The surface heating efficiency of atmospheric energy flux events during Arctic winter

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THE SURFACE HEATING EFFICIENCY OF ATMOSPHERIC ENERGY FLUX EVENTS DURING ARCTIC WINTER

by

Christopher J. Cardinale

A Dissertation
Submitted to the University at Albany, State University of New York
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The flux of moist static energy (MSE) into the polar regions plays a key role in the energy budget and climate of the polar regions. While usually studied from a vertically integrated perspective ($F_{\text{wall}}$), this dissertation examines its vertical structure, using the NASA-MERRA-2 reanalysis to compute climatological and anomalous fluxes of sensible, latent, and potential energy across 70°N and 65°S. This dissertation applies an energy budget analysis to winter-season synoptic periods of increased tropospheric ($F_{\text{trop}}$) and stratospheric ($F_{\text{strat}}$) energy flux convergence events and examines the processes that drive Arctic anomalous surface warming and sea ice loss during $F_{\text{trop}}$ events. This dissertation also quantifies the partial contribution to Arctic winter surface warming from changes in the MSE flux convergence and its coupling to the surface in the RCP8.5 warming scenario of the CESM Large Ensemble (CESM-LE).

The vertical structure of the climatological flux is bimodal, with peaks in the mid-to lower-troposphere and mid- to upper-stratosphere. The near zero flux at the tropopause defines the boundary between $F_{\text{strat}}$ and $F_{\text{trop}}$. Especially at 70°N, $F_{\text{strat}}$ is found to be important to the climatology and variability of $F_{\text{wall}}$, contributing 20.9 W m$^{-2}$ to $F_{\text{wall}}$ (19% of $F_{\text{wall}}$) during the winter and explaining 23% of the variance of $F_{\text{wall}}$. During winter, an anomalous poleward increase in $F_{\text{strat}}$ preceding a sudden stratospheric warming is followed by an increase in outgoing longwave radiation anomalies, with little influence on the surface energy budget of the Arctic. Conversely, a majority of the energy input by an anomalous poleward increase in $F_{\text{trop}}$ goes toward heating the Arctic surface. $F_{\text{trop}}$ is found to be a better metric than $F_{\text{wall}}$ for evaluating the influence of atmospheric circulations on the Arctic surface climate.

During an $F_{\text{trop}}$ event, a poleward anomaly in $F_{\text{trop}}$ initially increases the sensible and latent energy of the Arctic troposphere; as the warm and moist troposphere loses heat, the anomalous energy source is balanced by a flux upward across the tropopause and a downward net surface flux. A new metric for the Arctic surface heating efficiency ($E_{\text{trop}}$) is defined, which measures the fraction of the energy source that reaches the surface. Composites of high, medium, and low-efficiency events help identify key physical factors that drive...
$E_{\text{trop}}$, including the vertical structure of $F_{\text{trop}}$ and Arctic surface preconditioning. In high-efficiency events ($E_{\text{trop}} \geq 0.63$), a bottom-heavy poleward $F_{\text{trop}}$ occurs in the presence of an anomalously warm and unstratified Arctic—a consequence of decreased sea ice—resulting in increased vertical mixing, enhanced near-surface warming and moistening, and further sea ice loss. Smaller $E_{\text{trop}}$, and thus weaker surface impacts, are found in events with anomalously large initial sea ice extent and more vertically uniform $F_{\text{trop}}$. These differences in $E_{\text{trop}}$ are manifest primarily through turbulent heat fluxes rather than downward longwave radiation. The frequency of high-efficiency events has increased from the period 1980–1999 to 2000–2019, contributing to Arctic surface warming and sea ice decline.

An analysis of high, medium, and low-efficiency events in the CESM-LE verify the key physical factors that determine $E_{\text{trop}}$. In the RCP8.5 warming scenario, the winter mean $F_{\text{trop}}$ is found to decrease by 9.5 W m$^{-2}$, dominated by a decrease in the dry static energy (sensible and geopotential) component and only partially compensated by the latent component. Winter mean $E_{\text{trop}}$ is found to increase by 5.7%, indicating an increased coupling between $F_{\text{trop}}$ and the surface energy budget. The increase in $E_{\text{trop}}$ is consistent with an increased frequency of high-efficiency events, likely driven by decreased Arctic lower-tropospheric stability. The decrease in $F_{\text{trop}}$ is overcompensated by an increase in $E_{\text{trop}}$, contributing 0.7 W m$^{-2}$ of surface heating or 0.3 K of warming. From a surface energy budget perspective, changes in the poleward atmospheric energy flux contribute to the bottom-amplified warming of the Arctic despite a decrease in the vertical integral of the energy flux.
I would like to thank my advisor, Dr. Brian E. J. Rose. His mentorship, patience, and enthusiasm throughout my time in graduate school was greatly appreciated and helped me grow as a scientist. I am especially grateful for the summer job opportunity after completing my bachelor’s degree, which served as a springboard for all the research to follow.

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1.1 Motivation

Recent observations indicate a bottom-amplified warming of the Arctic, with surface air temperatures increasing at a faster rate than the global average, especially during the cold season (Screen and Simmonds, 2010a; Cohen et al., 2014). Winter sea ice has also declined in recent decades, especially in the Barents–Kara Seas, where relatively large reductions in sea ice concentration (SIC; D.-S. R. Park et al., 2015) and thickness (King et al., 2017) have been observed. The bottom-amplified warming is also a robust feature of climate model projections (e.g., Boeke and Taylor, 2018).

Arctic surface warming contributions are commonly based on a top-of-atmosphere (TOA) budget analysis. In this TOA framework, the vertically uniform temperature change to each forcing and feedback (water vapor, cloud, albedo, CO₂, Planck, atmospheric energy transport, and surface fluxes) is computed, with deviations from the vertically uniform warming denoted as the lapse-rate feedback. In an attribution study, Pithan and Mauritsen (2014) found that important contributions to future Arctic warming include the lapse-rate and albedo feedbacks; while the change in the atmospheric energy transport is only a small positive contributor. During winter, dominant contributors include the lapse-rate feedback and surface heat loss (i.e., the ice-insulation feedback; Screen and Simmonds, 2010b; Stroeve et al., 2012; Burt et al., 2016).

Attribution studies typically do not consider contributions to the lapse-rate feedback. Contributions from sea ice loss investigated in fixed sea ice experiments (e.g., Deser et al., 2010; Dai et al., 2019; Jenkins and Dai, 2021) conclude that the bottom-amplified warming of the Arctic, and thus the lapse-rate feedback, is tied to sea ice loss. However, the precise mechanisms that drive the bottom-amplified warming are still under active debate (e.g., Henry et al., 2021).

Model simulations disagree on the magnitude and sign of the energy transport, and the
role of atmospheric circulations in Arctic warming is still debated (e.g., Taylor et al., 2022). Using a TOA budget analysis, contributions to surface warming from changes in the vertically integrated polar cap–averaged moist static energy (MSE; sensible, latent, and geopotential) flux convergence ($F_{\text{wall}}$) are relatively small in a multimodel mean (Pithan and Mauritsen, 2014; Feldl et al., 2020). However, from a synoptic perspective, it has been suggested that an increased frequency of moist intrusion events (synoptic-scale events associated with intense poleward heat and moisture fluxes) is an important driver of recent trends of Arctic surface warming (Woods and Caballero, 2016; Gong et al., 2017; Lee et al., 2017) due to the enhanced warming effect associated with moisture fluxes.

The broad objective of this dissertation is to improve the understanding of the role of atmospheric circulations in winter Arctic warming.

1.2 Literature review

1.2.1 The vertical structure of the polar cap–averaged MSE flux convergence

The polar regions are marked by weak annual mean insolation and would be extremely cold were it not for the energetic input into the regions from atmospheric and oceanic energy transport. In the Arctic, poleward of 70°N, the annual average poleward energy flux convergence nearly balances the net radiative deficit of the region (110 W m$^{-2}$) and is dominated by $F_{\text{wall}}$ (100 W m$^{-2}$), while poleward oceanic energy flux convergence (10 W m$^{-2}$) is an order of magnitude smaller (e.g., Serreze et al., 2007). Improving the estimate and understanding of the fluxes contributing to $F_{\text{wall}}$ in the Arctic energy budget has been a recurring goal (e.g., Nakamura and Oort, 1988; Serreze et al., 2007; Porter et al., 2010; Mayer et al., 2019).

The role of the stratospheric contribution to $F_{\text{wall}}$ ($F_{\text{strat}}$) in the energy budget of the Arctic polar cap has not been carefully studied. It is reasonable to assume that $F_{\text{strat}}$ is relatively small compared to the tropospheric contribution ($F_{\text{trop}}$), as the stratosphere is dry and makes up a small percentage of atmospheric mass (10–30% depending on latitude). However, $F_{\text{strat}}$ could be important during periods of anomalous stratospheric conditions, such as sudden stratospheric warming (SSW) events, which can have impacts lasting on the order of months (Kidston et al., 2015). SSW events are known to be associated with large poleward heat flux anomalies at 100 hPa (e.g., Polvani and Waugh, 2004).
Comprehensive studies on the variability of $F_{\text{strat}}$ and its contribution to the polar cap energy budget are lacking. Overland and Turet (1994) showed that $F_{\text{strat}}$ is a non-negligible portion of $F_{\text{wall}}$ at 70°N [consistent with Fig. 13.10 in Peixoto and Oort (1992)] with a large seasonality and maximum values during the winter (NDJFM). The vertical structure of $F_{\text{wall}}$ reported by Overland and Turet (1994) was calculated from spatially coarse reanalysis data (2.5° x 5° horizontal resolution and 11 vertical levels) and has not been updated using a modern high-resolution reanalysis.

### 1.2.2 Synoptic-scale tropospheric energy flux events

Episodic $F_{\text{trop}}$ events—commonly referred to as moist intrusion events—are synoptic-scale events of poleward $F_{\text{trop}}$ anomalies that temporarily warm the Arctic surface and reduce sea ice growth (Doyle et al., 2011; Yoo et al., 2012; Woods et al., 2013; D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Baggett and Lee, 2015; Woods and Caballero, 2016; Luo et al., 2016; Baggett and Lee, 2017; Gong et al., 2017; Gong and Luo, 2017; Luo et al., 2017; Johansson et al., 2017; Zhong et al., 2018; Liu et al., 2018; Chen et al., 2018; Luo et al., 2019; Tyrlis et al., 2019; Graham et al., 2019b) via the following sequence:

1) Anomalous tropospheric sensible and latent heat flux convergence contribute to the warming and moistening of the troposphere over the region and at the pressure level of the anomalous convergence. These events are typically associated with atmospheric blocking over the Ural Mountains [i.e., Ural blocking (UB)].

2) The anomalously warm and moist troposphere and increased clouds result in increased downwelling longwave radiation (DLR) at the surface (mainly over sea ice), initiating surface heating (i.e., reduced surface heat loss and the reduced growth or melting of sea ice). Increased surface air temperatures and specific humidity also suppress upward surface turbulent sensible and latent heat fluxes (SHLH), reducing surface heat loss (mainly over open ocean). Storm driven mixing of warm Atlantic water causes further ice melt.

Positive $F_{\text{trop}}$ anomalies cause tropospheric heating and moistening. This warming results in increased longwave emission from the atmosphere, both upward and downward to the surface. The efficiency with which increased $F_{\text{trop}}$ heats the surface is intimately tied
to the partition of this radiative cooling between the upwelling flux across the tropopause and the downwelling flux to the surface. For a polar cap in “radiative-advective equilibrium” (Cronin and Jansen, 2016), the surface heating effect decreases monotonically with the vertical height of the advective heat source (i.e., MSE flux convergence); concentrating the atmospheric heating closer to the surface will result in a larger fraction of the anomalous poleward MSE flux convergence going into surface heating. The fraction of energy that reaches the surface is referred to as the surface heating efficiency in this dissertation. Changes in and drivers of the surface heating efficiency have only been studied in highly idealized scenarios (e.g., Cronin and Jansen, 2016).

The surface heating efficiency of $F_{\text{trop}}$ events are also potentially linked to tropical convection and the initial state of the Arctic surface. Tropical convection in the Pacific warm pool can trigger planetary scale waves that amplify the climatological stationary wave pattern, initially driving adiabatic warming in the Arctic and subsequently driving long duration surface warming associated with large eddy fluxes of sensible and latent heat and increased DLR (Yoo et al., 2012; Baggett and Lee, 2015, 2017). The response of the Arctic surface during an event appears sensitive to the type of atmospheric blocking pattern (Chen et al., 2018; Luo et al., 2019). Chen et al. (2018) found that more persistent UB events are associated with an initially lower SIC and a more intense surface response in the Barents–Kara Seas due to increased moisture and DLR.

$F_{\text{trop}}$ events associated with a greater surface warming are typically linked to increased DLR (e.g., Chen et al., 2018). However, Vargas Zeppetello et al. (2019) showed that in a mean climatic state, DLR is largely determined by surface temperatures, questioning the role of DLR in driving the Arctic surface response to an event. Therefore, the processes that drive the surface response are not fully understood.

### 1.2.3 The role of energy transport in Arctic surface warming

In traditional attribution studies, $F_{\text{wall}}$ changes are positive, but are relatively small, and contribute to little Arctic warming and Arctic amplification in a multimodel mean when the polar cap is defined from 60 to 90°N (e.g., Pithan and Mauritsen, 2014; Feldl et al., 2020; Hahn et al., 2021). Small contributions are found in both the winter and summer (Pithan and Mauritsen, 2014; Hahn et al., 2021). $F_{\text{wall}}$ changes are anti-correlated with
Arctic amplification and are found to dampen the inter-model spread; i.e., $F_{\text{wall}}$ decreases in models with relatively large Arctic amplification and increases in models with relatively weak Arctic amplification (e.g., Boeke and Taylor, 2018).

$F_{\text{wall}}$ is often partitioned into latent heat (LH) and dry static energy (DSE; SH+GP) components (e.g., Koenigk et al., 2013; Feldl et al., 2020; Hahn et al., 2021). LH has an amplified warming effect on the Arctic due to its relationship with cloud and atmospheric emissivity changes (Graversen and Burtu, 2016; Baggett and Lee, 2017; Graversen and Langen, 2019). The DSE component of $F_{\text{wall}}$ decreases in a future warming climate, compensated by an increase in the LH component (e.g., Hwang et al., 2011; Koenigk et al., 2013; Graversen and Burtu, 2016; Hahn et al., 2021). The degree to which the LH component compensates the DSE component depends on the definition of the Arctic domain [more models show a decrease in $F_{\text{wall}}$ when defined at 70°N rather than 60°N (Hwang et al., 2011)] and on the degree of the Arctic amplification (Boeke and Taylor, 2018).

Although not typically considered in attribution studies, contributions from energy transport changes to the lapse-rate feedback have been quantified using an alternative attribution method in a single-column model (Henry et al., 2021). In various 4 x CO$_2$ experiments, Henry et al. (2021) found that increased energy transport preferentially heats the upper troposphere—consistent with Feldl et al. (2020)—and results in a smaller surface warming contribution than in a TOA approach. In experiments where an Arctic surface heat source is added (i.e., mimicking the presence of sea ice loss in the single-column model), energy transport changes result in less surface warming due to reductions in the DSE component that preferentially cool the lower troposphere.

In apparent contrast to the results of attribution studies, it has been suggested that the increased frequency of synoptic-scale tropospheric heat and moisture fluxes into the Arctic [i.e., moist intrusion events; Woods and Caballero (2016)] is an important driver—perhaps the main driver—of recent trends in surface temperature (e.g., Woods and Caballero, 2016; Gong et al., 2017; Lee et al., 2017) and sea ice (e.g., D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Woods and Caballero, 2016). The increased frequency of events over recent decades—consistent with the increased duration (Rinke et al., 2017) or frequency (Valkonen et al., 2021) of Arctic cyclones—is potentially driven by the increased occurrence of tropical convection over the Pacific warm pool (Baggett and Lee, 2017) or an increased frequency
of UB events (Luo et al., 2016; Rinke et al., 2017; Gong and Luo, 2017; Chen et al., 2018; Luo et al., 2019). The increased UB frequency could be driven by reduced sea ice in the Barents–Kara Seas and the associated atmospheric circulation response, a potential positive feedback mechanism (Gong and Luo, 2017; Chen et al., 2018).

It’s possible that the increased frequency of these synoptic-scale weather events in a warming Arctic are not found in climate models. Peings (2019) found no significant atmospheric response in model simulations with perturbed ice in the Barents–Kara Seas. However, the divergence between modeling and observational studies potentially results from models underestimating the role of sea ice in driving an atmospheric circulation response (Overland et al., 2021) or from a misinterpretation of causality in observational studies—atmospheric circulation anomalies drive rather than respond to sea ice anomalies (Blackport et al., 2019).

1.3 Research questions and hypotheses

The following research questions and hypothesis are addressed in this dissertation:

1. Is $F_{trop}$ a better metric than $F_{wall}$ for the influence of atmospheric circulations on the Arctic surface climate? Hypothesis: $F_{trop}$ is a better metric due to the relationship between the vertical structure of the MSE flux convergence and the surface energy budget.

2. What drives the surface heating efficiency of synoptic-scale $F_{trop}$ events during winter? Hypothesis: Concentrating the MSE flux convergence closer to the surface increases the surface heating efficiency.

3. What is the role of changes in the atmospheric energy transport and surface heating efficiency in future Arctic warming? Hypothesis: An increase in the surface heating efficiency mitigates a decrease in $F_{trop}$.

Chapter 2 addresses the first research question, and Chapters 3 and 4 address the second and third research questions, respectively. The materials in Chapter 2 have been peer-reviewed and published in the Journal of Climate (JCLI) and can be found at Cardinale et al. (2021). The materials in Chapter 3, co-authored with Dr. Brian E. J. Rose, have been submitted to
JCLI and has received a “minor revision” decision. The materials in Chapter 4, co-authored with Dr. Brian E. J. Rose, are expected to be submitted to Geophysical Research Letters (GRL).
CHAPTER 2

Stratospheric and Tropospheric Flux Contributions to the Polar Cap Energy Budgets

2.1 Introduction

This chapter has peer-reviewed and published in JCLI and can be found at Cardinale et al. (2021). Minor changes to the format have been made.

The polar regions are marked by weak annual mean insolation and would be extremely cold were it not for the energetic input into the regions from atmospheric and oceanic energy transport. In the Arctic, poleward of 70°N, the annual average poleward energy flux convergence nearly balances the net radiative deficit of the region (110 W m\(^{-2}\)) and is dominated by atmospheric energy flux convergence (100 W m\(^{-2}\)), while poleward oceanic energy flux convergence (10 W m\(^{-2}\)) is an order of magnitude smaller (e.g., Serreze et al., 2007). The polar cap–averaged atmospheric flux convergence, hereafter \(F_{\text{wall}}\), is proportional to the zonally and vertically integrated moist static energy (MSE; sensible heat, latent heat, and geopotential) flux across a boundary defining the polar cap. Improving the estimate and understanding of the fluxes contributing to \(F_{\text{wall}}\) in the Arctic energy budget has been a recurring goal (e.g., Nakamura and Oort, 1988; Serreze et al., 2007; Porter et al., 2010; Mayer et al., 2019).

Using atmospheric reanalyses, the poleward flux of MSE has been linked to variability and long-term changes in Arctic surface and free tropospheric temperatures. On synoptic timescales, anomalies in MSE flux convergence have been linked to changes in Arctic sea-ice thickness (D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Graham et al., 2019b) and surface warming (Woods and Caballero, 2016) via the following sequence:

1) MSE flux convergence initially increases the moist enthalpy (latent and sensible heat) of the atmospheric column at the pressure level of the anomalous flux [see Eq. (4) in Trenberth and Solomon (1994)].

2) The warm and moist atmosphere subsequently fluxes longwave radiation downward to
the surface to initiate surface heating (i.e., surface warming and ice melt).

At cold Arctic temperatures, the latent component of the moist enthalpy storage is small; therefore, the moist enthalpy tendency is very nearly proportional to the temperature tendency.

Linking the poleward atmospheric energy flux to the vertical structure of multi-decadal trends in Arctic temperatures, Graversen et al. (2008) found that a significant proportion of the vertical structure of Arctic warming in the summer half-year can be explained by changes in $F_{\text{wall}}$ at 60°N. Yang et al. (2010) compared the vertical structure of total and $F_{\text{wall}}$-congruent temperature trends during decades of cooling and warming in the Arctic. Their study concluded that decadal variation of Arctic free troposphere temperature is heavily influenced by changes in the poleward flux of atmospheric energy at 65°N, associated with the changing intensity of the polar meridional circulation cell.

More recent studies using atmospheric reanalyses have linked different components of the poleward energy flux to variability in Arctic surface temperatures. Baggett and Lee (2015) found the winter Arctic warming (at 2 meters) associated with planetary-scale waves to be greater and more persistent than the warming associated with synoptic-scale waves. During the planetary wave life cycle, significant convergence of latent and sensible heat fluxes in the Arctic increases the downward longwave radiation, warming the surface (Baggett and Lee, 2017). The anomalous energy flux into the Arctic was associated with an amplification of the climatological stationary wave pattern forced by tropical convection in the Pacific warm pool. Graversen and Burtu (2016) also found that Arctic warming was associated with enhanced $F_{\text{wall}}$ (especially the latent heat component) by planetary-scale waves, whereas $F_{\text{wall}}$ by synoptic-scale waves was correlated with an enhanced meridional temperature gradient, and thus anti-correlated with Arctic temperature anomalies.

The impact of $F_{\text{wall}}$ on high-latitude climate variability and long-term changes are established; however, the analyses on this topic have focused on fluxes linked to tropospheric circulations, whereas the potential role of the stratosphere in $F_{\text{wall}}$ anomalies has not been investigated. It is reasonable to assume that the stratospheric contribution to $F_{\text{wall}}$ ($F_{\text{strat}}$) is relatively small compared to the tropospheric contribution ($F_{\text{trop}}$), as the stratosphere is dry and makes up a small percentage of atmospheric mass (10–30% depending on latitude). However, $F_{\text{strat}}$ could be important during periods of anomalous stratospheric conditions,
such as sudden stratospheric warming (SSW) events, which can have impacts lasting on the order of months (Kidston et al., 2015). SSW events are known to be associated with large poleward heat flux anomalies at 100 hPa (e.g., Polvani and Waugh, 2004). This linkage is suggested by the basic dynamical theory of SSWs (e.g., Limpasuvan et al., 2004), where meridional eddy heat flux is a measurable dynamical proxy for the vertical propagation of planetary wave activity (e.g., Edmon et al., 1980). Deceleration of the stratospheric vortex is accomplished through breaking of these upward-propagating planetary waves (e.g., Matsuno, 1971). Thus, there is a long tradition in the stratospheric literature of using lower-stratospheric horizontal eddy heat fluxes as a diagnostic for this coupling between the troposphere and stratosphere (e.g., Polvani and Waugh, 2004; Butler et al., 2017). However, the role of $F_{\text{strat}}$ in the energy budget of the Arctic polar cap has not been studied as carefully.

Comprehensive studies on the variability of $F_{\text{strat}}$ and its contribution to the polar cap energy budget are lacking. Overland and Turet (1994) showed that $F_{\text{strat}}$ is a non-negligible portion of $F_{\text{wall}}$ at 70°N [consistent with Fig. 13.10 in Peixoto and Oort (1992)] with a large seasonality and maximum values during the winter (NDJFM). The vertical structure of $F_{\text{wall}}$ reported by Overland and Turet (1994) was calculated from spatially coarse reanalysis data (2.5° x 5° horizontal resolution and 11 vertical levels) and has not been updated using a modern high-resolution reanalysis.

We speculate that $F_{\text{strat}}$ variability is only very weakly coupled to polar cap surface temperatures. Positive $F_{\text{wall}}$ anomalies cause air temperatures within the polar cap to increase. This warming results in increased longwave emission from the atmosphere, both upward (as outgoing longwave radiation, OLR) and downward to the surface. The efficiency with which increased $F_{\text{wall}}$ heats the surface is intimately tied to the partition of this radiative cooling between the upwelling flux to space and the downwelling flux to the surface. For a polar cap in “radiative-advective equilibrium” (Cronin and Jansen, 2016), the surface warming effect decreases monotonically with the vertical height of the advective heat source (i.e., MSE flux convergence). Concentrating the atmospheric heating closer to the surface will result in a larger fraction of the anomalous poleward MSE flux convergence going into surface heating versus longwave emission to space. Thus, the impact on the Arctic surface climate from stratospheric heating ought to be much smaller than the impact of a similar magnitude of tropospheric heating.
Much of the literature reviewed above (e.g., Graversen et al., 2008; Yang et al., 2010; Baggett and Lee, 2015; Graversen and Burtu, 2016; Baggett and Lee, 2017) includes relationships between vertically integrated energy fluxes and the climate of the Arctic polar cap. In this chapter, we will explicitly separate the stratospheric and tropospheric contributions to the climatology and variability of $F_{\text{wall}}$. The stratospheric component will be linked to the literature on stratospheric variability (e.g., Polvani and Waugh, 2004; Butler et al., 2017). Additionally, we will quantify the relationship between $F_{\text{wall}}$ and the Arctic surface climate after removing the effects of stratospheric variability.

To characterize the stratospheric and tropospheric contributions to $F_{\text{wall}}$ and compare their relative impacts on the Arctic surface climate, the analysis presented considers two key themes and associated research questions:

1) Climatology and Variability:

   (a) Using a modern reanalysis, the Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2), what is the vertical structure of the climatological $F_{\text{wall}}$ and can it be cleanly separated into contributions from the troposphere ($F_{\text{trop}}$) and stratosphere ($F_{\text{strat}}$)?

   (b) What is the variability of $F_{\text{wall}}$, $F_{\text{strat}}$, and $F_{\text{trop}}$ in both hemispheres?

   (c) How much of the variance of $F_{\text{wall}}$ does $F_{\text{strat}}$ distinctly explain?

2) Link to the Arctic Climate:

   (a) In the MERRA-2, what is the Arctic response following an anomalous poleward increase in $F_{\text{wall}}$ when dominated by either $F_{\text{strat}}$ or $F_{\text{trop}}$?

   (b) After removing $F_{\text{strat}}$ from $F_{\text{wall}}$, is there a stronger correlation between $F_{\text{wall}}$ and warming of the Arctic lower-troposphere?

The data and methods used for this analysis are described in section 2.2. Theme 1 will be addressed in section 2.3 of this chapter; theme 2 will be addressed in section 2.4.
2.2 Data & Methods

2.2.1 MERRA-2

We use atmospheric winds, temperature, specific humidity, geopotential, radiative fluxes at the top of atmosphere (TOA) and surface, and surface turbulent energy fluxes from the Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2), the latest atmospheric reanalysis (1980–present) produced by NASA’s Global Modeling and Assimilation Office (GMAO). MERRA-2 has a horizontal resolution of $0.5^\circ \times 0.625^\circ$, 72 vertical levels with output interpolated to 42 pressure levels up to 0.1 hPa, and a temporal resolution of 3-hours (GMAO, 2015). The period 1980–2016 (37 years) is used in this analysis.

Notable improvements from MERRA to MERRA-2 include assimilation of additional satellite observations, conservation of dry mass, and reduced spurious trends and jumps related to changes in the observing system (Bosilovich et al., 2015). Although many of the updates pertain to tropospheric processes, MERRA-2 improves ozone representation and gravity wave drag parameterization. The General Circulation Model (GCM) component uses a cubed-sphere grid; thus, eliminating computation instability issues near the poles, which can be important during SSW events where cross-polar flow can occur and for studies of the high-latitudes. Information on the initial evaluation of the climate in MERRA-2 can be found in Bosilovich et al. (2015), and information on input observations can be found in McCarty et al. (2016).

2.2.2 Contributions to $F_{\text{wall}}$

To calculate the MSE flux and $F_{\text{wall}}$, a method similar to Overland and Turet (1994) is followed. MSE is defined by:

$$\text{MSE} = c_p T + gz + L_v q,$$

where $c_p$ is the specific heat of dry air at constant pressure, $T$ is temperature, $g$ is the gravitational acceleration, $z$ is geopotential height, $L_v$ is the latent heat of vaporization for water, and $q$ is specific humidity. The contribution from the kinetic energy is small and has been neglected.
The meridional MSE flux is then \( v \)MSE, where \( v \) is the meridional component of the wind. \( F_{\text{wall}} \) is defined as the polar cap–averaged MSE flux convergence, equal to the zonally and vertically integrated flux through the bounding latitude divided by the area of the polar cap:

\[
F_{\text{wall}} = \frac{C}{2\pi A} \int_{0}^{2\pi} \int_{0}^{P_s} \left( c_p v T + g v z + L v v q \right) \frac{d\lambda dp}{g},
\]

where \( C \) is the circumference of the latitude defining the polar cap boundary and \( A \) is the area of the polar cap. Terms on the RHS correspond to the flux of sensible heat (SH), geopotential (GP), and latent heat (LH), respectively. \( F_{\text{wall}} \) is computed instantaneously from the 3-hourly data and averaged monthly and daily to define the climatological fluxes and anomalous \( F_{\text{wall}} \) events, respectively.

Each component of the MSE flux, SH, LH, and GP, can be expanded into an eddy flux (EF), a mean meridional circulation flux (MMC), and a net mass flux (NMF). The NMF has been removed from calculations of \( F_{\text{wall}} \) due to unphysical high-frequency noise associated with the net atmospheric mass flux into the polar cap. The NMF, by definition, has no vertical structure, meaning the results in this chapter are not sensitive to this term; this is further explored in section 2.2.3.

The MSE flux has units of J kg\(^{-1}\) m s\(^{-1}\). This flux can be written as the local contribution to the integrated flux convergence in terms of W m\(^{-2}\) (100 hPa\(^{-1}\)) with a conversion factor of \((C/Ag)10^4\). To recover units of W m\(^{-2}\), flux values are vertically integrated. This conversion factor is used to more explicitly compare the flux between the two hemispheres, since we define the latitude of the polar cap boundary differently in the Northern and Southern Hemisphere. 65°S is used rather than 70°S due to large differences in terrain between the latitudes (about 50% of 70°S is over high Antarctic terrain).

### 2.2.3 \( F_{\text{wall}} \) decomposition and quantifying the Net Mass Flux

The components of the moist static energy (MSE) flux, sensible heat flux (SH), latent heat flux (LH), and geopotential flux (GP), can be expanded into an eddy flux (EF), a mean meridional circulation flux (MMC), and a net mass flux (NMF). For example, the SH term
can be expanded as:

\[
c_p \{[vT]\}_{SH} = c_p \{[v^*T^*]\}_{EF} + c_p \{[v]'[T]''\}_{MMC} + c_p \tilde{v} \tilde{T},
\]

where

\[
[b] \equiv \frac{1}{2\pi} \int_{0}^{2\pi} bd\lambda,
\]

\[
\{b\} \equiv \frac{1}{\int dp/g} \int bd\lambda,
\]

departure from zonal mean

\[
b^* \equiv b - [b],
\]

departure from zonal mean

\[
b'' \equiv b - \{b\},
\]

departure from mass weighted vertical average

and

\[
\tilde{b} \equiv \frac{1}{\int dp/g} \int bd\lambda.
\]

zonal mean of the mass weighted vertical average

The NMF is defined as the MSE brought into the polar cap via a net mass transport. The NMF has no vertical structure (i.e., all the information on the vertical structure of the MSE flux is contained in the EF and the MMC). The NMF is written in terms of the vertically and zonally averaged meridional wind ($\tilde{v}$). This is in contrast to Overland and Turet (1994), where the NMF term is $c_p \{[\overline{v}]\} \{[\overline{T}]\}$ (the overline denotes a time average). Our definition of the NMF ensures that a longitude with relatively high terrain contributes less to the NMF.

The NMF term has been removed from $F_{wall}$ calculations due to unphysical high-frequency noise. $\tilde{v}$ (proportional to the NMF) has been subtracted from $v$ for the entire dataset to ensure mass balance and remove the NMF. Fig. 2.1 shows the 3-hourly instantaneous polar cap–averaged surface pressure and the correction to the meridional wind ($v$) that removes the NMF for the period January–July 2000. If $\tilde{v}$ is a physical signal, then it should be well correlated with the surface pressure. A correlation of 0.48 is found for the year 2000 across 70°N at a lag of about 1 day. Also shown is the low-pass filtered $\tilde{v}$, with a cutoff timescale of 4 days. The correlation between the surface pressure and the filtered $\tilde{v}$
is 0.65 at a lag of 1 day. We then conclude that there is a decent amount of high-frequency noise associated with \( \tilde{v} \) and the NMF. This noise may be associated with interpolation of the data to regular pressure levels. This contrasts with the Overland and Turet (1994) method of assuming that for monthly time averages \( \{[\bar{v}]\} \approx 0 \), a method that would remove any low frequency signal while retaining the high-frequency signal. In addition, Liang et al. (2018) showed that a majority of the MSE brought to the polar caps through high-frequency net mass transport does not increase the average energy of the polar caps. The increase in the energy storage of the polar cap is exactly balanced by the added mass of air masses at the same energy as the polar cap.

The mean and variability of \( \tilde{v} \) implies a large contribution from the NMF to \( F_{\text{wall}} \) in the dataset. \( \tilde{v} \) at 70\(^\circ\)N for the entire dataset is 0.006 m s\(^{-1}\) and is -0.004 m s\(^{-1}\) at 65\(^\circ\)S, indicating a small poleward and equatorward flux of mass, respectively. Root-mean-square deviation (RMSD) between \( \tilde{v} \) and the corrected \( \tilde{v} \) (0 m s\(^{-1}\)) at 70\(^\circ\)N using the entire dataset is 0.04 m s\(^{-1}\). At 65\(^\circ\)S, the RMSD is also 0.04 m s\(^{-1}\). Consider the NMF component of the SH term for a vertically averaged temperature of 250 K at 70\(^\circ\)N. Although the mean (0.006 m s\(^{-1}\)) and variability (given an RMSD of 0.04 m s\(^{-1}\)) of \( \tilde{v} \) both appear small, they correspond to a NMF convergence mean and variability of 14 and 91 W m\(^{-2}\), respectively. The NMF would have a much smaller contribution to \( F_{\text{wall}} \) if it were instead defined relative to the vertically averaged temperature of the polar cap (e.g., no contribution from the NMF component of the SH term convergence to \( F_{\text{wall}} \) for a polar cap–averaged temperature equal to the temperature at 70\(^\circ\)N). This example points out both the sensitivity of \( F_{\text{wall}} \) to the definition of the NMF and the difficulty of physically interpreting this term (Mayer et al., 2019).

Applying a low-pass filter to \( \tilde{v} \) and, as suggested by Liang et al. (2018), defining the NMF relative to the average energy of the polar caps would lead to a stronger correlation between the NMF and climate signals (e.g., the polar cap–averaged temperature tendency).
2.3 Climatology & Variability

2.3.1 Vertical structure of the poleward MSE flux from 1980 to 2016

Fig. 2.2 shows the vertical structure of the monthly averaged poleward MSE flux across 70°N and 65°S from 1000 to 0.1 hPa for the entire period (1980–2016). We use pressure rather than height as the vertical coordinate in order to visualize contributions from each level to the vertical integral (total convergence). At both latitudes, the level of smallest variability and climatological magnitude is found near 300 hPa, which we define as the boundary between tropospheric and stratospheric fluxes.

Across 70°N, the climatological poleward MSE flux and its variability occurs primarily in two distinct and vertically separated locations in the mid-stratosphere and lower-mid troposphere. The climatological MSE flux in the stratosphere across 70°N and its variability is almost exclusively a wintertime phenomenon (cf. Fig. 2.3b and 2.3c). The variability of the stratospheric flux across 65°S is small compared to the stratospheric flux across 70°N. Across 65°S, the MSE flux peaks in the lower-troposphere (975–800 hPa) during winter. Interestingly, the seasonality of the MSE flux in the lower-troposphere is more pronounced at 65°S than its counterpart at 70°N despite weaker seasonality of both the magnitude and location of the storm track in the Southern Ocean (Trenberth, 1991); this issue is further explored in section 2.3.2.

2.3.2 Seasonality of the MSE flux

Annual, winter, and summer means of the MSE flux and its components are shown in Fig. 2.3. For all results, we use the November–March (NDJFM) winter season and June–August (JJA) summer season in the Northern Hemisphere and the May–September (MJJAS) winter season and December–February (DJF) summer season in the Southern Hemisphere. We use a 5-month winter season for direct comparisons with Overland and Turet (1994) and to identify all sudden stratospheric warming (SSW) events (Polvani and Waugh, 2004). Across 70°N, local poleward maxima in the annual and winter mean MSE flux is located around 30 hPa and in the broad region of the lower- and mid-troposphere. During the summer, the tropospheric maximum is closer to the surface and the maximum in the stratosphere is an order of magnitude smaller in value and closer to the tropopause.
Across 65°S, local poleward maxima in the annual and winter mean MSE flux are located around 150 hPa and around 950 hPa. Evidence of a clean separation between tropospheric and stratospheric fluxes are clearly shown, especially during winter, by the minimum in flux magnitude and standard deviation near 300 hPa.

Flux values across 70°N and 65°S are directly compared in units of J kg⁻¹ m s⁻¹, with units of W m⁻² (100 hPa)⁻¹, the local polar cap convergence, used when comparing their impact on the climate of the polar regions. The most apparent difference between the hemispheres is in the stratosphere, where the flux convergence at 70°N is much larger than at 65°S both in the annual and winter mean (about 4 times larger in the winter mean). The seasonality of the tropospheric flux across 70°N is much weaker than across 65°S (Fig. 2.4), especially in the lower-troposphere (Fig. 2.3). Part of this difference can be explained by the trade-off between SH and LH fluxes, which are out of phase at 70°N (cf. green and cyan lines in Fig. 2.4) but are in phase at 65°S.

These results are consistent with Overland and Turet (1994) for the flux across 70°N. The vertical structure of the MSE flux is generally in agreement, except for the magnitude of the summer stratospheric flux. During the summer mean, Overland and Turet (1994) showed a maximum of approximately 10 W m⁻² (100 hPa)⁻¹ in the MSE flux at 50 hPa, the top-level of the GFDL dataset, while the flux is near 0 throughout much of the stratosphere using the MERRA-2 dataset, with a small local maximum around 200 hPa. The stratospheric maximum during the winter in the GFDL dataset is also slightly larger.

2.3.3 Stratospheric contribution to $F_{\text{wall}}$

The mean annual cycle of $F_{\text{wall}}$ and contributions from the stratosphere ($F_{\text{strat}}$), troposphere ($F_{\text{trop}}$), LH, SH, and GP fluxes are shown in Fig. 2.4, with climatological values at 70°N and 65°S included in Table 2.1 and Table 2.2, respectively. In this chapter, $F_{\text{wall}}$ is expressed as the polar cap–averaged MSE flux convergence (in W m⁻²) calculated as the zonally and vertically integrated poleward flux at 65°S and 70°N.

$F_{\text{strat}}$ contributions to $F_{\text{wall}}$, expressed in both W m⁻² and as a percentage, in the annual and winter mean are larger at 70°N than at 65°S. In the annual mean, $F_{\text{strat}}$ at 70°N is 14.4 W m⁻² or 15% of $F_{\text{wall}}$. $F_{\text{strat}}$ is largest during the winter, with a mean of 20.9 W m⁻² or 19% of $F_{\text{wall}}$ and weakest during the summer, with a mean of 6.0 W m⁻² or 7% of $F_{\text{wall}}$. $F_{\text{strat}}$ at
65°S is 9.7 W m\(^{-2}\) or 11% of \(F_{\text{wall}}\) in the annual mean, 7.1 W m\(^{-2}\) or 6% of \(F_{\text{wall}}\) in the winter mean, and 9.1 W m\(^{-2}\) or 15% of \(F_{\text{wall}}\) in the summer mean. The winter and summer seasons at 65°S include local minima in the annual cycle, with local maxima occurring in March and October (Fig. 2.4e). This is consistent with increased magnitude of stratospheric stationary waves associated with October final warming events in the Southern Hemisphere.

\(F_{\text{strat}}\) contributions to \(F_{\text{wall}}\) seasonality are larger at 70°N than at 65°S. The seasonal cycles in \(F_{\text{strat}}\) and \(F_{\text{trop}}\) at 70°N are generally in phase. The seasonal range, defined as the maximum minus minimum monthly flux convergence derived from Fig. 2.4, is larger in \(F_{\text{strat}}\) (22 W m\(^{-2}\)) than in \(F_{\text{trop}}\) (18 W m\(^{-2}\)). Thus, at 70°N, \(F_{\text{strat}}\) contributes more to the seasonal range of \(F_{\text{wall}}\). The SH component has the largest annual cycle in \(F_{\text{wall}}\), which is best explained by its seasonality in the stratosphere (opposed in part by the GP component). The seasonal range of \(F_{\text{wall}}\) at 70°N is 32 W m\(^{-2}\) smaller than at 65°S. In general, the seasonal cycles in \(F_{\text{strat}}\) and \(F_{\text{trop}}\) at 65°S are out of phase. At 65°S, the seasonality of \(F_{\text{wall}}\) is dominated by \(F_{\text{trop}}\), as there is little seasonality in \(F_{\text{strat}}\), except in the largely opposed SH and GP components.

### 2.3.4 Variability of the stratospheric contribution to \(F_{\text{wall}}\)

Fig. 2.5 shows the time series of monthly mean \(F_{\text{wall}}\), \(F_{\text{strat}}\), and \(F_{\text{trop}}\). Inter-annual variability in \(F_{\text{wall}}\) is larger at 70°N than at 65°S. \(F_{\text{wall}}\) is largest in both magnitude and variability during their respective winters, this can also be seen in Fig. 2.3 and in Fig. 2.4.

Fig. 2.6 shows the same monthly time series, but expressed as anomalies relative to the mean annual cycle from Fig. 2.4. \(F_{\text{strat}}\) variability is larger in the Northern Hemisphere and explains more of the variance of \(F_{\text{wall}}\) than in the Southern Hemisphere. At 70°N, the Pearson correlation between \(F_{\text{strat}}\) and \(F_{\text{trop}}\) monthly mean anomalies is approximately zero during all seasons. However, the correlation between \(F_{\text{strat}}\) and \(F_{\text{wall}}\) anomalies is +0.48. In other words, \(F_{\text{strat}}\), distinct from \(F_{\text{trop}}\), explains 23% of the variance of \(F_{\text{wall}}\) at 70°N. During the summer, the correlation between \(F_{\text{wall}}\) and \(F_{\text{strat}}\) decreases to +0.33 (11% of the variance), and during the winter, the correlation increases to +0.52 (27% of the variance). In contrast, at 65°S, \(F_{\text{strat}}\) only explains 10% of the variance of \(F_{\text{wall}}\). Similar results are found at 70°N when using daily as opposed to monthly anomalies. These results show the importance of \(F_{\text{strat}}\) to \(F_{\text{wall}}\) variability at 70°N and a lesser degree of importance at 65°S.
Section 2.4 will then focus on the variability of $F_{\text{strat}}$ and $F_{\text{trop}}$ at 70°N.

2.4 Link to the Arctic climate

2.4.1 Climate impacts of $F_{\text{strat}}$ and $F_{\text{trop}}$ anomalies

Given the vertical separation and temporal orthogonality of $F_{\text{trop}}$ and $F_{\text{strat}}$ anomalies seen in section 2.3.4, we now ask if $F_{\text{trop}}$ and $F_{\text{strat}}$ anomalies have distinct climate impacts on the troposphere and stratosphere, respectively. To accomplish this task we analyze the signature of $F_{\text{wall}}$ and its partitioning between $F_{\text{trop}}$ and $F_{\text{strat}}$ across composites of two different climate events: 1) sudden stratospheric warmings (SSWs) and 2) the atmospheric forcing of polar surface heating which we will quantify from downward surface flux events (DSFEs).

2.4.1.1 Definition of Sudden Stratospheric Warmings and Downward Surface Flux Events

An SSW is defined as the first day on which the 60°N or 60°S $[u]_{10 \text{hPa}}$ reverses from westerly to easterly during the winter (NDJFM in the Northern Hemisphere and MJJAS in the Southern Hemisphere). Additionally, $[u]_{10 \text{mb}}$ must return to westerly for at least 20 consecutive days between events. Table 2.3 lists these events, where the event date is defined as the central date (day of wind reversal). This definition of SSWs follows Charlton and Polvani (2007) except that, in this chapter, we include some early final warming events. These events are included because their dynamics are similar to mid-winter SSWs; early final warmings tend to be strongly wave driven, and thus associated with greater heat flux than climatological or late final warmings (Butler et al., 2019). The event on February 6 1995 is also not included in the MERRA-2 component of the SSW Compendium (Butler et al., 2017; Molod et al., 2015). This is likely due to only 1 day of easterlies during the event.

We expect that $F_{\text{strat}}$ is anomalously poleward prior to the central date of an SSW and is preceded by poleward anomalies in $F_{\text{trop}}$. Polvani and Waugh (2004) showed that the 40-day period prior to the central date of SSWs is associated with anomalously strong poleward meridional eddy heat fluxes, averaged over the 40-day period, at 100 hPa. The meridional eddy heat flux at 100 hPa, which is averaged between 45°N and 75°N in Polvani
and Waugh (2004), is proportional to the eddy component of the SH term in $F_{wall}$. The meridional eddy heat flux is also proportional to the vertical component of the planetary wave activity flux (e.g., Edmon et al., 1980), with origins in the troposphere (e.g., Matsumo, 1971; Polvani and Waugh, 2004). The expected poleward anomalies in $F_{trop}$ are associated with the tropospheric origin of SSWs, consistent with the weak but non-zero lagged correlation between monthly $F_{trop}$ and $F_{strat}$ anomalies, with $F_{trop}$ leading by 1-month (+0.31). However, a near tropopause-level (lower-stratospheric) planetary wave source may also play a role in the development of SSWs (Boljka and Birner, 2020). The associated heat flux from a lower-stratospheric planetary wave source may not be well captured by $F_{trop}$.

A DSFE is defined as the first day on which the surface downward energy flux averaged over the polar cap exceeds the 95th percentile threshold for the 5-month winter climatology. The net surface flux includes sensible and latent heat fluxes, the net longwave flux, and the absorbed shortwave flux, which is negligible during the winter. The central date of an event is defined as the day of the downward surface flux maximum. A 7-day rolling mean was applied to the surface flux data (3-hourly) to ensure that multiple maxima are not selected for one event. The linear trend was also removed from the surface flux data to ensure that events were selected over the entire dataset (red tick marks in Fig. 2.5 and Fig. 2.6). The 95th percentile threshold was chosen so that there are approximately the same number DSFEs (34) as SSWs (32).

We expect that $F_{trop}$ is anomalously poleward prior to the central date of a DSFE. A downward surface flux indicates that the surface is warming at the expense of the atmosphere. During the winter, this is a combination of an increased downward longwave flux and suppression of upward sensible and latent heat fluxes, which is expected to be preceded by an increase in polar cap–averaged sensible and latent heat (moist enthalpy) in the troposphere.

### 2.4.1.2 Composite analysis of SSWs and DSFEs

Table 2.4 shows composites of $F_{wall}$, $F_{strat}$, and $F_{trop}$ in the 30-day mean before and after an SSW and DSFE. Both the raw flux convergence and anomalous flux convergence are provided to emphasize contributions to $F_{wall}$. We note that there is a small contribution (1–3 W m$^{-2}$) from the climatological seasonal cycle to the change in the raw flux convergence before and after an event. SSWs tend to occur later in the winter, while DSFEs tend to
occur earlier in the winter; thus, there is a slight climatological decrease and increase in the flux convergence during SSWs and DSFEs, respectively. Prior to an SSW, there are statistically significant poleward anomalies in $F_{\text{wall}}$ (12.2 W m$^{-2}$) primarily due to anomalies in $F_{\text{strat}}$ (8.8 W m$^{-2}$). After the central date, statistically significant equatorward anomalies in $F_{\text{wall}}$ and $F_{\text{strat}}$ are found, possibly reflecting the lagged relationship between the stratosphere and troposphere during an SSW. Prior to a DSFE, there are statistically significant poleward anomalies in $F_{\text{wall}}$ (7.6 W m$^{-2}$) primarily due to anomalies in $F_{\text{trop}}$ (5.8 W m$^{-2}$). In other words, both types of events are preceded by significant poleward $F_{\text{wall}}$ anomalies, with the anomalous heating located in the stratosphere and troposphere for SSWs and DSFEs, respectively.

In order to more precisely investigate the temporal evolution of the energy flux during SSWs and DSFEs, a composite of the daily mean $F_{\text{wall}}$, $F_{\text{strat}}$, and $F_{\text{trop}}$ in the 30 days before and after these events are computed and are shown in Fig. 2.7a and c. The evolution of the stratosphere during SSWs includes two distinct periods associated with the weakening (breakdown) and strengthening (recovery) of the polar vortex centered around the central date. The period 30 days before and after the central date adequately captures the typical timescale of the deceleration of the zonal mean zonal winds (weakening) and the subsequent recovery of the polar vortex (black line in Fig. 2.7a). The observed increase in $F_{\text{strat}}$ is largest in the 8 days prior to the central date. In that 8-day mean, the corresponding poleward $F_{\text{strat}}$ anomaly is 25.2 W m$^{-2}$. The maximum anomaly in $F_{\text{strat}}$ (37.2 W m$^{-2}$) on day -3 is preceded by a maximum anomaly in $F_{\text{trop}}$ (20.5 W m$^{-2}$) on day -7. After the central date, $F_{\text{strat}}$ returns to near climatology (cf. solid and dashed lines in Fig. 2.7a). This reduction is consistent with the decrease in meridional eddy heat flux anomalies after SSW events (e.g., Butler et al., 2017).

Fig. 2.7b shows the vertical structure of the MSE flux convergence contributing to $F_{\text{wall}}$ during an SSW. Compared to the winter climatology, anomalous poleward fluxes in the 8-day mean prior to the central date are found in the entire stratosphere. The maximum in the mid- to upper-stratosphere is significant prior to the SSW, with a relatively smaller increase with respect to the winter climatology in the mid-troposphere. After the central date, much of the MSE flux in the column reduces to less than the winter climatology, with exception in the lower-troposphere. The composite of the 30-day mean after the SSW includes some dates
in April (not included in the winter climatology), which is associated with the downward progression of the MSE flux convergence maximum and a climatological increase in the lower-tropospheric MSE flux convergence (Fig. 2.3). However, this lower-tropospheric increase is still anomalous with respect to the mean annual cycle.

A composite of DSFEs show the importance of $F_{\text{trop}}$ anomalies in initiating these events (Fig. 2.7c). The observed increase in $F_{\text{trop}}$ is largest in the 8 days prior to the central date, where a mean poleward anomaly of 36.5 W m$^{-2}$ and a maximum anomaly of 47.9 W m$^{-2}$ were found. After the central date, $F_{\text{wall}}$ and $F_{\text{trop}}$ reduce to approximately climatological levels. During the entire period, the anomalies in $F_{\text{strat}}$ are not significantly different from zero.

Fig. 2.7d shows the vertical structure of the MSE flux convergence during a DSFE. In the mean 8-day period prior to the central date of a DSFE, anomalous poleward fluxes compared to the winter climatology are found throughout the entire troposphere and are maximized in the lower-troposphere. As suggested by the vertical structure, these events are associated with statistically significant poleward LH flux anomalies. In the 8-day mean prior to the central date, the contribution to the anomalous $F_{\text{trop}}$ from the LH anomaly is 6.6 W m$^{-2}$ (not shown). This is consistent with anomalous downward surface energy fluxes preceded by intense moisture flux events (Woods and Caballero, 2016).

2.4.2 Composite analysis of the Arctic response to SSWs and DSFEs

In section 2.4.1 we analyzed the signature of $F_{\text{wall}}$ and its partitioning into $F_{\text{strat}}$ and $F_{\text{trop}}$ across two different types of climate events: SSWs and DSFEs. This section focuses on the Arctic response to these two types of events.

Fig. 2.8 shows composites of the anomalous energy budget of the Arctic climate system over SSWs (Fig. 2.8a) and DSFEs (Fig. 2.8b). The terms in the budget are cumulative time integrals of anomalous $F_{\text{wall}}$, $F_{\text{trop}}$, and $F_{\text{strat}}$ and polar cap–averaged cumulative time integrals of anomalous moist enthalpy ($h_m$) tendency in the atmosphere (i.e., $h_m$ storage—which is subdivided into stratospheric and tropospheric components), outgoing longwave radiation (OLR), and net surface flux (NSF) in MJ m$^{-2}$. The cumulative integration allows for easier visualization and starts 20 days before the central date of the events, the approximate date when $F_{\text{wall}}$ becomes anomalously poleward for an extended period. In Fig. 2.8, an increasing
cumulative anomaly of a term indicates that there are positive anomalies of that term, with the slope indicating the magnitude of the anomaly on a particular lag day. Linear trends were removed for all anomalies in these composites.

In the SSW composite, first, there is a cumulative poleward increase in anomalous $F_{\text{strat}}$ and associated stratospheric $h_m$ storage. The increase in stratospheric $h_m$ storage slows and subsequently decreases, indicating a gradual cooling. This stratospheric cooling is accompanied by a gradual increase in the cumulative OLR anomaly, with little change in the NSF. Cumulative $F_{\text{strat}}$ is nearly balanced by the sum of $h_m$ and OLR suggesting that the energy input by anomalous poleward $F_{\text{strat}}$ during an SSW acts to increase the stratospheric $h_m$ storage and OLR, with little influence on the surface. The total anomalous energy budget of the Arctic is not necessarily constrained in our analysis since all terms are calculated independently. However, the sum of all terms nearly balance in both the stratosphere and troposphere (see Fig. B1), suggesting our analysis conserves energy in the column average, troposphere, and stratosphere. The NSF even becomes weakly anomalously upward (negative) after the central date, a response to equatorward $F_{\text{trop}}$ anomalies. These results are fairly consistent with the SSW life cycle as explored by Limpasuvan et al. (2004). During the SSW life cycle, poleward heat flux anomalies found in the troposphere and stratosphere during the breakdown of the polar vortex are followed by equatorward heat flux anomalies in the troposphere during the recovery of the polar vortex.

In the DSFE composite (Fig. 2.8b), first, there is a cumulative increase in the anomalous $F_{\text{trop}}$ and associated tropospheric $h_m$ storage. The increase in tropospheric $h_m$ storage slightly precedes the increase in $F_{\text{trop}}$ due to an anomalously upward NSF contributing energy to the troposphere at the beginning of the period. While the increasing tropospheric $h_m$ storage anomaly slows and subsequently decreases (cooling and drying), there is a cumulative downward (positive) increase in the NSF anomaly. This composite shows that the energy input from $F_{\text{wall}}$ primarily heats the atmosphere preceding the DSFE and this energy is subsequently fluxed downward from the warmed atmosphere to the surface. After the event is over, $F_{\text{wall}}$ has returned to climatology and the anomalous $F_{\text{wall}}$ over the duration of the event, almost entirely due to $F_{\text{trop}}$, has primarily gone into the surface (accumulated NSF), secondarily increased the energy content of the atmosphere, and only a small portion has been radiated back to space (little response in the OLR).
The composite analysis shown in Fig. 2.8 suggests that the Arctic surface climate is more sensitive to $F_{\text{trop}}$ than $F_{\text{strat}}$ variability. In other words, $F_{\text{trop}}$ is more efficient at warming the surface than $F_{\text{strat}}$. A schematic of the response to an anomalous increase in $F_{\text{strat}}$ and $F_{\text{trop}}$ is shown in Fig. 3.10. Although both events are associated with a similar increase in $F_{\text{wall}}$, the NSF term only shows a large anomalous response when $F_{\text{wall}}$ is dominated by $F_{\text{trop}}$. Although not reflected in the surface energy budget of the Arctic, these results do not suggest that SSWs have no climatic impact as surface impacts can result from dynamical stratosphere–troposphere coupling (Kidston et al., 2015). In addition, individual SSW events might show a more pronounced lagged relationship between $F_{\text{strat}}$, $F_{\text{trop}}$, and the NSF.

Fig. 2.8 also includes the energy budget residual in the total, troposphere, and stratosphere in MJ m$^{-2}$. The total residual is the difference between the energy input ($F_{\text{wall}}$), and Arctic response terms: NSF, OLR, and $h_m$ storage, with positive values indicating excess $F_{\text{wall}}$. The total residual could result from interpolation error, neglecting the contribution to $F_{\text{wall}}$ from the climatically relevant part of the net mass flux (NMF), or energy imbalances in the underlying MERRA-2 data. The total residual indicates a slight excess of $F_{\text{wall}}$ in the SSW composite (Fig. 2.8a) and, for most of the period, a slight deficit of $F_{\text{wall}}$ in the DSFE composite (Fig. 2.8b).

The approximate budget closure in the troposphere and stratosphere suggests that the energy exchanges between stratosphere and troposphere within the polar cap are relatively small. The stratospheric residual is the difference between $F_{\text{strat}}$ and combined OLR and stratospheric $h_m$ storage terms, while the tropospheric residual is the difference between $F_{\text{trop}}$ and combined NSF and tropospheric $h_m$ storage terms. These residuals provide an estimate of the vertical exchange of energy across the tropopause. In the SSW composite, the stratospheric residual gradually increases following poleward anomalies in $F_{\text{strat}}$, suggesting a flux of energy from the stratosphere to the troposphere. In the DSFE composite, the tropospheric residual increases, especially in days -5 to +5, following poleward anomalies in $F_{\text{trop}}$, suggesting a flux of energy from the troposphere to the stratosphere. The energy exchange across the tropopause appears to be larger during DSFEs. However, the exchange is small relative to the magnitude of $F_{\text{wall}}$ and the dominant Arctic response to $F_{\text{wall}}$ during SSWs and DSFEs. This justifies the simple two-box interpretation of the energy budget sketched in Fig. 3.10. A similar small vertical exchange of energy across the tropopause
is found in other stratospheric events described in Dunn-Sigouin and Shaw (2015): strong vortex and extreme heat flux events (not shown).

2.4.3 Metric for the influence of atmospheric Circulations on the Arctic surface climate

Results thus far have shown that $F_{\text{strat}}$ is an important contributor to $F_{\text{wall}}$ variability into the Arctic and that a poleward increase in $F_{\text{strat}}$ does not result in increased area-averaged heat flux to the Arctic surface. $F_{\text{trop}}$ should then be a better metric for the influence of atmospheric circulations on the Arctic surface climate than $F_{\text{wall}}$, especially during the winter when $F_{\text{strat}}$ variability is largest. Fig. 2.10 shows correlations between the lower-tropospheric (1000–900 hPa) polar cap-averaged $h_m$ tendency, $F_{\text{wall}}$, and $F_{\text{trop}}$. Correlations are plotted with respect to rolling means, applied to all data, up to 30 days. Correlations between $F_{\text{trop}}$ and the $h_m$ tendency are indeed larger than the correlations with $F_{\text{wall}}$, especially during the winter. Thus, $F_{\text{trop}}$ explains a larger proportion of the variance of the $h_m$ tendency. This result is quantitatively similar to the proportion of $F_{\text{wall}}$ variance explained by $F_{\text{strat}}$ provided in section 2.3.4.

During the 1980–2016 period considered in this data, the maximum correlation for any rolling mean window is larger when using $F_{\text{trop}}$ as opposed to $F_{\text{wall}}$. For the full dataset, the maximum correlation increases from 0.64 to 0.75, during the winter the maximum increases from 0.64 to 0.76, and during the summer the maximum increases from 0.64 to 0.71. During the summer, the maximum correlation between $F_{\text{trop}}$ and the $h_m$ tendency occur at a later rolling mean (11 days) than the winter (7 days) and correlations remain relatively high for longer timescales. One possible explanation for this result is that the ice albedo feedback results in a longer timescale of atmospheric response as the sea ice melts and additional solar energy is added to the Arctic climate system. During the winter, the coefficient of determination (correlation squared) increases from 0.41 to 0.58. Therefore, $F_{\text{wall}}$ explains 17% more of the variance of the $h_m$ tendency when $F_{\text{strat}}$ is filtered out.
2.5 Summary and discussion

In this analysis, the vertical structure of the poleward moist static energy (MSE) flux in the MERRA-2 across 70°N and 65°S was examined. This chapter sought to quantify the stratospheric (\(F_{\text{strat}}\)) and tropospheric (\(F_{\text{trop}}\)) contributions to \(F_{\text{wall}}\), and the Arctic response following events of significant increases in \(F_{\text{strat}}\) and \(F_{\text{trop}}\). In both hemispheres, local maxima in magnitude and variability of the poleward MSE flux are found at two vertically distinct locations in the mid- to upper-stratosphere and mid- to lower-troposphere, with a minimum near the tropopause. \(F_{\text{wall}}\) is separated into distinctly stratospheric and tropospheric components that have temporally uncorrelated anomalies. \(F_{\text{strat}}\) was found to be non-negligible, especially at 70°N during winter (NDJFM), where \(F_{\text{strat}}\) contributes 19% of the climatological \(F_{\text{wall}}\). \(F_{\text{strat}}\) distinctly explains 23% of the variance of \(F_{\text{wall}}\) when using monthly mean anomalies; this value provides an estimate of how much \(F_{\text{strat}}\) biases the part of \(F_{\text{wall}}\) that is relevant to the Arctic surface climate.

Motivated by the greater importance of \(F_{\text{strat}}\) variability to \(F_{\text{wall}}\) at 70°N, we focussed on the Arctic and argued that \(F_{\text{trop}}\) and \(F_{\text{strat}}\) have different impacts on the climate system, with \(F_{\text{trop}}\) associated with energy input to the surface of the Arctic and \(F_{\text{strat}}\) associated with sudden stratospheric warmings (SSWs). Fig. 3.10 provides a visual summary of the Arctic response to poleward anomalies in \(F_{\text{trop}}\) and \(F_{\text{strat}}\). In the 20 days preceding an SSW, significant poleward \(F_{\text{strat}}\) anomalies lead to stratospheric warming, with the majority of the \(F_{\text{strat}}\) anomaly going into stratospheric sensible energy storage, approximately one-third of the energy input radiated to space, and little change in the net surface flux. During winters with early (December–January) SSWs, Kuttippurath and Nikulin (2012) found minimal wintertime stratospheric ozone loss (i.e., increased ozone concentrations). This increase in ozone concentrations is associated with increased atmospheric emissivity, which may play a role in the increased OLR along with the warmer temperatures, though we have not attempted to separate these signals.

In the 15 days preceding a downward surface flux event in the Arctic (DSFE), significant poleward \(F_{\text{trop}}\) anomalies lead to the heating and moistening of the atmospheric column. Thereafter, \(F_{\text{trop}}\) anomalies are not sustained and the warmed atmosphere fluxes energy downward with the net effect of the event being a near balance of the time integrated \(F_{\text{wall}}\) anomaly and surface energy anomaly. Removing \(F_{\text{strat}}\) variability from \(F_{\text{wall}}\) resulted
in an increased correlation between $F_{\text{wall}}$ and the lower-tropospheric $h_m$ (sensible and latent energy) tendency. Therefore, the efficiency with which poleward anomalies in $F_{\text{wall}}$ heat the Arctic surface is increased during periods dominated by tropospheric anomalies. For a given poleward $F_{\text{trop}}$ anomaly, the surface heating efficiency is expected to increase with pressure (lower-tropospheric heating) and with the contribution from the LH component. LH flux convergence is associated with both atmospheric heating and moistening (increased atmospheric emissivity), and thus an increased downward longwave flux to the surface (Graversen and Burtu, 2016).

Our results suggest that, composited over many events, $F_{\text{strat}}$ and $F_{\text{trop}}$ variability is distinct (temporally orthogonal) and primarily impact the stratosphere (SSWs) and surface (DSFEs), respectively. These results, however, do not rule out the importance of troposphere–stratosphere interactions for individual events. There may exist events that, similar to events described in Baggett and Lee (2017), are associated with both large poleward anomalies in $F_{\text{wall}}$ dominated by $F_{\text{trop}}$ and high planetary wave activity. As a result of the increase in planetary wave activity, these events may be associated with a larger vertical exchange of energy across the tropopause than DSFEs and may precede some SSW events, which is consistent with the dynamical theory of SSWs. These events would likely be less efficient at heating the surface than DSFEs as a result of the larger troposphere to stratosphere energy flux. To better understand the relationship between the vertical structure of the MSE flux convergence on the surface heating efficiency, Chapter 3 will seek to determine drivers of the surface heating efficiency in events associated with an increase in $F_{\text{trop}}$.

We speculate that changes to the vertical structure of $F_{\text{wall}}$ in a warmer climate will change the surface heating efficiency of $F_{\text{wall}}$ in the Arctic due either to changes in $F_{\text{strat}}$ or $F_{\text{trop}}$. Recent work has looked at changes in SSWs in transient climate change simulations, which could impact variability and trends in $F_{\text{strat}}$. Ayarzaguena et al. (2018) found no statistically significant changes in SSW frequency or duration by the end of the 21st century, across 12 Chemistry–Climate Model Initiative (CCMI) models. This result suggests that robust changes in the surface heating efficiency of $F_{\text{wall}}$ will be linked to the troposphere. Comprehensive climate models project increased moisture transport (i.e., LH flux) into the Arctic under future global warming (Hwang and Frierson, 2010), which is thought to be an important driver of polar amplification of climate change (e.g., Alexeev and Jackson, 2013).
Changes in total $F_{\text{wall}}$ are anti-correlated with the amount of polar amplification, with the decrease in the flux of dry static energy dominating the inter-model spread in $F_{\text{wall}}$ (Hwang et al., 2011). We speculate that these changes are associated with a downward shift toward a more tropospheric-weighted $F_{\text{wall}}$ (since the LH component is all tropospheric, e.g. Fig. 2.4), which would increase the surface heating efficiency of $F_{\text{wall}}$. It is possible that this downward shift overwhelms any effects of changes in total $F_{\text{wall}}$, and this may be an under-appreciated mechanism for polar amplification.
2.6 Tables

Table 2.1: Climatological values of the vertically integrated polar cap–averaged moist static energy (MSE) flux convergence ($F_{\text{wall}}$) and contributions from the stratosphere ($F_{\text{strat}}$; 300–0.1 hPa) and troposphere ($F_{\text{trop}}$; 1000–300 hPa) for annual and seasonal means in W m$^{-2}$ at 70°N. Also provided are contributions from the sensible heat (SH), latent heat (LH), and geopotential (GP) flux convergence.

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Table 2.2: As in Table 2.1, but for 65°S.

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Table 2.3: Dates of Sudden Stratospheric Warmings or Weak Vortex Events. Dates of late-winter warmings not included in the MERRA-2 component of the SSW Compendium are denoted by *.

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<td>2016</td>
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<td>06*</td>
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Table 2.4: Composite of the MSE flux convergence and associated MSE flux convergence anomalies in W m\(^{-2}\) at 70°N averaged in the 30 days before and after the central date of an SSW and downward surface flux event (DSFE). * indicates anomalies significantly different from zero at the 95% confidence level. A two-sided t-test was used to determine significance. For p-values < 0.05, we reject the null hypothesis of equal averages.

<table>
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<td>(F_{\text{trop}})</td>
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<tr>
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<td></td>
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<tr>
<td>(F_{\text{strat}})</td>
<td>18.7</td>
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<td>(F_{\text{trop}})</td>
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<td>(F_{\text{wall}})</td>
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2.7 Figures

Figure 2.1: 3-hourly instantaneous polar cap–averaged surface pressure (blue) in hPa, $\tilde{v}$ at 70° N (light green), and low-pass filtered $\tilde{v}$ with a cutoff timescale of 4 days (green) between January 2000 and July 2000.
Figure 2.2: Time-pressure series of monthly mean moist static energy flux in J kg$^{-1}$ m s$^{-1}$ and local moist static energy flux convergence in W m$^{-2}$ (100 hPa)$^{-1}$ across (a) 70°N and (b) 65°S with positive defined as a poleward flux.
Figure 2.3: Annual, winter, and summer mean local MSE flux convergence (red) in W m$^{-2}$ (100 hPa)$^{-1}$ (lower axis), MSE flux (red) in J kg$^{-1}$ m s$^{-1}$ (upper axis), and standard deviation (light red fill) across (a,b,c) 70°N and (d,e,f) 65°S with positive defined as a poleward flux. Contributions from the latent heat flux (LH; cyan), sensible heat flux (SH; green), and geopotential flux (GP; blue) are shown.
Figure 2.4: Mean annual cycle of the vertically integrated, monthly and polar cap–averaged flux convergence ($F_{\text{wall}}$; red), and contributions from the stratosphere ($F_{\text{strat}}$; dotted red) and troposphere ($F_{\text{trop}}$; dashed red), and standard deviation (light red fill) at (a,b,c) 70°N and (d,e,f) 65°S in W m$^{-2}$. Contributions from LH (cyan), SH (green), and GP (blue) are shown.
Figure 2.5: Monthly mean $F_{\text{wall}}$ (red), $F_{\text{strat}}$ (green), and $F_{\text{trop}}$ (blue) at (a) $70^\circ$N and (b) $65^\circ$S in W m$^{-2}$. Sudden stratospheric warmings (SSWs) are denoted by black tick marks, and downward surface flux events (DSFEs) for the Northern Hemisphere are denoted by red tick marks.
Figure 2.6: Monthly mean anomalies in $F_{\text{wall}}$ (red), $F_{\text{strat}}$ (green), and $F_{\text{trop}}$ (blue) at (a) 70°N and (b) 65°S in W m$^{-2}$. SSWs and DSFEs are denoted by black and red tick marks, respectively.
Figure 2.7: (a) Composite of daily mean $F_{\text{wall}}$ (red), $F_{\text{strat}}$ (green), $F_{\text{wall}}$ (blue), and associated winter climatologies (dashed) during SSWs or weak vortex events (32) at 70°N. Also shown is the composite of the daily mean $|u|_{10\text{mb}}$ across 60°N (black). (b) Composite of the total MSE flux in the 8-day mean prior to the SSW central date (red solid), in the 30-day mean after the SSW central date (red dashed), the winter climatology of the total MSE flux (black) across 70°N in J kg$^{-1}$ m s$^{-1}$, and the corresponding local flux convergence in W m$^{-2}$ (100 hPa)$^{-1}$. (c) As in (a), but for DSFEs (34). Also shown is a composite of the linearly detrended anomalous polar cap surface flux (positive downward; black). (d) As in (b), but for DSFEs.
Figure 2.8: (a) Composite of the cumulative time integral of anomalous $F_{\text{wall}}$ (red), $F_{\text{strat}}$ (green dashed), $F_{\text{trop}}$ (blue dashed), outgoing longwave radiation (OLR; black dashed; positive upward), net surface flux (NSF; black dashed; positive downward), tropospheric moist enthalpy tendency ($h_m^{\text{storage}}$; blue dashed) and stratospheric moist enthalpy tendency ($h_m^{\text{storage}}$; green dashed) during SSWs in MJ m$^{-2}$. Also included are energy budget residuals in the total (dotted red), troposphere (dotted blue), and stratosphere (dotted green). (b) As in (a), but for DSFEs. Anomalies are linearly detrended.
Figure 2.9: Schematic of the response to an increase in the stratospheric ($F_{\text{strat}}$) and tropospheric ($F_{\text{trop}}$) flux convergence in the Arctic polar cap poleward of 70°N. The response to an increase in $F_{\text{strat}}$ (green) is an increase in the stratospheric sensible energy storage, followed by an increase in outgoing longwave radiation (OLR). The response to an increase in $F_{\text{trop}}$ (blue) is an increase in the tropospheric sensible and latent energy storage, followed by an increase in the downward net surface flux (NSF). Note that there is a relatively small vertical exchange across the tropopause.
Figure 2.10: Correlation between polar cap and lower-tropospheric (1000–900 hPa) averaged moist enthalpy ($h_m$) tendency, $F_{wall}$ (dashed), and $F_{trop}$ (solid), in the full dataset (black), winter (blue), and summer (red). Rolling means are applied to both the $h_m$ tendency and energy convergence.
CHAPTER 3

The Arctic Surface Heating Efficiency of Tropospheric Energy Flux Events

3.1 Introduction

This chapter has been submitted for publication in JCLI and has received a “minor revision” decision. Minor changes to the format have been made from the submitted manuscript. In Chapter 2, the Arctic response to increases in the polar cap–averaged tropospheric ($F_{\text{trop}}$) and stratospheric ($F_{\text{strat}}$) MSE flux convergence and found that a similar magnitude increase in $F_{\text{trop}}$ resulted in a greater surface response. Chapter 2 speculated that the surface heating efficiency of $F_{\text{trop}}$ is driven by its vertical structure and contribution from the latent component. This chapter further examines the drivers of the surface heating efficiency.

Observations over the past several decades indicate a bottom-amplified warming of the Arctic, with surface air temperatures increasing at a faster rate than the global average, especially during the cold season (Screen and Simmonds, 2010a; Cohen et al., 2014). Winter sea ice has also declined in recent decades, especially in the Barents–Kara Seas where relatively large reductions in sea ice concentration (SIC; D.-S. R. Park et al., 2015) and thickness (King et al., 2017) have been observed. Although there exists a strong negative correlation between winter sea ice growth and November sea ice thickness (i.e., thin ice grows thermodynamically faster than thick ice), the influence of atmospheric forcing on the thinning sea ice has been increasing (Stroeve et al., 2018). It has been suggested that increased surface downward longwave radiation (DLR) is an important driver—perhaps the main driver—of trends in surface temperature (e.g., Woods and Caballero, 2016; Gong et al., 2017; Lee et al., 2017) and sea ice (e.g., D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Woods and Caballero, 2016), while increased ocean-to-atmosphere fluxes (i.e., the ice-albedo or insulation feedback; Screen and Simmonds, 2010b; Stroeve et al., 2012; Burt et al., 2016) play a minor role. The DLR trend has been attributed to an increased frequency of intense tropospheric heat and moisture fluxes (i.e., moist intrusion events; Woods and Caballero, 2016) into the
Arctic (D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Woods and Caballero, 2016; Gong et al., 2017; Lee et al., 2017). A recent study showed that in a mean climatic state, DLR is largely determined by surface temperatures (Vargas Zeppetello et al., 2019). For example, an increase in surface temperature warms and moistens the boundary layer through upward turbulent energy fluxes, inducing increased DLR; thus, it is difficult to determine if increased DLR is the dominant driver or largely a diagnostic of surface temperatures. Therefore, the processes that drive the Arctic surface response during moist intrusion events are not fully understood.

The Arctic surface response—mainly in the Barents–Kara Sea region—associated with synoptic-scale winter-season poleward tropospheric energy flux events has been a focus of numerous studies (e.g., Doyle et al., 2011; Yoo et al., 2012; Woods et al., 2013; D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Baggett and Lee, 2015; Woods and Caballero, 2016; Luo et al., 2016; Baggett and Lee, 2017; Gong et al., 2017; Gong and Luo, 2017; Luo et al., 2017; Johansson et al., 2017; Zhong et al., 2018; Liu et al., 2018; Chen et al., 2018; Luo et al., 2019; Tyrlis et al., 2019; Graham et al., 2019b). In general, these events are associated with the following temporal structure:

1) Anomalous tropospheric sensible and latent heat flux convergence contribute to the warming and moistening of the troposphere over the region and at the pressure level of the anomalous convergence. These events preferentially occur over the Atlantic sector of the Arctic basin (approximately 20°W to 80°E) and are associated with atmospheric blocking over the Ural Mountains [i.e., Ural blocking (UB)].

2) The anomalously warm and moist troposphere with increased cloudiness increases the DLR at the surface (mainly over sea ice), initiating surface heating (i.e., reduced surface heat loss and the reduced growth or melting of sea ice). Moisture and clouds contribute to increased DLR through increased atmospheric emissivity (Graversen and Burtu, 2016), which reduces the longwave cooling efficiency of the Arctic atmosphere and surface (Hegyi and Taylor, 2018). Increased surface air temperatures and specific humidity also suppress upward surface turbulent sensible and latent heat fluxes (SHLH), reducing surface heat loss (mainly over open ocean). Storm driven mixing of warm Atlantic water likely causes further ice melt.
3) Reduced SIC and increased sea surface temperatures are followed by enhanced upward SHLH, contributing to the warming of the lower-troposphere at the expense of the surface.

The initial state and response of the Arctic surface can differ for individual tropospheric energy flux events. Tropical convection in the Pacific warm pool can trigger planetary scale waves that amplify the climatological stationary wave pattern, initially driving adiabatic warming in the Arctic and subsequently driving long duration surface warming associated with large eddy fluxes of sensible and latent heat and increased DLR (Yoo et al., 2012; Baggett and Lee, 2015, 2017). Additionally, the initial state and response of the Arctic surface during an event appears sensitive to the type of atmospheric blocking pattern (Chen et al., 2018; Luo et al., 2019). Chen et al. (2018) found that more persistent UB events are associated with an initially lower SIC and a more intense surface response in the Barents–Kara Seas due to increased moisture and DLR as opposed to suppressed SHLH. However, Chen et al. (2018) only focused on surface fluxes over areas of majority sea ice, likely reducing contributions from SHLH. More importantly, the potential large contributions to increased DLR from the surface warming itself—suggested by the work of Vargas Zeppetello et al. (2019)—questions the dominant role of increased moisture in driving the surface response and variability in the surface response.

The increased frequency of events over recent decades—consistent with the increased duration (Rinke et al., 2017) or frequency (Valkonen et al., 2021) of Arctic cyclones—is potentially driven by the increased occurrence of La Niña–like tropical convection over the Pacific warm pool (Baggett and Lee, 2017) or an increased frequency of UB events (Luo et al., 2016; Rinke et al., 2017; Gong and Luo, 2017; Chen et al., 2018; Luo et al., 2019). Interestingly, the increased UB frequency could be driven by reduced sea ice in the Barents–Kara Seas and the associated atmospheric circulation response, a potential positive feedback mechanism (Gong and Luo, 2017; Chen et al., 2018). However, Peings (2019) found no significant atmospheric response in model simulations with perturbed ice in the Barents–Kara Seas. The divergence between modeling and observational studies potentially results from models underestimating the role of sea ice in driving an atmospheric circulation response (Overland et al., 2021) or from a misinterpretation of causality—atmospheric circulation anomalies drive rather than respond to sea ice anomalies (Blackport et al., 2019).
The above review demonstrates that tropospheric energy flux events can drive Arctic surface warming and sea ice loss. Here we define these events as anomalous convergence of tropospheric fluxes of moist static energy (MSE; sensible, latent, and geopotential), denoted $F_{trop}$ poleward of $70^\circ\text{N}$. Such anomalous $F_{trop}$ events are associated with tropospheric warming and moistening and anomalously downward surface fluxes (increased DLR and suppressed SHLH). In addition, a fraction of the energy input by a poleward $F_{trop}$ anomaly is lost upward across the tropopause and does not impact the Arctic surface (Chapter 1; Cardinale et al., 2021). It is possible that the variability in the surface response to tropospheric energy flux events is explained by differences in the fraction of the total energy input that reaches the Arctic surface—a quantity we refer to as the surface heating efficiency. The processes that determine this heating efficiency are not well understood but are crucial for understanding how these events may change in a future warmer climate.

To this end, we define and compute a new metric for the Arctic surface heating efficiency. Specifically, this metric describes the fraction of the anomalous tropospheric energy source due to $F_{trop}$ that is balanced by an anomalous net surface flux ($\text{NSF}$; combined turbulent heat and longwave fluxes) over the entire Arctic polar cap. Here, “anomalous” refers to deviations from the climatological seasonal cycle. The NSF is used as a proxy for surface changes as it can drive both surface temperature and sea ice variability.

We are particularly interested in the vertical structure of $F_{trop}$ and its coupling to the Arctic surface. In Chapter 1, we analyzed the vertical structure of the MSE flux convergence and the Arctic surface response during $F_{trop}$ and stratospheric energy flux ($F_{\text{strat}}$) events (e.g., sudden stratospheric warmings) and found that a majority of the energy input by a poleward anomaly in $F_{trop}$—mainly in the lower troposphere—was balanced by the NSF, while a poleward anomaly in $F_{\text{strat}}$ had little influence on the NSF. This suggests that the surface heating efficiency is linked to the vertical structure of the MSE flux convergence. We hypothesize that concentrating the MSE flux convergence (i.e., atmospheric heating and moistening) closer to the surface increases the surface heating efficiency. It is possible that tropospheric energy flux events have a large variability in surface heating efficiency and that the role of atmospheric circulations on Arctic warming and sea ice loss during winter is best explained by changes in the efficiency of events as opposed to changes in the total frequency of events. Using the Modern-Era Retrospective Analysis for Research and Applications, version
2 (MERRA-2) reanalysis, we compute the Arctic surface heating efficiency of tropospheric energy flux events during winter from January 1980 to March 2020.

Several key findings of our study have not been suggested in previous literature. Here we give a brief summary of these results:

1) The vertical structure of the MSE flux convergence and the stability of the lower troposphere—associated with the initial state of the Arctic surface prior to the event—largely determine the surface heating efficiency of a tropospheric energy flux event.

2) Downward anomalies in surface turbulent heat fluxes (i.e., suppression of upward sensible and latent heat fluxes), as opposed to anomalous DLR, explain the majority of differences in the surface heating efficiency between events.

3) The surface heating efficiency of tropospheric energy flux events has increased, especially since 2004.

The efficiency metric will be introduced in section 2. An analysis of the efficiency of tropospheric energy flux events will be presented in section 3a, comparisons of events with different efficiencies will be presented in section 3b, and a trend analysis will be presented in section 3c.

3.2 Data and methods

3.2.1 MERRA-2 and the MSE flux

We use winds, atmospheric temperatures (not including skin temperature), specific humidity, geopotential, radiative fluxes at the surface and top of atmosphere [TOA; i.e., outgoing longwave radiation (OLR)], surface turbulent energy fluxes, sea ice concentration (SIC), and cloud concentrations from MERRA-2. MERRA-2, the latest atmospheric reanalysis (1980–present) produced by NASA’s Global Modeling and Assimilation Office (GMAO), has a horizontal resolution of 0.5° x 0.625°, 72 vertical levels with output interpolated to 42 pressure levels up to 0.1 hPa, and a temporal resolution of 3-hours (GMAO, 2015). Winter seasons (NDJFM) from January 1980 to March 2020 are used in this analysis. An evaluation of the performance of surface fluxes in the MERRA-2 reanalysis over Arctic sea ice during
winter can be found in Graham et al. (2019a). Compared to observations from the Norwegian Young sea ice campaign (N-ICE2015), MERRA-2 shows high skill in simulating downwelling longwave radiation and poor skill in simulating turbulent heat fluxes—similar to all reanalyses evaluated in Graham et al. (2019a). However, errors in the turbulent heat fluxes can be largely explained by differences in the point observations and the grid cell averages in MERRA-2, which contain a fraction of open ocean and larger mean upward turbulent fluxes relative to the point observations over sea ice (i.e., where the SIC is 100%). Despite large differences in mean cloud fractions between reanalyses and satellite products, MERRA-2 has demonstrated skill in estimating anomalous cloud fractions (Liu and Key, 2016). More information on the evaluation of the climate in MERRA-2 can be found in Bosilovich et al. (2015), and information on input observations can be found in McCarty et al. (2016).

Fluxes of moist static energy (MSE) and contributions from the sensible heat (SH), latent heat (LH), and geopotential (GP) fluxes on pressure levels at 70°N are computed following Chapter 1, where the flux was separated into contributions from the eddy and mean meridional circulation (MMC) flux, neglecting contributions from the net mass flux (NMF). Contributions to the total vertically integrated polar cap–averaged MSE flux convergence \( F_{\text{wall}} \) from the troposphere \( F_{\text{trop}} \) are obtained by integrating the flux from 1000 to 300 hPa. The MSE flux is computed instantaneously from the 3-hourly data and averaged daily. On pressure levels, the MSE flux and local contribution to the integrated MSE flux convergence have units of J kg\(^{-1}\) m s\(^{-1}\) and W m\(^{-2}\) (100 hPa)\(^{-1}\), respectively. Vertically integrated flux values (e.g., \( F_{\text{trop}} \)) have units of W m\(^{-2}\) (expressed as a polar cap–averaged flux convergence) and time-integrated values have units of MJ m\(^{-2}\).

### 3.2.2 A metric for the Arctic surface heating efficiency

We define the Arctic surface heating efficiency, \( E_{\text{trop}} \), as:

\[
E_{\text{trop}} = \frac{\langle NSF' \rangle}{F_{\text{trop}}' - \int_{300}^{1000} \left\langle \frac{\partial h_m'}{\partial t} \right\rangle dp}
\]  

(3.1)

where NSF is the net surface flux (the sum of turbulent energy fluxes and net longwave radiation; positive downward), \( F_{\text{trop}} \) is the polar cap–averaged tropospheric MSE flux convergence, and \( h_m \) is the moist enthalpy (latent and sensible heat). Angle brackets indicate
polar cap–averages and primes indicate anomalies relative to the daily mean annual cycle; $E_{\text{trop}}$ is computed with daily mean anomalies. Note that the contribution from the geopotential energy tendency is not explicitly written in the moist enthalpy tendency term in (3.1) as the dry enthalpy (sensible heat) accounts for changes in internal and potential energy—a consequence of hydrostatic balance (see Boer, 1982). Although the absorbed shortwave flux would typically be included in the NSF term in (3.1), we neglect it since our analysis is limited to winter.

The denominator in (3.1), which we refer to as the net tropospheric energy source (NTES), is the net excess energy available after accounting for the moist enthalpy storage in the Arctic atmosphere. The denominator thus represents the maximum energy available for anomalous surface heating. By taking the ratio of the anomalous NSF to NTES, our metric $E_{\text{trop}}$ approximates the fraction of the atmospheric source that heats the surface during periods of anomalously strong $F_{\text{trop}}$. We then assume that the residual fraction $1 - E_{\text{trop}}$ is lost upward due to fluxes across the tropopause. The moist enthalpy tendency is computed using second-order accurate central differences using 3-hourly data and then averaged daily. Contributions from the kinetic energy to NTES are small and have been neglected.

All anomalies have been linearly detrended and a low-pass filter with a 2-day cutoff frequency was applied to NTES anomalies prior to calculation of the metric. The potential impact of unrealistic long-term trends in polar cap–averaged fields in reanalyses (e.g., Taylor et al., 2018) on $E_{\text{trop}}$ prompted the use of detrended anomalies. Applying a low-pass filter increased the winter correlation between NTES [i.e., denominator of (3.1)] and the NSF (0.75 to 0.83) and increased the number of winter days with an $E_{\text{trop}}$ between 0 and 1 (an additional 60 days)—indicating a reduction in the energy imbalance.

Figure 3.1 shows a histogram of $E_{\text{trop}}$ over all winter days. $E_{\text{trop}}$ would be physically bounded between 0 and 1 if atmospheric MSE fluxes were the only potential drivers of anomalous NSF. In reality, the NSF is also dependent on processes internal to the Arctic, and $E_{\text{trop}}$ could become large on days when NTES (denominator) is small. Examples of these internal processes include energy fluxes across the tropopause that are forced by stratospheric processes and strong surface winds that drive sea ice (mechanical forcing) and surface flux changes. $E_{\text{trop}}$ falls between 0 and 1 on 63% of all winter days (not shown). Most days outside this range (i.e., $E_{\text{trop}}$ greater than 1 or less than 0) are associated with small anomalies in
NTES. These are days for which the atmosphere would have little potential to drive NSF anomalies. For example, days with an NTES magnitude less than 12 W m\(^{-2}\) (approximately the bottom 50% of days) make up 85% of days with an \(E_{\text{trop}}\) greater than 1 or less than 0. Additionally, Chapter 1 found that energy budget residuals in MERRA-2 are small relative to \(F_{\text{trop}}\) and the NSF during periods of anomalously large \(F_{\text{trop}}\), increasing our confidence that MERRA-2 provides an accurate \(E_{\text{trop}}\) during significant tropospheric heating events.

3.2.3 Definition of tropospheric energy flux events

Tropospheric energy flux (i.e., \(F_{\text{trop}}\)) events on synoptic timescales are identified and assigned a mean \(E_{\text{trop}}\) using the following procedure:

1) Isolate days when the NTES anomaly is greater than 0.

2) Calculate the cumulative time integral of the NTES anomaly over each event period—the consecutive days that meet the first criteria.

3) An event is defined as a cumulative increase of 8 MJ m\(^{-2}\), with the event date (also referred to as the central date) defined as the day on which the threshold is reached.

4) Calculate the mean \(E_{\text{trop}}\) over the event period (including the event date) for days when \(E_{\text{trop}}\) is between 0 and 1.

The 8 MJ m\(^{-2}\) threshold maximizes the amount of events that are 2–6 days apart [synoptic timescales; Gulev et al. (2002)], while reducing events that occur less than two days apart. Additionally, for all anomalously positive NTES periods, the average cumulative NTES anomaly is about 8 MJ m\(^{-2}\).

In contrast to the definition of \(F_{\text{trop}}\) events, composite analysis of events with various mean \(E_{\text{trop}}\) (section 3b) includes days with an \(E_{\text{trop}}\) greater than 1 or less than 0; however, results of this chapter are not sensitive to this choice. We decide to include these days for better visualization of the temporal and vertical structure of events.
3.3 Surface heating efficiencies of winter tropospheric energy flux events

3.3.1 High, medium, and low-efficiency events

Figure 3.1 shows histograms of both daily $E_{\text{trop}}$ for all winter days and mean $E_{\text{trop}}$ during winter-season synoptic-scale $F_{\text{trop}}$ events as defined in Section 3.23.2.3. The mean $E_{\text{trop}}$ over the subset of all days with $E_{\text{trop}}$ between 0 and 1 (63% of days) is 50%. This indicates that half of the anomalous net tropospheric energy source (NTES) goes toward anomalous heating of the surface and half is expressed as an upward loss across the tropopause. The separation into three efficiency-based categories (high, medium, and low-efficiency events) is sensible given the approximately normal distribution of the event mean $E_{\text{trop}}$. High and low-efficiency events are defined as events with an average $E_{\text{trop}}$ greater than or equal to the 75th percentile ($E_{\text{trop}} \geq 0.63$) and less than or equal to the 25th percentile ($E_{\text{trop}} \leq 0.42$) of events, respectively. The remainder of the events are classified as medium-efficiency events. These percentiles are calculated from the event mean $E_{\text{trop}}$, not from all days. In total, there are 312 events (7.8 per season), separated into 78 high, 156 medium, and 78 low-efficiency events.

Figure 3.2 provides an example of the identification of $F_{\text{trop}}$ events and their separation into high, medium, and low-efficiency during the 2009–2010 winter season. Eight total events are identified, with three of these positive NTES periods each containing two events. Multiple events are allowed in the same positive NTES period if the 8 MJ m$^{-2}$ threshold is crossed multiple times (e.g., at 8 and 16 MJ m$^{-2}$). In total, about half of all events occur in multiple event periods; in these periods, there is an average of about 4 days between events. 60, 58, and 32% of high, medium, and low-efficiency events follow another event in the same period (not shown).

The events in Fig. 3.2 are separated into four high, one medium, and three low-efficiency events indicated by the cumulative average $E_{\text{trop}}$ when a threshold is reached (red, gray, and blue bars in the upper panel of Fig. 3.2). The mean $E_{\text{trop}}$ of each event can also be seen by comparing NTES with the net surface flux (NSF); events with the highest $E_{\text{trop}}$ are associated with the smallest difference between the time-integrated NTES and NSF. Fig. 3.2 also shows several episodes of positive NTES that do not meet our 8 MJ m$^{-2}$ threshold and
are thus not counted as events. In total, there are 135 episodes that do not meet the 8 MJ m\(^{-2}\) threshold; if included, these events would make up about 36% of events.

Anomalies in NTES and \(F_{\text{trop}}\) are shown in the lower panel of Fig. 3.2. Events (i.e., periods of positive NTES anomalies) are typically preceded by positive \(F_{\text{trop}}\) anomalies. Distinct positive \(F_{\text{trop}}\) anomalies are also typically found before events that are closely preceded by another event (blue lines prior to the second and fourth events in Fig. 3.2); thus, it is sensible to consider these events as separate. However, some events (e.g., the eighth event in Fig. 3.2) are associated with neutral or negative \(F_{\text{trop}}\) anomalies and the tropospheric forcing is hard to distinguish from the previous event. For example, it appears that a single persistent \(F_{\text{trop}}\) anomaly forces both the seventh and eighth events; the troposphere remained anomalously warm and moist long enough for another event to be identified even when coincident with a negative \(F_{\text{trop}}\) anomaly.

Figures 3.1 and 3.2 reveal that events with similar tropospheric heating can have very different impacts on the surface energy budget. To improve our understanding of the processes that determine \(E_{\text{trop}}\) during these events, we compare composites of high, medium, and low-efficiency events. Composited quantities to be analyzed include atmospheric temperatures, sea ice concentration (SIC), lower-tropospheric stability [calculated as the potential temperature (\(\theta\)) difference between 850 hPa and 2-m; \(\theta_{850hPa} - \theta_{2m}\)], surface fluxes, local MSE flux convergence, and cloud concentrations in the 30 days before and after the central dates of events. For events that occur in November and March, the composite analysis includes some days in October and April. Results from the composite analysis in section 3b will mainly be provided in three periods: “before” (day \(-21\) to \(-7\)), “during” (day \(-7\) to 7), and “after” (day 7 to 21) the event. Based on the method for defining events described above, the maximum anomalous NTES occurs in the “during” period. The anomalous poleward MSE flux convergence, on the other hand, does not neatly fit into the “before” or “during” period as it tends to span days \(-14\) to 0.

### 3.3.2 Event Comparisons

#### 3.3.2.1 Before events—preconditioning

Figure 3.3 shows the polar cap–averaged and Atlantic sector–averaged (20°W to 80°E) anomalies in the 2-m temperature and SIC for high, medium, and low-efficiency events. The
Atlantic sector average is provided, as this region is generally associated with the largest anomalies. High-efficiency events are found to have a different preconditioning of the Arctic surface than medium and low-efficiency events; these differences are statistically significant mainly in the Atlantic sector (cf. dashed lines in Figs. 3.3a–c). In the high-efficiency composite prior to day $-7$ (i.e., prior to the onset of anomalous surface heating), 2-m temperature anomalies are positive and SIC anomalies are negative, especially in the Atlantic sector. In the same period for both medium and low-efficiency composites, Atlantic sector 2-m temperature and SIC anomalies are generally negative and positive, respectively. The statistically significant differences between events are not largely explained by the presence of antecedent events. The preconditioning signal is qualitatively similar in event composites that require events to be at least 21 days apart (not shown).

Figures 3.4a–c show spatial maps of SIC anomalies and lower-tropospheric stability anomalies during the “before” period. The largest SIC anomalies excluding the Canadian Archipelago are found in the Barents–Kara Seas. In the high-efficiency composite, decreased stability is found over the majority of the Arctic Ocean, with the largest negative anomalies found over areas of negative SIC anomalies (Fig. 3.4a), suggesting increased lower-tropospheric turbulent mixing. In the medium and low-efficiency composites, increased stability is found over much of the Arctic Ocean, especially over areas of increased SIC anomalies (Figs. 3.4b and 3.4c), suggesting decreased lower-tropospheric turbulent mixing. The association between stability and SIC anomalies is consistent with Deser et al. (2010) and Vihma (2014); enhanced turbulent heat fluxes are found in areas of reduced SIC and act to warm the air above the surface and decrease the stability of the lower troposphere, while suppressed turbulent heat fluxes are found in areas of increased SIC and act to increase the stability (not shown).

The vertical structure of anomalous atmospheric temperatures in the polar cap average before day $-7$ (Figs. 3.5a–c) reveal that temperature anomalies extend the depth of the troposphere. Polar cap-averaged lower-tropospheric stability anomalies are negative and positive in high and medium-efficiency events, respectively, consistent with anomalies in the Barents–Kara Seas (Figs. 3.4a and 3.4b). Low-efficiency events are associated with the lowest tropospheric temperatures; however, polar cap-averaged stability anomalies are negative (reduced) mainly due to contributions from the Canadian Archipelago (Fig. 3.4c).
During events

Figure 3.5 includes composites of the vertical and temporal structure of polar cap-averaged temperature anomalies (Figs. 3.5a–c) and the cumulative time integral of anomalous local MSE flux convergence (filled contours in Figs. 3.5d–f). The use of cumulative time integrals allows for easier visualization of the vertical structure of the flux anomalies and comparison with the temperature anomalies. Positive anomalies in the MSE flux convergence—which increase the sensible and latent energy of the Arctic column—emerge at approximately day $-14$ in all three composites. MSE flux convergence anomalies have a similar “bottom-heavy” profile in high and medium-efficiency events; anomalies maximize in the lower troposphere and increase with pressure until around 925 hPa (Figs. 3.5d and 3.5e). The MSE flux convergence anomalies in low-efficiency events are approximately vertically uniform above 925 hPa (Fig. 3.5f). In all three composites, the sensible (SH) and latent (LH) flux convergence anomalies maximized in the lower troposphere are compensated by geopotential (GP) flux divergence (negative) anomalies (cf. red and black contours in Figs. 3.5d–f). The lack of a bottom-heavy MSE flux convergence in the low-efficiency composite results from a weaker vertical gradient in positive SH and LH flux anomalies and stronger negative anomalies in the GP flux that extend deeper into the troposphere. Additionally, the near-surface layer of anomalous MSE flux divergence decreases in depth and intensity from low to high-efficiency (cf. brown-filled contours between events in Figs. 3.5d–f). Despite vertical structure differences, the cumulative $F_{\text{trop}}$ anomaly at day 0 is similar for each event composite, with values of 28.9, 26.2, and 24.5 MJ m$^{-2}$ in high, medium, and low-efficiency events, respectively (blue lines in Figs. 3.5g–i).

The vertical structure of anomalous warming in high, medium, and low-efficiency events can be explained by considering the vertical structure of the MSE convergence and the lower-tropospheric stability. The vertical structure of the flux convergence compares reasonably well with the warming structure (cf. red-filled contours in Figs. 3.5a–c and green-filled contours in Figs. 3.5d–f). Despite similarities in the flux convergence in high and medium-efficiency events, high-efficiency events are associated with larger lower-tropospheric temperature anomalies. High-efficiency events occur in the presence of reduced Atlantic sector SIC, which reduces the lower-tropospheric stability and increases turbulent mixing (Fig. 3.4a)—likely enhancing the lower-tropospheric temperature anomaly by mixing the heating anomaly.
to the surface (Kayser et al., 2017). Both medium and low-efficiency events occur in the presence of enhanced Atlantic sector SIC, which increases the lower-tropospheric stability and decreases turbulent mixing (Figs. 3.4b and 3.4c); thus, differences in lower-tropospheric temperature anomalies between medium and low-efficiency events can be explained by differences in the vertical structure of the MSE convergence. From a Lagrangian perspective, as anomalously warm and moist air propagates through a stratified Arctic, it ascends along or slightly less than the slope of isentropic surfaces (Komatsu et al., 2018; You et al., 2021). In a relatively unstratified Arctic (e.g., high-efficiency events), the anomalously warm and moist air may remain closer and better coupled to the surface.

Figure 3.5 also includes composites of cumulative time integrals of anomalous NTES and polar cap–averaged fluxes at the surface and tropopause. NTES and NSF begin to increase 1–2 weeks before the central date and peak 1–2 weeks after the central date. The interpretation that an NSF anomaly represents the fraction of an NTES anomaly that heats the surface should be approximately valid during these events in which $F_{\text{trop}}$ is anomalously large. In low-efficiency events, the NTES anomaly only becomes positive when the troposphere is anomalously warm and moist, around day $-5$ (black line in Fig. 3.5i). The noticeably smaller peak in the anomalous NTES in the low-efficiency composite mainly results from a larger fraction of the $F_{\text{trop}}$ anomaly going into moist enthalpy storage (not shown) and a shorter duration of positive temperature anomalies (cf. red-filled contours between events in Figs. 3.5a–c). Only a small increase in the downward (DLR) and net (NLR) longwave fluxes are found when moving from low to high-efficiency (cf. red lines in Figs. 3.5g–i). Differences in the NSF and $E_{\text{trop}}$ between events, visualized by the size of the shaded residual, are best explained by differences in the combined sensible and latent turbulent heat (SHLH) fluxes (cf. green lines). This suggests that near-surface warming and moistening over areas of open ocean or thin sea ice are key to reducing surface heat loss (i.e., positive NSF anomalies) and increasing $E_{\text{trop}}$. Additionally, the SHLH term is the dominant contributor to the anomalous polar cap–averaged NSF in high and medium-efficiency events from day $-14$ to 10. These results appear inconsistent with Chen et al. (2018), who suggested that turbulent heat fluxes play a secondary role in synoptic-scale events; however, their study mainly focused on sea ice variability and did not include areas of majority exposed ocean.

To investigate further the mechanisms determining $E_{\text{trop}}$, we show in Fig. 3.6 the time-
integrated polar cap–averaged anomalies in surface fluxes and the upward loss from day −14 to 10 for each event composite. These values are identical to those shown for day 10 in Figs. 3.5g–i. Also shown are contributions to the Arctic average from areas of open ocean (SIC < 15%; i.e., the ice edge), partial sea ice (98% < SIC < 15%), and the combined areas of nearly full ice cover (SIC < 98%) and land. In all events, SHLH anomalies are the dominant contributor to the anomalous NSF in areas of open ocean, while DLR is the dominant contributor to the anomalous NSF over sea ice and land, consistent with Woods and Caballero (2016). However, differences in the anomalous NSF, and thus $E_{\text{trop}}$, between events are best explained by SHLH anomalies in the Arctic average, areas of open ocean, and areas of partial sea ice up to 98% SIC (cf. green boxes between events). Changes in SHLH anomalies between events over areas of open ocean and partial sea ice appear of equal importance. Interestingly, downward SHLH anomalies in high-efficiency events in areas of partial ice cover result in part from increased SIC anomalies in the Canadian Archipelago, while the upward SHLH anomalies in low-efficiency events result in part from decreased SIC in the Canadian Archipelago (not shown). Changes in DLR anomalies between events are only dominant in areas with near 100% SIC and land, where absolute contributions to the polar cap–averaged NSF are relatively small. Additionally, when normalized by the time-integrated NTES anomaly from day −14 to 10, polar cap–averaged DLR anomalies are approximately equal in all three events and NLR anomalies only slightly increase from low to high-efficiency (not shown). During this period, longwave fluxes only contribute 8% of the change in $E_{\text{trop}}$.

The relatively large downward SHLH anomalies in high and medium-efficiency events in areas of open ocean and partial sea ice have implications for Atlantic water temperatures and sea ice growth, respectively. However, there exist substantial uncertainties in SHLH over areas of sea ice (e.g., Taylor et al., 2018; Graham et al., 2019a), especially during cold-stable periods in winter, where upward SHLH biases are found (Graham et al., 2019a). Although these biases are largely explained by differences in point observations and grid cell averages which contain a fraction of open ocean, upward biases are still found when compared to satellite-retrieved SHLH. Satellite-retrieved climatological SHLH is downward (i.e., the surface acts as an atmospheric heat sink) during winter over areas of sea ice (Taylor et al., 2018), while the climatological winter SHLH is weakly upward in MERRA-2 (not shown). Errors in SHLH should then be smaller in the warm periods associated with $F_{\text{trop}}$. 

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events; however, relative to the climatologically downward satellite-retrieved SHLH over sea ice, SHLH anomalies over sea ice (Fig. 3.6) would be reduced. These uncertainties would not greatly impact surface flux differences between events, assuming that the potential biases are equal in all events.

Figure 3.7 shows the zonal structure of the anomalous local MSE flux convergence and contributions from eddy, mean meridional circulation (MMC), and LH fluxes in the 14 day average before the central date of events. The majority of tropospheric heating in all three events can be attributed to eddy fluxes, dominated by SH and LH fluxes (not shown), consistent with past work (e.g., Yoo et al., 2012; Baggett and Lee, 2015, 2017). In high and medium-efficiency events, the local MSE flux convergence is concentrated in the Atlantic sector (Figs. 3.7a and 3.7c), much like the moist intrusion events in Woods and Caballero (2016), with a more zonally uniform flux in low-efficiency events (Fig. 3.7e). Negative MMC anomalies—dominated by GP fluxes (not shown)—indicate adiabatic cooling and act to cool the lower troposphere, especially in low-efficiency events (solid blue lines in Figs. 3.7b, 3.7d, and 3.7f). Positive contributions from the LH and SH components to the MMC flux indicate warm and moist inflow and partially compensate the adiabatic cooling (not shown). In high-efficiency events, positive anomalies in the lower-tropospheric MMC prior to day $-14$ (not shown) may also help explain the warm troposphere prior to the event—consistent with adiabatic warming preceding eddy fluxes during Arctic surface warming events initiated by tropical convection (Yoo et al., 2012). Lower-tropospheric MSE flux convergence anomalies in high and medium-efficiency events are significantly different from low-efficiency events (thick black lines in Figs. 3.7b and 3.7d) due to enhanced positive eddy flux anomalies and reduced negative MMC flux anomalies. All three events are associated with poleward LH flux anomalies, with a significant difference between high and low-efficiency events. However, increases in the LH flux from low to high-efficiency are small relative to increases in the total MSE flux (cf. solid and dashed black lines in Figs. 3.7b, 3.7d, and 3.7f) and only partially explain differences in lower-tropospheric and surface heating between events.

The 2-m temperature and SIC responses shown in Fig. 3.3 are consistent with the anomalous surface fluxes and zonal structure of each event. The increase in the Atlantic sector 2-m temperatures and the decrease in the Atlantic sector SIC in each event composite is larger than in the polar cap–average, especially in high-efficiency events (cf. dashed and
solid lines in Fig. 3.3a). Although the temperature and SIC tendencies are similar between events, the magnitude and duration of anomalous 2-m temperature and SIC in the Atlantic sector increase from low to high-efficiency. In all composites, lower-tropospheric stability anomalies are largely positive over areas of open ocean in the Atlantic sector (equatorward of SIC anomalies in the Barents–Kara Seas) and negative elsewhere over the Arctic basin (Figs. 3.4d–f). Relatively large downward SHLH anomalies over areas of open ocean act to slow the near-surface warming relative to 850 hPa, resulting in positive stability anomalies. Large differences in anomalous SIC in the Barents–Kara Seas can be seen in Figs. 3.4d–f, with relatively large negative anomalies in high, small negative anomalies in medium, and a mix of small negative and positive anomalies in low-efficiency events. Differences in the surface response to each type of event are not due to differences in total energy input into the Arctic column; for example, these results are nearly identical when normalized by the cumulative $F_{\text{trop}}$ from day $-14$ to 0 (not shown).

The cloud response is shown in Fig. 3.8. In all three event composites, high cloud fraction anomalies are generally positive, especially in the Atlantic sector (Figs. 3.8g–i). Differences between composites are apparent in middle and low clouds. Mainly negative cloud fraction anomalies are found in the middle and low cloud layers in high-efficiency events, a mix of positive and negative anomalies are found in medium-efficiency events, and mainly positive anomalies are found in low-efficiency events, except for low cloud anomalies in the Atlantic sector (Figs. 3.8a–c). The results for high and medium-efficiency events are inconsistent with Johansson et al. (2017), who found increased satellite-retrieved cloud fraction anomalies throughout the column during moist intrusions. Additionally, recent studies using satellite data find that reduced lower-tropospheric stability results in increased low cloud amounts (Barton et al., 2012; Solomon et al., 2011, 2014; Taylor et al., 2015; Yu et al., 2019). However, the negative lower-tropospheric stability anomalies over much of the Arctic in all events (Fig. 3.4a–f) only appear associated with positive cloud anomalies in low-efficiency events. Low cloud amount decreases in high and medium-efficiency events potentially result from areas of increased stability in the Atlantic sector and over land, or from errors that result from the parameterization of cloud physics in MERRA-2 (Graham et al., 2019a). The decrease in the Atlantic sector low cloud fractions, especially in high-efficiency events, may also result from strong heating in the boundary layer, which raises the inversion level and breaks up stratocumulus clouds (Eirund et al., 2020).
Figure 3.9 shows the variability in anomalous polar cap–averaged cloud fraction, top-of-atmosphere (TOA) and surface cloud radiative effects (CRE), and surface fluxes. As suggested by the spatial plots, polar cap–averaged cloud fraction anomalies are generally positive, with negative anomalies found in low and middle clouds in high-efficiency events. The relatively weak anomalous TOA (OLR\textsubscript{clear-sky}−OLR) and surface (NLR−NLR\textsubscript{clear-sky}) CRE compared to DLR suggests that anomalous cloud fractions do not play an important role in determining $E_{\text{trop}}$. This result is consistent with Sokolowsky et al. (2020), who found that temperature and water vapor impact DLR more than clouds during moist intrusion events over Utqiaġvik, Alaska. The mean CRE in high and medium-efficiency events are close to 0, likely the result of opposing cloud fraction anomalies in lower and upper levels. Additionally, the CRE decreases from low to high-efficiency, acting in the opposite direction of DLR. Figure 3.9 clearly shows that the variability in the NSF in each event and differences in the NSF between events are best explained by turbulent, not longwave fluxes. High and medium-efficiency SHLH anomalies are statistically different from low-efficiency events at the 95% confidence level. High-efficiency DLR anomalies are also significantly different from low-efficiency events; however, no significant difference is found when anomalies are normalized by the cumulative $F_{\text{trop}}$ anomaly at day 0.

### 3.3.2.3 After events

We now discuss impacts on the Arctic surface after day 7 (following the passage of the tropospheric heating event), with emphasis on sea ice recovery. Figure 3.3 shows that in all composites, 2-m temperatures are generally higher and the SIC is generally smaller when compared to the “before” period. SIC anomalies in Figs. 3.3 and 3.4 reveal a slower sea ice recovery in both the polar cap average and Atlantic sector average when moving from low to high-efficiency. The slow growth of sea ice following a high-efficiency event is consistent with positive 2-m temperature anomalies (Fig. 3.3a) that further suppress upward turbulent heat fluxes (Fig. 3.5g). Near climatological 2-m temperatures in medium-efficiency events are associated with faster sea ice growth relative to high-efficiency events. Negative 2-m temperature anomalies in the Atlantic sector in low-efficiency events are associated with a sea ice growth rate almost equal to the sea ice decline rate during the event—consistent with large negative daily mean SHLH anomalies after day 7 (slope of the green line in Fig. 3.5i).
After each event, lower-tropospheric stability anomalies generally decrease over the Arctic Ocean (especially in the Atlantic sector), consistent with Johansson et al. (2017) and Liu et al. (2018), associated with decreased SIC anomalies (Figs. 3.4g–i). Reduced stability over the Barents–Kara Seas is particularly apparent in the high-efficiency composite (Fig. 3.4g) and is associated with upward anomalies in SHLH in this region (not shown), a likely source for the positive 2-m temperature anomalies over the Arctic during this period—despite downward SHLH anomalies in the polar cap–average.

In Fig. 3.10, we summarize our understanding of the key distinguishing features of high, medium, and low-efficiency events in the periods “before”, “during”, and “after” an event. We will use this schematic to anchor a comprehensive discussion in section 3.4. First, however, we will look at trends in the MERRA-2 data.

### 3.3.3 Trends in the efficiency of tropospheric energy flux events

In section 3, we separated synoptic-scale $F_{trop}$ events during winter into three efficiency-based categories and found that Arctic surface warming and sea ice loss are indeed larger in events with a higher $E_{trop}$, despite similar magnitudes of atmospheric forcing (i.e., $F_{trop}$). Trends in the relative frequency of high, medium, and low-efficiency events should then be crucial in understanding the role of synoptic-scale events in Arctic winter warming and sea ice decline.

Figure 3.11 shows the cumulative sum of seasonal (winter) high, medium, and low-efficiency events from January 1980 to March 2020. Over the last two decades, the number of high-efficiency events has rapidly increased, while low-efficiency event frequency has declined. High-efficiency events have increased by about two events per year and low-efficiency events have decreased by about two events per year, with little change in the frequency of medium-efficiency events, as indicated by the decadal changes in the average number of events per year in Fig. 3.11. In general, the increase in high-efficiency events occurs in mid-winter (December–February), while the decrease in low-efficiency events occurs in the middle to late winter (January–March). Although the number of total events remains unchanged throughout the period, the surface heating efficiency of $F_{trop}$ events has increased.

It is important to note that we linearly detrended the source data prior to computing $E_{trop}$, but that these trends in counts of events in each efficiency bin are nonetheless evident.
in Fig. 3.11. The raw MERRA-2 data actually contain a decreasing $F_{trop}$ trend and increasing NSF trend; thus, our results based on the detrended data may represent an underestimate of the true $E_{trop}$ trends. We return to this point in the discussion. Additionally, the event count trends are not sensitive to the method of trend removal from NTES and NSF anomalies (e.g., a removal of the 5-year running mean; not shown).

### 3.4 Conclusions and discussion

In this chapter, we applied an Arctic energy budget perspective to understanding how winter-season synoptic-scale tropospheric weather events influence surface heating. We defined a new metric for surface heating efficiency, $E_{trop}$, which approximately measures the fraction of the anomalous net tropospheric energy source (NTES) during flux events that reaches the surface (land, ocean, or sea ice) in the Arctic-wide area average. It is important to note that this interpretation of $E_{trop}$ is most relevant to periods in which tropospheric energy flux ($F_{trop}$) anomalies are large. We computed the mean $E_{trop}$ over $F_{trop}$ events, periods where $F_{trop}$ is likely the dominant driver of anomalous surface fluxes, over the period January 1980 to March 2020. The $F_{trop}$ events were then separated into three efficiency-based categories: high, medium, and low-efficiency. A summary schematic of the vertical and temporal structure of each event composite is provided in Fig. 3.10.

In high-efficiency events, bottom-heavy (i.e., concentrated in the lower troposphere) anomalous poleward $F_{trop}$—mainly in the Atlantic sector between 20°W and 80°E—occurs in the presence of reduced Atlantic sector sea ice concentration (SIC). Reduced SIC results in reduced lower-tropospheric stability and the implied enhanced turbulent mixing contributes to further near-surface warming (i.e., mixing the warm and moist air aloft to the surface). A majority of the anomalous NTES goes into surface heating, dominated by surface sensible and latent heat fluxes (SHLH) as a result of reduced near-surface vertical gradients in temperature and specific humidity over areas of open ocean and partial sea ice. SIC is further reduced and is slow to recover as positive temperature anomalies persist.

In medium-efficiency events, the Arctic is subject to the same bottom-heavy poleward $F_{trop}$ profile as in high-efficiency events, but this occurs in the presence of anomalously high Atlantic sector SIC and lower-tropospheric stability; as a result, reduced turbulent mixing limits near-surface warming. Approximately half of the anomalous NTES goes into surface...
heating, with comparable contributions from DLR and SHLH anomalies. Lower-tropospheric stability reduces and negative SIC anomalies emerge; however, near climatological temperatures allow SIC to quickly recover relative to high-efficiency events.

In low-efficiency events, the anomalous poleward $F_{\text{trop}}$ is more uniform both zonally and vertically (above 925 hPa), and occurs in the presence of increased Atlantic sector SIC and an anomalously cold troposphere with increased lower-tropospheric stability. The combination of poleward $F_{\text{trop}}$ anomalies extending deep into the upper troposphere and reduced turbulent mixing results in an approximate uniform warming of the troposphere. A minority of the anomalous NTES goes into surface heating, mainly through increased DLR. Tropospheric temperatures cool to below climatology, and sea ice quickly recovers to above climatology after a brief reduction.

This work suggests an alternative chain of causality by which the atmosphere drives Arctic winter surface warming. Much of the recent literature on synoptic-scale tropospheric heating events (e.g., D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Woods and Caballero, 2016; Gong et al., 2017; Lee et al., 2017) argues that increased DLR, associated with the anomalously warm and moist atmosphere, primarily drives the surface response [i.e., the net surface flux (NSF)] and found little preconditioning of the Arctic surface. Our results instead show that the impact of individual $F_{\text{trop}}$ events on surface heating varies tremendously depending on two main factors: (1) the preconditioning of the Arctic surface, and (2) the vertical structure of $F_{\text{trop}}$. Together these determine the efficiency of surface heating primarily through the suppression of polar cap–averaged climatological upward turbulent heat fluxes. The anomalous SHLH is four times larger in the high-efficiency composite than the low-efficiency composite (see Fig. 3.9), due to the combination of bottom-heavy atmospheric heating and preconditioned weak stratification. Note that while individual high-efficiency events are associated with greater surface heating, the surface warming (i.e., temperature change) is similar in all three events (see Fig. 3.3). Instead, the greater surface heating in high-efficiency events results in reduced ocean cooling or slower sea ice growth. It is only in the longer term (seasonal or greater) that the accumulated effect of more high-efficiency events would manifest as anomalously warm temperatures.

Regional differences can help reconcile our results with recent literature (e.g., D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Woods and Caballero, 2016; Gong et al., 2017;
Lee et al., 2017). While DLR anomalies are found to dominate in areas of sea ice and land, consistent with the aforementioned literature, the anomalous polar cap–averaged DLR is in fact smaller than anomalous polar cap–averaged SHLH for all but the low-efficiency composite and is nearly the same across all event types. Additionally, differences in SHLH anomalies between events are equally large in areas of open ocean and partial sea ice. We suggest that DLR is primarily a diagnostic of lower-tropospheric air temperature and is not a very useful indicator of the causality of surface warming. Additionally, we found a weak preconditioning signal in composites of all events (not shown), possibly explaining why the initial state of the Arctic surface appears unimportant in the aforementioned literature. Our results are consistent with Chen et al. (2018), who found a larger and more persistent surface response during events preceded by low SIC anomalies.

A potential limitation of our results is the substantial uncertainty in SHLH over areas of sea ice in reanalysis data (e.g., Taylor et al., 2018; Graham et al., 2019a). Our results showing that SHLH anomalies explain the majority of $E_{\text{trop}}$ differences between events should be relatively insensitive to these biases so long as the biases are independently distributed across all event types. Evaluating this would be an interesting avenue for future work.

Our results also show that clouds are not a primary factor in determining surface heating efficiency. Interestingly, increased cloud fractions are largest in low-efficiency events, and both surface and TOA cloud radiative effects (CRE) are relatively small and decrease with increasing $E_{\text{trop}}$. Clouds are thus a small mitigating factor in the efficiency compared to the large range in SHLH. Negative low cloud anomalies in high and medium-efficiency events are inconsistent with recent work using satellite data that find increased cloud amounts in moist intrusion events (Johansson et al., 2017) and over areas of decreased stability (e.g., Yu et al., 2019). However, errors in the relatively small CRE anomalies would likely not impact our results.

The trends we found in the relative frequency of high vs. low-efficiency events may have played a role in the recent winter trends in Arctic surface temperature and sea ice. Despite a lack of trends in the total number of $F_{\text{trop}}$ events, from the period 1980–1999 to 2000–2019, high-efficiency events have increased by approximately two events per year while low-efficiency have decreased by approximately two events per year. This result indicates that the atmosphere is becoming increasingly efficient at heating the Arctic surface during
these winter-season events. We speculate that there is both a shift from low to medium-efficiency events (i.e., an increase in bottom-heavy $F_{\text{trop}}$ flux events) and a shift from medium to high-efficiency events (i.e., a decrease in lower-tropospheric stability). An increase in bottom-heavy $F_{\text{trop}}$ events would be consistent with an increase in moist intrusion events (e.g., Woods and Caballero, 2016), which have a similar vertical and zonal structure to high and medium-efficiency events. A decrease in lower-tropospheric stability—a signature of Arctic warming (e.g., Cohen et al., 2014)—combined with an increase in bottom-heavy $F_{\text{trop}}$ events could explain why high-efficiency events have increased in frequency, despite little change in medium-efficiency event frequency.

We found this shift towards high-efficiency events despite calculating $E_{\text{trop}}$ from detrended $F_{\text{trop}}$ and NSF anomaly data. We removed long-term trends in order to allay concerns about possible spurious trends in the reanalysis data due to changes in the observing system (e.g., Taylor et al., 2018). While we cannot rule out the possibility that the event-count trends in Fig. 3.11 are likewise spurious, the fact that they appear in our derived $E_{\text{trop}}$ metric after filtering based on event occurrence is strong evidence that they represent a real physical relationship that is captured in the MERRA-2 data. It is worth noting that the long-term winter-season trend in the MERRA-2 NSF data that we removed is actually downward (atmosphere to ocean) at a rate of 2.15 W m$^{-2}$ decade$^{-1}$ (largely driven by a downward trend in SHLH). If this trend were retained in our event calculations, it would have thus produced an even larger trend toward higher efficiency. In ongoing work, we are verifying whether similar trends toward more frequent high-efficiency events are found in historical model simulations, and whether this represents a response to anthropogenic forcing.

Our results suggest an underappreciated atmospheric driver for Arctic amplification of climate change. High-efficiency tropospheric heating events occur preferentially when the Arctic is anomalously warm and unstratified. These events in turn provide more energy to the surface and less upward loss, prolonging the warm and unstratified conditions. This represents a positive feedback loop by which Arctic winter surface warming can be amplified, consistent with our finding of a multi-decadal increase in high-efficiency events. Importantly, this amplification could in principle occur with no change in the large-scale circulation or increase in $F_{\text{trop}}$, nor in cloud feedbacks. Rather, it would manifest in a traditional attribution study as a positive lapse-rate feedback (Pithan and Mauritsen, 2014)—the mechanisms for
which remain under active debate in the Arctic (e.g., Henry et al., 2021). By taking an event-by-event synoptic view, our efficiency metric provides a new perspective on these processes.

To better understand the relationship between the poleward energy flux and polar amplification, Chapter 4 will focus on the changes in the surface heating efficiency in a warmer climate, including the complex causal relationships between $F_{trop}$ structure, Arctic stratification, and sea ice loss.
3.5 Figures

Figure 3.1: Daily efficiencies during winter (NDJFM) from January 1980 to March 2020 (black) and time mean efficiencies of high (red), medium (gray), and low (blue) efficiency events. Dashed vertical lines indicate the mean of each distribution, excluding days with an efficiency less than 0 or greater than 1. A linear scale is used between 0 and 1 and a logarithmic scale is used elsewhere.
Figure 3.2: (Upper) Cumulative time integrals of anomalies in the net tropospheric energy source (NTES; solid black; MJ m$^{-2}$) and net surface flux (dashed black; MJ m$^{-2}$) and the event mean efficiency ($E_{\text{trop}}$; bars) for each event during the 2009–2010 winter season. Dotted vertical lines indicate central dates of high (red bars), medium (gray bars) and low (blue bars) efficiency events and dotted horizontal lines indicate event thresholds. (Lower) Anomalies in $F_{\text{trop}}$ (blue; W m$^{-2}$) and NTES (black; W m$^{-2}$) during the 2009–2010 winter season.
Figure 3.3: Composite of polar cap–averaged (solid) and Atlantic sector–averaged (dashed) anomalies in 2 m temperature (K; red) and sea ice concentration (%) black in the 30 days before and after the central date of (a) high, (b) medium, and (c) low-efficiency events. The shading indicates anomalies statistically different from low-efficiency events at the 95% confidence level. For p values < 0.05, we reject the null hypothesis of equal averages.
Figure 3.4: Composite of anomalous time mean lower-tropospheric stability (K, fill) and sea ice concentration (%), SIC; black contours) during the period (a)–(c) before (day $-21$ to $-7$), (d)–(f) during (day $-7$ to 7), and (g)–(i) after (day 7 to 21) the central date of (left) high, (center) medium, and (right) low-efficiency events. SIC anomalies are contoured at 1.5, 3, 6, 12, and 24%, with line widths increasing with magnitude. Solid and dashed contours indicate positive and negative SIC anomalies, respectively. The boundaries of the Atlantic sector ($20^\circ$W to $80^\circ$E and $70^\circ$N to $90^\circ$N) are also shown (black longitude lines) in (b), (e), and (h).
Figure 3.5: Composite of the (a)–(c) polar cap-averaged anomalous temperature (K), (d)–(f) cumulative time integral of the anomalous local MSE flux convergence poleward of 70°N [fill; W m\(^{-2}\) (100hPa)\(^{-1}\)] and contributions from the combined SH and LH components (red) and the GP component (black) during (left) high, (center) medium, and (right) low-efficiency events. The SH, LH, and GP anomalies in (d)–(f) are contoured at 2, 4, 8, 16, and 32 W m\(^{-2}\) (100hPa)\(^{-1}\), with line widths increasing with magnitude. Solid and dashed contours indicate positive and negative anomalies, respectively. (g)–(i) show cumulative anomalies in \(F_{trop}\) (blue), the net tropospheric energy source (NTES; solid black), and net surface flux (NSF; dashed black). NSF anomalies (positive downward) are decomposed into additive contributions from sensible and latent surface turbulent heat fluxes (SHLH; green) and net longwave radiation flux (NLR; dotted red). NLR is further decomposed into anomalous upward (ULR; dashed red) and downward (DLR; solid red) fluxes. The gray shading shows the residual between the NTES and NSF, which we interpret as upward loss.
Figure 3.6: Time-integrated anomalies (MJ m$^{-2}$) in the polar cap–averaged DLR (dark red boxes), SHLH (green boxes), upward loss (gray boxes), and ULR (light red boxes) from day $-14$ to 10 in high, medium, and low-efficiency events. Also shown are absolute contributions to the polar cap–averaged NSF in areas of SIC $< 15\%$ (open ocean), $98\% <$ SIC $< 15\%$, and SIC $< 98\%$ and land. Polar cap–averaged anomalies are identical to the anomalies in (g)–(i) of Fig. 5 at day 10.
Figure 3.7: Composite of the (left) time mean eddy contribution to the local MSE flux convergence and (right) time and zonal mean total local MSE flux convergence (black), with contributions from the eddy (red) mean meridional circulation (MMC; blue), total latent heat (LH; dashed black), eddy LH (dashed red), and MMC LH (dashed blue) fluxes $[\text{W m}^{-2} (100\text{hPa})^{-1}]$ in the 14 days before the central date of (a)–(b) high, (c)–(d) medium, (e)–(f) low-efficiency events. Vertical dashed lines in (a), (c), and (e) indicate the longitude boundaries of the Atlantic sector ($20^\circ$W to $80^\circ$E). The shading in (b), (d), and (f) indicates anomalies statistically different from low-efficiency events at the 95% confidence level. A two-sided $t$ test was used to determine significance. For $p$ values $< 0.05$, we reject the null hypothesis of equal averages.
Figure 3.8: Composite of time mean (a)–(c) low cloud (below roughly 700 hPa), (d)–(f) middle cloud (about 700–400 hPa), and (g)–(i) high cloud (above about 400 hPa) anomalies (%) during the 7 days before and after the central date of (left) high, (center) medium, and (right) low-efficiency events.
Figure 3.9: Box-and-whisker plots of time and polar cap–averaged (left) anomalous (W m$^{-2}$) net surface flux (NSF), with contributions from the combined sensible and latent surface turbulent energy fluxes (SHLH) and net longwave flux (NLR), downward longwave flux (DLR) and the cloud radiative effect (CRE) at the top-of-atmosphere (TOA) and surface (SFC), and (right) anomalous high, medium, and low clouds (%) in the 7 days before and after the central date of high (red), medium (black), and low (blue) efficiency events. All boxes extend from the 25th to 75th percentile, with a horizontal line at the mean, and whiskers extend to the 5th and 95th percentiles.
Figure 3.10: Schematic of the anomalous energy flux convergence (black), temperature (red), Atlantic sector sea ice concentration (SIC; blue hashed), combined surface sensible and latent turbulent heat flux (green arrows), surface downward longwave flux (red arrows), and upward loss at the tropopause (blue arrows) during the period before, during, and after the central date of (upper) high, (middle) medium, and (lower) low-efficiency events.
Figure 3.11: (Left) Cumulative sum of high (red), medium (black), and low (blue) efficiency events for each winter season, and (right) decadal changes in the average number of events per year.
CHAPTER 4

The Increasing Arctic Surface Heating Efficiency of Tropospheric Energy Flux Events in the CESM Large Ensemble

4.1 Introduction

Chapter 3 examined the drivers of the surface heating efficiency ($E_{\text{trop}}$) during winter polar cap–averaged tropospheric MSE flux convergence ($F_{\text{trop}}$) events. Chapter 3 found that $E_{\text{trop}}$ is driven by the vertical structure of $F_{\text{trop}}$ and Arctic surface preconditioning, with high-efficiency events associated with a bottom-heavy $F_{\text{trop}}$ and reduced Arctic lower-tropospheric stability (associated with reduced sea ice). The main goal of this chapter is to use climate model simulations to test for significant changes in $E_{\text{trop}}$ in a future warming scenario and to quantify the partial contributions to the net surface flux from changes in $F_{\text{trop}}$ and $E_{\text{trop}}$.

The bottom-amplified warming of the Arctic is consistently found in recent observations (Screen and Simmonds, 2010a; Cohen et al., 2014) and climate model projections (Boeke and Taylor, 2018), with surface air temperatures increasing and projected to increase at a faster rate than the global average, especially during the cold season. However, models disagree on the change in atmospheric energy transport and the role of atmospheric circulations in Arctic warming is still debated (e.g., Taylor et al., 2022).

There are different metrics for the influence of atmospheric energy transport on the Arctic surface climate. These metrics have been used to quantify the contribution from energy transport to Arctic warming during winter. $F_{\text{wall}}$ is the total vertically integrated polar cap–averaged moist static energy [MSE; sensible (SH), latent (LH), and geopotential (GP)] flux convergence and is frequently used (e.g., Hwang et al., 2011; Koenigk et al., 2013; Pithan and Mauritsen, 2014; Goosse et al., 2018; Boeke and Taylor, 2018; Feldl et al., 2020; Hahn et al., 2021). A major problem with using $F_{\text{wall}}$ as a metric is the non-negligible stratospheric contributions ($F_{\text{strat}}$, integrated above 300 hPa)—19% of the winter mean $F_{\text{wall}}$ from 1980–2016 in the MERRA-2 (Chapter 2; Cardinale et al., 2021). In Chapter 2, $F_{\text{strat}}$ was found to have little influence on the Arctic surface energy budget, suggesting a relationship between the surface energy budget and the vertical structure of the MSE flux convergence.
In traditional attribution studies, $F_{\text{wall}}$ changes are positive, but relatively small, and contribute to little Arctic warming and Arctic amplification in a multimodel mean when the polar cap is defined from 60 to 90°N (e.g., Pithan and Mauritsen, 2014; Feldl et al., 2020; Hahn et al., 2021); small contributions are found in both the winter and summer (Pithan and Mauritsen, 2014; Hahn et al., 2021). $F_{\text{wall}}$ changes are anti-correlated with Arctic amplification and is found to dampen the inter-model spread; i.e., $F_{\text{wall}}$ decreases in models with relatively large Arctic amplification and increases in models with relatively weak Arctic amplification (e.g., Boeke and Taylor, 2018).

$F_{\text{wall}}$ is often partitioned into latent heat (LH) and dry static energy (DSE; SH+GP) components (e.g., Koenigk et al., 2013; Feldl et al., 2020; Hahn et al., 2021). LH has an amplified warming effect on the Arctic due to its relationship with cloud and atmospheric emissivity changes (Graversen and Burtu, 2016; Baggett and Lee, 2017; Graversen and Langen, 2019). The DSE component of $F_{\text{wall}}$ decreases in a future warming climate, compensated by an increase in the LH component (e.g., Hwang et al., 2011; Koenigk et al., 2013; Graversen and Burtu, 2016; Hahn et al., 2021). However, attribution studies typically do not consider the amplified warming effect of increased LH and contributions from energy transport changes to the lapse-rate feedback. The degree to which the LH component compensates the DSE component depends on the definition of the Arctic domain [more models show a decrease in $F_{\text{wall}}$ when defined at 70°N rather than 60°N (Hwang et al., 2011)] and on the degree of the Arctic amplification (Boeke and Taylor, 2018).

The MSE flux convergence integrated over the troposphere ($F_{\text{trop}}$: 1000–300 hPa) was found to be a better metric than $F_{\text{wall}}$ for lower-tropospheric temperature variability on daily to monthly time scales (Chapter 2; Cardinale et al., 2021). However, not all of $F_{\text{trop}}$ is available for surface heating. A fraction of $F_{\text{trop}}$ goes into the heating and moistening of the troposphere (e.g., the moist enthalpy tendency). We define the net tropospheric energy source, NTES, as:

$$\text{NTES} = F_{\text{trop}} - \int_{300}^{P_s} \left( \frac{\partial h_m}{\partial t} \right) \frac{dp}{g},$$

(4.1)

where $h_m$ is the moist enthalpy (latent and sensible heat) and angle brackets indicate polar cap–averages. (4.1) provides the maximum tropospheric energy available for surface heating. The $h_m$ tendency term is near zero in the winter mean, but is an important term on short time scales (see Chapter 3). Contributions from changes in $F_{\text{trop}}$ and NTES to Arctic warming
have not been carefully studied.

Episodic $F_{\text{trop}}$ events—commonly referred to as moist intrusion events—are synoptic-scale events of poleward $F_{\text{trop}}$ anomalies that temporarily warm the Arctic surface and reduce sea ice growth through increased downward longwave radiation and reduced upward turbulent heat fluxes (Doyle et al., 2011; Yoo et al., 2012; Woods et al., 2013; D.-S. R. Park et al., 2015; H.-S. Park et al., 2015; Baggett and Lee, 2015; Woods and Caballero, 2016; Luo et al., 2016; Baggett and Lee, 2017; Gong et al., 2017; Gong and Luo, 2017; Luo et al., 2017; Johansson et al., 2017; Zhong et al., 2018; Liu et al., 2018; Chen et al., 2018; Luo et al., 2019; Tyrlis et al., 2019; Graham et al., 2019b). The increased frequency of these events have been linked to recent winter surface warming and sea ice loss in reanalyses (e.g., Woods and Caballero, 2016). Potential drivers of the increased frequency include increased tropical convection in the Pacific warm pool (Baggett and Lee, 2017) and increased Ural blocking (Luo et al., 2016; Gong and Luo, 2017; Chen et al., 2018; Luo et al., 2019).

Only a fraction of the anomalous NTES during an $F_{\text{trop}}$ event reaches the surface, resulting in anomalous surface heating. In the MERRA-2, the variability in the fraction of energy that reaches the surface (i.e., the surface heating efficiency) mainly depends on the vertical structure of the anomalous $F_{\text{trop}}$ and the preconditioning of the Arctic surface (see Chapter 3). Events with a relatively high-efficiency are associated with initially lower sea ice concentration (SIC) and bottom-heavy $F_{\text{trop}}$ anomalies. Reduced SIC results in reduced lower-tropospheric stability and increased turbulent mixing—increasing the coupling between the surface and free troposphere and enhancing the near-surface warming and downward turbulent heat flux anomalies. High-frequency data (e.g., daily) is necessary to compute this surface heating efficiency, which is a non-nonlinear function of the net surface flux (NSF) and $F_{\text{trop}}$. The use of low-frequency data (e.g., monthly) would not capture synoptic-scale events. In Chapter 3, the frequency of high-efficiency events was found to increase in the MERRA-2 from 1980 to 2020, while the frequency of events with relatively low-efficiency decreased, indicating an increase in the surface heating efficiency of $F_{\text{trop}}$ events. A potential driver of the increased surface heating efficiency is a decrease in Arctic lower-tropospheric stability, suggesting a positive feedback loop (see Chapter 3).

To investigate the changes in $F_{\text{trop}}$ and $E_{\text{trop}}$ in response to greenhouse gas forcing, we use simulations of the Community Earth System Model Large Ensemble (CESM-LE) with
a strong radiative forcing scenario (RCP8.5) for the period 2071–2080. A large ensemble is used to separate the forced response to greenhouse gases from internal variability. The CESM-LE is associated with a relatively large Arctic amplification (Holland and Landrum, 2021). While $F_{\text{wall}}$ is found to decrease in models with a relatively large amplification factor (Boeke and Taylor, 2018), the relationship between $F_{\text{trop}}$ and Arctic amplification have not been carefully studied.

This chapter addresses the following research question and associated hypothesis: What is the role of changes in the atmospheric energy transport in future Arctic warming in the RCP8.5 warming scenario of the CESM-LE? We hypothesize that an increase in the surface heating efficiency mitigates the decrease in $F_{\text{trop}}$.

### 4.2 Atmospheric energy flux convergence in the CESM-LE

#### 4.2.1 CESM-LE output

We use output from the Community Earth System Model Large Ensemble (CESM-LE; Kay et al., 2015). The simulations provide a 40-member ensemble with a strong radiative forcing scenario to isolate the forced response to anthropogenic warming. Data from the historical simulations (years 1990–2005; referred to as 20C) are used and compared with simulations using years 2071–80 from RCP8.5 simulations. The year ranges were selected since they provide the three-dimensional model output at 6-hourly resolution, which is needed for the analysis. Energy fluxes are computed directly following Cardinale et al. (2021) and Chapter 2, using 6-hourly instantaneous data at 70°N (meridional winds, atmospheric temperature, specific humidity, and geopotential height). Prior to the calculation of energy fluxes, we interpolated the 30 hybrid sigma-pressure levels to 28 pressure levels from 1000 to 5 hPa; the interpolated output in the troposphere has a vertical resolution of 25 hPa below 700 hPa, 50 hPa between 500 and 700 hPa, and a 100 hPa resolution above 500 hPa. For comparison of the 20C climate average conditions, we use output from the Modern-Era Retrospective analysis for Research and Applications, Version 2 (MERRA-2; GMAO, 2015) from 1980 to 2020.
4.2.2 Winter mean energy flux convergence in the 20C ensemble

Winter (NDJFM) mean polar cap-averaged (70–90°N in this chapter) temperature profiles and energy flux convergence metrics in the 20C ensemble are shown in Figs. 4.1a and 4.1b. The climatological winter mean 20C temperature profile is characterized by a surface inversion from 1000 to 850 hPa (Fig. 4.1a). In the winter mean, $F_{\text{wall}}$ supplies 116 W m$^{-2}$ to the Arctic (Fig. 4.1b). The vertical structure of $F_{\text{wall}}$ is bimodal, with local maxima in the troposphere at the top of the stable boundary layer and in the stratosphere (not shown). The winter mean $F_{\text{trop}}$ is smaller than $F_{\text{wall}}$, supplying 95 W m$^{-2}$, with the latent heat flux convergence (LH) contributing 13.5 W m$^{-2}$ or 14%. The winter mean net tropospheric energy source (NTES) is slightly larger than $F_{\text{trop}}$ (97 W m$^{-2}$), due to a relatively small and negative winter mean moist enthalpy (latent and sensible heat; $h_m$) tendency. A negative winter mean $h_m$ tendency indicates that the troposphere, on average, is warmer and moister in November compared to March.

4.2.3 Comparison between the 20C ensemble and the MERRA-2.

Winter mean atmospheric temperatures are generally similar to the MERRA-2 except in the upper troposphere and stratosphere (Fig. 4.1a; cf. blue and black lines). The surface inversion is also stronger in the CESM-LE. The stronger surface inversion is consistent with the overestimated stability over Arctic sea ice in CMIP5 models compared to reanalyses (Pithan et al., 2014), which has been linked to the shortcomings in the representation of mixed-phase clouds.

All terms in Fig. 4.1b are overestimated by 2–6 W m$^{-2}$, when compared to the MERRA-2, with all MERRA-2 terms lying outside the ensemble range (cf. blue vertical lines and white circles). Note that the MERRA-2 terms are averaged between 1980 and 2020; however, these differences are slightly reduced when the 1990–2005 period is used (not shown). Consistent with the results from the MERRA-2 in Chapter 2, the winter mean $F_{\text{trop}}$ of 95 W m$^{-2}$ in the 20C ensemble mean is about 20% smaller than $F_{\text{wall}}$. 
4.2.4 A metric for the Arctic surface heating efficiency

The fraction of an NTES anomaly that reaches the surface can be calculated by the Arctic surface heating efficiency metric, $E_{\text{trop}}$. With (4.1), we can rewrite (3.1), $E_{\text{trop}}$, as:

$$E_{\text{trop}} = \frac{\langle \text{NSF}' \rangle}{\text{NTES}'},$$

(4.2)

where NSF is the net surface flux (the sum of turbulent energy fluxes and net longwave radiation; positive downward) and primes indicate anomalies relative to the daily mean annual cycle. The denominator in (4.2) represents the maximum energy available for anomalous surface heating. Anomalies are computed relative to the daily mean annual cycle (Fig. 4.2 for surface fluxes) and are linearly detrended—computed separately for each ensemble member to ensure a stationary time series. Only 10–15 years in each run are available, so a low-pass filter with a cutoff frequency of 12 days is applied to each term before calculating the mean annual cycle for smoothing.

The mean annual cycle of the NSF is not driven by NTES. Rather, the seasonality is best explained by the absorbed solar radiation during summer (large increase and decrease in dashed lines between April and November in Fig. 4.2a) and sea ice growth during winter, which acts to suppress upward turbulent heat fluxes (increase in blue and red lines from November to April in Fig. 4.2c). Thus, the winter mean $E_{\text{trop}}$, calculated with anomalies relative to the daily mean annual cycle, measures the coupling between $F_{\text{trop}}$ and the NSF and approximates the fraction of the raw NTES that reaches the surface. The partial contribution to the time mean NSF from the atmosphere can be written as:

$$\langle \text{NSF}_{\text{atm}} \rangle = E_{\text{trop}} \text{NTES},$$

(4.3)

where overbars indicate the winter mean. While $E_{\text{trop}}$ in (4.3) is still calculated with anomalies relative to the daily mean annual cycle, (4.3) approximates the total (i.e., raw) atmospheric source that heats the surface in the winter mean. Note that the winter mean $E_{\text{trop}}$ neglects periods where $E_{\text{trop}}$ is greater than 1 or less than 0 (about 40% of days); these are periods where NTES anomalies are generally small and have little potential to drive anomalous surface heating (not shown).

A winter mean $E_{\text{trop}}$ (calculated daily) and NTES of about 50% and 97 W m$^{-2}$ (Fig.
4.1b), respectively, in the 20C ensemble mean indicates that the atmosphere contributes about 49 W m$^{-2}$ of surface heating averaged over the Arctic. This value can be interpreted as the ocean heat flux convergence necessary to maintain the same surface temperatures in the absence of $F_{trop}$, or the amount that the NSF would decrease in the absence of $F_{trop}$. The winter mean NSF would decrease (increased surface cooling) from $-47$ W m$^{-2}$ to $-95$ W m$^{-2}$. An NSF of $-95$ W m$^{-2}$ is quantitatively similar to the y-intercept of the regression line between daily mean NSF and NTES.

4.2.5 Winter mean change in energy flux convergence metrics in the RCP8.5 warming scenario

Concurrent with a surface amplified warming and weaker surface inversion (Fig. 4.1a), winter mean $F_{wall}$ and $F_{trop}$ decrease in the RCP8.5 warming scenario by about 9.5 W m$^{-2}$ (Fig. 4.1c). A decrease in $F_{wall}$ is consistent with the relatively large winter Arctic amplification of approximately 3 (Boeke and Taylor, 2018); defined as the ratio between the polar-cap averaged temperature change to global temperature change. The difference between the change in $F_{trop}$ and $F_{wall}$ are relatively small; however, the ensemble spread in $F_{trop}$ changes are reduced compared to $F_{wall}$. An increase in the LH component appears to only partially compensate the decrease in the DSE component. However, the surface warming impact of increased LH is likely amplified due to increased clouds, atmospheric emissivity, and close proximity to the surface. NTES decreases by about 8.5 W m$^{-2}$ and has a smaller ensemble spread than both $F_{wall}$ and $F_{trop}$. All terms in Fig. 4.1c are significantly different from 0. The decrease in $F_{wall}$, $F_{trop}$, and NTES would contribute to Arctic surface cooling if we used a top-of-atmosphere (TOA) budget attribution method; e.g., divide the decrease in W m$^{-2}$ by the Planck feedback in W m$^{-2}$ K$^{-1}$ (Hahn et al., 2021).

4.3 The surface heating efficiency of $F_{trop}$ events

4.3.1 Definition of $F_{trop}$ events

The definition of $F_{trop}$ events follows Chapter 3, where an event is defined as a cumulative increase in the anomalous NTES of 8 MJ m$^{-2}$ during winter. The event mean $E_{trop}$ is then calculated over this period. There are 60.5 events per decade in the 20C ensemble.
mean, with an ensemble standard deviation of 4.5 events.

Events are separated into three efficiency-based categories: high (event mean $E_{trop} \geq 75$th percentile), low (event mean $E_{trop} \leq 25$th percentile), and medium (the remainder of events). Similar to the MERRA-2, high and low-efficiency event $E_{trop}$ thresholds are about 63 and 44%, respectively; i.e., a majority of the anomalous NTES goes into surface heating and is lost upward across the tropopause (upward loss) in high-efficiency and low-efficiency events, respectively.

### 4.3.2 Drivers of $E_{trop}$

As in Chapter 3, to understand the processes that determine $E_{trop}$ in the CESM-LE, we compare composites of the anomalous tropospheric energy budget and lower-tropospheric stability (LTS; the potential temperature difference between 850 hPa and 2 m) in high, medium, and low-efficiency events (Fig. 4.3) in the 20C ensemble. Composites of the vertical structure of the time-integrated anomalous local MSE and LH flux convergence in the 14 days before the central date (the date that 8 MJ m$^{-2}$ cumulative NTES anomaly threshold is reached) of an event is shown in Figs. 4.3a–c. High and medium-efficiency events are associated with bottom-heavy $F_{trop}$ anomalies that are significantly different from low-efficiency events.

The Arctic preconditioning before an event is assessed by the mean LTS between days $-21$ and $-7$ (Figs. 4.3d–f). High-efficiency events occur in the presence of reduced LTS; the mean LTS is significantly different from low-efficiency events. The LTS in medium-efficiency events is not significantly different from low-efficiency events. This suggests a greater coupling between the atmosphere and surface in high-efficiency events, resulting in enhanced near-surface warming and moistening due to increased mixing. Another interpretation of the differences between high and medium-efficiency comes from a Lagrangian prospective: as anomalously warm and moist air propagates through a stably stratified Arctic, it ascends along or slightly less than the slope of isentropic surfaces (Komatsu et al., 2018; You et al., 2021). In high-efficiency events, the reduced LTS allows for the warm and moist air to remain closer to the surface.

Anomalies in the LTS relative to the daily mean annual cycle are not found prior to high-efficiency events; however, high-efficiency events mainly occur in early winter when raw
sea ice concentration (SIC) and associated LTS is lower. The association between stability and SIC anomalies is consistent with Deser et al. (2010) and Vihma (2014) and can be explained through changes in surface turbulent heat fluxes.

Time-integrated anomalies in the tropospheric energy budget between days $-14$ and $10$ are shown in Figs. 4.3g–i. While both the net longwave radiation (NLR) and SHLH anomalies in high and medium-efficiency anomalies are significantly different from low-efficiency events, differences in the mean $E_{\text{trop}}$ between events (i.e., the ratio between the NSF and NTES terms) is best explained by differences in the turbulent heat fluxes (SHLH). Downward SHLH anomalies account for about 77% of the increase in $E_{\text{trop}}$ from low to high-efficiency events during this period (cf. green violin plots between composites). Although the surface cloud radiative effect (CRE; NLR$-\text{NLR}_{\text{clear-sky}}$) is positive in all three composites, only medium-efficiency events are associated with a significantly larger time-integrated CRE anomaly.

### 4.3.3 $E_{\text{trop}}$ event composite comparison between the MERRA-2 and 20C ensemble

Figs. 4.3a–c also shows the vertical structure of the local MSE and LH flux convergence in the MERRA-2 (dashed lines). MSE and LH flux convergence and surface flux anomalies are generally similar and lie within the ensemble range; e.g., a bottom-heavy vertical structure in high and medium-efficiency events. The tropospheric energy budget terms in the MERRA-2 (blue vertical lines in Figs. 4.3g–i) lie within the ensemble range, except for the surface CRE anomalies in high and medium-efficiency events (Figs. 4.3g and 4.3h). CRE differences between the CESM-LE and MERRA-2 potentially result from errors in the parameterization of cloud physics in MERRA-2 (Graham et al., 2019a) or from differences in the LTS prior to an event—the LTS is higher in the CESM-LE in all three event composites (not shown), consistent with Fig. 4.1a. CRE anomalies are consistent with Johansson et al. (2017), who found that cloud amounts increase during these episodic events.

The reduced SHLH anomalies in the CESM-LE in high and medium-efficiency events possibly result from the larger LTS and smaller upward SHLH during winter in the CESM-LE (Fig. 4.2); i.e., less potential for suppressed SHLH during an event. The larger upward SHLH in the MERRA-2 is likely partially explained by the upward biases over sea ice (Graham
et al., 2019a). NLR anomalies explain 23% of the change in $E_{\text{trop}}$ from low to high-efficiency, 15% higher than in the MERRA-2. This potentially results from a larger increase in the anomalous LH flux convergence and a larger contribution from the anomalous CRE.

There are about 10 fewer $F_{\text{trop}}$ events per decade in the ensemble mean, compared to the MERRA-2 (not shown). This result is consistent with the negative moist intrusion density bias in the Atlantic sector (Woods et al., 2017); $F_{\text{trop}}$ events most frequently impact the Atlantic sector (20°W to 80°; see Chapter 3). However, this bias shown in Woods et al. (2017) is compensated by a positive bias in the Pacific sector.

4.3.4 $F_{\text{trop}}$ event frequency and winter mean $E_{\text{trop}}$ changes

Fig. 4.4 shows changes in the relative frequency of high, medium, and low-efficiency events between the 20C and RCP8.5 ensembles. High-efficiency events increase, while low-efficiency events decrease in frequency in the ensemble mean and in the majority of members. All changes are significantly different from 0. The high-efficiency event increase of 13.4 events per decade—mainly during the early winter (not shown)—is nearly compensated by a decrease in low and medium-efficiency events; however, the change in medium-efficiency event frequency is relatively small.

The increased frequency of high-efficiency events is consistent with the increase in the winter mean $E_{\text{trop}}$ from 50.2 to 55.9% in the RCP8.5 warming scenario (Fig. 4.5a). We speculate that the increase in $E_{\text{trop}}$ and frequency of high-efficiency events results from decreasing LTS. Figure 4.6 shows that the increased surface heating efficiency in $F_{\text{trop}}$ events from 20C to RCP8.5 runs is best explained by an increase in downward SHLH anomalies, a consequence of increased coupling between the surface and free troposphere. An increase in the LH flux convergence and associated increases in NLR appears to play a secondary role. It is possible that the LH flux convergence plays a larger role in models with a larger increase in the winter mean LH flux convergence. Note that the contribution from the surface CRE decreases in the RCP8.5 ensemble mean despite a large increase in winter clouds (not shown), consistent with the decrease in LTS (e.g., Taylor et al., 2015). A potential explanation for this is that cloud concentrations are already high at the start of an event.
4.3.5 The implications of an increasing winter mean \( E_{\text{trop}} \)

The change in the contribution to the time mean NSF from the atmosphere can be written as:

\[
\Delta \langle \text{NSF}_{\text{atm}} \rangle = \Delta (E_{\text{trop}} \text{ NTES}) \approx E_{\text{trop}} \Delta (\text{NTES}) + \Delta (E_{\text{trop}}) \text{ NTES},
\]

where the uppercase delta indicates the winter mean change and the right-hand side (RHS) indicates the linear approximation. As in Hahn et al. (2021), the partial surface temperature change can be attributed by dividing the negative change in each term on the RHS of 4.4 by the Planck feedback. Here, we use the local Arctic Planck feedback factor (approximately \(-2.7 \text{ W m}^{-2} \text{K}^{-1}\) for the Arctic; Zhang et al., 2020). This attribution method provides only an approximate value for the partial surface temperature change, as it assumes that each term results in a vertically uniform warming.

Fig. 4.5b shows the contribution from NTES and \( E_{\text{trop}} \) changes to the NSF (i.e., the surface energy budget). The first term on the RHS of (4.4; blue histogram) indicates that a decrease in the NTES contributes to surface cooling (approximately 1.6 K), reducing the NSF by 4.3 W m\(^{-2}\). The second term on the RHS of (4.4; black histogram) indicates that an increase in \( E_{\text{trop}} \) contributes to surface heating (approximately 2 K), increasing the NSF by 5.5 W m\(^{-2}\).

The increase in \( E_{\text{trop}} \) compensates the decrease in \( F_{\text{trop}} \) (red histogram in Fig. 4.5b). A small, but statistically significant, increase in the energy gained by the surface (0.7 W m\(^{-2}\) or approximately 0.3 K of surface heating) results from a larger contribution from the change in \( E_{\text{trop}} \), with about 65% of members agreeing on the sign of the change.

4.4 Summary and discussion

In this chapter, we investigated the role of atmospheric circulations on Arctic (70–90\(^{\circ}\)N) warming during the winter (NDJFM) in the RCP8.5 warming scenario in the CESM large ensemble (CESM-LE). We argued that changes in the polar cap–averaged tropospheric energy flux convergence (\( F_{\text{trop}} \)) and changes in the coupling between \( F_{\text{trop}} \) and the surface energy budget should be considered when quantifying the contribution from atmospheric energy transport to Arctic surface warming. The coupling between \( F_{\text{trop}} \) and the surface
energy budget can be approximated by the surface heating efficiency ($E_{\text{trop}}$), which measures the fraction of an anomalous $F_{\text{trop}}$ that goes into anomalous surface heating after accounting for tropospheric heating and moistening. Importantly, when calculating $E_{\text{trop}}$, synoptic variability must be taken into account.

Winter mean $F_{\text{trop}}$ was found to decrease by 9.5 W m$^{-2}$ from the 20C to RCP8.5 runs, dominated by a decrease in the contribution from the dry static energy (DSE) and only partially compensated by the latent heat (LH) component. Winter mean $E_{\text{trop}}$ was found to increase by 5.7% (from 50.2 to 55.9%). Importantly, only days with an $E_{\text{trop}}$ between 0 and 1 were used to calculate winter means, which largely neglects periods where the atmosphere is not the dominant driver of NSF anomalies.

Composite analysis of synoptic-scale high, medium, and low-efficiency $F_{\text{trop}}$ events reveal that $E_{\text{trop}}$ is largest in events with reduced lower-tropospheric stability (LTS) and a bottom-heavy $F_{\text{trop}}$, consistent with Chapter 3. Consistent with the increase in the winter mean $E_{\text{trop}}$, the frequency of high-efficiency events was found to increase by 13.4 events per decade. Results suggest that the increase in $E_{\text{trop}}$ is a response to decreasing sea ice and the associated reduced LTS, which acts to increase the coupling between the surface and troposphere.

The net effect of the decrease in $F_{\text{trop}}$ and an increase in $E_{\text{trop}}$ is a small positive contribution to the change in the net surface flux and surface temperature during winter. This result supports the hypothesis that an increase in $E_{\text{trop}}$ mitigates the decrease in $F_{\text{trop}}$ and prevents the damping of Arctic amplification in the CESM-LE.

The use of the large ensemble allows us to interpret the efficiency increase as a forced response to greenhouse gases in the CESM-LE. It is also important to consider the robustness of these results to inter-model differences, particularly as the sign of the change of $F_{\text{wall}}$ is inconsistent across models (Hwang et al., 2011; Boeke and Taylor, 2018). Because our results suggest that $E_{\text{trop}}$ increases primarily due to loss of stratification, we expect that $E_{\text{trop}}$ will increase robustly with Arctic warming (and thus contribute to enhanced surface warming) independently of changes in $F_{\text{trop}}$. The mitigation that we have found in CESM-LE may be a model-specific result, but efficiency very likely increases in most models under future warming scenarios.
4.5 Figures

Figure 4.1: (a) Winter (NDJFM) mean polar cap-averaged (70–90°N) temperature profiles in 20C (1990–2005; blue) and RCP8.5 (2071–2080; red) runs, compared to the MERRA-2 reanalysis (1980–2020; black); the shading indicates the ensemble 5th–95th percentile range. (b) The winter mean total vertically integrated polar cap-averaged moist static energy (MSE) flux convergence ($F_{\text{wall}}$; red; W m$^{-2}$), the tropospheric contribution to $F_{\text{wall}}$ ($F_{\text{trop}}$; blue; W m$^{-2}$), contributions to $F_{\text{trop}}$ from the dry static energy (DSE; green; W m$^{-2}$) and latent heat (LH; cyan; W m$^{-2}$), and the net tropospheric energy source (NTES; gray; W m$^{-2}$) in 20C runs, compared to the MERRA-2 (blue vertical lines). (c) shows the winter mean change between 20C and RCP8.5 runs. Violin plots are used to show the ensemble distribution, where the shading indicates the probability density, thin horizontal lines indicate the 5th–95th percentile range, thick horizontal lines indicate the interquartile range, and white circles indicate the mean. Asterisks in (c) indicate where ensemble means are significantly different from 0 at the 95% confidence level.
Figure 4.2: Mean annual cycle of the polar cap–averaged (a) net surface flux (NSF; dashed) and contributions to the net surface flux from the combined net longwave and turbulent heat fluxes (NLR+SHLH; solid), (b) net longwave flux, and (c) additive contributions from latent and sensible turbulent heat fluxes in the 20C (blue; W m\(^{-2}\)) and RCP8.5 (red; W m\(^{-2}\)) runs of the CESM Large Ensemble, compared to the MERRA-2 (black; W m\(^{-2}\)). The shading indicates the ensemble 5th–95th percentile range.
Figure 4.3: Composite of the (a)–(c) time-integrated anomalous local MSE (red) and LH (blue) flux convergence (MJ m\(^{-2}\) (100 hPa\(^{-1}\)) from day \(-14\) to \(0\) in (left) high, (center) medium, and (right) low-efficiency events, compared to the MERRA-2 (dashed lines). (d)–(f) shows the mean lower-tropospheric stability (LTS; blue; K) between days \(-21\) and \(-7\). (g)–(h) shows time-integrated anomalies in the NTES (gray; MJ m\(^{-2}\)), net surface flux (NSF; light gray; MJ m\(^{-2}\)), combined sensible and latent surface turbulent heat fluxes (SHLH; green, MJ m\(^{-2}\)), net longwave radiation (red; MJ m\(^{-2}\)), and surface cloud radiative effect (CRE; MJ m\(^{-2}\)) from day \(-14\) to \(-10\), compared to the MERRA-2 (blue vertical lines). Violin plots follow the same convention as in Fig. 4.1. The shading in (a)–(c) indicates the ensemble 5th–95th percentile range. Asterisks indicate where ensemble means are statistically different from low-efficiency events at the 95% confidence level.
Figure 4.4: Winter mean change in the number of (a) high (red), (b) medium (black), and (c) low-efficiency (blue) events per decade between 20C and RCP8.5 runs. Dashed vertical lines indicate the ensemble mean of each distribution. Asterisks indicate where ensemble means are significantly different from 0 at the 95% confidence level.

Figure 4.5: (a) Winter mean surface heating efficiency ($E_{\text{trop}}$) in 20C (blue; %) and RCP8.5 runs (red; %) and the winter mean change in $E_{\text{trop}}$ (black; %). (b) Contribution to the winter mean change in the polar cap–averaged NSF from changes in NTES ($E_{\text{trop}}$ fixed at 20C values; blue; W m$^{-2}$), changes in $E_{\text{trop}}$ (NTES fixed at 20C values; black; W m$^{-2}$), and changes in both NTES and $E_{\text{trop}}$ (red; W m$^{-2}$). Asterisks indicate where ensemble winter mean changes are significantly different from 0 at the 95% confidence level.
Figure 4.6: Composite of the change (RCP8.5 − 20C) in the (a) time-integrated anomalous local MSE (red) and LH (blue) flux convergence (MJ m⁻² (100 hPa)^⁻¹) from day −14 to 0 in $F_{trop}$ events (combined high, medium, and low-efficiency events). (b) shows the change in mean lower-tropospheric stability (LTS; blue; K) between days −21 and −7. (c) shows the change in time-integrated anomalies in the NTES (gray; MJ m⁻²), net surface flux (NSF; light gray; MJ m⁻²), combined sensible and latent surface turbulent heat fluxes (SHLH; green, MJ m⁻²), net longwave radiation (red; MJ m⁻²), and surface cloud radiative effect (CRE; MJ m⁻²) from day −14 to −10. Violin plots follow the same convention as in Fig. 4.1. The shading in (a) indicates the ensemble 5th–95th percentile range. Asterisks indicate where ensemble mean changes are statistically different from 0 events at the 95% confidence level.
5.1 Key questions

The broad objective of this dissertation is to improve the understanding of the role of atmospheric circulations in winter Arctic warming. The following research questions were addressed in Chapters 2–4:

1. Is the vertically integrated tropospheric moist static energy (MSE) flux convergence \( F_{\text{trop}} \) a better metric than the total vertically integrated MSE flux convergence \( F_{\text{wall}} \) for the influence of atmospheric circulations on the Arctic surface climate?

2. What drives the surface heating efficiency of synoptic-scale \( F_{\text{trop}} \) events during winter?

3. What is the role of changes in the atmospheric energy transport and surface heating efficiency in future Arctic warming?

5.2 Summary of results

5.2.1 Chapter 2

In Chapter 2, the vertical structure of the polar cap–averaged MSE flux convergence and its seasonality were examined in the MERRA-2 reanalysis between 1980–2016, with emphasis on the Arctic winter (NDJFM). \( F_{\text{wall}} \) was separated into contributions from the stratosphere \( F_{\text{strat}} \) and troposphere \( F_{\text{trop}} \). To assess if \( F_{\text{trop}} \) and \( F_{\text{strat}} \) have different surface impacts in the Arctic winter, an Arctic energy budget analysis was applied to events associated with a poleward increase in \( F_{\text{strat}} \) [i.e., sudden stratospheric warmings (SSWs)] and events associated with a poleward increase in \( F_{\text{trop}} \). Correlations between \( F_{\text{trop}} \), \( F_{\text{wall}} \), and the polar cap–averaged lower-tropospheric sensible and latent energy (moist enthalpy; \( h_m \)) tendency were then compared.

In both hemispheres, the MSE flux convergence was found to be bimodal, with local maxima in the stratosphere and troposphere and is near-zero at the tropopause, consistent
with Overland and Turet (1994). This bimodal vertical structure is most evident during the
winter in the Northern Hemisphere. During the Arctic winter, $F_{\text{strat}}$ was found to be non-
negligible and explains 23% of the variance in $F_{\text{wall}}$. Importantly, $F_{\text{strat}}$ and $F_{\text{trop}}$ variability
are temporally distinct (uncorrelated at a lag of zero).

In Section 1.3 it was hypothesized that $F_{\text{trop}}$ is a better metric than $F_{\text{wall}}$ for the influence
of atmospheric circulations on the Arctic surface climate. This hypothesis is supported
by the results of Chapter 2. Composite analysis revealed that a majority of the energy input
from anomalous increase in $F_{\text{strat}}$ during SSWs went into stratospheric sensible energy
storage or was radiated to space, with little change to the net surface flux (NSF). In contrast, a majority of the energy input from an anomalous increase in $F_{\text{trop}}$ prior to periods of
downward anomalies in the NSF went into surface heating, with little indication of vertical
exchange of energy across the troposphere. Consistent with the composite analysis, $F_{\text{trop}}$
was found to be a better metric than $F_{\text{wall}}$ for the influence of atmospheric circulations on
the Arctic surface climate; $F_{\text{trop}}$ was found to explain 17% more of the variance in the $h_m$
tendency during winter.

5.2.2 Chapter 3

Having established that $F_{\text{strat}}$ variability has little influence on the surface energy bud-
get of the Arctic polar cap, Chapter 3 only considered the role of $F_{\text{trop}}$ variability on the
Arctic surface energy budget. Motivated by the potential for trends in synoptic-scale tropo-
spheric weather events (i.e., $F_{\text{trop}}$ events) to drive recent observed trends in Arctic surface
temperatures and sea ice (e.g., Woods and Caballero, 2016; Lee et al., 2017) during winter,
Chapter 3 investigated the potential role of trends in the surface heating efficiency of events
in recent Arctic temperature and sea ice trends.

Chapter 3 introduced a new metric for the Arctic surface heating efficiency ($E_{\text{trop}}$),
which measures the fraction of the anomalous $F_{\text{trop}}$—after accounting for the tropospheric
$h_m$ storage—that reaches the surface. $E_{\text{trop}}$ is calculated daily to ensure that it accounts
for synoptic variability. The $E_{\text{trop}}$ metric was used to categorize various synoptic-scale $F_{\text{trop}}$
events, identified in the MERRA-2 between 1980 and 2020, during winter. Composites of
high, medium, and low-efficiency were compared to understand the processes that determine
the $E_{\text{trop}}$ of an event. Trends in the relative frequency of high, medium, and low-efficiency
events were then calculated.

In Section 1.3, it was hypothesized that the $E_{\text{trop}}$ of $F_{\text{trop}}$ events is driven by the vertical structure of the anomalous MSE flux convergence; such that concentrating the MSE flux convergence closer to the surface increases the $E_{\text{trop}}$. This hypothesis is supported by the results of Chapter 3; however, it was found that the $E_{\text{trop}}$ of an event also depends on the preconditioning of the Arctic surface. High-efficiency events are associated with a “bottom-heavy” MSE flux convergence and reduced lower tropospheric stability (LTS) associated with reduced sea ice concentration in the Barents–Kara Seas. The implied enhanced turbulent mixing associated with reduced LTS enhances the near-surface warming. For example, the elevated anomalous warm and moist air is mixed to the surface, resulting in a prolonged period of high surface temperatures and reduced sea ice growth. Although anomalies in the downward longwave radiation (DLR) are a large contributor to the anomalous NSF in all three events [consistent with Woods and Caballero (2016)], differences in $E_{\text{trop}}$ are best explained by differences in downward sensible and latent turbulent heat flux (SHLH) anomalies. This highlights the importance of the vertical structure of atmospheric heating and moistening in driving $E_{\text{trop}}$.

These results suggest an underappreciated driver for the Arctic amplification of climate change. High-efficiency events were found to increase and low-efficiency events were found to decrease by approximately 20 events per decade, with little change in the total number of events. It was speculated that these trends result from a decrease in Arctic LTS and an increase in the frequency of bottom-heavy events. This result suggests a positive feedback loop, amplifying Arctic surface warming:

1) High-efficiency events preferentially occur when the Arctic is warm and unstratified.

2) A larger fraction of the energy input during an event is provided to the surface as a result of the unstratified conditions.

3) The warm and unstratified conditions are prolonged as a result of the suppressed upward heat loss.

In a traditional attribution study (e.g., Pithan and Mauritsen, 2014), this amplification would manifest as a positive lapse-rate feedback.
5.2.3 Chapter 4

Chapter 4 aimed to verify the results from the MERRA-2 (Chapter 3) on the drivers of $E_{trop}$ in $F_{trop}$ events in 20C runs of the CESM Large Ensemble (CESM-LE). Chapter 4 also sought to quantify the partial contributions to the NSF and Arctic surface temperature changes from changes in the winter mean $F_{trop}$ and $E_{trop}$ in the RCP8.5 warming scenario.

Chapter 4 first compares winter mean metrics for the atmospheric energy transport in the 20C runs of CESM-LE to the MERRA-2. Metrics include $F_{trop}$, $F_{wall}$, the net tropospheric energy source (NTES), and contributions to $F_{trop}$ from the latent (LH) and dry static energy (DSE). This comparison revealed that all terms are overestimated by 2–6 W m$^{-2}$ in the CESM-LE. High, medium, and low-efficiency events are then identified and are compared with the MERRA-2 results. In general, events are similar to the MERRA-2, with both the vertical structure of $F_{trop}$ and the initial LTS of the Arctic identified as important $E_{trop}$ drivers.

Winter mean changes in the aforementioned metrics and in $E_{trop}$ in the RCP8.5 warming scenario are computed. $F_{trop}$ and $F_{wall}$ are found to decrease by 9.5 W m$^{-2}$, dominated by a decrease in the DSE component and partially compensated by a small increase in the LH component. The winter mean $E_{trop}$ was found to increase by 5.7%, consistent with an increase in the frequency of high-efficiency events (13.4 more events per decade). Consistent with the results from the MERRA-2 in Chapter 3, the increase in $E_{trop}$ appears to be driven by a decrease in the LTS of the Arctic and indicates an increase in the coupling between the atmosphere and surface.

The third hypothesis in Section 1.3 stated that in a future warmer climate, an increase in $E_{trop}$ mitigates a decrease in $F_{trop}$. This hypothesis is supported by the results of Chapter 4. In a traditional TOA attribution study, a decrease in $F_{trop}$ indicates a vertically uniform cooling of the Arctic. However, the increase in $E_{trop}$ overcompensates the decrease in $F_{trop}$, resulting in a small positive—but statistically different from 0—contribution to the NSF (0.7 W m$^{-2}$). The contribution to the NSF from changes in $E_{trop}$ alone is an increase of 5.5 W m$^{-2}$, or approximately 2 K of surface heating. These results suggest that changes in the surface heating efficiency of $F_{trop}$ contributes to surface amplified warming of the Arctic, consistent with the conclusions from Chapter 3. Additionally, the increased frequency of high-efficiency events possibly contributes to decreased sea ice, indirectly contributing to the
ice-insulation feedback.

These findings are consistent with studies linking sea ice loss and the associated increase in upward surface heat loss during winter to the bottom-amplified Arctic warming and lapse-rate feedback (Deser et al., 2010; Dai et al., 2019; Feldl et al., 2020; Jenkins and Dai, 2021; Kaufman and Feldl, 2022). Additionally, this dissertation suggests a mechanism by which reduced sea ice and Arctic LTS drive an increase in the coupling between $F_{\text{trop}}$ and the surface energy budget (measured by $E_{\text{trop}}$), contributing to Arctic surface warming.

Although the third hypothesis in Section 1.3 is supported in the CESM-LE, changes in the surface heating efficiency likely do not mitigate changes in $F_{\text{trop}}$ in models where $F_{\text{trop}}$ increases. The results of Chapters 3 and 4 suggest that $E_{\text{trop}}$ increases in any model where the Arctic LTS decreases; thus, an increase in $E_{\text{trop}}$ would enhance the warming resulting from an increase $F_{\text{trop}}$.

5.3 Recommendations for future work

Changes in $E_{\text{trop}}$ and $F_{\text{trop}}$ need to be calculated in other climate models. This is needed to determine whether energy transport and surface heating efficiency changes enhance or dampen the inter-model spread in Arctic amplification. However, it is not common for data to be output at a sufficient temporal resolution to directly compute $F_{\text{trop}}$ and $E_{\text{trop}}$. Chapters 3 and 4 argue that calculations of $F_{\text{trop}}$ require relatively high-frequency data (e.g., daily) and that lower frequency data (e.g., monthly) may not capture synoptic variability. Comparisons between $E_{\text{trop}}$ calculated with various data frequencies should be made. Additionally, there exist biases in Arctic surface turbulent heat fluxes in observations and climate models when compared to satellite observations Taylor et al. (2018). The magnitude of these biases should be taken into account when comparing $E_{\text{trop}}$ between models.

The role of clouds in driving $E_{\text{trop}}$ changes needs to be carefully examined in climate models. Low cloud concentrations are expected to increase as a result of reduced lower-tropospheric stability (e.g., Yu et al., 2019). The role of the increased cloud cover on changes in $E_{\text{trop}}$ needs further study.

The findings in Chapters 3 and 4 need to be verified in other models, including carefully constructed mechanism-denial experiments. These experiments could hold either the vertical
profile of the MSE flux convergence or the Arctic stratification fixed in a single column model. The goal of the mechanism-denial experiments would be to determine the relative importance of changes in stratification and energy transport in driving surface heating efficiency and Arctic surface changes.
APPENDIX A

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