which turns out to be $-1.6 \times 10^{-3}$ hPa km$^{-1}$ per decade. This corresponds to approximately 1.7% of the DJF climatological value averaged over the same domain, indicating that over the 25-year period, the domain average $S_\theta$ has declined by 4.3%. We also calculated $S_\theta$ derived from the ERA-Interim products in isentropic coordinates. For this calculation, to test the sensitivity of the period chosen, we computed nineteen 20-yr segments of $S_\theta$, ranging from 1979/80–1998/99 DJF to 1997/98–2016/17 DJF. The result reveals that the slopes of individual isentropes throughout the troposphere (i.e., 265K to 330K) have systematically declined poleward of 50°N (not shown).

5.4.2 Relationship between eddy heat fluxes and isentropic slope during ISI event days

As discussed earlier, the interdecadal $S_\theta$ trend is mostly driven by the upward trend of large magnitude ISI days (Fig. 5.2). The decorrelation time scale of the ISI is 7 days, indicating
that intraseasonal time scale processes are linked to interdecadal variability. This perspective is supported by previous studies wherein changes in the frequency of occurrence of intraseasonal processes account for a substantial fraction of interdecadal variability (Johnson and Feldstein 2010; Feldstein and Lee 2014; Gong et al. 2017, 2020). Therefore, it is of our interest to examine a spatial map of intraseasonal eddy heat fluxes associated with the flattening trend of extratropical isentropic surfaces. We first compare the DJF climatological low-frequency (high-frequency) eddy heat flux with the climatological PF (SF), which are shown in Fig. 5.3. Compared with the high-frequency eddy heat flux (Fig. 5.3a), the SF (Fig. 5.3c) is present over a broader region in the lower troposphere and has a greater amplitude. This difference arises from intermediate scale waves (e.g., zonal wavenumber 4) that contribute to the zonal-mean heat transport in the lower troposphere (Solomon 1997).

Despite this difference, overall, the temporally filtered eddy heat fluxes (top row) and spatially filtered eddy heat fluxes (bottom row) show close resemblances, indicating that the result of TB12 can also be interpreted in terms of the difference between synoptic-scale and planetary-scale eddy fluxes. For the rest of the analyses in this study, it is more convenient to use the spatially filtered wave fluxes rather than the temporally filtered wave fluxes. Therefore, in the remaining analyses, the spatially filtered eddy fluxes will be examined. As discussed in the introduction, low-frequency eddy fluxes flatten the extratropical $S_\theta$ (TB12), and we have found a similar relationship using 700-hPa PF and SF as done in TB12 (not shown). This result is to be expected given the similarity between the two sets of eddy heat fluxes.

Next, we test the hypothesis that the extratropical $S_\theta$ anomalies associated with its interdecadal trend are driven by an enhancement of PF. Figure 5.4 depicts zonal-mean composites of anomalous $S_\theta$, PF, SF, and temperature during the ISI > 1.0 event days (n = 84). At lag day -6, $S'_\theta$ is essentially null (Fig. 5.4a), and so are the temperature anomalies (Fig. 5.4m), indicating that the anomalous baroclinity is negligible at the beginning of the $S'_\theta$ life cycle. Following the
positive PF anomalies (Fig. 5.4e), three days later, a positive-negative-positive $S'_\theta$ pattern (Fig. 5.4b) and warm-Arctic–cold-midlatitude temperature (Fig. 5.4n) emerge. At this lag, even stronger PF anomalies (Fig. 5.4f) are found. By lag 0, the $S'_\theta$ and temperature patterns have further intensified (Figs. 5.4c,o), the $S'_\theta$ pattern closely resembles the trend pattern (Fig. 5.1b), as required by construction, and the PF anomalies weaken (Fig. 5.4g). Three days later the PF anomalies become negative (Fig. 5.4h). Very different features are observed for SF anomalies, as negative values emerge at lag $-3$ (Fig. 5.4j) and intensify over the next three days (Fig. 5.4k). As the PF anomalies and SF anomalies weaken, the $S'_\theta$ pattern decays (Fig. 5.4d). At the same time, the warm-Arctic–cold-midlatitude pattern also diminishes (Fig. 5.4p). As expected from the thermal wind relationship, this temperature anomaly pattern is accompanied by negative zonal mean zonal wind anomalies between 50°–70°N and positive wind anomalies between 40°–50°N (green contours in Figs. 5.4n–p), indicating an equatorward shift of the midlatitude jet. This linkage between midlatitude jet shift and planetary-scale eddy fluxes supports the previous studies which showed that low-frequency eddies play a more dominant role than high-frequency eddies in the variability of the meridional displacement of the Southern Hemisphere midlatitude jet (e.g., Boljka et al. 2018; Lindgren et al. 2020). The opposite patterns of eddy fluxes and temperature anomalies (i.e., suppressed PF, enhanced SF, a poleward shift of the midlatitude jet, and Arctic cooling) are found during ISI $<-1.0$ events (not shown).

Although the domain of the ISI is confined to the north of 50°N, the composites also capture the positive $S'_\theta$ anomalies in 30°–50°N that resemble the $S'_\theta$ trend pattern (Fig. 5.1b). We suggest that this association represents the following relationships: enhanced PF which flattens isentropic slopes in 50°–80°N and cools the atmosphere to the south, centered at $\sim$50°N (Fig. 5.4n). This cooling at $\sim$50°N enhances baroclinicity and isentropic slopes to the south, between 30°–50°N. Using a new index designed to capture the steepening trend in 30°–50°N, we found that PF anomalies are enhanced 6 days prior (not shown). This PF anomaly warms the Arctic,
decreasing hemisphere-wide baroclinity, and subsequently weakens SF. This reduced SF activity is also likely to contribute to the $30^\circ$–$50^\circ$N steepening which subsequently enhances SF during positive lag days.

The results from Fig. 5.4 suggest that the trend-pattern-like $S'_\theta$ fields are caused by positive PF anomalies, and that these positive PF anomalies are not preconditioned upon anomalously strong baroclinity. This behavior is reminiscent of the findings by Yoo et al. (2012b) who find that in composites for all 8 phases of the Madden Julian Oscillation (MJO; Madden and Julian 1971), the anomalous baroclinity in midlatitudes responds to the eddy heat fluxes (e.g., their Fig. 8). This temporal relationship is opposite to what arises from baroclinic instability. These considerations also motivate the hypothesis stated in the introduction that the eddy fluxes that lead to the $S_\theta$ trend pattern might originate from anomalous tropical convection. This possibility is examined in the next subsection.

5.4.3 Relationship between tropical convection and isentropic slopes during ISI event days

Figure 5.5 shows lagged composites of the OLR anomalies meridionally averaged over the tropics ($30^\circ$–$30^\circ$N) as a function of longitude. Heightened (suppressed) tropical convection is denoted by negative (positive) OLR anomalies. For the ISI $> 1.0$ event days (Fig. 5.5a), we see enhanced tropical convection centered at $\sim120^\circ$E, near the Maritime Continent, starting from lag day $-23$. The convection center propagates eastward until day $-5$, reaching the western tropical Pacific. The same feature but of the opposite sign is found for the ISI $< -1.0$ event days (not shown), indicating a linear relationship between the extratropical $S_\theta$ variability and tropical convection.

Examining the top 1/3 bin of ISI events separately, we see that the convection anomalies over the western tropical Pacific and the Indian Ocean are stronger and better organized (Fig.
5.5b). The sign of the OLR anomalies over the Maritime Continent becomes opposite after lag 0, which corresponds to the timing of PF weakening and subsequent decay of the extratropical $S_\theta$ anomalies (Fig. 5.4). For the bottom 1/3 bin of the ISI events (Fig. 5.5c), there are signals of tropical convection anomalies, but they are more transient. At 130°E, for example, the OLR anomaly is positive between day –25 and –18, switches to negative until day –8, and then is close to zero by lag 0. During these event days, other sources of extratropical planetary-scale eddies, such as land-sea thermal contrasts in midlatitudes, orographic forcing, and upscale energy cascade from synoptic-scale eddies might play a role. The findings from Fig. 5.5 suggest that strong and persistent tropical convection anomalies over the Indo-western tropical Pacific lead to a greater PF and more forceful flattening of extratropical isentropes poleward of 50°N. An eastward propagation of the OLR anomalies raises the possibility that the MJO might play a role during the ISI event days. We tested this possibility by calculating the composites of the real-time multivariate MJO (RMM; Wheeler and Hendon 2004) indices. We found that for the top 1/3 bin of the ISI > 1.0 events, MJO is active, starting with phase 4 at lag day –25 and reaching phase 7 by lag day 0 (Fig. E1b). The RMM values are statistically significant throughout negative lag days. In the bottom bin, however, there is no evidence of active MJO. The result indicates that the MJO is an important contributor to the zonally asymmetric tropical heating that leads to the enhancement of planetary-scale eddy heat fluxes and the subsequent variations in extratropical isentropes. In fact, Yoo et al. (2011) showed that MJO phase 5 has been occurring more frequently while MJO phase 1 occurrence has been declining. Together with the fact that it is the top 1/3 days that contribute to the extratropical $S_\theta$ trend (Fig. 5.2), these results suggest that a source of the $S_\theta$ trend is the enhanced zonal asymmetry in the tropical Pacific sea surface temperature (Seager et al. 2019; Clark and Lee 2019) and the upward trend of western Pacific warm-pool deep convection (Jiang and Zhu 2020). The case study of 1984/85 DJF, when the
seasonal ISI index is large during the early years of the analyzed period (Figs. 5.2a,b), provide further support to the proposed mechanism (Figs. E2 and E3).

5.4.4 EP-flux diagnostics associated with isentropic slope variability

Given the evidence that planetary-scale wave fluxes excited in the tropics contribute to the extratropical $S_\theta$ trend, we employ Eliassen Palm (EP) flux diagnostics (Andrews and McIntyre 1976) to examine if there has been an increase in planetary-scale wave activity flux from the tropics during the same time period. We calculate the linear trend of the wave activity flux and its divergence in pressure coordinates (Edmon et al. 1980). Figure 5.6 shows that indeed there has been a significant increase in northward propagation of the total wave activity flux in the deep tropics at 200 hPa (Fig. 5.6a), and that this trend is mostly comprised of its planetary-scale component (Fig. 5.6b). This wave activity flux from the tropics contributes to the Eliassen-Palm (EP) flux convergence trend at the equatorward flank of the climatological subtropical jet (Fig. 5.1a) and helps explain the weakening of zonal-mean zonal wind reported in Maher et al. (2020). In the extratropics, we see a strengthening of upward wave activity flux (Fig. 5.6a). This trend is even stronger in the planetary-scale EP flux which shows significant EP flux convergence above the 2-PVU isoline (Fig. 5.6b).

To examine the evolution of the wave activity associated with ISI, we calculate composites of planetary-scale and synoptic-scale wave activity during ISI $> 1.0$ event days (Fig. 5.7). At lag $-6$, poleward propagation of planetary-scale wave activity is seen at the tropical and subtropical upper troposphere (Fig. 5.7a), and upward propagating wave activity centered at 60°N begins to develop, consistent with the simultaneous positive PF anomalies (Fig. 5.4e). In contrast, there are equatorward synoptic-scale EP flux anomalies in the tropical upper troposphere (Fig. 5.7e). For the lag $-3$ and 0 composites, the planetary-scale wave activity reaches farther
northward and propagates into the polar stratosphere, while the synoptic-scale wave activity anomalies are largely downward (Figs. 7b-c and 7e-f). The magnitude of the planetary-scale wave activity flux is larger than that of the synoptic-scale eddies, resulting in net upward propagation poleward of 50°N. At positive lags, the planetary-scale wave activity flux is downward, corresponding to the decay of the PF anomalies (Fig. 5.4h). These wave activity results are consistent with the findings in previous studies which show that the extratropical planetary-scale components of the EP-flux anomalies are often triggered by localized tropical convection (Yoo et al. 2012b; Baggett and Lee 2017). Together with the flattening $S_\theta$ trend (Fig. 5.1b) and the intensifying planetary-scale EP flux convergence trend (Fig. 5.6b), the quadrupole pattern in the zonal mean temperature composites at lag day 0 (Fig. 5.4o) and the associated anomalous planetary-scale EP flux convergence at 300 hPa and 65°N (Fig. 5.7c) support the idea that planetary-scale waves excited in the tropics can warm the high latitudes and cool the midlatitudes.

In order to further strengthen the argument that poleward propagating planetary-scale wave activity is induced by localized tropical convection, we computed the Rossby wave source (RWS) of Sardeshmukh and Hoskins (1988; their Eq. 4) at 200 hPa. We found that during ISI > 1.0 event days, there are positive planetary-scale RWS anomalies over southeast Asia in negative lags (not shown). Our results are consistent with previous findings that positive RWS anomalies in the subtropical jet exit region arise when a poleward propagating Rossby wave train is induced by enhanced convection over the western tropical Pacific (Lukens et al. 2017; Henderson and Maloney 2018).
5.5 Testing the impact of the heating on $S'_\theta$ using a numerical model

In this section, we test the hypothesis that anomalous latent heating causes the trend-like $S'_\theta$ pattern. The forcing of the model is the composite latent heating anomalies for the top bin of ISI > 1.0 event days. Figure 5.8a shows that two weeks prior to the maximum ISI days, there are significant latent heating anomalies in the tropics while extratropical latent heating anomalies are weak and scattered. However, over the following 10-day time period (Figs. 5.8b,c) significant heating anomalies develop over the central North Pacific and North Atlantic, which contribute to ridge development over Alaska and the Greenland Sea (not shown). This time evolution is reminiscent of the relay-circulation mechanism proposed by Park and Lee (2019); tropical latent heating induces circulation that transports moisture into the extratropics, where the moisture condenses, and the resulting latent heating in turn drives circulation farther downstream.

Figure 5.9 shows the model day-21 $S'_\theta$ response which corresponds to lag day 0 in the observations. The model response is defined as the zonal-mean $S'_\theta$ from the model minus the climatological $S'_\theta$. Negative (positive) values therefore indicate flattening (steepening). In the experiment forced with the global diabatic heating anomaly field (Fig. 5.9a), the $S'_\theta$ response can be characterized by 1) weak positive anomalies above 800 hPa in 30°–40°N and 2) negative anomalies throughout the extratropical troposphere in 50°–80°N. These anomalies show a close resemblance with the observed trend pattern shown in Fig. 5.1b. We next isolate the impact of the domain of diabatic forcing by examining the separate responses to tropical heating (30°S–30°N) and extratropical heating (30°–80°N). In response to the tropical heating, the model produces negative $S'_\theta$ anomalies mostly in the lower troposphere between 20°–60°N, with weaker negative anomalies in the upper troposphere centered at 50°N (Fig. 5.9b). In the experiment with extratropical heating only (Fig. 5.9c), the model response captures most of the features in Fig. 5.9a, indicating that the extratropical heating plays a critical role in driving the $S'_\theta$ anomalies.
When both tropical and extratropical heating (30°S–80°N) are inputted in the model calculation (Fig. 5.9d), the model $S_I$ response is essentially identical to the one shown in Fig. 5.9a. It is also found that the zonal-mean latent forcing in the lower extratropical troposphere is collocated with the isentropic slope anomalies, implying that there is also a direct contribution of diabatic heating to the extratropical isentropes. Examining the isentropic slope budget analysis over the North Atlantic for 2009 and 2010 winters, Papritz and Spengler (2015) found that latent heat release substantially contributes to the generation of the isentropic slope tendency over the North Atlantic. An analysis of the zonal-mean isentropic slope equation would be helpful in evaluating relative contributions by the direct and indirect effects of diabatic heating, but such an analysis is beyond the scope of this study.

In order to explore the possibility of whether the tropical heating used in this study can transport moisture poleward by circulation anomalies, we analyzed the model passive tracer field that represents the distribution of moisture advected by circulation. In this experiment, the initial passive tracer field corresponds to the zonal-mean climatological specific humidity field during boreal winter (e.g., Park and Lee 2019). We found that the model tracer response to tropical heating (Fig. E4a) shows positive anomalies over the central North Pacific and the Bering Sea (Baggett et al. 2016; Park and Lee 2019). For comparison, we also computed vertically integrated moisture flux convergence from the reanalysis data. We found that moisture flux convergence driven by planetary-scale eddies contributes to positive moisture anomalies poleward of 50°N in the North Pacific, particularly over the Bering Sea (Fig. E4b). This moisture anomaly feature is captured by the aforementioned model tracer response. Synoptic-scale eddy contribution to moisture flux convergence is confined to equatorward of 60°N and the eastern North Pacific (Fig. E4c). These results support the perspective that localized tropical heating plays a role in transporting anomalous moisture poleward and generating localized extratropical heating anomalies. However, there is also a limitation in the model results. Although the model $S_I$
response near 40°N reproduces positive isentropic slope anomalies, its magnitude is relatively weak compared to observed variations. This weak response may stem from several reasons. As discussed in Section 5.4.2, the $S_d'$ anomalies in 30°–50°N are strongly coupled to variability of synoptic-scale eddy fluxes, yet our model does not adequately represent synoptic-scale eddies.

In the observational analysis, the flattening of the extratropical $S_d'$ accompanies Arctic warming and midlatitude cooling (e.g., Fig. 5.4), especially in the lower troposphere. To examine if this anomalous temperature pattern can be driven by the observed diabatic forcing, we examine vertically averaged (700–1000 hPa) temperature anomalies. In response to the diabatic heating anomalies over the entire globe (Fig. 5.10a), the model lower-tropospheric temperature anomalies are positive in most regions poleward of 60°N, while negative anomalies are present over western Greenland, Eurasia, the Norwegian Sea, and North America. In Fig. 5.10b, we see that tropical heating contributes to warming over Alaska and the Kara Sea and cooling over central Asia and the Norwegian Sea. Figure 5.10c shows that extratropical heating also induces warming over the North Atlantic sector of the Arctic Ocean and cooling over Eurasia and western Canada. The model response to tropical and extratropical heating (Fig. 5.10d) can reproduce most of the temperature anomalies shown in Fig. 5.10a.

The model temperature responses are reminiscent of the warm Arctic–cold continent (WACC) pattern (Overland and Wang 2010; Overland et al. 2011; Inoue et al. 2012; Mori et al. 2014; Sorokina et al. 2016; Clark and Lee 2019). There is a large body of literature that proposes that the WACC pattern is driven by Arctic sea ice loss (e.g., Overland and Wang 2010; Overland et al. 2011; Inoue et al. 2012; Mori et al. 2014). However, Sorokina et al. (2016) find that the reduction of Barents Sea ice plays a minor role in driving cold anomalies over Siberia, because turbulent heat flux variability primarily depends on variability of the large-scale atmospheric circulation, rather than sea ice variability. Clark and Lee (2019) show that the enhanced east-west sea surface temperature gradient over the tropical Pacific Ocean contributes to the emergence of
the WACC pattern by driving a circulation pattern that advects warm air into the Arctic and cold air to the continents. Our model calculation supports this mechanism, as it indicates that atmospheric latent heating plays an important role in driving the atmospheric circulation and the WACC-like pattern.

5.6 Summary and Conclusions

The main findings of this study are: 1) Over the past few decades, the isentropic slope $S_\theta$ between 50°N and 80°N has been flattening, accompanying the occurrence of the warm Arctic cold continent (WACC) pattern during boreal wintertime. 2) On the intraseasonal time scale, the $S_\theta$ flattening is preceded by an anomalously strong planetary-scale heat flux (PF) and followed by anomalously weak synoptic-scale heat flux (SF). 3) The anomalously strong PF is preceded by anomalously strong tropical convection over the western tropical Pacific. 4) Over the past few decades, there has been an upward trend in the poleward propagation of EP flux in the tropical upper troposphere, and this flux trend is associated with planetary-scale waves. 5) Numerical model solutions forced by observed latent heating anomalies show an extratropical $S_\theta$ response that resembles the $S_\theta$ trend pattern. 6) The solution also shows the WACC-like pattern.

Result 2 indicates that planetary-scale eddies play a key role in regulating the extratropical isentropic slope. This finding is consistent with the earlier result by TB12, which showed that for isentropic variability, low-frequency eddies are more effective at flattening the isentropic slope than are the synoptic-scale eddies. Findings 1 and 2 further indicate that extratropical $S_\theta$ have been changing over the past few decades, and that it has been caused by the planetary-scale eddies. These findings are corroborated by prior studies which showed a pivotal role of planetary-scale eddies in driving poleward latent heat transport and subsequent Arctic warming during boreal winter, whereas synoptic-scale heat transport is rather reduced (Graversen
and Burtu 2016; Baggett and Lee 2017). From the diagnostics of the atmospheric energy transport, Graversen and Burtu (2016) further showed that zonal wavenumber 1–3 mostly accounts for latent energy transport associated with the surface air temperature variability across 70°N (e.g., their Fig. 6). Findings 3–5 provide evidence that these planetary-scale eddies are excited by anomalously strong convective heating over the western tropical Pacific. This picture is at odds with the baroclinic adjustment theory (Stone 1978), according to which there should be no secular trend in $S_{\theta}$. The reason for this disagreement is that the effect of the planetary-scale waves, especially those excited from zonally asymmetric tropical heating, is not considered in the theory. Because such planetary-scale waves arise independent of extratropical baroclinity (Baggett and Lee 2015), they form a viable source of eddy heat transport that can occur even when the background state corresponds to the marginal state. Model result 5 in this study suggests that latent heating plays a role in the $S_{\theta}$ flattening. However, we do not rule out contributions from other wave sources such as the orography of central Asia and North America (Held et al. 2002), upscale energy cascade from interactions with synoptic-scale eddies (Cai and Mak 1990), and the ice-albedo feedback from the Arctic sea ice loss (Overland and Wang 2010).

Results 1 and 6 indicate that the proposed mechanism accounts for both flattening of $S_{\theta}$ between 50°–80°N and warming of the Arctic throughout the troposphere. The same mechanism also steepens $S_{\theta}$ between 30°–50°N and cools the continents, again throughout the troposphere. Our findings further elucidate that the WACC-like pattern (e.g., the quadruple temperature field in Fig. 5.4o) can be induced by the atmospheric circulation forced by diabatic heating. Recent studies cast doubt on the argument that Arctic sea ice loss causes the WACC pattern. Blackport et al. (2019) found that strong cooling over continents and the associated positive sea level pressure anomaly pattern are found during winters when circulation drives the sea ice variability, whereas these patterns are absent during winters when sea ice drives the circulation variability. Another recent modeling study showed that warming induced by reduction in sea ice is largely confined to
surface, and cooling anomalies over Siberia cannot be captured by sea ice forcing alone (Labe et al. 2020). Thus, it will be worthwhile to pay attention to the recent variability of diabatic heating as a source of circulation anomaly associated with the WACC pattern.

The decline of extratropical $S_\theta$ may also provide insight into the variability of the Arctic stratosphere. For instance, TB12 examined the linkage between extratropical $S_\theta$ and the stratospheric variability by using the northern annular mode index that measures the strength of the stratospheric vortex. They found that flattened extratropical $S_\theta$ is associated with weakening of the polar stratospheric vortex indicated by the reduced northern annular mode index. Employing the zonal-mean temperature of four different reanalysis products, Cohen et al. (2020) show that the warming trend over the Arctic during boreal wintertime occurs throughout the column, with the strongest warming taking place near the surface and a secondary peak at the upper troposphere and stratosphere (e.g., their Fig. 1a). As seen in Figs. 5.4 and 5.7, the upward propagation of planetary-scale wave activity into the Arctic stratosphere and the subsequent stratospheric warming are preceded by 1) localized tropical convection over the tropical Pacific and then by 2) the extratropical isentropes being flattened over $50^\circ$–80$^\circ$N. Because there have been positive trends in both poleward propagating planetary-scale wave activity from tropics and upward propagating planetary-scale wave activity from extratropics (Fig. 5.6), the Arctic stratospheric temperature trend is likely to be associated with the same processes that drive the trend of the extratropical $S_\theta$. 
Fig. 5.1: (a) DJF Climatological isentropic slope (shading), isentropic surfaces (contours with an interval of 10 K), and zonal-mean zonal wind (green contours with an interval of 8 m s\(^{-1}\)). Black contours denote the climatological isentropic slope in (a) with an interval of 0.03 hPa km\(^{-1}\). Dotted areas indicate statistical significance at the 10% level. Statistical significance is evaluated by a Mann-Kendall test. Black dashed box represents the domain used to construct the isentropic slope index. In both panels, red contour indicates the 2-PVU isoline. The climatological isentropic slope at 500 hPa and 45\(^\circ\)N is approximately 0.13 hPa km\(^{-1}\), which corresponds to 130 hPa per 1000 km, or approximately 2.1 km per 1000 km (using the hypsometric equation with a scale height of 7 km).
Fig. 5.2: The linear trend of each DJF accumulated isentropic slope index for (a) the entire isentropic slope index $> 1.0$ days, (b) the top third of isentropic slope index $> 1.0$ days, (c) the middle third of isentropic slope index $> 1.0$ days, and (d) the bottom third of isentropic slope index $> 1.0$ days over the period spanning from 1979/80 to 2016/17 DJF. Black lines denote the linear regression line. Statistical significance is evaluated by a Mann-Kendall test. The domain used to construct the Isentropic Slope Index is 50°–80°N and 300–1000 hPa (the black dashed box in Fig. 5.1b). Positive values indicate that daily isentropic slope anomalies project positively onto the flattening trend of the extratropical isentropic slope in the isentropic slope index domain.
Fig. 5.3: DJF climatological eddy heat fluxes: (a) High frequency (less than 10 days) eddy heat flux, (b) Low frequency (greater than 10 days) eddy heat flux, (c) Synoptic-scale eddy heat flux that retains zonal wavenumbers 4 to 72, and (d) Planetary-scale eddy heat flux that retains zonal wavenumbers 1 to 3. The red line in each panel denotes the 2-PVU isoline, and green contours in each panel denote isentropic surfaces with an interval of 10 K.
Fig. 5.4: Zonal-mean composites of (a)–(d) anomalous isentropic slope, (e)–(h) planetary-scale eddy heat flux, (i)–(l) synoptic-scale eddy heat flux, and (m)–(p) temperature at (a),(e),(i),(m) lag day −6, (b),(f),(j),(n) lag day −3, (c),(g),(k),(o) lag day 0, and (d),(h),(l),(p) lag day 3 during ISI > 1.0 event days (n = 84). Green contours in temperature fields (m)–(p) denote composites of anomalous zonal-mean zonal wind with an interval of 1 m s$^{-1}$, and positive and negative values are denoted by dashed and solid contours, respectively. The red line in each panel denotes the 2-PVU isoline. A two-sided Monte Carlo simulation with 1000 random subsets is performed to evaluate the statistical significance, and the dotted areas indicate statistical significance at the 5% level.
Fig. 5.5: The longitude–lag plots of anomalous outgoing longwave radiation averaged over 30°S–30°N during (a) ISI > 1.0 events ($n=84$), (b) 1/3 top bin of ISI > 1.0 events ($n=28$), and (c) 1/3 bottom bin of ISI > 1.0 events ($n=28$). A Monte-Carlo simulation with 1000 random samples is performed for the statistical significance test, and the dotted areas indicate statistical significance at the 10% level. Prior to plotting, a nine-point local smoothing was applied twice to the composites.
Fig. 5.6: The linear trend of the E-P flux vectors and their divergence (shading) for the 1992/93-2016/17 DJF period for (a) all waves and (b) planetary-scale waves. In both panels, black solid (dashed) contours denote positive (negative) values of total climatological E-P flux divergence with a contour interval of 100 m$^2$ s$^{-2}$. The red line indicates the 2-PVU isoline. Prior to plotting, the vectors ($F_p$, $F_p$) are scaled by multiplying $(\alpha \pi)^{-1}$ and $(10^{-5}\cos \phi)$ to each component respectively, where $\alpha$ denotes the radius of the Earth. A Monte Carlo simulation with 1000 random samples is performed for the statistical significance test, and the dotted areas indicate statistical significance at the 10% level. Also, E-P flux vectors are plotted only if either one of the two vector components is statistically significant at the 10% level.
Fig. 5.7: Lag composites of anomalous (a)–(d) planetary-scale and (e)–(h) synoptic-scale E-P flux vectors and their divergence (shading) at (a),(c) lag day –6, (b),(f) lag day –3, (c),(g) lag day 0, and (d),(h) lag day 3 during ISI > 1.0 events (n = 84). The same scaling has been applied as in Fig. 5.6, and the red line indicates the 2-PVU isoline. A Monte Carlo simulation with 1000 random samples is performed for the statistical significance test, and the dotted areas indicate statistical significance at the 10% level. Also, EP flux vectors are plotted only if either vector component is statistically significant at the 10% level.
Fig. 5.8: Pentad composites of the 300–925 hPa vertically-averaged latent heating anomalies for the 1/3 top bin of ISI > 1.0 event days: (a) lag day –17 to –13, (b) lag day –12 to –8, and (c) lag day –7 to –3. A Monte Carlo simulation with 1000 random samples is performed for the statistical significance test, and the dotted areas indicate statistical significance at the 10% level. Prior to plotting, a nine-point local smoothing was applied twice to the composites for visualization.
Fig. 5.9: Composites of anomalous isentropic slope at model day 21 when the model is forced by (a) global latent heating anomalies, (b) tropical latent heating anomalies only, (c) extratropical latent heating anomalies only, and (d) tropical and extratropical latent heating anomalies. Black contours in each panel denote composites of latent forcing averaged over lag day $-7$ to $-3$, corresponding to Fig. 5.8c. Positive and negative values are denoted by dashed and solid contours with an interval of 0.05 K day$^{-1}$, respectively.
Fig. 5.10: Composites of the lower tropospheric (700–1000 hPa) temperature anomalies averaged over model days 19–21 when the model is forced by (a) global latent heating anomalies, (b) tropical latent heating anomalies only, (c) extratropical latent heating anomalies only, and (d) tropical and extratropical latent heating anomalies. Gray shading indicates the region where the surface pressure is below 700 hPa.
Chapter 6

Conclusions

Changes in atmospheric circulation play a critical role in driving regional climate variability, and their impact on regional climate extremes is expected to increase in future projections. In this dissertation, I propose a dynamic pathway of linking tropical and extratropical latent heating to extratropical circulation anomalies to understand both direct and indirect effects of tropical diabatic heating in extratropical circulation variability, which is termed the heating–circulation relay mechanism. In the following subsections, I briefly address the research questions posed in the introductory chapter. For each study, I also provide some remaining important research questions that I wish to address in future work and potential usefulness of the heating–circulation relay mechanism in investigating those questions.

6.1 Conclusion to Chapter 2: Relationship between Tropical and Extratropical Diabatic Heating and their Impact on Stationary-transient Wave Interference

1. During stationary-transient wave interference, to what extent are constructive interference and Arctic temperature anomalies driven by latent heating in the tropics, and by latent heating in the extratropics?

I find that a spatial pattern of wave interference and the associated temperature anomalies depends on the structure of tropical and extratropical latent heating. From observational analyses, different flavors of stationary wave interference are shown. The results show that constructive stationary wave interference tends to be preceded by enhanced tropical Pacific
warm-pool convection and then followed by enhanced heating over the North Pacific and
North Atlantic, leading to warming over the Arctic Ocean. If zonal asymmetries in both
tropical and extratropical heating become enhanced simultaneously, the magnitude of
constructive interference is significantly increased, and thereby poleward moisture flux is
the strongest among examined SWI events. If constructive interference takes place when
zonal asymmetries in both tropical and extratropical diabatic heating are weak, wave
interference is mostly limited to midlatitudes, subsequently with attenuated moisture flux
and substantially weaker Arctic warming. If zonal asymmetry is enhanced in either tropical
or extratropical heating only, maintenance of constructive interference is relatively short
and results in mild Arctic warming.

2. During stationary-transient wave interference, how might tropical latent heating and
extratropical latent heating be related to each other? Does the circulation driven by the
tropical heating organize extratropical latent heating?

I find that tropical and extratropical latent heating are related to each other during stationary
wave interference events. From the lead-lag relationships amongst the tropical and the two
extratropical heating indices, it is seen that although the latent heating anomalies in these
three domains can occur by themselves, they tend to occur together within 7-10 days of each
other, with the tropical heating anomaly leading the North Pacific heating anomaly which
in turn is followed by the North Atlantic anomaly. Next, results from idealized model
experiments show that the tropical heating (the North Pacific heating) is an important driver
of wave interference over the North Pacific (North Atlantic) Ocean, as well as an important
trigger of extratropical circulation that organizes extratropical latent heating. The heating–
circulation relay hypothesis is also tested by computing condensational heating from the
model’s passive tracer which represents specific humidity. This initial-value passive tracer experiment indeed supports the heating–circulation relay hypothesis that emerged from the observational analysis: tropical latent heating → circulation anomalies → latent heating in the North Pacific → circulation anomalies → latent heating in the North Atlantic.

The results from Chapter 2 imply that the impact of tropical heating should be taken into account even for a circulation feature attributable to extratropical heating. An important question that remains to be investigated is whether this heating–circulation relay mechanism can be employed for further understanding of atmospheric teleconnection patterns linking weather and climate anomalies at one location to those at other remote locations (Wallace and Gutzler 1981; Barnston and Livezey 1987). It is well established that tropical convection is an important forcing mechanism for teleconnections (Feldstein and Franzke 2017) such as the Pacific-North American pattern. It can be hypothesized that the heating–circulation relay processes may affect the growth, maintenance, and decay of teleconnections, if tropical heating is strong enough to promote extratropical latent heating through circulation anomalies, as seen from Chapter 2.

6.2 Conclusion to Chapter 3: A Mechanism for the Midwinter Minimum in North Pacific Storm Track Intensity from a Global Perspective

1. Does the midwinter minimum in storm track intensity occur on a hemispheric scale?

The presence of the midwinter minimum on a hemispheric scale is shown from the hemispheric-scale energetics which grounds the framework of the Lorenz energy cycle. The analyzed period is separated into severe suppression years and little-or-no suppression years using North Pacific storm-track intensity. The observation analysis depicts that during
severe suppression years, the global midwinter minimum is found in the hemispheric
synoptic-scale EKE with the relatively suppressed Eady growth rate over the extratropics.

2. Is there any relationship between tropical convection and suppression of global storm-track
intensity, and how is this relationship explained using the Lorenz energy framework?

The lead-lag relationships between energetics and western Pacific warm-pool convection
are examined through the composite analysis of the severe suppression and enhancement
years of North Pacific storm-track intensity. In the suppression years, the anomalous ZAPE
shows a minimum during early January, and this ZAPE minimum is preceded several days
by anomalously large planetary-scale EKE. The positive planetary-scale EKE anomaly is
in turn preceded two weeks by enhanced western warm-pool convection. The opposite
sequence of events is found during the enhancement years. In short, the lead-lag
relationships are as follows: enhanced warm pool convection $\rightarrow$ growth of planetary-scale
EKE $\rightarrow$ suppressed ZAPE $\rightarrow$ suppressed synoptic-scale EKE. These relationships can be
understood from the Lorenz energy cycle framework, similar to the planetary-scale EKE
life cycle of Baggett and Lee (2015). As summarized in the schematic diagram of Chapter
3, the tropical convection over the western warm-pool is enhanced during mid-December,
and the resultant increase in the zonal gradient of convective heating leads to excitation of
anomalous planetary-scale waves that propagate into the extratropics. As the planetary-
scale waves amplify, zonal winds and baroclinity increase locally, intensifying the
equatorward flank of the Pacific subtropical jet. However, at the expense of the planetary-
scale wave growth, on a hemispheric scale, the ZAPE and the hemisphere-wide baroclinity
are reduced. This reduction in ZAPE and baroclinity typically occurs from late December
to early January. Following this period of reduced hemispheric ZAPE and baroclinity, the January mean synoptic-scale storm track activity declines.

The results of Chapter 3 are obtained from observational analyses, and thus, the causal relationship between localized tropical convection and the midwinter minimum of synoptic-scale EKE on a hemispheric scale has not been yet demonstrated, which is left for future work. Yuval and Kaspi (2016) showed that the Pacific midwinter minimum in the EKE field can be reproduced by an idealized general circulation model with dry dynamics and the Newtonian relaxation scheme. In this regard, one might be able to design idealized model experiments to test if the midwinter minimum occurs on a hemispheric scale and if this midwinter minimum can be induced by localized tropical forcing. Also, from the heating–circulation relay mechanism, the suppression of storm tracks over the globe results in suppressed latent heating induced by storm track eddies, which in turn drive the subsequent circulation anomalies downstream. The impacts of extratropical latent heating anomalies during severe midwinter minimum years also merits further investigation to assess whether suppressed zonal asymmetry in extratropical latent heating leads to Arctic cooling, as seen in Chapter 2.

6.3 Conclusion to Chapter 4: Is the Stationary Wave Bias in CMIP5 Simulations Driven by Latent Heating Biases?

1. What are spatial patterns of stationary wave and precipitation biases in CMIP5 simulations during boreal winter?

It is found that in historical simulations of 28 CMIP5 models, there is a quadrupole pattern of stationary eddy meridional wind biases; negative biases over the East Siberian Sea and
the Greenland Sea, and positive biases over North America and northeastern Europe. Regarding climatological precipitation, systematic widespread biases occur over the tropics and extratropics in general. For tropical precipitation, there are positive biases over the Indian Ocean and Maritime Continent, while negative biases prevail over the western and central equatorial Pacific. For extratropical precipitation, the models have positive biases over most of the continents and coastal regions, except south of Greenland.

2. Can precipitation biases induce the stationary wave bias in climate models?

Idealized general circulation model experiments are conducted to test the hypothesis that biases in atmospheric latent heating, associated with precipitation biases in the CMIP5 models, reproduce the stationary wave bias. I find that the idealized model forced by precipitation biases reasonably reproduces the bias pattern of stationary eddy meridional wind described in the previous question.

3. Is there any relationship between tropical convection and suppression of global storm-track intensity, and how is this relationship explained from the framework of the Lorenz energy cycle?

Precipitation bias patch experiments are conducted to investigate regional precipitation bias that is responsible for regional eddy meridional wind bias. Each patch corresponds to a 30° × 30° grid box with a 10° interval in longitude and latitude. It is also allowed to partly overlap with other patches nearby. From these 360 patch experiments, I identify a region of precipitation bias that largely contributes to the development of regional stationary wave bias. For the bias over the Greenland-Scandinavia domain, the eastern North Pacific,
western North Atlantic, and tropical Atlantic precipitation biases are contributors, whereas for the bias over the East Siberian-Beaufort Sea, the central tropical Pacific and western subtropical Pacific are key regions of precipitation bias. Another important finding from the idealized model analysis is that the extratropical precipitation biases can be partly induced by tropical and upstream precipitation biases through moisture transport, which is consistent with the heating–circulation relay mechanism. More details can be found in the schematic diagram of Chapter 4.

Although researchers have devoted much effort to estimating variations in regional temperature and precipitation extremes by using future projections of climate models, the causal relationship between changes in the structure of stationary waves and regional climate variability has received less attention. It was recently shown (Garfinkel et al. 2020) that in future projections, zonally symmetric processes play a minor role in precipitation variability over some regions such as the United States, and this finding highlights the importance of zonally asymmetric, local processes in regional climate. Moreover, model projections of atmospheric circulation have substantial uncertainty on a regional scale, partly due to uncertainties in the stationary wave response (Shepherd 2014; Wills et al. 2019). Studying regional changes in the modeled stationary wave and their associated forcing are therefore essential to identify sources of model uncertainties in atmospheric circulation. In light of Chapters 2 and 4, future work includes identifying the mechanism of regional stationary-transient wave interference and quantitatively evaluating prediction skills in the associated temperature and precipitation extremes on seasonal to multiseasonal time scales.
6.4 Conclusion to Chapter 5: The Role of Planetary-scale Eddies on the Recent Isentropic Slope Trend during Boreal Winter

1. Does the extratropical isentropic slope exhibit a significant trend over the past few decades? If so, is this recent trend mainly driven by an enhancement of planetary-scale eddy activity?

Examining the observed trend of the isentropic slopes during boreal winter, I find that the extratropical isentropes from 50°N to 80°N have been flattening significantly for the recent 25-yr period. Projecting this trend pattern onto daily isentropic slope anomalies reveals that the trend pattern fluctuates at intraseasonal time scales. It is found that synoptic-scale eddy heat fluxes are indeed enhanced shortly after the isentropes are abnormally steep. However, following the peak of the synoptic-scale eddy heat fluxes, the expected decline in isentropic slope is rather modest. Instead, an enhancement of planetary-scale eddy heat fluxes leads to a significant decline in the isentropic slope. With this lead-lag relationship, there are also increasing trends of poleward propagating planetary-scale wave activity from the tropics and upward propagating planetary-scale wave activity from the extratropics. Overall, these results support the argument that the recent trend of extratropical isentropes is driven by planetary-scale eddy activity.

2. Is there a physical linkage between tropical convection and extratropical isentropic slopes?

From the composite analysis for days when daily isentropic slope anomalies match with the recent trend pattern, it is found that western Pacific warm-pool convection is enhanced prior to the growth of planetary-scale eddy heat flux and isentropic slope anomalies. Within the heating-circulation relay mechanism, it is hypothesized that the observed tropical and extratropical heating mostly accounts for these heat flux and slope anomalies. The causal
linkages between latent heating and the isentropic slope anomalies are tested by performing initial value calculations with the dynamical core of a general circulation model. It is concluded that the model calculations, forced by the observed tropical and extratropical latent heating, capture the structure of the extratropical isentropic slope trend pattern reasonably well.

3. Does an idealized model perturbed by latent heat forcing reproduce a warm-Arctic–cold-continent-like temperature pattern associated with the extratropical isentropic slope trend pattern?

It is found that the extratropical isentropic slope trend accompanies a warm-Arctic–cold-continent-like temperature pattern from the zonal-mean perspective. With the same model calculations forced by the observed latent heating, the model temperature response also depicts a spatial pattern of warm-Arctic–cold-continent-like temperature.

The results from Chapter 5 suggest that a warm-Arctic–cold-continent temperature pattern can be induced by remote diabatic heating anomalies, rather than local feedbacks such as sea-ice albedo feedback. An ongoing debate is what is the most important contributor to Arctic warming at unprecedented rates: local versus remote processes. Because the idealized model used in this study does not include local contributions from surface oceanic flux or sea-ice anomalies, using a coupled model would be helpful for a direct comparison of contributions from local surface properties and remote atmospheric diabatic heating. Moreover, another important research topic is whether the frequency of a warm-Arctic–cold-continent temperature pattern would increase in future projections. Mori et al. (2014) examined the 22 CMIP5 model simulations under RCP 4.5 scenario and found that the frequency of severe winters over the Eurasian continent may not
increase but rather decrease due to the circulation trend over the Arctic Ocean. It would be interesting for future work to update this result with the latest climate model simulations using more ensembles and updated physical parameterization schemes. If the relay mechanism described in Chapter 5 plays a significant role in driving the Arctic temperature variability, the frequency of severe winters over the continents with anomalous warming over the Arctic Ocean is likely to increase, corresponding to the intensifying trend of the zonal asymmetry in tropical convective heating.
Appendix A

Estimation of condensational heating from model passive tracer

We first compute saturation vapor pressure at each grid point using the solution to the Clausius-Clapeyron equation shown by Bohren and Albrecht (1998; their Equation 5.67):

\[ \ln \frac{e_s}{e_{s0}} = 6808 \left( \frac{1}{T_0} - \frac{1}{T} \right) - 5.09 \ln \frac{T}{T_0}, \]

where \( e_{s0} = 6.11 \text{mb} \) and \( T_0 = 273 \text{K} \). Next, we compute the amount of condensation by assuming that condensation takes place when the relative humidity reaches 100 \% (Manabe et al. 1965):

\[ q_{\text{cond}} = \begin{cases} q_{tr} - w_s & \text{if} \ r \geq 1 \\ 0 & \text{if} \ r < 1 \end{cases} \tag{1} \]

where \( q_{tr} \) is the tracer in the model and \( w_s \) is the saturation mixing ratio which can be computed from the saturation vapor pressure using

\[ w_s = \frac{0.622 e_s}{p - e_s}. \]

Because specific humidity, which is the tracer in our model, is very small (<<1) in the atmosphere, it can be approximated as the mixing ratio. The relative humidity can then be computed using the following relation,

\[ r \approx \frac{w}{w_s} = \frac{q_{tr}}{w_s}. \]

With the \( q_{\text{cond}} \) obtained from (1), the corresponding heating rate was computed using the following expression:

\[ Q = \frac{1}{\rho c_p D} \left[ \frac{1}{g} \int_{p_1}^{p_2} l_v q_{\text{cond}} dp \right], \]
where \( \rho \) is density of air, \( C_p \) is the specific heat of dry air at constant pressure, and \( D \) is the vertical depth of the heating, corresponding to pressure level between \( P_2 \) and \( P_1 \). To define the model anomaly, we subtract the heating rate of the control experiment from that of the perturbation experiments.
Appendix B

Supporting Information for Chapter 2

Lag-composites of PSI300 and Diabatic heating during SWI < −1.0 (468 days)

Fig. B1: (a)–(g) DJF Total 300-hPa streamfunction (contours with interval of $15 \times 10^6$ m$^2$ s$^{-1}$) and anomalies (shading), and (h)–(n) vertically averaged latent heating anomalies during the SWI–days. Dotted areas indicate statistical significance at the 10% level. Statistical significance is evaluated by employing a Monte Carlo simulation with 1000 random samples.
SWI and Lower tropospheric temperature north of 70N

Fig. B2: Lag composites of SWI values (blue) and lower tropospheric (700-1000 hPa) temperature anomalies averaged over north of 70°N (red) from lag days −20 to +20 in subsets: (a) $E_T | T_T^+$, (b) $E_T | T_B^+$, (c) $E_B | T_T^+$, (d) $E_B | T_B^+$, (e) $E_T | T_T^-$, (f) $E_T | T_B^-$, (g) $E_B | T_T^-$, and (h) $E_B | T_B^-$. The green (skyblue) line denotes temperature anomalies averaged over the Pacific (Atlantic) sector—90°E−270°E (90°W−90°E) at north of 70°N. For lower tropospheric temperature anomaly, the thick lines indicate statistical significance at the 5% level, evaluated by a Monte Carlo simulation with 1000 random samples.
Appendix C

Supporting Information for Chapter 3

Fig. C1: A histogram of January-mean (1980–2017), zonal-wavenumber 5-8 EKE over the North Pacific domain (20°N–60°N, 160°E–160°W). The statistics of the histogram—mean, standard deviation, skewness, kurtosis, and median – are indicated at the upper right corner.
Fig. C2: The longitude-time plots of OLR anomaly composites during (a) suppression years and (b) enhancement years. OLR anomaly is meridionally averaged over 15°S–15°N, and a Monte Carlo simulation with 5000 random samples of 13-yr (11-yr) for suppression (enhancement) years has been performed to test the statistical significance. Dotted areas indicate statistical significance at the 90% level. For visualization, a spatial smoothing has been applied by performing two iterations of a nine-point smoothing scheme provided by the NCAR Command Language.
Fig. C3: Composites of 500-hPa Eady growth rate anomaly as in Fig. 3 (shading) for (a) suppression years and (b) enhancement years. Black contours denote corresponding composites of 300-hPa zonal wind anomaly retaining wavenumber 1 to 3, with contour interval of 2 m s$^{-1}$. A spatial smoothing has been applied to Eady growth rate anomaly composites by performing two iterations of a nine-point local smoothing scheme provided by NCAR Command Language.
Composites of OLR / PSI300 / EKE (k=5-8) anomaly during suppression years

Fig. C4: Composites of (a-c) outgoing longwave radiation anomalies, (d-f) 300-hPa streamfunction from 31 December to 6 January, and (g-i) synoptic-scale EKE vertically integrated from 300 hPa to 1000 hPa. Dates used for composites are (a,d,g) from 10 December to 30 December, (b,e,h) 31 December to 6 January, and (c,f,i) 7 January to 22 January during suppression years. Black contours in second and third row denote the 300-hPa climatological streamfunction and climatological synoptic-scale EKE during December-January-February, while the contour intervals are $15 \times 10^6$ m$^2$ s$^{-1}$ and $0.5 \times 10^5$ J m$^{-2}$, respectively.
Composites of OLR / PSI300 / EKE (k=5-8) anomaly during enhancement years

**Fig. C5:** As in Fig. C4, but during enhancement years.
Appendix D

Supporting Information for Chapter 4

D.1 Experimental Design

In this study, the 1979-2005 DJF climatological state in ERA-Interim is inputted as the initial state of the model. Because this climatological mean flow does not correspond to a steady solution of the model equations, without an additional forcing term, the climatological fields freely evolve with time. To prevent this drift, we compute a forcing term $\mathcal{F}$ by integrating the model equations one-time step forward from the initial state. $\mathcal{F}$ is the negative of the time tendency of the initial model fields. If a perturbation forcing is added to the model, however, the model solution deviates from the initial climatological state. This deviation is regarded as the response to the perturbation. In this study, the perturbation forcing is the estimated CMIP5 minus ERA-Interim heating field.

D.2 Passive Tracer Analysis

The goal of the passive tracer analysis is to investigate the impact of circulation driven by upstream and tropical heating biases on moisture transport downstream, and to see if the resulting passive tracer field resembles the CMIP5 precipitation bias field in the downstream region. The initial state of the model passive tracer is the winter climatological specific humidity field, and as was discussed in D.1, this initial state is maintained throughout model integration if there is no additional forcing. A passive tracer, by construction, does not provide any thermodynamic and dynamic feedback onto the model fields. Therefore, the model passive tracer field represents the moisture field transported by a circulation in question. In our case, the circulation is the model response to the CMIP5 latent heating.
Fig. D1: CMIP5 multimodel mean biases in 250-hPa $\bar{v}^*$ during 1979-2005 DJF, computed with (a) ERA-Interim dataset as in Fig. 4.1c and (b) JRA-55 reanalysis dataset.
Fig. D2: (a) CMIP5 multimodel mean biases in precipitation, computed with daily GPCP climatology over the period of 1997-2005 DJF. Stippling indicates mean biases where at least 80% of the CMIP5 models have the same sign of bias. (b) GFDL model 250-hPa $\vec{v}'$ response averaged over model days 10-13 forced by CMIP5 latent heating biases converted from precipitation biases shown in (a).
Table D.1: Summary of the 28 historical CMIP5 simulations used in this study.

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Appendix E

Supporting Information for Chapter 5

Fig. E1: MJO phase-space diagram for (a) ISI > 1.0 event days, (b) 1/3 top bin of ISI > 1.0 event days (black) and 1/3 bottom bin of ISI > 1.0 event days (red). Each lag day has been denoted by circles, and 5-day labels are overlaid on each phase line. A two-sided Student’s t-test has been performed for statistical significance test at the 5% (10%) level, denoted by large (medium) circles. For visualization, axis extents 0.5 and 1.2 are used in (a) and (b), respectively.
Fig. E2: Zonal-mean composites of (a)–(d) anomalous isentropic slope, (e)–(h) planetary-scale eddy heat flux, (i)–(l) synoptic-scale eddy heat flux, and (m)–(p) temperature at (a),(e),(i),(m) lag day –6, (b),(f),(j),(n) lag day –3, (c),(g),(k),(o) lag day 0, and (d),(h),(l),(p) lag day 3 for a case study of 1984/85 DJF. The first row represents the instantaneous field on December 26th, 1984, the second row on December 29th, 1984, the third row on January 1st, 1985, and the fourth row on January 4th, 1985. The red line in each panel denotes the 2-PVU isoline. Note that the range of the color bars is four times larger than that in Fig. 5.4.
**Fig. E3:** The longitude-lag plot of anomalous outgoing longwave radiation averaged over 30°S–30°N ranging from December 7th, 1984 to January 26th, 1985. Lag day 0 corresponds to January 1st, 1985. Prior to plotting, a nine-point local smoothing was applied twice to the outgoing longwave radiation field. Note that the range of the color bar in this plot is three times larger than that in Fig. 5.5.
Fig. E4: (a) Composite of vertically integrated model passive tracer concentration anomalies averaged over model days 14 to 18, when the model is forced by tropical heating. (b) Composite of the ERA-Interim vertically integrated moisture flux convergence anomalies driven by planetary-scale waves \( \equiv -\frac{1}{\hat{g}} \nabla \cdot \int \hat{v} \, \hat{q} \, dp \) averaged over lag days \(-7\) to \(-3\). (c) Same as (b), but for synoptic-scale waves \( \equiv -\frac{1}{\hat{g}} \nabla \cdot \int \hat{v} \, \hat{q} \, dp \). A two-sided Monte Carlo simulation with 1000 random subsets is performed to evaluate the statistical significance in (b) and (c), and the dotted areas indicate statistical significance at the 5% level.
Appendix F

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